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SEMIEMPIRICAL MODEL OF THE THERMAL REGIME OF THE ATMOSPHERE AND REAL CLIMATE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 5-17

[Article by M. I. Budyko, Corresponding Member USSR Academy of Sciences, State Hydrological Institute, submitted for publication 31 August 1978]

Abstract: The article examines the results of application of a simplified model of the thermal regime of the atmosphere for clarifying the laws of genesis of climate and study of the physical mechanism of climatic changes.

[Text] Introduction. Simple semiempirical models of climatic theory were proposed about ten years ago. They describe the mean latitudinal distribution of air temperature at the earth's surface (Budyko, 1968; Sellers, 1969). Although the problem of the physical content of models of such a type was later discussed in many studies, up to the present time it cannot be considered adequately clarified. In this connection we will present a concise review of the principal results of application of the first of the above-mentioned semiempirical theories of the thermal regime and we will discuss the problem of the correspondence between these results and conditions of real climate. In this review we will limit ourselves to an examination of the most important conclusions.

1. A high sensitivity of the thermal regime to relatively small variations of the heat influx, especially to variations over long time intervals, during which there is an intensification of the feedback effect between air temperature at the earth's surface and the area of polar ice.
2. Indeterminacy of modern climate, that is, the possibility of the existence, with existing external climate-forming factors, not only of the observed meteorological regime, but at least one regime differing considerably from the modern regime.
3. The possibility of a total glaciation of the earth with a relatively small decrease in the heat influx.
4. The development, during the Cenozoic, of a cooling caused by a decrease in the quantity of carbon dioxide gas in the atmosphere.

5. The appearance in the Pleistocene of extensive glaciations whose cause was periodic decreases in radiation sums in the high latitudes during the warm periods of the year from variations of elements of the earth's orbit and inclination of the earth's axis.
6. Variations of mean air temperature in the modern era caused by changes in atmospheric transparency and increase in the carbon dioxide concentration.

Four of these six conclusions (1 and 4-6) can be checked directly on the basis of empirical data. For evaluation of the correctness of two conclusions (2-3) it is necessary to use indirect methods. Since the mentioned conclusions follow from computations based on a semiempirical model of the thermal regime of the atmosphere, whose results are dependent on the values of the empirical parameters entering into the used model, we will discuss the problem of determining these parameters.

Determination of Model Parameters

Structure of model of thermal regime. The mentioned model of thermal regime of the atmosphere (Budyko, 1968) is based on a solution of the heat balance equation for the earth-atmosphere system

$$Q(1-\alpha) = I + C. \tag{1}$$

Here Q is the radiation incident on the outer boundary of the atmosphere, α is system albedo, I is outgoing long-wave radiation, C is the receipts and losses of heat exchange when using the following empirical expressions in this solution:

$$I = A + BT, \tag{2}$$

where T is air temperature at the earth's surface, A and B are numerical coefficients dependent on cloud cover,

$$C = \beta(T - T_p), \tag{3}$$

where T_p is the mean temperature for the earth as a whole, β is a numerical coefficient,

$$\alpha = \alpha(T), \tag{4}$$

(the dependence of albedo on air temperature, used for taking into account the influence exerted on albedo by the formation of a snow-ice cover at low temperatures).

To these equations we add expressions following from (1)-(4),

$$Q_p\ (1 - \alpha_p) = A + bT_p, \tag{5}$$

$$\alpha_p = \alpha_p\ (T_p) \tag{6}$$

(the subscript p denotes values relating to the earth as a whole or one of the hemispheres).

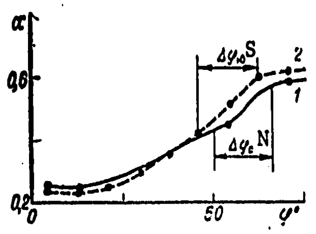
In the computations on the basis of these equations of the distribution of the mean latitudinal air temperatures the dependence of solar radiation on latitude was stipulated using tabular data.

A solution of equations (1)-(6) can be obtained using different approximate methods which have been discussed in a number of investigations, including studies by North (1975, 1975a).

The model presented above relates to a stationary case and can be used in determining the mean annual temperatures. For computations of the thermal regime under nonstationary conditions the equations (1) and (5) must include an additional term characterizing the change in heat content of the earth-atmosphere system with time. Such a method was used in a study by Budyko and Vasishcheva (1971) for determining the latitudinal distribution of temperature in the warm and cold seasons. Since the changes in heat content of the earth-atmosphere system in the annual variation are determined for the most part by fluctuations of temperature of the upper layer of ocean waters, in the mentioned study for taking this effect into account use was made of the equations for the heat balance of the oceans. The problem of computing the changes in the thermal regime with time when using a semiempirical model of the thermal regime was later examined in studies by Schneider and Gal-Chen (1973), Held and Suarez (1974), North and Koukli (1978) and other authors.

Albedo. The results of computations of the thermal regime of the atmosphere are essentially dependent on the assumed values of albedo of the earth-atmosphere system. In the investigation mentioned above use was made of two albedo values: for the region free of ice cover, 0.32, for the region of arctic polar ice 0.62 (Budyko, 1968). It was assumed that in the case of a change in the area of the polar ice the albedo in the zone occupied by it remains constant. In the discussion of this assumption it was noted that the albedo of the earth-atmosphere system is dependent on the angle of incidence of the sun's rays, decreasing with an increase in this angle. Such a dependence, however, is relatively weak for cases of high albedo values, characteristic for regions with an ice cover. At the same time, with an increase in the area of the ice zone a tendency arises for an increase in albedo due to a change in climatic conditions in the ice zone. It has been postulated that these factors more or less compensate one another, as a result of which it is possible to limit ourselves to a determination of the mean albedo value for the zone of the ice cover regardless of its extent (Budyko, 1969).

The actual albedo values for different latitude zones, obtained by Ellis and Vonder Haar (1976) on the basis of data from satellite observations, are represented in Fig. 1, where the curves α_N (1) and α_S (2) characterize the albedo values for the northern and southern hemispheres. The latitude intervals, denoted $\Delta \varphi_N$ and $\Delta \varphi_S$ in the figure, correspond to the regions of transition from an ice-free zone to a zone with a continuous ice cover in each of the hemispheres.



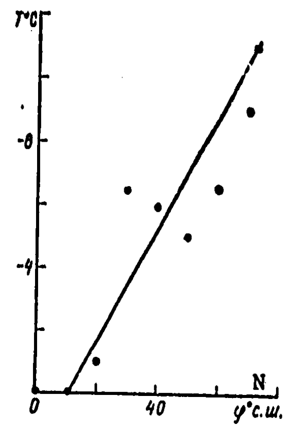


Fig. 1. Dependence of albedo on latitude. 1) Northern Hemisphere, 2) Southern Hemisphere

Fig. 2. Air temperature at boundary of ice cover.

The use of the data on albedo presented in Fig. 1 in computations of the present-day temperature distribution presents no difficulties. For estimating the albedo values with a change in ice area it is possible to take into account the relationship of the albedo values to the position of the ice, which is clarified when using these data. In both hemispheres, with a considerable difference in the area of the polar ice and in the nature of glaciations the albedo values in the zone of a permanent ice cover differ rather little and are close to 0.6 (in the southern hemisphere not much greater than this value, in the northern hemisphere not much less). In the transition zones, where the ice cover occupies only part of the space, the albedo decreases with increasing distance from the pole. These zones occupy about 20° of latitude in each of the hemispheres.

It can be postulated that with a change in ice area the albedo of the earth-atmosphere system will be 0.6 in the zone of continuous ice, whose boundary is situated at a distance of 10° in a poleward direction from the mean boundary of the ice cover, determined from its area.

In the transition zones, which extend 20° in latitude, the albedo can be determined by interpolation between the values for the zone of continuous ice and the value on the inner boundary of the transition zone under the assumption that the albedo does not change in the ice-free zone.

In order to validate the parameterization of albedo in the case of a variable ice area it is necessary to examine the dependence of the mean latitudinal albedo values on cloud cover. A study by Cess (1976) gave data on a substantial increase in cloud albedo with an increase in latitude due to a decrease in solar zenith angle, as a result of which the question arises: to what degree is the increase in albedo in the high latitudes dependent on the presence of the ice cover and to what degree on the conditions for the reflection of solar radiation on the upper surface of the clouds? In discussing this problem it must be remembered that stable regions of high pressure with a low air temperature and humidity form over quite extensive polar ice covers in the high latitudes and the cloud cover in these usually is small. Examples of such a type of climatic conditions are the central part of Antarctica and the regions of occurrence of Quaternary glaciations in the northern hemisphere. In this connection it can be surmised that with the movement of ice into the lower latitudes a decrease in the mean albedo of the zone of a continuous ice cover is improbable. It is possible that for a "white earth" with very low temperatures at all latitudes the atmosphere should be virtually transparent for radiation, as a result of which the earth's albedo would approach the albedo value for a pure snow surface (Budyko, 1974).

In computations of the latitudinal distribution of albedo it is necessary to take into account the position of the mean boundary of the permanent ice cover, which is dependent on thermal conditions. Since the temperature of the earth's surface usually differs little from the temperature of the lower air layer, it can be postulated that in the absence of an annual temperature variation the boundary of the permanent snow and ice cover will coincide on the continents with the isotherm 0°C. This assumption can be compared with data from computations of temperature at the boundary of the permanent snow and ice cover for different latitudes of the northern hemisphere, carried out using data on the elevation of the snow line in the mountains at these latitudes, taking into account the mean values of the vertical temperature gradients. The results of such computations are presented in Fig. 2. This figure shows that the mentioned temperature in actuality approaches zero in the equatorial zone. At the same time, in the higher latitudes the boundary of the snow cover corresponds to a negative mean annual air temperature, which in the northern hemisphere attains -11°C. Approximately the same mean annual temperature is observed on the boundary of the permanent ice cover in the high latitudes of the oceans.

Since the conditions of a continental climate predominate over the surface of an ice cover, it can be assumed that the mean annual air temperature at the boundary of the ice cover in different latitude zones corresponds to the similar temperature for the snow line on the continents, represented in Fig. 2.

Outgoing long-wave radiation and meridional heat exchange. The coefficients of the empirical formula for determining long-wave radiation (2) were initially determined using data from approximate computations of the mean monthly values of outgoing radiation in different geographic regions. Later it was possible to determine these coefficients on the basis of data from satellite observations of outgoing radiation (Cess, 1976). Similar computations were made by K. Ya. Vinnikov and I. M. Beyeva, who made use of more precise data on cloud cover for different latitude zones in comparison with the work of Cess. Taking the results of these computations into account, it is possible to represent formula (2) in the form

$$I = 16{,}2 + 0{,}106\,T - 4{,}75\,n, \tag{7}$$

where I is in Cal/(cm^2·sec), T is in degrees Celsius, n is cloud cover, expressed in fractions of unity.

Although the coefficients in formula (2) differ somewhat for the northern and southern hemispheres, this difference is not very great, as a result of which in formula (7) we used the mean values of the coefficients for the earth as a whole.

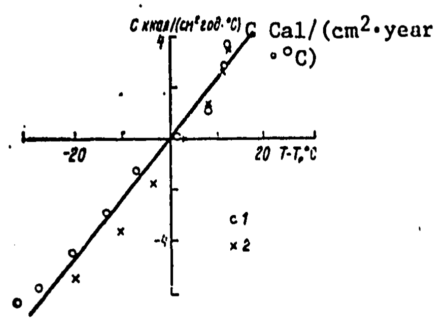


Fig. 3. Dependence of meridional redistribution of heat on temperature difference. 1) northern hemisphere, 2) southern hemisphere. A) C Cal/(cm^2·year·°C)

The β coefficient was computed using data on the relationship between the temperature difference for individual latitude zones and the earth as a whole ($T - T_p$) and the value of the meridional redistribution of heat C, which for mean annual conditions is equal to the radiation balance of the earth-atmosphere system, known from satellite observation data. The corresponding dependence using computed data for the C value when using formula (7) is shown in Fig. 3, where the points characterize the values for all 10° latitude zones, except the region of the Antarctic ice cover, rising high above sea level. This graph shows that the hypothesis of proportionality of the C values to the $T - T_p$ values is confirmed well by empirical data; the β coefficient is equal to 0.232 Cal/(cm^2·year·°C), which almost precisely coincides with the value obtained in the first investigation (Budyko, 1968).

Results of Application of Model

Distribution of mean latitudinal temperatures. Using the values of the parameters cited here, it is possible to compute the distribution of mean air temperatures at different latitudes. The results of such computations were presented in Fig. 4, where they are represented by curve 2. By comparing these results with the empirical data (curve 1) it can be found that the mean deviation of the measured and computed air temperatures is 1.2°C. This value is small in comparison with the interval of changes of mean latitudinal air temperatures, which indicates a satisfactory description of the conditions of real climate in the employed model.

Figure 4 does not show the mean air temperatures at the earth's surface in the central regions of Antarctica, where this temperature is considerably below the results of computations using the particular model of the thermal regime. The reason for such a difference is the great elevations of the central antarctic regions above sea level and the absence in this zone of heat transport by ocean currents.

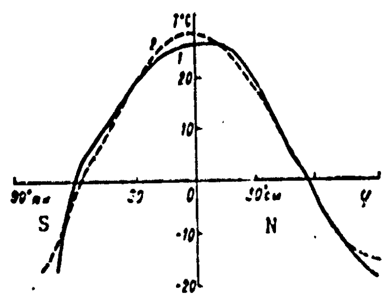


Fig. 4. Latitudinal distribution of mean air temperature.

Dependence of thermal regime on heat influx. The problem of the relationship between mean air temperature at the earth's surface and the heat influx, having a great importance for understanding the genesis of climate and its changes, has been discussed in many investigations. At the present time there are no less than four independent methods for determining the ΔT_1 parameter -- the difference in mean air temperature at the earth's surface with a change in the heat influx (solar constant) by 1%. One of these methods -- computations on the basis of parameterized models of thermal regime theory -- gives ΔT_1 values equal to 1.2°C (Manabe, Wetherald, 1967) and 1.5°C (Budyko, 1968).

It follows from computations using a model of general circulation of the atmosphere that $\Delta T_1 = 1.5$°C (Wetherald, Manabe, 1975). An analysis of empirical data on modern changes in climate gives $\Delta T_1 = 1.1$-1.2°C (Budyko, 1969, 1977). It was found on the basis of data from satellite observations of outgoing radiation that $\Delta T_1 = 1.1$-1.4°C (Budyko, 1975) and $\Delta T_1 = 1.45$°C (Cess, 1976).

Computations using the variant of the semiempirical model of the thermal regime of the atmosphere set forth here give a value $\Delta T_1 = 1.4°C$.

All these values relate to the case of absence of influence of a feedback between the thermal regime and the area of polar ice or to conditions of a relatively weak influence of this relationship. It follows from the data cited above that for this case the real ΔT_1 value falls in the range 1.1-1.5°C. It is probably closer to the greater of these values.

Investigations have been made in which the authors introduce an allowance for the feedback between air temperature and cloud cover in simplified models of the thermal regime of the atmosphere; as a result, the ΔT_1 value is appreciably reduced. As noted earlier, such models are obviously incorrect because the conclusions drawn from them contradict both the results of computations based on the most detailed models of the theory of climate (Manabe, 1976) and empirical analyses of the influence of cloud cover on the thermal regime, carried out on the basis of satellite observation data (Budyko, 1975; Cess, 1976).

In order to study the regularities in variation of climate in the geological past it is necessary to evaluate the dependence of mean air temperature on variations of the heat influx with a change in the area of the ice cover. in accordance with thermal conditions.

Since with an increase in ice area the earth's albedo increases, it is obvious that the $\Delta T'_1$ parameter relating to this case must be greater than the ΔT_1 value.

It is clear that the parameter $\Delta T'_1$ computed on the basis of a semiempirical model of the thermal regime must be dependent on the parameterization of the latitudinal distribution of albedo. For clarification of this dependence computations were carried out and the results are given in Table 1.

In this table α_1 corresponds to the albedo for an ice-free zone ($\alpha_1 = \alpha_1 (\varphi)$ denotes allowance for the albedo value at different latitudes on the basis of data from Ellis and Vonder Haar), α_2 is the albedo of the ice zone. It was assumed in variants A and B that the albedo at the ice boundary changes in a jump; in variants C and D it is assumed that the change in albedo occurs in a transition zone whose width is 20° in latitude. The ice boundary in variants A, B and C corresponds to a temperature $T_0 = -11°$; in variant D it varies with latitude in accordance with the dependence represented in Fig. 2.

The data cited in Table 1 were obtained in computations of outgoing radiation using formula (7). They correspond to the case of a decrease in the solar constant and relate to northern hemisphere conditions.

In a study by Lian and Cess (1977), with use of this same semiempirical model of the thermal regime of the atmosphere and determination of the dependence of albedo on the angle of incidence of the sun's rays, the authors found the value $\Delta T'_1 = 1.85°C$. This value is probably somewhat too low because in its determination, in particular, no allowance was made for the above-mentioned dependence of the albedo of the ice cover region on the extent of this region.

Table 1

Influence of Parameterization of Albedo on Response of Thermal Regime to Heat Influx

	Альбедо (1)		$\Delta T'_1$ °C	Δq_0%
A	А. $a_1 = 0,32$ (2) Без переходной зоны $T_0 = -11°C$	$a_2 = 0,62$	3,9	2,1
B	Б. $a_1 = a_1(\varphi)$ (2) Без переходной зоны $T_0 = -11°C$	$a_2 = 0,60$	1,90	4,3
		$a_2 = 0,55$	1,65	5,5
C	В. $a_1 = a_1(\varphi)$ (3) Переходная зона 20° $T_0 = -11°C$	$a_2 = 0,60$	1,73	7,3
		$a_2 = 0,55$	1,59	10,3
D	Г. $a_1 = a_1(\varphi)$ (3) Переходная зона 20° $T_0 = T_0(\varphi)$	$a_2 = 0,60$	3,3	3,2
		$a_2 = 0,55$	2,0	5,0

KEY:

1. Albedo
2. Without transition zone
3. Transition zone 20°

Due to the difficulty in determining the $\Delta T'_1$ parameter on the basis of model computations it is of interest to consider the possibility of evaluating it by an empirical method. For this purpose it is possible to compare data on change in mean air temperature in the northern hemisphere with variations of the carbon dioxide concentration in the modern epoch and in the Neogene (Budyko, 1977a). It follows from this comparison that $\Delta T'_{mean}/\Delta T_{mean} = 1.5$, where $\Delta T'_{mean}$ is the change in the mean air temperature as a result of doubling of the carbon dioxide concentration with a corresponding change in the area of the polar ice, ΔT_{mean} is the change in mean temperature as a result of doubling of the concentration of carbon dioxide in the case of a constant ice area. Assuming that $\Delta T'_1/\Delta T_1$ is also equal to 1.5, and assuming that $\Delta T_1 = 1.4\text{-}1.5°C$, we find $\Delta T'_1 = 2.2°C$, which agrees with the result of the above-cited model computations for the most realistic variant D.

It follows from the available evaluations of the ΔT_1 and $\Delta T'_1$ parameters that the mean temperature at the earth's surface varies in the case of variations in the heat influx to a considerably greater degree than the temperature of a black body deprived of an atmosphere, for which the value of the similar parameter is about 0.6°C. As demonstrated by our studies, without allowance for this effect it is impossible to explain the physical mechanism of changes in modern climate and the climates of the geological past.

Inhomogeneity of climate. The assumption of the possibility of the existence, with the present-day value of the solar constant, of a second variant of climate -- a "white earth," that is, a planet completely covered with ice, with very low temperatures at all latitudes, was initially expressed on the basis of simple physical considerations (Budyko, 1962). The conclusion that such a possibility exists was drawn on the basis of the first semiempirical models of the thermal regime (Budyko, 1968; Sellers, 1969), and then from a model of general circulation of the atmosphere (Wetherald, Manabe, 1975), and also from a series of other models in climatic theory.

Taking into account the results of these investigations, the conclusion that today's climate is uncertain must be considered extremely probable. However, the problem of the possibility of the existence of other variants of climate, other than today's climate and a "white earth," cannot be regarded as solved. From the equations of the considered model of the thermal regime it must be concluded that for the modern value of the solar constant there is a third solution, intermediate between a "white earth" and modern climate. This solution, however, corresponds to an unstable regime which cannot exist for a long time.

Using some semiempirical models of the thermal regime it was possible to obtain other variants of an uncertain climate, but a more detailed analysis of these models does not confirm the reality of such variants.

It is worthwhile to examine the problem of the possibility of the prolonged existence, under modern conditions, of an ice-free Arctic Ocean, which would correspond to still another variant of an uncertain climate. This problem, already formulated rather long ago (Budyko, 1961; Donn, Show, 1966), has still not been clarified adequately. Although a solution corresponding to an ice-free regime of the Arctic Ocean does not follow from the considered model, such a solution can be obtained with relatively small changes in the model, probably falling within the limits of computation accuracy. Attempts at the use of other semiempirical models of the thermal regime of the polar zone for study of this problem led to the conclusion that an ice-free regime in the Arctic Ocean is possible, but in the modern epoch would not be very stable, as a result of which its long-term persistence cannot be expected (Budyko, 1962, 1971). Since sea polar ice is very sensitive to relatively small climatic changes it is probable that it is difficult to solve this problem on the basis of application of present-day highly schematic models.

Conditions for total glaciation. In the first investigations of stability of modern climate, carried out with use of semiempirical models of the thermal regime, the conclusion was drawn that a total glaciation of the earth would occur with a decrease of the solar constant by a very small value -- about 2%.

Table 1 gives the results of computations of the relative decrease in the solar constant, leading to the earth's glaciation Δq_0, for different parameterizations of the latitudinal distribution of albedo. These values fall in the range 2-10%; the more probable values (variant D), equal to 3-5%, are close to the similar value found in computations using a model of general circulation of the atmosphere (Wetherald, Manabe, 1975).

The agreement of the critical values of the decrease in heat receipts at which the earth's glaciation occurs, found in the computations using different models, makes probable a correspondence between this conclusion and the conditions of real climate. However, it is necessary to point out the need for a further study of the considered problem, which includes inadequately clarified problems. For example, among these is the need for explaining the absence of traces of general glaciation of our planet during the Precambrian, when, in the opinion of a number of authors, the solar constant was appreciably less than its present-day value. Without question, at such a remote time the chemical composition of atmospheric air differed greatly from that of today. This difference, evidently, considerably intensified the greenhouse effect of the atmosphere. In addition, it is possible that in the considered epoch the quantity of water in the hydrosphere was inadequate for the forming of an ice cover over the entire planet. Although these factors probably explain the above-mentioned contradiction, there is obviously a need for a special investigation of this process.

The second problem requiring study is: to what extent is it possible, with the present-day structure of the earth's surface, that ice will appear which covers the entire planet? As indicated by paleoclimatic data, during the development of Quaternary glaciations over the parts of the continents which were free of ice there was development of extensive regions of inadequate moistening, which in some regions impeded the further advance of glaciers. It can be surmised that in the southern hemisphere, whose surface for the most part is covered with water, this mechanism cannot be of substantial importance and that with total glaciation of the southern hemisphere the retention of ice-free oases in some intracontinental regions of the northern hemisphere will be improbable due to the sharp reduction in mean air temperature at the earth's surface. Nevertheless, this problem must be investigated in greater detail.

It must be emphasized that the problem of the conditions for the appearance of total glaciation in no way coincides with the problem of uncertainty of present-day climate. Thus, in particular, the relatively simple validation of the possibility of existence of a "white earth" can be considered demonstrated despite the lack of clarity in study of the mechanism of transitions from one variant of an uncertain climate to another.

Cenozoic cooling. In proceeding to an examination of the results of application of the semiempirical model of the thermal regime for a quantitative explanation of the changes in climate, we will first discuss the cooling process which developed beginning at the end of the Mesozoic era and extending to the recent past. The most probable cause of this cooling was a decrease in the mass of atmospheric carbon dioxide, occurring over the extent of the last hundred million years (Budyko, 1977).

Using the dependence of mean air temperature at the earth's surface on the concentration of carbon dioxide, based on the semiempirical model of the thermal regime (Budyko, 1974), using data on the quantity of carbon dioxide for different parts of the Cenozoic it is possible to compute the mean air temperature for each part. The results of such computations (Budyko, 1977a) virtually coincide with data on mean air temperature in the northern hemisphere found from paleogeographic data, which indicates a satisfactory correspondence between the considered model and the climatic changes which transpired over the course of tens of millions of years.

Quaternary glaciations. Although the first of the semiempirical models of the thermal regime of the atmosphere was developed for explaining the physical mechanism of development of Quaternary glaciations (Budyko, 1968), however, it could not be used for this purpose, since it was found that in accordance with the Milankovich hypothesis (1930) glaciations are not dependent on the mean annual, but on the seasonal conditions of the thermal regime (Budyko, 1971). Only after creating a variant of a semiempirical model, making it possible to compute the distribution of mean latitudinal air temperature for the warm and cold seasons of the year, was it possible to employ this model for explaining the relationship between Quaternary glaciations and external climate-forming factors (Budyko, Vasishchev, 1971). It was established in that study that the variations in radiation values caused by changes in the elements of the earth's orbit and inclination of the earth's axis must lead to periodic movements of the ice covers from the polar into the middle latitudes by a distance close to the values established in paleogeographic investigations. At the same time, the computations have indicated that in the zone of ice distribution the air temperature was appreciably reduced, whereas outside this zone it changed very little: the mean air temperature for the earth as a whole in the epochs of maximum glaciations was reduced only by a value of about 1°C. In its time this conclusion caused doubts among specialists in the field of the paleoclimatology of the Pleistocene (they assumed that the glaciations lead to a great decrease in mean global temperature), but later this conclusion was fully confirmed by research results (GLIMAP, 1976) in which empirical temperature maps were constructed for the time of the last glaciation.

The conclusion of a decisive importance of astronomical factors for the development of Quaternary glaciations was recently also confirmed in an investigation of a group of authors (Hays, et al., 1976). They obtained data which were in very good agreement on the chronology of glaciations with the results of computations of variations in the receipt of radiation in latitudes where an ice cover exists.

Studies have been published in recent years in which a similar conclusion was drawn when using other parameterized theories of climate (Berger, 1973; Suarez, Held, 1977 and others). The above-mentioned investigations make it possible to consider the problem of the reason for the development of Quaternary glaciations to be clarified to a considerable degree.

We will note one perplexity arising in a study by Lindzen and Farrell (1977) in a discussion of this problem. Noting that according to data from modern empirical investigations the temperature in the tropics during the development of glaciations changed relatively little and assuming that computations in accordance with this model give considerable temperature variations in the low latitudes, Lindzen and Farrell proposed that changes be introduced into this model which would ensure a decrease in variations of the temperature in the tropics computed in accordance with the model for periods of glaciations. A secondary result of these changes was an increase in the critical value of the decrease in the heat influx with which there is a total glaciation of the earth.

The opinion of Lindzen and Farrell that on the basis of our model it must be concluded that there is a change in temperature at the equator by several degrees with a movement of the boundary of the ice cover from its present position at 72° to 60°N is based on the simple computations which they made, which give rise to two objections.

As noted above, on the basis of the astronomical theory of Quaternary glaciations it can be concluded that there is a correlation of these glaciations not with the mean annual, but with the seasonal conditions of the meteorological regime. Accordingly the Lindzen and Farrell model for mean annual conditions was not feasible in study of the dependence of the thermal regime on the area of glaciation.

The second and principal objection is that on the basis of the astronomical theory of glaciation it follows that with the decrease in radiation in the high latitudes characteristic for these glaciations there will inevitably be an increase in radiation in the low latitudes because the mean global radiation despite variations in the elements of the earth's orbit and inclination of the earth's axis remains constant. Since Lindzen and Farrell did not take this circumstance into account, the results of their calculations of the influence of glaciations on the thermal regime of the equatorial zone were incorrect. This invalidated the proposed modifications of the semiempirical model of climatic theory in their investigation.

Recent climatic changes. Observational data show that during the last century the mean air temperature at the earth's surface for the northern hemisphere varied in a range equal to 0.6°C. The hypothesis that these changes were caused by variations in the quantity of radiation reaching the lower layers of the atmosphere was confirmed as a result of computations of the secular variation of mean air temperature for the northern hemisphere in accordance with the semiempirical theory of the thermal regime of the atmosphere (Budyko, 1969, 1974). It was noted in subsequent studies that recent changes in mean air temperature for the most part are dependent on variations in the transparency of the atmosphere associated with changes in the mass of atmospheric aerosol (especially aerosol in the lower stratosphere, where the lifetime of its particles can attain many months) and on the increase in the concentration of atmospheric carbon dioxide, intensifying the greenhouse effect in the atmosphere (Budyko, 1972, 1977).

A similar point of view was expressed in a series of other recent investigations (Pollack, et al., 1975; Oliver, 1976; Karol', 1977, and others), in which the authors also made computations of the secular variation of mean air temperature at the earth's surface.

The results of the computations cited above agree well with observational data on air temperature, evidence of the possibility of a quantitative explanation of recent climatic changes on the basis of use of semiempirical models of the thermal regime of the atmosphere.

A fact worthy of attention is that these models satisfactorily describe not only the secular variation of the mean air temperature for the northern hemisphere, but also the peculiarities of secular variation in different latitude zones, in particular, an appreciable increase in temperature changes in the high latitudes in comparison with the low and middle latitudes (Budyko, 1977).

Summary

As can be seen from the above, the formulation of semiempirical models of the thermal regime of the atmosphere made it possible to proceed to a clarification of the regularities in the genesis of climate which had not been known on the basis of data from meteorological observations. In a number of cases such regularities were first established in investigations based on semiempirical theories and then were confirmed by observational data. Some of the conclusions drawn from semiempirical theories of climate were later found in investigations using the models of the theory of general circulation of the atmosphere. Among the results of use of semiempirical theories a result of particular importance is a clarification of the reasons for climatic changes in the geological past and in the modern epoch, which makes it possible to develop a method for predicting climatic changes.

It must be emphasized that the successful solution of a number of problems in climatic theory with the use of semiempirical models does not mean that it is possible to replace these models with more detailed climatic theories, and in particular, theories of general circulation of the atmosphere, which can answer many questions falling outside the limits of the content of semiempirical models. It can be surmised that for further progress in physical climatology it is desirable to carry out complex investigations of the genesis of climate based on joint use of the methods of empirical analysis, parameterized models of climatic theory (including semiempirical models) and detailed models of the theory of general circulation of the atmosphere and oceans.

The author expresses appreciation to M. A. Vasisncheva, who made a number of computations for this study.

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PRESENT STATUS OF THE METHOD FOR SEASONAL WEATHER FORECASTS AND PROSPECTS FOR ITS DEVELOPMENT

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[Article by Candidate of Geographical Sciences N. A. Aristov and Doctor of Geographical Sciences D. A. Ped', USSR Hydrometeorological Scientific Research Center, submitted for publication 14 July 1975]

Abstract: The authors have defined three stages in the development of synoptic and synoptic-statistical methods for seasonal weather forecasts in the Soviet Union (1917-1977). The basic directions for future research, directed to improving the official method for predicting weather and seasonal phenomena are defined. [This paper was presented at the anniversary session of the Scientific Council of the USSR Hydrometeorological Center on 11 October 1977.]

[Text] Weather forecasts for a season in advance are most important for the national economy of the country. Investigations for the purpose of creating a method for such forecasts began after the resolutions of the Second Meteorological Congress in Russia in 1909. At this congress representatives of the Ministries of Agriculture and Internal Affairs energetically raised the question before meteorologists about the creation of a method for seasonal weather forecasts, making it possible to predict droughts, which were responsible for great underharvests of grains in Russia and misery for the population. However, only after the Great October Socialist Revolution did favorable conditions develop for the development of research in this direction. Work on creation of a synoptic method for seasonal weather forecasts was headed by B. P. Mul'tanovskiy, under whose direction a great body of researchers on this problem worked.

As a result of study of peculiarities of atmospheric circulation, beginning in the winter of 1922/23, seasonal weather forecasts began to be issued sporadically. Although at that time meteorologists had extremely limited initial data and only at the earth's surface; nevertheless, it was possible to discover the existence of natural synoptic (n.s.) periods, n.s. seasons, n.s. regions, rhythmic activity in the atmosphere and the phases of macroprocesses. Despite the first successful forecasts, the author of the method and his successors met with enormous difficulties in solving this problem, although the selected direction has not caused doubts. A major service of B. P. Multanovskiy was the exposition of this method in the monograph OSNOVNYYE POLOZHENIYA SINOPTICHESKOGO METODA DOLGOSROCHNYKH PROGNOZOV POGODY (Fundamental Principles of the Synoptic Method for Long-Range Weather Forecasts), published in 1933.

The initial period of this work for the most part was characterized by intensive exploratory research, participating in which were B. P. Multanovskiy and his successors E. S. Lir, M. V. Loveyko, G. Ya. Vangengeym, S. T. Pagava, N. A. Shirkina, A. A. Rozhdestvenskiy, E. A. Isayev, T. A. Duletova, Ye. K. Anninskaya. Later these investigations were included in a book published in 1940 -- OSNOVY SINOPTICHESKOGO METODA DOLGOSROCHNYKH PROGNOZY POGODY (Principles of the Synoptic Method for Long-Range Weather Forecasting) (T. A. Duletova, S. T. Pagava, A. A. Rozhdestvenskiy, N. A. Shirkina). This gave a generalization of all attainments in the field of theory and practice of long-range weather forecasting during this period. It was a reference book for meteorologists working in this field and ended the first stage in development in this direction.

We note that the procedures presented in this monograph played a major role in preparing forecasts during the period of the Great Fatherland War (1941-1945). Guided by these methods, meteorologists carried out successful support of the command of the Red Army and Navy with forecasts.

The second stage in the development of methods for seasonal weather forecasts began during the post-war period and was characterized by the creation of more reliable routine procedures for their preparation. In this a major service was rendered by S. T. Pagava, who after B. P. Mul'tanovskiy, beginning in 1938, headed this direction. An improvement in the quality of forecasts at this time occurred as a result of the developing favorable objective conditions: the use of pressure pattern charts began, meteorological information began to be received from the entire northern hemisphere, there was a broadening of the front of research, in which new professional meteorologists were included.

The method for seasonal weather forecasts, created by S. T. Pagava and his successors in the late 1940's and then somewhat improved, was based primarily on allowance for the peculiarities of atmospheric circulation. Six n.s. seasons were defined, each of which is characterized by a definite predominance of a certain type of macroprocesses; in n.s. seasons about 75% or more identical processes are observed. However, atypical processes are also observed -- "disruptions," the first of which are used as the precursors for subsequent n.s. seasons. As a result, using the precursors it was possible to prepare, for a period of 1-3 months in advance, predictions of the seasonal air temperature anomaly and the quantity of precipitation, and also determine the prevailing type of synoptic processes for the coming seasons and construct the computed geopotential fields for the 500-mb surface.

However, the choice of precursors is the most complex and responsible stage in the S. T. Pagava method. Ususally for this purpose use is made of about seven different procedures; not one of them is universal and only in combination is it possible to obtain a correct determination of the precursor (n.s. period). Naturally, it would be desirable to have one method for ascertaining the precursor which would substantially facilitate work in preparing weather forecasts. Only recently work began on the quantitative (objective) procedures for establishing precursors.

As an example we will cite one of the principal procedures for determining a precursor. If in the first n.s. region we compare the H_{500} field on the basis of n.s. periods, characteristic of the current n.s. season, with any comparable process, the disruption (precursor) will occur in the following cases:

a) in a region where there is a predominance of cyclones (anticyclones) there is an anticyclone (cyclone) or a ridge (trough);

b) an anticyclone (cyclone) will be in the region of localization of troughs (ridges);

c) in the territory where there were ridges (troughs) a trough (ridge) appears.

It is evident that the disruption will appear in all cases when the high-altitude deformation fields of the precursor and the prevailing processes in the n.s. season are substantially different from one another.

Computations of the anticipated field $\tilde{H}^s_{500}$ are made using 126 points uniformly situated in the space of the first n.s. region, bounded by the meridians 50°W-80°E and on the south by the parallel 20°N. The value of the predicted geopotential is determined using the formula $\tilde{H}^s_{500} = \Delta \tilde{H}^s_{500} + \overline{H}^s_{500}$, where $\Delta \tilde{H}^s_{500} = (H^p_{500} + \delta \overline{H}_{500} - \overline{H}^s_{500})$, that is, $\Delta \tilde{H}^s_{500}$ is equal to the sum of the geopotential anomaly of the precursor $(H^p_{500} - \overline{H}^s_{500})$ and the long-term change in $\overline{H}_{500}$ from the initial n.s. season to the season for which the forecast is prepared $(\delta \overline{H}_{500})$. Here there is adherence to the condition that the $\Delta \tilde{H}^s_{500}$ value does not exceed half the amplitude of H500 for this season $(\leqslant 0.5 A_s)$. If it is found that $\Delta \tilde{H}^s_{500} > 0.5 A_s$, in the computations it is assumed only that $0.5 A_s$, but when $\Delta \tilde{H}^s_{500} \leqslant 0.5 A_s$ a value is obtained which was obtained by computations. Similarly a forecast is written for the season for relative geopotential $(\tilde{H}^{s500}_{1000})$.

The computed values obtained by the indicated method, after being plotted on a blank map and a corresponding analysis, give an idea about the anticipated geopotential field $(\tilde{H}^s_{500})$ and the deviations from the norm $\Delta \tilde{H}^s_{500}$ and $\Delta \tilde{H}^{s500}_{1000}$. They are used in a preparation of a forecast of a seasonal temperature anomaly $(\Delta \tilde{T})$ and the quantity of precipitation $(\Delta \tilde{R})$. It has been established that in most cases there is satisfaction of the following relationships between $\Delta\tilde{H}^{s}_{500}$ and $\Delta\tilde{T}$: if $\Delta\tilde{H}^{s}_{500} > 4$ gp dam, then $\Delta\tilde{T} > 0$; when $\Delta\tilde{H}^{s}_{500} < -4$ gp dam $\Delta\tilde{T} < 0$ and in cases $-4 < \Delta\tilde{H}^{s}_{500} < 4$ gp dam we will have the inequality $-1°C < \Delta\tilde{T} < 1°C$. In addition, for predicting seasonal $\Delta\tilde{T}$ we will use a series of rules taking into account the advection of temperature as a function of the relative positioning of the isohypses of the computed $\tilde{H}^{s}_{500}$ and normal $\overline{H}^{s}_{500}$ fields, and also $\Delta\tilde{H}^{500}_{1000}$ for the season for which the forecast is prepared.

The preparation of the forecast of the seasonal anomaly of the quantity of precipitation ($\Delta\tilde{R}$) is based on a determination of the peculiarities of circulation for the surface pressure field (from the composite kinematic chart) of the precursor and the field of computed $\Delta\tilde{H}_{500}$ for the natural synoptic season. We will be guided by the following considerations. If the region of the cyclonic (anticyclonic) pressure field of the precursor coincides with the computed negative (positive) $\Delta\tilde{H}^{s}_{500}$ values, in the corresponding regions a precipitation excess (deficit) is expected. With other combinations of the considered fields a precipitation norm (80-120%) is assumed.

In preparing $\Delta\tilde{T}$ and $\Delta\tilde{R}$ forecasts the onset of a new n.s. season is determined from the precursor, using rules based on the fact that it appears one or two months before the onsetting n.s. season. A statistical study of the time intervals ($\Delta\tau$) between the first days of the n.s. period of the precursor (t_{pre}) and the new n.s. season (t_{new}) enabled D. A. Ped' and N. A. Sevalkina to discover a good linear feedback. For most seasons it is $r = -0.7$- -0.9 and satisfies the inequality condition $r \geq 3\sigma_r$ (σ_r is the mean square error of the correlation coefficient). This served as a basis for writing the regression equation $\Delta\tau = At_{pre} + B$, by means of which on the basis of the computed parameters A and B, and also t_{pre} of the current year it is possible to compute $\Delta\tau$. Using this parameter we then determine the onset of the new n.s. season (day of month):

$$t_{new} = t_{pre} + \Delta\tau.$$

With refinement of the forecast for the n.s. season, using data for the first n.s. period typical for it we compute the mean seasonal field H^{s}_{500} using the formula

$$\tilde{H}^{c}_{500} = \overline{H}^{c}_{500} + r\frac{\sigma H^{c}}{\sigma H}(H_{500} - \overline{H}_{500}),$$

[c = s(easonal)] where H_{500} and $\overline{H}_{500}$ is the current and mean long-term H_{500} value at a particular point for the first n.s. period typical for the incipient season; $\tilde{H}^{s}_{500}$ and $\overline{H}^{s}_{500}$ is the anticipated and mean long-term H_{500} value at a particular point during the entire incipient n.s. season; r is the correlation coefficient between $\tilde{H}^{s}_{500}$ and H_{500}, σ is the mean square value for H^{s}_{500} and H_{500}.

Supplementing the official method for seasonal weather forecasts there are the individual methods of Ye. I. Borisova, L. I. Blyumina, A. L. Kats, M. Ya. Kist, D. A. Ped', Ye. I. Tsepkanova, V. G. Shishkov; for forecasting n.s. seasons for the European USSR extremal in temperature respects; for studying and predicting droughts and stable transitions of air temperature through 0, ±5°C in spring and autumn; for predicting precipitation excesses and deficits in n.s. seasons, etc. All these studies were directed to an improvement and detailing of the seasonal weather forecast.

The third stage in improvement and development of seasonal weather forecasts began in the very late 1960's. This is continuing to the present time. It is characterized by attempts to employ data on the thermal peculiarities of the underlying surface, primarily oceanic, and the parameters of stratospheric circulation when developing weather forecasting methods. Electronic computers and small calculators have come into extensive use.

During this period S. T. Pagava, D. A. Ped', N. A. Bagrov, N. M. Zakharova, N. A. Sevalkina, N. I. Zverev, Kh. Kh. Rafailova and others carried out investigations devoted to a physical validation of the existence of n.s. regions and n.s. seasons; use of water temperature of the North Atlantic for determining the precursors and formulating synoptic-statistical methods for predicting air temperature; use of stratospheric parameters for determining precursors, spring and autumn changes in circulation and preparation of synoptic-statistical methods for predicting air temperature and the quantity of precipitation; study and forecasting of droughts and excess moistening, and also large and extremal air temperature anomalies; prediction of stable transitions of air temperature through 0, ±5°C; analysis and prediction of inhomogeneous n.s. seasons; seeking of similarity parameters and their use in scientific and practical work; study of the rhythmic activity in the atmosphere; detection of dynamically active regions in the northern hemisphere for the purpose of formulating synoptic-statistical methods for seasonal air temperature forecasts. All these investigations entered as component parts into the arsenal of a unified method for seasonal weather forecasts.

Most of the mentioned studies were included in a monograph written by S. T. Pagava, N. A. Aristov, L. I. Blyumina, Z. L. Turketti, entitled OSNOVY SINOPTICHESKOGO METODA SEZONNYKH PROGNOZOV POGODY (Fundamentals of the Synoptic Method for Seasonal Weather Forecasts) (1966), being a guide for the preparation of seasonal weather forecasts. Since 1963 there has been regular issuance of official seasonal weather forecasts. Their success for air temperature for the parameter $\rho = 0.35$, and for the quantity of precipitation $P = 76\%$.

In 1977 the seasonal weather forecasts for spring, the first half of summer and the second half of summer, employing the adopted criteria, had good success: with respect to air temperature $\rho = 0.42$ with a relative error 0.7, and with respect to precipitation -- 75%. Spring was as expected and actually was warm over the greater part of the European USSR, in Western Siberia and Kazakhstan; the precipitation deficit in the eastern European USSR was also correctly predicted. The first half of summer, as expected, was warm in the European USSR, in Kazakhstan, in most regions of Central Asia and Western Siberia; there was also a good prediction of the precipitation deficit at the center and in the northeastern part of Kazakhstan and also a precipitation excess in the western Ukraine. In the second half of the summer there was expected to be a colder season than usual with a great quantity of precipitation in the Baltic republics, Belorussia, the northern half of the Ukraine, and a warm season with precipitation near the norm in the eastern European USSR, in the Urals, in Western Siberia, Central Asia and Western Kazakhstan.

We assume that the results make it possible to consider the employed method to be promising. Incidentally, according to the evaluation of the quality of the seasonal weather forecasts it is the best in the world. At the same time, it is entirely clear that in order to ensure the ever-increasing needs of the national economy the level of reliability of forecasts must be increased still more.

In order to improve the quality of seasonal forecasts and supply them to the national economy, in the future it is necessary to do the following: improve and objectivize the official method for making seasonal weather forecasts, bringing it to routine use in the eastern part of the USSR; seek new procedures for the preparation of seasonal weather forecasts, basing this on more physically sound hypotheses and concepts; develop methods for predicting the most important seasonal weather phenomena -- atmospheric droughts and excess moistening, stable transitions of air temperature through 0, ±5°C, periods of hot and cold weather, anomalies of the number of days with thaws and severe frosts.

For the long-range improvement of seasonal weather forecasts it is necessary: to apply hydrodynamic methods in order to create a reliable scheme for predicting the H_{500} field for a season (two months) and separately by months for 1-2 months in advance. This anticipated H_{500} field will be included as a component part in a complex scheme of a computation method for a seasonal weather forecast. If such forecasts were available, then in the first approximation there would be solution of the problem of predicting atmospheric droughts, there would be a substantial improvement in the quality of forecasts of anomalies of air temperature and the quantity of precipitation; it would be possible to clarify the role of heat exchange between the subtropical and polar latitudes in the formation of an anomalous type of weather in the temperate latitudes, heat and moisture exchange between the oceans and continents in the formation of dry and moist weather in the temperate latitudes; c) it would be possible to determine the interrelationship between the processes in the troposphere and stratosphere and stratospheric influences on weather formation in the temperate latitudes; d) it would be possible to carry out an analysis of the possibility of using the two-year cycle in preparing long-range weather forecasts; e) in computation schemes it would be possible to make a fuller allowance for the effect of the thermal peculiarities of the underlying surface in the preparation of seasonal weather forecasts.

In our opinion, the solution of these problems will lead to a substantial improvement in the servicing of national economic organizations in the country. The results of these investigations will serve as the basis for writing a Manual on the Preparation of Seasonal Weather Forecasts in 1980-1985.

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NUMERICAL MODEL OF LOCAL ATMOSPHERIC PROCESSES

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[Article by Doctor of Physical and Mathematical Sciences V. V. Penenko, Candidate of Physical and Mathematical Sciences A. Ye. Aloyan and G. L. Lazriyev, Computation Center Siberian Department USSR Academy of Sciences, submitted for publication 23 May 1978]

Abstract: The authors describe a numerical model of local atmospheric processes over an orographically and thermally inhomogeneous underlying surface in a quasistatic approximation. The interaction between the air mass and the underlying surface is described using a model of the quasistationary surface layer. The results of numerical experiments are given.

[Text] For the purpose of numerical modeling of local atmospheric processes and solution of some practical problems it is necessary to use models which satisfactorily describe the development of local processes in the mesometeorological boundary layer of the atmosphere over a thermally and orographically inhomogeneous underlying surface.

There are now several such models [6, 7, 9-12]. However, the assumptions made in these models limit their use for prognostic purposes.

This paper describes a numerical model of local processes in which the interaction between an air mass and the underlying surface is described using a model of the quasistationary surface layer. The basis used was the model from [6], and the surface layer equations are transformed in such a way that allowance for it in model [6] was accomplished by means of stipulation of the boundary condition at the lower computation level, which coincides with the upper boundary of the surface layer.

Formulation of Problem

We will assume that the boundary layer has the constant height H and that its lower part is a quasistationary surface layer with the height h = const. For describing the model we will use a system of equations from [6], supplementing it with the moisture transfer equation:

$$\frac{\partial u'}{\partial t} + \operatorname{div} \vec{u} u = -\frac{\partial \pi'}{\partial x} + \lambda \vartheta' \delta_x + l v' + \frac{\partial}{\partial z} \nu_u \frac{\partial u}{\partial z} + \Delta u, \quad (1)$$

$$\frac{\partial v'}{\partial t} + \operatorname{div} \vec{u} v = -\frac{\partial \pi'}{\partial y} + \lambda \vartheta' \delta_y - l u' + \frac{\partial}{\partial z} \nu_u \frac{\partial v}{\partial z} + \Delta v, \quad (2)$$

$$\frac{\partial \vartheta'}{\partial t} + \operatorname{div} \vec{u} \vartheta' = -S(w' + u'\delta_x + v'\delta_y) + \frac{\partial}{\partial z} \nu_\vartheta \frac{\partial \vartheta'}{\partial z} + \Delta \vartheta' - u'\Theta_x - v'\Theta_y + \frac{L_w}{c_p} \Phi + Q_r, \quad (3)$$

$$\frac{\partial q'}{\partial t} + \operatorname{div} \vec{u} q' = -\frac{\partial Q}{\partial z}(w' + u'\delta_x + v'\delta_y) + \frac{\partial}{\partial z} \nu_\vartheta \frac{\partial q'}{\partial z} + \Delta q' - u'Q_x - v'Q_y - \Phi, \quad (4)$$

$$\frac{\partial \pi'}{\partial z} = \lambda \vartheta' \quad \left(\Delta = \frac{\partial}{\partial x} \mu_1 \frac{\partial}{\partial x} + \frac{\partial}{\partial y} \mu_2 \frac{\partial}{\partial y}\right), \quad (5)$$

$$\frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} + \frac{\partial w'}{\partial z} = 0, \quad (6)$$

$$u = U + u', \; v = V + v', \; w = W + w', \; \vartheta = \Theta + \vartheta', \; q = Q + q', \; \pi = \Pi + \pi'.$$

Here we have used a curvilinear coordinate system (x, y, z); x, y are mutually orthogonal coordinates directed along the relief (x -- to the east, y -- to the north); $z = z_1 - \delta(x, y)$, where z_1 is elevation above sea level; $z_1 = \delta(x, y)$ describes relief which has first been smoothed in such a way that its characteristic horizontal scales are not less than 10 km and the slopes to the horizon are small ($\delta_x \ll 1$, $\delta_y \ll 1$); x_1, y_1, z_1 are Cartesian coordinates; $\vec{u}$ is the wind velocity vector, u, v, w are its components along the coordinate axes x, y, z.

The remaining notations are: t is time, λ, l, $S = \partial\Theta/\partial z$ are the convection, Coriolis and stratification parameters μ_1, μ_2 are the horizontal turbulence coefficients, ν_u, ν_ϑ are the vertical coefficients of turbulent exchange for momentum and heat, ϑ is potential temperature, q is specific humidity, π is a value proportional to pressure, U, V, W, Θ, Q, Π are the background values of the meteorological fields, u', v', w', ϑ', q', π' are the deviations from the background (background values of the fields of meteorological elements and S, μ_1, μ_2 are assumed to be known functions of the space coordinates and time), Θ_x, Θ_y and Q_x, Q_y are horizontal gradients of background potential temperature and background specific humidity, L_w is the latent heat of condensation, c_p is the specific heat capacity of air, Φ is the rate of formation of the liquid phase, Q_r is the radiation component of the heat influx.

Adhering to the Monin and Obukhov similarity theory, the equations for the surface layer are written in the following form:

$$\varkappa z \frac{\partial |V|}{\partial z} = u_* \varphi_u(\zeta), \quad z \frac{\partial p}{\partial z} = p_* \varphi_\theta(\zeta) \quad (p = \vartheta, q), \tag{7}$$

$$\varkappa |V| = u_* f_u(\zeta, \zeta_0), \quad p - p_0 = p_* f_\theta(\zeta, \zeta_0), \quad \zeta = \frac{z}{L}, \zeta_0 = \frac{z_0}{L}, \tag{8}$$

$$\nu_i = \frac{u_* \varkappa z}{\varphi_i(\zeta)}, \quad (\nu_i)_h = \frac{u_* \varkappa h}{\varphi_i(\zeta_h)} \quad (i = u, \vartheta), \quad \zeta_h = \frac{h}{L}, \quad L = \frac{u_*^2}{\lambda \varkappa^2 \vartheta_*}, \tag{9}$$

where $|V| = (u^2 + v^2)^{1/2}$ is the modulus of the velocity vector, u_* is friction velocity, ϑ_*, q_* are the scales of potential temperature and specific humidity, $\varkappa$ is the Karman constant, z_0 is the roughness parameter (the subscripts 0 and h denote the meteorological fields with $z = z_0$ and $z = h$ respectively), L is the length scale, ζ is dimensionless length, φ_i, f_i are continuous universal functions [3].

A numerical model of solution of the problem (1)-(6) is obtained in such a way that the inclusion of the surface layer into model [6] was reduced only to a change in the method for stipulating the boundary conditions with $z = h$ without a change in the structure of the computation algorithm itself. For this purpose the equations (6)-(8) are transformed in such a way that from them it is possible to obtain the corresponding conditions with $z = h$ [3].

Finally, the boundary conditions for system (1)-(6) assume the following form: $\frac{\partial u'}{\partial x} = \frac{\partial v'}{\partial x} = 0, \ \frac{\partial \vartheta'}{\partial x} = 0, \ \frac{\partial q'}{\partial x} = 0$

when $x = \pm X$, (10)

$$\frac{\partial u'}{\partial y} = \frac{\partial v'}{\partial y} = 0, \quad \frac{\partial \vartheta'}{\partial y} = 0, \quad \frac{\partial q'}{\partial y} = 0 \qquad \text{when } y = \pm Y, \tag{11}$$

$$u' = v' = 0, \quad \vartheta' = 0, \quad q' = 0, \quad \pi' = 0 \qquad \text{when } z = H, \tag{12}$$

$$w = 0, \quad h \frac{\partial}{\partial z} \left\{ \begin{matrix} u \\ v \end{matrix} \right. = a_u \left\{ \begin{matrix} u \\ v \end{matrix} \right., \quad h \frac{\partial p}{\partial z} = a_\theta (p - p_0) \qquad \text{when } z = h, \tag{13}$$

where $a_i = \varphi_i(\zeta_h)/f_i(\zeta_h, \zeta_0)$.

Conditions (10), (11) give one of the methods for closing the problem. However, to a higher degree they express the requirement of smoothness of perturbations in the neighborhood of the boundary of the region than the physical sense of the modeled processes. Therefore, together with conditions (10), (11) we will examine the conditions determining the outcome of the processes transpiring within the $\tilde{D}$ region in a background flow regime. These conditions can be written as follows:

$$u = U, \ v = V, \ \vartheta = \Theta, \ q = Q \qquad \text{when } x = \pm X, \ \text{when } y = \pm Y. \tag{10a}$$

We will assume the functions ϑ_0 and q_0 over water to be stipulated:

$$\vartheta_0 = f_0(t), \quad q_0 = 0{,}622\, E_0(\vartheta_0)/P, \tag{14}$$

where E_0 is the elasticity of saturation of water vapor at the temperature ϑ_0, P is atmospheric pressure, $f_0(t)$ is the temperature of the water surface, which is assumed to be a known function of horizontal coordinates and time.

Over land the temperature is determined from the heat balance equation and the function q_0 is computed using the formula

$$q_0 = 0{,}622\, \eta_0 E_0(\vartheta_0)/P, \tag{15}$$

where η_0 is relative humidity at the soil surface, which is assumed to be a known function of x, y and t.

The heat balance condition at the atmosphere-soil discontinuity has the form

$$G_0 - \rho c_p \left(\nu_\vartheta \frac{\partial \vartheta}{\partial z} \right)_0 - \rho L_w \left(\nu_\vartheta \frac{\partial q}{\partial z} \right)_0 = I_0 (1 - A) - F. \tag{16}$$

Here

$$G_0 = C_s \rho_s k_s \left(\frac{\partial T}{\partial z} \right)_0$$

is heat transfer through the soil surface, ρ_s, c_s, k_s, T are density, specific heat capacity, thermal diffusivity coefficient and absolute soil temperature, ρ is air density, I_0 is short-wave solar radiation, A is the albedo of the underlying surface, F is effective long-wave radiation.

The solution of equation (16) can be represented in the following form [7]:

$$\vartheta_0 = (\tilde{\vartheta}_0 + D_1 \vartheta_h)/(1 + D_1), \quad D_1 = A_1 F_1 c_p, \tag{17}$$

where

$$\begin{gathered} A_1 = \rho c_u c_\vartheta |V|_h, \quad F_1 = [B + A_1 \tilde{\mu} L_w + 4 (F/T_0)^{l-1}]^{-1}, \\ \tilde{\vartheta}_0 = \vartheta_0^{l-1} + F_1 \left[I_0 (1 - A) - F^{l-1} - B \left(0{,}414\, \tau_0^{l-1} + \sum_{m>1} k_m \tau_0^{l-m} \right) \right] \\ + (q_h - q_0^{l-1}) A_1 F_1 L_w, \\ \tau = T - \overline{T}, \; \tilde{\mu} = 0{,}622\, \eta_0 \left[\frac{dE_0(\vartheta_0)}{d\vartheta_0} \right]^{l-1}, \quad B = 2 \rho_s c_s \left(\frac{k_s}{\pi \Delta t} \right)^{1/2}. \end{gathered} \tag{18}$$

$$k_0 = 1, \quad k_m = (m+1)^{1/2} - 2 m^{1/2} + (m-1)^{1/2} \quad \text{when } m > 0,$$

c_u, c_ϑ are the friction and heat transfer coefficients, j is a time index, Δt is the time interval, τ_0^{j-m} is the deviation of soil surface temperature from its mean daily value $\overline{T}$ = const at the moment in time t = (j-m)Δt. An analysis of equation (18) indicated that when Δt = 20 min in order to attain a satisfactory accuracy in the sum it is sufficient to take the first four terms.

Substituting (17) into (13), we finally obtain the boundary condition for ϑ over the land

$$h\frac{\partial\vartheta}{\partial z}=\alpha_0\frac{\vartheta-\tilde{\vartheta}_0}{1+D_1} \quad \text{when } z = h. \tag{19}$$

As the initial conditions we will use the stipulated values of the perturbations u', v', w', ϑ', q' and τ_0^{j-m} (m = $\overline{1,5}$) at the moment in time t = 0. The determination of these perturbations on the basis of measurement data requires the development of special methods making it possible to describe correctly the characteristic scales of the processes to be modeled. In the numerical experiments, assuming the values of the background fields to be stipulated, as a simplification we will stipulate the initial conditions in the following form:

$$u'=v'=w'=0,\quad \vartheta'=0,\quad q'=0,\quad \tau_0^{j-m}=0\ (m=\overline{1,5}) \quad \text{when } t = 0. \tag{20}$$

In the computations of the surface characteristics we used the following empirical formulas:

1. Albrecht formula [1]

$$I_0=a_0\sin h_c-b_0\sqrt{\sin h_c},\quad I_0\geq 0, \tag{21}$$

$$\sin h_c=\sin\varphi\sin\psi-\cos\varphi\cos\psi\cos\left(\frac{t-12}{12}\pi\right),$$

[c = sun] where h_s is the solar zenith angle, φ is latitude of the place, ψ is solar declination, a_0 = 2 cal/(cm^2·min), b_0 = 0.3 cal/(cm^2·min). The time t here and henceforth is reckoned in hours from midnight.

2. Brent formula [5]

$$F=\sigma f_s T_0^4(a_3-b_3\sqrt{e}), \tag{22}$$

where σ is the Stefan-Boltzmann constant, f_s is the coefficient of soil grayness, a_3, b_3 are empirical constants, e is water vapor elasticity.

3. Magnus formula [5]

$$E_0(T)=6{,}11\exp[17{,}55(T-273{,}15)/(T-31{,}25)]. \tag{23}$$

4. Charnok formula [2] for determining the roughness parameter over water

$$z_0 = 0{,}035\ u_*^2/g, \tag{24}$$

where g is the acceleration of free falling.

Solution Method

The system of equations (1)-(6) with the boundary conditions (10)-(13) will be solved by the splitting method on the basis of physical processes [4]. As in [6], the solution in the time interval $t^j \leqslant t \leqslant t^{j+1}$ is obtained as a result of successive solution of two simpler systems.

a) A system of equations describing the transport of the substance along the trajectory and turbulent exchange,

$$\frac{\partial u}{\partial t} + \operatorname{div} \vec{u} u - \frac{\partial}{\partial z} \nu_u \frac{\partial u}{\partial z} - \Delta u - \lambda \theta' \delta_x = \frac{\partial U}{\partial t}, \tag{25}$$

$$\frac{\partial v}{\partial t} + \operatorname{div} \vec{u} v - \frac{\partial}{\partial z} \nu_u \frac{\partial v}{\partial z} - \Delta v - \lambda \theta' \delta_y = \frac{\partial V}{\partial t}, \tag{26}$$

$$\frac{\partial \theta'}{\partial t} + \operatorname{div} \vec{u} \theta' - \frac{\partial}{\partial z} \nu_\theta \frac{\partial \theta'}{\partial z} - \Delta \theta' + u (\Theta_x + S \delta_x) + v (\Theta_y + S \delta_y)$$
$$= U (\Theta_x + S \delta_x) + V (\Theta_y + S \delta_y), \tag{27}$$

which is solved with the following boundary conditions:

$$\frac{\partial u}{\partial x} = \frac{\partial v}{\partial x} = 0, \quad \frac{\partial \theta'}{\partial x} = 0 \qquad \text{with } x = \pm X, \tag{28}$$

$$\frac{\partial u}{\partial y} = \frac{\partial v}{\partial y} = 0, \quad \frac{\partial \theta'}{\partial y} = 0 \qquad \text{with } y = \pm Y, \tag{29}$$

$$u = U, \quad v = V, \quad \theta' = 0 \qquad \text{with } z = H, \tag{30}$$

$$h \frac{\partial}{\partial z} \begin{Bmatrix} u \\ v \end{Bmatrix} = a_u \begin{Bmatrix} u \\ v \end{Bmatrix}, \quad h \frac{\partial \theta'}{\partial z} = a_\theta \frac{\theta' + \Theta - \tilde{\theta}_n}{1 + D_1} - S \qquad \text{with } z = h. \tag{31}$$

b) The system of equations describing the process of initialization of meteorological fields, whose type and solution method was given in [6]. We note only that the initial condition for the system of initialization equations for meteorological fields is the solution of problem (25)-(31) at the time t^{j+1}. After the initialization stage, using the determined fields u^{j+1}, v^{j+1} we solve the equation for specific humidity.

In order to obtain finite-difference approximations we introduce in the field $\tilde{D}$ the grid D^h $\{(x_i, y_n, z_k),\ (i = \overline{1.19};\ n = \overline{1.19},\ k = \overline{1.14})\}$ with the intervals Δx_i, Δy_n, Δz_k respectively. Approximations in spatial variables are obtained by analogy with [6], whereas for integration of the transfer problem (25)-(31) in time we use the symmetrized splitting method, based on successive application of the Crank-Nicholson scheme [4].

Numerical Experiment

As an illustration we will cite an example of computations carried out using the model described above. The numerical experiment includes four variants of the computations and was organized in such a way that beginning with a relatively simple problem additional factors are gradually introduced into the model. This makes it possible to examine the influence of different factors on the development of processes in the boundary layer. We have limited ourselves to solutions governed by periodic changes of the external parameters with time, such as the influx of heat from the sun, relative humidity at the earth's surface and the background wind.

In all the numerical computations presented below we have used a model with the following values of the input parameters: $X = Y = 85$ km, $h = 50$ m, $H = 2{,}050$ m, $\Delta x = \Delta y = 10$ km, $\Delta z = \{100$ m, if $z \leqslant 300$ m; 150 m, if 300 m $z \leqslant 750$ m; 200 m, if $750 < z < 2150\}$; $\lambda = 0.035$ m/(sec$^2\cdot$°C), $l = 10^{-4}$ sec^{-1}, $\mu_1 = \mu_2 = 10^4$ m^2/sec, $S = 3\cdot10^{-3}$ °C/m, $z_0 = 0.01$ m, $\varkappa = 0.35$, $\rho = 1300$ g/m^3, $P = 1000$ mb, $c_p = 0.24$ cal/(g$\cdot$(°C), $L_w = 530$ cal/g, $A = 0.3$, $\rho_s c_s = 44\cdot10^4$ cal/(m$^3\cdot$°C), $k_s = 3\cdot10^{-7}$ m^2/sec, $g = 9.8$ m/sec^2, $f_s = 0.9$, $a_3 = 0.56$, $b_3 = 0.08$ (in variant II $a_3 = 0.2$, $b_3 = 0$), $\varphi = 42.5°$, $\psi = 11°$.

$$f_0(t) = \left\{300 + \cos\left[\frac{\pi}{12}(t-22)\right]\right\} \text{°C}.$$

For closing system (1)-(6), for the sake of simplicity, the vertical coefficients of turbulence in the interval $h \leqslant z \leqslant H$ were stipulated in the form of linear functions decreasing with altitude from $(\nu_i)_h$ to zero.

The relative humidity at the soil surface (in %) was also stipulated as in [8]

$$\eta_0 = 59 + 22\cos\alpha + 11\sin\alpha - 3\cos 2\alpha + 4\sin 2\alpha, \quad \alpha = \pi(t-4)/12.$$

The background temperature and specific humidity fields were stipulated in the following way:

$$\Theta = \overline{\Theta} + Sz_1, \; \Theta_x = \Theta_y = 0, \quad Q = \overline{Q}\exp(-\beta z_1), \quad Q_x = Q_y = 0,$$

$$\frac{\partial Q}{\partial z} = -\beta Q,$$

where $\overline{\Theta} = 300$ K, $\overline{Q} = 12$ g/kg, $\beta = 6\cdot10^{-5}$ m^{-1}. In the computations it was assumed that $T = \Theta_0$, $Q_r = 0$, $W = 0$, $\Phi = 0$.

Variant I. We will examine a mesometeorological process which develops in the absence of a background wind over a round island with a radius of 30 km without allowance for humidity processes in the atmosphere and moisture content processes in the soil. At sunrise the island begins to be heated; a temperature gradient is formed between the land and sea which favors the appearance of a breeze circulation whose intensity attains a maximum at 1400 hours (1540 hours). [The corresponding values for variant II are given in parentheses.]

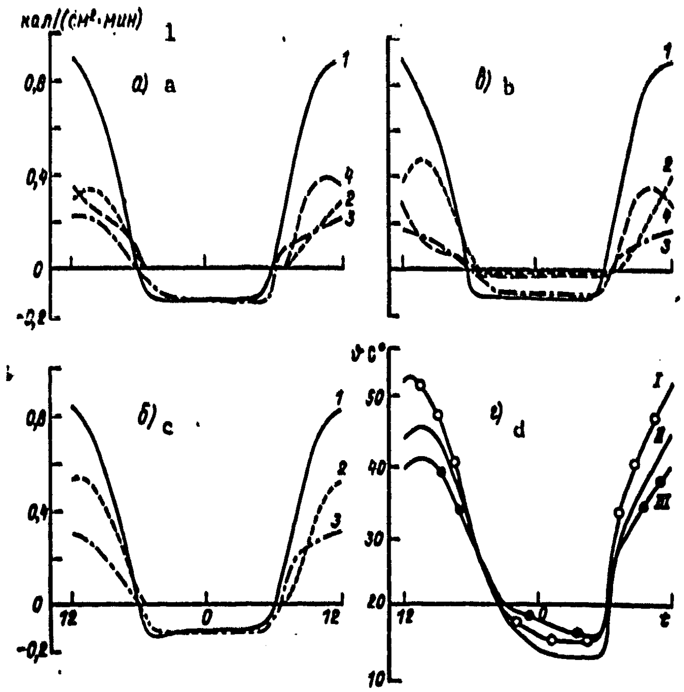

Fig. 1. Diurnal variation of components of the heat balance and temperature of the underlying surface at the center of an island. The curves correspond to:

$$1)\ I_0(1-A)-F,\quad 2)\ -\rho c_p\left(k_\vartheta \frac{\partial \vartheta}{\partial z}\right)_0,\quad 3)\ Q_0,\quad 4)\ -\rho L_w\left(k_q \frac{\partial q}{\partial z}\right)_0.$$

KEY:

1. cal/(cm^2·min)

The maximum of the wind velocity modulus of 9 m/sec (7 m/sec) is obtained at a height of 100 m 20 km from the center of the island. The ascending currents, which are an order of magnitude greater than the descending currents, attain a maximum value of 29 cm/sec (20 cm/sec) at the center of the island at a height of 600 m. The thickness of the sea breeze is 1,400 m (1,200 m); aloft there is an antibreeze with a maximum velocity of 2.6 m/sec (2 m/sec) at a height of 1,700 m (1,500 m). In the evening the sea breeze dies down and then a shore breeze begins to blow which is weaker than the sea breeze. By 2400 hours the intensity of the shore breeze attains a maximum. The greatest velocity of 2.2 m/sec (2.5 m/sec) is observed at a height of 100 m near the shore over the water. The descending currents are weak (-2 cm/sec at the center of the island). The thickness of the shore breeze is 600 m; aloft there is a weak antibreeze with a velocity of 0.5 m/sec.

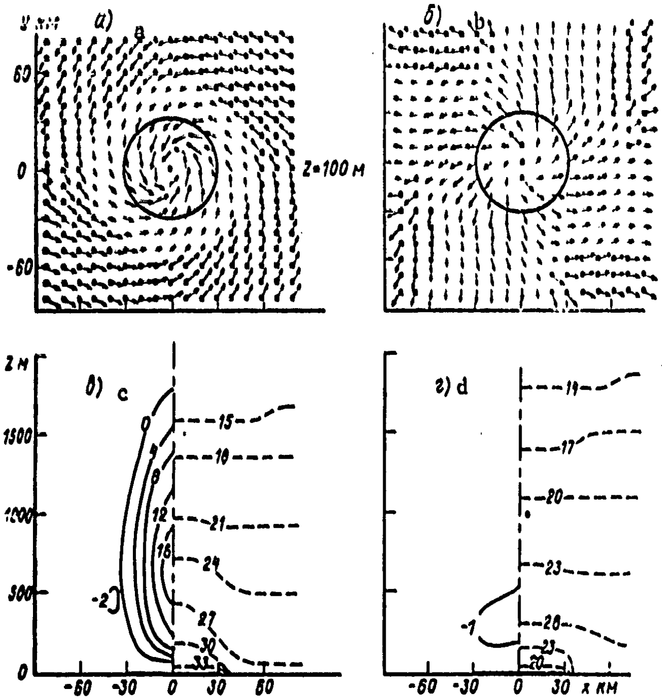

Fig. 2. Pattern of breeze circulation for 1600 hours (a,c) and for 0000 hours (b,d) in sections z = 100 m and y = 90 km in absence of background wind.

Variant II. In contrast to variant I, we took into account humidity, which considerably changed the pattern of development of breeze circulation over the island. This change is evidently associated with the diurnal variation of temperature of the earth's surface with given mesoprocesses.

Figure 1 gives the computed temporal variation of the components of the heat balance and temperature of the earth's surface with humidity (Fig. 1a) and without humidity (Fig. 1b) at the center of the island, where due to axial symmetry the wind velocity is equal to zero.

During the daytime, when a considerable part of the heat is expended on evaporation, soil with moisture is heated to a lesser extent than absolutely dry soil, as a result of which, in comparison with variant I, there is a decrease in the turbulent flow of heat into the atmosphere and the breeze circulation develops less intensively.

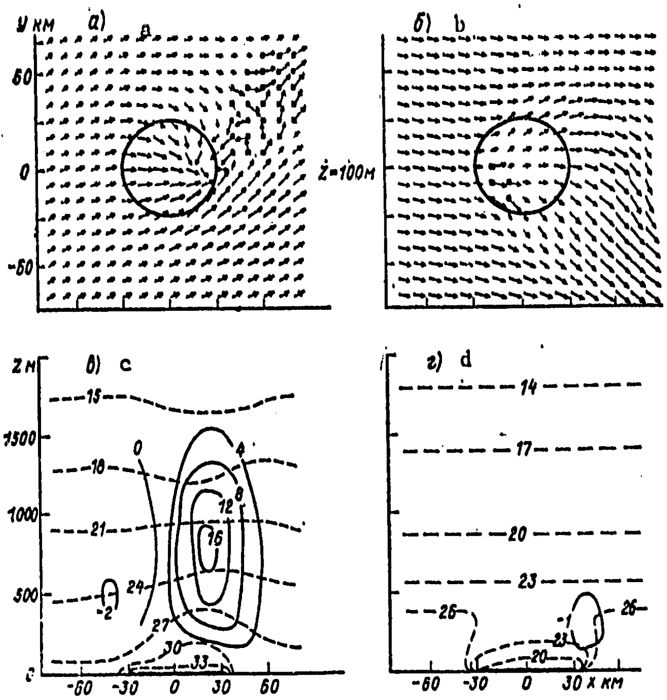

Fig. 3. Pattern of breeze circulation for 1600 hours (a,c) and for 0000 hours (b,d) in sections z = 100 m and y = 90 km with allowance for external background wind.

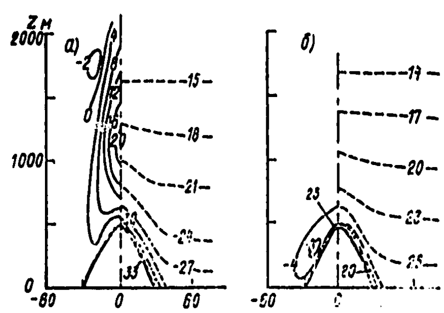

Fig. 4. Distribution of vertical currents (cm/sec) and temperature (0°C) with allowance for relief in the absence of the background wind for 1600 hours (a) and for 0000 hours (b) in section y = 90 km.

At nighttime the turbulent flows of heat and moisture are equal to zero; only two components, G_0 and F, remain in the heat balance and if it is taken into account that the constants a_3 and b_3 in the Brent formula were selected in such a way that the radiation balances in variants I and II were found to be approximately identical, it becomes clear why at nighttime processes with and without account for humidity transpire virtually identically.

Figure 2 gives some results of computations for 1600 hours (left half of the figure) and for 0000 hours (right half of figure) in the sections z = 100 m and y = 90 km. The wind fields are represented by arrows (the arrows with the square ends correspond to small velocities). The contour of the island is surrounded with a thick line; the isolines of vertical velocities, in cm/sec, are represented by solid lines; the temperature isolines, in °C, are represented by a dashed line.

Variant III. In comparison with variant II, we have introduced the external wind, dependent on time, which in the interval 3 hours $\leqslant t \leqslant$ 15 hours is given by the expression

$$U(t) = \left[3 + 5 \sin\left(\frac{t-12}{12}\pi\right)\right] \text{ m/sec}, \quad V(t) = 0,$$

whereas at the remaining moments in time we assume to be constant U(t) = 3 m/sec, V(t) = 0.

During the daytime over the western part of the islar' to a height of 1,200 m the local circulation and the external flow are directed in one direction, which leads to an intensification of the wind, whereas over the eastern part of the island, on the other hand, the wind attenuates. By 1600 hours, when the background wind lessens to 3 m/sec, and the local circulation attains a maximum, currents appear which are directed opposite the oncoming flow; near the eastern shore a convergence zone is formed, in whose region the vertical currents and temperature assume extremal values. Above 1,200 m there is no wind over the island.

At nighttime the direction of the local circulation changes to the opposite in comparison with daytime. For this reason in the lower layers the flow, approaching the island, is slowed down and seemingly flows around it on two sides. Beyond the island the wind is deflected to the right and with increasing distance from the shore again begins to intensify. In the upper part of the mesometeorological boundary layer the perturbations of the main flow are insignificant.

Both during the daytime and during the nighttime the fields of vertical velocities and temperatures are displaced in the direction of the external wind.

Figure 3 illustrates the results of computations similar to those cited in Fig. 2.

Now we will examine the influence of the external wind on the diurnal variation of the components of the thermal balance and temperature of the earth's surface at the center of the island (Fig. 1c). During the daytime, when due to the wind there is an intensification of turbulent mixing, the temperature maximum of the earth's surface is 4°C lower than in variant II. At nighttime, also due to the external wind, turbulent flow of heat and moisture appear, which, to be sure, assume small values, but in the course of the entire night do not change sign and therefore the minimum ϑ_0 value is 3°C higher than without a wind.

Variant IV. The influence of relief on the development of breeze circulation was taken into account. In comparison with variant I an axially symmetric mountain with an elevation of 512 m is introduced on the island; its surface is described by the expression $\delta(r) = (212 + 300 \cos (\pi r/40 \text{ km}))$ m, where r is the distance from the center of the island.

The course of intensity of the process with time approximately coincides with variant II. There are some differences which are entirely caused by relief.

During the daytime the intensity of circulation changes little. The wind velocity increases by only 1 m/sec. Ascending currents, whose region narrows with altitude, attain 23 cm/sec at the center of the city. Descending currents assume the same values as in variant II, but in contrast to a flat island they are situated directly over the land.

At nighttime the influence of the mountain is more significant. The intensity of the shore breeze is almost doubled, at midnight attaining 4 m/sec. The descending currents are intensified to -8 cm/sec.

Figure 4 shows isolines of vertical velocities and temperature for 1600 hours (Fig. 4a) and for 0000 hours (Fig. 4b) in the section y = 90 km. The notations are the same as in Fig. 2.

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SPATIAL-TEMPORAL STATISTICAL STRUCTURE OF MEAN MONTHLY GEOPOTENTIAL OF THE 500-mb SURFACE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 35-43

[Article by Professor G. V. Gruza, Doctor of Physical and Mathematical Sciences I. I. Polyak and Candidate of Physical and Mathematical Sciences V. A. Shakhmeyster, All-Union Scientific Research Institute of Hydrometeorological Information-World Data Center and Main Geophysical Observatory, submitted for publication 31 August 1978]

Abstract: For a period of 28 years from 1948 to 1975 the authors evaluate the periodograms and spectra of one-dimensional time series with a month-long time resolution for the mean latitudinal mean monthly geopotential of the 500-mb isobaric surface. The article gives evaluations of the annual variation of norms, standard deviations and also two-dimensional correlation and spectral functions. The possible long-term trend of the time series is analyzed.

[Text] In studying the most general patterns of atmospheric processes over a considerable territory and for sufficiently long time intervals it is customary, in particular, to use zonally averaged circulation characteristics [8].

A number of evaluations of the characteristics of the state and circulation of the atmosphere, averaged by circles of latitude, are given in [1-3]. In article [3] an attempt was made to investigate the spatial statistical structure of the integral characteristics of general circulation of the atmosphere and their representation by means of natural orthogonal functions. All these evaluations were made on the basis of relatively short samples of observations.

Now it is possible to analyze data for the free atmosphere for periods of 25-30 years. This makes it possible to approach the problem of evaluating the structure of modern climate of the free atmosphere and the patterns of climatic change.

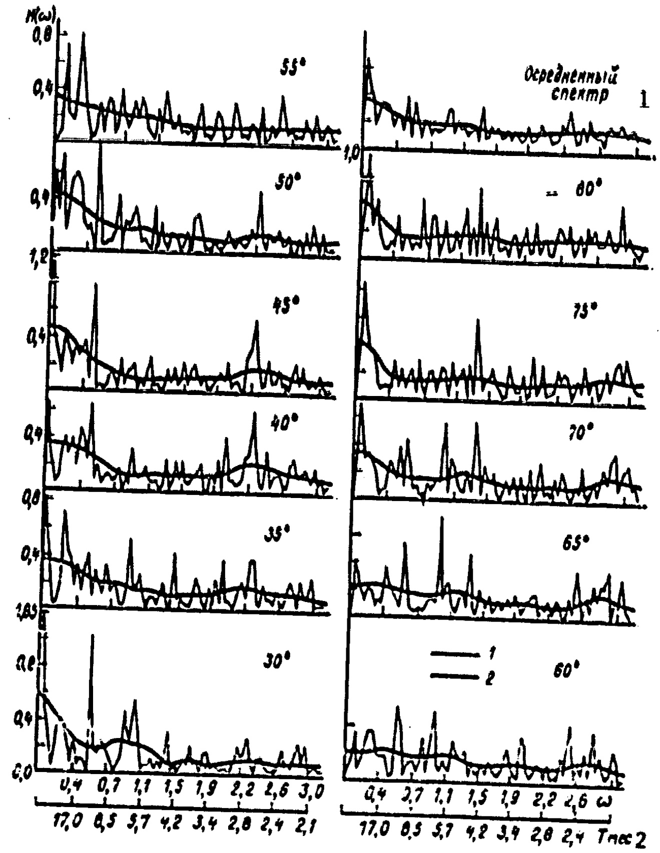

Fig. 1. Periodograms (1) and spectra (2) of time series of normalized geopotential anomalies AT_{500} for different latitudes and averaged by latitudes.

KEY:

1. Averaged spectrum
2. T months

As the first step it is natural to study long-period oscillations of the middle-latitude mean monthly geopotential values for the 500-mb isobaric surface in the northern hemisphere. For this purpose we used data for 1948-1975 at the points of intersection of a regular coordinate grid with a 5° latitude interval and a 10° longitude interval, visually interpolated from daily pressure pattern charts (AT_{500}). The data averaged for each month were then averaged by longitude (36 points on each circle of latitude). These data for 11 latitudes (30, 35,...,80°) and 336 readings in time (12 months x 28 years) were used for a statistical analysis in this study. The statistical structure of the initial daily geopotential fields was studied earlier in [4].

Table 1

Annual Variation of Middle-Latitude Geopotential Norms AT_{500} (dam)

Широта, град 1	Январь 2	Февраль 3	Март 4	Апрель 5	Май 6	Июнь 7	Июль 8	Август 9	Сентябрь 10	Октябрь 11	Ноябрь 12	Декабрь 13
80	507,89	508,93	511,17	518,79	531,71	542,46	547,87	545,23	533,94	519,91	514,92	509,60
75	510,18	511,83	513,91	521,30	534,35	544,88	551,57	548,31	537,17	523,37	516,88	512,33
70	513,72	515,16	517,90	525,32	537,81	548,34	555,48	552,21	541,33	527,98	520,02	515,53
65	517,37	518,81	522,49	530,22	541,51	551,89	558,78	555,93	545,74	533,05	524,10	519,21
60	521,93	522,91	527,37	535,39	545,26	555,15	561,48	559,42	550,54	538,91	529,41	523,95
55	527,70	528,00	532,73	541,01	549,76	559,15	565,20	561,15	556,61	546,38	536,32	530,18
50	534,95	534,75	538,98	547,72	556,14	561,67	570,67	570,45	564,34	555,56	545,41	538,26
45	543,93	543,35	546,63	555,21	563,37	571,10	577,12	577,39	572,37	561,95	555,34	547,96
40	552,99	552,28	554,96	562,62	570,11	576,98	582,79	583,37	579,21	572,61	564,05	557,21
35	562,68	562,04	561,22	570,28	576,41	582,36	587,01	587,43	584,39	578,91	572,23	566,56
30	571,95	571,40	572,93	576,96	581,66	586,76	589,71	589,83	587,75	583,72	578,96	574,90

Table 2

Annual Variation of Standard Deviations of Middle-Latitude Mean Monthly Geopotential AT_{500} (dam)

Широта, град 1	Январь 2	Февраль 3	Март 4	Апрель 5	Май 6	Июнь 7	Июль 8	Август 9	Сентябрь 10	Октябрь 11	Ноябрь 12	Декабрь 13
80	6,13	5,59	6,55	3,89	4,04	3,40	3,01	3,38	3,32	2,79	4,27	4,18
75	5,09	4,08	4,91	3,10	3,24	2,20	2,10	2,42	2,40	2,41	3,55	3,82
70	4,09	3,31	3,81	2,54	2,57	1,72	1,90	1,75	2,04	2,10	2,77	3,53
65	3,26	3,13	2,90	2,11	2,18	1,54	1,58	1,55	1,95	1,58	2,01	3,00
60	2,63	2,80	2,10	1,86	2,05	1,43	1,29	1,38	1,62	1,27	1,87	2,31
55	2,14	2,13	1,69	1,65	1,89	1,18	1,33	1,21	1,14	1,63	2,04	1,80
50	2,23	1,64	1,83	1,64	1,50	1,17	1,44	1,10	0,97	1,96	1,89	[illegible]
45	2,30	2,05	1,96	1,72	1,39	1,15	1,15	1,10	1,01	1,72	1,70	1,93
40	1,90	2,14	1,72	1,56	1,25	0,87	0,72	0,95	0,90	1,06	1,09	[illegible]
35	1,60	1,93	1,39	1,17	1,01	0,76	0,67	0,82	0,82	0,84	0,74	1,10
30	1,58	1,70	1,31	1,15	0,99	0,99	1,00	0,98	0,90	0,92	0,91	1,15

KEYS TO TABLES 1 and 2

1. Latitude, degrees
2. January
3. February
4. March
5. April
6. May
7. June
8. July
9. August
10. September
11. October
12. November
13. December

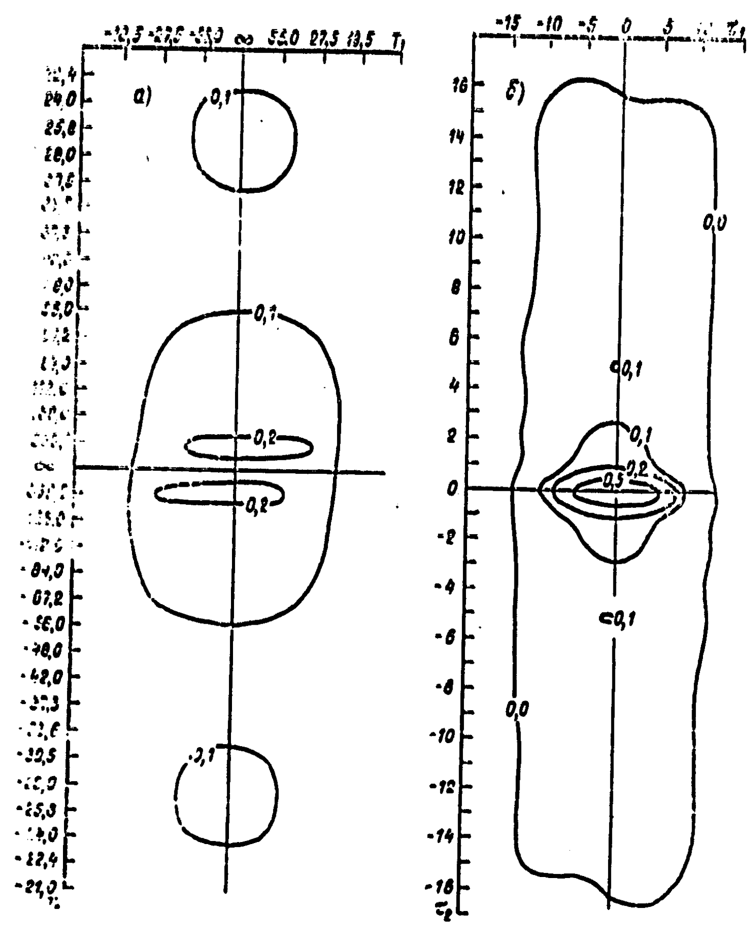

Fig. 2. Evaluation of two-dimensional spectrum (a) and two-dimensional correlation function (b) of changes in mean monthly middle-latitude geopotential AT_{500}. T_1 wavelength in space (degrees of latitude), T_2 -- period (months), τ_1 spatial shift (degrees of latitude), τ_2 time shift (months).

Table 1 gives evaluations of the mean monthly norms for each latitude zone. Similar results were given in [1, 2, 5, 9]. The results cited in [5], determined using data for a five-year observation period (1958-1963), only in individual cases differ (by 3-5 dam) from the corresponding values in Table 1. Most of the values differ by not more than 1 dam.

The amplitude of the annual variation varies in a wide range and the amplitude increases sharply with transition from the subtropics to the temperate latitudes, attains a maximum at latitude 70° and slowly decreases poleward. According to the data in the table, it is possible to trace as well the earlier onset of the extrema of the annual variation in the high latitudes, described in [1, 2].

Table 2 gives estimates of the mean monthly values of the standard deviations of middle-latitude mean monthly geopotential values AT_{500}. The same as the norms, the standard deviations have an annual variation with a minimum in the warm season of the year and a maximum in the cold season. Considerably greater standard deviations correspond to northern latitudes than to southern latitudes.

Now we will carry out a spectral and correlation analysis of the considered observations [6]. In particular, we will filter out the annual variation by means of finding the deviations of each value from the corresponding evaluation of the mean (Table 1) and by dividing these deviations by the corresponding evaluations of the standard deviations (Table 2). The normalized zonal anomalies obtained in this way for each latitude zone separately constitute time series which it is possible to analyze by the methods of the theory of stationary random processes. The periodograms and spectra (in percent) found for each of the 11 latitude zones are given in Fig. 1. Some periodograms have statistically significant maxima in the low frequencies (with periods of several years), but their positioning on the frequency axis is random and they do not have precisely coinciding periods for data for different latitudes.

Without citing evaluations of the spectral functions, we point out (see the last column of Table 3) that for 9 of the 11 time series the deviations of the computed values from the theoretical spectral density of white noise are statistically significant (fall beyond the 95% confidence interval).

The spectra obtained by smoothing of the periodograms usually have a small singularity in the low frequencies with periods of several years. They are close to one another in the sense that the main fraction of the variability falls on white noise.

Table 3 gives evaluations of the correlation functions for lags of 1 and 2 months. For large shifts the correlations were found to differ insignificantly from zero.

Due to the fact that the periodograms and evaluations of the spectra corresponding to different latitude zones are close to one another, we carried out averaging for a set of periodogram values. The values thus obtained are given in Fig. 1 (upper graph at right). Smoothing of the averaged periodogram made it possible to find a spectrum whose Fourier transform gives the correlation function. The spectrum and correlation function are close to the evaluations which were obtained for each latitude zone separately. The spectrum has a small singularity in the low frequencies and the correlation function (which we do not cite) with a lag of one month is equal to 0.2.

White noise on the average accounts for about 80% of the dispersion of the standard anomalies of the time series of middle-latitude geopotential values. We also note that the maximum level of white noise corresponds to latitudes 60 and 65° (approximately 90% dispersion) and the minimum to a latitude of 30° (approximately 65% dispersion).

Now we will examine the latitudinal-temporal changes in normalized zonal geopotential anomalies and we will find their two-dimensional spectral and correlation characteristics.

A detailed description of the employed mathematical approach is given in [7]. We note that the numerical models considered in [7] represent a generalization for a two-dimensional case of the method used in a spectral analysis of stationary processes [6]. The sequence of computations is as follows.

Using a two-dimensional realization

$$Y_{\theta t}\ (\theta = 0, 1, \ldots, N-1,\ \ t = 0, 1, \ldots, k-1)$$

($\Delta\theta$ is the interval along the θ- axis, Δt is the interval along the t axis) of the homogeneous random field we will compute the coefficients of the two-dimensional Fourier transform-

$$A_{pq} = \frac{1}{Nk} \sum_{\theta=0}^{N-1} \sum_{t=0}^{k-1} Y_{\theta t} e^{-i(\Omega_p \theta + \omega_q t)},$$

where

$$(p = 0, 1, \ldots, N-1,\ \ q = 0, 1, \ldots, k-1),$$

$$\Omega_p = \frac{2\pi}{N\Delta\theta} p,\quad \omega_q = \frac{2\pi}{k\Delta t} q.$$

The values

$$S_{pq} = \frac{N\Delta\theta k\Delta t}{4\pi^2} |A_{pq}|^2$$

determine a two-dimensional periodogram whose smoothing using the formula

$$\overline{S}_{pq} = \frac{1}{(2s+1)(2r+1)} \sum_{l=-s}^{s} \sum_{j=-r}^{r} S_{p+l,\ q+j}$$

gives evaluations of the spectral density. The two-dimensional correlation function $M_{\tau_1 \tau_2}$ is equal to

$$M_{\tau_1 \tau_2} = \frac{4\pi^2}{N\Delta\theta k\Delta t} \sum_{p=0}^{N-1} \sum_{q=0}^{k-1} S_{pq} e^{i(\Omega_p \tau_1 + \omega_q \tau_2)}.$$

The values

$$T_1(p) = \frac{2\pi}{\Omega_p} \quad\text{and}\quad T_2(q) = \frac{2\pi}{\omega_q}$$

Table 3

Values of Correlation Function for Lag of One and Two Months and Presence of Significant Differences in Spectral Functions from White Noise

Широта, град[1]	K (1)	K (2)	Δ
80	0,17	0,09	+
75	0,15	0,13	+
70	0,15	0,14	±
65	0,11	0,10	−
60	0,10	0,06	−
55	0,25	0,11	+
50	0,33	0,21	+
45	0,24	0,21	+
40	0,17	0,14	+
35	0,21	0,16	+
30	0,37	0,23	+

KEY:
1. Latitude, degrees

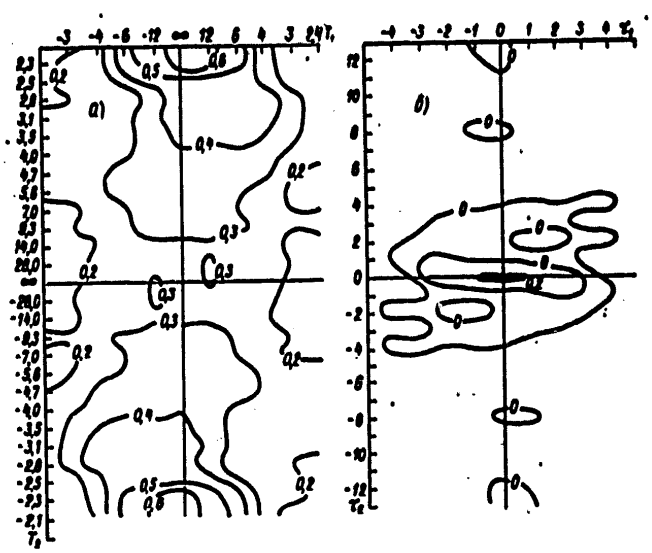

Fig. 3. Evaluation of two-dimensional spectrum (a) and two-dimensional correlation function (b) of time series of mean monthly middle-latitude geopotential AT500. T_1 -- period (months), T_2 -- period (years), τ_1 -- time shift (months), τ_2 -- time shift (years).

determine the periods (or wavelengths) of the oscillations, in accordance with the dimensionality of the θ-axis and t. The spectral evaluations considered below are considered in percent of the dispersion.

Table 4

Evaluations of Statistical Characteristics of Linear Trend for Middle-Latitude Geopotential AT_{500} for Different Latitude Zones for 28 Years (1948-1975)

Широта, град (1)	σ_1	β_1	s_{β_1}	$t\ (k=1)$	$s_{\bar{y}}$	s_0	$\bar{y}$
80	1,727	−0,074	0,0404	1,8	0,340	1,801	524,368
75	1,360	−0,052	0,0318	1,6	0,265	1,401	527,193
70	1,003	−0,046	0,0235	2,0	0,199	1,054	530,893
65	0,631	−0,027	0,0148	1,8	0,124	0,657	534,918
60	0,497	−0,018	0,0116	1,6	0,097	0,511	539,314
55	0,643	−0,013	0,0150	0,9	0,121	0,641	544,771
50	0,673	−0,063	0,0157	4,0	0,158	0,838	551,025
45	0,617	−0,081	0,0144	5,6	0,170	0,902	559,889
40	0,466	−0,035	0,0109	2,2	0,102	0,539	567,429
35	0,423	−0,014	0,0098	1,4	0,081	0,430	574,548
30	0,537	−0,031	0,0126	2,5	0,110	0,584	580,532

KEY:
1. Latitude, degrees

Figure 2 represents the central parts of the fields of spectral and correlation evaluations respectively. The values omitted in these figures are close to zero. It follows from an analysis of the periodogram that the main intensity falls on temporal changes. There are several small peaks corresponding to points on the periodograms close to the origin of coordinates. With smoothing a small spectral maximum is formed in the neighborhood of the origin of coordinates. The correlation of values from one latitude zone to another (0.53) is considerably greater than the time correlation from month to month (0.21). It must be remembered that the time and latitude scales in no way agree with one another.

Two-dimensional spectral and correlation characteristics can be obtained for time series of mean monthly values of individual latitude zones together with one-dimensional spectra (Fig. 1). In this case each times series is arbitrarily considered as a two-dimensional random field where the changes along the x-axis transpire with an interval of one month and along the y-axis with an interval of one year. By analyzing such a field we seemingly divide the intraannual and year-to-year variations and we have the possibility of studying their interaction. For one of the time series of normalized zonal anomalies of geopotential 500 mb (latitude 55°) Fig. 3 shows the two-dimensional spectrum and correlation function respectively. An interesting peculiarity of the periodogram is a maximum in its high-frequency part corresponding to periods of about two years. There are also a number of other peculiarities of a somewhat lesser importance.

The correlation function shows the presence of small positive correlations (0.19) from month to month and still lesser negative correlations (-0.12) from year to year.

Similar evaluations were obtained for each of the considered time series. However, their examination would occupy too much place.

We will give an analysis of the trends in the change in geopotential during the considered 28-year time period. For this purpose the mean annual values for each of the latitudes are approximated by a straight line $\bar{y} + \beta_1 t$ and the statistical significance of the mean rate β_1 of change in geopotential 500 mb is evaluated. The results are given in Table 4.

The notations used in Tables 4-6 are as follows: $\bar{y}$ and β_1 are the coefficients of the straight line $\bar{y} + \beta_1 t$, σ_1^2 is the dispersion of the deviation from a straight line, $\sigma^2\beta_1$ is the dispersion of the coefficient β_1, $t(k - 1) = |\beta_1| / \sigma\beta_1$ are Student statistics, $\sigma^2/\bar{y}$ is the dispersion of the mean, σ_0^2 is the dispersion of the deviations from the mean.

With $t(k - 1) > 2$ the β_1 evaluations are statistically significant with the 95% level.

The results in Table 4 show that during the considered time intervals at all latitudes on the average there was a small decrease in the geopotential values. In five cases (of 11) this decrease was statistically significant with the 95% level.

Table 5

Evaluations β_1 for Different Periods (Numerator) and Corresponding Student Statistics (Denominator)

Широта, град (1)	1948—1955	1956—1965	1966—1975
80	—0,245/0,9	—0,339/1,8	—0,005/0,02
75	—0,037/0,2	—0,258/1,7	—0,041/0,2
70	0,068/0,5	—0,243/2,3	—0,076/0,6
65	0,031/0,4	—0,196/3,0	—0,089/1,4
60	—0,018/0,8	—0,1[illegible]5/2,4	—0,121/2,2
55	—0,114/1,4	—0,057/0,9	—0,132/1,7
50	—0,254/2,9	—0,099/1,6	—0,112/1,3
45	—0,274/4,4	—0,156/2,8	—0,073/0,9
40	—0,210/4,6	—0,117/2,9	0,012/0,2
35	—0,064/3,5	—0,132/2,5	0,050/1,3
30	—0,020/0,6	—0,196/3,3	0,075/1,9

KEY:

1. Latitude, degrees

For a more detailed characterization of the change in geopotential the evaluations of the β_1 parameter were found for three different time intervals (1948-1955), (1956-1965), (1966-1975).

Table 5 gives the computed values and the corresponding Student statistics. The results show that a decrease in geopotential occurred particularly intensively during the period from 1956 through 1965, when there were seven statistically significant β_1 evaluations. During the period 1948-1955 there are four, and during the period 1966-1975 only one statistically significant evaluation.

It should be noted that during the last 10 years observations of geopotential are considerably more precise than during the preceding years.

The statistical characteristics of geopotential AT_{500} considered above make it possible to draw the following conclusions.

1. Most of the variability of the normalized zonal geopotential anomalies is due to white noise, which on the average is about 80% of the dispersion.

2. The spectra of the time series for each of the considered latitude zones have a small singularity in the low frequencies with periods of several years. Individual oscillations of relatively great intensity, showing the possibility of the existence of long-period oscillations, correspond to this sector of frequencies.

However, due to the fact that during the considered time interval the number of aerological stations substantially increased, we cannot make any categorical assertions that the discovered oscillations are natural. An increase in the number of stations inevitably leads to a refinement of the mean latitudinal values, that is, the appearance of long-period trends caused by a constant decrease in the systematic error of the mean. This trend can be the reason for the long-period oscillations discovered both in an analysis of the spectra and in an examination of the linear trend. Such oscillations are a result of methodological, not natural physical factors.

It is clear that at the present time it is impossible to say definitely whether the discovered peculiarities are caused by methodological or natural factors.

3. Even using relatively small observation periods the climatic means (annual variation of the norms) of different elements of the free atmosphere can be determined with a sufficiently high accuracy, as follows from the coincidence of our results with [1, 2, 5, 9].

We note in conclusion that it is of great interest to make a similar statistical analysis of geopotential data at other levels and also for other characteristics of state and circulation of the atmosphere. A comparative study of the evaluations, and also the reciprocal statistical characteristics, makes it possible to clarify the patterns of change in the statistical structure of the meteorological elements in the free atmosphere.

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SOME RESULTS OF COMPUTATIONS OF THE PRINCIPAL ENERGY CHARACTERISTICS OF THE ZONAL CIRCULATION IN THE UPPER ATMOSPHERE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 44-49

[Article by Candidates of Geographical Sciences I. V. Bugayeva and L. S. Minyushina and Candidate of Physical and Mathematical Sciences G. R. Zakharov, Central Aerological Observatory, submitted for publication 3 July 1978]

Abstract: The paper gives the results of computations of kinetic energy per unit volume, available potential energy and generation of kinetic energy in the atmospheric layer from the earth to 60 km. The computations were made on the basis of mean monthly sections for individual years and long-term meridional sections of the wind and temperature fields along 75°W from 82°N to 20°S, constructed using data from rocket and radiosonde observations during 1967-1974.

[Text] Investigation of energy processes in the atmosphere is one of the important directions in modern physics of the atmosphere. It is sufficient to recall how important it is to know the energy characteristics in formulating prognostic models and models of general circulation of the atmosphere [6, 12, 13]. There is now an extensive literature [1, 2, 5, 8, 17] on matters relating to problems involved in the levels of energy in the troposphere and lower stratosphere. There has been a far less detailed investigation of energy processes in the higher layers of the atmosphere [14, 15, 18].

The principal source of information on the processes transpiring in the upper stratosphere and lower mesosphere until recently was data from rocket sounding. During recent years ways have been developed for using satellite information [9, 10, 19]. The presently existing world rocket network makes it possible to obtain an adequate amount of information for constructing meridional sections of the wind and temperature fields on the basis of which it is possible to study zonal circulation of the upper atmosphere.

It is of interest to investigate zonal circulation of the upper atmosphere from the point of view of its energy characteristics. For computations of the energy characteristics we used the mean monthly meridional sections of the zonal (V_{zon}) and meridional (V_{mer}) wind and temperature along 75°W from 82°N to 20°S in the atmospheric layer from the ground to 60 km during the period from 1967 through 1974. At the same time, we also constructed the mean monthly sections on the basis of long-term data.

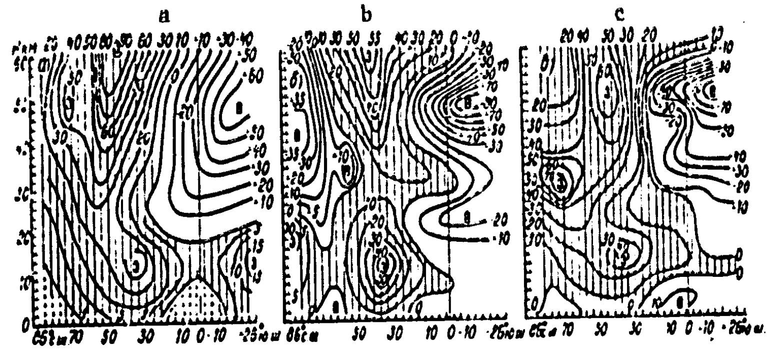

Fig. 1. Spatial sections of zonal wind velocity component u m/sec along 75°W. a) for January on basis of long-term data; b) for January 1971; c) for January 1967. Westerly wind -- positive. The dots represent areas of absence of data.

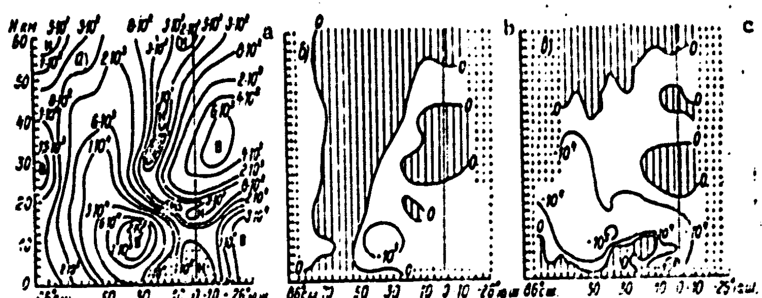

Fig. 2. Spatial sections of field of kinetic energy per unit volume -- KE ($g/(m \cdot sec^2)$). a) for January on the basis of long-term data; b) deviations of kinetic energy for January 1971 from KE on the basis of long-term data; c) deviations of KE for January 1967 from KE on the basis of long-term data. The regions of positive values are shaded. The dots represent fields of absence of data.

In constructing the sections we used observational data on temperature and wind from the network of rocket and radiosonde stations published in [7, 16]. For the stratosphere and lower mesosphere we took data from 11 stations in the American rocket network. In latitude the intervals between individual stations attain 15-20° (1,500-2,000 km) and therefore the sections cannot reflect the circulation systems, whose characteristic dimensions are less than this value. For the troposphere we have used data for 20 radiosonde stations. In the altitude range 20-30 km we carried out initialization of rocket and radiosonde data. As an example, Fig. 1 shows meridional sections of zonal wind fields.

The values of the zonal and meridional wind and temperature were taken from the sections with an altitude interval 2 km and a latitude interval 4°. In those cases when in the high latitudes of the northern hemisphere and in the low latitudes of the southern hemisphere there was a lack of data, no computations are given for these regions.

We will begin an examination of the energy characteristics with an analysis of the spatial distribution of kinetic energy (KE) per unit volume

$$KE = 1/2\ (V_{zon}^2 + V_{mer}^2)\rho, \qquad (1)$$

where ρ is density.

Figure 2a shows the KE field constructed for January on the basis of long-term data. The structure of the KE fields is determined by the distribution of the fields of maxima and minima of wind velocity and the vertical ρ profile. The maximum KE values ($2\cdot10^5$ g/(m·sec^2)) are related to the jet stream region in the troposphere: in latitude the region of the maximum V_{zon} and KE values coincide, whereas in altitude the KE maximum is approximately 2 km below the V_{zon} maximum. The region of increased KE values in the latitude zone 54-34°N is traced to 60 km, but in contrast to the field of the zonal wind without maxima -- 46 and 60 km. In the southern hemisphere in the latitude zone 10-20°S there is a KE maximum at an altitude of 34-40 km, equal to $0.7\cdot10^4$ g/(m·sec^2) (the focus of maximum velocities V_{zon} is situated at altitudes 48-54 km).

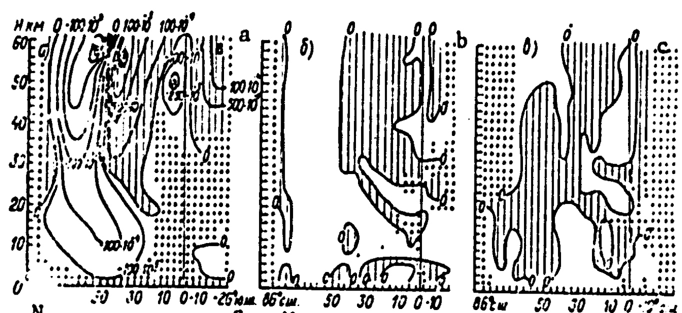

Fig. 3. Spatial sections of field of generation of kinetic energy of zonal motion GKE (m^2/sec^3). a) for January on basis of long-term data; b) deviations of GKE for January 1971 from GKE according to long-term data; c) deviations of GKE for January 1967 from GKE according to long-term data. Regions of positive values shaded. Dots denote regions without data.

Zones of decreased KE values are localized in regions of change in circulation sign, that is, in regions with a zero isotach and in regions adjacent to it with low wind velocities. In July, when over the northern hemisphere in the stratosphere and mesosphere there are easterly winds with maximum velocities in the middle latitudes at an altitude of 58-60 km, in the kinetic energy field at these same latitudes, but somewhat lower, at an altitude of 48-55 km, there is a maximum equal to $0.2 \cdot 10^4$ g/(m·sec^2). A second KE maximum is situated in the troposphere in the jet stream zone and is $2 \cdot 10^4$ g/(m·sec^2). The zone of minimum KE values passes at the level of separation of the stratospheric easterly circulation and the tropospheric westerly circulation, that is, at the velopause [3], and also in the region of separation between the easterly summer circulation of the northern hemisphere and the westerly winter circulation of the southern hemisphere.

For the remaining months there is also a coincidence in the position of the maximum V_{zon} and KE values in latitude and a lowering of the latter in altitude relative to the V_{zon} maxima in depth to 2-4 km in the troposphere and to 10-20 km in the stratosphere. The maximum KE values fall in March and the minimum values fall in the summer months.

Now we will discuss the available potential energy (APE) of zonal movement:

$$\mathrm{APE} = \int \frac{\gamma}{2} [T - \overline{T}_p]^2 \, g \rho \, d\sigma, \tag{2}$$

where

$$\gamma = -\frac{R}{gp\left(\frac{\partial T}{\partial p} - \frac{RT}{C_p p}\right)},$$

R is the universal gas constant of air, $\overline{T}_p$ is mean temperature at the level p, g is the acceleration of free falling, C_p is the specific heat capacity of air at a constant pressure, $d\sigma$ is an element of volume.

Integration is carried out for the entire considered region of the atmosphere.

It is known that the available potential energy is the difference between the potential energy of a particular state and the total potential energy of the conditional state, obtained from the initial state as a result of an adiabatic process in which the isobaric surfaces and the surfaces of equal potential temperature coincide, and accordingly, the rate of reciprocal transition from the total potential energy to kinetic energy and back is equal to zero or is minimum. Together with the integral APE, it is interesting to examine the spatial distribution of "specific APE," under this understanding that part of the total potential energy at a point which enters into the APE.

The principal characteristics of structure of the field of "specific APE" in all months remain similar: the maximum values -- in the region of the pole and equator, the minimum values -- in the middle latitudes, a decrease in "specific APE" with altitude (due to a decrease in ρ). The reserves of APE in winter are $3\cdot10^{23}$ g/(m·sec^2). In summer the APE reserves are less than in winter, approximately by a factor of 2-4 and constitute about $1\cdot10^{23}$ g/m·sec^2, which is associated with the attenuation of contrast between the pole and the equator.

Now we will examine the peculiarities of the spatial distribution of the generation of kinetic energy (GKE) of zonal movement per unit mass:

$$\mathrm{GKE} = -\frac{1}{\rho}\frac{V_{\text{мер}}}{R_з}\frac{\partial p}{\partial \varphi}, \tag{3}$$

[з = E; мер = mer] where R_E is the earth's radius, p is pressure, φ is latitude.

The results of computations of GKE for January on the basis of long-term data are presented in Fig. 3a. Since the transformation from potential energy to kinetic energy as a result of meridional movement is a reversible process, in the meridional section there are both regions of kinetic energy (positive regions) and regions of transition from the kinetic energy of zonal movement to zonal potential energy (regions of negative values).

It follows from Fig. 3a that the regions of negative values are localized in the troposphere and stratosphere of the high and temperate latitudes, with a maximum of -0.24 m^2/sec^3, at an altitude of 56-60 km, and in the equatorial latitudes with a maximum at 48-52 km, which is $-0.25\cdot10^{-2}$ m^2/sec^3. The region of generation of kinetic energy is noted in the stratosphere in the latitude zone 52-18°N with a maximum value 0.15 m^2/sec^3 at an altitude 54-58 km. In July in the stratosphere and in the lower mesosphere, as far as the equator, there is a region of negative GKE values with a maximum in the middle latitudes above 54 km (-0.34 m^2/sec^3).

An analysis of the GKE field on the basis of long-term data in the course of the year demonstrated that in the stratosphere and in the lower mesosphere of the high and temperate latitudes in summer and winter the GKE values are negative, whereas in the transitional seasons they are positive. Accordingly, in these regions in the main seasons of the year there is a transformation from the kinetic energy of zonal movement to zonal potential energy. In the tropical latitudes the negative GKE values are observed only in the summer months; in winter, however, and in the transition seasons there is a transition from potential and kinetic energy. In the troposphere and lower stratosphere the GKE values are negative during the entire year. This indicates that by limiting the frameworks of zonal circulation it is impossible to explain the maintenance of stable zonal movements in this region of the atmosphere. The source of kinetic energy here is the kinetic energy of eddies.

In addition to computations of the energy characteristics on the basis of long-term data, we carried out computations of KE, APE and GKE using mean monthly sections for 1967-1974. We will consider how the influence of winter stratospheric warmings is manifested in the fields of energy characteristics. For an analysis of the influence of winter warmings we selected Januaries as months which are sufficiently representative for an investigation of this phenomenon. All Januaries from 1967 through 1974 are broken down into two groups: the first includes Januaries of those years when there were strong warmings with a pressure restructuring of the field in the high latitudes (on the mean monthly sections this was manifested in the setting-in of an easterly circulation); to the second group we assigned the Januaries of those years when the warmings were weak, without restructuring of circulation on the mean monthly sections [4, 11, 19].

Figure 2b,c represents the KE deviations determined using data for January 1971 (strong warming) and January 1967 (weak warming) from the KE computed using long-term data. In 1971 there is a well-expressed latitudinal variation of KE deviations. In the high latitudes, to the north of 72°N, there is a region of excess of KE over the long-term values in virtually the entire thickness of the atmosphere, that is, there is an increase in KE. This is attributable to the fact that in the high-latitude stratosphere during the warming period there is a high-altitude frontal zone and a considerable intensification of the wind, especially the meridional component, which increases by a factor of 2-3 in comparison with the long-term values. To the south of 72°N there is a zone of negative deviations and in the troposphere and in the lower stratosphere it is bounded by 50°N; with increasing altitude it expands and in the mesosphere attains 25°N, that is, in the temperate latitudes during the period of warmings there is a decrease in the kinetic energy of zonal movement. Thus, during a period of warming there is a redistribution of kinetic energy with altitude and latitude -- an increase in KE is noted in the high latitudes and a decrease is observed in the middle latitudes.

In 1967 such a latitudinal variation of KE deviations is absent. We note that such a structure of the distribution of deviations is also maintained in other years with strong and weak warmings.

Now we will show how the warming processes are reflected in the fields of GKE deviations. Figure 3b,c shows the deviations of GKE for the Januaries of 1971 and 1967 from the GKE for January, computed on the basis of long-term data. In 1971, in the atmospheric layer above 30 km, in the latitude zone 78-50°N, and below 30 km -- virtually in the entire latitude range, there is a GKE excess over the long-term values by a value of about 0.1-0.2 m^2/sec^3. Since in this region the GKE, according to long-term data (Fig. 3a), is negative and varies in the range from -0.1 to -0.01 m^2/sec^3, that is, there is a transition from kinetic energy to potential energy, the GKE for January 1971 in the stratosphere and mesosphere for the high latitudes of this zone becomes positive, whereas in the remaining part of this zone, remaining negative, it decreases in absolute value. This means that for the stratosphere and mesosphere during the period of warming in the high latitudes (80-70°N) there is a transition from potential to kinetic energy; however, in the region of the temperate latitudes (60-50°N) the rate of transition from kinetic energy to potential energy decreases in comparison with the rate of transition on the basis of long-term data. Such a restructuring of the GKE field is determined by the restructuring of the pressure field: in the high latitudes there is a change in the sign of the pressure gradient (an anticyclone is situated in the pole region), in the lower latitudes the pressure gradients decrease. A similar structure of the GKE field is characteristic for all Januaries when there were strong warmings (1968, 1969, 1970, 1971). In January with weak warmings (1967, 1972, 1973) there were no such regularities in the GKE field.

Thus, our computations of the energy characteristics (KE, GKE, APE) demonstrated a substantial difference in their spatial distribution in dependence on season and the nature of circulation processes.

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DYNAMIC INITIALIZATION OF INITIAL FIELDS FOR A BAROCLINIC PROGNOSTIC MODEL

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[Article by Candidate of Physical and Mathematical Sciences M. S. Fuks-Rabinovich, USSR Hydrometeorological Scientific Research Center, submitted for publication 30 May 1978]

Abstract: The author proposes a method for taking into account information on surface pressure trends during the formation of the fields of initial information for a baroclinic prognostic model. For this purpose the author uses a definite modification of the procedure for dynamic initialization of fields. A study was made of the proposed variant of the dynamic initialization procedure. The article cites the results of numerical experiments with a baroclinic regional prognostic model indicating that the use of this approach exerts a positive influence on the quality of the forecast.

[Text] Introduction. It is known that use of the dynamic initialization of initial fields by use of the procedure of so-called pseudoforecasting [6, 9] makes it possible to increase the success of numerical forecasting [6, 10]. The dynamic initialization of fields is also used in a four-dimensional analysis of meteorological information [6]. However, until recently the dynamic initialization method on the basis of a pseudoforecast was rigorously developed only for a barotropic model of the atmosphere. It is evident that with the mentioned considerations taken into account it is of interest to generalize this method applicable to a baroclinic model of the atmosphere. Such a generalization, however, involves considerable difficulties, examined in different studies [6, 8-10]. Below we propose an approach making it possible, as initial information, in a hydrodynamic forecast, to take into account data on surface pressure trends by means of a simple modification of the traditional procedure of dynamic initialization of fields by the pseudoforecasting method.

It is well known that the use for forecasting purposes of data on surface pressure trends, which is common in routine synoptic practice, leads to positive results in a number of prognostic models [1, 3, 4]. However, we note that in the mentioned studies use is made of synoptic, statistical or mixed statistical-hydrodynamic or synoptic-hydrodynamic approaches. It seems desirable to examine the possibility of taking additional information into account on pressure trends within the framework of a hydrodynamic model on the basis of an approach in which direct use is made of the prognostic equations of this model. However, we meet with a difficulty in that data on the surface pressure trend cannot be substituted directly into the system of prognostic equations since this leads to an overdetermination of the latter. It will be demonstrated below that by using the pseudoforecasting procedure in a definite modification it is possible, uniformly within the framework of a hydrodynamic model, to take into account data on surface pressure trends and form an initial state which would "absorb" this useful additional information in itself.

Investigation of the Procedure of Initialization of Fields in a Simple Model

We will examine the idea of the method in a simple example for a model of a one-dimensional flow of an ideal heavy fluid in a linear approximation:

$$\frac{\partial u}{\partial t} + \frac{\partial \Phi}{\partial x} = 0, \qquad \frac{\partial \Phi}{\partial t} + H\frac{\partial u}{\partial x} = 0, \tag{1}$$

where H = const is the mean height of the free surface, u is velocity, Φ is geopotential.

A knowledge at the initial moment of the values of the trends (in this case the Φ trend) is equivalent to the fact that we know not only the initial value $\Phi_0 = \Phi|_{t=0}$, but also, for example, the value $\Phi_1 = \Phi|_{t=\Delta t}$, that is, the future Φ value in the first (or, generally speaking, in the n-th) time integration interval. We note that under real conditions the future Φ value is determined from the field of surface pressure trends by means of linear time interpolation.

We will formulate the pseudoforecasting procedure in such a way that there will be a forced adaptation of the Φ field to the Φ_0 and Φ_1 values, that is, to data on geopotential at the ends of some time interval.

In the "step ahead" procedures of pseudoforecasting, using the Euler model with scaling, for system (1) (the values of the predicted parameters in the first iteration are noted by the superscript "*")

$$\begin{aligned} u^*_{n+1} &= u_n - (\Phi_0)_x\,\Delta t, \\ \Phi^*_{n+1} &= \Phi_0 - H(u_n)_x\,\Delta t, \\ u_{n+1} &= u_n - (\Phi^*_{n+1})_x\,\Delta t, \end{aligned} \tag{2}$$

$$\Phi_{n+1} \cong \Phi_1. \tag{2}$$

Performing the "step backward" with this same integration model, we obtain

$$\begin{aligned} u_n^* &= u_{n+1} + (\Phi_1)_x \Delta t, \\ \Phi_n^* &= \Phi_1 + H (u_{n+1})_x \Delta t, \\ u_n^{(\nu+1)} &= u_{n+1} + (\Phi_n^*)_x \Delta t, \\ \Phi_n &\cong \Phi_0, \end{aligned} \tag{3}$$

where the subscript x is the operator for differentiation for x, n is the number of the time interval, Δt is the time interval, ν is the number of the iteration in the cyclic procedure of pseudoforecasting (as a simplification it is noted only when obtaining the new value $u_n^{(\nu+1)}$, in other places the ν superscript has been omitted).

After a step forward (2) we have

$$u_{n+1} = u_n - (\Phi_0)_x \Delta t + H (u_n)_{xx} (\Delta t)^2, \tag{4}$$

and after a step backward (3) we obtain

$$u_n^{(\nu+1)} - u_n = (\Phi_1 - \Phi_0)_x \Delta t + 2 H (u_n)_{xx} (\Delta t)^2 - \\ - H (\Phi_0)_{xxx} (\Delta t)^3 + H^2 (u_n)_{xxxx} (\Delta t)^4. \tag{5}$$

The first term on the right-hand side of (5) is a derivative of x from the known field trend $\bar{\Phi}$. Accordingly, additional information has obviously entered into the result of integration and furthermore in the form of the main term determining the change in the wind field in the particular modification of the pseudoforecasting procedure.

We note that this modification of the pseudoforecasting procedure with forced adaptation to the $\bar{\Phi}_0$ and $\bar{\Phi}_1$ values is approximately twice as economical as the traditional procedure, since for practical purposes it is necessary to carry out integration only for the second and third equations of systems (2) and (3), and not for all four.

Investigation of the Initialization Procedure for a Baroclinic Model

We will generalize the procedure considered above for the initialization of fields applicable to a baroclinic model. For this we will examine a linearized system of equations in a p-system of coordinates for a two-layer baroclinic model in full equations (this linearized system of equations was derived in [7], but was used for other purposes):

$$
\begin{aligned}
&\frac{\partial u_1}{\partial t} + \frac{\partial \Phi_1}{\partial x} = 0,\\
&\frac{\partial u_3}{\partial t} + \frac{\partial \Phi_3}{\partial x} = 0,\\
&\frac{\partial T_1}{\partial t} + \overline{S_1}\,\omega_2 = 0,\\
&\frac{\partial T_3}{\partial t} + \overline{S_3}\,\omega_2 = 0,\\
&\frac{\partial p_*}{\partial t} - \omega_* = 0,\\
&\omega_2 + \frac{\partial u_1}{\partial x}\Delta p = 0,\\
&\omega_* - \omega_2 + \frac{\partial u_3}{\partial x}(\overline{p}_* - p_2) = 0,\\
&\Phi_3 - AT_3 - B\pi_* = 0,\\
&\Phi_1 - \Phi_3 - ET_1 - FT_3 = 0,\\
&\pi_* = Pp_*.
\end{aligned}
\tag{6}
$$

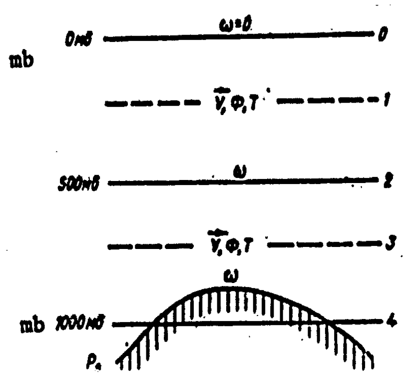

Fig. 1. Vertical structure of two-layer model.

The vertical structure of the model is represented in Fig. 1, from which the annotations are understandable. The system of ten equations (6) consists of prognostic equations relative to the parameters u_1, u_3, T_1, T_3, p_* (first five equations) and five diagnostic equations relative to the parameters ω_2, ω_*, Φ_1, Φ_3, π_*. Here we use the following notations: ω is the analogue of vertical velocity,

$$\overline{S}_1 = \pi_1 (\overline{\theta}_3 - \overline{\theta}_1)/2\, p_3; \quad \pi = \left(\frac{p}{p_4}\right)^{\varkappa};$$

$$\varkappa = R/c_p = \frac{2}{7}; \quad \Delta p = p_3 = 500 \text{ мб};$$

$$\theta\pi = T; \quad \overline{S}_3 = \pi_3 (\overline{\theta}_3 - \overline{\theta}_1)/2\, p_3;$$

$$A = C_p (\overline{\pi}_* - \pi_3)/\pi_3; \quad B = C_p \overline{T}_3/\pi_3;$$

$$E = C_p (\pi_3 - \pi_1)/2\, \pi_1; \quad p = \varkappa \overline{\pi}_* / \overline{p}_*;$$

$$F = C_p (\pi_3 - \pi_1)/2\, \pi_3.$$

In deriving the system of equations (6) we made the assumption

$$\pi_* = \left(\frac{p_* + p_*'}{p_4}\right)^{\varkappa} \cong \left(\frac{\overline{p}_*}{p_4}\right)^{\varkappa} + \varkappa \left(\frac{\overline{p}_*}{p_4}\right)^{\varkappa - 1} \frac{p_*'}{p_4}.$$

We will formulate an initialization procedure, assuming that we know additional data on surface pressure trends, or, in other words, that we know the values p^0 and p^1 of surface pressure at the ends of some time interval (which in this examination is equal to the time interval Δt). As above, using the Euler scheme for time integration (with scaling), in the step ahead we have (the field values in the first iteration are designated by the superscript "*")

$$
\begin{aligned}
& u_i^{*\,n+1} = u_i^n - \Phi_{ix}^n \Delta t, \\
& T_i^{*\,n+1} = T_i^n - \overline{S}_i \omega_2^n \Delta t; \quad (i = 1, 3), \\
& p_*^{*\,n+1} = p_*^0 + \omega_*^n \Delta t, \\
& \omega_2^{*\,n+1} = - u_{1\,x}^{*\,n+1} \Delta p, \\
& \omega_*^{*\,n+1} = \omega_2^{*\,n+1} - u_{3\,x}^{*\,n+1} (\overline{p}_* - p_3), \\
& \Phi_3^{*\,n+1} = A T_3^{*\,n+1} + B \pi_*^{*\,n+1}, \\
& \Phi_1^{*\,n+1} = \Phi_3^{*\,n+1} + E T_1^{*\,n+1} + F T_3^{*\,n+1}, \\
& \pi_*^{*\,n+1} = P p_*^{*\,n+1},
\end{aligned}
\qquad (7)
$$

and then, for the second iteration in the step ahead

$$
\begin{aligned}
& u_i^{n+1} = u_i^n - \Phi_{ix}^{*\,n+1} \Delta t, \\
& T_i^{n+1} = T_i^n - \overline{S}_i \omega_2^{*\,n+1} \Delta t; \quad (i = 1, 3), \\
& p_*^{n+1} = p_*^1, \\
& \omega_2^{n+1} = - u_{1\,x}^{n+1} \Delta p,
\end{aligned}
$$

$$\begin{aligned}
&\omega_*^{n+1} = \omega_2^{n+1} - u_{3\,x}^{n+1}(\overline{p}_* - p_2),\\
&\Phi_3^{n+1} = A T_3^{n+1} + B \pi_*^{n+1},\\
&\Phi_1^{n+1} = \Phi_3^{n+1} + E T_1^{n+1} + F T_3^{n+1},\\
&\pi_*^{n+1} \equiv P p_*^1.
\end{aligned} \tag{8}$$

and then for the second iteration in the step backward:

$$\begin{aligned}
&u_i^{*n} = u_i^{n+1} + \Phi_{ix}^{n+1}\Delta t,\\
&T_i^{*n} = T_i^{n+1} + \overline{S}_i \omega_2^{n+1}\Delta t; \quad (i = 1,\ 3).\\
&p_*^{*n} = p_*^1 - \omega_*^{n+1}\Delta t,\\
&\omega_2^{*n} = - u_{1\,x}^{*n}\Delta p,\\
&\omega_*^{*n} = \omega_2^{*n} - u_{3\,x}^{*n}\ (\overline{p}_* - p_2),\\
&\Phi_3^{*n} = A T_3^{*n} + B \pi_*^{*n},\\
&\Phi_1^{*n} = \Phi_3^{*n} + E T_1^{*n} + F T_3^{*n},\\
&\pi_*^{*n} = P p_*^{*n},
\end{aligned} \tag{9}$$

We will write similar expressions for the step backward:

$$\begin{aligned}
&(u_i^n)^{(\nu+1)} = u_i^{n+1} + \Phi_{ix}^{*n}\Delta t,\\
&(T_i^n)^{(\nu+1)} = T_i^{n+1} + \overline{S}_i \omega_2^{*n}\Delta t; \quad (i = 1,\ 3),\\
&(p_*^n)^{(\nu+1)} \equiv p_*^0,\\
&(\omega_2^n)^{(\nu+1)} = -(u_{1\,x}^n)^{(\nu+1)}\Delta p,\\
&(\omega_*^n)^{(\nu+1)} = (\omega_2^n)^{(\nu+1)} - (u_{3\,x}^n)^{(\nu+1)}\ (\overline{p}_* - p_2),\\
&(\Phi_3^n)^{(\nu+1)} = A\,(T_3^n)^{(\nu+1)} + B\,(\pi_*^n)^{(\nu+1)},\\
&(\Phi_1^n)^{(\nu+1)} = (\Phi_3^n)^{(\nu+1)} + E\,(T_1^n)^{(\nu+1)} + F\,(T_3^n)^{(\nu+1)},\\
&(\pi_*^n)^{(\nu+1)} \equiv P p_*^0.
\end{aligned} \tag{10}$$

We will now show how the considered fields change as a result of the full cycle of the pseudoforecasting procedure (7)-(10) in a forced adaptation regime. After rather unwieldy, but simple transformations we obtain the following expressions for the resultant fields:

$$(u_1^n)^{(\nu+1)} = u_1^n + BP\,(p_*^1 - p_*^0)_x\,\Delta t + \{[2\,BP - (A+F)\,\overline{S}_3 - E\overline{S}_1]\,u_{1\,xx}^n\,\Delta p +$$
$$+ 2\,BP\,(\overline{p}_* - p_2)\,u_{3\,xx}^n\}\,(\Delta t)^2 + O\,[(\Delta t)^3],$$
$$(u_3^n)^{(\nu+1)} = u_3^n + BP\,(p_*^1 - p_*^0)_x\,\Delta t + [(2\,BP - A\overline{S}_3)\,u_{1\,xx}^n +$$
$$+ 2\,BP\,(\overline{p}_* - p_2)\,u_{3\,xx}^n]\,(\Delta t)^2 + O\,[(\Delta t)^3],$$
$$(T_1^n)^{(\nu+1)} = T_1^n - [\overline{S}_1\,\Delta p\,\Phi_{1\,x}^n + A\overline{S}_3\,(\Delta p)^2\,u_{1\,xxx}^n]\,(\Delta t)^2 + O\,[(\Delta t)^3],$$
$$(T_3^n)^{(\nu+1)} = T_3^n - [\overline{S}_3\,\Delta p\,\Phi_{1\,x}^n + A\overline{S}_3^2\,(\Delta p)^2\,u_{1\,xxx}^n]\,(\Delta t)^2 + O\,[(\Delta t)^3],$$
$$(\omega_2^n)^{(\nu+1)} = -u_{1\,x}^n\,\Delta p - BP\,\Delta p\,(p_*^1 - p_*^0)_{xx}\,\Delta t -$$
$$-\{[2\,BP - (A+F)\,\overline{S}_3 - E\overline{S}_1]\,u_{1\,xxx}^n\,\Delta p + \tag{11}$$
$$+ 2\,BP\,(\overline{p}_* - p_2)\,u_{3\,xxx}^n\}\,\Delta p\,(\Delta t)^2 + O\,[(\Delta t)^3],$$
$$(\omega_*^n)^{(\nu+1)} = -u_{1\,x}^n\,\Delta p - u_{3\,x}^n\,(\overline{p}_* - p_2) - BP\,(p_*^1 - p_*^0)_{xx}\,(\Delta p + \overline{p}_* - p_2)\,\Delta t -$$
$$-\{2\,BP\,(\Delta p + \overline{p}_* - p_2) - [A\,(\Delta p + \overline{p}_* - p_2) + F\,\Delta p]\,\overline{S}_3 -$$
$$- E\overline{S}_1\,\Delta p\}\,u_{1\,xxx}^n\,\Delta p\,(\Delta t)^2 - 2\,BP\,(\Delta p + \overline{p}_* - p_2)(\overline{p}_* -$$
$$- p_2)\,u_{3\,xxx}^n\,(\Delta t)^2 + O\,[(\Delta t)^3],$$
$$(\Phi_3^n)^{(\nu+1)} = AT_3^n + BPp_*^0 -$$
$$- A\,[\overline{S}_3\,\Delta p\,\Phi_{1\,x}^n + A\overline{S}_3\,(\Delta p)^2\,u_{1\,xxx}^n]\,(\Delta t)^2 + O\,[(\Delta t)^3],$$
$$(\Phi_1^n)^{(\nu+1)} = (A+F)\,T_3^n + ET_1^n + BPp_*^0 -$$
$$- \{\Phi_{1\,x}^n\,\Delta p\,[(A+F)\,\overline{S}_3 + E\overline{S}_1] + A\overline{S}_3\,(\Delta p)^2 \times$$
$$\times\, u_{1\,xxx}^n\,(A+F+E)\}\,(\Delta t)^2 + O\,[(\Delta t)^3].$$

It can be concluded from the resultant expressions (11) how the additional information on surface pressure trends modifies the wind fields u_1 and u_3 and the fields of vertical velocities ω_2 and ω_* in the result of a full cycle of pseudoforecasting in a forced adaptation regime. In addition, a validation is obtained for the regime of forced adaptation of the wind field to the temperature (geopotential) and surface pressure fields.

Results of Numerical Experiments

Numerical experiments for testing of the described method for taking into account information on surface pressure trends were carried out with the use of a regional baroclinic prognostic model [5], which replaced the earlier developed routinely used hemispherical model [2].

The numerical experiments were carried out in collaboration with A. D. Spektorman, who was responsible for their preparation and realization. Below we give the most general results. More detailed information will be presented in a separate joint publication.

Table 1

Mean Relative Errors ε and Correlation Coefficients r of Regional Forecasts (for Nine Cases of Initial Data in January and April 1977)

1 Заблаговременность прогнозов, r	2 Вариант прогноза	3 Уровень, мб											
		Земля 4		850		700		500		300		200	
		ε	r	ε	r	ε	r	ε	r	ε	r	ε	r
12	1	0,80	0,66	0,86	0,50	0,87	0,53	0,76	[illegible]	0,51	0,58	0,89	0,55
	2	0,70	0,69	0,79	0,56	0,81	0,55	0,71	[illegible]	0,75	0,62	0,79	0,58
24	1	0,82	0,65	0,88	0,52	0,89	0,49	0,75	0,65	0,76	0,64	0,92	0,57
	2	0,74	0,69	0,80	0,55	0,78	0,57	0,69	0,63	0,71	0,67	0,84	0,62
36	1	0,95	0,63	0,93	0,47	0,90	0,46	0,77	0,62	0,80	0,56	0,90	0,49
	2	0,81	0,66	0,88	0,49	0,86	0,51	0,75	0,63	0,78	0,59	0,86	0,52

KEY:

1. Advance time of forecasts, r
2. Variant of forecast
3. Level, mb
4. Ground

Table 2

Mean Improvements (%) of Relative Errors $\delta\varepsilon$ and Correlation Coefficients δr for Routine Regional Forecasts for 24-Hours in 1977 in Comparison With 1976

Регион 1	2 Улучшение оценки	3 Уровень, мб					
		Земля 4	850	700	500	300	200
5 Ташкентский	$\delta\varepsilon$	4	4	4	6	7	8
	δr	12	4	3	5	7	10
6 Московский	$\delta\varepsilon$	10	7	6	6	6	6
	δr	9	6	6	6	7	7

7 Примечание. Оперативные испытания в московском регионе в 1977 г. проводились только в первом полугодии.

KEY ON NEXT PAGE

KEY:

1. Region
2. Improvement in evaluation
3. Level, mb
4. Ground
5. Tashkentskiy
6. Moskovskiy
7. Note. Routine tests in the Moscow region in 1977 were carried out only during the first half-year

Table 1 gives mean evaluations for a series of regional forecasts on the basis of initialized data computed in two variants: 1) without taking into account information on surface pressure trends and 2) with this information taken into account. It can be concluded from the data in Table 1 that allowance for this information leads to a systematic improvement of forecasts at all surfaces.

In addition to the above-cited methodologically direct comparison of the two variants it is also possible to carry out a direct comparison: it appears desirable to compare the success of forecasts computed in the course of routine tests of regional models applicable to the Moscow and Tashkent regions, separately grouping forecasts in which (during the first stages of the tests) no allowance was made for additional information on surface pressure trends and those forecasts where this information was taken into account. We will cite evaluations for these groups of forecasts (Table 2) on the basis of mass routine data. They integrally indicate, in our opinion, the comparative level of success of forecasts with allowance for additional initial information. We note that the principal conclusions which can be drawn from an analysis of the data in Tables 1 and 2 coincide and are indicative of a positive influence of the considered method for taking into account data on surface pressure trends on the quality of the forecast.

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UDC 551.510.42

INVESTIGATION OF DEPENDENCE OF FALLOUT OF GLOBAL RADIOACTIVE AEROSOLS ON METEOROLOGICAL FACTORS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 58-64

[Article by Candidate of Geographical Sciences A. I. Osadchiy and Candidate of Physical and Mathematical Sciences S. G. Malakhov, Institute of Experimental Meteorology, submitted for publication 3 July 1978]

Abstract: The multiple correlation method is used in studying the dependence of fallout of a global passive aerosol impurity (in the example of global radioactive aerosols -- the products of nuclear tests) on the parameters of atmospheric circulation and meteorological factors. Data are given on the total, partial and multiple coefficients for linear correlation of the intensity of fallout of radioactive aerosols with the principal forms of atmospheric circulation (as defined by A. A. Girs), the characteristics of the pressure-circulation regime of the atmosphere (as defined by L. A. Vitel's) and precipitation for four macroregions.

[Text] The processes of global transport of passive impurities in the atmosphere and the patterns of their fallout are of interest in connection with unceasing tests of nuclear weapons, and also an increase in atmospheric contamination from industry and aviation. At the present time there are many cases when atmospheric contaminations of industrial origin are propagated beyond the limits of countries within whose territories the sources of contamination are situated and which under definite conditions can be entrained into global transport. Contaminations of the stratosphere and also the upper and middle troposphere from the operation of aviation have a tendency to be propagated globally, causing a substantial part of the anthropogenic component of the background contaminations of the atmosphere.

With global propagation of impurities, having the character of passive aerosols, such as the global products of nuclear explosions, meteorological factors are capable of leading to large-scale redistribution of their fallout on the earth's surface [6-8].

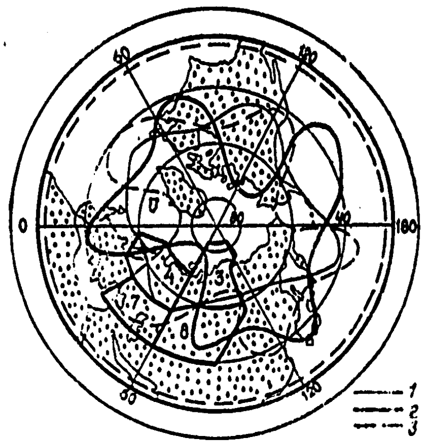

Fig. 1. Diagram of position of high-altitude (AT_{500}) ridges and troughs during macroprocesses W_3 (1), EM_2 (2), CM_1 (3) and map of macroregions.

The purpose of this study is a statistical investigation of the correlation between the intensity of fallout of a passive aerosol impurity (global radioactive aerosols -- the products of nuclear tests in the atmosphere) with some parameters of the state of the atmosphere, describing the effect of large-scale meteorological factors, with large spatial and temporal averaging.

As the basic material for the study we use the mean monthly values of the total beta-activity of global fallout of radioactive products of nuclear explosions, territorially averaged for natural synoptic regions 3, 4, 7, 8 (Fig. 1), defined in accordance with [1]. Measurements of the mean monthly global radioactive fallout were carried out at a number of places in the Soviet Union during 1963-1972 [6-8].

For the purpose of eliminating the influence of the seasonal effect on the process of entry of radioactive aerosols from the stratosphere into the troposphere and the dependence of the intensity of fallout on the reserve of radioactive products in the atmosphere, in this paper we examine the ratios of the mean monthly (mean seasonal) values of fallout of radioactive aerosols, averaged for the above-mentioned regions, to the mean monthly (mean seasonal) values of radioactive fallout, averaged over the territory of the USSR.

Table 1, in relative units (relative to the mean value), gives the values of the extremal mean monthly fallout of passive radioactive aerosols (in the territory of the USSR) in regions 3, 4, 7, 8. It follows from Table 1 that considerable changes in the relative intensity of global fallout of radioactive aerosols are observed in all the considered regions. There is every basis for assuming that these variations are caused by the peculiarities of atmospheric circulation determining the processes of entry of radioactive aerosols from the stratospheric reservoir into the lower layers of the troposphere and also their tropospheric transport and the meteorological factors determining the intensity of the processes of fallout of radioactive aerosols from the atmosphere onto the earth's surface.

Table 1

Extremal Values of Mean Monthly (Mean Seasonal) Fallout of Radioactive Aerosols in Regions 3, 4, 7 and 8 Relative to Corresponding Fallout in the Area of the USSR During the Period 1963-1972

1 Район	2 Минимум	3 Месяц (сезон)	4 Год	5 Максимум	6 Месяц (сезон)	7 Год
3	0,29 (0,56)	Февраль 8 Зима 9	1969 1969	1,92 (1,56)	Июнь 13 Лето 10	1964 1964
4	0,36 (0,44)	Январь 11 Зима 9	1972 1972	1,29 (1,11)	Июль 12 Осень 14	1969 1967
7	0,69 (0,91)	Июль 12 Лето 10	1971 1971	2,02 (1,66)	Апрель 15 Весна 16	1963 1963
8	0,67 (0,83)	Январь 11 Лето 10	1963 1963	1,67 (1,63)	Апрель 15 Весна 16	1971 1971

KEY:
1. Region
2. Minimum
3. Month (season)
4. Year
5. Maximum
6. Month (season)
7. Year
8. February
9. Winter
10. Summer
11. January
12. July
13. June
14. Autumn
15. April
16. Spring

The peculiarities of atmospheric circulation and the macroscale effect of the combined action of meteorological factors can be described in general using the pressure-circulation characteristics of the atmosphere. The characteristics of the pressure-circulation regime, cited in the studies of L. A. Vitel's and S. V. Kasogledova [1, 2], characterize pressure fields at sea level and for the most part reflect the conditions for the fallout of atmospheric impurities onto the earth's surface as a result of meteorological factors, the most important of which are precipitation, turbulent transfer and descending air movements. In describing the state of the atmosphere using the G. Ya. Vangengeym-A. A. Girs [3] circulation forms the emphasis is on the distribution and nature of long waves in the middle and upper troposphere. The nature of the long waves in the troposphere is closely tied in to the position and intensity of centers of action in the atmosphere, the geographic position of jet streams, the peculiarities of cyclogenesis and surface frontal zones, macroscale distribution of ascending and descending vertical air movements and other large-scale meteorological factors determining the horizontal and vertical transport of air masses and with them, passive impurities. Thus, the Vangengeym-Girs forms of atmospheric circulation for the most part reflect the global transport of radioactive aerosols in the lower stratosphere and in the upper troposphere and the conditions for entry of radioactive aerosols into the lower layers of the troposphere.

Table 2

Correlation of Intensity of Fallout of Global Radioactive Aerosols and Precipitation and Frequency of Recurrence of Forms of Surface Pressure Field. N -- Number of Cases

1 Район		Осадки 2		3 Приземное [illegible]											
				циклоническое 4			антициклоническое 5			глубокий циклон 6			мощный антициклон 7		
	N	K	ΔK	N	K	ΔK	N	K	ΔK	N	K	ΔK	N	K	ΔK
3	120	0,11	0,09	96	0,12	0,10	96	−0,13	0,10	52	[illegible]	0,11	[illegible]	0,09	0,15
4	120	0,21*	0,09	96	0,30*	0,09	96	−0,31*	0,09	81	[illegible]*	0,16	52	−0,12	0,14
7	120	0,32*	0,08	96	0,17	0,10	96	−0,17	0,10	23	−0,02	0,21	46	−0,03	0,15
8	120	0,23*	0,09	96	0,06	0,10	96	−0,15	0,19	35	0,06	0,17	63	−0,11	0,12

8 * Статистически значимые (с доверительной вероятностью 95%) коэффициенты корреляции.

KEY:

1. Region
2. Precipitation
3. Surface pressure field
4. Cyclonic
5. Anticyclonic
6. Deep cyclone
7. Powerful anticyclone
8. Statistically significant (with confidence interval 95%) correlation coefficients

Table 2 gives the coefficients (K) and the probable errors (Δ K) of the linear correlation coefficients of the mean monthly values of radioactive fallout with precipitation (during 1963-1972) and the forms of the pressure fields as defined by L. A. Vitel's (during 1963-1970) for the third, fourth, seventh and eighth macroregions (Fig. 1). In the computations we used the territorially averaged (for the corresponding macroregions) mean monthly quantities of precipitation selected parallely with samples of fallout of global radioactive aerosols. The characteristics of the surface pressure fields for the considered regions were taken from [1, 2]. In accordance with the method described in [1], pressure situations related to a deep (with a pressure at the center $\leqslant$ 990 mb), temperate or weak cyclone, with a trough or periphery of a cyclone, and also with a blurred cyclonic field or with a saddle nature of the low-pressure corridor, were assigned to a cyclonic pressure field. We assigned to the anticyclonic pressure field those situations related to a powerful (with a pressure at the center $\geqslant$ 1035 mb), moderate or weak anticyclone, with the ridge or periphery of an anticyclone, and also with a blurred anticyclonic field or with a saddle character of the high-pressure field.

It follows from Table 2 that the positive correlation between the intensity of fallout of radioactive aerosols, the quantity of precipitation and the frequency of recurrence of cyclonic forms of the pressure field (accordingly a negative correlation between the intensity of fallout and anticyclonic forms of the pressure field) is observed for all the defined microregions. It can be postulated that this correlation for the most part is a result of the washing-out effect of precipitation, in turn associated with the cyclonic pressure field. In regions 3, 7, 8 the intensity of the fallout of radioactive aerosols is unrelated to the characteristics of the pressure formations; in region 4 there is a positive correlation between the intensity of fallout of radioactive aerosols and the frequency of occurrence of deep cyclones, the reasons for which will be considered below.

Table 3 gives the coefficients and probable errors of the linear correlation coefficients for the mean monthly values of the intensity of fallout of the radioactive products of nuclear explosions with a frequency of recurrence of forms of atmospheric circulation as given by A. A. Girs. Data on the frequency of recurrence of the forms of atmospheric circulation W, E. C, З, M_1, M_2 for the period 1963-1972 were taken from [3]. Circulation processes, relating to forms W and C, are characterized by the presence in the troposphere of waves of relatively small amplitude moving from west to east. The processes relating to circulation forms E, C, M, M_2 are types of the meridional state of atmospheric circulation and in the troposphere they correspond to stationary waves of a great amplitude whose positioning, in accordance with [3], is given in Fig. 1.

It follows from Table 3 that the intensity of radioactive fallout in the considered regions is most closely related to circulation forms З , M_1, M_2 of the Pacific Ocean-American (second) sector. It is possible to trace a clear tendency of the dependence of the sign (direction of influence) on the geographic position of the region. For regions 3 and 4, situated to the north, the correlation between the intensity of fallout of radioactive aerosols and circulation forms З and M_1 is negative or there is no dependence, whereas for the more southerly regions 7 and 8 the correlation of the intensity of fallout with these circulation forms is positive. The form of atmospheric circulation M_2 is related positively with the intensity of fallout of radioactive aerosols in regions 3 and 4 and negatively in regions 7 and 8.

Table 3

Correlation Between Intensity of Fallout of Radioactive Aerosols and the Principal Forms of Atmospheric Circulation

1 Район	2 Число случаев	W		C		E		З 3		M_1		M_2	
		K	ΔK	K	ΔK	K	ΔK	K	ΔK	K	ΔK	K	ΔK
3	120	0,04	0,09	−0,22*	0,09	0,09	0,09	−0,22*	0,09	−0,28*	0,08	0,39*	0,08
4	120	0,15	0,09	−0,05	0,09	−0,07	0,09	−0,21*	0,09	0,02	0,09	0,13	0,09
7	120	−0,11	0,09	0,25*	0,08	−0,06	0,09	0,12	0,09	0,18	0,09	−0,25*	0,09
8	120	−0,0	0,09	−0,05	0,09	0,05	0,09	0,18	0,09	0,20*	0,09	−0,29*	0,08

4 * Статистически значимые (с доверительной вероятностью 95%) коэффициенты корреляции.

KEY:

1. Region
2. Number of cases
3. З = 3
4. Statistically significant (with confidence interval 95%) correlation coefficients

The data from Table 3 make it possible to note that the intensity of the fallout of global radioactive aerosols from the atmosphere in the northern regions with circulation forms 3 and M_1 is relatively lower and with M_2 is relatively higher, whereas in the southern regions the dependence is the reverse. The fact of a dependence of the intensity of the processes of fallout of global passive atmospheric impurities of stratospheric origin in the macroregions situated in the first (Atlantic-Eurasian) sector on the atmospheric processes transpiring in the second (Pacific-American) sector is unexpected and exceedingly interesting. The detection of the reasons for this dependence should be the subject of special investigations going beyond the framework of this study.

As a result of the dependence of the process of entry of radioactive aerosols from the stratosphere into the troposphere and the process of their transport in the troposphere on the forms of atmospheric circulation, on the one hand, and the dependence of meteorological factors determining the intensity of fallout of radioactive products from the atmosphere onto the earth's surface on the pressure-circulation characteristics of the atmosphere, on the other, the data in Tables 2 and 3 do not make it possible to discriminate the most important meteorological factors determining large-scale spatial-temporal fluctuations of the fallout of global radioactive aerosols. Therefore, it is interesting to examine the special correlation coefficients (K_x, K_y, K_z) and the multiple correlation coefficients (R) for the forms of atmospheric circulation and meteorological factors favoring the fallout of radioactive aerosols from the atmosphere with the intensity of their fallout in the considered macroregions (Table 4). As the parameters determining the positive correlation between the intensity of fallout of radioactive aerosols with meteorological factors we examined precipitation (x), the frequency of recurrence of cyclonic forms of the pressure field (y) and the type of circulation $M_2(z)$ for the third and fourth regions, precipitation (x), frequency of recurrence of cyclonic forms of the pressure field (y) and type of circulation $M_1(z)$ for the seventh and eighth regions.

Table 4

Correlation Coefficients (Partial and Multiple) for the Intensity of Fallout of Global Radioactive Aerosols With Meteorological Factors Favoring the Fallout of Impurities from Atmosphere

1 Район	K_x	K_y	K_z	K
3	0,07	0,01	0,40*	0,42*
4	0,03	0,21	0,21*	0,50*
7	0,32*	—0,02	0,23*	0,39*
8 .	0,23*	—0,04	0,11	0,26*

* Статистически значимые (с доверительной вероятностью 95%) коэффициенты корреляции.

*Statistically significant (with confidence coefficient 95%) correlation coefficients

On the basis of the data in Table 4 it can be concluded that regions 3, 4, 7 and 8 with respect to the effectiveness of meteorological factors differ from one another and these differences are quite clearly expressed for the northern and southern regions. In the northern regions (regions 3 and 4) the principal factor favoring an increase in the intensity of radioactive fallout is atmospheric circulation (partial correlation coefficients $K_{z3} = 0.40$ and $K_{z4} = 0.21$). The role of precipitation is insignificant ($K_{x3} = 0.07$; $K_{x4} = 0.03$). The influence of cyclonic forms of the surface pressure field on the process of fallout of global radioactive aerosols from the atmosphere is absent ($K_{y3} = 0.01$), at the same time that in region 4 cyclonic activity is associated with an intensification of fallout ($K_{y4} = 0.21$). The reason for this is the peculiarities of cyclonic activity in these regions. In region 3, relating to the Siberian sector (according to the B. L. Dzerdzeyevskiy classification) [4], cyclonic activity as a rule does not lead to the transport of air masses from southern regions containing higher concentrations of radioactive aerosols [5].

In region 4, relating to the European sector, cyclonic activity to a considerable degree is associated with a meridional southern type of circulation in which in the territory of region 4 there is a transport of air masses from the regions of their entry from the stratosphere into the troposphere [5]. Thus, the correlation of the intensity of fallout of radioactive aerosols with the frequency of recurrence of cyclonic forms of the surface pressure field in region 4 indicates that for this region the transport of radioactive aerosols in the lower layers of the atmosphere is significant.

In the southern regions 7, 8 the principal factors favoring the intensification of fallout of radioactive aerosols is precipitation (K_{x7} = 0.32, K_{x8} = 0.23) and the forms of atmospheric circulation (K_{z7} = 0.23, K_{z8} = 0.11). The role of cyclonic forms of the surface pressure field is insignificant (K_{y7} = 0.02, K_{y8} = -0.04). The influence of precipitation on the intensity of fallout of radioactive aerosols in these regions is evidently a result of the constant presence of relatively high concentrations of radioactive aerosols in the air due to their entry from the stratospheric reservoir. The oppositeness of sign of the effect of circulation factors on the intensity of fallout of global radioactive contaminations in southern (7,8) and northern (3, 4) regions makes it possible to assume that the tropospheric transport of radioactive aerosols from the southern into the northern regions can be substantial in comparison with the direct transport of radioactive impurities from the stratosphere in northern regions.

Thus, all the considered meteorological factors should exert an influence on the intensity of fallout of passive aerosols from the stratospheric reservoir. The direction of the effect and the effectiveness of their influence in different regions can be different. The intensity of fallout of global radioactive aerosols has a greater influence on the peculiarities of processes of global transport and entry of aerosols from the upper layers of the troposphere and the stratospheric reservoir into the lower layers than the meteorological factors directly washing impurities out of the atmosphere (precipitation) or favoring the process (surface cyclonic forms of pressure fields).

It should be noted that all the correlations considered in this study as linear under real conditions are nonlinear and therefore the reliability of their quantitative characteristics is evidently small. The results are of interest for the most part from the point of view of clarifying the most general trends in the global redistribution of fallout of passive aerosol impurities from the stratospheric reservoir or from the upper layers of the troposphere.

Conclusions

In the case of a great spatial and temporal averaging there is a statistical correlation between the intensity of fallout of passive aerosols from the atmosphere -- the global products of nuclear explosions and meteorological factors.

The most important influence on the intensity of fallout of global aerosols from the stratospheric reservoir or from the upper layers of the troposphere is from the peculiarities of atmospheric circulation, in particular, the geographic distribution of troughs and ridges of tropospheric long waves. At the same time, in regions of the entry of aerosols from the stratospheric reservoir into the troposphere a definite role can be played by precipitation, and evidently by other meteorological factors directly favoring the fallout of passive aerosol impurities from the troposphere.

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DEPENDENCE OF AEROSOL ATTENUATION OF OPTICAL RADIATION ON AIR HUMIDITY

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 65-69

[Article by Candidate of Physical and Mathematical Sciences V. L. Filippov, submitted for publication 7 July 1978]

Abstract: The article gives the results of a statistical analysis of experimental data on changes in the spectral optical density of aerosols in dependence on the relative and absolute humidity and air temperature for the transitional seasons of the year (spring-autumn). Confirmation was obtained for known data on absence of an interrelationship between aerosol attenuation of radiation and absolute humidity. The close correlation between the optical state of aerosol and relative humidity makes it possible to speak of the possibility of creating corresponding prognostic models.

[Text] Investigations carried out during recent years (for example, see [1-6,8]) have demonstrated the fundamental importance of humidity in forming the optical state of substances dispersed in the air. Most of these investigations have been devoted to the problems involved in a qualitative interpretation of the corresponding experimental data and very frequently there is no possibility, at the present-day level of knowledge, of obtaining any quantitative information. On the other hand, despite the complexity of the discussed problem, its practical significance explains the attempts of a number of authors (for example, see [8], and others) to formulate a model of prediction of visibility in the atmosphere as a function of relative humidity with the use of some stylized representations of the physicochemical properties of natural aerosols (for example, it is postulated that all particles in a dry state have an identical chemical composition, the transformation of aerosols is not dependent on their size spectrum, etc.). Such an approach, without question, is not now ensured with the initial information and cannot even be argued with the mentioned degree of reliability from the point of view of the completeness of the considered concepts on the mechanism of transformation of aerosol matter [1, 2, 6]. Taking this circumstance into account and also adhering to the conclusions in [6], in which it was demonstrated that with a differentiated approach to determination of the type of "optical weather" [4] it is possible to establish, at least, the stochastic regularities between the variations of the spectral coefficients of aerosol attenuation α_λ and the relative humidity f, the corresponding experimental investigations were continued and we give below some results of their analysis.

Table 1

λ, мкм (1)	$f = 30 \div 100\%$	$f = 60 \div 100\%$	
	$r_{\lambda f}$	$\rho_{\lambda T}$	$\rho_{\lambda e}$
0,55	0,85±0,013	−0,22±0,050	−0,03±0,053
0,83	0,84±0,013	−0,26±0,049	−0,08±0,052
1,06	0,83±0,014	−0,28±0,048	−0,10±0,052
1,18	0,81±0,015	−0,29±0,048	−0,11±0,052
1,66	0,82±0,015	−0,32±0,047	−0,14±0,052
2,09	0,81±0,015	−0,32±0,047	−0,15±0,051
3,97	0,81±0,015	−0,32±0,047	−0.14±0,052
2 Количество реализаций	500	350	

KEY:

1. microns
2. Number of records

Table 2

f %	λ мкм (1)						
	0,55	0,83	1,06	1,18	1,66	2,09	3,97
95	0,0465	0,0492	0,0485	0,0450	0,0402	0,0369	0,0345
85	0,0146	0,0165	0,0160	0,0150	0,0136	0,0119	0,0114
75	0,0064	0,0078	0,0064	0,0078	0,0078	0,0064	0,0045
65	0,0064	0,0074	0,0053	0,0053	0,0053	0,0037	0,0037
55	0,01	0,0112	0,01	0,0087	0,0071	0,0051	0,0051
45	0,0087	0,01	0,0071	0,0071	0,0051	0,0051	0,005

KEY:

1. microns

The initial material for this discussion was the results of optical sounding of the surface layer of the atmosphere under controlled meteorological conditions, carried out by the authors in the development of a program of complex optical-microphysical experimentation [5] during the period from 1973 to 1976.

Table 3

λ мкм (1)	$K'_{0\lambda}$	$K_{1\lambda}$	$K'_{2\lambda}$
0,55	0,18	0,96	11,7
0,83	0,12	1,03	13,00
1,06	0,07	0,99	13,0
1,18	0,05	0,92	13,5
1,66	0,03	0,77	12,7
2,09	0,02	0,70	13,5
3,97	0,02	0,66	14,3

KEY:
1. microns

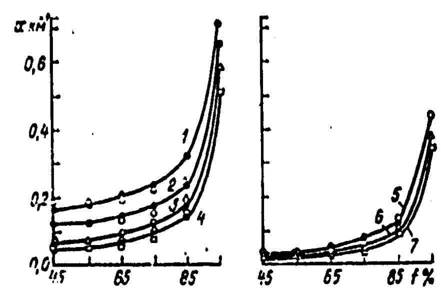

Fig. 1. Mean profiles $\bar{\alpha}_\lambda$ (f) for different parts of spectrum. 1) 0.55; 2) 0.83; 3) 1.06; 4) 1.18; 5) 1.66; 6) 2.09; 7) 3.97μm.

As a characteristic of the optical state of the aerosol we used data on the volume coefficients of aerosol attenuation in the windows of atmospheric transparency in the range α = 0.55-14μm (α = 0.55; 0.83; 1.06; 1.18; 1.66μm, etc. [4]) for weather conditions typical for the transitional seasons of the year (spring-autumn). These materials in the form of samples of n spectra were subjected to statistical processing, the results of which for λ = 0.55-3.97μm are illustrated in Tables 1-3.

The results of a correlation analysis are given in Table 1, which illustrates, in particular, the interrelationship between variations of the coefficients of aerosol attenuation $\Delta\alpha_\lambda$, and accordingly, the relative Δ f and absolute humidity Δe, air temperature Δ T.

It should be noted that the interrelationship between the variations $\Delta\alpha_\lambda$ and Δ f has a nonlinear character; therefore, the closeness of the correlation between the indicated parameters must be evaluated using the correlation ratio $\eta_{\lambda f}$. In Table 1 the results of computations of $\eta_{\lambda f}$ are given for the range f $\in$ 30-100%, which for all practical purposes takes in the entire region of change in relative humidity under conditions typical for the temperate latitudes; here also we have given the matrix of errors for the considered correlation ratio. It can be seen that in the mentioned region for the investigated windows of transparency a close interrelationship is retained between the changes $\Delta\alpha_\lambda$ and Δ f ($\eta_{\lambda f}$>0.8), which justifies the optimism of the authors of [5, 6, 8], indicating the possibility of stochastic prediction of the optical characteristics of an atmospheric aerosol on the basis of data on air humidity.

Table 1 also gives the correlation coefficients $\rho_{\lambda e}$, $\rho_{\lambda T}$, determining the closeness of the interrelationship, between the changes $\Delta\alpha_\lambda$ and Δ e, $\Delta\alpha_\lambda$ and Δ T respectively, which have extremely low values; here also for the considered correlation coefficients we have given the matrices of errors. Thus, the establishment of clearer correlations between aerosol turbidity of the natural atmosphere and absolute humidity, noted in a number of studies, as rightfully observed by the authors of [2], evidently is of an indirect nature, which contradicts the conclusions given in [3].

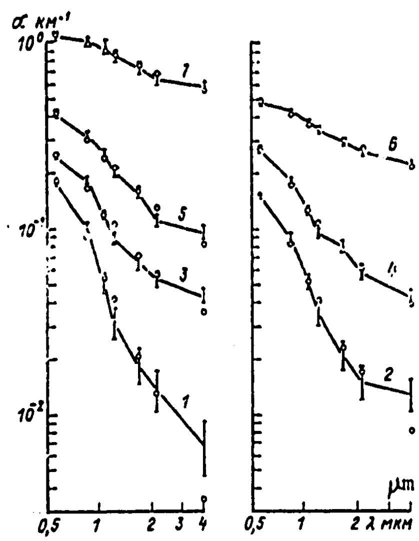

Fig. 2. Mean dependences $\bar{\alpha}(\lambda)$ (Table 4 illustrates weather conditions).

For the purpose of determining the type of dependence $\alpha_\lambda(\underline{f})$ for each of the parts of the spectrum we determined the mean profiles $\bar{\alpha}_\lambda(f)$, which is illustrated in Fig. 1. In order to decrease the influence of the error in measuring relative humidity, which in our experiments on the average was equal to ±3%, it was deemed desirable to carry out averaging of data for α_λ for Δ f intervals with a width of 6%, and the mean value

$\bar{\alpha}_\lambda(f_i)$ obtained in this way is related to the middle of the considered interval, which on the $\bar{\alpha}_\lambda(f)$ profiles is denoted by dots. (The regression equation $\bar{\alpha}_\lambda(f)$ in such a form henceforth is approximated by a corresponding analytical expression). The corresponding error matrices $d_{\lambda f}$ for each profile are given in Table 2.

Table 4

№ спектра (a)	Тип оптической погоды (b)	f, %	S_M км	n_0	n_1	n_2
1	(c) Дымка весны и осени	30—60	15—20	0,00	0,35	2,0
2	(d) то же	50—75	20—50	0,03	0,35	2,0
3	«	50—75	10—20	0,09	0,44	1,5
4	«	75—90	10—20	0,10	0,40	1,5
5	«	90—95	5—15	0,07	0,54	1,1
6	(e) Туманная дымка	90—100	5—10	0,22	0,57	0,6
7	(d) то же	90—100	1—5	0,06	0,79	0,4

KEY:

a) No of spectrum
b) Type of optical weather
c) Spring and autumn haze
d) Same
e) For haze

Taking into account that the dependence $\alpha_\lambda(f)$ has a nonlinear character, the analytical expression for the regression equation, from our point of view, is conveniently determined in the form

$$\alpha'_\lambda(f) = K_{0\lambda} + K_{1\lambda}\left(\frac{f}{100}\right)^{K_{2\lambda}}, \qquad (1)$$

where $K_{0\lambda}$, $K_{1\lambda}$, $K_{2\lambda}$ are empirical coefficients dependent on wavelength. The values of the latter, determined by the least squares method for λ = 0.55-3.97 μm, are given in Table 3. The results of reconstruction of the initial $\alpha_\lambda(f)$ profiles by means of expression (1) are shown by symbols (circles, triangles, squares respectively) in Fig. 1. In turn, in [5], on the basis of a complex analysis of data from optical and microphysical measurements, the authors noted a tendency to a parallel shift of individual records of the dependence $\alpha_\lambda(f)$ (in the range $f \in$ 30-95%), which indicates a possibility of using an expression of type (1) for predicting the nature of transformation of the optical state of aerosol formations under specific weather conditions.

It should be noted that the considered type of dependence $\alpha_\lambda(f)$ is most characteristic for conditions of quasistationary anticyclonic formations in the temperate latitudes formed in continental temperate air.

Thus, the state of the aerosol, and accordingly, to a considerable degree also the optical properties of the atmosphere, is regulated by a whole complex of geophysical factors; under definite conditions the decisive link in this process is relative humidity. Therefore, any prerequisites on construction of the corresponding models must rest, in particular, on information on a particular meteorological parameter. Taking this requirement into account, it was deemed desirable to trace in the development of the ideology [4] the change in the spectral dependences α_λ with stipulated limits f and the meteorological range of visibility S_M. For this purpose for different regions of change in the indicated characteristics (see Table 4) we computed the mean profiles of the coefficients of aerosol attenuation $\bar{\alpha}_\lambda$ (Fig. 2); here also for each profile we have graphically shown the corresponding matrix of errors. An analysis of the mean profiles $\bar{\alpha}_\lambda$, in particular, indicates (for example, compare curves (1,2) and (3,4) (Fig. 2) that for practical purposes the generalization of experimental data must be carried out in relatively broad limits of f and S_M. Nevertheless, in this study in a description of the mean statistical curves $\bar{\alpha}_\lambda$ on the basis of an expression from [7]

$$\alpha_\lambda = \alpha_{0,55} \; (n_0 + n_1 \lambda^{-n_2}) \tag{2}$$

for completeness of the primary information the corresponding coefficients n_0, n_1, n_2 were computed by the LSM for all f and S_M limits (see Table 4). Their values are represented in this same table. The corresponding results of computations of $\alpha_{\lambda,\Delta f}$ on the basis of $\{n\}$ and (2) are represented by the small circles in Fig. 2.

Thus, special investigations of the interrelationship of the coefficients of aerosol attenuation and air humidity, carried out by the authors, again confirmed the decisive role of relative humidity in the formation of the optical state of the disperse phase. The resulting statistical dependences unambiguously indicate the fundamental possibility of predicting the optical state of an aerosol on the basis of information on the meteorological parameters.

The latter conclusion sounds more convincingly taking into account that the first [6] and present stages in the investigation were carried out in different regions.

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UDC 556.(51/.54+166)

CRITERION OF CORRESPONDENCE BETWEEN THE DISTRIBUTION CURVES OF PROBABILITIES OF MAXIMUM WATER DISCHARGE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 70-76

[Article by Candidate of Technical Sciences D. M. Mamatkanov and N. Sh. Kudaybergenov, Kirgiz Scientific Research Power Section, submitted for publication 5 June 1978]

Abstract: The authors demonstrate the inadequate correctness of use of existing criteria for correspondence to the empirical distribution curve of maximum runoff in the presence of observations of limited duration, which leads to systematic errors in computations. Using as a point of departure the limited nature of the observations and the peculiarities of construction of the empirical curves of the probabilities of maximum discharge of water, the authors propose a correlation criterion which is a modification of the Cramer-Mises criterion.

[Text] On the basis of the computed maximum water discharge it is possible to calculate the spillways of dams, the handling capacities of bridges, canals and other hydraulic structures. This same parameter to a great extent determines both the construction and the makeup of the entire hydraulic complex. Therefore, the computed maximum water discharge is a highly important parameter of the hydraulic complex, and the correctness of its determination governs the safety of the structures and their economic effectiveness. Accordingly, the objective of hydrological computations of the maximum discharges is a combination of the requirements of safety and economic effectiveness of the hydraulic structures.

The combination of these two requirements can be accomplished by means of use of the principle of stochastic computations, based on determination of the hydrological conditions for the formation of maximum runoff, on the one hand, and on allowance for its stochastic excess, on the other hand [11].

With use of the stochastic method for computations of the maximum water discharge the primary and most important problem is the correspondence between the empirical and theoretical (hypothetical) probability distribution curves, on the basis of which it is possible to judge the degree of guarantee of safety of structures in connection with the use of different computed probabilities of excess.

This problem has still been inadequately solved and requires further improvement of existing and development of new, simpler and more effective methods from the practical point of view.

As is well known, the construction of an empirical curve for the probability of maximum water discharge is accomplished by means of determination of the probabilities of excess of its elements, represented in the form of the variation series

$$x_1 > x_2 > x_3 > \ldots > x_n, \tag{1}$$

according to the S. N. Kritskiy and M. F. Menkel' formula

$$P_m = \frac{m}{n+1}, \tag{2}$$

where m = 1, 2,..., n is the sequence number of the elements of the variation series (1) in the sequence of decrease.

This formula is recommended by the "Instructions on Determination of Computed Hydrological Characteristics" SN 435-72, since its use with a limited volume of a sample leads to some reserve in the computations, satisfying practical requirements.

As the theoretical (hypothetical) probability curve P(x), approximating the empirical function $P_n(x)$, it is possible to use a two- and three-parameter gamma distribution or a log-normal distribution, a Gumbel' distribution, a system of Johnson distribution curves, etc.

For selecting the types of hypothetical distribution curves of probabilities in the theory of probabilities and mathematical statistics specialists have developed several criteria of the correspondence of empirical and theoretical distribution curves, such as the Kolmogorov and Cramer-Mises ($n\omega^2$) tests and the Pearson chi-square test. The application of these tests to the probability distribution curves for the observed data on maximum water discharge for a limited duration is inadequately correct and gives systematic errors including the following.

All existing correspondence criteria have been developed applicable to the empirical function (stepped curve) for the distribution of the observed data $P_n(x)$, the probability of excess of the elements of the sample, which is determined using the formula for the classical theory of probabilities,

$$P_m = \frac{m}{n}. \tag{3}$$

As is well known, this formula gives acceptable results only with sufficiently great n and is applicable to the terms of a variation series situated in a zone adjacent to the center of the distribution. Indeed, the very center of the distribution, in the case of limited n, is displaced. For terms of a ranked series, occupying the last place, with any finite n we will always have $P_m = 1$, whereas for the first term of the series $P_m = 1/n$, which, to be sure, is an extremely rough evaluation. In addition, in all existing computation criteria for all practical purposes one seeks an agreement between the theoretical and empirical distribution curves, the values of the probabilities of excess of whose sample elements are determined using the expression

$$P_m = \frac{2m-1}{2n} = \frac{m-0,5}{n}, \tag{4}$$

that is, the A. Khazen formula.

It is also known that this formula under conditions of a restricted volume of the sample always gives reduced values of the probabilities of high and exaggerated values of the probabilities of low runoff values in comparison with formula (2), that is, understates the ordinates of the runoff distribution curves in the case of small probabilities and exaggerates the ordinates in the case of high probabilities. This can also be seen from Table 1.

Accordingly, the use of the above-mentioned criteria applicable to the distribution curves for maximum water discharges in the case of a restricted volume of the sample will always give systematic errors, since one seeks an agreement between the theoretical curves and the empirical curve, constructed using formula (4), and not (2). The errors will go in the direction of an understatement of the maximum runoff values, which is related to a decrease in the safety guarantee for hydraulic structures. Therefore, the computation criterion for the correspondence of empirical and theoretical distribution curves invariably must take into account the form of the formula used in obtaining the empirical distribution functions for the observed data for the regimes of hydrological characteristics. Taking this into account, now we will discuss the problems involved in obtaining the correspondence criteria for the empirical and theoretical distribution functions for the maximum water discharge applicable to the formula for determining probabilities (2).

Table 1

Comparison of Empirical Probabilities (in %) Computed by Kritskiy-Menkel' (2) and Khazen (4) Formulas

Formula	$n = 10$			$n = 20$			$n = 40$			$n = 60$		
	$m=1$	$m=2$	$m=n$	$m=1$	$m=2$	$m=n$	$m=1$	$m=2$	$m=n$	$m=1$	$m=2$	$m=n$
(2)	9,09	18,18	90,91	4,76	9,52	95,54	2,44	4,88	97,56	1,64	3,28	98,36
(4)	6,73	16,35	93,27	3,43	8,33	96,56	1,72	4,21	98,27	1,16	2,81	98,84

As the basic derived criterion we used the criterion $n\omega^2$, expressed using the formula

$$n\omega^2 = \frac{1}{12n} + \sum_{m=1}^{n} \left[P(x_m) - \frac{2m-1}{2n} \right]^2, \tag{5}$$

since it, being based on the basis of the observed data, more fully uses the information contained in them and considerably more rapidly converges to a limiting law, especially in the region of large ω^2 values, which is important in a stochastic evaluation [2, 10]. Accordingly, the proposed criterion is a modification of the Cramer-Mises criterion.

Since these problems in the theory of probabilities and mathematical statistics have been developed applicable to the distribution functions of excess probabilities, we will also use this approach, which exerts no influence on the results of the problem to be solved and only creates conveniences in the integration.

The considered Cramer-Mises criterion applicable to our problem is written in general form:

$$\omega^2 = \int_{-\infty}^{\infty} [P(x) - P_n(x)]^2 \, dP(x); \tag{6}$$

and $dP(x) = P'(x)dx$, that is, it is assumed that the function $P(x)$ has the derivative (probability density function) $P'(x)$. Otherwise it is necessary to understand integration in the Stieltjes sense. Here $P_n(x)$ is the empirical probability of nonexcess of the elements x of an existing sample of the volume n. These values are determined using formula (2) and are used as their true values. $P(x)$ is the hypothetical probability of the value $P_n(x)$, being a random function dependent not only on the x value, but also on the type of the tested hypothetical distribution curves and their statistical parameters.

Then, assume that there is a variation series of the maximum water discharge with the duration n

$$x_1 < x_2 < x_3 < \ldots < x_n. \tag{7}$$

In this case the x values are essentially positive, that is, $x \geqslant 0$.

If the integration limit for the integral (5) is broken down into intervals $(-\infty, x_1)$, $(x_1, x_2), \ldots, (x_{n-1}, x_n)$, $(x_n, +\infty)$, then expression (6) can be written as follows:

$$\omega^2 = \int_{-\infty}^{x_1} [P(x) - P_n(x)]^2 \, dP(x) + \sum_{m=1}^{n-1} \int_{x_m}^{x_{m+1}} [P(x) - P_n(x)]^2 \, dP(x) + \tag{8}$$

$$+\int_{x_n}^{+\infty}[P(x)-P_n(x)]^2\,dP(x). \tag{8}$$

We will examine the solution (8) applicable to the formulas for determining the empirical probability of nonexcess (2), which with stipulated n divides the interval (0.1) by (n+1) equal parts with the width 1/(n+1). In this case at the discontinuity point $x = x_m$ the function $P_n(x)$ passes in a jump from the value 2 m - 1/2(n + 1) (in the interval $x_{m-1} \leqslant x \leqslant x_m$) to the value 2 m + 1/2(n + 1), maintaining the latter value in the next interval. With the above taken into above, the empirical function (stepped curve) for the distribution of hydrological series of maximum water discharge (7) is determined by the equations

$$
\begin{aligned}
&P_n(x)=0 && \text{when } x\leqslant 0\\
&P_n(x)=\frac{1}{2(n+1)} && \text{when } 0\leqslant x\leqslant x_1\\
&P_n(x)=\frac{2m-1}{2(n+1)} && \text{when } x_m\leqslant x\leqslant x_{m+1}\\
& && \text{where } m = 1,2,\dots,(n-1)\\
&P_n(x)=1 && \text{when } x\geqslant x_n
\end{aligned}
\tag{9}
$$

In this case, from (8), with (9) taken into account, we have:

$$\int_{-\infty}^{x_1}[P(x)-P_n(x)]^2\,dP(x)=\frac{[P(x)-P_n(x)]^3}{3}\Bigg|_{x=-\infty}^{x=x_1}=$$

$$=\frac{1}{3}\left[P(x_1)-\frac{1}{2(n+1)}\right]^3;$$

$$\int_{-x_m}^{x_{m+1}}[P(x)-P_n(x)]^2\,dP(x)=\frac{[P(x)-P_n(x)]^3}{3}\Bigg|_{x=x_m}^{x=x_{m+1}}=$$

$$=\frac{1}{3}\left[P(x_{m+1})-\left(\frac{m}{n+1}+\frac{1}{2(n+1)}\right)\right]^3;$$

$$\int_{x_n}^{+\infty}[P(x)-P_n(x)]^2\,dP(x)=\frac{[P(x)-P_n(x)]^3}{3}\Bigg|_{x=x_n}^{x=+\infty}=$$

$$=\frac{1}{3}\left[P(x_n)-\left(\frac{n}{n+1}+\frac{1}{2(n+1)}\right)\right]^3.$$

Substituting the determined values of the integrals into (8) and solving them, we obtain the final computation formula

$$(n+1)\,\omega^2 = \frac{n}{12\,(n+1)^2} + \sum_{m=1}^{n}\left[P(x_m) - \frac{m}{n+1}\right]^2. \qquad (10)$$

The computed expression (10) can be called the criterion of correspondence between the empirical and theoretical (hypothetical) distribution curves for maximum runoff or simply $(n + 1)\omega^2$, which takes into account the maximum possible information contained in a sample of limited volume, with formula (2) taken into account, adopted for constructing the empirical function of the actually observed data.

It can be seen from (10) that with a full correspondence between the hypothetical and empirical distribution curves we have

$$(n+1)\,\omega^2 = \frac{n}{12\,(n+1)^2}. \qquad (11)$$

Condition (11) can be satisfied only in the presence of a general set or in the presence of "ideal" samples of a limited volume. In the remaining cases

$$(n+1)\,\omega^2 > \frac{n}{12\,(n+1)^2}. \qquad (12)$$

In this case with an increase in the volume of the sample, tending to infinity, the value $(n + 1)\omega^2$ tends to $n/12(n + 1)^2$.

For the purpose of comparing the considered criteria we will assume that there are "ideal" series with the duration n from the general set. In this case the hypothetical probabilities $P(x_m)$, considered as random values, will coincide with the empirical values $P_n(x) = P_m$, determined from (2). Therefore, substituting $P(x_m) = P_m$ from (2) into (5), we have

$$n\,\omega^2 = \frac{1}{12\,n} + \sum_{m=1}^{n}\left(\frac{m}{n+1} - \frac{2\,m-1}{2\,n}\right)^2 = \frac{1}{6\,(n+1)}. \qquad (13)$$

With respect to the proposed criterion (10), in this case it is expressed by formula (11). In this case their difference

$$\Delta\omega^2 = n\,\omega^2 - (n+1)\,\omega^2 = \frac{n+2}{12\,(n+1)^2} \qquad (14)$$

characterizes the value of the systematic errors allowed by the criterion $n\omega^2$ with its application to the empirical distribution curve for maximum runoff, of the stipulated duration n, constructed using the Kritskiy-Menkel' formula (2).

It can therefore be seen that the values of the systematic errors (14), allowed by the criterion $n\omega^2$, exceed the values $(n + 1)\omega^2$ obtained using (11), and the value (13) exceeds the latter by a factor of $2(n + 1)/n$, that is, greater than by a factor of two. Accordingly, the application of the criterion $n\omega^2$ to the empirical distribution curve for the probability of maximum runoff with a limited duration of the sample is inadequately correct.

As is well known [2, 10], the limiting distribution of $n\omega^2$ statistics is very complex and with $n \to \infty$ is close to the function

$$a_1(x) = \lim_{n\to\infty} P\{n\omega^2 < x\} = \tag{15}$$

$$= \frac{1}{\sqrt{2x}} \sum_{j=0}^{\infty} \frac{\Gamma\left(j + \frac{1}{2}\right)}{\Gamma\left(\frac{1}{2}\right)\Gamma(j+1)} \sqrt{4j+1}\, e^{-\frac{(4j+1)^2}{16x}} \left\{ J_{-\frac{1}{4}}\left[\frac{(4j+1)^2}{16x}\right] - J_{\frac{1}{4}}\left[\frac{(4j-1)^2}{16x}\right]\right\},$$

where $J_k(Z)$ is a modified Bessel function (Bessel function of a fictitious argument). It was demonstrated in [10] that when $n > 40$ the distribution, independently of the type of initial hypothetical distribution, is close to the limiting distribution (15), for which the values of the critical points, corresponding to some (frequently used in practical computations) significance levels [2, 10], are presented in Table 2.

Table 2

Value of Upper Limit of $n\omega^2$ in Dependence on Significance Level q%

	Уровни значимости $q = P[n\omega^2 > Z_q] \cdot 100\%$ (1)									
	50	40	30	20	10	5	3	2	1	0,1
Критические точки Z_q (2) . . .	0.1184	0,1467	0,1843	0,2412	0,3473	0,4614	0,5489	0,6198	0,7435	1,1679

KEY:

1. Significance levels $q = P[n\omega^2 > Z_q] \cdot 100\%$
2. Critical points Z_q

This table in equal degree is also applicable for the criterion $(n + 1)\omega^2$, since it was obtained from the limiting function (15) with $n \to \infty$ and is not dependent on the type of initial hypothetical distribution curve. Indeed, formulas (2) and (4) with large $n \to \infty$ give identical results. This also leads to legitimacy of use of Table 2 for $(n + 1)\omega^2$. The method for practical application of the proposed criterion is precisely the same as the use of the criterion $n\omega^2$, jointly with the values of the critical points corresponding to the stipulated significance level, obtained from its limiting distribution function (15), and applicable to river runoff is set forth in [8].

As a practical example we will examine series of maximum water discharges of the Tissa River at Delovoy village during 1933-1963 ($n = 31$, $\bar{Q} = 307\ m^3/sec$) taken from [13]. The statistical parameters C_v and C_s of this series were obtained by an evaluation by the method of the moments of probabilities of excess of the initial values, on the basis of the criterion of the minimum of the sum of the absolute errors for the evaluated parameters [8], taking into account the weights of the parameters to be summed. They were equal to: $C_v = 0.51$, $C_s/C_v = 5$ and coincided with the values obtained by the maximum similarity method [13].

Then, using the Kritskiy-Menkel' curve with the determined values of the parameters C_v and C_s/C_v we determined the value of the criterion $(n + 1)\omega^2$, which was equal to 0.0542.

Assuming a significance level $q = 5\%$ and employing the values of the critical points Z_q from Table 2, we obtain $Z_q = 0.4614$.

Now we will construct the critical region for checking the correspondence between the sample data to the hypothetical (theoretical) distributions of a three-parameter gamma distribution with $C_v = 0.51$ and $C_s/C_v = 5$ in the form $(n + 1)\omega^2 > 0.4616$.

The determined value of the criterion $(n + 1)\omega^2 = 0.0542$ falls in the region of admissible values and substantially to the left of the critical limit $Z_q = 0.4616$. Accordingly, a three-parameter gamma distribution with the parameters $C_v = 0.51$ and $C_s = 5C_v = 2.55$ with the significance level $q = 5\%$ does not contradict the observational data.

It should be noted that the proposed criterion can also be employed for evaluating the statistical parameters of a hydrological series of maximum water discharge if it is represented in the form

$$(n+1)\,\omega^2 = \frac{n}{12\,(n+1)^2} + \sum_{m=1}^{n}\left[\Pr(x_m, \overline{X}, C_v, C_s) - \frac{m}{n+1}\right]^2 = \min. \qquad (16)$$

Thus, using data from the above-mentioned series we obtained the same values of the parameters as above. This shows the possibilities of its use in problems of evaluation of parameters. However, for the final derivation it is necessary to carry out large-scale computations and compare the results with other widely employed methods. We plan to devote a separate study to this.

In conclusion it can be noted that the use of the proposed correspondence (conformity) criterion $(n + 1)\omega^2$ in application to curves of distribution of the probabilities of maximum water discharge, with a restricted sample volume n, is the most correct and ensures the maximum possible use of information available in actual observation series and also the adopted method for constructing empirical functions.

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INTRADIURNAL CHANGE IN STRENGTH OF THE ICE COVER OF RIVERS AND RESERVOIRS DURING SPRING

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 77-81

[Article by V. M. Timchenko, Ukrainian Weather Bureau, submitted for publication 21 June 1978]

Abstract: Using field measurements, the author has made a quantitative estimate of the intradiurnal variation in the strength of a thawing ice cover. On rivers and water bodies situated in regions with a solar type of springs and marked variations of meteorological conditions the amplitude of the intradiurnal changes in strength attains 60-70% of its total decrease during the entire period of the spring thawing. An empirical dependence of the intensity of the intradiurnal changes in the limiting breaking point of the ice cover on the air temperature sums was obtained.

[Text] A quantitative evaluation of the strength of the melting ice cover has come into extensive use in hydrological forecasts and ice engineering computations. There are definite recommendations and also methods for computing the strength of ice on rivers and water bodies during spring. In particular, according to the TECHNICAL INSTRUCTIONS (SN 76-66), the computed strength of ice for any type of load on hydraulic structures in spring is assumed to be constant. A major step forward in taking into account the mechanical qualities of spring ice was the development (by S. N. Bulatov [1]) of a method for computing the diurnal values of strength of the melting ice cover, which opened up broad prospects for developing methods for computing and predicting the times of opening up of rivers and reservoirs, prediction of ice jam-induced rises in water level, etc.

The practice of using the mean daily strength values shows that they rather precisely reflect the change in this characteristic with time. Nevertheless, sometimes against the general background of such a change there is a superpositioning of intradiurnal variations commensurable with the degree of general decrease in ice strength during the entire time of its thawing.

The first quantitative data on the magnitude of the intradiurnal changes in strength of the melting ice cover were obtained in field investigations on the Amur at Khabarovsk [6]. During the morning hours, as a result of the nighttime freezing of the upper layer, the strength of the ice cover on this river increased by 10-30%. Later, in a study of the mechanical and radiation properties of the thawing ice cover on the Amur River at Komsomol'sk [7], on the Razdol'naya River at Terekhovka (Primorskiy Kray) and on Kiyevskoye Reservoir [5] special experiments were devoted to an investigation of the intradiurnal variations of ice strength.

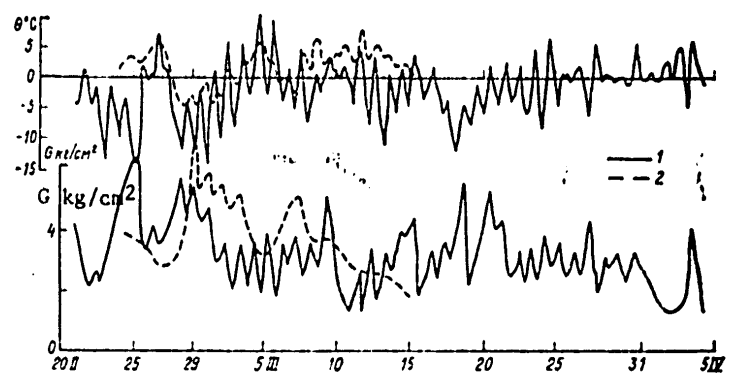

Fig. 1. Air temperature θ and strength of ice cover σ. 1) Razdol'naya River, 1976; 2) Kiyevskoye Reservoir, 1977.

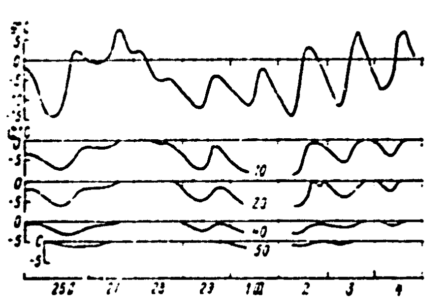

Fig. 2. Complex graph of the variation of air temperature (θ) and ice cover (ϑ_{ice}) on Razdol'naya River at different depths. The figures on the curves represent the depth of measurement of ice cover temperature.

The choice of objects for investigations was determined to a considerable degree by the difference in natural conditions of the regions. At the same time the principle of comparability of results is adhered to. In particular, it is of definite interest to compare the regime of intradiurnal change in the strength of ice of identical thickness on the Razdol'naya River and on Kiyevskoye Reservoir, where the meteorological conditions during spring are very different. It is sufficient to recall that the maximum possible intensity of receipts of solar radiation on the ice during the period of its thawing on this river and reservoir differs by approximately 150-170 cal/(cm^2·day). At Terekhovka village there are great intradiurnal changes in the meteorological elements. The amplitude of the variation, for air temperature, for example, here attains 15-20°C.

The additional investigations for each of the mentioned water bodies included a determination of the strength of the ice cover during a 24-hour period, measurement of the thickness and temperature of the ice cover, meteorological and actinometric observations. As a strength index we used the limiting breaking stress [1] obtained during testing (downward imparting of force) of cantilevers prepared in the natural ice cover. For the purpose of excluding the scale effect on the results of comparison of strength of the ice cover, we tested samples of an approximately identical size with a cross-sectional area of 3,000-4,000 cm^2. In those cases when for any reason a sample of a different size was tested, the result of the testing was corrected taking into account the difference between the sample section and the adopted size. We used the recommendations and formulas of I. P. Butyagin [2].

The frequency of measurements was established in dependence on the amplitude of the intradiurnal variations of ice strength. In individual cases the experiments were carried out at nighttime.

We carried out a total of more than 630 tests of ice cantilevers for breaking stress. This made it possible to obtain some quantitative indices of the intradiurnal change in strength of the thawing ice cover.

The intradiurnal variations in strength of the ice cover are a result of change in the meteorological conditions of thawing in the course of the day. As an illustration of this, Fig. 1 shows time-collated curves of the change in strength of the ice cover on the Razdol'naya River and Kiyevskoye Reservoir and the variation of the principal element of the heat balance at the ice surface -- air temperature. In addition to the great total amplitude of the variations in air temperature in the Razdol'naya River region we note a very frequent (almost daily) transition of air temperature through zero degrees in both directions: to positive and to negative values. Taking into account that during the daytime an average of up to 240 cal/cm^2 of solar radiation heat enters the ice cover thickness on the Razdol'naya River during the period of its melting, it becomes completely understandable that there is a marked change in ice strength during the course of the day on this river. On individual days the amplitude of variation of limiting breaking stress attains 3-4 kg/cm^2, which is 60-70% of the total decrease in strength of the ice cover during the entire thawing period.

As indicated by measurements of ice temperature which were carried out using extensible soil thermometers frozen in at different depths (10, 20, 40, 50 cm), the variations in air temperature, intensified by the heat influx during the daytime due to solar energy, leads to a significant temperature variation of the ice layer itself. It goes without saying that in the upper layers these variations are greater and in the lower layers are less (Fig. 2). With an increase in depth there is a lengthening of the time interval between the onset of the air temperature and ice temperature extrema. The temperature waves reach a depth of 20-30 cm in 1-2 hours, whereas they penetrate to a depth of 50-60 cm only after 3-5 hours.

During the initial period of thawing of the ice cover the thickness of the layer of nighttime freezing can exceed 50 cm. Later, when the thickness of the fluid intercrystalline intercalations becomes greater, nighttime decreases in air temperature lead to freezing only of the upper ice layer with a thickness of not more than 1-5 cm. However, the appearance of even such a thin layer of solid ice leads to a considerable increase in the strength of the entire ice cover. For example, on the Razdol'naya River after 7-8 March a total freezing together of the ice crystals in the morning hours was noted only in a thin (up to 5 cm) upper layer of the ice cover, but the total strength of the ice in this case increased by 1.5-2.0 kg/cm^2, that is, by approximately 30-40% of that observed in the evening (see Fig. 1).

On the Kiyevskoye Reservoir, where the change in air temperature during the course of the day during the observation period averaged 3-5°C, there were no systematic transitions of air temperature through zero. An appreciable dependence of ice cover stregth on air temperature was observed only with negative values of the latter. The variations in ice strength during the course of the day here averaged 0.5-0.8 kg/cm^2, which does not exceed 10-15% of the limiting breaking stress of the ice cover.

With a positive air temperature in the course of a 24-hour period there were virtually no variations in strength of the ice cover on Kiyevskoye Reservoir. During such periods (24-27 February, 4-6 and 8-15 March) there was a gradual decrease in ice strength (see Fig. 1).

Among the factors determining this considerable difference in the intradiurnal change in strength of the melting ice cover on the Razdol'naya River and Kiyevskoye Reservoir it is necessary to note the different intensity of the influx of solar radiation. During the period of observations on Kiyevskoye Reservoir during the daytime hours the ice layer daily received an average of 110-120 cal of heat as a result of radiant energy. This is approximately half the figure for the Razdol'naya River. In addition, an investigation of the structure of the ice cover on these water bodies demonstrated that the size of the crystals on the Razdol'naya River is somewhat greater than on Kiyevskoye Reservoir. And this, according to [3], is the reason for the increase in the probability of disruption of the continuity of crystals under identical deforming forces and leads to a more intensive decrease in ice strength.

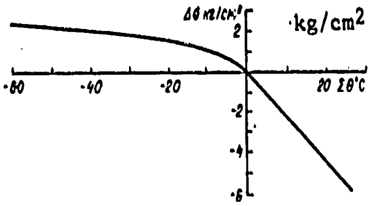

Fig. 3. Dependence of the intensity of change in strength of the ice cover during the course of the day ($\Delta\sigma$) on air temperature sums ($\Sigma\theta$).

Comparison of observational data on the Razdol'naya River and on the Amur River, where meteorological conditions during the period of melting of the ice cover are almost identical and where the intensity of solar radiation, mechanical and radiation properties of the ice cover also do not particularly differ, made it possible to detect a small difference in the intradiurnal regime of strength of melting ice covers of different thicknesses, especially during the first half of the thawing period. This is expressed for the most part in a somewhat lesser amplitude of variations in the strength of the ice cover on the Amur River. For the time being we have not obtained quantitative criteria. More extensive experimental data are necessary.

Two approaches are possible in solving the problem of computing the intradiurnal variations in strength of the thawing ice cover. The first is a decrease in the computation time unit when making computations by the existing method [1]. Such computations for the Razdol'naya River in 1976 for 3-, 6- and 12-hour time intervals indicated that the best convergence of the computed and measured strength was observed in computations on the basis of semidiurnal data. This is attributable to the fact that the adoption of shorter computation time units greatly hinders and reduces the accuracy in determining some of the parameters entering into the computations. This applies to the intensity of solar radiation, the coefficient of absorption of radiant energy by the ice cover, albedo of the ice surface, etc.

A second approach to solution of the problem of computing intradiurnal changes in strength of the melting ice cover can be the derivation of empirical dependences of the strength index on the factors determining it. With such an approach there must be a great volume of field investigations, which at the present time we do not have. As a first approximation, on the basis of the data obtained in investigations on the Amur River, on the Razdol'naya River and on Kiyevskoye Reservoir, an empirical dependence is found for the change in the strength of the melting ice cover ($\Delta\sigma$) on temperature conditions in the course of the day (Fig. 3). As an argument we use the sum of air temperatures accumulated after each temperature transition through 0°C ($\Sigma\theta$). The computation time interval in this case is assumed to be 2 hours.

The derived dependence is characterized by a good density of the field of empirical points in the region of positive temperature sums, that is, a determination of the decrease in strength during the daytime can be made with a sufficiently high accuracy. Analytically this dependence can be represented in the form of the straight-line equation

$$\Delta\sigma = -0{,}22\Sigma\theta. \qquad (1)$$

The field of points on the graph is somewhat blurred in the region of negative temperature sums. Nevertheless, the general tendency to an increase in the strength of the ice cover with an increase in the sums of negative temperatures can be traced. The empirical curve drawn through this field of points is described fairly well by the formula

$$\Delta\sigma = \sqrt{0{,}1\,|\Sigma\theta_-|}. \qquad (2)$$

Control computations made using the dependence $\Delta\sigma = f(\Sigma\theta)$ revealed a good accuracy in zones of strength decrease and a somewhat reduced accuracy in the zones of its increase.

The derived empirical dependence can be used in formulating recommendations on the regime of ice-breaking work and other measures related to the use of water bodies when there is a melting ice cover.

This dependence is also promising with respect to prediction of ice cover strength. It will make it possible, on the basis of use only of an air temperature forecast, to propose a quantitative range of intradiurnal change in the strength of the melting ice cover predictable using existing methods [1, 4].

Additional field investigations in different regions are needed for a further refinement of the scheme for computing and precomputing the intradiurnal variations of ice cover strength. The most careful attention must be devoted to water bodies whose ice cover during the spring period is subjected to considerable thermal destruction before the spring breakup.

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RETENTION OF SNOW AND MELTWATER IN AGRICULTURAL FIELDS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 82-87

[Article by Doctor of Geographical Sciences V. N. Parshin and A. K. Alekseyeva, USSR Hydrometeorological Scientific Research Center, submitted for publication 11 August 1978]

Abstract: The authors examine the problem of the effectiveness of plowing of the snow cover over enormous steppe expanses of the southeastern European USSR in the agricultural regions of Western Siberia and Northern Kazakhstan. The article gives an evaluation of the drift of snow from fields into gullies and ravines, evaporation from the snow surface and percolation of melt water into the soil.

[Text] Work on the retention of melt water in agricultural fields is widespread and is carried out over extensive expanses in the steppe regions of the southeastern European USSR, Kazakhstan and Western Siberia. This work is mechanized and is carried out by means of special snowplows attached to tractors. It is assumed that the banks formed as a result of plowing of the snow cover favor the retention of snow and meltwater in the fields. However, there are no suitable experimental data which would indicate the effectiveness of this work in years with different moistening of the soil from autumn and with different winter precipitation. The few materials from small experimental fields available along these lines can scarcely be applied to large steppe expanses. Even now there is a possibility, using aviation and special surface investigations, to carry out a comparison of snow accumulation and the moisture reserves in the soils in fields with and without plowing of the snow cover. For the time being the following considerations can be expressed along these lines.

The plowing of the snow cover over great steppe areas almost completely occupied by agricultural fields should favor the retention of the snow cover in place and impede its drifting into gullies and ravines, but it does not pursue the objective of snow accumulation in one sector of a field at the expense of another. If this is so, then it is especially of interest to have data on the magnitude of snow drift into the gully-ravine network. Quite a few investigations [1, 5, 6, 9, 10] have been made of this problem, and they all indicate that the drift of snow from fields into gullies and ravines in Northern Kazakhstan and in the southern part of Western Siberia is 5-10%, in the Lower Volga and the Central Chernozem regions -- 10-15% and only in places with highly dissected relief -- up to 20%. As an example we have cited our computations for the basin of the Don River (Central Chernozem region) and materials from expeditions of the State Hydrological Institute in the basin of the Toguzak River (Northern Kazakhstan).

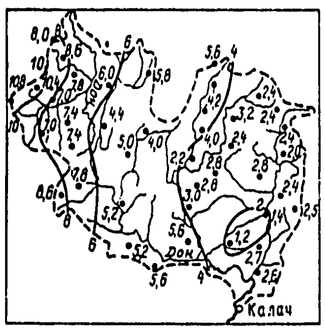

Fig. 1. Area occupied by gullies and ravines (in %) in Don River basin.

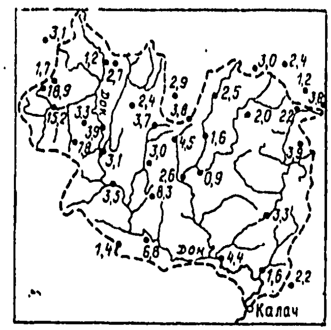

Fig. 2. Ratio of snow reserves in gullies and ravines to snow reserves in field.

Table 1

Бассейн 1	Площади под балками и оврагами, % 2	3 Норма максимальных снегозапасов, мм		4 Норма сноса снега в балки и овраги	
		5 без учета переноса снега в балки и овраги	6 с учетом переноса снега в балки и овраги	мм	%
7 Быстрая Сосна—Елец	8,6	75	92	17	18,5
8 Воронеж—Воронеж	5,6	88	99	11	11,1
9 Хопер—Бесплемяновский	4,1	85	93	8	9,4
10 Медведица—Арчединский	2,2	70	74	4	5,7

KEY:

1. Basin
2. Area of ravines-gullies,%
3. Norm of maximum snow reserves, mm
4. Norm of snow drift into gullies and ravines
5. Without allowance for snow drift into gullies and ravines
6. With allowance for snow drift into ravines-gullies
7. Bystraya Sosna-Yelets
8. Voronezh-Voronezh
9. Khoper-Besplemyanovskiy
10. Medveditsa-Archedinskiy

We carried out an analysis of snow accumulation in the gully-ravine network of the Don basin to Kalach, whose area is 221,600 km^2. This basin, especially in its upper part, is one of the regions of the European USSR most dissected by a gully-ravine network. In order to evaluate the magnitude of snow drift into gullies and ravines it is necessary, first of all, to know the area occupied by gullies and ravines, and second, have observational data on snow accumulation in them. The percentage of the area occupied by gullies and ravines was determined using a large-scale map on which the gullies and ravines were represented by special hachuring. The results of the calculations are shown in Fig. 1. The figure shows that the relative fraction of the area occupied by ravines and gullies varies from 8 to 10% at the Don headwaters to 2-3% on the interfluve of the Khoper and Medveditsa.

An estimate of the snow accumulation in ravines and gullies can be made on the basis of special snow-measuring surveys in these relief forms, carried out in the network of hydrometeorological stations and posts regularly since 1952. The processing of this material indicated that on the average the snow reserve in gullies and ravines is three times greater than on the adjacent fields.

As an example, Fig. 2 gives the ratio of snow reserves in gullies or ravines to the snow accumulation in the adjacent field sector at the end of one of the winters.

Table 1 gives the norms for the maximum snow reserves for a number of watersheds, calculated taking into account and not taking into account the drift of snow into ravines and gullies, the difference between which gives the norm for the drift of snow from fields into gullies and ravines.

The table shows that the quantity of snow carried into gullies and ravines from fields is dependent on the fraction of the area occupied by gullies and ravines and varies from 6 to 20%.

Table 2 gives data from a continuous snow-measuring survey in the basin of the Toguzak River with an area of 7,970 km^2 (Kustanayskaya Oblast), made in 1962. These materials show that the drift of snow from fields into gullies, ravines, lakes, river channels and into forest clearings is about 6%, since the area of these land use areas is about 12%, whereas the snow accumulation in them is about 18%.

When evaluating the snow retention effect during plowing of the snow cover it must be remembered that the use of snow plows for the banking of snow is possible when the depth of the snow is not less than 12 cm [3], that is, the snow falling at the beginning of winter will be subject to transport under ordinary conditions. On the average the water reserve in the snow cover in many steppe regions by the onset of spring is 70-100 mm, whereas by the time of the possible first plowing of the snow cover (snow depth 15 cm) it is equal to approximately 30 mm. Thus, 30-40% of the precipitation falling in winter does not come under the influence of plowing of the snow cover. Naturally, in winters with little snow this percentage increases. Thus, if it is assumed that after plowing of the snow cover all the snow remains in the fields, in this case the additional snow accumulation in the fields, even with a maximum area of the gullies and ravines, is about 10%, only 10-15 mm, whereas with the most frequently encountered gullying in Northern Kazakhstan and in the southern part of Western Siberia -- about 5% -- 5-10 mm.

Table 2

	Вид угодий 1	2 Относительная площадь, %	3 Снегозапас снегонакопления в бассейне, %	4 Отношение слоя воды на угодье к слою воды на водосборе
5	Пашня	39,3	31,4	0,8
6	Целина	33,4	33,4	1,0
7	Стерня	12,7	14,0	1,1
8	Луг	1,9	2,3	1,2
9	Озера	1,3	0,8	0,6
10	Камыш	1,4	2,2	1,6
11	Колки, лесопосадки, кустарник	6,4	10,2	1,6
12	Болота	0,7	0,8	1,2
13	Русла рек, балки, овраги	2,0	4,0	1,6
14	Прочие угодья	0,9	0,8	—

KEY:

1. Type of land use
2. Relative area, %
3. Snow reserve of snow accumulation in basin, %
4. Ratio of water layer in land-use area to water layer in basin
5. Plowed land
6. Virgin land
7. Stubble
8. Meadow
9. Lakes
10. Rushes and reeds
11. Forest clearings, forest plantings, scrub
12. Swamps
13. River channels, gullies, ravines
14. Other land use areas

It is also necessary to investigate the problem of evaporation from the snow cover for a level surface or a surface broken by snowplows. Experimental investigations carried out in this direction by the Kazakh Scientific Research Hydrometeorological Institute [2] made it possible to draw the conclusion that a surface broken by a snowplow evaporates almost three times more than an even surface because there is an increase in the evaporating surface and the deflatability of the snow. According to data from the State Hydrological Institute [4], evaporation from the snow cover, with condensation taken into account, in the steppe regions averages 10-15 mm, whereas according to the data of the Kazakh Scientific Research Hydrometeorological Institute, for Western and Northern Kazakhstan [2, 8] it varies from year to year from 14 to 26 mm. Even if we take the minimum evaporation value 10 mm and assume, as indicated above, that evaporation increases threefold with plowing of the snow, the additional moisture losses in evaporation after plowing of the snow is 20 mm. At the same time, as indicated above, a positive effect from plowing of the snow as a result of cessation of deflation of snow into gullies and ravines and its retention in the fields, even in the case of highly dissected relief, is 10-15 mm. Thus, the additional snow accumulation in the fields as a result of plowing of the snow cover is completely set off by the increase in evaporation from the snow. However, all this must be checked by broad field experiments.

It is also necessary to introduce clarity into the process of retention of melt water in the fields and percolation of melt water into the soil as a result of plowing of the snow cover. On the one hand, the small depth of the snow cover in furrows favors a deep freezing of the soil, which retards the thawing of the soil below and the penetration of melt water in depth. On the other hand, in spring thawed patches rapidly appear in the furrows, the upper soil layer thaws and absorbs melt water. However, with the appearance of soil thawed patches during sunny weather the soil is rapidly heated and there is a marked increase in evaporation not only from the open sectors of the land, but also from the edges of the snow banks [2]. This increases the losses in evaporation. Here also there is a need for field experiments.

If we speak in general about the percolation of melt water into the soil in the steppe regions, the long-term data on this problem [3, 5-7] indicate that when the soil is poorly moistened from the autumn, virtually all the melt water infiltrates into the soil without its working by snowplows. For example, with a water reserve in the snow cover of 100 mm, only 1-5 mm enters the rivers, and there are many such years in the zone of inadequate moistening [5]. Evidently, in such years the plowing of the snow cover does not give a significant effect in the sense of an increase in the percolation of melt water into the soil.

As an example, we will examine the formation of spring runoff in the basin of the Tobol River to the gaging station Grishenka. The area of this basin is 13,400 km^2 and it is situated in the limits of Kustanayskaya Oblast. During the last 40 years the following annual values were determined for the Tobol basin to Grishenka: entry of water into the basin from the melting of snow and rains during this period (S), an indirect characteristic of the moistening of soil before its freezing in autumn (W_{in}) and runoff during the period of spring high water (Y).

The S-Y value is the quantity of melt water which remains on the basin surface and is expended in percolation into the soil and the filling of differently closed deepenings in the surface. Below we give the probability of this value for the Tobol basin to Grishenka. It is expressed in fractions of the quantity of melt water entering into the basin (S).

$\frac{S-Y}{S}$	0.95	0.90	0.80	0.60	0.40	0.20
Probability, %	25	45	70	85	95	100

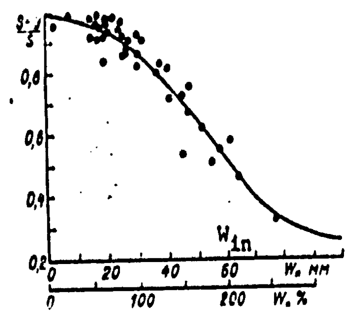

Fig. 3. Dependence of relative value of melt water expended on percolation into the soil and retention on the surface in closed deepenings on soil moistening before its freezing.

The cited data can be interpreted in the following way: during the 40 considered years during the period from 1936 through 1975 in 10 cases 95-99% of the melt water remained on the surface of the basin and in almost half the years -- 90-99%. During these years the basin received 60-110 mm of water from snow melting and rains, whereas river runoff during the high-water period was only 1-6 mm. It can be assumed that there was virtually no runoff from the fields and all the melt water was expended in percolation into the soil and filling of different kinds of deepenings on the surface. The river received only the melt water forming as a result of melting of the snow in the gullies and ravines, as is readily confirmed by the following computations. We will assume that during the spring 80 mm of water entered into the basin and there was no runoff from the fields, and in the gullies and ravines, whose total area was 3%, the snow reserve was 160 mm. We will assume the coefficient of runoff from the gullies and ravines to be 0.8; then the ravines and gullies will give a runoff into the river, scaled to a layer from the entire basin 160 x 0.8 x 0.03 = 4 mm.

Is it possible in advance, for example, at the onset of winter, to establish what will be the percolation of melt water into the soil in spring? An answer to this question is given by Fig. 3, which clearly shows the correlation between the relative fraction of melt water which is retained and which percolates into the soil (S - Y/S), as well as moistening of the soil before its freezing in the autumn (W_{in}), the computation method for which we gave in [7].

In the steppe regions, where the soil freezes deeply, the percolation of melt water in spring is determined by the degree of filling of its pores with ice, which in turn is dependent on the moistening of the soil from autumn. The process of replenishment of soil moisture by melt water in spring is extremely rational, specifically, a very dry frozen soil is capable of absorbing almost all the melt water, whereas with an increase in initial moisture there is absorption of such a quantity of melt water as will replenish the moisture reserve deficit, as can be seen clearly from the computations cited below, made for the Tobol River basin to Grishenka. The computations were made for different cases of moistening of the soil from autumn, but with one and the same layer of entry of melt water into the basin (100 mm).

$W_н$, мм	10	20	40	60	80
Y мм	3	7	28	50	70
ΔW мм	85	75	45	20	0
P мм	12	18	27	30	30
$W_н+\Delta W+P$, мм	107	113	112	110	110

[W_H = W_{in}]

Here ΔW is the percolation of melt water into the soil; P is the retention of melt water on the soil surface in various kinds of undraining deepenings.

All the values are given in millimeters of the layer in the basin area. The retention on the surface of the basin was determined separately because this water does not participate in the runoff during the period of high water and later is expended in percolation and evaporation.

Thus, the analysis presented here and the computations revealed that the plowing of the snow cover is of no significant advantage with respect to an increase in snow accumulation and percolation of melt water in the fields. However, the conclusions drawn must be checked by means of broad field experiments. Plowing of the snow cover evidently favors a more uniform distribution of snow over the area, but if as a result of the experiments there is confirmation of a considerable increase in evaporation from a snow cover broken by snowplows, this positive conclusion cannot serve as a basis for carrying out the plowing of snow.

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OCCURRENCE OF FROZEN GROUND IN THE CHANNELS AND FLOODPLAINS OF RIVERS AND THEIR INFLUENCE ON THE CHANNEL PROCESS

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[Article by Candidate of Geographical Sciences A. A. Levashov, Leningrad Hydrometeorological Institute, submitted for publication 3 July 1978]

Abstract: The paper presents the results of field investigations of manifestations of permafrost in channel sandy formations and floodplains and the peculiarities of the channel process in rivers of the permafrost zone.

[Text] In the hydrological literature little attention has been devoted to the occurrence of frozen ground in the channels and floodplains of rivers. The shortage of field data is particularly significant. At the same time, the authors of [1, 2, 4, 5] pointed out a rather complex pattern of occurrence of permafrost in the channels. However, the influence of permafrost on the channel process of numerous rivers in the permafrost zone, occupying about 48% of the territory of the USSR [3], and on bottom dredging, has been poorly studied or not studied at all. However, this influence on the channel process is specific and substantially affects the production of dredging work: when encountering frozen ground the buckets of dredging equipment are broken and the rate of working of the channels and basins is sharply reduced. This is attributable to the fact that the excavated, cemented, permanently frozen ground is not drawn into the suction apparatus and the sand concentration in the mass is sharply reduced. With removal of the unconsolidated thawed ground from the frozen surface and creation of artificial circulation of water over it the thawing of the ground and the destruction of the cement bonds by the ice occurs much more rapidly. As a result, bottom dredging by suction equipment, although at lower rates, is possible.

The freezing and thawing of bottom material causes processes which impart singular characteristics to the relief of channels and floodplains. A specific micro- and mesorelief of the river channel and floodplain arises.

For example, the thawing of frozen ground, the thawing out of ice concentrations on floodplains, is accompanied by the appearance of negative relief forms: thermokarst lakes, settlings, depressions, collapsed sectors, cracks, solifluction on slopes, etc. The freezing of unconsolidated, moistened and disperse ground is accompanied by positive forms: heavings, uplifts, ice encrustations.

The author succeeded in carrying out field investigations on the Nadym, Poluy and Pur Rivers and partially on some shallow-water reaches of Obskaya Guba. In the course of these investigations we obtained detailed data on the occurrence of permafrost in channels and on floodplains, on the depth of the permafrost surface in different channel and floodplain morphological formations, and on the rate of thawing of the seasonally frozen layer. A very simple method has been developed for detecting the surface of the frozen ground on mid-channel islands, spits, in shoals and on floodplains. This method involves the plunging of a metal rod with a length of 5-7 m and with a section of about 1.0 cm^2 to the permafrost surface by means of successive frequent sinking of the rods by one or two field workers. The permafrost surface is determined rather reliably because the contact between the rod and the permafrost is sensed as an impact against a solid object. A manual geological drill with screwed-together joints and attachments for performing the same operation requires five times more time and considerably greater efforts. By this method it was possible to detect frozen ice layers on a number of mid-channel islands and spits on the Nadym River and also the surface of the frozen ground on the floodplains and in the shoals of the mentioned rivers.

Fig. 1. Manifestation of thermokarst on floodplain near river mouth.

High mid-channel islands and spits, not held in place by vegetation, from which the snow is deflated by the wind, freeze for seven months to a great depth (3-4 m) and frequently will not succeed in thawing during the short summer season. For example, on a mid-channel island, situated at km 115 on the Nadym River, on 12 July 1975 the frozen core was situated in its high part at a depth of 1.6 m. The mean rate of thawing during the period from the moment of stable transition of air temperatures through zero was 3.2 cm/day. The rising and falling of high water passed when the ground was frozen. The flow, encountering such frozen formations on its path, makes a forced turn, and in the event of their flooding the frozen ground is encumbered by on-creeping ridges and the rate of "creeping" of spits and mid-channel islands is sharply slowed.

In order to determine the rate of thawing of frozen sandy floodplain shore at kilometer 113 from the mouth of the Nadym River we carried out clearing of the frozen skeleton of the shore from thawed ground. The measurements of the depth of thawing were made each day. The thawed layer was removed from the horizontal and lateral surfaces after each measurement of its thickness. The mean daily air temperature during the time of the measurements varied in the range 12-15°C. The rate of thawing of the shore slope in this case was 3.0 cm/day. The frozen layer sank below the low-water level and then unexpectedly disappeared. The shore in the measurement sector was steep, undermined at a rate of 2-3 m/year. It can therefore be assumed that as the channel moves horizontally frozen lenses persist beneath it for some time at a shallow depth. Such phenomena are encountered on the Lena River [5].

The surface of the frozen ground was discovered beneath a water layer of 1.3 m at a depth of 1 m from the bottom surface in numerous shallow-water sectors of Obskaya Guba in the middle of July with a water temperature at the time of measurement of 17°C. With a silty bottom surface the surface of the permafrost was at a depth of 0.8 m.

Investigations of the depth of the frozen ground on the shore part of the floodplain, directly interacting with the flow, were made for a distance of 130 km along the length of the Nadym River and 180 km along the Poluy River. These investigations indicated that on the floodplain permafrost is encountered along the entire extent of the reaches, but has an "insular" nature of distribution, that is, there are also thawed spots. The surface of the frozen ground is situated at different depths and is dependent on a complex of natural conditions, in particular, on vegetation. For example, under mosses at the end of August the frozen ground is situated at a depth of 0.2-0.3 m. In the mouth region on the floodplain, overgrown with scrub, the permafrost is found everywhere, and its surface at the end of August lies at an average depth of 1.0 m. However, here there is also widespread occurrence of thermokarst formations: lakes, settlings, depressions (Fig. 1).

Detailed observations of the position of the surface of the frozen mass and the intensity of its seasonal thawing were made during 1972-1976 on profiles of the part of the floodplain along the channel. One of these profiles is situated at kilometer 8 from the mouth of the Nadym River on a low flat floodplain overgrown with sedge-cotton grass vegetation. The mean intensity of seasonal thawing of the ground during the investigated period was 0.8 cm/day; at points situated closer to the channel the intensity of thawing was greater and at the water line attained 3.0 cm/day. The surface of the frozen ground, for the most part, duplicates the surface relief.

Here, already in 1972, we noted a case showing the sensitivity and variability of the strength properties of frozen ground in relation to the change in the thermal regime over their surface [1]. During the thawing the finely disperse ground began to liquefy and acquired floating properties. In August frozen ground was discovered on islands, mid-channel islands and shallow-water sand bars in the mouth reaches of rivers. Mid-channel islands containing frozen ground, in contrast to mid-channel islands without permafrost, have steeper slopes, evidence of their greater resistance to erosion. During the period of high water there was evidently erosion of the thawed sectors of the channel.

At the time of winter low water levels with a duration of about six months there is deep freezing not only of the high mid-channel islands and spits, but also a considerable part of the shore shallow waters and shores. During the freezing of water in the shallow waters the ice is frozen together with the bottom surface and safeguards it against erosion during the spring water rise.

We constructed standard graphs of variations in water level and ice thickness and selected typical channel profiles in order to judge the extent of through freezing of river channels across the width of the river. Determinations of the mean values of the water levels, the dates on which typical ice phenomena are observed, freezing levels, ice movements and going-out of the ice were made using curves of the probability of these values. Along the lower surface of the ice on a profile we determined the width of the nonfreezing part of the channel. The determined value was related to the width of the high-water channel, measured between the brows of the banks and assigned the value 100%. By having long-term data on the water levels and state of the ice it is possible to make a quantitative evaluation of the most probable value of channel freezing in different reaches.

A comparison of the widths of the nonfreezing parts of river channels, developing in conformity to different types of channel process, reveals their considerable difference. For example, the multibranched split channel of the Nadym River freezes on the average in 66% of its width and only 34% is warmed by the flow. The single-branch meandering channel of the Poluy River freezes in 35% of its width; the remaining part of the channel is warmed by the flow during the entire winter. The small distributaries of the Nadym River freeze by 82% of their width and sometimes completely.

A standard graph of levels, constructed jointly with a graph of ice phenomena for the split channel of the Nadym River, shows that the entire rise in levels (except for 1-2 days with going-out of the ice) occurs during the setting-in of the ice. But during going-out of the ice at the beginning the river does not open up simultaneously in its entire width. The shallow-water ice, freezing to the surface, rises later than the central channel ice, shielding the bottom from the direct effect of the flow. There can be a partial piling-up of such ice by moving sand ridges.

On meandering, nonsplitting, single-branch rivers the breaking-up occurs simultaneously in the entire width of the channel and almost the entire channel is subjected to the influence of the flow, the same as in rivers of the nonpermafrost zone. The morphological effect of freezing on mid-channel forms is the formation of specific features on them: funnels, fissures, cave-ins, and the formation of a frozen core in the body, this consisting of cemented frozen sand. Real active movement of such a form with a frozen core is possible only after its thawing.

The thawing of the frozen core in homogeneous sandy deposits is not accompanied by settling. This is attributable to the fact that the formation of different types of pore ice in sands occurs without an appreciable heaving and the water, during freezing of the sands, is pressed out from the freezing front. Fissures, funnels and cave-ins are associated most frequently with backwater sectors behind spits where there is thawing of the freezing finely dispersed deposits and the thawing out of ice carried by sediments which has frozen to the bottom surface. The mean intensity of thawing of frozen sands of mesoforms in the investigated rivers (according to observational data collected in 1969-1973) was 2.8-3.5 cm/day.

The determination and analysis of the principal schemes for development of the river channel, carried out for the investigated rivers of the permafrost zone, indicated that these schemes retain features in common with rivers in the nonpermafrost zone. The forms of channels created by the flow for the most part are described by the classification used by the State Hydrological Institute.

Together with the similarity of the channel process, with respect to the basic general characteristics typical for lowland rivers of the nonpermafrost zone, the permafrost zone is characterized by a number of characteristic phenomena, a knowledge of which is necessary in the planning and construction of structures and the straightening of channels on these rivers.

Depending on the state of the permanently frozen ground, they can play the role of a limiting factor -- the frozen material with respect to its strength properties is close to those of bedrock -- and a factor accelerating deformation -- during a period of degradation of permafrost the liquefied mass of disperse deposits can acquire the properties of fluidity.

Permafrost phenomena favor the development of specific micro- and mesorelief on floodplains. The thawing and freezing of floodplain deposits is accompanied by the appearance of negative and positive forms.

The reworking of concave shores on rivers of the permafrost zone occurs for the most part due to the thawing out and undermining of the thawed lower part of the shore and the subsequent collapse of its upper sandy loam, clayey loam and sodded part.

River bends on rivers in the permafrost zone pass through the same cycles of development as river bends in the nonpermafrost zone: however, the rate of horizontal deformations and the intraseasonal reformation from year to year here transpire more nonuniformly than on the rivers of the nonpermafrost zone. This is associated with the permafrost regime.

Thus, the use of hydromorphological theory as a methodological basis for evaluating the channel process on rivers flowing in zones of deep freezing and occurrence of permanently frozen ground has proven itself completely.

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GENERALIZED DEPENDENCE FOR HYDRAULIC COARSENESS OF PARTICLES OF DIFFERENT CONFIGURATION

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[Article by Candidate of Technical Sciences V. I. Yefimov and V. A. Rusin, Tambovskiy Institute of Chemical Machine Building, submitted for publication 10 July 1978]

Abstract: The article gives a method for computing the hydraulic coarseness of particles of different configuration. The derived computed dependence involves laminar, transient and turbulent regimes.

[Text] In the solution of a number of known practical problems in different branches of the national economy the hydraulic coarseness of particles is one of the principal hydrodynamic characteristics of heterogeneous systems. It is known that the rate of free falling of particles is dependent on their diameter, configuration and density, and also on the density and viscosity of the fluid. At the present time the total development of this dependence is absent. An interpolation formula was derived by O. M. Todes and associates [1] for the entire range of regimes of free flow around a sphere and further generalization of experimental data for particles of different configurations was carried out by Z. R. Gorbis [3]. However, these generalizations have a limited field of applicability and the interpolation formula gives substantial errors in the transitional region. With this circumstance taken into account we carried out an additional analysis of experimental data and developed a method for computing the hydraulic coarseness of particles for later obtaining a similar computation method under constrained conditions.

The rate of free falling of a particle of spherical configuration in the case of uniform steady movement is determined from the equation of equilibrium between the weight of the sphere in the medium and the resistance exerted against this movement,

$$\frac{1}{6}\pi d^3 g(\rho_T-\rho)=\lambda_\text{ш}\frac{\pi d^2}{4}\frac{\rho w_\text{ш}^2}{2} \tag{1}$$

or

[T = tur(bulent); Ш = sph(erical)]
$$\omega_{ш} = \sqrt{\frac{4}{3} g \frac{(\rho_т - \rho) d}{\lambda_ш \rho}}, \qquad (2)$$

where ρ_{sol} is particle density; ρ is fluid density; d is particle diameter; λ_{sph} is the coefficient of resistance to flow.

Expression (2) is correct for the entire range of regimes of free flow around a sphere. However, as is well known, a theoretical λ_{sph} value can be obtained only for a laminar flow regime.

In order to obtain a generalized dependence for the hydraulic coarseness of particles we introduce the parameter φ, equal to the ratio of the critical value of the velocity of free falling of a particle of any configuration ω_{cr}, corresponding to the critical value of the Reynolds number Re_{cr} with transition to a turbulent flow regime, to the velocity of free falling ω of a particle in regimes $Re < Re_{cr}$, that is

$$\varphi = \frac{\omega_{кр}}{\omega}. \qquad (3)$$

[KP = cr(itical)]

The configuration of irregular particles will be evaluated by the geometrical form factor f, equal to the ratio of the particle surface F to the surface of a spherical particle F_{sph} with an identical volume V of the solid particles, that is

$$f = \left(\frac{F}{F_ш}\right)_{V = \text{idem}}.$$

If we denote by d the determining geometrical dimension of the solid particle, equal to the diameter of the sphere, equivalent in surface to a given particle (determined, for the most part, using data from a sieve analysis for round particles), and by d_{equiv} the diameter of a sphere equivalent to the particle with respect to its volume, then, using the formulas of elementary geometry we obtain expressions for particles of any configuration in the form

[Э = equiv(alent)]
$$d_э = 1{,}24 \sqrt[3]{V}; \quad d = f^{0{,}5} d_э. \qquad (4)$$

With the introduction of the value of the equivalent diameter of the particles d_{equiv} the dependence (2) will be correct for particles of any configuration; then using (3) the φ parameter will be equal, with (4) taken into account,

$$\varphi = \frac{\omega_{кр}}{\omega} = \sqrt{\frac{d_{э.\,кр}}{d_э}} \sqrt{\frac{\lambda}{\lambda_т}} = f^{0{,}25} \sqrt{\frac{d_{э.\,кр}}{d}} \sqrt{\frac{\lambda}{\lambda_т}}, \qquad (5)$$

[Э = equiv; KP = cr(itical)]

where $d_{equiv\ cr}$ is the value of the equivalent critical diameter of the particle corresponding to the critical Reynolds number Re_{cr}; λ_T is the coefficient of resistance to flow in a turbulent regime, that is, when $Re \geqslant Re_{cr}$.

Table 1

Value of φ Parameter in Dependence on $d_{equiv\ cr}/d$

$\frac{d_{э.\ кр}}{d}$ $d_{equiv\ cr}$	0	0,1	0,2	0,3	0,4	0,5	0,6	0,7	0,8	0,9
1	1	1,086	1,171	1,254	1,337	1,420	1,500	1,560	1,649	1,738
2	1,827	1,916	2,004	2,092	2,181	2,269	2,357	2,445	2,718	2,621
3	2,708	2,796	2,883	2,971	3,058	3,146	3,233	3,320	3,407	3,494
4	3,581	3,720	3,843	3,958	4,075	4,192	4,309	4,428	4,547	4,666
5	4,787	4,908	5,029	5,151	5,274	5,398	5,522	5,646	5,771	5,897
6	6,023	6,150	6,277	6,405	6,533	6,662	6,792	6,921	7,052	7,183
7	7,314	7,446	7,578	7,711	7,845	7,978	8,113	8,247	8,383	8,518
8	8,645	8,791	8,928	9,065	9,203	9,341	9,480	9,619	9,759	9,899
9	10,04	10,18	10,32	10,46	10,90	11,09	11,28	11,47	11,66	11,86
10	12,06	12,25	12,45	12,65	12,85	13,06	13,26	13,46	13,67	13,88
11	14,09	14,30	14,51	14,72	14,93	15,15	15,36	15,58	15,80	16,02
12	16,24	16,46	16,68	16,91	17,13	17,36	17,59	17,81	18,04	18,28
13	18,51	18,74	18,98	19,21	19,45	19,68	19,92	20,16	20,40	20,65
14	20,89	21,13	21,38	21,63	21,87	22,12	22,37	22,62	22,88	23,13
15	23,38	23,64	23,89	24,15	24,41	24,67	24,93	25,19	25,45	25,72
16	25,98	26,25	26,52	26,78	27,05	27,32	27,59	27,87	28,14	28,41
17	28,69	28,97	29,24	29,52	29,80	30,08	30,36	30,64	30,93	31,21
18	31,50	31,78	32,07	32,36	32,65	32,94	33,23	33,52	33,82	34,11
19	34,41	34,70	35,00	35,30	35,60	35,90	36,20	36,50	36,81	37,11
20	37,42	37,72	38,03	38,34	38,65	38,96	39,27	39,58	39,89	40,21
21	40,52	40,84	41,15	41,47	41,79	42,11	42,43	42,75	43,07	43,40
22	43,72	44,05	44,37	44,70	45,03	45,36	45,69	47,51	47,91	48,31

KEY:

1. $d_{equiv\ cr}$

The dependence (5) shows the hydromechanical sense of the φ parameter, determining the relative value of the resistance coefficient for any regime of flow around a particle to the resistance coefficient in the self-similar region with a known relationship of the diameters $d_{equiv\ cr}$ and d, and on the basis of experimental data it is possible to represent the dependence $\varphi = \psi\ (d_{equiv\ cr}/d)$. The desirability of introducing the φ parameter in such a dependence is attributable to the fact that since transition to a turbulent flow regime for particles of different configuration occurs with a definite Re_{cr} value, then, determining $d_{equiv\ cr}$ from it, we thereby take into account the temperature of the fluid and we obtain the single dependence $\varphi = \psi\ (d_{equiv\ cr}/d)$ for particles of different configuration.

For constructing the dependence of the φ parameter we used experimental data for a sphere in the range of numbers Re = 0.1-500, cited in tabular form in [3], the tabulated data for river alluvium cited in [4], and also experimental data for particles of other configurations taken from [3]. For the purpose of retaining accuracy and simplifying computations the values of the φ parameter are given in Table 1 and for a laminar regime are represented by the empirical dependence (8).

It appears possible to compute the value of the critical Reynolds number Re_{cr} for particles of any configuration and compare the determined values with experimental data. For particles of identical density and identical physical properties of the fluid we can write:

$$\frac{Re_{кр.\,ш}}{Re_{кр}} = \frac{\omega_{кр.\,ш}\, d_{кр.\,ш}}{\nu} \Big/ \frac{\omega_{кр}\, d_{э.\,кр}}{\nu} = \sqrt{\frac{\lambda_т}{\lambda_{т.\,ш}} \frac{d^{1,5}_{кр.\,ш}}{d^{1,5}_{э.\,кр}}}$$

[КР = cr; Ш = sph; Т = tur]

or, taking the dependence (4) into account,

$$Re_{кр} = Re_{кр.\,ш} \frac{1}{f^{0,75}} \sqrt{\frac{\lambda_{т.\,ш}}{\lambda_т}}. \qquad (6)$$

We note that according to different sources in the literature, the turbulent regime for a sphere arises when $Re \geqslant 500$-1000 and for this region it is assumed that $\lambda_{tur\ sph}$ = const = 0.43-0.48. This is attributable to the fact that the setting-in of the self-similarity region for a sphere occurs more smoothly than for particles of an irregular configuration. In the method which we have proposed the initial value is the initial critical value of the Reynolds number. Therefore, the value of the resistance coefficient for a sphere in this case must assume the values of the Reynolds number. When $Re_{cr\ sph} = 500$ $\lambda_{tur\ sph} = 0.55$.

For example, we will determine Re_{cr} for river sands and alluvium ($f = 1.16$, $\lambda_{tur} = 1.092$). Using (6) we obtain $Re_{cr} = 317$. For particles with $f = 1.50$ and $\lambda_{tur} = 1.80$ (coal) $Re_{cr} = 205$, which agrees well with the experimental data.

Thus, the conclusion follows from (6) that for particles of an irregular configuration the self-similarity region sets in earlier than for a sphere, which confirms the conclusions from an analysis of experimental data, cited in [3], not only from the qualitative point of view, but also from the quantitative point of view.

The computation dependence for the hydraulic coarseness of particles of any configuration using (3), with (2) taken into account, has a form with its determination from the diameter d

$$\omega = \frac{f^{0,25}}{\varphi} \sqrt{\frac{4}{3} g \frac{(\rho_т - \rho)\, d_{э.\,кр}}{\lambda_т \rho}}. \qquad (7)$$

With the availability of experimental values of the resistance coefficient λ_{tur} in the self-similar region and the form factor f for particles, the $d_{equiv\ cr}$ value in (7) can be determined using the dependence (6) with the substitution of $Re_{cr\ sph} = 500$ and $\lambda_{tur\ sph} = 0.55$ using the formula

$$Re_{кр} = \frac{\sqrt{\frac{4}{3} \cdot g \frac{(\rho_т - \rho)}{\lambda_т \rho} d^{1,5}_{э.\,кр}}}{\nu},$$

where ν is the kinematic coefficient of fluid viscosity.

The φ values are determined using Table 1. With $d_{equiv\ cr}/d > 22.9$ φ is determined using the formula

$$\varphi = 0{,}126 \left(\frac{d_{э.\ кр}}{d}\right)^{1{,}9}. \qquad (8)$$

[э = equiv; KP = cr]

The form factor f for a series of materials of particles is given in [3].

In the absence of λ_{tur} and f for a given material of the particles the dependence (7) makes it possible to determine them. For this purpose, using experimental data, we determined the dependence of the resistance coefficient λ_{tur} in the self-similar region on the form factor of particles, since the form of particles unambiguously determines their resistance. This dependence is determined by the following empirical formulas:

$$\text{when} \quad f = 1 \div 1{,}16 \quad \lambda_т = 0{,}55\, f^{4{,}62};$$
$$\text{when} \quad f > 1{,}16 \quad \lambda_т = 0{,}826\, f^{1{,}90}. \qquad (9)$$

Experimental data show that the abrupt change in the dependence $\lambda_{tur} = \psi(f)$ when $f > 1.16$ is associated with a change in the behavior of the resistance coefficient for particles whose form differs considerably from a sphere. In order to obtain the dependence (9) to $f \approx 2.0$ it is possible to limit ourselves to use of experimental data.

The sequence of determination of λ_{tur} and f is as follows. First for one diameter of a particle of a particular material we experimentally determine the rate of its settling ω. Then, stipulating a series of f_i values, using (9) we determine λ_{tur} and from (7) we compute the rate of its settling ω for a stipulated f_i value.

Constructing the computation dependence $\omega = \psi(f)$ and plotting the experimental value ω on it, we determine the form factor for a particular material of the particles f and from (9), λ_{tur}, which make it possible to determine the value of hydraulic coarseness from (7) for particles of any size.

Comparison of computed data using (7) with the experimental data indicated a sufficiently good agreement.

We note that it is not necessary that our parameter and the similar V. N. Goncharov parameter [2] be considered similar, since they have a radical difference. The solution of this problem, proposed by Goncharov, seems to us virtually infeasible because even for particles with an identical form factor it is necessary to have experimental data for φ in dependence on the temperature of the fluid and its kind.

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AGROCLIMATIC BASIS FOR DISTRIBUTION OF LENTIL OVER THE TERRITORY OF THE USSR

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 98-102

[Article by G. G. Vasenina, Northern Caucasus Administration of the Hydrometeorological Service, submitted for publication 3 October 1978]

Abstract: The author examines the problems relating to climatic evaluation of the resources of the European territory of the USSR applicable to the cultivation of lentil. It was possible to determine the climatic ranges of possible and desirable (with respect to agrometeorological conditions) cultivation of the crop for grain.

[Text] A decisive role in ensuring a fully sound nutrition of people is played by protein. However, at the present time protein is produced at less than half the necessary norm, 21-22 kg per man per year [3]. In this connection the problem of increasing production of plant protein is one of the most timely.

Among the legumes it is of interest to examine the high-protein crop lentils -- Lens culinaris. This crop has great possibilities and is of universal importance (food, fodder, industrial, agronomic). On the basis of the content of protein in seeds (up to 36%), rate of digestibility, high taste qualities and nutrition it surpasses other legumes. In addition, lentils are an advantageous export crop. The cost of its grain on the world market is considerably greater than the cost of peas, and other legumes, 3-4 times greater than the cost of the grain crops: wheat, rye and barley [8].

In the USSR the area of lentil cultivation before the Second World War attained almost 1 million hectares (1937), which was 66% of the entire world area under this crop. With respect to the production of lentil and its export the USSR occupied first place in the world. Among the legumes lentil with respect to sown area was smaller before 1941 only relative to peas. After the Second World War the area in lentil was sharply reduced, in 1961-1963 attaining its minimum. In 1963 the area under lentil was reduced by a factor of almost 30 in comparison with 1941.

In all the main regions of lentil production this crop is inferior in yield to peas, but with a high level of agricultural engineering its yields constitute 15-25 centners/ hectare. The record yields of lentil in the USSR exceeded 30 centners/hectare.

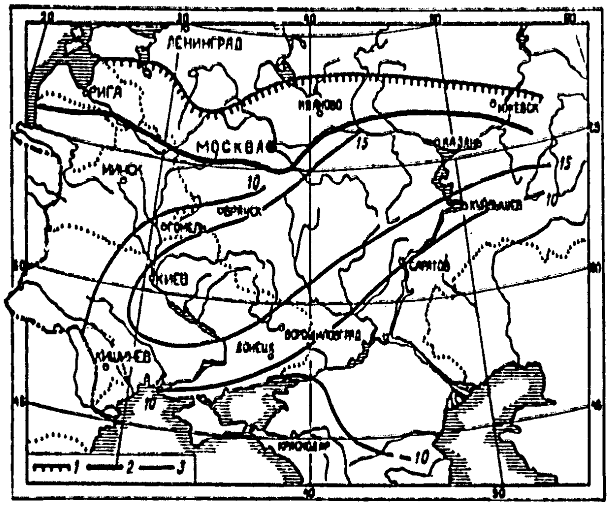

Fig. 1. Regions of cultivation of lentil in the European USSR. Limits of ripening of lentil for grain with the following probability: 1 -- 50%, 2 -- 80%; 3 -- mean yield of lentil, centners/hectare.

The interests of the national economy require a considerable expansion in the coming years of plantings of lentil and an increase in its yield.

One of the ways for solving the formulated problems is a rational distribution of lentil over the territory of the USSR.

The results of our investigations made it possible to determine the degree of correspondence between the climatic conditions of the European USSR and the biological peculiarities of lentil. It was possible to establish climatically sound areas of the crops for grain which are possible and desirable with respect to agrometeorological conditions.

The method for formulating experimental investigations and developing agrometeorological indices for the development and productivity of lentil have been examined in our studies [1, 2]. The agrometeorological indices of the rate of development of lentil, the dependence of its yield on air temperature and precipitation totals, and the results of study of meteorological conditions in the European territory of the USSR were used for agroclimatic validation of the zone of cultivation of lentil and evaluation of the agroclimatic conditions of the territory, determining the rate of development and yield of the studied crop.

The information on the influence of weather conditions on the yield of lentil in large part gives the qualitative characteristics of this phenomenon [4, 6, 7]. Our developments of the quantitative characteristics (indices) of correlation between air temperature, precipitation and yield of lentil made it possible on the basis of air temperature and the precipitation sum to determine the possible value of the grain yield. This study was preceded by an analysis of weather conditions under which each yield was formed (from available cases of determination of yield in seed selection sectors).

The results of the investigations indicated that lentil is most productive (mean yield of 16-20 centners/hectare) when it is cultivated under conditions of moderately warm weather, with a mean air temperature of 15-18°C during the growing season and a precipitation sum during the period from sprouting to maturity at harvesting equal to 100-180 mm.

The harvesting maturity of lentil sets in with the accumulation of a temperature sum of 1,500°C. The isotherms of these sums have a probability of 50%, that is, the maturity of lentil will take place in 5 years out of 10.

For evaluating the general trends in thermal resources in the practice of hydrometeorological support of agriculture use is made of the sum of active temperatures above 10°C, since at a temperature of 10°C and above most plants have an active growing season. However, the autumn transition of temperature through 10°C coincides with the period of onset of autumn freezes during which there can be damage and even total loss of the harvest. Therefore, we took into account the conditions of the growing season from the date of transition of temperature through 10°C in spring to the date of transition through 15°C in autumn. In this connection a need arose for conversion from some sums to others. Such a conversion (reduction) was carried out by introducing a so-called climatic correction obtained for the period from the date of transition of mean daily air temperature in autumn through 15°C to the date of transition to 10°C. The climatic correction was about 300°C.

Despite the fact that lentil is an early sown crop, in agricultural practice its sowing is carried out at times of transition of the mean daily air temperature through 10°C. Therefore, there was no need to introduce an additional climatic correction for the onset of the period of calculation of the sum of active temperatures. The sum of active temperatures during the period between the transition of air temperature through 10°C in spring and 15° in autumn, equal to 1,800°C, determines the northern boundary of growth of this crop. For accumulation of this sum of temperatures with a probability of 80% and 90% it is necessary that their mean long-term sum be higher. Using the A. N. Lebedev graphs [5], for computing the sums of mean daily air temperatures above 10°C with a different probability we find the final sum of active temperatures ensuring reliable cultivation of the investigated crop. It was found that the sum of temperatures, equal to 1,800°C, is ensured 80% with a mean climatic sum of temperatures of 2,000°C, by 90% -- with a sum of temperatures 2,100°C. Therefore, the isotherm of the sum of temperatures 2,000°C is the northern boundary of stable cultivation of lentil.

It has been established that the northern boundary of stable cultivation of lentil for grain in the European USSR runs along the line Riga-Moscow-Ivanovo-Izhevsk (Fig. 1).

In the European USSR the cultivation of the crop is feasible in regions where during the period from sprouting to harvesting maturity less than 100 mm of precipitation falls with a mean air temperature during this period of more than 19°C and in regions of heavy rains, when in the course of this same period more than 210 mm of precipitation fall. Under such conditions the yields of lentil are low, at state seed-selection stations -- less than 10 centners/hectare, in production fields -- about 5 centners/hectare. Assuming this yield volume to be the minimum admissible, we will assume that the isoline of 10-centner yields at seed-selection stations, governed by the above-mentioned meteorological conditions, is the boundary for the cultivation of lentil for grain feasible with respect to agrometeorological conditions. The northern boundary of cultivation of lentil (isoline of yields 10 centners/hectare) passes through Khmel'nitskiy, to the north of Gomel', to Bryansk and then coincides with the boundary of maturing of the crop. A considerable decrease in the yield of lentil in the moist zone of the European USSR, where the annual quantity of precipitation exceeds possible evaporation, is observed as a result of an excess of moisture, a great number of rainy days and a decrease in soil fertility. The sum of precipitation in these regions during the period from sprouting to maturity exceeds 200 mm and the mean temperature is not above 16-17°C.

The southern boundary of feasible cultivation of lentil for grain passes somewhat to the south of Kherson, Donetsk, Voroshilovgrad, to Saratov and then through the southern part of Kuybyshevskaya Oblast.

In the very arid zone, corresponding to the zone of dry steppes, plantings of lentil can exist only to a limited extent, primarily in regions bounding on the moderately arid zone. In the southeastern part of the European USSR the lentil yields are reduced due to inadequate supply of sown areas with moisture with high mean air temperatures during the growing season (more than 20-21°C). Despite the fact that the crop is drought-resistant, its cultivation without irrigation is irrational in a very arid zone. With irrigation, however, in early plantings, it is possible to achieve a considerable increase in its yields.

During the last decade the areas sown in lentil have been concentrated primarily in the RSFSR (98% of all the areas sown in lentil in the USSR). The Volga region accounts for 54,800 hectares (93% of the entire sown in lentil in the Soviet Union), the Central Chernozem region accounts for 2,800 hectares (slightly less than 5%), and other regions less than 1%. The greatest sown areas in lentil are in Saratovskaya Oblast (76%), in the Tatarskaya ASSR -- about 10%, in Voronezhskaya Oblast -- 3%, in Kuybyshevskaya, Penzenskaya, Ul'yanovskaya, Tambovskaya Oblasts -- approximately 2% in each oblast.

Lentil is most productive in the following oblasts: Gor'kovskaya (southern half), Ural'skaya, Kuybyshevskaya (northern half), Penzenskaya, Ryazanskaya, Orlovskaya (southeastern half), Kurskaya, Belgorodskaya, Sumskaya, Chernigovskaya, Kievskaya, Lipetskaya, Tambovskaya, Voronezhskaya (northwestern half), Poltavskaya, Tul'skaya (southeastern part), Cherkasskaya, Kirovogradskaya, Dnepropetrovskaya (northwestern part), in the autonomous republics: Tatarskaya, Mordovskaya, Chuvashskaya, Mariyskaya. The mean yields at the seed-selection stations in the mentioned oblasts and autonomous republics attain 15-20 centners/hectare (13-14 centners/hectare along the Dnepr River valleys). In the remaining territory of lentil cultivation within the limits of the European USSR the mean grain yields at the seed-selection stations are about 10-14 centners/hectare.

Judging from agrometeorological conditions, an increase in sown areas of lentil is feasible in the Central Chernozem region (in Lipetskaya, Tambovskaya, Kurskaya, Belgorodskaya, partially in Voronezhskaya, Oblasts, in the Volga region (in Penzenskaya, Ul'yanovskaya, northern half of Kuybyshevskaya Oblast, Saratovskaya (in the irrigated eastern regions of the oblast) and in the Tatarskaya ASSR. In the enumerated regions the weather conditions ensure mean yields of lentil of 15-20 centners/hectare at the level of agricultural techniques employed at state seed selection stations during the last decades, but the record yields in production exceed 30-34 centners/hectare. These regions are favorable in climatic respects for the cultivation of lentil for seed. In addition, it is rational to broaden the sown areas of lentil in the northeastern part of the Ukraine (Chernigovskaya, Sumskaya, Khar'kovskaya, Poltavskaya Oblasts), in the Volgo-Vyatskiy Rayon (in the southern half of Gor'kovskaya Oblast, Mordovskaya and Chuvashskaya ASSRs), and also in Ryazanskaya Oblast. The soil-climatic conditions in the indicated zones of lentil cultivation make it possible to achieve high indices in lentil cultivation and to obtain high income from it.

Thus, our agrometeorological studies make it possible to draw the following conclusions:

1. The northern boundary of lentil maturing (80% probability) corresponds to the isotherm of sums of mean daily (above 10°C) temperatures 2000°C.

2. The boundary of rational cultivation of lentil (yield at seed selection stations greater than 10 centners/hectare, in production fields -- more than 5 centners/hectare) passes through the northwestern part of the zone through Khmel'nitskiy village, to the north of Gomel', to Bryansk, Moscow, Ivanovo, Izhevsk. The southern boundary runs along the line Kherson-Donetsk-Voroshilovgrad-Saratov and then through the southern part of Kuybyshevskaya Oblast. With irrigation and rigorous adherence to sowing times the southern boundary can be somewhat displaced to the south.

3. The air temperature during the period from sprouting to harvesting maturity, equal to 15-18°C, and a precipitation sum of 100-180 mm with a level of agricultural techniques equal to that at state seed selection stations

ensure obtaining yields of lentil seeds of 15-20 centners/hectare. Such conditions are created in the European USSR in the Central Chernozem and Volga economic regions and in Tatarskaya ASSR. In the mentioned regions the organization of lentil seed cultivation is most rational. In addition, on the basis of agrometeorological conditions it is desirable to expand the areas sown in lentil in the northeastern part of the Ukraine, southern half of Gor'kovskaya Oblast, Mordovskaya and Chuvashskaya ASSRs.

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HYDROLOGICAL PECULIARITIES IN THE DESIGNING OF CANALS IN FLOODPLAINS

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[Article by Candidate of Geographical Sciences A. V. Plashchev, All-Union Scientific Research Institute for Water Preservation, submitted for publication 7 June 1978]

Abstract: Dangerous hydrological phenomena are examined (plane deformations in a channel and floodplain, wind waves, ice phenomena) which must be taken into account in the designing of canals in river floodplains.

[Text] In the Ukrainian Division of Gidroproyekt specialists have developed a downstream variant of a route for the Dnepr-Donbass Canal, in accordance with which the canal is not run along the water divide, but along the floodplain of the tributaries of the Dnepr and Severskiy Donets. In accordance with a similar variant, specialists at this institute have also planned a Don-Oskol canal. The route of this canal passes along the floodplain of the Don and Oskol tributaries -- the Potudan' and Kotel Rivers. The water level in such a canal is close to the ground water level of the adjacent territory. The downstream variant, in comparison with the water divide variant, has a number of advantages, of which the most important are a decrease in the losses of water in filtration without special antifiltration measures and the entry of ground water into the canal [4].

In the designing of canals on river floodplains it is necessary to take into account a number of dangerous hydrological phenomena, among which we must first include plane deformations in the river channel and floodplain, wind waves and ice phenomena in the flooded floodplain.

When laying out the route a factor of considerable importance is the closeness of a meandering river channel to the canal planned in the floodplain. As a result of plane deformations of the channel there can be destruction of the dikes safeguarding the canal from inundation by high waters, pumping stations, pipelines, electric transmission lines and other structures.

In selecting the route for a canal it is necessary to visualize clearly the essence of the channel processes transpiring in the river and to know the regularities in their development. A particularly important problem is the preparation of a prognostic evaluation of horizontal deformations of the river channel and floodplain. It can be made on the basis of the method for hydrological-morphological analysis of the channel process developed at the State Hydrological Institute (N. Ye. Kondrat'yev, I. V. Popov, B. F. Snishchenko) [2]. This method makes it possible to establish the type of channel process, with an adequate accuracy to determine the rate of horizontal deformations and accordingly make a valid choice of a route for the canal or collapsing which may occur along segments of great extent. At the present time this method is being used extensively in planning river crossings for high-tension power lines [3], pipelines and water intake structures. However, the hydrological-morphological analysis method has still not been developed applicable to the choice of a canal route on floodplains.

At the present time this work is being done at the State Hydrological Institute. It has formulated the basic points in the Instructions for Evaluating Channel and Floodplain Processes in the Planning of Canals in Floodplains.

Among the types of floodplains defined by the State Hydrological Institute the most convenient and diversified conditions for the laying out of canals are on the floodplains of freely meandering rivers. Such floodplains, having a great width, are characterized by the fact that the current zone of meandering frequently does not occupy the entire expanse from one valley slope to the other. Beyond the limits of the extensive floodplain areas there are sectors of floodplain which are not subjected to reformation and which are suitable for the laying out of canals. Under these conditions it is possible to get by with a minimum number of intersections of river channels by the canal route and there is a minimum volume of earth work; this can be done by laying out the canal on the lowest terraced part of the floodplain. In the laying out of a canal there is a possibility for making use of ox-bow lakes and floodplain distributaries when laying out routes. However, the complex relief of the floodplain surface favors the appearance of currents on the floodplain during its inundation and these currents can then cause erosion of the slopes of protective dikes.

Maps of the distribution of types of channel process contained in the handbooks RESURSY POVERKHNOSTNYKH VOD SSSR (Surface Water Resources of the USSR) can be used for a preliminary judgment concerning the principal types of floodplains. At the same time using the classification of floodplains formulated by the State Hydrological Institute and these maps it is possible to draw preliminary conclusions concerning the types of floodplains encountered in different segments along the canal route. On the basis of a preliminary analysis of the cartographic material it is possible to establish the type of channel process. Then it is possible to define morphologically uniform segments and the general tendencies in channel reformations, as well as the peculiarities of development of individual channel (meander) macroforms. Particular attention is devoted to reaches within which the development of the meanders can constitute a danger for a canal and its structures. It is possible to determine the principal indices of the channel process, to establish the morphometric and hydromorphological dependences and to estimate the rates of horizontal and vertical deformations. The meandering zone is established for each reach; the canal route is laid out taking the development of these processes into account.

When constructing a canal on a river floodplain its dikes can split the floodplain flow and lead to an increase in the discharges in any of the cut-off parts of the floodplain. This can cause an increase in the water level and the current velocity in the floodplain sector. In such segments of the canal bordered by dikes with an increase in the water level the dikes themselves and the floodplain can be subjected to erosion and other unfavorable consequences. Information on the magnitude of increase in water level in the diked segment, the appearance of currents with considerable velocities and the going-out of the ice is necessary for laying out the route of a canal, ascertaining how high the dikes should be and making a decision on the reinforcement of their outer slopes.

Wind waves are also among the dangerous hydrological phenomena which must be taken into account in the planning of canals.

Rather high waves can be formed on the inundated broad floodplains of rivers and the engineering structures situated there can be subjected to destruction. The greatest danger arises when a particularly strong wind coincides with particularly strong flooding, although the probability of coincidence of these two phenomena can be beyond the limits of the normalized and taken-into account probabilities.

Information on the height of the wind waves, especially on broad and deeply inundated floodplains, is necessary when calculating the height of the dikes protecting the canal.

The hydrological investigations carried out by the Hydrometeorological Service in lakes and reservoirs as a rule include observations of waves. Such observations are not being made on floodplains inundated at high water.

As a result of lack of field data and experimental data until now there are no reliable procedures and methods for computing the height of wind waves on an inundated floodplain. Most frequently for this purpose specialists use methods developed for lakes and reservoirs, especially the A. P. Braslavskiy method [5].

In order to determine the computed height of wind waves on an inundated floodplain it is most acceptable to use the methods developed at the State Hydrological Institute [3, Appendix VI]. In selecting the computed wave height an expression is derived directly for the probability of wave height. In computations by this method it is necessary to have regime stochastic characteristics of the factors determining wind waves -- curves of the probability or frequency of recurrence of wind velocity and the same characteristics for the water level.

The Hydrological Institute has proposed a simplified and a detailed method for constructing the wave height probability curve. The simplified computation method does not ensure obtaining precise results in the region of low probabilities. In order to plot a probability curve it is necessary to have considerable material over a long observation period. The detailed method is based on use of probability density curves for wind velocities and level positions.

However, the methods for computing wind waves in an inundated zone, proposed by the State Hydrological Institute, should be improved for the purpose of their simplification.

In the hydrological validation of the plan for a canal on a floodplain it is necessary to have detailed characteristics of the river ice regime, the ice regime on the floodplain and at the source feeding the canal. For a river flowing near the canal and a source from which the water is fed into the canal the description can be prepared on the basis of actual observations and data obtained by computations. With respect to river floodplains, information on their winter regime is usually lacking. However, on a floodplain, a considerable part of which is occupied by a canal and its dikes, dangerous hydrological phenomena can occur which must not be overlooked when planning canals here.

In investigating the ice-thermal regime of a projected canal use is made of observational data from the Hydrometeorological Service network and the results of an analysis of data on the winter operation of existing canals. Use is also made of a theoretical computation method for determining different elements of the winter regime (determination of the setting-in of ice, thickness of the ice cover, slush formation, etc.). As a result it is possible to clarify the thermal and ice regimes of the canal and possible ice difficulties and measures are laid out for eliminating them and ensuring normal operation of the structure under winter conditions.

The going-out of the ice in spring on a floodplain occupied by a canal and protective dikes can be classified as a dangerous hydrological phenomenon. However, for our lowland rivers the going-out of the ice in spring occurs in the channel. It is therefore usually assumed that it can inflict damage only to structures situated in the main channel of the river, where when there are high levels, and accordingly, a great current velocity, the currents have a maximum destructive force. However, provision must be made for the possibility of damage to structures by ice forming directly on the floodplain itself, on which high waters can be formed during winter. This is characteristic for rivers in the southern regions of the country.

At ordinary high water the floodplain is flooded for a relatively short period. However, cases have been noted when in winter, as a result of thaws and rains, the floodplain will be inundated for a long time. We observed such a high winter water on the Severtskiy Donets River at Mayaki village in Donetskaya Oblast [1]. As a result of thaws and rains the river floodplain was inundated for a period of two months. Then there was a drop in air temperature and an ice cover with a thickness up to 20 cm was formed on the floodplain. During movements the ice inflicted considerable damage on the supports of electric transmission and communication lines situated here.

In the planning of canals it is necessary to take into account the ice loads on structures on the floodplain [6]. In computing them it is necessary to know the ice thickness. The following two cases can be considered:

1. Rivers with emergence of the ice onto the floodplain. The going-out of the ice in spring coincides with floodplain inundation. In this case the ice is carried from the river onto the floodplain and the ice thickness is determined by the conditions of its accretion in the rivers; computations of ice thickness are made using the formulas employed for rivers.

2. The ice goes out on the floodplain itself. For computing the floe thickness it is possible to use formulas derived for reservoirs because floodplains can be regarded as water bodies with a weak flowthrough.

When lakes are present it is necessary to take into account the possibility of formation of drifting ice fields.

The dimensions of the floes and their thickness are determined by direct observations in the field, by analogy or by computations.

On low sandy floodplains the ice can be carried onto them at the time when the ice is going out and it is piled up against the slopes of the protective dikes. A destructive effect on the slopes of the dikes is also exerted by the ice cover freezing to them.

In this article we have only discussed a small fraction of the hydrological phenomena which must be taken into account when selecting the route for a canal on a floodplain. In the future, taking into account the requirements of professional planners, it is necessary to develop methods for computing the horizontal deformations of river channels for different types of meandering. It is also necessary to develop a method for computing the depth of inundation of the floodplain during high waters and flood waters of different probability. The dimensions of protective dikes cannot be computed without having such data.

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UDC 556.166:627.81

THE ANDREYEV FORMULA FOR TAKING INTO ACCOUNT THE ACCUMULATION OF INFLOW IN CALCULATING WATER DISCHARGE PIPES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 105-107

[Article by Candidate of Technical Sciences L. G. Rabukhin, Volgograd Civil Engineering Institute, submitted for publication 5 June 1978]

Abstract: On the basis of the results of computations made using the water balance equation the author gives an analysis of accuracy and the field of applicability of the O. V. Andreyev formula, used for taking into account the influence of accumulation of inflow in the case of curvilinear hydrographs. It was possible to determine the values of the coefficients making it possible to use this formula for computations of regulation of runoff in the case of hydrographs constructed in accordance with the INSTRUCTIONS ON DETERMINING COMPUTED HYDROLOGICAL CHARACTERISTICS SN 435-72.

[Text] O. V. Andreyev [1], using data from computations for constructing hydrographs and determining the openings in structures, with accumulation taken into account, work done by A. A. Kurganovich at the Kiev Highway Institute, obtained the dependence

$$q_m = Q_m\left(1 - \frac{V_m}{0,7\ W}\right), \tag{1}$$

where q_m is the maximum water discharge passing through the structure, Q_m is the maximum water discharge flowing in from the basin, V_m is the maximum water volume accumulated in the pond-reservoir, W is the runoff volume.

The ordinates of the inflow hydrographs for six time intervals were computed by A. A. Kurganovich on an electronic computer using the equation for the balance of runoff volumes.

In the derivation of formula (1) O. V. Andreyev, on the basis of data from A. A. Kurganovich, established ([1], p 27) that with the relationship $q_m/Q_m >$ 1/3 the triangular hydrograph of water inflow to the structure, introduced into the computations, does not correspond to the entire runoff volume, but only to 0.7 W.

Table 1

Values of the k Coefficient

λ	При трубе диаметром d, м [1]				Средние значения k [2]
	1,0	1,25	1,5	2,0	
0,3	0,70	0,72	0,72	0,73	0,72
0,8	0,86	0,86	0,87	0,88	0,87
1,5	0,88	0,89	0,90	0,91	0,90
2,6	0,91	0,92	0,92	0,93	0,92

KEY:
1. With pipe having diameter d, m
2. Mean values

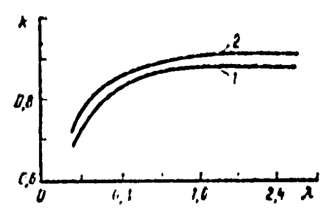

Fig. 1. Dependence of the k coefficient on the inflow hydrograph form coefficient λ. 1) with curve of dependence of volume of pond-reservoir on depth of type V_1; 2) with curve of dependence of volume of pond-reservoir on depth of type V_2.

The process of accumulation of a high water is influenced to a considerable extent by the form of the inflow hydrograph. According to the INSTRUCTIONS ON DETERMINING COMPUTED HYDROLOGICAL CHARACTERISTICS, the form of the inflow hydrograph is taken from models of observed high and flood waters (most disadvantageous for operation of planned structures) or from a standard equation ([5], section 4.41). The configuration of the hydrograph is characterized by the hydrograph form coefficient λ or the asymmetry coefficient k_{sa}. The λ and k_{sa} values are given in editions of the monograph RESURSY POVERKHNOSTNYKH VOD SSSR (Surface Water Resources of the USSR).

The O. V. Andreyev formula (1) is frequently cited in the literature [2, 4 and elsewhere] and is used in planning. [In [1, 2 and 4] different letter notations are used: for example, in [1] the q_m value is denoted Q_s and the W parameter is designated W_{st}.] It is therefore important to know which of the types of hydrographs constructed in accordance with the INSTRUCTIONS [5] corresponds to this formula. For clarification of this problem we used the results of computations of round pipes, taking accumulation into account, which we made using the water balance equation at the Vologograd Civil Engineering Institute on an "Odra-1204" electronic computer. The computations were made using different values of the maximum inflowing discharges Q_m, duration of inflow T, pipe diameters d, regimes of water movement in pipes without a head, with partial and full heads, and three types of curves of the dependence of pond volume V on depth H: $V_1 = 3330\ H^3$, $V_2 = 4600\ H^{5/2}$, $V_3 = 5000\ H^2$ with indices of curvature of the volume curve η equal to 0.385, 0.325 and 0.25 respectively. The η value represents the ratio of the greatest degree of curvature of the volume curve $V = \varphi(H)$ in the interval between the initial and maximum water horizons in the pond-reservoir to the difference in the readings of these horizons.

The computations were made for hydrographs with form coefficients λ from 0.3 to 2.6, which takes in the entire range of change in λ provided for in accordance with the INSTRUCTIONS SN 435-72 for the territory of the USSR. More than 1,200 computations were made.

As a result, it was established that formula (1) gives precise results in the case of hydrographs with a form coefficient $\lambda = 0.3$. With other forms of hydrographs the values of the maximum discharges passed through the structures are too low; for example, with $\lambda = 0.8$ -- on the average by 26%, with $\lambda = 1.5$ -- by 34%, and with $\lambda = 2.6$ -- by 40%. With a decrease in the duration of inflow and a stipulated pipe diameter this discrepancy increases and particularly strongly in the case of hydrographs with high λ values. Therefore, in place of the constant value of the coefficient 0.7 in formula (1) it is desirable to use the variable k value, that is, represent this formula in the form

$$q_m = Q_m\left(1 - \frac{1}{k}\frac{V_m}{W}\right).$$

Table 1 gives the k values determined on the basis of computation data for a volume curve of the type V_2 and on the basis of averaged k values it was possible to construct (Fig. 1) the dependence $k = f(\lambda)$ for volume curves of the types $V_1(\eta = 0.385)$ and $V_2(\eta = 0.325)$.

With a pond-reservoir volume curve of the type V_1 the k values are only 3-5% less than for a volume curve of the V_2 type. These types of volume curves take in the most frequently encountered forms of the regulating capacity of ponds. We note that I. A. Zheleznyak [3] considers to be most probable the value $\eta = 0.35$ for computations of bridges, under-road pipes and other structures situated in the lower part of a ravine or deep gully. This conclusion is based on the results of investigations for generalizing and classifying the volume curves for 274 reservoirs made by Ye. A. Topchiyev and A. I. Moldovanov.

It follows from the data in the table that the value of the k coefficient remains almost constant in dependence on pipe diameter.

The use of the O. V. Andreyev formula (1) with the mentioned values of the k coefficient (1) makes it possible to increase the accuracy of the results of computations of pipes, with accumulation taken into account, in the case of hydrographs constructed in accordance with the INSTRUCTIONS SN 435-72. In this work it is possible to use the graph analysis method proposed by O. V. Rasskazov [1, 3], but on the y-axis of the $q_m = f(H^3)$ graph instead of 0.7W/a it is necessary to plot kW/a, where

$$a = \frac{m_1 + m_2}{6i_{rav}};$$

m_1 and m_2 are coefficients of steepness of ravine slopes; i_{rav} is the slope of the ravine bottom.

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UDC 551.(510.3+524.7)

SOME RESULTS OF DEVELOPMENT OF A METHOD FOR MEASURING DENSITY AND TEMPERATURE USING FALLING SPHERES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 108-111

[Article by Professor P. Dyubua, A. N. Melnikov, S. V. Pakhomov and M. B. Pinskiy, Central Aerological Observatory and Aerological Observatory, Meteorological Service GDR, submitted for publication 22 June 1978]

Abstract: The article describes some results of development of a method for measuring density and temperature of the upper atmosphere using inflated falling spheres, carried aloft by a M-100B meteorological rocket. The mean square errors in measuring temperature and density vary from 3 to 9%; wind is in the range from 10 to 4 m/sec in the altitude range 80-40 km. Experimental density and temperature profiles are given.

[Text] One of the methods for measuring density and temperature of the stratosphere and mesosphere used in rocket sounding of the upper atmosphere is the "falling spheres" method [4]. It is based on rocket sounding of the elements of the trajectory of a body of spherical configuration delivered to a stipulated altitude by means of a meteorological rocket and performing motion in the free atmosphere under the influence of gravity and wind. The joint solution of the equations of motion of the sphere and the equations of state of the gas will make it possible to compute density, temperature, pressure and wind velocity components.

The development of the falling spheres method applicable to M-100B meteorological rockets was carried out by the Central Aerological Observatory of the State Committee on Hydrometeorology in collaboration with the Aerological Observatory of the GDR Meteorological Service under the program of work of the socialist countries in the "Interkosmos" space meteorology field.

The sphere (Fig. 1) is a soft envelope fabricated from polyethylene terephthalate film with a thickness of 12 μm. In order to ensure the necessary radar reflectivity, within the envelope there is an octahedral metallized corner reflector. The envelope, packed in a special container, is placed aboard an M-100B meteorological rocket; it is shot from the rocket when it reaches the peak of its trajectory at an altitude of about 90 km. In the free atmosphere the container is automatically opened and the envelope is filled with vapors of isopentane C_5H_{12}, situated in a capsule with calibrated apertures for evaporation. The diameter of the envelope in a filled state is 1.5 m; the total mass of the system is 359 g. The evaporation regime and the quantity of isopentane were selected in such a way that in the altitude range from 90 to 35-40 km an excess pressure of 8-10 mb was ensured.

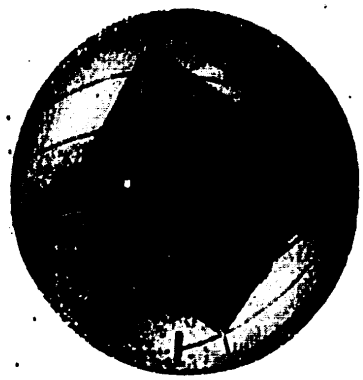

Fig. 1. General view of sphere.

Trajectory measurements are made using a pulsed radar station operating in the centimeter range with automatic tracking in all coordinates and representation of output data in digital form suitable for direct input into an electronic computer. The processing of data is with a "Minsk-32" electronic computer using a special program.

The basis for computing density is a well-known expression for sphere drag

$$m\vec{a} = -\frac{1}{2}\rho|\vec{V}|\vec{V}CA,$$

where m is sphere mass; $\vec{a}$ is acceleration; ρ is atmospheric density; $\vec{V}$ is the sphere velocity; C is the drag coefficient; A is the sphere presentation area.

If the wind velocity W is not equal to 0, the equations of motion are written in the form

$$ma_x = -\frac{1}{2}\rho CA|\vec{V}|(V_x - W_x),$$

$$ma_y = -\frac{1}{2}\rho CA|\vec{V}|(V_y - W_y),$$

$$ma_z = mg - \frac{1}{2}\rho CA|\vec{V}|(V_z - W_z).$$

where

$$|\vec{V}| = \sqrt{(V_x - W_x)^2 + (V_y - W_y)^2 + (V_z - W_z)^2}.$$

It follows from these equations that

$$W_x = V_x - a_x \frac{V_z - W_z}{g - a_z},$$
$$W_y = V_y - a_y \frac{V_z - W_z}{g - a_z}, \qquad (1)$$
$$\rho = -\frac{2m(a_z - g)}{CA|V|(V_z - W_z)}.$$

Here V_x, a_x, V_y, a_y, V_z, a_z are the components of velocity and acceleration of the sphere; W_x, W_y, W_z are the wind velocity comonents.

Taking into account the gravity components along the axes and the Archimedes force the expressions (1) assume the form

$$W_x = V_x - V_z \frac{a_x + \left[g_0 \frac{X}{r}\left(1 - \frac{V_{c\Phi}\rho}{m}\right)\right]}{a_z + \left[g_0\left(1 - \frac{2Z}{r}\right)\left(1 - \frac{V_{c\Phi}\rho}{m}\right)\right]},$$
$$W_y = V_y - V_z \frac{a_y + \left[g_0 \frac{Y}{r}\left(1 - \frac{V_{c\Phi}\rho}{m}\right)\right]}{a_z + \left[g_0\left(1 - \frac{2Z}{r}\right)\left(1 - \frac{V_{c\Phi}\rho}{m}\right)\right]}, \qquad (2)$$
$$\rho = \frac{2m\left[g_0\left(1 - \frac{2Z}{r}\right) - a_z\right]}{CA|V|(V_z - W_z) + g_0\left(1 - \frac{2Z}{r}\right)V_{c\Phi}},$$

where r is the earth's radius, V_{sph} is the sphere volume [cΦ = sph(ere)]

The pressure at the first altitude level is determined from the expression

$$P_1 = T_1 \rho_1 \frac{R}{\mu}.$$

Here R and μ are the gas constant and the molecular weight of the air respectively; T_1 is the model value of temperature for a particular level, which is taken from the CIPA-72 reference atmosphere.

At all subsequent points the following expression is used

$$P_i = P_{i-1} + \frac{\rho_i + \rho_{i-1}}{2} \Delta Z g_0 \left(1 - \frac{2Z}{r}\right),$$

where ΔZ is the vertical processing interval.

Temperature is computed using the expression

$$T_i = \frac{\mu}{R}\frac{P_i}{\rho_i}.$$

Computations are made by the successive approximations method until there is a convergence between adjacent density values of 1%.

For satisfying the self-consistency condition for the computation scheme the determination of the aerodynamic Reynolds number Re and the Mach number M is accomplished in accordance with the expressions

$$\mathrm{Re} = \frac{\rho_i |\vec{V}| d\,(T_i + 110{,}4)}{1{,}458\cdot 10^{-6}\, T_i^{3/2}},$$

$$M = \frac{|\vec{V}|}{340{,}3}\sqrt{\frac{288{,}16}{T_i}},$$

which also includes the density and temperature values. On the basis of the computed Re and M numbers the most reliable [3] experimental values of the drag coefficient C are selected.

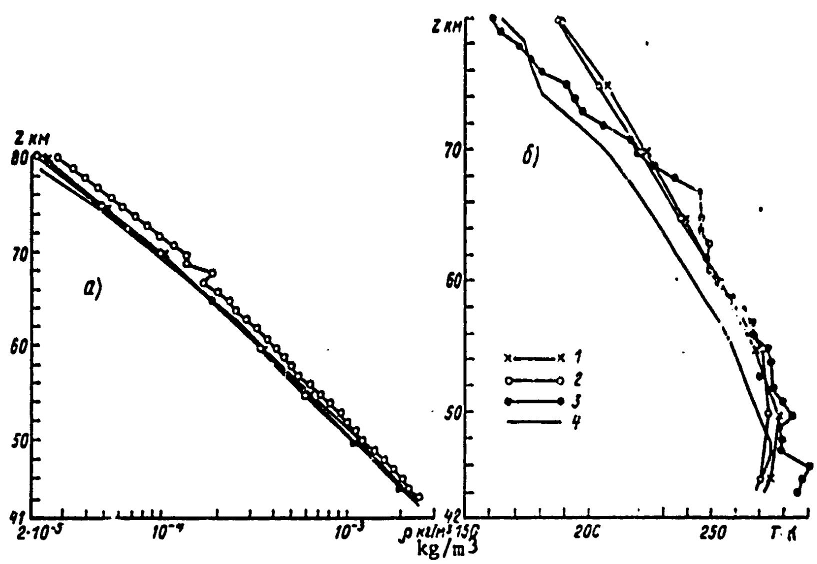

Fig. 2. Profiles of density (a) and temperature (b) obtained in an experimental launching at Volgograd. 1) CIRA-72, 2) CA-64, 3) sphere 21 May 1976, 4) M-100B 21 May 1976.

Table 1

Components of Measurement Errors (%) for Density, Temperature and Wind

Error components	Altitude, km 80-55	55-35
Mean square error in measuring density	2.5-3.0	3.0-7.0
Mean square error in measuring temperature	2.0-3.0	3.0-9.0
Systematic error in measuring density	2.0-3.0	1.0-2.0
Systematic error in measuring temperature	1.0-2.0	1.0-2.0
Random error of aerodynamic nature for density and temperature	1.0	1.0
Random error in measuring density due to errors in determining C	3.0	2.0
Mean square error in determining wind components, m/sec	10.0-4.0	4.0-2.0
Random error in measuring density as a result of neglecting vertical wind	0.5-1.5	1.5-5.0

In order in the final data to eliminate high-frequency noise with periods 2-3 sec, caused by the specific nature of operation of the radar autotracking systems, the primary trajectory data are subjected to filtering using digital filters. The parameters of the filters are selected in such a way that on the one hand there is a minimizing of the component of error of radar origin, and on the other hand, so that there are no distortions of parameters of the measured trajectory. The choice of the optimum parameters of the filters -- the cutoff period and the differentiation period, equal to 25 and 55 sec, was accomplished experimentally by numerical modeling on an electronic computer.

As a result of filtering of the initial data it is possible to achieve an optimum relationship between the noise and systematic measurement errors.

An analysis of errors of an aerodynamic nature (peculiarities of detached flow in a wake, oscillatory motion due to displacement of the center of mass relative to the pressure center) [2] indicated that errors of this class are negligible in the entire investigated altitude range.

We also analyzed the errors in determining density and temperature as a result of inexact determination of the drag coefficient and errors in tie-in to the model value of temperature at the first computation level. The components of errors in measuring density, temperature and wind are given in Table 1 for two altitude ranges.

It can be seen that the random errors in determining density and temperature as a result of errors of radar origin substantially exceed the remaining error components, especially at altitudes 35-40 km. This is attributable to a decrease in the vertical velocity of the sphere at low altitudes. On the other hand, the errors in determining the wind components decrease as the sphere descends.

At an altitude of about 35 km the sphere slows down, after which only the wind velocity components are computed.

Figure 2 shows the results of one of the experimental launchings of a spherical envelope on an M-100B meteorological rocket at Volgograd. The same figure gives the temperature and density values obtained in this same launching of a meteorological rocket by the resistance thermometer method [1]. As a comparison we have shown the vertical distributions of density and temperature in accordance with the standard atmosphere CA-64 and the International Reference Atmosphere CIRA-72. With the apogee of the sphere at an altitude of 90 km correct measurements begin at an altitude of about 80 km, when the vertical acceleration of the sphere begins to differ from the acceleration of free falling. In the region of altitudes 67-69 km the increased scatter of experimental points is attributable to the fact that this region corresponds to Mach numbers approximately equal to unity, for which the C values are determined with a maximum error. In general, the resulting profiles agree, within the limits of measurement accuracy, both with independently measured density and temperature values and with standard data.

It is important to note that the developed method, which is relatively simple and inexpensive, can be used at altitudes up to 100 km, where standard methods for measuring density and temperature are inapplicable.

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USE OF PILOT BALLOON MEASUREMENTS FOR NUMERICAL ANALYSIS AND WEATHER FORECASTING

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 112-118

[Article by Candidate of Physical-Mathematical Sciences P. K. Dushkin, Candidate of Technical Sciences A. I. Belov and S. I. Grigorov, submitted for publication 4 April 1978]

Abstract: A study was made of the possibility of using the results of sounding of the atmosphere using constant-volume pilot balloons having small dimensions and simple equipment (temperature or pressure sensor, radio transmitter). An attempt is made to show the information possibilities of constant-volume pilot balloons.

[Text] Constant-volume pilot balloons, in flight having a constant mass, move in the atmosphere along isosteric trajectories, the altitude of which is dependent on the state of the fields of meteorological elements, in particular, on temperature, pressure and wind.

The nature of the movement of pilot balloons in space is to a certain degree an indicator of state of the atmosphere. At the same time, a pilot balloon is a singular integrator of the equations of atmospheric dynamics, since it gives information on the coordinates of the trajectory, whose determination relates to one of the complex and fundamental problems of classical hydromechanics. In itself the Lagrangian study of a fluid, as is well known, is characterized by a higher information content in comparison with the Eulerian method. Indeed, from the economic point of view the Eulerian method is less attractive. For example, for constructing the trajectory of an air particle over a quite extensive area use is made of a considerable volume of information from several hundred meteorological stations.

Such an approach has never been alien to hydrodynamic forecasting methods. One of the first hydrodynamic schemes, developed by M. I. Yudin in the 1950's and which has not lost its importance at the present time, was formulated fundamentally on the basis of computation of trajectories [7].

A pilot balloon, such as a meteorological satellite or a weather reconnaissance aircraft, carries out asynchronous measurements, for the processing and analysis of whose data it is necessary to use some methods from four-dimensional analysis [1]. However, this asynchronicity of pilot-balloon information from the positions of hydrodynamic forecasting methods is ordered and natural, since the trajectories of pilot balloons, in contrast to the trajectories of satellites or aircraft, are determined for the most part by physical processes transpiring in the atmosphere.

Using the constant density conditions ρ = const, the equation of motion for a pilot balloon in the isosteric coordinate system x_ρ, y_ρ, ρ, t can be represented in the form

$$\frac{du}{dt_\rho} = -\frac{1}{\rho}\frac{\partial P}{\partial x_\rho} - RK\frac{\partial z}{\partial x_\rho} + lv, \tag{1}$$

$$\frac{dv}{dt_\rho} = -\frac{1}{\rho}\frac{\partial P}{\partial y_\rho} - RK\frac{\partial Z}{\partial y_\rho} - lu, \tag{2}$$

where u,v are the velocity vector components, P is pressure, R is the universal gas constant of the air, $K = g/R - \gamma$, g is the acceleration of free falling, $\gamma = -\partial T/\partial z$, T is temperature, Z is the altitude of the surface ρ = const, l is the Coriolis parameter.

The ρ index will henceforth be omitted and the K parameter is assumed to be constant along the trajectory.

Taking into account that on the isosteric surface

$$\frac{\partial P}{\partial x} = K\rho\frac{\partial T}{\partial x},$$

in place of (1) and (2) we obtain the equations

$$\frac{du}{dt} = -R\frac{\partial \varphi}{\partial x} + lv, \tag{3}$$

$$\frac{dv}{dt} = -R\frac{\partial \varphi}{\partial y} - lu, \tag{4}$$

where $\varphi = T + KZ$.

If we use the natural coordinate system $\vec{S}$, $\vec{n}$ (where $\vec{S}$ and $\vec{n}$ are the direction along the tangent and the normal to the pilot balloon trajectory respectively) on an isosteric surface, which in a general case is not horizontal, we obtain the equations

$$\frac{dV}{dt} = -R\frac{\partial \varphi}{\partial S}; \tag{5}$$

$$K_\tau V^2 + lV = -\frac{\partial \varphi}{\partial n}. \tag{6}$$

where V is the velocity modulus, S is an element of the stream line, n is the normal to the stream line, K_T is trajectory curvature.

Assume that on pilot balloons there is only a temperature sensor. It is natural to require that it, like a radiosonde, has a radio transmitter transmitting the results of temperature measurements at the receiving point. In addition, using the radio transmitter signals it is possible to get a fix on the pilot balloon coordinates, which are necessary not only for its geographic tie-in, but also for determining the velocity and acceleration of the air currents.

It is impossible to obtain direct precise computations of pilot balloon altitude using equations (1)-(6) and information on velocity and acceleration. Therefore, we will examine variants of approximate computation of altitude on the basis of the mentioned equations.

Using equation (5) it is easy to derive its following analogue:

$$V\frac{dV}{dt} = -R\frac{d\varphi}{dt} + R\frac{\partial\varphi}{\partial t}. \qquad (7)$$

Integrating (7) along the trajectory, we obtain the expression

$$\frac{V_t^2 - V_0^2}{2R} = -(\varphi_t - \varphi_0) + \int_{s(t)} \frac{\partial\varphi}{\partial t}\,dt, \qquad (8)$$

where V_0, φ_0, V_t, φ_t are the values of the functions at the initial and final points of the trajectory respectively; the integral on the right-hand side

$$\int_{s(t)} \frac{\partial\varphi}{\partial t}\,dt$$

must be computed along a segment of the trajectory $S(t_0)$, $S(t)$. This integral can be computed only over an area with sufficient coverage with meteorological data by means of use of the results of numerical forecasting of the field φ. More approximate computations over a territory poorly covered with meteorological data are possible on the basis of the dynamic analysis method.

For relatively short time intervals not exceeding 6 hours, computations of the φ function are possible from equation (5) if it is assumed that in the course of the mentioned interval the field is stationary and the streamlines coincide with the trajectories.

The validation of such a method follows from a synoptic experiment in constructing trajectories, which on each pressure pattern chart are drawn by the weatherman along the isohypses. It goes without saying that such a method is more precise the shorter is the time interval after which new charts are plotted. However, these intervals are usually not less than six hours.

From the equation of motion it is easy to compute the mean value of the parameter $\partial\varphi/\partial S$, being a projection of the φ gradient onto the streamline.

In actuality, integrating equation (5) along the trajectory, we obtain the expression

$$\frac{V_t^2 - V_0^2}{2R} = -\int_{S(t_0)}^{S(t)} \frac{\partial\varphi[S(t),t]}{\partial S}\,dS,$$

from which we determine the mean value $\partial\varphi/\partial S$, equal to

$$\overline{\frac{\partial\varphi[S(\tau),\tau]}{\partial S}} = -\frac{V_t^2 - V_0^2}{2R\,\delta S}, \tag{9}$$

where $\delta S = S(t) - S(t_0)$, τ is some time interval within the interval t_0, t.

Expression (9) can be used for extrapolating the φ value if we introduce one of the two assumptions: 1) the φ field in the interval t_0, t is stationary, or to be more precise, its nonstationarity can be neglected; 2) the time τ can be related to some specific moment t_k, for example, coinciding with the middle of the interval. Then the extrapolation formula will have the form

$$\varphi[S(t_0) + \delta S(\tau)] = \varphi[S(t_0),\tau] + \overline{\frac{\partial\varphi}{\partial S}}\,\delta S, \tag{10}$$

where $\varphi[S(t_0), \tau]$ is determined at the launching point at the time $\tau = t_k$. In actuality, however, the point $S(t)$, τ, to be sure, will not coincide with the launching point due to disruption of stationarity of the field. However, we will considered the t_0, t interval to be quite small (not more than 6 hours) in order to neglect nonstationarity.

With respect to atmospheric pressure, its computations will not be discussed here in detail, since the algorithm is rather simple. Atmospheric density is determined from the equation for equilibrium of a pilot balloon in the atmosphere, that is, is expressed through a determination of the characteristics of each specific pilot balloon (weight, lift). After computing the "flight" density of the atmosphere the pressure value can be determined from the equation of state $P = R\rho T$, where T is the temperature measured during the time of flight.

After computing the pilot balloon flight velocities it is possible to compute some meteorological elements in the flight zone situated along the trajectory. If the trajectory is projected onto a horizontal plane, then from the equation

$$K_\tau V^2 + lV = -\frac{1}{\rho}\frac{\partial P}{\partial n_0}$$

it is easy to determine $\partial P/\partial n_0$, and accordingly, the $P(\delta n_0)$ value at the point situated on the normal to the plane of the trajectory:

$$P(\delta n_0) = P_0 + \frac{\partial P}{\partial n}\delta n_0,$$

where n_0 is the normal to the plane trajectory.

The interval δn_0 of linear extrapolation in a general case will be different in dependence on the degree of trajectory curvature. However, it can be asserted with great assurance that this interval will not be less than 300-500 km on both sides of the trajectory.

Similarly, using (6), we determine the value $\varphi(\delta n)$, which is of interest only in a case when at the point S, δn one of the functions determining φ is known, that is, T or Z, or P or Z. Then the second unknown function (Z or T) is computed when satisfying this condition.

Thus, if the φ value at some point on the trajectory is known, for example, after we have applied one of the methods described above for computing Z, it is possible to compute φ at a point situated on the normal to S(t).

An approximate determination of $P(\delta n)$ on an isosteric surface is made in the following way:

$$P(\delta n) = P(\delta n_0) + \frac{\partial P}{\partial Z}\delta Z = P(\delta n_0) - \rho g \delta Z, \tag{11}$$

where δZ is a still unknown value.

For computing δZ we introduce a formula determining the slope of the isosteric surface in the direction of the normal. From the differential equation for the isosteric surface, using the assumption of polytropicity of the atmosphere, we obtain the expression

$$\frac{\delta Z}{\delta n_0} = \frac{2\nu}{1-\nu^2}, \tag{12}$$

where $\nu = \frac{R\Gamma}{a_n}$, $a_n = K_\tau V^2 + lV$.

Thus, by computing the φ and P values at the point S, δn, we can make computations at this same point Z. Such an approach is also applicable in the negative direction of the normal $\vec{n}$.

Thus, in addition to obtaining a series of elements on the pilot balloon trajectory, it is possible to compute some elements (P, T, Z) to one side of the trajectory.

One of the effective applications of pilot balloon measurements involves the use of different expressions invariant along the trajectory. Adiabatic invariants (potential vorticity, potential temperature, different combinations containing entropy) are well known in dynamic meteorology.

However, for an isosteric trajectory it is possible to obtain invariants unrelated to the assumption of an adiabaticity of the processes. For this we will use the Ertel equation [4]:

$$\frac{d}{dt}\left(\frac{1}{\rho}\,\mathrm{grad}\,\psi\cdot\vec{\Omega}'\right)-\frac{1}{\rho}\,\vec{\Omega}'\,\mathrm{grad}\,\frac{d\psi}{dt}=\frac{1}{\rho}\,\mathrm{grad}\,\psi\cdot\left(\mathrm{grad}\,P\times\mathrm{grad}\,\frac{1}{\rho}\right), \tag{13}$$

where ψ is an arbitrary field function, $\vec{\Omega}' = \vec{\Omega} + 2\vec{\omega}$ is absolute vorticity, $\vec{\omega}$ is the vector of angular velocity of the earth's rotation.

If $\psi = \psi(P, \rho)$, then the right-hand side of (13) is equal to zero. For example, in place of ψ substituting the potential temperature, in an adiabatic approximation we obtain

$$I=(\gamma_a-\gamma)\,\frac{\theta}{T}\,(\Omega_z+l), \tag{14}$$

where γ_a is the adiabatic temperature gradient, Ω_z is the vertical component of the vorticity Ω vector.

If it is assumed that $\psi = \rho$, on the isosteric trajectory we obtain

$$\frac{d}{dt}\left(\frac{1}{\rho}\,\mathrm{grad}\,\rho\cdot\Omega'\right)=0, \tag{15}$$

where the operator d/dt is written for the standard coordinate system x, y, Z.

If in (15) we retain only terms having a maximum order of magnitude, that is, the vertical components of the scalar product, we obtain the expression

$$I=\frac{1}{T}\left(\gamma-\frac{g}{\kappa}\right)(\Omega_z+l), \tag{16}$$

in which, in contrast to (14), the adiabaticity condition was not used.

In the absence of moisture outflow and a turbulent influx of water vapor along the trajectory, the following equation is correct

$$\frac{dq}{dt}=0, \tag{17}$$

where $q = 0.622\,E(T_d)/P$ is specific humidity, E is the maximum elasticity of water vapor, T_d is the dew point.

After simple transformations, from (17) we obtain the invariant

$$I_q=\frac{\alpha}{T_d}+\ln T, \tag{18}$$

where $\alpha = L/AR_n$, L is the phase heat influx, A is the thermal equivalent of work, R_n is the gas constant for water vapor.

It is entirely obvious that expression (18) can be used for computing the dew point deficit $S = T - T_d$. In an adiabatic approximation on an isosteric trajectory $dT/dt = 0$. In actuality, the equation for heat influx to a unit air mass with $d\rho/dt = 0$ has the form $\varepsilon = c_v \frac{dT}{dt}$; then $dS/dt = 0$.

The quantity of condensing vapor is determined by the following equation [2]:

$$m = -\frac{dq^*}{dt} + \frac{q^*}{T}\frac{dT}{dt}, \tag{19}$$

where m is the quantity of water vapor condensing in a unit mass of air per unit time,

$$q^* = 0.622\frac{E(T)}{p}$$

is the specific humidity of saturation.

The value dq^*/dt in (19) for an isosteric trajectory is determined in the following way:

$$\frac{dq^*}{dt} = -q^*\left(a\frac{d}{dt}\left(\frac{1}{T}\right) + \frac{d\ln T}{dt}\right) = -q^*\frac{d}{dt}\left(a\frac{1}{T} + \ln T\right). \tag{20}$$

Accordingly, equation (19) can be represented in the form

$$m = q^*\frac{d}{dt}\left[\frac{a}{T} + \ln T^2\right], \tag{21}$$

where the right-hand side is determined on the basis of measurements.

In the diagnosis equation (21) can be used for approximate identification of zones of water vapor condensation, that is, cloud masses.

Since on the trajectory, in accordance with (16), we determined the value $\Omega_z + l$, then for some humidity characteristics it is desirable to use the invariant expressions obtained by Ye. Shvets and L. A. Kuznetsov for a plane trajectory [5, 6]

$$I_S = \ln(\Omega_z + l) - aS, \tag{22}$$

$$I_{T_d} = E(T_d)(\Omega_z + l)^b, \tag{23}$$

where a, b are some pseudoconstants whose values are given in the cited studies.

As already noted, expressions (22) and (23) were derived for a plane trajectory, since computations of the spatial trajectories are difficult. With the use of pilot balloons, strictly speaking, it is necessary to use invariants for a spatial (three-dimensional) trajectory. In these invariants, as we have seen, vorticity is not involved because in their derivation no need arose for excluding the derivative of vertical velocity using the vorticity equation.

Since on the trajectory, in accordance with (18), the dew point is computed and the temperature is measured, the relative humidity is easily determined using the formula

$$r = \frac{E(T_d)}{E(T)}. \qquad (24)$$

It follows from the above that measurement of temperature on the isosteric trajectory provides all the necessary information for computing the principal humidity characteristics. Due to neglecting of the turbulent influx of moisture these computations will contain errors whose value in summer will be maximum, and in winter minimum.

However, the final evaluations of the errors are possible only after carrying out the corresponding experiments with launchings of pilot balloons.

Using temperature and pressure at the flight altitude we compute the potential temperature

$$\theta = C(p)\, T^{1-\lambda}, \qquad (25)$$

where

$$C(p) = \left(\frac{1000}{Rp}\right)^{\lambda}, \quad \lambda = \frac{AR}{c_p}.$$

On the other hand, the θ value can be determined using the formula

$$\theta = T + \gamma_a h, \qquad (26)$$

where $\gamma_a = \frac{Ag}{c_p} \frac{T_i}{T_e}$.

T_i is the temperature of a particle adiabatically descending from the level P to the level P_0 = 1000 mb, h is the thickness of the layer from the level P_0 = 1000 mb to the flight level, T_e is the temperature of the air surrounding the particle. Evidently, T_i and T_e are functions of altitude.

Usually it is assumed that $T_i/T_e = 1$, then $\gamma_a = Ag/c_p$ = 0.98°C/100 m. Experience has shown that for computations of h the indicated simplifications are too approximate and lead to great errors. Thus, according to radiosonde data for August, September and November 1977 (Dolgoprudnaya) the mean absolute error in computing h with $\gamma_a = Ag/c_p$ for P = 500 mb exceeds 10 dam.

We will consider methods for refining computations of h. In place of (26) we will write the expression

$$\theta = T + \frac{Ag}{c_p} \int_0^h \frac{T_i}{T_e}\, dZ. \qquad (27)$$

Using the identity $\frac{T_i}{T_e} = 1 + \frac{\tau(Z)}{T_e}$, $\tau(Z) = T_i - T_e = (\gamma_a - \gamma) Z$,

we obtain the following expression for γ_a in the polytropic atmosphere:

$$\gamma_a = \gamma_a^0 \frac{T_p}{T_p - (\gamma_a^0 - \gamma) Z}, \tag{28}$$

where $\gamma_a^0 = \frac{Ag}{c_p}$, $\frac{T_i}{T_e} = \frac{T_p}{T_p - (\gamma_a - \gamma) Z}$,

T_p is temperature at flight altitude, Z is altitude.

Using (27) and (28) we obtain

$$h = \frac{T_p (1 - e^{-\beta})}{\gamma_a^0 - \gamma}, \tag{29}$$

where $\beta = \frac{\gamma_a^0 - \gamma}{\gamma_a^0}\left[\left(\frac{1000}{P}\right)^\lambda - 1\right]$, $\gamma \neq \gamma_a$.

With those same simplifications as in the derivation of (29), it is possible to obtain γ_a in the form of the following series:

$$\gamma_a = \gamma_a^0 + (\gamma_a^0 - \gamma) \sum_{k=1}^{n} \left(\frac{\gamma_a^0 Z}{T_e}\right)^k + (\gamma_a - \gamma)\left(\frac{\gamma_a^0 Z}{T_e}\right)^{n+1}, \tag{30}$$

where in the residual term it is feasible to assume that $\gamma_a = \gamma_a^0$.

An evaluation shows that it is sufficient to limit ourselves to the first two terms of series (30) so that the error in computation of γ_a does not exceed 1%. Special investigations are required for selecting the γ value in dependence on season, type of air mass, etc.

In the integration it is convenient to introduce into (27) the mean temperature T_m and for its exclusion use the known formula of atmospheric statics

$$h = \frac{R}{g} T_m \ln \frac{P}{P_0}.$$

Finally, taking into account the comments made above, we obtain the computation formula

$$h = \frac{\theta - T_p}{\gamma_a^0 \left[1 - \frac{(\gamma_a^0 - \gamma) R}{2g} \ln \frac{P}{1000}\right]} \tag{31}$$

or

$$h = \frac{T_p \left[\left(\frac{1000}{P}\right)^\lambda - 1\right]}{0.98\left[1 - \frac{0.98 - \gamma}{2g} R \ln \frac{P}{1000}\right]}. \tag{32}$$

The mean absolute error in computing h on the basis of formula (31) for P = 500 mb (August, September, November 1977, Dolgoprudnaya) was 2.3 dam.

It follows from the preceding that T_m is a function only of temperature at the flight level. If we neglect the slope of the isosteric surface, that is, we assume

$$\frac{\partial \rho}{\partial x} = \frac{\partial \rho}{\partial y} = 0,$$

then the following expressions are correct

$$\frac{1}{\rho}\frac{\partial P}{\partial x} = R\frac{\partial T}{\partial x}, \quad \frac{1}{\rho}\frac{\partial P}{\partial y} = R\frac{\partial T}{\partial y}, \tag{33}$$

in which the components of the pressure gradient are determined through the wind velocity values, measured on the trajectory.

Assuming the right-hand sides of (33) to be known, we will determine the components of the thermal wind using the known formulas

$$u_t = -\frac{gh}{lT_0}\frac{\partial T_m}{\partial y}, \quad v_t = \frac{gh}{lT_0}\frac{\partial T_m}{\partial x}, \tag{34}$$

in which $\partial T_m/\partial y$ and $\partial T_m/\partial x$ are determined using (31), and $T_0 = T + \gamma h$.

Using (32), we compute the velocity components for the geostrophic wind at the surface P = 1000 mb

$$u_0 = u + \frac{gh}{lT_0}\frac{\partial T_m}{\partial y}, \quad v_0 = v - \frac{gh}{lT_0}\frac{\partial T_m}{\partial x}. \tag{35}$$

Using measurements with pilot balloons it is easy to determine the components of the ageostrophic wind deviations

$$u' = -\frac{1}{l}\frac{dv}{dt}, \quad v' = \frac{1}{l}\frac{du}{dt}. \tag{36}$$

It must be emphasized that at the present time the values u' and v' are determined frequently using the formulas

$$u' = u + \frac{1}{l}\frac{\partial \Phi}{\partial y}, \quad v' = v - \frac{1}{l}\frac{\partial \Phi}{\partial x}, \tag{37}$$

in which the components of the geopotential gradient are computed using finite differences. Taking into account the large interval (250-500 km), the well-known peculiarities of the pressure field, causing difficulties in a sufficiently precise determination of the gradient, it must be assumed that the considered method is incorrect. It is not very clear how one should interpret the computations of u' and v' on the basis of formulas (37). By the finite differences method do we actually compute the ageostrophic components of the wind or only evaluate the error in computing the geostrophic wind?

We feel that the use of measurement data from pilot balloons is a more natural method for evaluating the ageostrophic wind.

Thus, preliminary evaluations show that the use of constant-volume pilot balloons make it possible to obtain a great volume of information on state of the atmosphere.

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SIXTIETH BIRTHDAY OF YEVGENIY MIKHAYLOVICH DOBRYSHMAN

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 p 119

[Unsigned article]

Professor Yevgeniy Mikhaylovich Dobryshman, Doctor of Physical and Mathematical Sciences, member of the editorial board of the journal METEOROLOGIYA I GIDROLOGIYA, a well-known professional meteorologist, author of a number of original and profound investigations in the field of dynamic meteorology and numerical weather forecasting methods well known in our country and abroad, marks his 60th birthday on 17 April.

The editorial board of this journal sincerely congratulates him on this memorable date and wishes him long years of life, good health and today's inexhaustible creative energy.

SEVENTY-FIFTH BIRTHDAY OF KHOREN PETROVICH POGOSYAN

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 p 120

[Article by workers at the USSR Order of Lenin Hydrometeorological Scientific Research Center]

[Text] Doctor of Geographical Sciences Professor Khoren Petrovich Pogosyan, outstanding professional meteorologist, Meritorious Scientific Worker RSFSR, on 2 April 1979 marked his 75th birthday and 55 years of scientific-pedagogic and social activity.

The scientific activity of Kh. P. Pogosyan began in 1932. His first studies were devoted to an investigation of the dynamics of atmospheric processes. He was one of the first to recognize the exceptional importance of using aerological observations for an analysis of atmospheric processes and weather forecasting and was one of the initiators of the creation of a network of aerological stations in the Soviet Union (1936-1937).

Investigation of atmospheric processes at the earth's surface and aloft enabled Kh. P. Pogosyan in 1943 for the first time to introduce routine compilation of prognostic surface pressure charts and pressure pattern charts, for which he was awarded a prize by the USSR Council of Ministers.

In the early 1940's Kh. P. Pogosyan, in collaboration with N. L. Taborskiy, developed a method of advective-dynamic analysis, on the basis of which it was possible to formulate a number of qualitative rules for the prediction of synoptic processes.

In 1950 Kh. P. Pogosyan was named director of the Main Geophysical Observatory, and somewhat later the first deputy chief of the Hydrometeorological Service of the USSR Council of Ministers. Despite a great amount of administrative work, during these years he continued scientific research and developed a new theory of the moisture cycle in the atmosphere.

His investigations of general circulation of the atmosphere and jet streams are well known, as are his investigations of the interrelationships between processes in different regions of the Northern Hemisphere, the relationships between vertical processes, and many others.

During recent years Kh. P. Pogosyan has devoted great attention to study of the quasi-two-year wind cyclicity in the equatorial stratosphere. He proposed a new hypothesis of the origin of this cyclicity.

During the time of his scientific activity Kh. P. Pogosyan published about 200 scientific studies, including 20 monographs, collections of articles and atlases. His studies have been translated into foreign languages and have been published abroad.

Kh. P. Pogosyan devotes great attention to the training of young scientists and specialists, as well as to the popularization of meteorological knowledge. His brochures and the study aid ATMOSFERY ZEMLI (Earth's Atmosphere) have been awarded all-union prizes and first- and second-degree diplomas. Many results of his investigations have been included in textbooks and study aids in the field of meteorology.

Kh. P. Pogosyan does much scientific-social work, being a member of the boards of the Central Scientific-Methodological Council of the "Znaniye" (Knowledge) Society, a member of the Central Scientific Council of the Geographical Society of the USSR and the presidium of its Moscow Affiliate, a member of scientific councils and editorial boards of journals.

For his great contribution to the development of Soviet meteorology Kh. P. Pogosyan was awarded two orders and five medals.

We sincerely congratulate Khoren Petrovich on his anniversary and wish him good health, good spirits and further successes in his fertile work.

SEVENTIETH BIRTHDAY OF IOSIF ADAMOVICH YANKOVSKIY

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 p 121

[Article by his work comrades]

[Text] Iosif Adamovich Yankovskiy, Candidate of Geographical Sciences, docent, and senior scientific specialist at the Main Geophysical Observatory, marked his 70th birthday in February 1979.

I. A. Yankovskiy began his work activity in 1926. In 1931 he graduated from the air force military school and in 1940 from the military faculty at the Moscow Hydrometeorological Institute.

In June 1941 I. A. Yankovskiy was named to the post of senior inspector of the division of academic institutions of the Main Administration of the Hydrometeorological Service of the Red Army. In April 1942, in accordance with his request he was assigned to the active army and served there until 1945 in the posts of head of the meteorological service of an attack aviation group, an aviation division and an aviation corps of the High Command. In 1953-1966 he was a department head at the Leningrad Military Engineering Academy imeni A. F. Mozhayskiy.

Beginning in March 1966 I. A. Yankovskiy worked as a senior scientific specialist at the Main Geophysical Observatory. In this post he carried out much scientific and scientific-methodological work in creating a national network for monitoring the contamination of atmospheric air. He is the author of 80 published studies, including a number of textbooks and academic aids.

A member of the Communist Party since 1932, Iosif Adamovich engages in much social and political work. He is secretary of a Party organization and is Chairman of the Council of Veterans of the Great Fatherland War.

The Party and government have a high regard for his services and scientific activity, presenting him with a series of government awards: the orders of Lenin, Red Banner, Fatherland War First and Second Degree and Red Star and the medals "For Valor" and "For Military Services," as well as 12 others.

Iosif Adamovich meets his 70th birthday full of vigor and creative thought. We wish him good health and further successes.

CONFERENCES, MEETINGS AND SEMINARS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 121-127

[Article by I. A. Yankovskiy, O. I. Vozzhennikov, V. P. Gavrilov, B. S. Ustyuzhanin and A. V. Shcherbak]

[Text] An All-Union Seminar on "Meteorological Support of Work on Protecting Atmospheric Air Against Contamination" was held during the period 9-12 October 1978 in Moscow at the All-Union Exhibition of Achievements in the National Economy. The seminar was attended by more than 160 representatives from 80 organizations in the country. Forty-two reports and communications were presented on meteorological, hygienic and technical aspects of the problems involved in contending with atmospheric contamination.

By request of the State Committee on Hydrometeorology the seminar was opened by M. Ye. Berlyand. In his introductory words he noted the substantial attainments in study of the state of the air basin, which are gradually being introduced into the work practice of the National Service for Observing and Monitoring Contamination of the Environment. An example of the effective use of research results is the development of systematic instructions and manuals on monitoring atmospheric contamination and State Standards in the series "Preservation of Nature. Atmosphere."

A lengthy report by M. Ye. Berlyand, entitled "Principles for Scientific-Technical Prediction of Atmospheric Contamination," gave an expanded description of the status of the problem at the modern stage. It gave the principles for formulating a forecast far in advance, gave the basis for the scientific-technical forecasting method developed at the Main Geophysical Observatory, and defined the problems facing the National Service for Observing and Monitoring Contamination of the Environment.

A report entitled "Organization of a Service for Predicting Air Contamination and Regulation of Effluent," presented by L. R. Son'kin and B. B. Goroshko, discussed the problems relating to routine prediction of air contamination. It gave the results of tests of the methods developed and examined the status of work at the Administrations of the Hydrometeorological Service and Scientific Research Hydrometeorological Institutes. It was noted that at a number of Administrations of the Hydrometeorological Service groups have been established for the prediction of air contamination. Specialists have examined the scientific fundamentals and principles for the development of measures for reducing effluent during periods of dangerous meteorological conditions.

I. N. Ponomarenko presented a report entitled "Short-Range Forecasting of Air Contamination." He presented a scheme for a synoptic-statistical method for the forecasting of air contamination by sulfur gas, nitrogen dioxide and dust.

L. M. Neronovoy and S. P. Ponomarenko, in a report entitled "Improved Method for the Short-Range Forecasting of the Meteorological Potential of Atmospheric Contamination," revealed the essence of potential. It was shown that it includes synoptic conditions favoring the accumulation and dissipation of impurities, and also the meteorological parameters determining turbulent mixing and horizontal transfer.

A report by E. Yu. Bezuglaya, L. I. Yelekoyeva and Ye. A. Razbegayeva, entitled "Climatic Conditions and Atmospheric Contamination," included maps of regionalization of the territory of the USSR with respect to the climatic conditions for the dissipation of impurities from low sources characteristic for urban conditions. A method for evaluating these conditions is given. In another report E. Yu. Bezuglaya dealt with the results of comparisons of the results of measurement of the concentration of carbon monoxide using a GMK-3 automatic gas analyzer and a SV-7633 titrometric gas analyzer.

Problems involved in the status and development of the network of the National Service for Observations and Monitoring of Atmospheric Contamination were the subject of a report by I. V. Tsvetkov and I. A. Yankovskiy. It was noted that in the system of the Hydrometeorological Service there has been monitoring of the state of the air basin in 255 cities in the USSR at 1,300 first and second category posts. A major place in the report was devoted to the introduction of new technical means for the monitoring of atmospheric contamination, an increase in the quality and effectiveness of operation of the observation network.

The principal directions in improvement of methods for analysis of atmospheric contamination were examined in a report by N. Sh. Vol'berg. Here particular attention was devoted to the prospects for replacing fluid absorption instruments with thin-film solid sorbents and the feasibility of development of centralized laboratories.

The report of R. I. Onikul was entitled "Study of Sources of Contamination and Normalization of Effluent of Contaminating Substances in the Atmosphere." This report dealt with measures for increasing the quality of determination of effluent during the planning and operation of enterprises, and also computations of atmospheric contamination on an electronic computer using a standardized program.

Ye. L. Genikhovich familiarized seminar participants with methods for computing atmospheric diffusion. He analyzed the results of recent studies carried out at the Main Geophysical Observatory on creation of methods for computing the mean annual and one-time concentrations for sources of different type, on the ground and aloft, and also in the field of modeling of the propagation of impurities in hilly terrain. I. G. Gracheva told about allowance for the influence of local relief on the nature of propagation of impurities from a source.

A report by Ye. S. Selezneva, entitled "Sulfur Balance in the Atmosphere of an Industrial Region," examined the results of computations of dry precipitation of sulfates with use of some experimental results and climatic characteristics of precipitation (its quantity, intensity and duration).

O. P. Petrenchuk and N. V. Nesterova reported on the transport of impurities over great distances and changes in the chemical composition of atmospheric precipitation.

Ye. N. Rusina analyzed the possibility of using the AT-50 actinometer with wide-band filters for evaluating aerosol turbidity in different regions of Eurasia.

I. I. Solomatina reported on the state, analysis and generalization of data on the discharge of harmful substances into the atmosphere.

F. V. Korshenko told about the dependence of contamination of the Kiev air basin on meteorological conditions.

A report by a group of authors from the Khar'kov Institute of Radioelectronics, S. I. Babkin, Yu. I. Pakhomov, Ye. G. Proshkin and Yu. N. Ul'yanov examined acoustic methods for the remote study of meteorological parameters for predicting atmospheric contamination.

On the basis of experimental data on the scattering of impurities in the atmosphere, emerging from the stacks of regional thermoelectric power stations with a height of 320 m, B. B. Goroshko in his report mentioned the satisfactory agreement of experimental and computed data on the concentrations of the investigated impurities. The results confirm the possibility of a broad application of the method for computing concentrations set forth in the UKAZANIYE PO RASCHETU RASSEIVANIYA V ATMOSFERE VREDNYKH VYBROSOV (Instructions on Computation of Scattering of Harmful Effluent in the Atmosphere) (SN 369-74), including cases of the highest stacks.

A report by N. S. Burenin was devoted to a study of air contamination by auto transport. It gave a validation of a complex of observations on highways, including measurements of the content of harmful substances ejected by auto transport.

A report by G. V. Dmitriyeva and V. P. Petrov gave an analysis of some peculiarities of year-to-year changes in air contamination in the European USSR during the period from 1973 through 1975. The report gives the relative characteristics of the mean annual concentrations of different impurities in cities. G. V. Dmitriyeva also reported on the principal types of synoptic conditions favorable for increasing the contamination of urban air from low sources.

Ye. B. Steklogorov told about means and methods for a complex evaluation of the total contamination of atmospheric air when carrying out monitoring of the state of the air basin.

V. V. Kiselev reported on computation of the mean concentrations of harmful impurities in the atmosphere on the basis of network data. He proposed that computations be made using the mean values of a linear unbiased evaluation with a minimum dispersion. It was shown that its use leads to a decrease in the error in computations by a factor of 1.5-2.

The results of an aircraft investigation of atmospheric contamination in the Ukraine were discussed in a report by A. V. Tkachenko, V. M. Shoshin and L. A. Ramenskiy.

A report by I. B. Pudovkina, et al. was devoted to a statistical processing and analysis of routine information on atmospheric contamination in priority cities.

The results of prediction of atmospheric contamination for the industrial region of Uzbekistan were reported by N. V. Koroleva. It was shown that in definite synoptic situations there is a coincidence of the degree of air contamination in a number of cities of the industrial region. This made it possible to organize prediction of background contamination as a whole for several cities.

V. I. Kuznetsov told about experience in organizing and carrying out an expeditionary investigation of air contamination of cities in the Northwest in 1976-1977.

Considerable interest was shown in a communication by the deputy chief of the Administration for Study of Environmental Contamination of the State Committee on Hydrometeorology I. V. Tsvetkov. It gave an analysis of the present status of the National Service for Observing and Monitoring Atmospheric Contamination and characterized measures for its long-range development. Attention was given to the need for a further increase in the effectiveness and quality of operation of all branches of the service. In conclusion he gave thorough responses to numerous questions from seminar participants.

At the end of the seminar there was a special session of the section of professional forecasters concerned with prediction of the contamination of atmospheric air. Those speaking at this session, L. R. Son'kin, L. M.

Neronova, I. P. Ponomarenko, I. A. Shevchuk, M. A. Gol'dberg, Ye. L. Genikhovich, S. I. Ponomarenko and others related to the problems involved in routine prediction of air contamination and evaluation of the probable success of forecasts.

I. A. Yankovskiy

A Second Interdepartmental Seminar on Applied Problems in Atmospheric Diffusion was held at the Institute of Experimental Meteorology during the period 23-25 October 1978. The seminar was attended by 60 persons from 15 organizations. Nineteen reports were heard at the seminar. Their subject matter can be arbitrarily divided into three groups: theoretical aspects of description of turbulent diffusion, measurements of diffusion parameters and problems related to practical problems of turbulent diffusion.

Important aspects of the problem of preserving the environment are the prediction and determination of the statistical characteristics of air contamination, averaged for different time intervals, including the mean long-term values. The solution of these problems involves the necessity for creating a method for computing the concentrations of impurities using data from standard observations or on the basis of information present in weather forecasts. Specifically these problems were dealt with in a report by D. L. Laykhtman, F. A. Gisina, G. A. Natanzon and S. M. Ponomareva (Leningrad Hydrometeorological Institute). The method proposed in the report was based on the joint solution of equations for the atmospheric boundary layer and turbulent diffusion and makes it possible to determine the distribution of the impurity in the lower layer of the atmosphere with different external parameters, a method for determining which has been developed. The method makes it possible to take into account the peculiarities of vertical structure of the atmospheric boundary layer.

A theoretical study of the influence of dangerous conditions on the propagation of impurities from high and surface sources was the subject of a report by N. Berlyand, R. I. Onikul and S. S. Chicherin (Main Geophysical Observatory). They proposed a method for computing the mean annual concentrations of harmful substances using data from gradient observations in different climatic zones of the country. In addition, the authors told about methods for norm-setting for industrial effluent into the atmosphere of cities, setting of the maximum admissible effluent from enterprises and calculations of the extent of sanitary-protective zones.

A study by V. P. Gavrilov (Institute of Experimental Meteorology) examined the transverse scattering of an impurity in the neutrally stratified surface layer of the atmosphere from an instantaneous point surface source. On the basis of two variants of a semiempirical model of turbulent diffusion with coefficients dependent only on altitude, or only on time, using the moments method it was possible to derive expressions for transverse dispersions σ_y^2 (z,t). The author has obtained analytical evaluations of the constants determining the behavior of dispersions in accordance with similarity theory and the relationship between the constants of the two variants of the semiempirical theory is found.

O. I. Vozzhennikov (Institute of Experimental Meteorology) told about a model for computing the dispersion of the concentration of an impurity from a local source in the surface layer of the atmosphere. The closing of the system of initial equations was accomplished using formulas similar to the formulas for approximate Kolmogorov similarity. The scale length, determining mixing in the diffusion layer, is dependent both on diffusion time and on the turbulence scale. The author devoted the second part of his report to computation of the number of excesses of a stipulated level by concentration fluctuations at a fixed point in space.

A great number of studies were devoted to an experimental investigation of transverse diffusion in the atmospheric surface layer. These gave an evaluation of the absolute σ_y^2 or relative S_y^2 dispersion in the form of power functions of the distances from a source and less frequently the time of diffusion with experimentally determined constants. Usually the theoretical validation of the determined empirical dependences rests on the theory of homogeneous turbulence. However, in the atmospheric surface layer, for conditions close to neutral, the empirical data agree best with the relationships of the hypothesis of Lagrangian turbulence similarity:

During 1975-1977 specialists at the Institute of Experimental Meteorology carried out experiments with smoke plumes in the steppe zone of the Ukraine, in Southeastern Kazakhstan and on the shores of the Sea of Azov. The results of these experiments, and in particular, evaluations of the constants α_y and β_y, were examined in a report by Ye. K. Garger, A. V. Naydenov and D. B. Uvarov (Institute of Experimental Meteorology) entitled "Transverse Diffusion in the Atmospheric Surface Layer."

The influence of the rate of dry precipitation v on the field of concentration of an impurity in the atmosphere was examined in a report by N. L. Byzova, N. A. Krotova and G. A. Natanzon (Institute of Experimental Meteorology, Leningrad Hydrometeorological Institute). A numerical solution of the semiempirical equation of turbulent diffusion indicated that all the characteristics of the surface concentration and flow experience changes in the region $0.01 < v/u_* < 5$; when $v/u_* < 0.01$ there is a virtually total reflection, whereas with $v/u_* > 5$ there is total absorption. In addition, in the report, on the basis of an approximate solution of the stationary diffusion equation with $K = \varkappa u_* z$, the authors proposed a physical interpretation of the rate of dry precipitation and two methods for its determination are given.

A report by F. A. Gisina and G. A. Natanzon (Leningrad Hydrometeorological Institute), entitled "Comparative Analysis of Some Methods for Computing the Concentrations of Impurities Entering from Industrial Sources," gave experimental data on the vertical and horizontal dispersions. They gave an estimate of the mean square values of the scatter of data under different meteorological conditions. In addition, a study was made of the dependence of the maximum concentrations obtained on the basis of different computation methods for the basic parameters -- height of source, stratification and wind velocity.

A report by V. B. Kiselev (Main Geophysical Observatory) was devoted to an analytical study of the influence of the diffusion coefficient profile and wind velocity on the maximum surface concentration $q_m(z = 0)$. Representing the change $q_m(z = 0)$ in the form of a linear functional of wind velocity and the diffusion coefficient, the author succeeded in finding the profiles of these parameters exerting the greatest influence on the $q_m(z = 0)$ values.

A report by V. P. Gavrilov and M. A. Novitskiy (Institute of Experimental Meteorology) examined the influence of stratification on the diffusion characteristics of a cloud of impurity during scattering in the atmospheric boundary layer. The scattering of an impurity from an instantaneous point source was analyzed on the basis of a numerical solution of a system of equations for the atmospheric boundary layer and equations for the moments of concentration. As a model of the atmospheric boundary layer the authors selected the Wipperman model. The horizontal diffusion coefficients were assumed to be proportional to the coefficient of vertical turbulent exchange. The report also covered the problem of the dependence of dispersions on the height of the impurity source.

A study of the processes of heat and moisture transfer, with horizontal diffusion taken into account, was presented by B. G. Vager and Ye. K. Nadezhina (Main Geophysical Observatory). With a stipulated wind velocity field it was possible to obtain the temperature distribution in the neighborhood of warm spots. The study also dealt with the problem of taking into account the terms in the heat transfer equation containing nondiagonal components of the tensor of exchange coefficients.

The seminar participants displayed great interest in a report by N. L. Byzova entitled "Review of Methods for Obtaining Parameters Using Measurement Results." On the basis of data in the literature the author analyzed different methods for determining mesoroughness $\hat{z}_0$ of the upper boundary of the atmospheric surface layer and stability parameters. It is noted that at the present time there are still no adequately convincing proofs as to the advantage of one method or another for determining the mentioned parameters.

The possibility of use of spline functions in solving practical problems relating to the atmospheric boundary layer was discussed by B. G. Vager (Leningrad Institute of Civil Engineers). Spline function methods are used in reconstructing the fields of meteorological elements on the basis of experimental data and also for obtaining a numerical solution of the equations of motion for the atmospheric boundary layer. In the author's opinion, spline solutions have some advantages in comparison with a finite-difference solution of the equations, since spline solutions can be represented in analytical form, which is impossible for finite-difference solutions.

The Division of Experimental Research of the Ukrainian Scientific Research Hydrometeorological Institute carried out a series of flights using an IL-14 aircraft laboratory for the purpose of studying the propagation of smoke from the stacks of thermoelectric power stations. The results of an investigation of the total plume forming as a result of merging of the plumes from individual stacks were presented in a study by V. S. Antonenko, L. A. Geyko, L. A. Ramenskiy, A. V. Tkachenko. During the flights it was possible to determine the geometrical parameters of the plume, measure the concentration of aerosols and study the temperature and turbulence fields. On the basis of measurements and computations a study was made of the peculiarities of formation of the total plume under different meteorological conditions and there was a discussion of the possibility of taking them into account in practical diffusion problems.

A report by E. P. Volkov, V. I. Kormilitsin and L. A. Ramenskiy contained methodological procedures and experimental data on investigation of the rising of gases from stacks at a thermoelectric power station. The registry of meteorological characteristics of the atmosphere was carried out at the levels 0.5, 2, 10, 23 and 45 m aboard an IL-14 flying laboratory. The trajectories of the plumes from stacks were determined using phototheodolites and the aircraft sounding of plumes. Experimental data were cited on the rising of a smoke plume in a layer up to 1,500 m.

A report by N. V. Vonorova, L. R. Orlenko and O. B. Shklyarevich (Main Geophysical Observatory) was devoted to a method for computing the vertical wind profile in the lower 300-m layer on the basis of synoptic information for different types of underlying surface. For computing the wind profile the authors proposed quantitative relationships between the external parameters, on the one hand, and surface winds and wind coefficients, on the other.

G. B. Mashkova and L. M. Khachaturova (Institute of Experimental Meteorology) presented materials from comparisons of different methods for determining stability classes. The comparison was made using measurement data registered on the high mast at the Institute of Experimental Meteorology at Obninsk. The authors concluded that all the considered classifications are close to one another. In more than 50% of the cases the discrepancy does not exceed one class, and in 86% of the cases -- two classes.

I. L. Levitin (Leningrad Division, Heat Engineering Institute) told about the influence of water vapor discharge from industrial coolers on the microclimate of the surrounding territory.

A report by V. S. Cherednichenko (Kazakh Scientific Research Hydrometeorological Institute) proposed a method for computing the characteristics of vertical movements in the free atmosphere on the basis of data from rawin observations.

A report by Yu. M. Zhavoronkov and A. A. Bykov dealt with approaches to formulation of a climatological model of atmospheric diffusion. This is a basic computation tool for developing a long-range strategy of measures for safeguarding the atmosphere.

As a result of the work, the seminar adopted a resolution in which it was noted that during the two years which had elapsed since the first seminar (1976) there has been a significant development of studies of a theoretical research, experimental and applied character. The seminar took note of the inadequate development of experimental field work. Due to the complexity and great cost of such work it is felt that there is a need for cooperation among different institutes both for the purpose of planning experiments and also for their organization and implementation.

O. I. Vozzhennikov and V. P. Gavrilov

A conference-seminar on the national water inventory and water use in the Transcaucasian republics was held in Baku during the period 3-6 October 1978. It was attended by professional hydrologists of the State Committee on Hydrometeorology, different ministries and departments, a total of 75 persons.

In opening the conference-seminar, I. A. Aliyev, head of the Hydrometeorological Administration Azerbaydzhan SSR, emphasized that this was the first regional conference of specialists directly responsible for the introduction of a unified system for a national water inventory and water use. It is of great importance for more precise determination of the water resources of Transcaucasia and their most rational distribution both among the republics and within them.

In his introductory words a representative of the State Committee on Hydrometeorology, E. V. Buryak, defined the principal tasks of the conference-seminar. He brought the attention of the conferees to the timeliness of discussion of a broad range of problems involved in the inventory of water resources and their use in the national economy, the importance of increasing accuracy in this inventory and also the need for a further improvement in the contacts and interrelationships among the professional hydrologists of different ministries and departments.

The participants in the conference-seminar, in discussing the state of the national water inventory and water use in the Transcaucasian republics, noted that in accordance with the "Instructions on the State Inventory of Waters and Their Use," the organizations and institutes of the State Committee on Hydrometeorology, USSR Geology Ministry and USSR Water Management Ministry in the Transcaucasian republics have carried out a number of organizational and scientific-technical measures. In particular, the administrations of the hydrometeorological services of the republics of Transcaucasia are carrying out an inventory of surface water resources, determining channel water balances for individual river reaches (Kura, Alazan', Razdan, Vorotan, Sulak, and others), and also evaluating the accuracy of water inventories and water use in the national economy (reports of B. S. Ustyuzhanin (State Hydrological Institute), D. A. Baindurashvili (Tbilisi Hydrometeorological Observatory), M. O. Oganisyan (Yerevan Hydrometeorological Observatory), Sh. M. Agayev (Baku Hydrometeorological Observatory)).

The agencies of the USSR Water Management Ministry, in collaboration with the agencies of the State Committee on Hydrometeorology and the USSR Geology Ministry, have prepared lists of enterprises, organizations and institutes the use of water by which is subject to state inventory. They have determined the water users taking water directly from surface and underground water sources or dumping waste water into them, as well as the sequence for transmitting to the State Committee on Hydrometeorology and its agencies data from the inventory of water use. Joint work has been completed on determining the siting of water-intake and water-discharge structures and their tie-in to the mouth of rivers or other fixed points.

The agencies of the USSR Water Management Ministry have introduced interim instructions on conducting an initial inventory of water use, on the preparation of statistical reports, on the reduction of the data received by water users to digital form and their processing on an electronic computer, and on the use of water-gaging instruments for inventorying water use (reports of the head of the section on water inventory and recording P. V. Kovsh, representatives of the State Committee of the Council of Ministers Armenian SSR on Use and Preservation of Surface and Ground Water Resources V. M. Petrosyan and M. A. Kamelyan, the head of the section on inventory of water resources and the water balance of the Water Management Ministry Azerbaydzhan SSR F. D. Suleymanov and a representative of the State Committee of the Council of Ministers Georgian SSR on the Preservation of Nature).

The agencies of the USSR Geology Ministry have carried out an inventory of ground water resources, have developed and introduced standardized forms to be employed in the initial ground water survey, have carried out regionalization of the territory of the USSR (including the Transcaucasian republics) with respect to ground water resources for developing a system for inventorying ground water resources and have coordinated with the State Committee on Hydrometeorology in relation to the range of data to be collected in the inventory of ground water resources (I. A. Barkalov, All-Union Scientific Research Institute of Hydrology and Engineering Geology).

The territorial administrations and scientific institutes of the USSR Geology Ministry have carried out an analysis of the status of inventorying of the discharge of ground water at operating water intakes, technical state of water-intake boreholes and their outfitting with gages and the operating regime of water intakes (reports of representatives of the Geology Administration Council of Ministers Azerbaydzhan SSR F. Sh. Aliyev, B. M. Samedov and others and a section head in the Geology Administration of the Council of Ministers Armenian SSR S. O. Martirosyan).

The scientific institutes of the State Committee on Hydrometeorology, USSR Geology Ministry and USSR Water Management Ministry are developing unified methodological instructions on the national inventory of waters and water use, the first variant of which, according to plans, is to be prepared in 1979.

At the same time, the speakers and discussion participants mentioned a number of shortcomings holding back the introduction of the system for the inventorying of waters and their use in the Transcaucasian republics.

For the purpose of more successful adoption of the national inventory of waters and their use in conformity to a unified system for the USSR in the Transcaucasian republics the conference-seminar recommended:

-- that the local agencies of the State Committee on Hydrometeorology, USSR Geology Ministry and USSR Water Management Ministry establish closer interaction with one another;

-- that it be brought to the attention of the USSR Water Management Ministry that there is a need for increasing the accuracy in inventorying water use with respect to both qualitative and quantitative indices and that there be technical outfitting of observation stations at water-using enterprises and also there must be tighter checking on the completeness and quality of reporting on form 2-tp (water management) submitted by water users;

-- that the Geology Ministry be asked to examine the problem of formulation, in the next few years, of a general scheme for the use of ground water in the Transcaucasian republics;

-- that the local agencies of the State Committee on Hydrometeorology, USSR Geology Ministry, USSR Water Management Ministry, USSR Power Ministry and other interested organizations carry out special hydrological studies for investigating the conditions for the formation and loss of runoff in the Kura basin, taking into account that in computations of channel water balances in individual reaches of rivers and canals, and also in evaluating water resources and their use, there are considerable nonclosures of data.

B. S. Ustyuzhanin

A preparatory meeting for the Tenth Conference of the Danubian Countries on Hydrological Forecasts was held at Vienna (Austria) during the period 11-13 October 1978. The conference was attended by delegations from Austria, Bulgaria, Hungary, USSR, West Germany, Czechoslovakia and a representative of the Danubian Commission Secretariat.

The preliminary meeting approved the proposal of the Austrian delegation on the holding of the Tenth Conference of the Danubian Countries on Hydrological Forecasts during the period 10-15 September 1979 at Vienna, and also discussed organizational problems and the agenda for carrying out the Tenth Conference and its scientific program.

At the Tenth Conference plans call for the presentation and discussion of reports on the following subjects applicable to the Danube basin:

-- short-range forecasting of precipitation, water levels and discharges, especially for high waters;
-- long-range and superlong-range prediction of water levels and discharges;
-- prediction of water low levels and discharges;
-- prediction of ice phenomena and elements of the thermal regime;
-- prediction of channel processes and runoff of sediment;
-- evaluation of the accuracy and national economic effectiveness of hydrological forecasts.

The principal objective of the Tenth Conference is an exchange of experience in the practical use of forecasting methods among all the Danubian countries and presentation of communications from individual researchers on new results and work which has been done.

The retention of a practical direction for the scientific program of the Tenth Conference (as for the preceding Ninth Conference) is motivated, in the opinion of the Soviet delegation, by the need for reflection, in the reports presented at the conference, of matters involved in the practical use of forecasts, their social importance, probable success and economic effectiveness, as well as the organization of a notification and warning service for dangerous hydrological phenomena.

Also to be presented at the conference will be reports devoted to mathematical models of runoff and their use for the monitoring of water resources, mathematical models of rain-induced high water and others with an indication of the possibility of their use in the development of forecasting methods in the Danube basin.

The preliminary meeting drew attention to the need for the reports to take into account man's influence on hydrological processes in the development of forecasting methods.

It is also proposed that the program for the Tenth Conference include national reports of all the Danubian countries on the methods which they use for hydrological forecasting applicable to the rivers of the Danube basin.

The working languages of the Tenth Conference will be Russian and German. The preparation of Soviet scientists and specialists for participation in the Tenth Conference of Danubian Countries on Hydrological Forecasts has been delegated by the USSR State Committee on Hydrometeorology to an organizing committee, set up at the USSR Hydrometeorological Center. All questions concerning the impending conference should be directed to that committee.

NOTES FROM ABROAD

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 pp 127-128

[Article by B. I. Silkin]

[Text] As reported in ANGEWANDTE CHEMIE, Vol 17, p 366, 1978; NEW SCIENTIST, Vol 79, No 1114, p 342, 1978, until now it has been assumed that molecules of chlorofluorocarbon are inert, in no way participating in reactions transpiring in the lower layers of the atmosphere and are accumulated there in ever-greater quantities.

Such an opinion, formulated in 1974 by the American scientists S. Rowland and M. Molina, gave rise to lively discussion and calls for the universal banning of the use of aerosols into which these substances enter.

Recently a group of specialists at the Institute of Ecological Chemistry (Munich, West Germany), headed by Professor F. Korte, demonstrated that the chlorofluorocarbons CCl_3F and CCl_2F_2 have the capacity for being absorbed by silica gels (silicon dioxide gels) and, being subjected to the influence of solar UV radiation, are photodissociated, releasing carbon monoxide and dioxide and hydrogen chloride in the presence of oxygen.

In their experiments the researchers selected silica gels due to their adsorption characteristic, similar to many varieties of dust and sand really existing in the lower layers of the earth's atmosphere. In several cases irradiation was carried out over a period of several weeks and was accompanied by prolonged mixing of the adsorbent and gas phases.

It was also established that the intensity of processes of breakdown of molecules is essentially dependent on the wavelength of the UV radiation. The irradiation carried out through quartz glass caused the formation of considerable quantities of molecular chlorine and intensification of the rate of breakdown of molecules.

As reported in INDUSTRIAL RESEARCH AND DEVELOPMENT, Vol 20, No 5, p 58, 1978, it was established more than 200 years ago that along the eastern shores of Australia, washing the states of Queensland and New South Wales, there is a current which moves from north to south. This stable flow is clearly traced from the northern part of the Coral Sea to the southern regions of the Tasman Sea. However, the factors responsible for such a significant current have remained unknown because investigations have been impeded by the dangers of navigation in the waters of the Great Barrier Reef, abounding with coral formations.

In 1977, the Organization of Scientific and Industrial Research of Australia, in collaboration with the Australian Navy, undertook a study, lasting more than 11 months, of movements of water masses using a system consisting of twelve freely drifting buoys and an artificial earth satellite collecting data accumulated by instruments mounted on these buoys.

As a result it was established that the current has a considerably more complex structure than had been assumed earlier. It consists of three funnel-shaped eddies with diameters attaining 200 km. The direction of movement of these annular currents is counterclockwise. The largest of these is observed on the traverse Coffs Harbour and Jervis Bay (Queensland), the others in the region of Cape Howe and along the northern shore of Tasmania.

The reasons for such a phenomenon were also clarified. The heated tropical waters in the northern region of the Coral Sea experience the influence of the earth's rotation, which leads to their "overflowing" over a great area over the colder underlying layers. They are replaced, from the deep layers, by cooled water masses surrounding the warm "islands" formed in such a way.

Measurements have shown that at the center of such "islands" the sea level is approximately 1 m higher than the level of the surrounding space, which is probably directly related to the surge of warm waters. The forces of terrestrial gravitation and planetary rotation bring such warm "lenses" into rotation, which acquires a counterclockwise direction. It was found that the communications received during recent decades indicating that supposedly this current does not exist are attributable to the discontinuity of the current: when the water level at the center of a warm "island" comes into equilibrium with the surrounding medium it dies out until the arrival of new heated water masses from the Coral Sea.

OBITUARY OF IRAKLIY IL'ICH KHERKHEULIDZE (1908-1978)

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 4, Apr 79 p 128

[Article by Z. I. Tskvitinidze and G. N. Khmaladze]

[Text] Candidate of Technical Sciences Irakliy Il'ich Kherkheulidze, head of the section on mudflows and channel processes at the Transcaucasian Scientific Research Institute of Hydrometeorology, died on 29 August 1978 in the 71st year of his life.

I. I. Kherkheulidze was born in 1908 at Kars. After graduating in 1932 from the civil engineering faculty of the Transcaucasian Institute of Transportation Engineers, until 1937 he worked in the system of the People's Commissariat of Railroads. Between 1937 and 1958 he worked at the Tbilisi Affiliate "Soyuzdorproyekt" in the post of senior engineer, as section head and as chief hydrologist, and in the years of the Great Fatherland War, from 1942 through 1944, in the frontal planning brigades of the Voyenmosttranproyekt on the Crimean front. Between 1958 and 1963 he occupied the post of head of the section on mountain melioration of the Georgian Scientific Research Institute of Hydrology and Melioration, and from 1963 to the end of his life he worked at the Transcaucasian Scientific Research Hydrometeorological Institute.

The results obtained by I. I. Kherkheulidze, extremely timely for science and important for practical application, were published in numerous Soviet and foreign publications. He was the author of 70 published studies, including four monographs.

While working in planning organizations, I. I. Kherkheulidze drew up a plan for a broad city bridge for Rustavi, and also successfully solved the problem of regulation of industrial water intake and protection of the territory against the slag and ash waste heaps of the Rustavi Metallurgical Plant.

During 1959-1962, under the direct leadership of I. I. Kherkheulidze, a 10-volume scientific research study was written on the problem of constructing a storm sewer system in Tbilisi.

The investigations and recommendations of I. I. Kherkheulidze have been used in instructions on planning and in implemented projects for bridge crossings, by the planning institute "Soyuzgiprogaz" in the planning of gas lines across mountain rivers on the routes Karadag-Tbilisi, Ordzhonikidze-Tbilisi, Yevlakh-Nakhichevan', Trans-Iranian, etc.

An excellent generalization of the enormous experience in planning was the creation of mudflow-protection structures and the scientific work "Through Protective and Regulating Structures of Reinforced Concrete on Mountain Rivers" (SKVOZNYYE ZASHCHITNYYE I REGULIRUYUSHCHIYE SOORUZHENIYA IZ SBORNOGO ZHELEZOBETONA NA GORNYKH REKAKH) (1967), for which the first prize of the Council of Ministers Georgian SSR was awarded in 1974.

I. I. Kherkheulidze devoted the last years of his life to a profound study of the processes of formation and movement of mudflows. He drew upon attainments in different sciences (hydrological engineering, fluvial hydraulics, soil science, dynamics of channel flows, hydraulic engineering, etc.), with the carrying out of laboratory experimental investigations, which enabled Irakliy Il'ich to develop effective constructions of through protective and regulating structures made of reinforced concrete.

The construction of these structures on the basis of the plans of the author himself was carried out on the mountain rivers of Durudzhi for protection of Kvareli and in Yugoslavia and was begun in Armia and Kazakhstan. He also developed recommendations for projects for antimudflow protection for Telavi, the rayon center Lentekhi in the Georgian SSR, Tyrnyauz in Kabardino-Balkaria and elsewhere.

At the request of the USSR Gosstroy I. I. Kherkheulidze and specialists at the State Hydrological Institute for the first time drew up the "Instructions on Determining the Computed Characteristics of Rain-Induced Mudflows" (INSTRUKTSIYA PO OPREDELENIYU RASCHETNYKH KHARAKTERISTIK DOZHDEVYKH SELEY) (VSN-03-76).

In 1946 I. I. Kherkheulidze was awarded third prize and special diplomas by the All-Union VNITOS (All-Union Scientific Technical Society of Builders) for the best scientific research work and in 1954, the third prize of the all-union competition for the best results in introducing advanced technology.

The productive field and scientific activity of I. I. Kherkheulidze was recognized by government and departmental awards.

The personnel of the Transcaucasian Scientific Research Hydrometeorological Institute will long remember this leading specialist and scientist, Irakliy Il'ich Kherkheulidze, this lover of work who spared no energy and time, who was demanding on himself and others, but at the same time a sensitive man, and whose untimely end is a great loss for the Hydrometeorological Service.

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