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METEOROLOGY AND HYDROLOGY

No. 1, January 1980

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WATER VAPOR AND PRECIPITATION BUDGET IN SYNOPTIC FORMATIONS IN THE TEMPERATE LATITUDES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 5-11

[Article by Candidate of Physical and Mathematical Sciences T. P. Kapitanova and Professor N. Z. Pinus, Central Aerological Observatory, submitted for publication 4 July 1979]

Abstract: The article presents the results of investigation of the moisture reserves and water vapor budget under cyclonic and anticyclonic conditions observed in a polygon with an area of $4 \cdot 10^{11}$ m^2 with its center at Perm' (December 1973 and November-December 1977). The authors evaluate the role of horizontal and vertical transfer and phase transformations in the water vapor budget. It is shown that on the basis of computations of the water vapor budget it is possible to make satisfactory quantitative estimates of steady precipitation over an area with the passage of cyclonic formations.

[Text] In our study [2] we presented the results of investigation of the water vapor budget and energy of phase transformations in a deep cyclone penetrating into the European USSR from the northwest. It was shown that the maximum water vapor reserves during all stages of its evolution are contained in the layer 850-800 mb; the maximum in the vertical profile of moisture reserves is more clearly expressed the greater the total moisture content. An important role in the water vapor budget of a cyclone is played by horizontal transfer and vertical redistribution associated with the transport of water vapor from the boundary layer into the higher layers of the atmosphere. In the central part of the cyclone the leakage of water vapor into the higher layers is essentially compensated by an inflow of water vapor through the lateral boundaries. The inflow of water vapor is expended primarily in condensation with a maximum in the layer 800-600 mb. In the layer P_e-600 mb on the southern periphery of the cyclone there is a leakage of water vapor due to horizontal advection; vertical redistribution and condensation processes are more weakly expressed. In general, as indicated by computations, in the central part of a cyclone the local changes are less than 10% of the budget, advection and vertical redistribution is no more than 40% and about 50% is attributable to changes related to condensation processes. On the cyclone periphery the main contribution is determined by local changes and condensation processes.

In this paper we give the results of study of the water vapor budget in different parts of cyclones and anticyclones observed in a fixed region -- in a polygon (Kazan' - Ufa - Sverdlovsk - Ivdel' - Syktyvkar - Kirov - Kazan' with its center at Perm') during the periods from 12 through 24 December 1973, from 11 through 17 November and from 1 through 9 December 1977. The area of the polygon in which the Central Aerological Observatory over a period of years has been carrying out flight investigations of atmospheric energy and turbulence is approximately $4 \cdot 10^{11}$ m^2.

As in [2], use was made in the computations of an equation describing the water vapor budget in the atmosphere.

$$\int \frac{\partial q}{\partial t}\, dX = \int -\nabla \cdot q\vec{v}\, dX + \int \frac{\partial qw}{\partial p}\, dX + \int -(q_{\text{к}} - q_{\text{и}})\, dX + \int D_q\, dX, \qquad (1)$$

where

$$\int dX = \frac{1}{gS} \iiint dx\, dy\, dp.$$

[K = condensation; И = evaporation]

where q is specific humidity, $\vec{v}$ is the wind velocity vector, p is pressure, w is the vertical component of wind velocity in a p coordinate system, g is the acceleration of free falling, q_{con} is the rate of water vapor condensation, q_{evap} is the rate of evaporation, S is the area of the base of a column of the atmosphere,

$$\nabla = \frac{\partial}{\partial x} + \frac{\partial}{\partial y}.$$

The term at the left in expression (1), which we will denote by q_t, describes the rate of local change in the content of water vapor in a column of the atmosphere with an area of the base S and the height $p_2 - p_1$; the first term on the right (q_{hor}) is the rate of change in the content of water vapor as a result of advection through the lateral boundaries of a column of the atmosphere; the second on the right (q_{ver}) is the rate of change as a result of the vertical redistribution of water vapor; the third on the right (q_Δ) is the rate of change in water vapor content as a result of the total effect of "condensation minus evaporation" of water vapor (effective condensation), the fourth on the right is the total rate of turbulent and diffusion transfer of water vapor.

All the elements of equation (1) were computed for layers of equal air mass with a thickness Δp = 50 mb in a column of the atmosphere from the earth's surface (P_e) and to p = 50 mb. Averaging of the elements in (1) was carried out by integration in area, where as the elementary areas dS we used all six triangles making up the polygon. In admissible cases the integration in area by means of the Ostrogradskiy-Gauss theorem was replaced by integration along the polygon perimeter.

Vertical velocities were computed on the basis of plane divergence from the continuity equation. It was assumed in the computations that $w = 0$ at the surface of the earth and at the upper level $p = 50$ mb.

In order to decrease the influence of errors caused by both the computation method and by errors in the measurement of meteorological elements we introduced corrections using the formulas cited in [5]. It was assumed that the computation errors increase linearly with altitude and the correction was carried out in such a way that the integral of divergence in the entire investigated layer is equal to zero.

Radiosonde data were used in computing all the terms in expression (1) except the terms q_Δ and $\bar{D}_q$, the sum of which is denoted by Δ. The Δ value was obtained as the residual term in expression (1). If it is assumed that D_q is small and the errors in radiosonde measurements and computations of the budget elements on an electronic computer are also small, the Δ value represents the total effect of the processes of condensation (sublimation) of water vapor and evaporation of cloud particles and precipitation particles in the considered volume. The Δ value will be positive in the case of a predominance of the processes of evaporation or negative in the case of a predominance of condensation (sublimation) of water vapor processes.

The assumption of a minor role of turbulence and diffusion of water vapor in the budget and especially the assumption of a small role of errors in measurements of meteorological elements is, to be sure, extremely approximate.

In order to evaluate the value and nature of the dependence of the water vapor budget components on errors in initial data, we carried out a numerical experiment. In particular, superposed on the initial radiosonde data were the disturbances $X \pm \delta X$ (X is any meteorological element) of a different character and in different combinations (for a wind velocity V, wind direction d, temperature t, relative humidity v we took the mean values of the errors in measurements in accordance with [1]). Table 1 gives the relative error (%) of the Δ value; it, like the residual term in equation (1), contains errors of all the budget elements. As a standard we used the Δ value, which was computed without superposing the disturbing values.

As we see, the relative error is maximum when the disturbances are superposed on all the meteorological elements forming the water vapor budget. With altitude the errors increase, as do the errors in measurements.

In particular, we should note the influence of errors in measurements of wind direction in the layers 600-400 and 400-200 mb. These large errors can exert a substantial influence on the results of computations of the horizontal fluxes of water vapor, playing a major role in the water vapor budget. On the average, the relative error is approximately 10%. An analysis of the results of the numerical experiment indicated that the measurement errors have little influence on the results of investigations of the general patterns of changes of elements in the water vapor budget.

Table 1

Relative Error Δ (%)

Слой, 1 мб	«Возмущены» все элементы 2	$V \pm \delta V$	$d \pm \delta d$	$t \pm \delta t$	$v \pm \delta v$
400—200	22	2	39	12	12
600—400	28	2	16	2	5
800—600	8	4	6	0	2
3 $P_з$—800	6	11	2	2	2
$P_з$—200	16	5	16	4	5

KEY:
1. Layer, mb
2. All elements disturbed
3. P_e

Table 2

Synoptic Situations

Год Year	A	B	C	E	Д
1973	7	19	—	26	10
1977	8	8	4	20	14

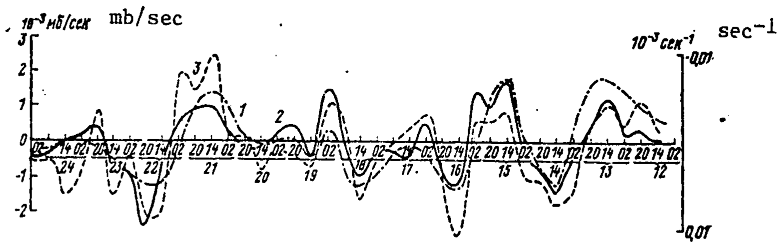

Fig. 1. Divergence in layer P_e - 600 mb (1) and vertical velocity in layers P_e - 600 mb (2) and 600-200 mb (3).

Now we will examine the synoptic situation in the polygon, which experienced substantial changes during the investigated periods. On the basis of an analysis of surface synoptic charts and pressure pattern charts all the observed atmospheric processes were assigned to four characteristic situations: A) frontal part of cyclone, B) southern part of cyclone, C) frontal wave, E) combination of situations A, B and C, D -- the periphery of an anticyclone. In order to refine the nature of the synoptic situation we used the results of computations of kinematic characteristics (divergence, vertical velocity, vorticity) on an electronic computer.

Table 3

Total Water Vapor Content (W) in Synoptic Formations in Temperate Latitudes (kg/m2) and Standard Deviation σ_q (kg/m^2)

Год (1)	Слой, мб (2)	A	B	C	E	D	W_E/W_D
1973, XII	(3) $P_э$—50 $P_э$—600	8,6 6,8	7,4 6,0	— —	8,0 6,4	5,5 4,4	1,55 1,50
1977, XI—XII	$P_э$—50 $P_э$—600	9,9 8,1	10,2 8,7	10,0 8,3	10,0 8,4	6,6 5,4	1,52 1,56
1973	σ_q	1,2	1,8	—	—	1,2	—
1977	σ_q	2,1	2,8	2,8	—	1,1	—

KEY:

1. Year
2. Layer, mb
3. P_e

In the analysis we excluded cases when it was hard to decide if the polygon was situated on the periphery of a cyclone or the periphery of an anticyclone. The number of computed cases for each situation was determined by the number of soundings during the particular period.

Table 2 gives data on the number of cases for each of the synoptic situations observed in the polygon.

Figure 1, as an example, shows the changes in divergence and vertical velocity observed in the atmospheric layer P_e - 600 mb over the polygon during the period 12-24 December.

In Fig. 1 we see that the computed kinematic characteristics have a wavelike character with a period of about 2.5 days. On days when the polygon was under the influence of cyclonic activity (12, 13, 15, 21, 23 December) there was a convergence of flow and ascending movements with a velocity of about $2 \cdot 10^{-3}$ mb/sec; on 16 and 22 December, when there was a well-expressed ridge in the neighborhood of the polygon there was divergence and descending movements with a velocity of about $2 \cdot 10^{-3}$ mb/sec.

Table 3 gives the total water vapor content in a unit air column for the layers P_e - 50 mb and P_e - 600 mb for different synoptic groups and also the standard deviations (σ_q) for the layer P_e - 600 mb. Although the cases for which the computations were made were few in number, the σ_q value does not exceed 30%. The cited mean water vapor reserves can be considered satisfactory.

Table 4

Water Vapor Budget in Synoptic Formations in Temperate Latitudes (10^{-2} g/(m^2·sec)) and Energy of Phase Transformations (W/m^2)

	Слой, мб (1)	1977 г.					1973 г.				
		$\bar{q}_t$	$\bar{q}_г$ (2)	$\bar{q}_в$ (3)	Δ	вт/м² (4)	$\bar{q}_t$	$\bar{q}_г$ (2)	$\bar{q}_в$ (3)	Δ	вт/м² (4)
A (5)	$P_з$—800	0,46	2,87	—1,30	—1.10	27,59	0,95	1,72	—0,96	0,20	—5,02
	800—600	0,09	1,85	—0,01	—1,75	43,89	0,65	1,98	—0,12	—1,23	30,85
	600—400	—0,02	0,98	1,20	—2,19	54,92	—0,22	0,68	0,97	—1,88	47,15
	$P_з$—400	0,53	5,70	—0,11	—5,05	126,40	1,38	4,38	—0,11	—2,91	72,98
B	$P_з$—800	—0,97	2,50	—1,40	—2,10	52,67	—0,01	0,47	—0,32	—0,17	4,26
	800—600	—0,85	0,86	0,96	—2,66	66,71	—1,20	—1,51	0,16	0,16	—4,01
	600—400	—0,12	—0.62	0,30	0,20	5,02	—0,32	—0,27	0,41	—0,45	11,28
	$P_з$—400	—1,94	2,74	—0,14	—4,56	114,36	—1,53	—1,31	0,25	—0,46	11,54
C	$P_з$—800	—0,16	2,70	—2,10	—0,76	19,09	—	—	—	—	—
	800—600	—0,21	0,30	1,50	—2,01	50,41	—	—	—	—	—
	600—400	—0,14	- 0,50	0,57	—0,21	5.27	—	—	—	—	—
	$P_з$—400	—0.51	2,50	—0,03	—2,98	74,77	—	—	—	—	—
D	$P_з$—800	—0,08	—0,68	0,22	0,36	—9,03	0,24	—0,56	1,10	—0,29	7,27
	800—600	—0,04	—0,15	0,24	—0,13	3,26	1,03	0,27	—0,14	0,91	—22,82
	600—400	—0.05	—0,15	—0,34	0,44	—11,04	—0,67	1,03	—0,78	—0.94	23,58
	$P_з$—400	—0,17	—0,98	0,12	0,67	—16,80	0,60	0,74	0,18	—0,32	8,02

KEY:

1. Layer, mb
2. q_{hor}
3. q_{ver}
4. W/m^2
5. P_e

Data in Table 3 show that the total water vapor content varies in the range from 6 to 10 kg/m^2 and almost 80% of this quantity is in the layer P_e - 600 mb. The water reserve in cyclonic situations (group E) on the average is 1.5 times greater than in anticyclonic situations (group D).

The data on the full moisture content give some idea concerning the supplies of latent heat; in cyclones it falls in the range (15-22)·10^6 J/m^2 in the atmospheric column P_e - 600 mb and in anticyclones -- (10-15)·10^6 J/m^2.

Now we will proceed to an examination of the water vapor budget. Table 4 gives quantitative estimates of components of the water vapor budget for all the considered synoptic situations. They show what factors are responsible for the replenishment of water vapor supplies, at what levels the greatest moisture exchange occurs and what the energy of phase transformations is. In the water vapor budget an important role is played by horizontal transfer through the lateral boundaries (q_{hor}). In the frontal part of the cyclone (group A) a horizontal influx as a result of convergence of the main flow is observed in the layer P_e - 400 mb; in the southern part of the cyclone and in the frontal wave region in the layer P_e - 600 mb the horizontal transfer decreases with altitude. The transport of water vapor into the higher layers takes place primarily from the layers P_e - 600 mb in group A and P_e - 800 mb in groups B and C.

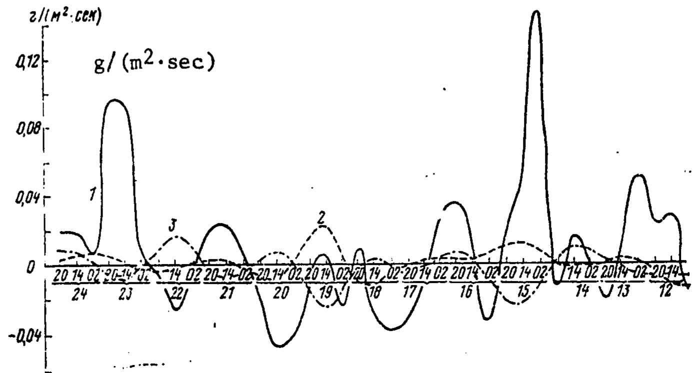

Fig. 2. Horizontal transfer of water vapor in the layer P_e - 600 mb (1) and vertical transfer in the layers P_e - 600 mb (2) and 600-200 mb (3).

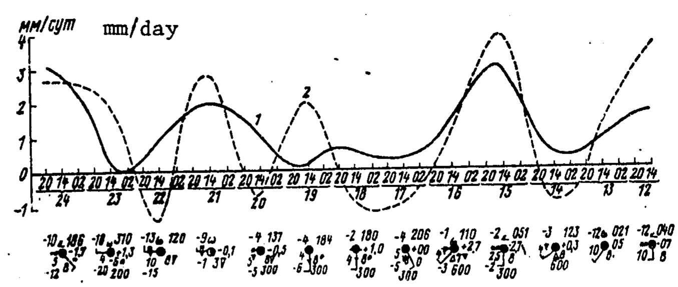

Fig. 3. Measured (1) and computed (2) quantities of precipitation during period from 12 through 24 December 1973.

The nature of the horizontal and vertical transfer of water vapor can be seen more clearly in Fig. 2, where as an illustration we have shown their change in this polygon area during the period from 12 through 24 December 1973. In particular, on 15 December 1973, with approach of a warm front to the polygon region, on the q_{hor} curve there is a pronounced maximum ($q_{hor} > 0.12$

$g/(m^2 \cdot sec)$). At the same time there is transport of water vapor into the higher layers ($q_v > 0.02$ $g/(m^2 \cdot sec)$). As a result of the processes in the layer 600-400 mb there was a local decrease in moisture reserves. In groups A, B, C there was predominance of processes of water vapor condensation over evaporation processes ($\Delta < 0$) and the energy of phase transformations in the layer P_e - 400 mb on the average attained 70-125 W/m^2. On the periphery of an anticyclone and in the region of high-pressure ridges (for example, 16 and 22 December) there was transport of water vapor through the lateral boundaries, especially in the atmospheric boundary layer, and the transport of water vapor from the higher layers (400-600 mb) into the lower layers (P_e - 800 mb) was a result of the descending movements occurring here.

The computations which we made on the basis of the data cited in Table 4 show that 80-85% of the Δ value, being to a definite degree a quantitative evaluation of effective condensation of water vapor, for the frontal part of cyclones (A) and the region of the frontal wave (C) was formed due to an influx of water vapor from an outer source through the lateral boundaries of the atmospheric layer and only 20-15% due to existing characteristic moisture supplies. Thus, in the processes of natural condensation of water vapor transpiring in the mentioned synoptic formations there was virtually no entrainment of water vapor supplies. However, on the southern periphery of the cyclone (B) only 35-40% of the Δ value was associated with the horizontal influx of water vapor. We note that according to [2], in the central part of a cyclone, in its development stage, 85-90% of the Δ value was formed due to lateral influxes of water vapor; in the central part of a cyclone, but in the filling stage, and on its southern periphery in the development stage, and in the filling stage 80-100% of the Δ value was associated with a decrease in moisture supplies.

A comparison of the results indicates a universality of the conclusions which we drew in [2] about the peculiarities of the water vapor budget in cyclonic formations in the temperate latitudes.

It is of particular interest to examine the matter of use of data on the water vapor budget for estimating the mean quantity of precipitation falling on the polygon. The residual term in expression (1) with small $\bar{D}_q$ values and measurement errors characterizes the effective quantity of condensing moisture (condensation minus evaporation). In this case it can be postulated that the value of the effectively condensing moisture must be close to the quantity of precipitation falling in the observation region.

Figure 3 shows the quantity of precipitation measured in the precipitation gage network of the polygon during the period from 12 through 24 December 1973 and the quantity of precipitation computed for this period on the basis of the residual term Δ. We see that the agreement between these curves is entirely satisfactory. The maximum quantity of precipitation was observed on days when the polygon was under the influence of cyclones (13, 14, 21 and 23 December 1973). The existing quantitative differences between the measured precipitation and the precipitation computed from the water vapor budget can be the result of both an inadequately precise determination of precipitation in the rain-gaging network and its low density, as well as (negative precipitation values) a result of nonallowance for "subgrid" processes in the budget equation, errors in calculations and errors in aerological data. The high correlation between the measured and computed quantities of precipitation are in good agreement with the results of similar comparisons published in [3, 4, 6]. The best agreement between the computed and measured precipitation was obtained in [3], which gives data for comparative purposes for a region with a dense network of precipitation gages (1 precipitation gage per 10 km^2).

An analysis of the results shows that an increase in precipitation is essentially associated with the horizontal influx of moisture in the lower half of the troposphere. We feel that data on the water vapor budget can be used for estimating the mean quantity of steady precipitation over an area for different practical applications.

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CHOICE OF DECISIONS WITH THE AVAILABILITY OF FORECASTS WITH DIFFERENT VALIDITY TIMES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 12-23

[Article by Candidate of Technical Sciences Ye. Ye. Zhukovskiy, Agrophysical Institute, submitted for publication 3 May 1979]

Abstract: The author examines the problems involved in the adoption of economically optimum decisions with the availability of meteorological forecasts which become more precise with time. The article includes a detailed study of the "climatological information - alternative forecasting" and "forecasting with a long validity time - forecasting with a short validity time" situations. It proposes a method for geometrical interpretation of the results using optimality criteria making possible quite simple determination for each user of the best economic strategy and formulation of sound requirements on prognostic information.

[Text] General principles. The principles of the economic approach to the problem of optimum use of meteorological forecasts were already laid in the 1930's by M. A. Omshanskiy [9] and were further developed in a whole series of later investigations [1, 7, 8, and others]. A result of these studies was solution of a broad range of problems directed to increasing the effectiveness of use of meteorological information in the control of the economy [5]. However, it must be noted that virtually all the problems considered to this time have had in a certain sense a static character and have pertained exclusively to those situations when the choice of economic alternatives was made by the user in one time interval. At the same time, it is possible to cite many examples when the process of adoption of a decision is more complex and is very closely related to the time factor. Indeed, the most common situation encountered is that which can be arbitrarily called an "either-either" situation. The essence of this situation is that in order to attain the desired effect the user in principle can make use of one of n economic decisions available to him, these differing from one another with respect to the moment of their realization. With an increase in the validity time of the measures taken there are usually two tendencies observed: the economy of the actions taken by the user increases, whereas the reliability of the meteorological information used in choosing decisions decreases. The opposite picture is observed with a decrease in validity time: the reliability of information about anticipated weather conditions increases, but the possibility of routine "adaptation" to these conditions, and accordingly, the results of the economic measures taken, is reduced. Thus, the need arises for finding some most rational "time" strategy for the adoption of decisions. In this communication we will examine this problem for the simplest case, when n = 2, that is, the user has the possibility of making a choice between two types of economic measures having different validity times.

In formulating the mentioned problem we will use as a point of departure that the user is dealing with some dangerous or simply unfavorable meteorological phenomenon which can occur or not occur; if it does occur, the economy is inflicted a loss equal to L units of cost. Adhering to Anderson [10], who evidently was the first to give attention to the necessity for a deeper study of the adoption of optimum decisions taking the time factor into account, we will assume that in order to prevent losses from unfavorable weather it is sufficient for the user to carry out one of two protective measures: d'_1, having a relatively great validity time T' and a relatively small cost C' or d''_1, having a lesser validity time T''(T''< T'), but greater than d'_1, cost C''. Thus, it is assumed that

$$C' < C''. \tag{1}$$

It is clear that any precautionary measures can be used only under the condition that expenditures in carrying them out do not exceed the losses caused by unfavorable weather. It therefore follows that in all situations which are of any practical interest the following expressions should be observed

$$\left.\begin{aligned} 0 \leq \frac{C'}{L} \leq 1 \\ 0 \leq \frac{C''}{L} \leq 1 \end{aligned}\right\}. \tag{2}$$

In addition to what has been said, we will assume that both types of considered protective measures d''_1 and d' are "ideal" in the sense that any of them completely prevent losses from the occurring harmful phenomenon.

Table 1

General Form of Matrix of Losses

Погода F (1)	Действия потребителя (2)		
	$d'_1\ d''_2$	$d'_2\ d''_1$	$d'_2\ d''_2$
F_1	C'	C''	L
F_2	C'	C''	0

KEY:
1. Weather F
2. Actions of user

F_1 and F_2 here are used to denote the states of weather corresponding to cases of realization of a harmful phenomenon (F_1) and its absence (F_2); $d_1'd_2''$, d_2', d_1'', $d_2'd_2''$ are variants of possible two-stage actions of the user (d_1', d_2'' -- in the first interval protective measures d_1'' are carried out, in the second nothing is done; $d_2'd_1''$ -- in the first interval nothing is done; in the second protective measures d_1'' are carried out; $d_2'd_2''$ -- protective measures are not taken in either the first or second intervals). It is easy to understand that the actions d_1d_1'', involving successive carrying out both types of protective measures, do not make sense by virtue of the assumption made concerning the idealness of the protective measures taken.

Analysis of strategies: analytical conclusions and geometrical interpretation. The nature of the economic decisions taken by the user will be essentially dependent on what meteorological information he has, or at least can get. From this point of view there are three cases which are of the greatest practical interest:

a) the availability to the user of only climatological information, stipulated in the form of the value of the natural frequency of occurrence of the considered harmful phenomenon $p_1 = P(F_1)$;

b) the availability to the user first (in the stage of adoption of beforehand decisions d') of climatological information alone, and then (with the adoption of short-range decisions d") a meteorological forecast with a short time validity;

c) availability of two successively issued forecasts with a long and short validity time.

It should be noted that the terms "long" and "short" validity times are employed here and in the text which follows exclusively for the purpose of emphasizing that the corresponding prognostic information has different validity time and is associated with the moments for carrying out different protective measures.

Case a). If the only meteorological information which the user has is information on the natural frequency of recurrence of a harmful phenomenon, interest is only in one of two climatological strategies $S_{cl\ 1}$, involving a constant (regardless of specific weather conditions) carrying out of protective measures with a long validity time d' and $S_{cl\ 2}$, consisting in the full neglecting of any protective measures whatsoever. The decision to "constantly carry out protective measures for a shore validity time" will not make sense because by virtue of inequality (1) such a strategy will inevitably be worse than $S_{cl\ 1}$.

Adhering henceforth to the usual Bayes approach, we will assume that the most advantageous strategy among any set of considered strategies is that which ensures minimizing of the mean losses U (in the statistical sense).

For two variants of the possible economic strategies $S_{cl\ 1}$ and $S_{cl\ 2}$ the U values will be equal (respectively) to

$$U_{cl\ 1} = p_1 C' + (1 - p_1) C' = C' \tag{3}$$

and

$$U_{cl\ 2} = p_1 L. \tag{4}$$

Converting from the absolute U values to the normalized indices E = U/L, these formulas can be rewritten in the form

$$E_{cl\ 1} = C'/L, \tag{5}$$

$$E_{cl\ 2} = p_1, \tag{6}$$

and similarly to U, relative to E, henceforth we will also use the term "mean losses" without the word "normalized," since this does not change the sense and nature of the reasonings.

Comparison of equations (5) and (6) makes it possible to conclude that with any values

$$\frac{C'}{L} < p_1 \tag{7}$$

the user must use the strategy $S_{cl\ 1}$, but with all

$$\frac{C'}{L} > p_1 \tag{8}$$

-- the stragey $S_{cl\ 2}$. In other words, if the conditions (7) are satisfied, then the climatologically optimum strategy is a constant carrying out of protective measures d_1', whereas with condition (8) one must be content with a possible loss from unfavorable weather.

For a graphic evaluation of the results it is then convenient to have a geometrical interpretation involving the construction of so-called optimum strategy diagrams.

As indicated in Fig. 1a, by virtue of conditions (1) and (2), in the coordinates (C"/L, C'/L) the needs of different users correspond to points lying within the "base" triangle formed by the x-axis C"/L, the vertical straight line C"/L = 1 and the bisector C'/L = C"/L. Solution of the problem of choice of the best economic strategy is reduced to breakdown of the area of this triangle into a number of nonintersecting regions, each of which corresponds to the optimality conditions of some one strategy, and subsequent finding of the region in which the point

$$N[(C''/L)_N,\ (C'/L)_N]$$

characterizing this user falls. For a case when we have only climatological information, such a breakdown is shown in Fig. 1b. All the points lying above the horizontal straight line $C'/L = p_1$ correspond to the climatologically optimum strategy $S_{cl\ 2}$ and all the points below it -- to the strategy $S_{cl\ 1}$.

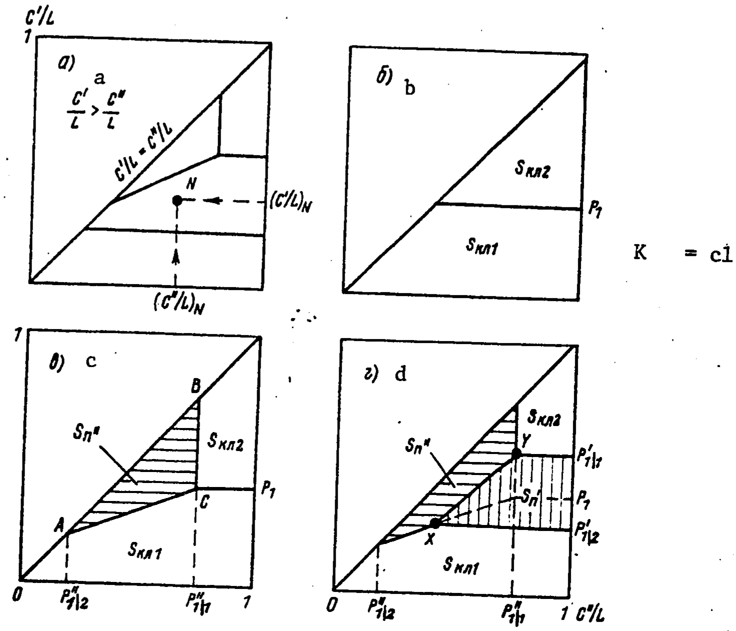

Fig. 1. Diagrams of optimum strategies. a) general case; b) presence of only climatological information; c) "climatology - alternative forecasting" situation; d) "forecast with a long validity time - forecast with a short validity time" situation.

Case b). Before turning directly to this case, we will make several general comments relating to the character of the used forecast. In particular, we will assume that this forecast, henceforth designated Π", from the point of view of the time of compilation and issuance to the user, can be employed in the process of adoption of decisions on carrying out protective measures with a short validity time. The second assumption made hereafter is that the forecast Π" has a categoric character, that is, its text contains one of two assertions: "the phenomenon is expected" or "the phenomenon is not expected."

As is well known, a thorough idea concerning the success of any alternative forecast Π is given by a square conjugation matrix whose elements are $p_{ij} = P(F_i \Pi_j)$ $(i, j = 1, 2)$ -- the joint probabilities of the predicted (Π_j) and actually occurring (F_i) weather conditions. The general form of such a matrix is given in Table 2, where p_1^* and p_2^* designate the probabilities of the texts of the forecast $\Pi = \Pi_1$ and $\Pi = \Pi_2$.

Table 2

Conjugation Matrix for Alternative Forecast

Погода F (1)	Прогноз П (2): Π_1	Π_2	Σ
F_1	p_{11}	p_{12}	p_1
F_2	p_{21}	p_{22}	p_2
Σ	p_1^*	p_2^*	1

KEY:

1. Weather F
2. Forecast

In addition to the joint probabilities p_{ij}, for the subsequent analysis it is useful to introduce into consideration the values $p_{i|j} = p_{ij}/p_j^*$ $(i, j = 1, 2)$ representing the conditional probabilities of occurrence of the weather phase F_i with the predicted state F_j (that is, under the condition $\Pi = \Pi_j$). It is obvious that for any methodologically sound prognostic schemes there must be satisfaction of the inequalities

$$\left.\begin{array}{l} p_{1|1} > p_1,\ p_{1|2} < p_1 \\ p_{2|1} < p_2,\ p_{2|2} > p_2 \end{array}\right\}, \tag{9}$$

and it is easy to demonstrate that the satisfaction of any of them immediately leads to satisfaction of all the others.

With the availability of climatological information and a successful alternative forecast Π'' with a short validity time, in addition to the strategies considered above $S_{cl\ 1}$ and $S_{cl\ 2}$, there is still another possibility -- proceeding to the strategy $S_{\Pi''}$, involving the carrying out of protective measures d_1'' in accordance with the anticipated (predicted) weather conditions. The strategy of adoption of decisions, despite a forecast, on the assumption of correctness of the conditions (9), can be immediately discarded since it will always yield to the better of the strategies $S_{cl\ 1}$ or $S_{cl\ 2}$.

Turning to Tables 1 and 2, we can write that the mean losses corresponding to the use of the strategy $S_{\Pi''}$ are computed using the formula

$$E_{\Pi''} = p_1^{*''} \frac{C''}{L} + p_{12}'', \tag{10}$$

where the double primes on p_1^* and p_{12} denote that the corresponding probabilities are related to the forecast Π''.

A comparison of the $E_{\Pi''}$, $E_{cl\ 1}$ and $E_{cl\ 2}$ values shows that the differentiation of economic decisions in forecasting with a short validity time is more advantageous than the climatological strategies $S_{cl\ 1}$ and $S_{cl\ 2}$ in those cases when there is satisfaction of the inequalities

$$\frac{C'}{L} > p_1^{*''}\frac{C''}{L} + p_{12}'' \qquad \frac{C''}{L} < p_{1|1}'' \tag{11}$$

Examining these relationships jointly with (7) and (8) it is simple to construct a corresponding optimality diagram and ascertain the regions of greatest effectiveness of each of the three strategies $S_{cl\ 1}$, $S_{cl\ 2}$ and $S_{\Pi''}$. In particular, as can be seen from Fig. 1c, the region where it is desirable to use the strategy $S_{\Pi''}$ is a triangle adjacent to the straight line $C'/L = C''/L$ and having vertices at the points

$$A\,(p_{1|2}'',\ p_{1|2}''),\quad B\,(p_{1|1}'',\ p_{1|1}''),\quad C\,(p_{1|1}'',\ p_1). \tag{12}$$

Here the first coordinate is C''/L and the second is C'/L.

The region in which there is dominance of the strategy $S_{cl\ 2}$ is situated in the right upper corner of the base triangle and corresponds to the points $(C''/L, C'/L)$, satisfying the conditions

$$\frac{C''}{L} > p_{1|1}'',\ \frac{C'}{L} > p_1. \tag{13}$$

In all the remaining cases the best solution will be the constant carrying out of protective measures with a long validity time, that is, the strategy $S_{cl\ 1}$.

It should be noted that in the general methodological plan the considered problem can be interpreted as the problem of the choice of an optimum economic strategy when a stochastic climatological forecast is available as well as a categorical forecast with a short validity time.

Case c). This differs from the preceding in that in addition to climatological information on the frequency of occurrence of a harmful phenomenon and an alternative forecast Π'' with a short validity time the user also has an alternative forecast with a long validity time Π' which from the point of view of the moment of compilation and issuance to the user can be employed in the stage of adoption of decisions on the carrying out of long-range protective measures d_1'.

With the availability of two forecasts in principle there can be so-called complex strategies whose realization assumes the simultaneous use of all the prognostic information [2-4]. As an example, it is easy to visualize the following variant of adoption of economic decisions.

In the case of poor weather ($\Pi' = \Pi_1$) anticipated from a forecast with a long validity time the protective measures d_1' are carried out and the forecast Π'' is not further taken into account. However, in the case of good weather ($\Pi' = \Pi_2$) anticipated under the first forecast the final decision is adopted in the second stage, to wit: if the forecast Π'', like Π', indicates favorable conditions ($\Pi'' = \Pi_2$), no measures are taken; however, if according to the second forecast a harmful phenomenon is anticipated ($\Pi'' = \Pi_1$), the protective measures d_1' are carried out. Thus, precautionary measures are taken when predicting a harmful phenomenon by at least one method. Another situation can also occur when protective measures, on the other hand, are carried out only in a case when the harmful phenomenon is anticipated only on the basis of both forecasts. It follows from an analysis of the different variants for the adoption of decisions that that in the particular case there is a total of ten conceivable complex strategies.

As was pointed out in [2-4], for finding the optimum algorithm for a complex approach it is not enough to know only the characteristics (conjugation matrices) of the individual alternative forecasts introduced above and it is necessary to have information on the probabilistic characteristics $P(F\Pi'\Pi'')$ which determine the joint frequency of recurrence of different combinations of texts of both forecasts Π', Π'' and the actually occurring weather conditions F. The seeking of these probabilities requires an analysis of considerable statistical material and frequently involves certain difficulties. Taking this into account, in the first stage we will examine only a simpler problem involving the choice of the best of four elementary (not complex) strategies $S_{cl\ 1}$, $S_{cl\ 2}$, $S_{\Pi''}$ and $S_{\Pi'}$. The first three of them are already known to us: they correspond to the mean losses determined by equations (5), (6) and (10). With respect to the strategy $S_{\Pi'}$, meaning adoption of decisions in accordance with an alternative forecast with a long validity time, by analogy with (10) the mean losses with its use will be

$$E_{\Pi'} = p_1^{*'} \frac{C'}{L} + p_{12}'. \qquad (14)$$

Here $p_1^{*'}$ and p_{12}', like the other probabilities with a prime presented hereafter, relate to the forecast Π' and have the very same sense as the similar values with a double prime for the forecast Π''.

It should be noted that the strategies of the operations, despite the forecasts Π' and Π'', which also are not complex, simply need not be considered, since, as already noted, with methodological soundness of the prognostic methods they always will be less effective than $S_{cl\ 1}$ and $S_{cl\ 2}$.

The seeking of regions of optimality of each of the four enumerated elementary strategies is reduced to joint analysis of expressions (5), (6), (10) and (14) and construction of a corresponding diagram of optimum strategies. The most natural case will be when a forecast with a short time validity Π'' has better success indices both from the point of view of the probability of a correct prediction of the occurring dangerous phenomenon ($p''_{1|1} > p'_{1|1}$) and with respect to the frequency of recurrence of different errors of the type "missing of a phenomenon" ($p''_{1|2} < p'_{1|2}$). The breakdown of the area of the base triangle into regions of optimality of individual strategies, corresponding to this condition, is shown in Fig. 1d. It is easy to see that for carrying out the corresponding constructions it is necessary and adequate to know five indices: conditional probabilities $p'_{1|1}$, $p'_{1|2}$, $p''_{1|1}$, $p''_{1|2}$ and the climatological frequency of recurrence of the phenomenon p_1.

Equating the right-hand sides of the expressions for $E_{\Pi''}$ and $E_{\Pi'}$ to one another, it can be shown that the boundary separating the region of dominance of the strategy $S_{\Pi'}$ from the region of greater effectiveness of the strategy $S_{\Pi''}$ will be the straight line

$$\frac{C'}{L} = p'_{1|1} + \frac{\overset{\bullet}{p}{}''_1}{\overset{\bullet}{p}{}'_1}\left(\frac{C''}{L} - p''_{1|1}\right), \tag{15}$$

passing through the points

$$\left.\begin{aligned} &X\left(\frac{p'_{1|2} - p''_{1|2}}{\overset{\bullet}{p}{}''_1},\ p'_{1|2}\right) \\ &Y\left(p''_{1|1},\ p'_{1|1}\right) \end{aligned}\right\}, \tag{16}$$

which in the particular case (that is, when the forecast Π'' is better than Π') lie below the straight line $C'/L = C''/L$. The regions of optimality of the strategies $S_{\Pi'}$ and $S_{\Pi''}$ in this case have the form of rectangles; the first of these is a right-angle trapezium. Easily confirmed also is the fact that when $p'_{1|1} \to p_1$ and $p'_{1|2} \to p_1$ — when the forecast with a long validity time Π' tends to a random forecast, the points X and Y approach one another and there is a gradual transition to the optimality diagram shown in Fig. 1.

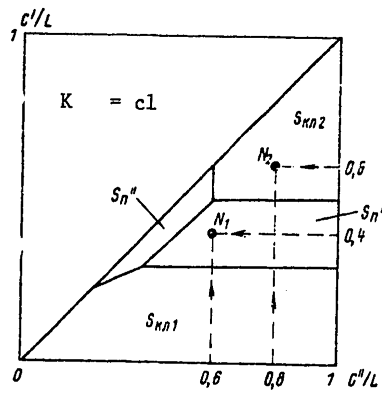

Fig. 2. Finding of optimum strategy in presence of two alternative forecasts (example).

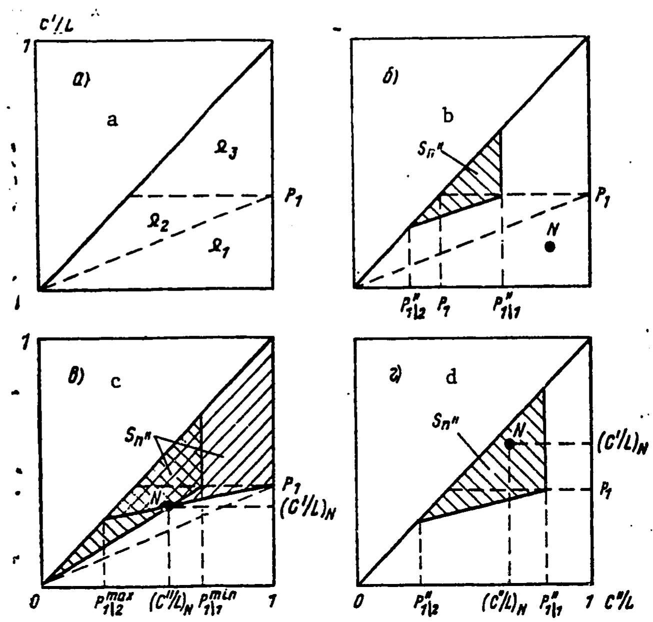

Fig. 3. Geometric interpretation of formulation of requirements on forecast on part of user. a) zones Ω_1, Ω_2 and Ω_3, differing in nature of requirements on the success of predictions: b) $N \in \Omega_1$; c) $N \in \Omega_2$; d) $N \in \Omega_3$.

Table 3

Percentage Frequency of Recurrence of Different Combinations of Predicted and Actually Occurring Weather Conditions for Two Forecasts of Different Validity Times

F	Π′		Π″		Σ
	Π_1	Π_2	Π_1	Π_2	
F_1	25	15	27	13	40
F_2	25	35	18	42	60
Σ	50	50	45	55	100

Discussion of results. The described geometric method can be used successfully in solving two mutually inverse problems. We have already dealt with the first in the process of the analysis of strategies carried out above. The problem is formulated as follows.

We will assume that we know the climatological frequency of recurrence of the harmful phenomenon and the probabilistic characteristics of forecasts of different time validity, determined by the corresponding conjugation matrices. It is necessary to determine what economic strategy must be adhered to by the users of this meteorological information with different values of the economic indices $(C''/L)_N$ and $C'/L)_N$ (N is the number of the user).

In order to answer the formulated problem it is sufficient to carry out a construction similar to that shown in Fig. 1d and then determine in which of the regions the point of interest to us

$$N\left[\left(\frac{C''}{L}\right)_N, \left(\frac{C'}{L}\right)_N\right].$$

falls.

We will illustrate this procedure in one specific example. Table 3 gives the percentage frequencies of recurrence of possible combinations of predicted and actually occurring weather conditions for two alternative forecasts Π' and Π'' differing with respect to validity time. On the basis of these data, for both prognostic methods we will find the characteristics of success $p_{1|1}$ and $p_{1|2}$. In this case we will have:

for Π'

$$p'_{1|1} = \frac{25}{50} = 0{,}50, \quad p'_{1|2} = \frac{15}{50} = 0{,}30;$$

for Π''

$$p''_{1|1} = \frac{27}{45} = 0{,}60, \quad p''_{1|2} = \frac{13}{55} = 0{,}24.$$

The climatological frequency of recurrence of the predicted phenomenon is $p_1 = 0.40$ (40%). Thus, both forecasts are methodologically sound and the forecast with the lesser validity time has the higher success. Figure 2 shows the diagram of optimality of different strategies corresponding to the considered case.

We will assume that there are two users, for the first of which $C''/L = 0.6$ and $C'/L = 0.4$, whereas for the second -- $C''/L = 0.8$ and $C'/L = 0.6$. The corresponding points in Fig. 2 are N_1 and N_2. Taking their positions into account, it can be concluded that for the first user the optimum strategy is that of carrying out of protective measures d_1 in accordance with the forecast Π', and for the second -- the dispensing with precautionary measures and being content with possible loss from unfavorable weather. In a similar way (without any computations whatsoever) it is possible to give recommendations on the adoption of economic decisions for any other users.

Now we will turn to the inverse problem. In this case the economic characteristics $(C''/L)_N$ and $(C'/L)_N$ are assumed to be stipulated, these determining the users, and the question is raised as to what requirements must be satisfied by forecasts, in particular, a forecast with a short time validity Π'' so that its use will give a positive effect in comparison with a climatologically optimum strategy or (if there is a methodologically sound forecast with a long validity time Π') in comparison with the best of the strategies $S_{cl\ 1}$, $S_{cl\ 2}$ and $S_{\Pi'}$. The solution of this problem, to be sure, could be obtained purely analytically, on the basis of a joint analysis of the expressions for the mean losses $E_{cl\ 1}$, $E_{cl\ 2}$, $E_{\Pi'}$ and $E_{\Pi''}$. However, these same results are obtained in a more graphic form if we have recourse to a corresponding geometrical interpretation. We will illustrate this in the example of formulation of the requirements on prognostic information in a "climatology - forecast with a short time validity" situation.

As demonstrated in Fig. 3a, there are three zones Ω_1, Ω_2 and Ω_3 which differ in the nature of the requirements imposed on prognostic information.

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For any points lying below the straight line

$$\frac{C'}{L} = p_1 \frac{C''}{L} \quad (\text{region } \Omega_1), \tag{17}$$

the very formulation of the problem of seeking the conditions for a change from the climatologically optimum strategy here $S_{c1\ 1}$ to the use of the forecast Π'' does not make sense, since with any probabilities $p''_{1|1} \in [p_1, 1]$ and $p''_{1|2}$ and $p''_{1|2} \in [0, p_1]$ the region of optimality of the strategy $S_{\Pi''}$ (shaded in Fig. 3b) is situated above the considered region Ω_1.

When the point $N\left[\left(\frac{C''}{L}\right)_N, \left(\frac{C'}{L}\right)_N\right]$

falls over the straight line (17) there can be two cases:

$$p_1 \left(\frac{C''}{L}\right)_N < \left(\frac{C'}{L}\right)_N < p_1 \quad (\text{region } \Omega_2) \tag{18}$$

and

$$\left(\frac{C'}{L}\right)_N > p_1 \quad (\text{region } \Omega_3) \tag{19}$$

The first of these is more complex. As can be seen from the constructions shown in Fig. 3c, in the region Ω_2 the strategy $S_{\Pi''}$ can be optimum only with $p''_{1|1}$ values which without question exceed the critically small level

$$p_{1|1}^{\min} = \frac{p_1\left(\frac{C''}{L}\right)_N}{\left(\frac{C'}{L}\right)_N}, \tag{20}$$

and $p''_{1|2}$ values unquestionably less than the critically high level

$$p_{1|2}^{\max} = \frac{\left(\frac{C'}{L}\right)_N - p_1\left(\frac{C''}{L}\right)_N}{\left(\frac{C'}{L}\right)_N - \left(\frac{C''}{L}\right)_N - p_1 + 1}. \tag{21}$$

It is also easy to demonstrate that with stipulation of a specific $p''_{1|2}$ value belonging to the interval $[0, p^{\max}_{1|2}]$ the use of the Π'' forecast will be feasible only with probabilities $p''_{1|1}$ exceeding the value

$$p_{1|1}^{\min} = p''_{1|2} + (p_1 - p''_{1|2}) \frac{\left(\frac{C''}{L}\right)_N - p''_{1|2}}{\left(\frac{C'}{L}\right)_N - p''_{1|2}}, \tag{22}$$

and, on the other hand, if a $p''_{1|1}$ value lying within the interval $[p^{\min}_{1|1}\ 1]$ is registered, the probability $p''_{1|2}$ must be less than

$$p_{1|2}^{max} = \frac{p''_{1|1}\left(\frac{C'}{L}\right)_N - p_1\left(\frac{C''}{L}\right)_N}{\left(\frac{C'}{L}\right)_N - \left(\frac{C''}{L}\right)_N - p_1 + p''_{1|1}}. \tag{23}$$

The limiting expressions (20) and (21) in fact represent special cases of the general expressions (22) and (23). In actuality, with $p''_{1|2} = 0$ (20) follows from (22), whereas with $p''_{1|1} = 1$ (21) follows from (23). Thus, in the region Ω_2, stipulated by the bilateral inequality (18), a change from an indicated $S_{cl\ 1}$ strategy which is climatologically optimum here to the strategy $S_{\Pi''}$ requires the satisfaction of definite limitations imposed on the admissible values of the probabilities $p''_{1|1}$ and $p''_{1|2}$.

With respect to the region Ω_3, situated above the horizontal straight line $C'/L = p_1$ (Fig. 3d), here a change to the strategy $S_{\Pi''}$ will be desirable with any values

$$p''_{1|1} > \left(\frac{C''}{L}\right)_N, \tag{24}$$

that is, in contrast to the preceding case no special requirements are imposed on the probability $p''_{1|2}$. This, to be sure, does not mean that the $p''_{1|2}$ value in the Ω_3 region has no importance at all. It is not important only from the point of view of fundamental solution of the problem of what is more advantageous: be guided by the forecast or not take any protective measures. [The mentioned peculiarity is a reflection of the general principle, discussed in [6], of the one-sidedness of the requirements imposed on an alternative forecast by some specific user.] However, with respect to the degree of usefulness of prognostic information, all other conditions being equal, it will always be the higher the lesser the probability $p''_{1|2}$.

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EVALUATION OF THE STAGE OF DEVELOPMENT OF A CUMULUS CLOUD

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[Article by G. N. Nikitina and G. I. Skhirtladze, Institute of Experimental Meteorology, submitted for publication 20 April 1979]

Abstract: This paper presents the results of measurement of the optical density of cumulus clouds. The authors demonstrate the differences in the distributions of the size of optical inhomogeneities in developing and decaying clouds (on the basis of visual observations). A method for probabilistic evaluation of the stage of evolution of a cumulus cloud is proposed which is based on the horizontal extent of optically homogeneous sectors.

[Text] The physical processes transpiring in cumulus clouds are essentially dependent on the stage of cloud development. It is possible to differentiate three stages in evolution of a cumulus cloud: development, stationary state and decay [9]. In making observations of clouds from the ground these stages can be determined using photogrammetric measurements of the rate of development of the tops and vertical and horizontal dimensions of the cloud [1, 10]. It is difficult to make such measurements aboard an aircraft laboratory. In this case there are limitations on visual observations of clouds, making it possible for the most part to separate decaying clouds from nondecaying clouds, that is, those in the stage of development or in a stationary state [11]. However, visual observations cannot be made in flight when there is a high degree of cloud coverage when the aircraft laboratory rapidly passes from one cloud into another. Accordingly, for aircraft investigations of cumulus clouds it is necessary to have a method making it possible to evaluate the stage in development of a cloud when processing experimental data. In this paper an attempt is made to evaluate the stage in development of a cumulus cloud on the basis of the nature of variability of the optical density of the cloud medium in a horizontal direction.

Experimental investigations [4-6] have shown that clouds of different species can be characterized by a definite mean horizontal scale of non-uniformity of the attenuation index -- a parameter characterizing the optical density of the medium. For cumulus clouds this scale is 150 m [6].

In the studies enumerated above there was no mention of the stage in development of the investigated cumulus cloud; averaging was for all clouds. Using an IL-14 aircraft laboratory we carried out measurements of the optical density of cumulus clouds, accompanied by visual evaluations of the aerologist aboard who determined the stage of cloud development directly before the aircraft passed through the cloud. The flights took place during 1976-1978 (in summer) over the territory of Moldavia and the Ukraine. The measurements were made in droplet clouds with a vertical thickness of 0.8-2.5 km. The cloud was considered to be in the development stage (or in a stationary state) under the following conditions: well-expressed turbulently developing top, clear outlines, and insofar as possible in observations in the cloud cover field, an even base [9, 11]. A cloud with a "fibrous" top and indistinct outlines was considered to be in a stage of decay. A cloud with a decaying uneven base, but retaining evidences of growing upper and middle parts, was also regarded as being in the development stage [1]. The flights through clouds were made for the most part in the middle or upper parts. In 1978 the measurements were made in growing and decaying clouds on each flight; in 1977 -- only in growing clouds. In 1976 repeated flights were made through clouds which before the fly-through were evaluated as growing clouds, but which in subsequent fly-throughs gradually decayed and measurements were made separately only in decaying clouds. The number of clouds investigated on each flight and the characteristics of the synoptic situations on measurement days are given in Table 1.

During flight through the cloud measurements were continuously made to determined optical density. A "Boomerang" instrument was used [7]; it had a time constant $\tau \approx 0.01$ sec. Temperature was measured using a low-inertia electric thermometer with $\tau \approx 0.03$ sec. Flight altitude and velocity were measured using EDPD-1 modified sensors. Measurement data were registered with a loop oscillograph with a rate of movement of the phototape of 8 mm/sec. Figure 1 shows examples of oscillograms with registry of instrument readings during flight through cumulus clouds.

The primary processing of experimental data involved the measurement of the lengths of segments in clouds which are uniform in optical density. As in [6], those sectors of a cloud were considered uniform in which the attenuation index γ m^{-1} (or the value of the range of visibility S m, inversely proportional to it [6]) did not change by a factor of more than 2. Cloud inhomogeneities were grouped by sizes. The range of dimensions of inhomogeneities in the group was 50 m. The data in each measurement series (each flight) were used in plotting the distributions of inhomogeneities by sizes in developing and decaying clouds. As a rule, the scales of the inhomogeneities in the decaying clouds were less than in the developing clouds. A comparison of the distributions was made using the Kolmogorov test $K(\lambda)$ [8]. Table 2 gives the computed values $1 - K(\lambda)$ for a comparison of the distributions of inhomogeneities from four arbitrarily selected series of measurements with one another. The high $1 - K(\lambda)$ values (insignificance of differences) for clouds in one stage of evolution and the low values for clouds in a different stage indicate that the optical structure of the cumulus clouds is slightly dependent on the conditions under which the cloud developed and exists in comparison with the dependence on development stage. Accordingly, it is possible to combine data from series of measurements relating to different synoptic conditions for the purpose of increasing the statistical back-up for the distributions.

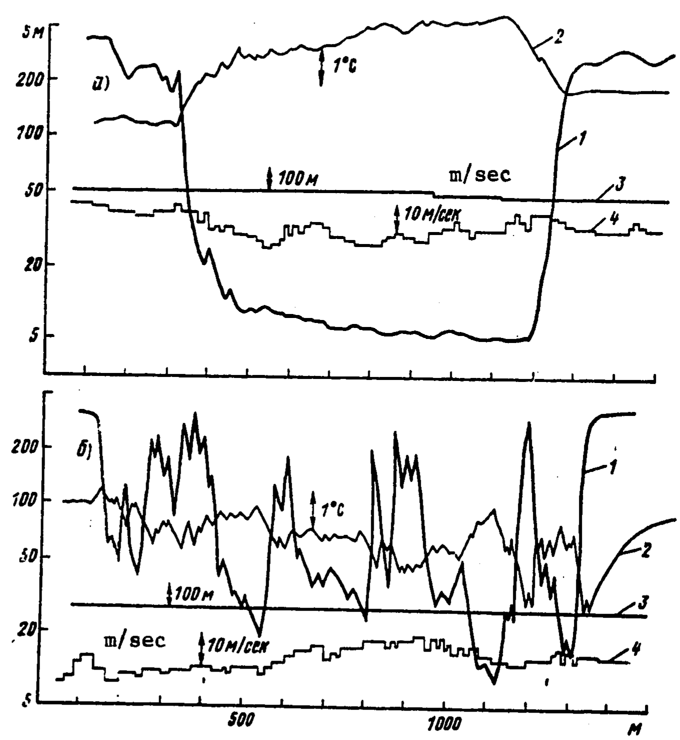

Fig. 1. Variation of optical density (1) and temperature (2) along the horizontal in developing (a) and decaying (b) cumulus clouds; 3 and 4 -- altitude and flight speed of aircraft respectively.

Table 1

Year	Flight No	Synoptic situation	Time, hours	Number of investigated clouds: growing	Number of investigated clouds: decaying
1976	1	Eastern periphery of anticyclone, ridge	10-14	25	--
	2	Western periphery of poorly expressed anticyclone	12-14	13	--
	3	Western periphery of anticyclone	9.5-12	10	--
	4	Cold sector of cyclone, trough	12-14	15	--
	5	Western periphery of anticyclone, ridge	14-16	--	12
	6	Western periphery of anticyclone, ridge	15-17	--	15
1977	7	Secondary cold front, trough	15-18	27	--
	8	Cold sector of cyclone	16-18	33	--
1978	9	Warm sector of cyclone	12-14	11	5
	10	Low-gradient pressure field of high pressure	14-16	8	14
	11	Eastern periphery of anticyclone	15-17	10	22
	12	Northern part of cyclone, low-gradient pressure field	16-18	7	13

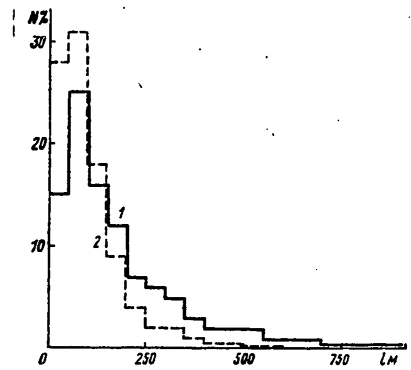

Fig. 2. Size distribution of optical inhomogeneities in growing (1) and decaying (2) clouds.

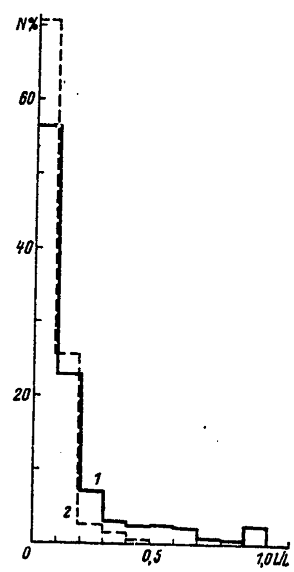

Fig. 3. Distributions of dimensions l of optical inhomogeneities in growing (1) and decaying (2) clouds relative to extent of cloud L.

Figure 2 shows the size distributions of optical inhomogeneities in clouds in the growth stage (1061 inhomogeneities) and the stages in decay (683 inhomogeneities) constructed on the basis of results of all measurements. The most probable size of inhomogeneities in both cases is about 100 m. However, there are significant differences between the two distributions. The distribution relating to decaying clouds (2) is more symmetric; it contains virtually no uniform sectors of more than 500 m in length and sectors with a length > 250 m constitute only 10% of the total number. Distribution (1) is sharply asymmetric; inhomogeneities in it greater than 250 m constitute 35% of the total number; individual inhomogeneities attain a size 1000-1500 m. The mean size of the optical inhomogeneities in growing clouds is 194 m, and in decaying clouds -- 102 m. The mean horizontal sizes of clouds in different stages were approximately identical -- about 1200 m. Estimates of the significance of differences in distributions (1) and (2), carried out using the $K(\lambda)$ test, show that these differences are significant with a significance level less than 5%.

Table 2

Values $1 - K(\lambda)$ in Intercomparison of Distributions of Optical Inhomogeneities from Different Measurement Series (Flights)

1 № полета (табл. 1)	5*	7	11	11*
3	0,379	0,999	0,965	0,064
5*	—	0,065	0,117	0,954
7	—	—	0,711	0,003
11	—	—	—	0,025

* series relating to decaying clouds.

KEY:

1. No of flight (Table 1)

In 90% of the cases the optically densest parts of the growing clouds were the longest homogeneous sectors. No such dependence was observed in decaying clouds. Figure 2 shows that the greater (lesser) are the dimensions of the inhomogeneities, the more probable it is that the cloud can be assigned to the category of growing (decaying), in the absence of visual observations. Table 3 gives the numerical values of the probability of correspondence of evaluations of the stage of development of clouds by sizes of optical inhomogeneities to visual observations, computed from the distributions of sizes of the most extensive cloud sectors.

Similar results can be obtained without making visual observations of a cloud, but by comparing the optical structure of clouds with repeated passes of the aircraft. It is known that the aircraft, when flying through, exerts a destructive effect on a cloud with small vertical development; for the IL-14 aircraft the number of flights through the cloud adequate for cloud decay is 1-5 [2]. It can be noted on the basis of measurement data for 1976, when repeated flights were made through clouds, that with an increase in the number of passes through the cloud the dimensions of the optical inhomogeneities in general decrease. In the first flights through clouds the mean dimensions of the inhomogeneities were 212 m, in the second -- 187 m, and in successive passes -- 135 m.

Table 3

Probabilities of Correspondence (%) of Evaluations of Stage of Development of Clouds by Sizes of Optical Inhomogeneities to Visual Observations

Стадия развития 1	2 Максимальные размеры неоднородностей, м									
	> 100	100—150	150—200	200—250	250—300	300—350	350—400	400—450	450—500	> 500
3 Растущие	0	13	20	25	35	50	60	75	90	100
4 Распадающиеся	100	87	80	75	65	50	40	25	10	0

KEY:
1. Development stage
2. Maximum size of inhomogeneities, m
3. Growing
4. Decaying

In an examination of the synchronous records of temperature and optical density variability in clouds we note a correlation between them (Fig. 1). According to [3], in a cumulus cloud the zones with a higher temperature correspond to ascending convective flows. The size distribution of zones of increased temperature (convective flows) in growing and decaying clouds was compared. These distributions differed from one another similar to the distributions of optical inhomogeneities (Fig. 2). Accordingly, the difference in the optical structure of clouds in different stages of evolution is attributable to the difference in the structure of cloud convection.

A comparison of the distributions of the relative (to the horizontal extent of the cloud) dimensions of optical inhomogeneities, represented in Fig. 3, is of definite interest. This figure shows that in decaying clouds the relative dimensions of the inhomogeneities are substantially less than in growing clouds. In the latter there are many cases when the cloud is almost as uniform horizontally as, for example, in Fig. 1a. A noteworthy point is the presence of a minimum in the distribution relating to growing clouds (Fig. 3 (2)). It is possible that this minimum is the limit separating growing and stationary clouds. This is supported by the fact that individual measurements of the structure of convective flows [12] show that in small cumulus clouds with a vertical thickness of about 2 km in the growth stage there is a predominance of one ascending flow whose dimensions are comparable with the extent of the cloud. Accordingly, convective flows determine the optical structure of a cloud. The position of the minimum in the distribution (1) is evidently governed by the nature of the separation of the main flow with transition of a cloud into a stationary stage. The quantity of clouds for which the relative dimensions of homogeneous sectors are close to unity is 10-15% of the total number of clouds evaluated as "growing." These evaluations agree with measurements of the relationship between the quantity of growing and stationary clouds in the field of cumulus clouds [3]. Nevertheless, despite indirect confirmations of the possibility of separating growing clouds from stationary clouds on the basis of the ℓ/L value, where ℓ is the dimension of a homogeneous sector, L is the extent of the cloud, additional investigations are required for its rigorous demonstration.

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UDC 551.510.534(47+57)

FEATURES OF OZONE DISTRIBUTION WITH ENTRY OF SOUTHERLY CYCLONES INTO THE EUROPEAN USSR

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[Article by Candidate of Geographical Sciences L. A. Uranova, USSR Hydrometeorological Scientific Research Center, submitted for publication 23 March 1979]

Abstract: The author examines the relationship between the forming of southerly cyclones and their entry into the European USSR and regions of a minimum total ozone content. It was found that on the average two days prior to the formation of a cyclone over the Mediterranean or Black Seas a region of minimum ozone is formed in this region. The trajectory of cyclone entry passes along a zone of reduced total ozone content, being deflected from the trajectory of motion of the ozone minimum by an average of 300-400 km. The entry of the cyclone into the European USSR usually occurs the day after the arrival of the ozone minimum.

[Text] As is well known, a prediction of entry of southerly cyclones into the European USSR is one of the most difficult problems and as yet does not have a high degree of success. Until now all studies of this problem have been based exclusively on synoptic interrelationships [1, 3, 5, 6].

In this paper we examine the possibility of an influence of the distribution of the total ozone content and its changes on the entry of southerly cyclones into the European USSR. During the past decade there have been quite a few studies devoted to the interrelationships between the distribution of the total ozone content and atmospheric circulation [2, 4, 7-10]. Many of them have related to the influence of atmospheric circulation on ozone. During recent years several studies have been devoted to the influence of the distribution of total ozone content on circulation [4, 7, 8]. In particular, in one of the recent studies [8] it was demonstrated that there is a rather good correlation between the day-to-day ozone variability

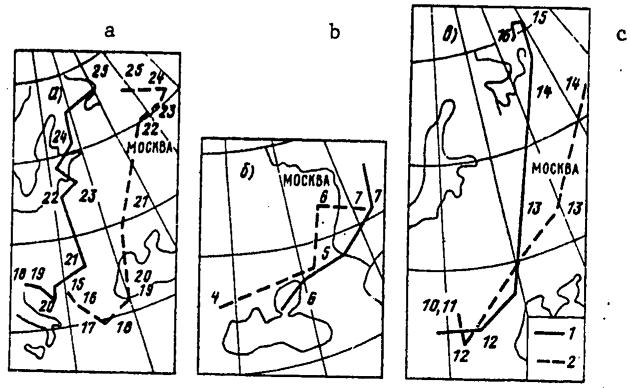

Fig. 1. Trajectories of motion of southerly cyclones (1) and accompanying regions of ozone minima (2). a) 18-25 August 1972; b) 6-7 October 1971; c) 12-16 September 1971.

Table 1

Distribution of Entries of Southerly Cyclones into European USSR by Months During 1971-1975

Год 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	Год 1
1971	1	1	0	1	0	0	3	0	4	1	3	0	14
1972	0	0	1	3	0	3	3	2	3	4	1	0	20
1973	0	5	0	4	4	0	1	1	0	1	2	3	21
1974	0	1	1	3	1	0	1	0	0	5	0	2	14
1975	0	0	3	4	0	2	0	0	0	1	0	2	12
15 Всего	1	7	5	15	5	5	8	3	7	12	6	7	81

KEY:

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

and also day-to-day variability of pressure and temperature at sea level. It was found on the basis of a rather long series of years that on the average after three days the day-to-day variability of pressure at sea level has the same sign as the day-to-day variability of ozone. In addition, a correlation was found between marked fluctuations of the total content of ozone from day to day and the appearance of tropical cyclones.

On the basis of the mentioned studies, in this article we have examined cases of the development of cyclones in the Mediterranean and Black Seas which later entered the European USSR and the field of the total ozone content during this same period. We analyzed a total of 81 entries of southerly cyclones into the European USSR during 1971-1975.

Also used in the analysis were daily maps of the total content of ozone, weather maps of the northern hemisphere for 0300 and 1500 hours Moscow time and pressure pattern charts of the isobaric surfaces 700 and 500 mb. We plotted the trajectories of motion of cyclones and the ozone minima regions associated with them, as well as graphs of the temporal distribution of the day-to-day variability of the total content of ozone and pressure at sea level from the moment of development to filling in the territory of the European USSR.

As demonstrated by an analysis of 544 cyclones arising over the Mediterranean and Black Seas during 1971-1975, each of the developing cyclones corresponded to a definite ozone minimum (Fig. 1). However, by no means all cyclones existed more than one day and only 15% of all the cyclones entered the European USSR.

The distribution of entries of southerly cyclones into the European USSR by months (Table 1) indicated that the greatest number of entries is observed in spring and autumn; in winter and summer there are considerably fewer. However, in 1972 the maximum of entry of cyclones into the European USSR occurred in summer and autumn, and in 1973 -- in winter and spring.

As indicated earlier [1], a necessary condition for the entry of southerly cyclones into the European USSR is a meridional transformation of the high-level pressure field in the lower troposphere. During the considered period the development of cyclones over the Mediterranean and Black Seas occurred at the sea surface simultaneously or somewhat earlier, with meridional transformation of the pressure field at the isobaric surface 500 mb. Usually, after the formation of cyclones, along the western peripheries of high-level troughs there is an inflow of cold air masses, penetrating to the Mediterranean Sea, which favors the further development of the cyclones developing here and their entry into the European USSR along the eastern periphery of this trough.

Here we will give a detailed analysis of one of the 81 cases of entry of southerly cyclones into the European USSR, also carefully analyzed.

For example, on 18 August 1972, a cyclone developed in the northern part of the Apennines Peninsula. Later it entered the European USSR and on 25 August reached the Kola Peninsula. The pressure at its center at the time of development was 1008 mb and during the entire time of entry changed insignificantly. The chart for the isobaric surface 500 mb for 0300 hours on 18 August indicated that over Eurasia there was a zonal flow and only on 19 August was there a restructuring of the high-altitude pressure field, as a result of which a high-altitude trough was formed over Europe. This trough reached the southern shores of the Mediterranean Sea. The inflow of cold air from the northern part of the Atlantic along the western periphery of this trough led to a deepening and activation of the cyclone developing the day before and favored its entry into the European USSR along the eastern periphery of the trough. Thus, the development of this cyclone cannot be attributed to meridional transformation of the high-altitude pressure field.

Table 2

Development of Southerly Cyclones and Minima of Total Ozone Content During Period 1971-1975

	Одновременно 1	2 Минимум озона раньше циклона на				
		1 день 3	2 дня	3 дня	4 дня	5 дней
4 Число случаев	0	19	184	277	48	16
%	0	4	33	51	9	3

KEY:
1. Simultaneously
2. Ozone minimum earlier than cyclone by...
3. 1 day, 2 days...
4. Number of cases

An analysis of the field of total ozone content for the second half of August 1972 indicated that from 14 to 15 August the ozone field was transformed in such a way that a zone with a minimum quantity of ozone passed through the whole of Europe from southwest to northeast (an ozone minimum of 0.240 cm was observed over the Baltic Sea). The region of the ozone maximum, up to 15 August occupying the temperate latitudes, was divided into two centers. One of them was situated over America (0.360 cm) with a "spur" to Western Europe; the other was over the southern part of Eastern Siberia (0.440 cm). This led to the formation of an ozone minimum region on 16 August in the ozone trough over the Balkan Peninsula; this ozone minimum later began to shift into the European USSR. The trajectory of motion of this ozone minimum passed from the southwest to the northeast to the shores of the Arctic Ocean. A marked decrease in the quantity of ozone occurred along the future trajectory of motion of the cyclone during the period 14-15 August. For example, at Kiev, Riga, Leningrad and Arkhangel'sk the ozone quantity in 24 hours decreased on the average by 0.060 cm. The region of the ozone minimum was formed three days before the development of the surface cyclone (Fig. 1). The trajectory of motion of the cyclone almost duplicates the trajectory of motion of the ozone minimum region, deviating 500-600 km to the left of it.

As already mentioned before, a high percentage of the entries of southerly cyclones occurred against a background of a reduced total content of ozone since the ozone minimum in the annual variation is observed in September-October. In addition, to the south of 40°N the field of a reduced quantity of ozone predominates the entire year. If in this case a zone of reduced ozone content, oriented from north to south or from northeast to southwest is observed over Europe, it can be postulated that a developing ozone minimum will be displaced into the European USSR and on the average after two days a cyclone is formed in the surface pressure field. This cyclone will also enter the European USSR. The entry of a cyclone usually occurs a day after entry of the ozone minimum.

The formation of a cyclone two or three days later in the region where the ozone minimum developed before this can be attributed to the fact that the decrease in the quantity of ozone favors an increase in UV penetrating into the lower layers of the atmosphere, which, in turn, leads to additional heating of the air in this region, the appearance of a heat ridge, and accordingly, the great temperature contrasts necessary for the genesis of a cyclone.

Analysis of all 544 cases of entry of southerly cyclones into the European USSR, occurring during 1971-1975, indicated that they always developed after the formation of an ozone minimum zone in this region and that in 84% of the cases they developed after two or three days (Table 2). The centers of cyclones, upon entry into the European USSR, lagged a little and deviated from the centers of the ozone minima by an average of 300-400 km.

All the entries of cyclones into the European USSR, on the basis of their relationship to the distribution of the total content of ozone, can first be divided into three groups.

The first group includes cyclones whose trajectories, upon entry into the European USSR, were deflected to the left from the trajectories of motion of regions of an ozone minimum. During the considered period 30 trajectories of cyclones or 37% of all the cases were deflected to the left from the ozone trajectories. In all these cases southerly cyclones persisted for an average of 5-6 days. These cyclones usually advanced far to the north or northeast (Fig. 1a).

The second group includes cyclones whose trajectories were deflected to the right from the trajectories of motion of regions of ozone minima. During 1971-1975 there were 14 such cyclones or 17% of all the cases. These cyclones were rapidly filled (they did not exist more than two days) and most did not advance to the north of 50°N (Fig. 1b).

The third group includes cyclones whose trajectories intersected the trajectories of motion of ozone minima (37 cyclones or 46% of the cases). In these cases the lifetime of the cyclones did not exceed 5 days. It should be noted that if after intersection the trajectory of motion of a cyclone was deflected to the right from the trajectory of the ozone minimum, the cyclone is rapidly filled; however, if the trajectory of the cyclone is deflected to the left of the trajectory of the ozone minimum, the cyclone is intensified and exists for several more days. However, in all cases (81) the ozone minimum is filled a day or two earlier than a cyclone.

During the considered period there were several cases when cyclones, after entry into the southern European USSR, changed the direction of motion and turned abruptly to the west. The region of the ozone minimum was also displaced to the west, somewhat outrunning the cyclone. The cyclone trajectory in this case is deflected to the right of the trajectory of the ozone minimum. One such entry occurred during the period 7-11 September 1971.

A joint analysis of data on the day-to-day variability of the total content of ozone and pressure at sea level during the development of cyclones and regions of ozone minima and along the trajectories of their motion indicated that there is a rather close correlation between them (Table 3). Such a correlation was obtained for the first time in [8] in 1975. As indicated by the data in this table, in only 37 cases, that is, in 46% of the cases, the pressure during the development of cyclones decreased by more than 5 mb; in all the remaining cases the pressure changes during one day were insignificant. The data in Table 3 also indicate that the formation of a cyclone occurred on the average two days after the formation of a region of a minimum of the total ozone content in this same region and the entry of a cyclone into the European USSR begins on the average a day after the onset of arrival of the ozone minimum.

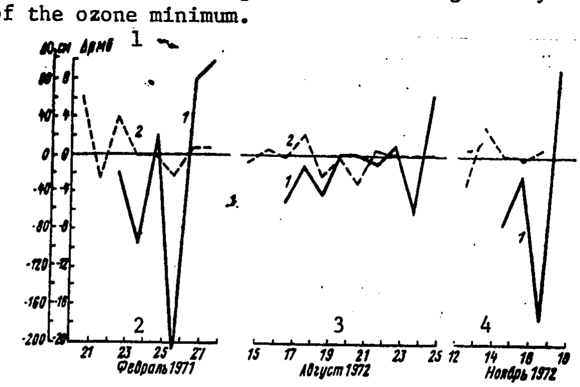

Fig. 2. Curves of the distribution of the day-to-day variability of pressure (1) and ozone (2) with the formation and entry of cyclones and ozone minima regions.

KEY:

1. p mb
2. February
3. August
4. November

Table 3

Day-to-Day Variability of Ozone Content and Pressure With the Development (A, B) and With the Arrival of Ozone Minima (A_1) and Southerly Cyclones (B_1) in European USSR in 1971-1972

Даты возникновения и затухания *мин* O_3 1	*A*	Дата выхода *мин* O_3 2	A_1	Даты возникновения и заполнения циклона 3	*Б*	Дата выхода циклона 4	$Б_1$
			1971 г.				
31 XII—5 I	—0,030	2 I	—0,025	2—5 I	—2	3 I	—17
21—28 II	—0,091	25 II	0	23—28 II	—2	26 II	—20
30 III—2 IV	—0,020	1 IV	—0,015	1—3 IV	0	2 IV	—10
28 VI—10 VII	—0,084	1 VII	0,050	1—10 VII	—5	2 VII	—5
19—25 VII	0,060	20 VII	0,060	21—25 VII	—2	21 VII	—2
22—28 VII	—0,026	23 VII	—0,011	24—28 VII	—3	24 VII	—3
3—11 IX	—0,034	4 IX	0,008	7—11 IX	—10	7 IX	—10
10—14 IX	—0,015	11 IX	—0,034	12—16 IX	—6	12 IX	—6
15—21 IX	—0,020	16 IX	—0,030	17—22 IX	—14	17 IX	—14
18—23 IX	0	18 IX	0	20—24 IX	—6	20 IX	—6
4—7 X	0	5 X	0	6—7 X	0	6 X	0
7—13 XI	—0,020	7 XI	—0,020	11—14 IX	0	11 XI	—5
11—15 XI	0	12 XI	0,020	14—16 IX	—5	14 XI	0
16—22 XI	—0,007	18 XI	—0,011	20—23 XI	0	20 XI	0
			1972 г.				
26—30 III	0	28 III	—0,009	29—30 III	—9	29 III	—9
19—24 IV	—0,016	21 IV	—0,013	21—25 IV	—1	21 IV	—1
19—25 IV	—0,010	21 IV	0,020	22—25 IV	—3	22 IV	—3
22—27 IV	0	25 IV	0	24—28 IV	—10	26 IV	8
30 V—3 VI	—0,020	1 VI	0,020	2—3 VI	—3	3 VI	—3
15—18 VI	—0,011	15 VI	—0,020	16—18 VI	0	16 VI	0
21—25 VI	—0,011	24 VI	—0,003	24—25 VI	0	24 VI	0
27 VI—5 VII	—0,010	30 VI	—0,015	30 VI—5 VII	—6	30 VI	—6
2—7 VII	0	4 VII	0	5—8 VII	1	5 VII	1
8—16 VII	—0,002	8 VII	—0,002	11—17 VII	0	11 VII	0
16—25 VIII	0	19 VIII	—0,014	18—25 VIII	—1	21 VIII	0
22—28 VIII	0	24 VIII	—0,015	25—28 VIII	0	25 VIII	0
31 VIII—4 IX	—0,010	1 IX	0,020	2—5 IX	0	2 IX	0
12—20 IX	—0,020	15 IX	0,040	15—21 IX	—10	15 IX	10
21—26 IX	0	23 IX	—0,015	24—27 IX	—1	24 IX	—1
15—17 X	0	16 X	—0,005	17—18 X	—1	17 X	—1
18—23 X	0,010	19 X	0,020	20—24 X	0	20 X	0
19—24 X	0	21 X	0	21—24 X	1	22 X	—7
20—22 X	—0,010	21 X	0,020	22—23 X	1	22 X	1
12—17 XI	—0,029	13 XI	0,032	16—18 XI	—1	16 XI	—1

KEY:

1. Dates of development and attenuation of O_3 minimum
2. Date of arrival of O_3 minimum
3. Dates of formation and filling of cyclone
4. Date of arrival of cyclone

Since the pressure changes accompanying the generation of southerly cyclones are in large part insignificant (Table 3), which does not always make it possible to detect them at once, the appearance of a region of an ozone minimum and an ozone trough, directed from northeast to southwest, is a good predictor for prediction of the entry of southerly cyclones onto the European territory of the USSR.

An analysis of graphs of the variability of the total content of ozone and pressure at sea level with their generation and along cyclone tracks, as is the case for ozone minima, indicated that the course of day-to-day variability of pressure in a cyclone duplicates the course of variability of ozone in the region of the minimum with an average lag of two days (Fig. 2). The characteristic course of day-to-day variability of both ozone and pressure with the appearance of cyclones is first a marked decrease and then an increase and again a decrease of ozone.

Thus, an analysis of maps of the total ozone content is also becoming necessary for daily weather forecasts because the changes in the total ozone content can serve as a predictor in predicting the entry of southerly cyclones into the European USSR.

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EVALUATION OF THE EFFECTIVENESS OF REMOTE SENSING OF THE GEOPOTENTIAL FIELD OVER THE NORTHERN HEMISPHERE

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[Article by A. Ya. Kazakov and Candidate of Physical and Mathematical Sciences O. M. Pokrovskiy, Leningrad State University, submitted for publication 30 May 1979]

Abstract: The authors propose a method for evaluating the information yield of global systems for observation of the geopotential field with the use of empirical orthogonal functions. The article gives a comparison of the characteristics of the effectiveness of the network of aerological stations, systems for remote sensing of the atmosphere and their combination. The spatial structure of errors in analysis of the geopotential field H_{500} in the northern hemisphere is studied for the mentioned observation systems. Zones are defined in which the analytical errors contain anomalous mesoscale components. The local and global information contribution of remote measurements carried out in individual polar orbits is investigated. The effect of "information saturation" is discussed.

[Text] The matters of use of data from remote thermal sensing of the atmosphere in numerical analysis and weather forecasting occupy an important place in investigations carried out in the area of overlap between atmospheric optics and dynamic meteorology [8]. A great many studies have been devoted to the problems involved in the four-dimensional assimilation of asynchronous meteorological information. These results have been summarized in [5]. The prospects for increasing the effectiveness of the procedures for assimilation are related to methods for the direct use of spectrometric information in objective analysis and its matching with ordinary aerological data [2, 3, 9].

In addition to the development of methods for the interpretation and analysis of the results of remote sensing it is of considerable interest to carry out investigations for evaluating the information yield from different systems of meteorological observations. It is of special importance to compare the data contributions of the existing aerological network of observations and the planned system for remote sensing of the atmosphere in numerical analysis and short-range forecasting. Also of practical interest are investigations of the role of both observation systems in the analysis of the components of meteorological fields of different scales and also determination of the information availability for individual geographical regions. Evaluations of the spatial distribution of errors in analysis are necessary as a point of departure for studying the effectiveness of the collected data from the point of view of the efficiency of numerical forecasting.

There can be two approaches from the methodological point of view. The first is based on use of the traditional objective analysis approach in which there is discrimination of limited spatial regions of "points of influence" (stations) for carrying out optimum interpolation and assimilation. In [1] this approach was employed in computing theoretical evaluations of analytical errors for both observation systems.

An alternative approach is related to realization of a numerical model of global data analysis. This model was employed in [7] for objective analysis purposes. Its effectiveness is determined to a considerable degree by the adequate choice of base functions.

In this paper such a method is proposed for evaluating the information yield of observation systems with respect to the geopotential field. As the basis we used empirical orthogonal functions (EOF) representing the statistical structure of the most important pressure field fluctuations. It is clear that the EOF base will not always be sufficiently flexible for objective analysis purposes due to limitation on the wave spectrum represented in it. However, evaluations of information content are averaged values and therefore must be determined by the statistical structure of the set of records, and not by the characteristics of its individual representatives.

Another prerequisite for the use of EOF in this investigation is the circumstance that due to the presence of spatial correlation of errors in remote sensing data the possibilities of detecting mesoscale disturbances of the geopotential field are limited. The EOF basis for the most part contains long-wave components.

Characteristics of Information Yield of Observation Systems

We will examine the formalism of evaluation of the information yield of systems for observing fluctuations of the geopotential field. We will use a statistical formulation of the problem. In the case of direct observations carried out in the network of aerological stations, the observation equation has the following form:

$$\tilde{x}(s)=x(s)+\varepsilon(s). \tag{1}$$

Here x(s) is the true geopotential value for some fixed isobaric surface at a point of intersection in a horizontal grid s ($s = 1,\ldots,N$), $\varepsilon(s)$ is the observation error, $\tilde{x}(s)$ is a value obtained from measurement.

The observation equation written in terms of deviations from the mean values $\tilde{x}(s)$ has a similar form. It can be assumed that $\varepsilon = (\varepsilon(1),\ldots,\varepsilon(N))$ is the random vector of observation errors having zero mean components and the stipulated matrix of covariations $K_\varepsilon = \sigma^2 \cdot I$ (I is the unit matrix). The latter expression means that the observation errors at network stations are statistically independent.

It is known that the interpretation of data from remote measurements is related to solution of the linearized algebraic system

$$\tilde{\Delta I}(s)=A\cdot\Delta T(s)+\delta(s). \tag{2}$$

Here $A\cdot\Delta T = \Delta I$ and ΔT are the vectors of deviations in values of the intensity of outgoing radiation and the temperature profile from the corresponding means $\bar{I}(s)$ and $\bar{T}(s)$, $\delta(s)$ is the vector of observation errors. We will assume that the δ components have zero means and the known covariation matrix K_δ.

We will assume that the set of subsatellite points does not intersect with the set of points in the aerological network. In this case $s = N+1,\ldots,M$.

The deviation of the relative geopotential value $\Delta x(s)$ at the point s is a linear functional of the type $r^* \Delta T(s)$ (* is the transposition symbol), where the vector-row $r^* = (b_1,\ldots,b_p, 0,\ldots,0)$; b_i are quadrature coefficients used in approximation of the integral, which arises when computing height (geopotential). In this case the observation equation assumes the following form:

$$\tilde{\Delta x}(s)=\Delta x(s)+\varepsilon(s). \tag{3}$$

Here $\Delta x(s)$ is the true deviation of geopotential from the mean value $x(s)$ at the point s. The $\varepsilon(s)$ value in (3) is the sum of the error in evaluating the relative geopotential $\varepsilon_1(s)$ and the contribution of uncertainty in information on surface pressure $\varepsilon_2(s)$. It follows from (2) that

$$\varepsilon_1(s) = q^*\cdot\delta(s) - r^*\cdot P\cdot\Delta T,\quad q^* = r^*\cdot A^+,\quad A^+$$

is a matrix which is pseudoinverse to A, $P = I - A^+\cdot A$. The corresponding dispersion is equal to

$$\sigma^2_{\varepsilon_1} = q^*\cdot K_\delta\, q+r^*\cdot P\cdot K_T\cdot P\cdot r.$$

Evaluations for the $\sigma^2_{\varepsilon_2}$ values are given in [8].

Now we will rewrite equations (2), (3) in spectral form with use of the EOF basis. Assume that the matrix

$$Z = \{Z_{ij}\}_{i=1}^{R}{}_{j=1}^{K}$$

is formed by the values of the first K EOF in a grid containing R points of intersection. We will assume that $K \ll M < R$. For the vector of deviations of geopotential at the points of grid intersection $\Delta X = (\Delta x(1), \ldots, \Delta x(R))^*$ the following representation is correct

$$\Delta X = Z \cdot c + \gamma. \tag{4}$$

Here $c = (c_1, \ldots, c_k)^*$ is a set of unknown coefficients, γ is the vector of the "remainders" of the series.

In the set of points of grid intersection we will discriminate the full set of observation points and we will denote it ω. Then, for these points from system (4) it is possible to discriminate a subsystem which, with equations (1)-(3) taken into account, is rewritten in the form

$$\Delta X(\omega) = Z_\omega \cdot c + \tilde{\varepsilon}(\omega). \tag{5}$$

Here

$$\tilde{\varepsilon}(\omega) = \gamma(\omega) + \varepsilon(\omega),$$

$\varepsilon(\omega)$ is a vector whose components were formed from the observation errors entering into (1) and (3).

The model (5) makes it possible to evaluate the accuracy in determining the coefficients c on the basis of the existing observation system. When using the EOF as a system of base functions the components of the vector c are independent random values with zero means and the stipilated dispersions λ_i^2 $(i = 1, \ldots, k)$. The dispersion of the norm of the vector of "remainders" γ is determined by the "tail" of the series $\sum_{i=k+1}^{R} \lambda_i^2$.

The covariation matrix of the vector c has the diagonal form $\Lambda = \mathrm{diag}\,(\lambda_1^2, \ldots, \lambda_k^2)$. We will denote the covariation matrix of the vector $\tilde{\varepsilon}(\omega)$ by $\tilde{K}_\varepsilon$.

It was demonstrated in [6, 10] that the use of 40-50 EOF ensures an adequate degree of approximation of the geopotential field in the sense that the value of the relative accumulated dispersion

$$\left(\sum_{i=1}^{k} \lambda_i^2 \Big/ \sum_{i=1}^{R} \lambda_i^2\right)$$

exceeds the established "threshold of significance" of the order 0.95-0.98.

Therefore the k value does not exceed 40-50. The sum M+N can be a value of several hundreds or thousands. Two identical representations are correct for the remaining covariation matrix $\tilde{\Lambda}$ of evaluations of the vector c. In one of them it is necessary to invert a matrix of the order M x M; in the other - a matrix of the dimensionality k x k. Taking into account that M k, we will make use of an approach in which it is necessary to invert a matrix of lesser dimensionality. This formula has the form

$$\tilde{\Lambda} = (Z^*_\omega \cdot K_\varepsilon^{-1} \cdot Z_\omega + \Lambda^{-1})^{-1}. \qquad (6)$$

Conversion to a regular grid makes it possible to obtain evaluations of accuracy on the basis of computations of the residual matrix of covariations

$$\tilde{\Sigma} = Z \cdot \tilde{\Lambda} \cdot Z^*. \qquad (7)$$

It is desirable to carry out an analysis of the results on the basis of absolute and relative accuracy characteristics. The values of the standard deviations $\tilde{\sigma}_{ll}$ ($l = 1,\ldots,R$), representing the square roots of the diagonal elements $\tilde{\Sigma}$, characterize the accuracy in analysis of the geopotential field. Convenient characteristics of information yield relative to the set of measurement data are the values of the relative residual dispersions $d_i^2 = \tilde{\sigma}^2_{ii}/\sigma^2_{ii}$ ($i = 1,\ldots,R$). Here σ^2_{ii} are the diagonal elements of the matrix of natural covariations of the geopotential values in a regular grid $\Sigma = Z \cdot \Lambda \cdot Z^*$. It is also desirable to examine the mean values of the dispersion

$$\tilde{d}^2 = \left(\sum_{i=1}^{R} d_i^2/R\right)$$

and the standard deviation $\sqrt{\tilde{d}^2}$ which give some idea concerning the general effectiveness of the observation system.

In the case of an aerological network of observations the errors in individual measurements can be considered statistically independent values. In this situation the computations of the product of the matrices in formula (6) are carried out using single summation of the components

$$p_{nm} = \sum_j z_{jn} \cdot \tilde{k}^\varepsilon_{jj} \cdot z_{jm}$$

($\tilde{k}^\varepsilon_{ij}$ are elements of the matrix K^{-1}). The errors in the results of remote sensing are [2, 12] spatially correlated. In this case the computation of the elements of the product of the matrices in (6) must be accomplished on the basis of double summing

$$p'_{nm} = \sum_i \sum_j z_{in} \cdot \tilde{k}^\varepsilon_{ij} \cdot z_{jm} = \sum_i z_{in} \sum_j \tilde{k}^\varepsilon_{ij} \cdot z_{jm}.$$

It is clear that the dependence k_{ij} as functions of the subscripts i, j has a unimodal structure with a maximum with i = j. In addition, it can be assumed that the EOF components $\{z_{in}\}$ vary quite smoothly as a function of the grid index i. Accordingly, proceeding on the basis of the relationships cited above it can be concluded that the structures of the p_{nm} and p'_{nm} data masses must be similar. However, due to the presence of additional summation the nature of the dependence of the index must be smoother in the case p'_{nm}. The absolute values p'_{nn} on the average can exceed the components p_{nn}. This means that the effective dispersion of errors in correlated observations is greater than for statistically independent observations. The noted circumstance in this case is associated with the fact that with addition of a symmetric matrix with zero diagonal and non-negative nondiagonal matrices to a diagonal positively determined matrix its norm will not decrease.

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Comparative Analysis of Effectiveness of Observation Systems

The computation scheme presented above can serve as a basis for evaluating the effectiveness of existing and planned observation systems. The object of consideration in this paper is the geopotential field H_{500}. The purpose of the investigation is a comparison of the information yield of data from the aerological network and the system of satellite observations.

Existing satellites in polar orbits make it possible to obtain indirect information on the global distribution of meteorological parameters over a period of 10-12 hours; this requires a definite time correction of data [8]. In the case of a planned system of 2-3 satellites, functioning simultaneously, the time interval is reduced to 3-6 hours and the problem of the asynchronicity of the collected information ceases to be so significant. Accordingly, we will not deal with the time aspect of the problem. As the EOF we used the J. Rinne grid base [10], describing the region of the high and temperate latitudes in the northern hemisphere. The mentioned EOF were obtained on the basis of trial and error and an analysis of a sample for 1965-1968 containing 228 records. This base very satisfactorily reproduces the principal peculiarities of the H_{500} field and is employed effectively for numerical analysis and short-range forecasting purposes [10]. The base is stipulated in a grid constructed using a polar stereographic projection and containing 1080 points of intersection. The distance between the grid points of intersection is about 300 km. In the basic computations we used the first 20 functions from the set of EOF provided us through the kindness of Doctor J. Rinne. This set of functions corresponds to the 96% level of relative accumulated dispersion when using dependent statistics and the 91% level when using an independent sample [10]. For the considered set of EOF the lower limit of effective wavelengths is 4000 km. Therefore, the corresponding evaluations of analytical accuracy to a considerable degree reflect large-scale fluctuations of the geopotential field.

The basis for the computations which were made was the actual positioning of the network of stations for aerological sounding and the set of subsatellite points for polar orbits, one following the other each 20° in longitude. It was assumed that the width of the band scanned by the spectrometer from a satellite is 500 km at the surface [8]. For a simplification of the computations as the points of data input we used the closest points of intersection in the grid at which the EOF are stipulated. However, it must be remembered that the mathematical scheme considered above does not impose any limitations on the totality of points in the ω set.

Now we will proceed to an examination of the results of these computations. First we will discuss the average characteristics of the effectiveness of three observation systems: a) ground network of stations, b) system for global remote sensing, c) combined system of aerological sounding and remote sensing. The mean values of the relative analytical errors

$$\sqrt{\overline{d^2}}$$

of the H_{500} field are given in Table 1. It follows from the presented data that in the absence of spatial correlation in the errors in remote measurements the system of satellite monitoring could be not less effective than the network of ground aerological stations. According to the evaluations in [2, 11], the correlation radius of such errors is 1000 km or more. The use of a correlation radius of 900 km in the computations leads to a decrease in the effectiveness of analysis of the geopotential field by a factor of 1 1/2. But even in this case the level of analytical errors when the analysis is based only on satellite data is only 30-50% higher than on the basis of aerological information. This circumstance makes it desirable to have a combined observation system. These evaluations indicate that the contribution of data from remote sounding (with use of optimum assimilation schemes) ensures an increase in analytical accuracy by a factor of approximately 1 1/2. It should be noted that the inclusion of mesoscale geopotential field components in the analysis probably can somewhat change the relationship of the effectiveness indices in favor of the aerological network of stations.

Table 1

Mean Characteristics of Effectiveness $\sqrt{\overline{d^2}}$ of Analysis of H_{500} Field for Different Observation Systems

Система наблюдений 1	Аэрологическая сеть 2	Дистанционное зондирование 3					Объединенная система наблюдений 4	
5 Точность данных наблюдений, м	15	30	30	50	50	15/30	15/30
6 Корреляция ошибок	нет 7	нет	да 8	нет	да	9 нет/нет	10 нет/да
$\sqrt{\overline{d^2}} \times 10^2$	3,1	2,4	3,6	4,0	6,0	1,6	2,0

KEY:

1. Observation system
2. Aerological network
3. Remote sensing
4. Combined
5. Accuracy of observation data, m
6. Correlation of errors
7. no
8. yes
9. no/no
10. no/yes

The relative accumulated dispersion, corresponding to the set of EOF [10], converges quite rapidly to 1. For example, for sets consisting of 16, 20, 24 and 28 first elements of the EOF base, the mentioned values are 0.935, 0.961, 0.973, 0.982 respectively. Therefore, the dependence of the mean relative analytical errors (Table 2) on the number of functions with $k \approx 20$ is quite weak. The cited data indicate that even 20 EOF make it possible to obtain reasonable evaluations of the effectiveness of the considered observation systems.

Table 2

Mean Characteristics of Effectiveness $\sqrt{\overline{a^2}} \times 10^2$ of Analysis in Dependence on Number of EOF

Система наблюдений 1	Точность данных, м 2	Корреляция ошибок 3	4 Число ЭОФ			
			16	20	24	28
5 Дистанционное зондирование	30	да 7	3,56	3,63	3,65	3,67
6 Объединенная система	15/30	нет/да 8	2,01	2,04	2,05	2,06

KEY:

1. Observation system
2. Accuracy of data, m
3. Correlation of errors
4. Number of EOF
5. Remote sensing
6. Combined system
7. yes
8. no/yes

Spatial Structure of Information Yield Characteristics

Now we will discuss the spatial distribution of the mean analytical error $\tilde{\sigma}$ on the basis of data from the observation systems which we considered above. First we will discuss the structure of the values of the mean natural fluctuations σ of the geopotential field. The distribution represented in Fig. 1b is characterized by two important anomalous zones which are associated with the presence of standing waves in the northeastern part of Canada and Siberia. The indicated regions are joined by a "saddle" extending out in the polar zone. We should note the complex spatial structure of the statistics of fluctuations of the geopotential field in the synoptically active zone of the northeastern Atlantic.

The use of data from aerological observations (Fig. 1a) ensures uniformity in the coverage with data for a large part of the continental regions. There is an extensive region of maximum analytical errors in the central part of the North Atlantic which extended into the polar regions. The deficit of data availability is related to "uncovered regions."

Now we will examine the case of a remote measurements system. Different variants of the computations are given in Figures 2, 3a. According to the estimates in [8], the σ_{ε_1} values for the random value ε figuring in formula (3) fall in the range 30-40 m. Taking into account that σ_{ε_2} is 5-15 m [8], the computations were made for the cases σ_ε = 30 and 50 m. The spatial correlation of errors in remote observations exerts no influence on the structure of the $\tilde{\sigma}$ distribution (Fig. 2). We can note only some smoothing

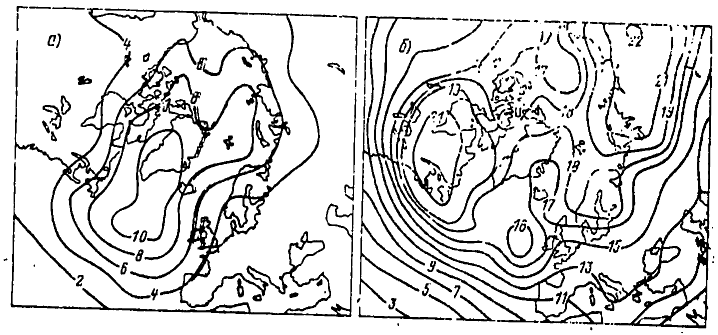

Fig. 1. Spatial distribution of analytical errors $\tilde{\sigma}$ of H_{500} field on the basis of data from the aerological network (σ_ε = 15 m) (a) and "natural" fluctuations of the field H_{500} σ (dam) (b).

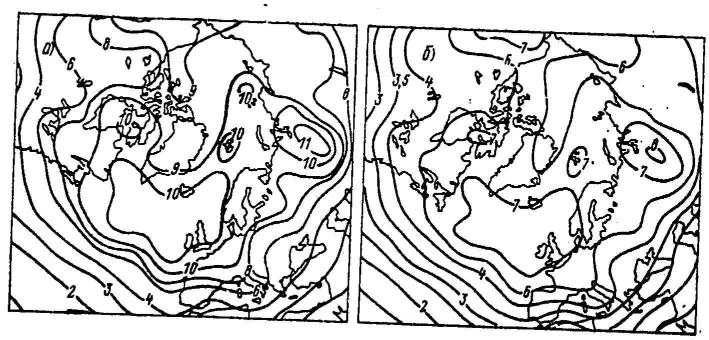

Fig. 2. Spatial distribution of analytical errors $\tilde{\sigma}$ of H_{500} field according to remote sensing data. a) σ_ε = 50 m (with correlation taken into account), b) σ_ε = 50 m (without correlation taken into account).

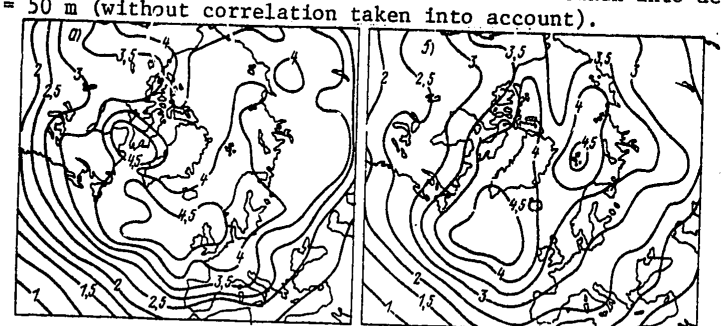

Fig. 3. Spatial distribution of analytical errors $\tilde{\sigma}$ of H_{500} field. a) according to data from remote sensing (σ_ε = 30 m), b) according to data from combined observation system: σ_ε = 15 m (aerology), σ_ε = 30 m (remote sensing).

effect. The anomalous zones of maximum values in analytical errors correspond to the position of regions of standing waves, where mesoscale disturbances dominate, which cannot be satisfactorily indicated on the basis of remote measurements [3]. The anomalous zone in the North Atlantic is most extensive. Its area is somewhat reduced with an increase in measurement accuracy (Fig. 3a). The spatial distribution of analytical errors has two anomalous regions of maximum values. One of them is situated on the polar periphery of the Siberian zone of standing waves; another is situated in the North Atlantic (Fig. 3b). It can be noted that the entire polar basin is poorly supplied with data even with the availability of a combined observation system.

The noted nonuniformities in the spatial structure of data coverage characteristics indicate the desirability of formulating the problem of the optimum planning of an observation system. Below we will examine some aspects of this problem applicable to a remote sensing system.

Evaluations of Effectiveness of Contribution of Remote Measurements in Individual Orbits

The volume of information stored in the on-board memory devices is always limited. Therefore, it is necessary to clarify the most important regions of collection of data ensuring their maximum information contribution. In connection with the specifics of obtaining satellite information it is desirable to examine the role of individual orbits. For this purpose we computed two characteristics of the effectiveness of measurements in individual orbits, separated from one another by 20° in longitude. The information contribution of the s-th orbit, as a result of analysis in the entire network is described by the value

$$r_1^s = \sqrt{\sum_{i=1}^{R} (\tilde{\sigma}_{ii}^s/\sigma_{ii}^s)^2/R}.$$

The effectiveness of these same data for the group of subsatellite points for a particular orbit is characterized by the ratio

$$r_2^s = \sqrt{\sum_{i \in L_s} (\tilde{\sigma}_{ii}^s/\sigma_{ii}^s)^2/R_s}.$$

In these expressions σ_{ii}^s are the standard analytical errors after input of observational data in the s-th orbit, L_s is a set consisting of R_s indices corresponding to the subsatellite points in a particular orbit. Thus, averaging in the formula for r_1^s is accomplished for the entire grid, but in the expression for r_2^s is done only for the subsatellite points of the current orbit.

According to the cited data (Table 3), in a case when the presence of an aerological network of stations is not taken into account it is orbit No 6 which is most informative with respect to both characteristics. It passes over anomalous regions of standing waves over the northeastern part of Canada and Siberia where there are maximum mean fluctuations of the geopotential field. Computations show that orbit No 9, situated to the west of the Atlantic coast of Europe, is least informative. The excess of the maximum values over the minimum r_1 and r_2 values is about 40%. Although the data obtained from individual orbits ensure some refining of the analytical results over the entire area, this information is of particular importance for the subsatellite points. Here the accuracy indices exceed the mean values for the grid by a factor of 4.

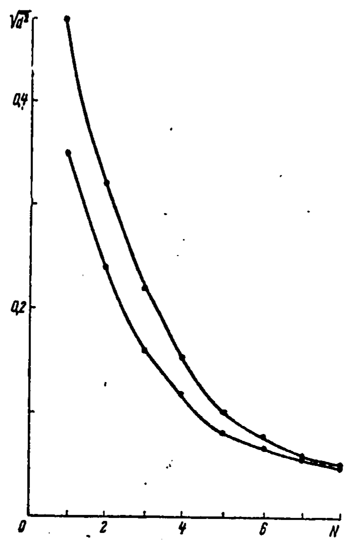

Fig. 4. Dependence of $\sqrt{\overline{d^2}}$ on number of orbital revolutions in which measurements were made (the two extreme curves are given).

Table 3

Global and Local Characteristics of Effectiveness of Measurements in Individual Orbits

№ орбиты (1)	Долгота, град (2)		Без учета аэрологической сети (3)		С учетом аэрологической сети (4)	
	в. д. (5)	з. д. (6)	$r_1 \times 10$	$r_2 \times 10$	$r_1 \times 10^2$	$r_2 \times 10^2$
1	0	180	4,89	1,29	2,23	2,15
2	20	160	4,71	1,30	2,22	2,18
3	40	140	4,87	1,28	2,21	2,25
4	60	120	4,67	1,23	2,22	2,08
5	80	100	4,81	1,33	2,24	1,96
6	100	80	3,95	1,15	2,17	2,11
7	120	60	4,44	1.45	2,15	2.53
8	140	40	4,58	1,40	2,12	2,97
9	160	20	5,20	1,49	2,24	2,57

KEY:

1. No of orbit
2. Longitude, °
3. Without allowance for aerological network
4. With allowance for aerological network
5. E
6. W

In a case when the presence of a network of aerological stations is taken into account the dependences of r_1 and r_2 on longitude are in antiphase. The most informative orbit (with respect to r_1) is No 8, which passes over the Atlantic Ocean, but is least effective for local analysis purposes. On the other hand, orbit No 5, associated with Western Siberia, while being least informative for global analysis purposes, is most effective for obtaining evaluations of geopotential at subsatellite points.

Now we will discuss the possible reasons for this phenomenon. We will turn to the spatial distribution of analytical errors when using data from the aerological network (Fig. 1a). In continental regions there is a dominance of the long-wave component. In the Atlantic Ocean area the wave spectrum to a considerable degree contains mesoscale fluctuations, whose remote sensing is difficult [3]. Therefore, from the measurements made over the ocean we for the most part extract information only on the long-wave component. Thus, the low accuracy in reconstructing the mesoscale components causes an appreciable increase in r^2 values for orbits passing over ocean areas.

Figure 4 shows the dependence of the $\sqrt{d^2}$ value on the number of orbits whose measurements were used in the analysis. The figure shows two extremal curves from nine possible sets of successive orbits. It can therefore be seen that "data saturation" already sets in after 4 to 5 revolutions. This is attributable to the fact that the collected data relate primarily to macroscale components of the geopotential field. A further increase in the density and volume of data does not lead to a substantial refinement of the analytical results.

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CALCULATING THE ZONE OF CLEARING FROM A LINEAR HEAT SOURCE IN CLOUDS

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[Article by Candidate of Physical and Mathematical Sciences A. S. Kabanov and M. M. Troyanov, Institute of Experimental Meteorology, submitted for publication 22 May 1979]

Abstract: The authors determined the quantitative relationships between the extent of zones of clearing formed by the temperature field from a linear heat source and the parameters of the cloud medium. The heat source was situated both far from the underlying surface and near it. When the heat source was situated near the underlying surface it was assumed that the velocity of the oncoming flow and heat conductivity increase with altitude in accordance with a power law. The influence of convection on regular deformation of the clearing zone was analyzed.

[Text] The heat method for dissipating a fog has long been in use. In a number of countries (United States, Great Britain, France, Japan) studies have been made of the energy scattering of a fog. In France a system for clearing a fog using turbojet engines situated on a landing strip has been introduced. There are theoretical and experimental studies devoted to the clearing of clouds under the influence of different heat sources, such as [2, 7, 8]. However, theoretical computations of the heat method essentially involve an estimate of the quantity of heat for the evaporation of a fog in a definite volume from the balance ratio: in order to evaporate a fog in a particular volume it is necessary that this volume be heated by 1-2°C. In this case no allowance is made for the nonuniformity of the field of temperature disturbances formed by the local heat source, which can lead to substantial errors. A recently published study [1] gives a numerical model for computing the zone of clearing from a point heat source with a constant diffusion coefficient and wind velocity and without convection taken into account. In this study there is no clarification of the interrelationship between different characteristics of the clearing zone and the zone of undisturbed effect of the medium.

This paper has the objective of determining the principal quantitative relationships between the parameters of the clearing zone in the cloud medium from a linear heat source and the characteristics of the medium, and also the source intensity. Although it must be admitted that the most realistic allowance for all the interrelated processes is possible on the basis of development of a numerical model, nevertheless the simplified model taken here as a basis makes it possible to attain a satisfactory understanding of the principal interrelationships and to obtain quantitative evaluations important for practical purposes.

Formulation of problem. Assume that a linear heat source of infinite length with the intensity Q (Q is the quantity of heat released into a unit length in a unit time) is operative in a fog at the height h from the earth's surface. We will select a Cartesian coordinate system in such a way that the y-axis coincides with the source, the z-axis is directed upward. An external flux moving with the velocity $U>0$ is stipulated along the x-axis. In a cloud medium undisturbed by any phenomenon the specific moisture content $P_0 = W_0+q_0(T_0)$ is a stipulated constant value. Here W_0 is specific liquid-water content, $q_0(T_0)$ is specific humidity, equal to saturated humidity at the undisturbed temperature T_0. The problem involves determination of the boundary of the region of clearing S at which the liquid-water content of the fog becomes equal to zero.

We will assume that the vapor in the cloud medium outside the clearing zone is in a saturated state and that the liquid-drop moisture is completely entrained by the air flow. Then the distribution of specific liquid-water content outside the clearing zone is instantaneously adjusted under the temperature field formed by the heat source, that is (T' is a disturbance of background temperature)

$$W(x, z, t) = P_0 - q[T_0 + T'(x, z, t)], \tag{1}$$

At the boundary S the specific liquid-water content is equal to zero and therefore

$$q[T_0 + T_s'(x, z, t)] = P_0. \tag{2}$$

Equation (2) is fundamental for determining the boundary S; for computing S it is necessary to know the temperature field T'.

For liquid-water contents observed in a fog there is satisfaction of the condition

$$\mu = \frac{LT_s'}{R_v T_0^2} \ll 1,$$

where L is the specific latent heat of condensation of water, R_v is the water vapor gas constant. Expansion of the q value in (2) with an accuracy to the linear term containing μ makes it possible to represent equation (2) in the form

$$T_s' = \frac{R_v T_0^2}{Lq_0} W_0. \tag{3}$$

Expression (3) was derived with neglecting of small changes in moisture content due to compressibility of the medium caused by a temperature change.

Assuming the coefficient of turbulent diffusion for water vapor to be equal to the coefficient of turbulent thermal conductivity $\varkappa$, the stationary equation for temperature in the cloud medium, taking phase transitions into account, in accordance with [4], can be represented by transformation to the form

$$U\frac{\partial T'}{\partial x}+\varkappa\Delta T'=\alpha\frac{Q}{c_p\rho}\,\delta(x)\,\delta(z),\quad \Delta=\frac{\partial^2}{\partial x^2}+\frac{\partial^2}{\partial z^2}, \tag{4}$$

where Q = const, ρ is air density, c_p is the specific heat capacity of the air at constant pressure,

$$\alpha=\left(1+\frac{L^2 q_0}{c_p R_v T_0^2}\right)^{-1}<1.$$

With T_0 = 283 K $\alpha \approx 0.45$.

The parameter $\alpha<1$, represented in (4) by the factor before the intensity of the heat source, takes into account heat expenditures on phase transitions outside the clearing zone. Equation (4) can be used in computing the S surface. A proof of this is the fact that (4) is equivalent to the equation

$$U\frac{\partial A}{\partial x}+\varkappa\Delta A=\frac{Q}{c_p\rho}\,\delta(x)\,\delta(z)$$

for the known invariant $A = T' - L/c_p\,(W - W_0)$ relative to phase transitions, which is applicable everywhere, including the boundary of the clearing zone. Taking (3) into account, we find that at the boundary of the clearing zone T' and A are related by the expression $T' = \alpha A$, that is, the α value in actuality takes into account the decrease in the level of the temperature field due to phase transitions.

Solution of the stationary problem without a boundary. We will assume that the heat source is situated far from the underlying surface so that its influence can be neglected. The $\varkappa$ and U parameters are assumed to be constant. Then the stationary solution of equation (4) with the boundary conditions $T' = 0$ with $x = \pm\infty$, $z = \pm\infty$ is the function [6]

$$T'(x,\ z)=\frac{\alpha Q}{2\pi c_p\rho\varkappa}\exp\left(\frac{Ux}{2\varkappa}\right)K_0\left(\frac{Ur}{2\varkappa}\right). \tag{5}$$

where $K_0(\xi)$ is a Bessel function of the second kind (zero order) of the fictitious argument: $r = (x^2+z^2)^{1/2}$. Now we will examine quite intense heat sources for which the distance r_s from the origin of coordinates to the S surface is considerably greater than the $2\varkappa/U$ value. This makes it possible to use an asymptotic expression for the function

$$K_0:\ K_0\left(\frac{Ur_s}{2\varkappa}\right)=\sqrt{\frac{\pi\varkappa}{Ur_s}}\exp\left(-\frac{Ur_s}{2\varkappa}\right).$$

Then (5) and (3) make it possible to find the boundary S:

$$\frac{1}{\sqrt{r}} e^{x-r} = c \equiv 2\sqrt{2\pi} \frac{R_v T_0^2 c_p \rho \varkappa}{\alpha Q L q_0} W_0, \tag{6}$$

where x and r are dimensionless values measured in the units $2\varkappa/U$.

Converting in (6) to the polar coordinates (r, φ), $x = r \cos \varphi$, we find

$$r = -\frac{\ln \frac{c}{a}}{1 - \frac{1}{a}\cos\varphi}, \tag{7}$$

where $a = a(r) = 1 + 1/2r \ln r$; we have $1 < a < 1.5$. If we neglect the changes in the background values T_0 and W_0 within the limits of a region commensurable with the clearing zone and arbitrarily assume that a = const, then (7) will represent a second-order curve. Since $1/a < 1$, then (7) is the equation for an ellipse. Thus, S is an ellipselike figure whose "focus" is situated at the origin of coordinates.

Equation (6) makes it possible to estimate the horizontal and vertical dimensions of the clearing zone. Thus, we find (in dimensional units) that:

a) the distance along the horizontal from the origin of coordinates to the most remote point of the clearing zone "downstream" ($x > 0$) is equal to:

$$l_x = A_0 \left(\frac{Q}{W_0}\right)^2 \frac{1}{\varkappa U}, \quad A_0 = \frac{1}{4\pi}\left(\frac{\alpha L q_0}{R_v T_0^2 c_p \rho}\right)^2; \tag{8}$$

b) the maximum height of the clearing zone is accordingly equal to

$$l_z = B_0 \frac{Q}{U W_0}; \quad B_0 = \sqrt{\frac{2}{\pi e}} \frac{\alpha L q_0}{R_v T_0^2 c_p \rho}, \tag{9}$$

where e is the base of natural logarithms.

With T_0 = 283 K we have $A_0 = 4 \cdot 10^{-23}$, $B_0 = 1.1 \cdot 10^{-11}$ (in esu units).

It follows from (9) that the maximum vertical dimension of the clearing zone is not dependent on the thermal conductivity coefficient. In order to explain this fact we will use a more simplified model of heat transfer. We will visualize that an air particle, passing through the point (0, 0) and then moving along the x-axis with the velocity U, propagates heat in a zone parallel to the z-axis. As is well known, in the one-dimensional problem of heat conductivity from an instantaneous point source the temperature maximum is not dependent on $\varkappa$ at any point. Only the time during which this maximum is attained is dependent on $\varkappa$. In the zone associated with a moving air particle there is a point $z_s = l_z/2$, for which T_s', the temperature disturbance determining the S isotherm, will be maximum. This z_s value is also the maximum section. The T_s' value is attained during the time

$t \simeq \ell_z^2/8\varkappa$. Accordingly, the maximum altitude of the zone does not change, but the distance $x_{max} = U\ell_z^2/8\varkappa$ from the source to the position of the maximum altitude will increase with a decrease in $\varkappa$. With a change in $\varkappa$ and U there is also a change in the characteristic time that the solution reaches a stationary regime $t_{st} \simeq \ell_z^2/2\varkappa$. As $\varkappa$ or U tends to zero there is an unlimited increase in the time for the solution to become stationary.

Formulas (8) and (9) make it possible to evaluate the intensity of the heat source clearing a region of stipulated dimensions. If it is necessary to create a zone of clearing with the horizontal dimension $\ell_x = 4 \cdot 10^4$ cm, then with $\varkappa = 2 \cdot 10^4$ cm^2/sec, T_0 = 283 K, $W_0 = 2 \cdot 10^{-7}$ g/cm^3 we obtain $Q = 1.25 \cdot 10^{10}$ erg/(cm·sec). This intensity is equivalent to the combustion of 2.8 g of kerosene per second per meter length of source. With such an intensity of the source the vertical dimension of the zone is $\ell_z = 3.4 \cdot 10^3$ cm.

Now the condition $r \gg 2\varkappa/U$ can be replaced by the condition $Q \gg Q^*$. Q^* is found from the condition $2\varkappa/U = r(Q^*)$, where as r we use the vertical distance from the source to the boundary of the clearing zone. We note that in this case we do not evaluate the dimensions of the clearing zone with negative x. For our values of the parameters $Q^* = 2.4 \cdot 10^9$ erg/(cm·sec).

Influence of underlying surface. The thermal conductivity coefficient and wind velocity in the atmospheric boundary layer can be parameterized in the following way [3]:

$$U(z) = U\left(\frac{z}{z_1}\right)^m, \quad \varkappa(z) = \varkappa\left(\frac{z}{z_1}\right)^n, \tag{10}$$

where m, $n \geq 0$ are determined experimentally. It has been established that $m \approx 0.15$-0.25; $n \approx 1$ (for example, see [3]).

In clarifying the influence of the underlying surface on the dimensions of the clearing zone it was assumed (by virtue of the fact that $\varkappa \gg \varkappa_{soil}$, $\varkappa_{soil}$ is the thermal conductivity of the soil) that there is no heat flux through the underlying surface. We will place the heat source near the underlying surface. The solution of equation (4) with the boundary condition

$$\left.\frac{\partial T}{\partial z}\right|_{z=0} = 0$$

is the following function (for example see [9, 10]; the diffusion heat transfer along the horizontal was neglected):

$$T'(x, z) = \frac{z_1^q \alpha Q U^{s-1}}{c_p \rho \Gamma(s)(m-n+2)^{2n-1} \varkappa^s x^s} \exp\left\{-\frac{U z^{m-n+2}}{(m-n+2)^2 \varkappa z_1^{m-n} x}\right\}, \tag{11}$$

$$m - n + 2 > 0.$$

where

$$s=\frac{m+1}{m-n+2}, \quad q=\frac{m+n}{m-n+2},$$

$\Gamma(s)$ is the gamma function.

In this case the boundary S is stipulated by the following equation

$$\exp\left[-\frac{Uz^{m-n+2}}{(m-n+2)^2\varkappa z_1^{m-n}x}\right]=\frac{R_v T_0^2 W_0}{Lq_0}\frac{c_p\rho\Gamma(s)(m-n+2)^{2s-1}\varkappa^s}{z_1^q\alpha QU^{s-1}}\equiv c_2,$$

or, in explicit form

$$z^{m-n+2}=-\frac{(m-n+2)^2\varkappa z_1^{m-n}}{U}(s\ln x+\ln c_2). \qquad (12)$$

In this case the clearing zone is similar in form to the upper half of the zone, obtained for velocities of the external flow and thermal conductivity, constant with altitude, but is considerably more elongated along the x-axis.

Equation (12) makes it possible to find the horizontal and vertical dimensions of the clearing zone. The horizontal dimension of the clearing zone is equal to

$$l_x=\frac{1}{c_2^s}=A_1\left(\frac{QU^{s-1}}{W_0\varkappa^s}\right)^s, \quad A_1=\left[\frac{Lq_0z_1^q\alpha}{R_vT_0^2\Gamma(s)(m-n+2)^{2s-1}c_p\rho}\right]^s. \qquad (13)$$

The maximum vertical dimension is equal to

$$l_z=B_1\frac{Q^sU^{s^2-s-1}}{W_0\varkappa^{s^2-1}}, \quad B_1=\left[\frac{(m-n+2)^2z_1^{m-n}}{e}\right]^{\frac{1}{m-n+2}}\times$$
$$\times\left[\frac{Lq_0z_1^q\alpha}{R_vT_0^2\Gamma(s)(m-n+2)^{2s-1}c_p\rho}\right]^{\frac{s}{m-n+2}}. \qquad (14)$$

We will assume that $U = 2\cdot10^2$ cm/sec, $\varkappa = 2\cdot10^4$ $cm^2\cdot$sec, $z_1 = 10^4$ cm, $W_0 = 2\cdot10^{-7}$ g/cm^3, $Q = 1.25\cdot10^{10}$ erg/(cm·sec), n = 1. For m = 0.15, 0.2, 0.25, the vertical dimensions of the clearing zone are equal to $\ell_z = 3.6\cdot10^3$ cm, $3.9\cdot10^3$ cm, $4.4\cdot10^3$ cm. It can be seen that the height of the clearing zone is slightly dependent on m. The horizontal dimension of the clearing zone with n = 1 is not at all dependent on m and is equal to $\ell_x = 6.2\cdot10^5$ cm.

Influence of convection on clearing zone. Convection, caused by a heat source, can raise up the clearing region. In evaluating this effect we will make use of the results in [5], where the author describes the convection induced by a linear heat source in a stably stratified medium. If the influence of the underlying surface on convection is not taken into account, then at the point where the heat source is operative, the vertical velocity [5] is

$$V_z = \frac{g\alpha Q}{2\pi^3 \rho c_p \varkappa T_0 \Omega} \operatorname{arctg} \frac{4\varkappa\Omega}{U^2}, \quad \Omega = \left[\frac{g}{T_0}(\gamma_a - \gamma)\right]^{1/2}, \tag{15}$$

where $\gamma_a = \frac{g}{c_p}, \quad \gamma = -\frac{\partial T_0}{\partial z},$

g is the acceleration of free falling.

Then, at the distance l_x from the heat source the clearing zone rises up to a height of about $h_{con} = V_z\ l x/U$. Using (8) and (15), we find

$$h_\kappa \simeq a_1 \frac{Q^3}{(W_0 \varkappa U)^2 \Omega} \operatorname{arctg} \frac{4\varkappa\Omega}{U^2}, \quad a_1 = \frac{g\alpha A_0}{2\pi^3 \rho c_p T_0}. \tag{16}$$

With T_0 = 283 K $a_1 = 2.6\cdot10^{-28}$ (esu). For $Q = 1.25\cdot10^{10}$ erg/(cm·sec), $W_0 = 2\cdot10^{-7}$ g/cm^3, $\varkappa = 2\cdot10^4$ cm^2/sec, $U = 2\cdot10^2$ cm/sec, $\gamma = 0.4\cdot10^{-4}$ °C/cm, in accordance with (16), $h_{con} = 1.7\cdot10^3$ cm. It follows from (15) and (16) that the oncoming flow decreases the intensity of convection. Such an effect of the oncoming flow is effective when $4\varkappa\Omega_0/U^2 \ll 1$.

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IRREGULARITY OF THE EARTH'S ROTATION AS POSSIBLE INDICES OF GLOBAL WATER EXCHANGE

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[Article by Candidate of Physical and Mathematical Sciences N. S. Sidorenkov, USSR Hydrometeorological Scientific Research Center, submitted for publication 18 April 1979]

Abstract: A redistribution of moisture between the world ocean and the glacial covers changes the components of the tensor of the earth's inertia and probably explains the year-to-year changes in the rate of the earth's rotation and the secular motion of the poles. These irregularities of the earth's rotation are registered by astronomical methods and can be used as integral characteristics of moisture redistribution over the earth. The year-to-year changes in the rate of the earth's rotation quantitatively characterize the increment of water mass in the world ocean, whereas the secular motion of the pole indicates the territory from which the moisture exchange occurs.

[Text] The vector of the instantaneous angular velocity of the earth's rotation can be expanded into three components: one along the mean axis of rotation and two others in the equatorial plane. The first component sets the duration of day and its changes set the nonuniformity of the earth's rotation. The other two determine the coordinates of the instantaneous pole. The purpose of this communication is to direct attention to the fact that the year-to-year changes in the earth's rate of rotation and motions of the poles evidently reflect the redistribution of moisture between the world ocean and the glacial covers of Antarctica and Greenland and can be used as indices of change in level of the world ocean and the state of these glacial covers.

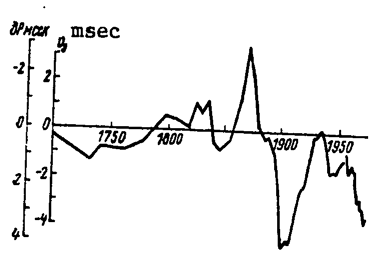

Fig. 1. Changes in rate of earth's rotation during last 300 years.

Year-to-year irregular changes in the rate of the earth's rotation were postulated at the end of the last century as a result of processing data from observations of the moon and other bodies of the solar system. In the 20th century their existence was finally demonstrated in [2]. Since 1955 they have been registered by a fundamentally new method which ensures a very high accuracy (up to $\pm 3 \cdot 10^{-11}$) [4].

Figure 1 illustrates the changes in the earth's rate of rotation during the last 100 years. Plotted along the y-axis are the relative deviations of angular velocity ν_3 in units of the ninth decimal place and the corresponding deviations of length of day δP. It is easy to see the complex irregular changes in the rate of the earth's rotation with characteristic times of about several tens of years. The earth rotated most rapidly in about 1870 and most slowly in about 1903. Between 1903 and 1935 there was an acceleration of the earth's rotation. Between 1935 and the present time there has been a slowing of rotation, sometimes replaced by periods of small acceleration. The latter occurred during 1948-1953, 1958-1960 and 1973-1975.

We note that the slowing of the rate of the earth's rotation noted during the last 100 years is not associated with tidal friction because the rate of tidal slowing is considerably less than that observed. This is also indicated by the fact that in the two centuries preceding 1870 the rate of the earth's rotation was not slowed, but was accelerated. The rate of tidal slowing has virtually no dependence on time.

Fluctuations of the level of the world ocean and the melting of polar ice were some of the first phenomena by which scientists attempted to explain the irregular changes in the earth's rate of rotation [2].

However, quantitative estimates already made during the last century by the noted British physicist Kelvin [10] indicated that in order to explain the irregular changes in the earth's rate of rotation there would have to be improbably great increments in the level of the world ocean. Therefore, since that time the opinion has gained great acceptance that a redistribution of moisture cannot cause the observed fluctuations in the rate of the earth's rotation.

It was demonstrated in our last study on investigation of the role of the atmosphere in the excitation of year-to-year changes in the earth's rate of rotation that the year-to-year changes in the rate of the earth's rotation that have already been described above are caused by the flux of small portions of sometimes positive and sometimes negative moments of momentum passing from the atmosphere to the earth through the surface layer [5]. The most probable supplier of these portions of moment of momentum in the atmosphere is water vapor, that is, it is found that the year-to-year changes in the rate of the earth's rotation in the last analysis are caused by a redistribution of moisture between the world ocean and the glacial covers primarily of Antarctica and Greenland (the role of the remaining glaciers is relatively small due to their small area).

With a redistribution of moisture there is a change in the components of the earth's tensor of inertia, which results in an observable irregularity of the earth's rotation. The dimensionless components ν_i of angular velocity in this case are described in the following way:

$$-\frac{1}{\sigma_0}\frac{d\nu_2}{dt}+\nu_1=-\frac{R_0^2}{2(C-B)}\iint_S \zeta \sin 2\theta \cos\lambda \, dS, \tag{1}$$

$$-\frac{1}{\sigma_0}\frac{d\nu_1}{dt}+\nu_2=-\frac{R_0^2}{2(C-B)}\iint_S \zeta \sin 2\theta \sin\lambda \, dS, \tag{2}$$

$$\nu_3=-(1-k)\frac{R_0^2}{C}\iint_S \zeta \sin^2\theta \, dS. \tag{3}$$

Here σ_0 is the angular frequency of motion of the pole with the Chandler period (1.18 year); t is time; C and B are the earth's polar and equatorial moments of inertia; R_0 is the earth's radius; S is the earth's entire area; $dS = R_0^2 \sin\theta \, d\theta \, d\lambda$; θ is the complement of latitude to 90°; λ is longitude; k = -0.3 is a coefficient taking into account the earth's deformation from a load; $\zeta = \zeta(\theta, \lambda, t)$ is the increment of the specific (that is, relating to a unit area) quantity of moisture (water or ice).

Expressions (1)-(3) make it possible to compute the coordinates of the pole ν_1 and ν_2 and the deviation of the earth's rate of rotation ν_3 from the field of increments of the specific quantity of moisture $\zeta(\theta, \lambda)$ at a fixed moment in time t. Unfortunately, global observations of change in the field $\zeta(\theta, \lambda)$ are lacking. Therefore, there is no possibility of directly computing the effect of redistribution of moisture on the earth's rotation and we will limit ourselves to consideration of a simplified model of redistribution of moisture between the world ocean and the glacial covers of Antarctica and Greenland. The influence of other mountain glaciers and ground water will be neglected.

Since we are interested in changes in the earth's rotation and the field $\zeta(\theta, \lambda)$ with characteristic times of approximately decades, the terms containing derivatives in equations (1) and (2) can be neglected.

Assume that as a result of melting of the glacier covers the specific quantities of ice decrease on the average over the entire Antarctic continent by ζ_A and over the whole of Greenland by ζ_Γ. Then the moisture being set free passes into the world ocean and its level will be increased by the value ζ'_0, which coincides approximately with the increment of the specific quantity of moisture ζ_0 (because $\zeta_0 = \rho_0 \zeta'_0$, where the density $\rho_0 \approx 1$ g/cm^3),

$$\zeta_0 = \frac{S_A \zeta_A + S_\Gamma \zeta_\Gamma}{S_0}. \tag{4}$$

[Γ = Greenland] Here S_0, S_A and S_Γ are the areas of the world ocean and the glacial covers of Antarctica and Greenland. We note that S_A is approximately seven times greater than S_Γ.

With such a model of the global redistribution of moisture the integrals on the right-hand sides of equations (1)-(3) are represented in the form of linear polynomials

$$\nu_i = A_i \zeta_A + \Gamma_i \zeta_\Gamma, \tag{5}$$

where Q_i^A, Q_i^Γ and Q_i^0 are constants and ζ_A, ζ_Γ and ζ_0 are time variables; the subscript i assumes the values 1, 2 and 3. Taking into account expression (4), equations (5) can be simplified, determining the components ν_i either through the state of the glacial covers

$$\nu_i = O_i \zeta_0. \tag{6}$$

or through the level of the world ocean (in the case $\zeta_A = \zeta_\Gamma$)

$$\nu_i = Q_i^A \zeta_A + Q_i^\Gamma \zeta_\Gamma + Q_i^0 \zeta_0, \tag{7}$$

The values of the constants Q_i^A, Q_i, Q_i^0, A_i, Γ_i and O_i were computed and given in the table in units 10^{-10}.

i	Q_i^A	Q_i^Γ	Q_i^0	O_i	A_i	Γ_i	$(A+\Gamma)_i$
1	4,426	—6,956	79,96	135,8	1,27	—7,42	—6,15
2	12,646	6,323	124,88	—294,1	7,72	5,60	13,3
3	—0,0287	—0,0093	—8,856	—8,0	0,32	0,04	0,36

[A = Antarctica; Γ = Greenland; 0 = world ocean]

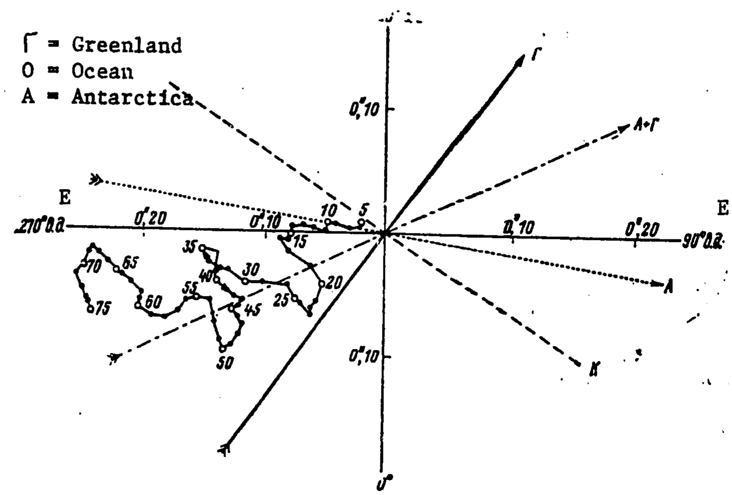

Fig. 2. Directions of displacements of "mean" north pole of the earth's rotation with the accumulation of moisture in Antarctica (A), Greenland (Г), Antarctica and Greenland simultaneously (A + Г), on the continent (K). The broken curve indicates the observed secular motion of the pole.

Expressions (6) and (7) show that with a fixed position of the land with which the exchange of moisture occurs the coordinates of the pole ν_1 and ν_2 are related by a linear dependence to the increment of the specific quantity of ice in this place or water in the world ocean. In other words, each territory corresponds to its meridian, along which there should be displacement of the mean pole of the earth's rotation with a redistribution of moisture between the world ocean and this territory. Figure 2 illustrates the positioning of the meridians along which the mean north pole of the earth's rotation should be displaced. In this figure the arrows indicate the directions of displacements of the pole with the accumulation of moisture in the considered part of the landmass. It can be seen, for example, that with the accumulation of ice in Antarctica the pole must be displaced equatorward along the meridian 81°E and on the entire land as a whole -- along 237°E. With the melting of the ice the pole must be displaced in the opposite direction and the meridians are: for Antarctica -- 261°E and for the land as a whole -- 57°E.

According to data from astronomical observations [7], the earth's north pole of rotation from 1903 through 1975 moved as indicated by the broken curve in Fig. 2. On the average the pole was displaced at a rate of about 10 cm/year along the meridian 285°E, which is close to the trajectory which must be observed with the simultaneous melting of ice in Antarctica and

Greenland. This fact confirms the opinions of glaciologists that during recent decades there has been a reduction of ice reserves in Greenland and Antarctica [1, 9]. The observed rate of displacement of the pole of 10 cm/year should correspond to thawing of the glacial covers of Antarctica and Greenland with the rate $\zeta_{A+\Gamma} \approx 10$ g/(cm^2·year). An analysis of the sinuous character of motion of the pole could tell much about the variability of the global redistribution of moisture in the 20th century because the system of three equations (5) in principle makes it possible to determine the three unknown parameters of global water exchange ζ_A, ζ_Γ and ζ_0 from the three components of angular velocity ν_i obtained from astronomical observations. However, the accuracy in determining the secular motion of the pole at the present time is so low that many astronomers in general deny its reality [8].

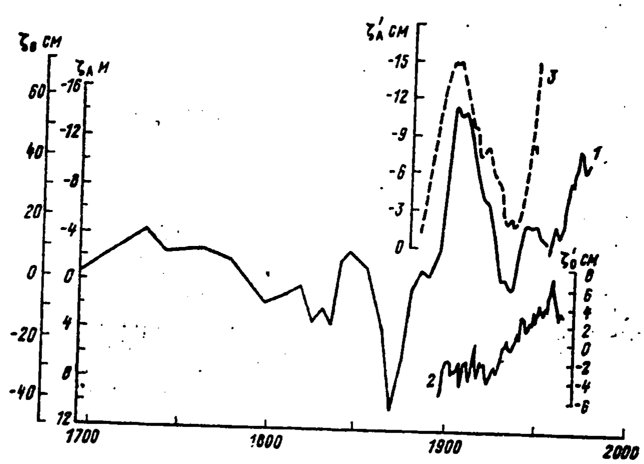

Fig. 3. Changes in some characteristics of the global redistribution of moisture. 1) theoretical curve of specific masses of ice in Antarctica ζ_A or water in the world ocean ζ_0; 2) observed fluctuation of level of world ocean; 3) observed snow accumulation in Antarctica.

As follows from expression (6), the change in the state of the glacial cover of Antarctica must be closely associated with the year-to-year variations of the earth's rate of rotation because the contribution of Antarctica is 8 times greater than the contribution of Greenland. Neglecting the latter, we have

$$\zeta_A = \frac{\nu_3}{A_3}. \tag{8}$$

[3 = earth]

The curve of change in the specific quantity of ice in Antarctica during the last 300 years, computed using formula (8), is shown in Fig. 3. The scale of values ζ_A is given at the right in meters (it is assumed that the ice density is not equal to 0.92 g/cm^3, but 1 g/cm^3, so that 1 m is equivalent to 100 g/cm^3). It can be seen that the thickness of the glacial cover in Antarctica must be decreased from 1870 through 1903 by more than 25 m and then increased by 1935 by 15 m. From 1935 to the present time the thickness of the glacial cover must have decreased.

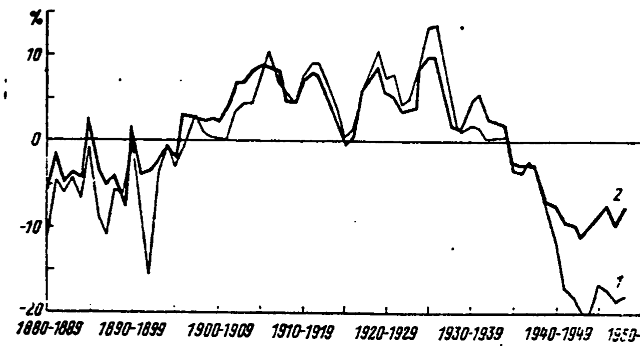

Fig. 4. Curve of moving 10-year anomalies in rate of accumulation of snow in Antarctica [3]. 1) data from simple averaging; 2) data from computations with use of weighting factors.

Through the efforts of the Antarctic expeditions of a number of countries some factual material has now been collected concerning the "input" part of the glacial cover - snow accumulation budget in Antarctica. This has been generalized in a monograph by V. N. Petrov [3]. He processed the annual snow accumulation values for nine stations. Three of them (Amundsen-Scott, Little America and Wilkes) have series since 1880. The remaining stations were activated in the 20th century.

The rate of snow accumulation in Antarctica was extremely different. At the intracontinental stations the mean snow accumulation values, averaged by decades, expressed through water equivalent, fluctuate from 6 to 10 g/(cm^2·year), and at the coastal stations -- from 20 to 40 g/(cm^2·year). V. N. Petrov computed the anomalies of the rate of snow accumulation for each of these stations using moving 10-year periods. On the basis of the averaged data for these nine stations he obtained a graph showing variation of the moving 10-year anomalies of the rate of snow accumulation in Antarctica from 1885 through 1953. We reproduce this in Fig. 4. Along the y-axis we have plotted the anomaly values (values of the relative deviations of the rate of snow accumulation from the norm, for which we use the mean value for the entire observation period) in percent. Curve 1 was obtained by averaging of the rates for all stations without the introduction of weights and curve 2 -- using weighting factors proportional to the area represented by the particular station.

The graph shows that at the end of the 19th century the rate of snow accumulation was below the norm; from the beginning of the 20th century through the end of the 1930's it was above the norm; in the 1940's-1950's the snow accumulation again decreased. This variation agrees well with the fluctuations in acceleration of the earth's rotation. In Fig. 3 for a comparison with curve 1 it is necessary to have an integral curve of snow accumulation in Antarctica. Accordingly, we carried out additional computations. Curve 2 in Fig. 4 was used in determining the values of the anomalies for each year. We subtracted 3% from each determined value (that is, it was assumed that the ice losses exceed ice accumulation by 3%). Then the percentages were converted into absolute values (as the value 100% we used the mean rate of snow accumulation in Antarctica, which according to these same data is equal to 15 g/(cm^2•year)) and we calculated the integral values of snow accumulation, assuming that at the initial moment (1884) it was equal to zero.

The resulting curve of ice accumulation in Antarctica from 1885 through 1950 is shown in Fig. 3 (curve 3). The y-axis for it is situated at the center of the figure in centimeters (1 cm is equivalent to 1 g/cm^2). A comparison of curve 3 with curve 1 reveals a surprising qualitative agreement. Only in the late 1940's does the decrease in ice thickness occur more rapidly than according to curve 1. But, as indicated by an analysis of the initial materials presented in the cited monograph, this is caused by the inclusion of Lazarev and Maudheim stations during recent decades; in comparison with other stations these give a large decrease in the rate of snow accumulation in the 1940's. One must remember the distortions which could be introduced by Greenland and the waters of the land, which were not taken into account here.

A quantitative comparison of the values represented by curves 3 and 1 reveals that the observed characteristics of ice accumulation in Antarctica are several tens of times less than the theoretical values.

The changes in the specific quantity of water in the world ocean also must be closely related to the year-to-year variations in the rate of the earth's rotation, because from equation (7) we have

$$\zeta_0 = \frac{v_3}{O_3}. \tag{9}$$

The variation of the specific quantities of water in the world ocean, computed using formula (9), is given in Fig. 3. The scale of values is given at the left in centimeters.

It can be seen that the specific quantity of water in the world ocean must have increased from 1870 through 1903 by approximately 100 g/cm^2 and then have decreased by 1935 by 60 g/cm^2. From 1935 through the present time it must have increased. Only during the periods from 1950 through 1954, from 1958 through 1961 and from 1972 through 1975 there must have been brief decreases of the ζ_0 value.

Unfortunately, the specific quantity of water in the world ocean is not measured. As an indirect index of change in the water mass of the world ocean we can use its level. A change in the quantity of water by 1 g/cm^2 is equivalent to an increment of the level by 1 cm. During recent years several studies have been published which give curves of variation of the level of the world ocean in the 20th century [6]. We reproduce one of them in Fig. 3 (curve 2). We note that this curve was constructed by simple averaging of measurements of 72 level-measuring posts situated along different shores of the world ocean.

It can be seen that from 1900 through 1930 the level of the world ocean varied in the range from -2 to -5 cm. In this time interval if the level did decrease it was by not more than 3 cm instead of the 60 cm required for explaining the acceleration of the earth's rotation (it is assumed that water density and ocean capacity are constant). After 1930 the level of the world ocean increased. In 1958 it was +7 cm, that is, it became 8 cm higher than in 1930. The nonuniformity of rotation indicates an increase in level only since 1935 and a value of about 20 cm. It is interesting to note that after 1949 and 1958 observations give, as is required by computations, a level decrease.

Thus, level variations of the world ocean during recent decades are in qualitative agreement with the level changes required for explaining the nonuniformity of the earth's rotation. True, the noted agreement is considerably poorer than that which was noted in the examination of ice accumulation in Antarctica. This is understandable because the sea level is unambiguously related to the specific quantity of water in a column. To a considerable degree it is dependent on water temperature and salinity. For example, with a constant value of the specific quantity of water the increase in the mean temperature of a column of water with a thickness of 4 km by 1°C can increase the level due to volumetric expansion by 0.5 m.

A quantitative estimate of the effect of redistribution of moisture on the earth's rotation both in the case of ice accumulation and in the case of variations in the level of the world ocean indicates that the observed redistribution of moisture is several tens of times less than the theoretical value. It is possible that this disagreement arises due to a noncorrespondence between the actual and theoretically computed properties of the earth. In formulating the theory it was assumed that the earth rotates as a single unit and therefore we used its planetary moments of inertia C and B. In the case of short-period effects (not more than 1 year) the earth's subcrustal matter completely satisfies this hypothesis. In the case of a redistribution of moisture, effects of the same sign last for decades. It is not impossible that with such prolonged effects the earth's subcrustal matter does not behave as a solid body but flows like a viscous fluid. Then the influence of moisture redistribution is not propagated to the entire earth, but only to its uppermost layer, evidently lying on the asthenosphere, and in the theory it is necessary to use the moments of inertia only for this upper layer of the earth. They are several tens of times less than the C and B values used in equations (1)-(3). Accordingly, all the estimates of the specific quantities of ice in Antarctica and water in the world ocean cited above on the basis of formulas (8) and (9) must be decreased by this number of times. Then there will be not only a qualitative, but probably, also a quantitative agreement between observations and theory.

In summary, it can be concluded that in any case the year-to-year changes in the earth's rate of rotation and secular motion of the pole can serve as an integral characteristic of the global redistribution of moisture between the world ocean and the continents (in particular, the glacial covers of Antarctica and Greenland). The direction of the secular motion of the pole indicates the place of thawing or accumulation of the ice and the change in the earth's rate of rotation characterizes, from the quantitative point of view, the variations in thickness of the glacial covers or variations in water mass in the world ocean.

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EVALUATION OF PARAMETERS OF A COMPLETELY BROKEN WAVE FLOW

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 60-68

[Article by Candidate of Physical and Mathematical Sciences B. A. Shulyak, Central High-Elevation Hydrometeorological Observatory, submitted for publication 11 May 1979]

Abstract: This article gives the results of approximate solution of the plane problem of the parameters of an "offset" flow created by a broken wave on the filtering slope along the shore. The derived expressions relate the limit of penetration of the "offset" flow on the shore slope, the instantaneous and mean velocities of the fluid particles in the offset flow and time of movement of the offset flow in forward and reverse directions. Despite the approximate nature of the derived expressions they are of practical interest in formulating a physico-mathematical model of the process of working of the accumulative sand shore of seas, lakes and reservoirs.

[Text] The problem of the propagation of waves on a slope and their deformation is of more than theoretical interest. It is also of great practical importance, for example, in the problem of the dynamics of shore processes. However, despite its timeliness, it still has not been solved due to major purely technical difficulties associated with the nonlinearity of the equations of motion and boundary conditions, and also the discontinuity of the latter for the case of deformation of waves with total breaking and transition into a plane-parallel flow.

Meanwhile, not one of the problems relating to the dynamics of the processes in the shore zone of the sea can be solved without taking into account the parameters of a broken wave flow. Therefore, it is desirable to seek an approximate solution, using, for example, the conservation equations and drawing on necessary additional information from general considerations and observational data.

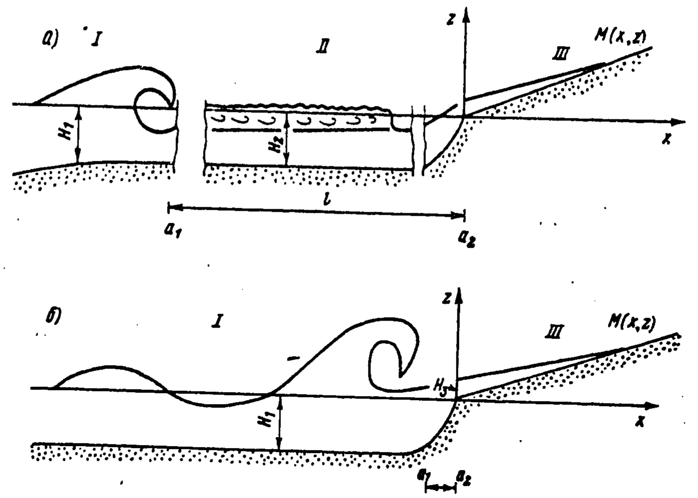

Fig. 1. Schematic diagram of process of wave breaking and formation of offset flow.

Such an approach naturally precludes the possibility of a complete and accurate description of the movement of flows, but with clever use it is easy to obtain expressions for a number of basic characteristics of movement parameters.

First we will in general form examine the plane problem of the breaking of waves before a filtering slope and transition of wave movement of a fluid into an unsteady plane-parallel flow ending in an offset flow on the part of the shore slope above the water (Fig. 1).

We will assume that:

1. A wave flow with the height h, with the period τ, with the wavelength λ and the phase velocity C is propagated on the surface of a fluid with the depth H_1, from left to right from the region I in the direction of region III. The breaking of the flow occurs in the neighborhood of the point a_1, to the right of which arises a highly turbulent plane-parallel unsteady current with the velocity $u_2 = u_2(x, z, t)$, with the depth H_2 (x, t), whereas in the neighborhood of point a_2 the flow is transformed into an offset flow (region III) having the parameters $U_3(x, z, t)$, H_3 (x, t), the equation for the free surface is $\zeta = \zeta(x, t)$, and moving upward along the slope, completely stops. Due to the complex dependence in which the parameters of these flows are situated, it is difficult to use the correlation between them for an arbitrary moment in time. However,

for some characteristic moments in time the energy relationships will extremely simply relate the principal parameters of the flows.

2. The nature of breaking of the waves is dependent on the angle of the underwater slope and the wave parameters themselves. The running of the offset flow onto the shore slope can be preceded by the single or multiple collapsing of the crests. In addition, the last collapsing can be accompanied by the movement of the flow in the form of a bore (prior to running onto the shore slope, see Fig. 1a) or directly undergo transition into an offset flow (Fig. 1b). Without here discussing the reasons for the differences in movement, we will discriminate these two characteristic cases for separate examination.

3. The energy losses in turbulent pulsations during the breaking of the wave flow and filtration losses will be taken into account only integrally -- through the dissipation coefficients $\varkappa^*$ and $\varkappa_1^*$ respectively or coefficients inverse of them -- unscattered energy $\varkappa = 1 - \varkappa^*$ and $\varkappa_1 = 1 - \varkappa_1^*$.

4. The parameters of the flows in all three regions are functions of time. However, since the degree of these dependences is different, we will assume that in region II it can be neglected (in comparison with region III), that is, we will assume $u_2 = \overline{u}_2(x, z, t)$, $\overline{H}_2(x, t) = H_2 \simeq \overline{H}_1$. But for region III we will consider both the averaged values and those dependent on time $U_3(x, t)$, $H_3(x, t)$, etc.

5. In the breaking of waves we will assume that only a part of the wave energy, specifically, that beneath the crest, that is, $1/2\ E_1$, makes a contribution to the directed movement of a plane-parallel flow.

In a general case, from the condition of conservation of energy and momentum, for a breaking wave and a plane-parallel flow (Fig. 1a) we have

$$\frac{\rho}{\tau}\int_0^{\tau}\int_{-H_1}^{\zeta}\int_0^{\lambda} v(x, z, t)\, dx\, dz\, dt = \rho\int_0^{l}\int_{-H_2}^{0} u_2\, dx\, dz. \tag{1}$$

The left integral in (1) gives the momentum, averaged by period (or length), in a volume $1\cdot\lambda\cdot H_1$, region I, whereas the right integral gives the momentum value in a broken flow with the extent l and the volume $1\cdot l\cdot H_2 \approx 1\cdot l\cdot H_1$.

From the condition of energy conservation with transition into region II (Fig. 1a) or directly into region III, when $l \sim 0$ (see Fig. 1b), we have, in the first case

$$\frac{1}{2}\varkappa E_1 \equiv \frac{\varkappa\rho}{2\tau}\int_0^{\tau}\int_{-H_1}^{\zeta}\int_0^{\lambda} v^2\, dx\, dz\, dt = \frac{\rho}{2}\int_0^{l}\int_{-H_2}^{0} \overline{u}_2^2\, dx\, dz, \tag{2}$$

and accordingly, in the second

$$\frac{1}{2}\varkappa E_1 \equiv \frac{\varkappa\rho}{2\tau}\int_0^{\tau}\int_{-H_1}^{\zeta}\int_0^{\lambda} v^2\,dx\,dz\,dt = \frac{\rho}{2}\int_0^{x_d(t)}\int_{z_1}^{z_s(t)} U_3^2\,dx\,dz + \rho g\int_0^{x_d(t)}\int_{z_1}^{z_3(t)} z\,dx\,dz, \tag{2'}$$

where the left integral gives the averaged wave energy $E_1 = \rho g h^2 \lambda / 8$, whereas the integrals at the right give the kinetic energy E^{kin}_{off} and the potential energy E^{pot}_{off} of the offset flow at arbitrary moments in time.

In the latter equation U_3, as well as the integration limits x_d and z_3 are functions of time. Accordingly, it is exceedingly complex to use (2) for determining the sought-for values. However, at some characteristic moments in time, such as at $t = t_0$, which corresponds to the onset of running of the offset flow onto the shore slope, or $t = t_1$, which corresponds to total stoppage of the offset flow on the slope, E^{pot}_{off} and E^{kin}_{off} independently become equal to zero. As a result, we have:

a) when $t = t_1$

$$\frac{1}{2}\varkappa E_1 \equiv \frac{\varkappa}{2}\frac{\rho g h^2 \lambda}{8} = \rho g \int_0^{x_d}\int_{z_1}^{z_3} z\,dx\,dz; \tag{2''}$$

b) when $t = t_0$

$$\frac{1}{2}\varkappa E_1 \equiv \frac{\varkappa}{2}\rho g\frac{h^2\lambda}{8} = \frac{\rho}{2}\bar{U}_3^2\int_0^{x_d}\int_{z_1}^{z_3} dx\,dz. \tag{2'''}$$

Here $\bar{U}_3 = U_3(x_0 t) \simeq U_3(x_0)$ is the mean velocity with which the fluid volume Ω passes through the section $x = x_0 = 0$, $t_0 \leqslant t < t_1$, $\Omega = \iint dx dz$ and x_d is the coordinate of the limiting point which the offset flow reaches.

The equation for the shoreline slope onto which the offset flow moves can be stipulated independently of the flow characteristics, and generally, independently of time:

$$z_1 = f_1(x,\ t) = f_1(x). \tag{3}$$

[This is unquestionably correct for the given problem, but not for the problem of shore dynamics.]

But for the equation for the free surface such an approximation is impossible: it is a function of the parameters of waves (energy), slope profile and time:

$$\zeta_3 \equiv z_3 = f_2(E_1, \varkappa, x, z, t, \ldots). \qquad (4)$$

In this case if the functions f_1 and f_2 have been established, it is possible to determine one of the principal sought-for characteristics -- the coordinates of the point M (x_d, z_d):

$$x_d = f_3(f_1, f_2), \qquad (5)$$

$$z_d = f_4(f_1, f_2). \qquad (6)$$

The joint use of conditions (2)-(6) makes it possible to obtain expressions also for other flow parameters. Therefore, the main factor complicating the problem is the successful choice of approximations for f_1 and f_2.

As an illustration of such an approach for solution of the considered problem we will consider the special case of destruction of waves in front of the shore slope when the wave flow after destruction is immediately transformed to an offset flow (region III, Fig. 1b).

In this case it is impossible to use equation (1); we are left with system (2)-(4).

We introduce a Cartesian coordinate system in accordance with Fig. 1b and represent the line equation (3) by the linear function

$$z_1 = f_1(x) = ax, \qquad (7)$$

in which, for ordinary conditions, a = 0.2.

As a result of the smallness of the slope the equation for the surface of the offset flow ζ_3 can also be represented in a linear approximation

$$\zeta_3 = f_2(x, t) = b(t, E_1, \ldots)ax + H_3(t, E_2, \ldots), \qquad (8)$$

which for the moment in time $t = t_1$ will correspond to the limiting position of an offset flow with the fixed values b and H_3:

$$\zeta_3 = bax + H_3. \qquad (8')$$

The boundary conditions (8') and (7) determine the coordinates of the points M(x, z) of a moving offset flow front at arbitrary moments in time, whereas when $t = t_1$ -- the upper limit of penetration of the offset flow upslope:

$$x(t) = \frac{H_3(t)}{a(1-b(t))}, \tag{9}$$

$$z(t) = \frac{H_3(t)}{1-b(t)}, \tag{10}$$

and

$$x_d = x(t_1) = \frac{H_3}{a(1-b)}. \tag{9'}$$

$$z_d = z(t_1) = \frac{H_3}{1-b}. \tag{10'}$$

In these expressions the values $b(t_1, E_1, \ldots)$ and $H_3(t_1, E_1, \ldots)$ remain undetermined. Both these values could be determined if in addition to the energy equation it was possible to write an equation for the volume of fluid flowing frcm region I (Fig. 1b) into region III. Since in this article we will not consider this problem, we will limit ourselves only to a determination of the b value from the energy conservation condition.

For the moment t_1 the kinetic energy in the remaining part of the offset flow is equal to zero; therefore, from (2") we have

$$\frac{\varkappa}{2}\frac{\rho g h^2 \lambda}{8} = \rho g \int_0^{x_d}\int_{z_1}^{z_3} z\,dx\,dz = \tag{11}$$

$$= \frac{\rho g}{2}\left(\frac{ab^2 x_d^3}{3} + abx_d^2 H_3 + x_d H_3^2 - \frac{a^2 x_d^3}{3}\right),$$

or, with (9') taken into account,

$$\frac{1}{2}\varkappa\frac{\rho g h^2}{8} = \frac{1}{6}\rho g H_3^3 \frac{2+b}{a(1-b)}. \tag{12}$$

As a convenience, introducing the notation

$$\mu = \frac{3a\varkappa E_1}{\rho g H_3^3} \equiv \mu_1 \frac{1}{H_3^3}, \tag{13}$$

we rewrite (12) in the form

$$\mu b^2 + (1-2\mu)b + \mu - 2 = 0, \tag{14}$$

in which for b, in accordance with the limitations imposed on the free surface equation by the line (7), the following condition must be satisfied

$$0 < b < 1. \tag{15}$$

From the quadratic equation (14), with (15) taken into account, we obtain

[redacted]

$$1 - b = \frac{(1 + 4\mu)^{1/2}}{2\mu} + \frac{1}{2\mu}. \tag{16}$$

Now we will evaluate the order of magnitude of μ. For the case of not very small parameters of waves approaching the shore itself before final breaking, that is, with $h \gtrsim 30$ cm, $\lambda \gtrsim 1000$ cm, $H_3 \sim 10$ cm, the μ evaluation gives:

$$\mu \gg 2. \tag{17}$$

Therefore, in (16) it is possible to drop the second term, as well as the "1" in the numerator of the first term. Therefore

$$1 - b \approx \mu^{-1/2}, \tag{18}$$

hence for $x_d\ (t_1)$, $z_d\ (t_1)$ and $\zeta_3(x,\ t_1)$ we obtain

$$x_d \equiv x_d(t_1) \approx \frac{h}{2}\left(\frac{3}{2} \varkappa \frac{\lambda}{aH_3}\right)^{1/2}, \tag{19}$$

$$z_d \equiv z_d(t_1) \approx \frac{h}{2}\left(\frac{3}{2} \varkappa \frac{a\lambda}{H_3}\right)^{1/2}, \tag{20}$$

$$\zeta_3 \approx (1 - H_3^{3/2}\mu_1^{-1/2})\, ax + H_3. \tag{21}$$

From the derived expressions for the coordinates of the extreme point of the offset flow $M(x_d,\ z_d)$ and the equation of its free surface at the time $t = t_1$ it is easy to obtain expressions also for the arbitrary time t if use is made of an expression for instantaneous velocity (see below).

A somewhat more complex problem is that of determination of velocities in the offset flow; these are not only a function of time, but also the two coordinates x and z.

An offset flow is unsteady movement on a slant plane. Therefore, the different layers must move at different velocities. However, if the thickness of the offset flow is small and as a result of strong turbulent mixing it is possible to neglect the dependence on z, then for determining the instantaneous velocity of an arbitrary i-th particle U_3^i it is necessary to take into account only the limiting point $M^i(t_1)$ with the coordinate $x_d^i(t_i)$, which it attains at the time $t = t_1$.

Among all the particles in the offset flow the ones of greatest interest in our problem are those moving at the leading edge of the offset flow. They move with the velocity $U_3^i = U_d^i$ and attain the coordinates x_d.

For each of the offset flow particles it is of interest to consider both their mean velocity on the slope $\overline{U}_3^i$ and as a general characteristic -- the mean velocity of all the particles on the slope $\overline{\overline{U}}_3$.

First we will write an expression for the mean velocity $\overline{U}_3 = U_3$ (0) with which the particles pass through the point x = 0. From condition (2") we have the equation

$$\varkappa \frac{\rho g h^2 \lambda}{16} = \frac{\rho \overline{U}_3^2}{2} \int_0^{x_d} \int_{z_1}^{z_2} d\Omega = \frac{\rho \overline{U}_3^2}{2} \Omega. \tag{22}$$

Since in the considered approximation for f_1 and f_2

$$\Omega = \int_0^{x_d} \int_{z_1}^{z_2} d\Omega = \frac{1}{2} x_d H_3 (t_1), \tag{23}$$

then

$$\overline{U}_3(0) \approx \left(\frac{\varkappa}{6} a g^2 h^2 \lambda H_3^{-1} \right)^{1/4}. \tag{24}$$

We will also cite the velocity, averaged for all particles on the slope

$$\overline{\overline{U}}_3 \approx \frac{1}{3} \left(\frac{3\varkappa}{2} a g^2 h^2 \lambda H_3^{-1} \right)^{1/4}. \tag{25}$$

In order to determine instantaneous velocity it is sufficient to know only its Lagrangian coordinate x = x(t) at an arbitrary moment in time and the limiting point $M^i(x_d^i, z_d^i)$ to which it rises along the slope. From simple kinematic expressions for an arbitrary moment in time t, satisfying the condition

$$t_0 < t < t_1, \tag{26}$$

we have

$$U_3^i (x, t) = (2 ag)^{1/2} [(x_d^i)^{1/2} - x_i^{1/2} (t)]. \tag{27}$$

where $(2ag\ x_d^i)^{1/2}$ is the maximum velocity of the i-th particle which it has with passage through the point $x_0 = 0$ in order to attain the limiting point x_d^i at the time t_1, whereas $x_i(t)$ is its Lagrangian coordinate at an arbitrary moment in time.

For the leading particles, in accordance with (27) and (19), for the instantaneous velocity value we have

$$U_3 (t) = (2 ag)^{1/2} \left[\left(\frac{3}{8} \frac{\varkappa h^2 \lambda}{a H_3} \right)^{1/4} - x^{1/2} (t) \right] = \left(\frac{3}{2} \varkappa a g^2 \frac{h^2 \lambda}{H_3} \right)^{1/4} - gt, \tag{28}$$

Therefore, the maximum $U_3 = U_3(x_0)$ value is equal to:

$$U_3 = (2 a g x_d)^{1/2} = \left(\frac{3}{2} \varkappa a g h^2 \lambda H_3^{-1} \right)^{1/4}. \tag{29}$$

We will also cite the mean velocity of fluid particles moving at the edge of the offset flow which will be necessary for evaluating the time of movement of the flow on the slope:

$$\overline{U}_3 = \frac{1}{2} U_3(x_d) \approx \frac{1}{2}\left(\frac{3}{2} \varkappa a g^2 h^2 \lambda H_3^{-1}\right)^{1/4}. \quad (30)$$

Now we will discuss the problem of evaluating the influence of filtering of the fluid on the flow parameters of the offset flow during its "return" movement.

Filtering leads to an energy loss. It is easily taken into account through the losses of potential energy transported by the volume of fluid filtered on the slope. If the fraction of the remaining volume of fluid by the time of onset of "return" movement $t + \delta t$ is equal to $\varkappa_1$, then by this time the following part remains from the potential energy

$$\varkappa_1 \varkappa \frac{1}{2} E_1. \quad (31)$$

From a comparison of this value with (2") it follows that with allowance for filtration the coefficient $\varkappa$ everywhere must be replaced by $\varkappa_1 \varkappa$. Therefore, at the time $t_1 + \delta t_1$ the coordinates of the offset flow are displaced by the value δx_d and $\delta z_d \sim \varkappa^{1/2} (1 - \varkappa_1)^{1/2}$ and we can write approximately as follows [see note]:

$$x_d(t_1 + \delta t_1) \approx \frac{h}{2}\left(\frac{3}{2} \frac{\varkappa \varkappa_1 \lambda}{a H_3}\right)^{1/2}, \quad (32)$$

$$z_d(t_1 + \delta t_1) \approx \frac{h}{2}\left(\frac{3}{2} \frac{\varkappa \varkappa_1 a \lambda}{H_3}\right)^{1/2}. \quad (33)$$

There will also be a similar change for the free surface equation

$$\zeta_3 = \zeta_3(t_1 + \delta t_1) \approx (1 - H_3^{3/2} \varkappa_1^{-1/2} \mu_1^{-1/2})\, ax + H_3, \quad (34)$$

Finally, for the averaged velocity $\overline{U}_3^*$ with return movement we obtain

$$\overline{U}_3^* = \frac{1}{2}\left(\frac{3 \varkappa_1 \varkappa a g^2 h^2 \lambda}{2 H_3}\right)^{1/4} = \varkappa_1^{1/4} \overline{U}_3. \quad (35)$$

As follows from the expression written above, if the filtration losses are small (about 20%, $\varkappa_1 \sim 0.8$), this effect can be neglected, since $\varkappa_1$ enters in the power 1/4 and is reflected only in the range 5%, which is considerably less than the level of errors stipulated by the formulas themselves.

Naturally, the expressions for $\overline{\overline{U}}_3$, U_3, $\overline{U}_d$, etc. will be similar.

Note: In actuality, filtration occurs over the entire surface of the slope. However, it transpires most intensively in the uppermost part of the offset flow.

The results also make it possible to evaluate the time of movement of the offset flow in forward and backward directions -- T and T* respectively.

In the forward direction for T we have

$$T \approx \int_0^{x_d} (1 + z_3'^2)\, dx \cdot \left[x_d^{-1} \int_0^{x_d} U_3\, dx \right]^{-1} = x_d \left(1 + \frac{1}{2} a^2\right) \overline{U}_3^{-1} = \tag{36}$$

$$= \left(1 + \frac{1}{2} a^2\right) \left(\frac{3}{2} \frac{\varkappa h^2 \lambda}{a^3 g^2 H_3}\right)^{1/4} \approx \left(\frac{3}{2} \frac{\varkappa h^2 \lambda}{a^3 g^2 H_3}\right)^{1/4},$$

where $\int_0^{x_d} (1 + z_3'^2)\, dx$

is the length of the shore slope under the offset flow

$$x_d^{-1} \int_0^{x_d} U_3\, dx$$

is the mean velocity of motion of the edge of the offset flow.

In accordance with the comment expressed on the influence of $\varkappa_1$ we immediately write an expression for the time of the return movement:

$$T^* = \varkappa_1^{1/4} T \tag{37}$$

or since $\varkappa_1^{1/4} \sim 1$ with small losses in filtration, it can be assumed that

$$T^* \approx T. \tag{38}$$

Now we will also discuss the correlation between the time of movement of the offset flow T and T* and the wave period τ. As a simplification we will assume that T = T*, as a result of which we replace the sum T + T* by 2T. Then

$$2T \approx 2\left(1 + \frac{1}{2} a^2\right) \left(\frac{3}{2} \frac{\varkappa h^2 \lambda}{a^3 g^2 H_3}\right)^{1/4} \approx 2\left(\frac{3}{2} \frac{\varkappa h^2 \lambda}{a^3 g^2 H_3}\right)^{1/4} \tag{39}$$

or

$$2T \approx \frac{\lambda}{c} \left(\frac{6}{\pi^2} \frac{\varkappa \delta (kH_1)^2}{a^3 H_3}\right)^{1/4} = \frac{\lambda}{c} \eta, \tag{39'}$$

where δ, kH_1 are the steepness of the waves and the dimensionless depth prior to breaking of the waves. Therefore

$$2T \approx \eta \tau. \tag{40}$$

Now we will evaluate the order of magnitude of η for some special values of the flow parameters. For example, with h = 50 cm, H_1 = 50 cm, λ = 10^3 cm, H_3 = 10 cm and a = 0.2 we have

$$\eta \sim 1. \tag{41}$$

Since δ and kH_1 at the shore slope itself do not change very greatly and the dependence on them is quite weak, η will change little with a change in the wave parameters. Therefore, with a certain approximation it can be assumed that

$$\eta = \text{const}, \tag{42}$$

which gives

$$2T \approx \tau, \tag{43}$$

or

$$T \approx T^* \approx \frac{\tau}{2}. \tag{44}$$

The resulting approximate dependences make it possible to give an approximate expression also for the remaining H_3 value undetermined above. Substituting (43) into the left-hand side of (39), we obtain

$$H_3 \approx \frac{6}{\pi^3} \frac{\varkappa\delta (\kappa H_1)^2 \cdot h}{a^3}. \tag{45}$$

It is of interest to use (45) in order to exclude H_3 from the expressions derived above for the parameters of the offset flow. By substituting (45), for example, into (19), (20), (21), (24) and (28) we obtain

$$x_d \approx \pi a \lambda (4\, \kappa H_1)^{-1} = a \lambda^2 (8\, H_1)^{-1}, \tag{46}$$

$$z_d \approx \pi a^3 \lambda (4 \tag{47}$$

$$\zeta_3 \approx 1 - \mu^{-1/2}\, \kappa H_1^3 (6\, \varkappa\delta h)^{3/2} a^{-1/2} x + 6\, \varkappa\delta\, h (\kappa H_1)^2 \pi^{-2} a^{-3}, \tag{48}$$

$$\overline{U}_3 \approx a \lambda g^{1/2} (2 \sqrt{3}\, H_1^{1/2})^{-1} = 3^{-1/2} \pi a C_\infty^2 C_H^{-1}), \tag{49}$$

$$U_3 \approx \frac{1}{2} a \lambda g^{1/2} H_1^{-1/2} \simeq \pi a C_\infty^2 C_H^{-1}. \tag{50}$$

Important comments must be made concerning the derived approximate expressions (46)-(50), as well as the initial approximate expression (45) itself. The H_3 value enters in the power 1/4 in the expression for η.

Therefore, an inaccuracy, and even a substantial inaccuracy in the H_3 evaluation can exert no significant influence on η and all the remaining expressions containing H_3 in the power 1/4 and even -1/2. In (45) the dependence was reversed and H_3 already varies greatly with a change in the flow parameters. Therefore, its determination from (45) can lead to substantial errors if η in (41) differs appreciably from unity. This must be taken into account when using formulas (45)-(50).

UDC 551.465.41(520)

INFLUENCE OF TURBULENCE AND BOTTOM RELIEF ON THE DYNAMICS OF CURRENTS IN THE BLACK SEA

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[Article by Candidates of Physical and Mathematical Sciences Ye. V. Stanev, Ye. Kh. Donev and T. Z. Dzhioyev, USSR Hydrometeorological Scientific Research Center and Sofia University, submitted for publication 29 March 1979]

Abstract: This article is devoted to an investigation of the influence of turbulence and bottom relief on the dynamics of currents in the Black Sea on the basis of the A. S. Sarkisyan model by means of numerical experiments. For different values of the coefficients of horizontal and vertical turbulent mixing the authors give diagnostic and prognostic computations of the level surface, current and density fields. These computations indicated that turbulence and bottom relief exert a considerable influence on the distribution of the fields of sought-for physical characteristics in a deep baroclinic sea. The variation of the coefficient of lateral exchange exerts a particularly strong influence on the results of these prognostic computations. The kinetic energy and the ratio of kinetic energies for different values of the input parameters, averaged for each horizon, were computed as an illustration of the influence of the integral effect of lateral exchange on dynamics of the sea.

[Text] Introduction. Theoretical investigations of recent years have indicated that there is a close interrelationship between the principal physical factors forming currents in a baroclinic basin. In particular, such factors include turbulence and bottom relief. It is known that the width of coastal boundary layers in model computations is highly dependent on the selected value of the coefficient of horizontal turbulent mixing. But, as indicated in [3], allowance for the coefficient of lateral exchange is also necessary in the internal parts of the sea, especially when there are small intervals of the computation grid.

This paper is devoted to an investigation of the influence of turbulence on the dynamics of currents in the Black Sea on the basis of the A. S. Sarkisyan theoretical model [3] by means of numerical experiments. It follows from diagnostic and prognostic computations carried out for different values of the input parameters that turbulence and bottom relief exert a considerable influence on the distribution of the current and density fields in a deep baroclinic sea, although in the first case the results of the computations to a certain degree were predetermined in advance because the density field, having a decisive importance for the formation of currents, is considered stipulated from observations.

Formulation of Problem. Boundary Conditions

We will examine a deep baroclinic basin with a real bottom and an arbitrary form of the shorelines. Atmospheric pressure P_a at sea level and bottom relief H(x, y) are considered stipulated from observations. The three-dimensional density field ρ (x, y, z) is either stipulated from observations (diagnostic problem) or is computed from the density diffusion equation (prognostic problem).

We will use the following system of equations for hydrodynamics of the sea for investigating the stationary large-scale circulation of waters and the density field:

$$l(v_g - v) = \nu \frac{\partial^2 u}{\partial z^2} + A_l \Delta u, \tag{1}$$

$$-l(u_g - u) = \nu \frac{\partial^2 v}{\partial z^2} + A_l \Delta v, \tag{2}$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0, \tag{3}$$

$$P = \rho_0 g \zeta + g \int_0^z \rho\, dz, \tag{4}$$

$$\frac{\partial \rho}{\partial t} + u \frac{\partial \rho}{\partial x} + v \frac{\partial \rho}{\partial y} + w \frac{\partial \rho}{\partial z} = \mu \Delta \rho + \varkappa \frac{\partial^2 \rho}{\partial z^2}. \tag{5}$$

Here u, v, w are the components of current velocity along the coordinate axes x, y, z (the x axis is directed to the east, the y axis is directed to the north, z is directed vertically downward); u_g, v_g are the components of velocity of the geostrophic current; P, ρ are the pressure and density anomalies; $l = 2\omega \sin\varphi$ is the Coriolis parameter; ω is the angular velocity of the earth's rotation; φ is geographic latitude;

$$\zeta = \zeta_* + \frac{P_a}{\rho_0 g}$$

is reduced sea level; ζ_* is the rise of the free sea surface; ρ_0 is the mean density value; g is the acceleration of free falling; A_l, ν are the coefficients of horizontal and vertical turbulent mixing of the fluid; μ, $\varkappa$ are the coefficients of horizontal and vertical diffusion of density.

Boundary conditions. At the sea surface $z = -\zeta_*(x, y)$ we have:

$$\rho_0 \nu \frac{\partial u}{\partial z} = -\tau_x, \quad \rho_0 \nu \frac{\partial v}{\partial z} = -\tau_y, \quad w = 0, \quad \rho = \rho^0, \tag{6}$$

where τ_x, τ_y are the components of wind shearing stress.

At the sea bottom $z = H(x, y)$ the following conditions are stipulated:

$$u = v = w = 0, \quad \rho = \rho^H. \tag{7}$$

At the lateral boundaries Σ we have:

for the liquid part of the boundary

$$\frac{1}{H}\int_0^H u dz = V_1, \quad \frac{1}{H}\int_0^H v dz = V_2, \quad \rho = \rho^\Sigma; \tag{8}$$

for the solid part of the boundary

$$V_1 = V_2 = 0. \tag{9}$$

The zonal density ρ (y. z) is stipulated at the initial moment in time.

From equations (1)-(4), using the boundary conditions, after simple transformations, we obtain the vorticity equation $\Omega = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}$:

$$\begin{aligned} \beta S_y = {} & [\nu + A_l (\nabla H)^2]\left(\frac{\partial \Omega}{\partial z}\right)_H + \frac{1}{\rho_0}\,\mathrm{rot}\,\bar{\tau} + \\ & + A_l\left[\underline{\Delta\left(\frac{\partial S_y}{\partial x} - \frac{\partial S_x}{\partial y}\right)} - \Delta H \Omega_H - 2\nabla H(\nabla\Omega)_H\right], \end{aligned} \tag{10}$$

where $\beta = dl/d\nu$ is the change in the Coriolis parameter with latitude; S_x, S_y are the components of total flow;

$$\mathrm{rot}\,\bar{\tau} = \frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y}$$

is the vertical component of vorticity of wind shearing stress;

$$\nabla = \frac{\partial}{\partial x}\bar{i} + \frac{\partial}{\partial y}\bar{j}$$ is the Hamiltonian.

All the terms in equation (10) with the coefficient A_l, except those underlined, were obtained as a result of allowance for the variability of bottom relief and horizontal turbulent mixing and are related to bottom topography. Similar terms were obtained by V. F. Kozlov [2].

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Equation (10) serves as the initial expression for obtaining the integral auxiliary function -- the level surface. Using the equation of hydrostatics and retaining terms with a secondary β-effect and the secondary bottom relief, for the adynamic correction

$$\zeta_1 = \zeta + \frac{1}{\rho_0}\int_0^H \rho\, dz$$

we obtain:

$$\frac{1}{2\alpha}\Lambda\zeta_1 + A\frac{\partial\zeta_1}{\partial x} + B\frac{\partial\zeta_1}{\partial y} = \Sigma_1 + \Sigma_2, \tag{11}$$

where

$$\Lambda \equiv F_1\frac{\partial^2}{\partial x^2} + F_2\frac{\partial^2}{\partial x\,\partial y} + F_3\frac{\partial^2}{\partial y^2},$$

$$F_1 = 1 + \frac{A_l}{\nu}(E^2 + 2C), \quad F_2 = 1 + \frac{A_l}{\nu}(F^2 + 2D),$$

$$F_3 = \frac{2A_l}{\nu}(E^2 - 2C),$$

$$A = \frac{\beta H}{l} - \left[1 + \frac{A_l}{\nu}(\nabla H)^2 - \frac{1}{2\alpha}\frac{A_l}{\nu}\left(\Delta H + 2\frac{\partial E}{\partial y}\right)\right]\cdot D + \\ + \frac{1}{2\alpha}\frac{A_l}{\nu}\left(\Delta H + 2\frac{\partial E}{\partial x}\right)\cdot C,$$

$$B = \left[1 + \frac{A_l}{\nu}(\nabla H)^2 - \frac{1}{2\alpha}\frac{A_l}{\nu}\left(\Delta H + 2\frac{\partial E}{\partial x}\right)\right]C + \frac{1}{2\alpha}\frac{A_l}{\nu}\left(\Delta H - 2\frac{\partial E}{\partial y}\right)D,$$

$$\Sigma_1 = \frac{\beta}{\rho_0 l}\int_0^H z\frac{\partial\rho}{\partial x}dz + \frac{A_l}{\nu}\Delta\left(H\Delta\zeta_1 + \{H, \zeta_1\} - \frac{1}{\rho_0}\int_0^H z\,\Delta\rho\right),$$

$$\Sigma_2 = \frac{1}{2\alpha\rho_0 g}\sqrt{\frac{\nu'}{\nu}}\,\Delta P_a, \quad \{H, \zeta_1\} = C\frac{\partial\zeta_1}{\partial x} + D\frac{\partial\zeta_1}{\partial y}, \quad \alpha = \sqrt{\frac{\omega\sin\varphi}{\nu}},$$

$$C = \frac{\partial H}{\partial x}, \quad D = \frac{\partial H}{\partial y}, \quad E = \frac{\partial H}{\partial x} + \frac{\partial H}{\partial y}, \quad F = \frac{\partial H}{\partial x} - \frac{\partial H}{\partial y}.$$

If it is assumed that $A_l = 0$, then from equation (11) we derive an equation for the adynamic correction for the A. S. Sarkisyan D_1 model [3].

Numerical Model

For a numerical solution of the equation for the adynamic correction ζ_1 we use a directed differences model, frequently employed in the practice of diagnostic computations, this scheme having the first order of accuracy. As indicated by numerous computations [3], even with horizontal grid intervals greater than 40-50 km the results obtained using the directed differences scheme virtually do not differ from the results of computations using schemes with a second order of accuracy.

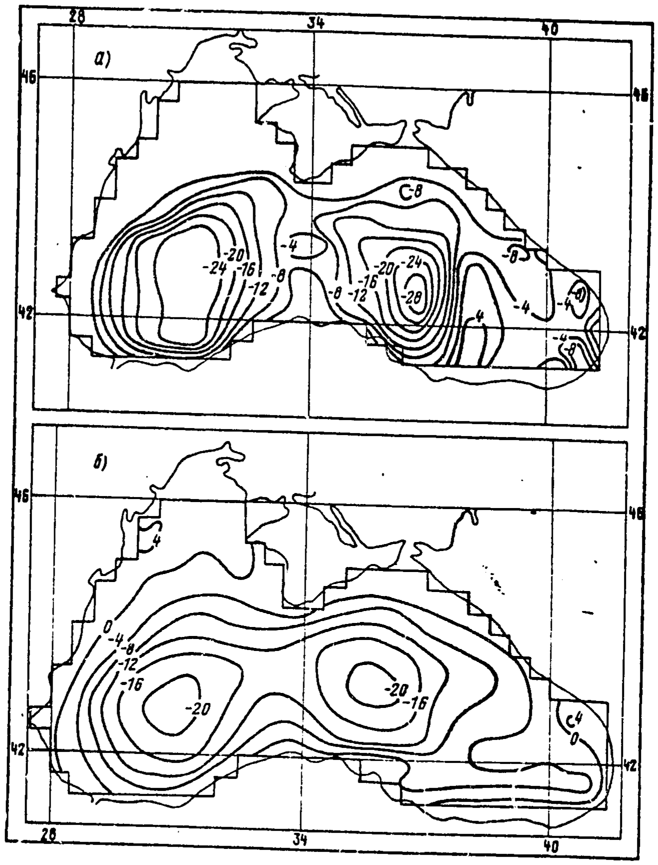

Fig. 1. Diagnostic (a) and prognostic (b) level surface fields for the Black Sea in centimeters ($A_\ell = \mu = 10^6$ cm^2/sec).

The equation for density diffusion (5) is approximated by two methods: the A. M. Il'in scheme [1] modified for three-dimensional problems and the Fiadeiro-Veronis weighted mean scheme [5]. Both schemes come close to the central differences scheme (approximation error $O(h^2+\tau)$), when the arguments of the hyperbolic cotangents participating in the schemes are far less than unity, that is, in cases of rigorous diffusion. With values of the arguments greater than unity the schemes are transformed into a directed differences scheme. In general, the A. M. Il'in scheme and the weighted mean scheme have the first order of accuracy, since the time derivative is replaced by one-sided finite difference ratios. Preference can be given to the weighted scheme, which has a greater computation stability with different values of the input parameters, that is, the range of variation of the values of the coefficients of turbulent exchange for this scheme is broader.

After the corresponding transformations the difference analogue of the density diffusion equation (5) can be written in the following form, convenient for iterations:

$$\rho_{ijk}^{n+1} = [\rho_{ijk}^{n} + \tau\,(a_{ijk}^{n}\rho_{i+1,jk}^{n+1} + b_{ijk}^{n}\rho_{i-1,jk}^{n+1} + c_{ijk}^{n}\rho_{i,j+1,k}^{n+1} + d_{ijk}^{n}\rho_{i,j-1,k}^{n+1} + e_{ijk}^{n}\rho_{ij,k+1}^{n+1} + f_{ijk}^{n}\rho_{ijk-1}^{n+1}]/(1 + g_{ijk}^{n}\tau), \quad (12)$$

where

$$a_{ijk}^{n} = \frac{u_{ijk}^{n}}{2h}\left(\mathrm{cth}\,\frac{u_{ijk}^{n}h}{2\mu} - 1\right), \quad b_{ijk}^{n} = \frac{u_{ijk}^{n}}{2h}\left(\mathrm{cth}\,\frac{u_{ijk}^{n}h}{2\mu} + 1\right),$$

$$c_{ijk}^{n} = \frac{v_{ijk}^{n}}{2h}\left(\mathrm{cth}\,\frac{v_{ijk}^{n}h}{2\mu} - 1\right), \quad d_{ijk}^{n} = \frac{v_{ijk}^{n}}{2h}\left(\mathrm{cth}\,\frac{v_{ijk}^{n}h}{2\mu} + 1\right),$$

$$e_{ijk}^{n} - \frac{w_{ijk}^{n}}{2\Delta z_k}\left(\mathrm{cth}\,\frac{w_{ijk}^{n}\Delta z_k}{2\varkappa} - 1\right), \quad f_{ijk}^{n} = \frac{w_{ijk}^{n}}{2\Delta z_k}\left(\mathrm{cth}\,\frac{w_{ijk}^{n}\Delta z_k}{2\varkappa} + 1\right),$$

$$g_{ijk}^{n} = -(a_{ijk}^{n} + b_{ijk}^{n} + c_{ijk}^{n} + d_{ijk}^{n} + e_{ijk}^{n} + f_{ijk}^{n}), \quad \Delta z_k = z_{k+1} - z_k \quad (13)$$

for the A. M. Il'in scheme.

The coefficients written above also have a similar form for the weighted mean scheme with the single difference that the velocity values with whole subscripts are replaced by velocities with fractional subscripts relating to the points between the principal points of intersection of the spatial grid

($u_{ijk}^{n} \sim u_{i+1/2,jk}^{n}$ in a_{ijk}^{n}, $u_{ijk}^{n} \sim u_{i-1/2,jk}^{n}$ in b_{ijk}^{n}, $v_{ijk}^{n} \sim v_{i+1/2,jk}^{n}$ in c_{ijk}^{n}, $v_{ijk}^{n} \sim v_{i-1/2,jk}^{n}$ in d_{ijk}^{n}, etc.)

Discussion of Results

In the computations we used the following values of the input parameters: $\rho_0 = 1.018$ g/cm^3, $\mu = 10^5$–10^7 cm^2/sec, $A_l = 10^5$–10^7 cm^2/sec, τ (time interval) $= 2.6 \cdot 10^6$ sec.

Specific computations were made for the Black Sea at 15 horizons: 0, 10, 25, 50, 100, 150, 200, 300, 400, 500, 600, 800, 1000, 1500 and 2000 m. The horizontal grid interval is h = 0.5°. Atmospheric pressure and the density field corresponded to the averaged long-term data for winter. The zonal density field ρ (y, z) was stipulated at the boundary and as the initial field.

Now we will proceed to a discussion of the computation results. Figure 1a shows the level surface, computed on the basis of the stipulated density field, taking into account the coefficient of horizontal turbulent mixing

(A_l = 10^6 cm^2/sec). The cyclonic rings in the western and eastern parts of the sea, known from observations, are clearly visible.

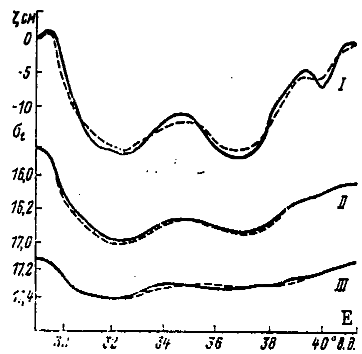

Fig. 2. Zonal section along parallel 43°N for level surface (curve I) and density σ_t at horizons 100 and 300 m (curves II, III), obtained using A. M. Il'in scheme (dashed curve) and weighted mean scheme (solid curves).

A comparison of the determined level surface field with the results of earlier computations [4], without allowance for the lateral exchange coefficient (A_l = 0), definitely indicates that the discrepancy between them is observed for the most part in the coastal regions. In the central regions of the sea the differences between the compared results were small (approximately 5-7%).

The level surface field obtained from the computed density field (Fig. 1b) has a somewhat smoothed character. We will call it the prognostic field. It follows from an analysis of the figure that the prognostic computations make it possible to determine the principal peculiarities characteristic of the level surface of the Black Sea obtained by diagnostic methods. The smoothed prognostic field of the level surface leads to a decrease in the values of the horizontal current velocity components (35-40 cm/sec in the coastal regions versus 40-50 cm/sec in diagnostic computations).

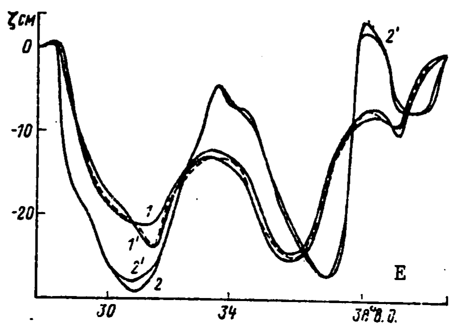

Fig. 3. Zonal section along parallel 43°N of level surface for different values of coefficients A_l and μ. 1) $A_l = \mu = 10^6$ cm^2/sec, prediction; 2) $A_l = 10^6$ cm^2/sec, diagnosis; 1') $A_l = 0$, $\mu = 10^6$ cm^2/sec, prediction, 2') $A_l = 0$, diagnosis.

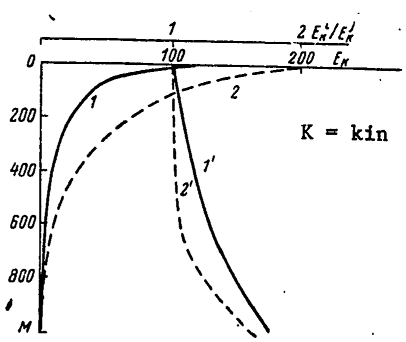

Fig. 4. Mean (for each horizon) kinetic energy E_k (1, 2) and ratio of kinetic energies E_k^i/E_k^j (1', 2'). 1) $A_l = \mu = 10^6$ cm^2/sec, prediction; 2) $A_l = 10^6$ cm^2/sec, diagnosis; 1') $E_k^i - A_l = 0$, $\mu = 10^6$ cm^2/sec, prediction; $E_k^j - A_l = \mu = 10^6$ cm^2/sec; 2') $E_k^i - A_l = 0$, diagnosis; $E_k^j - A_l = 10^6$ cm^2/sec, diagnosis.

As we have already mentioned, the A. M. Il'in scheme and the weighted mean scheme are close to one another, as is confirmed by computations made using both schemes. In particular, this is indicated by Fig. 2, which shows the zonal section of the level surface along the parallel 43°N (curve I) and the density field (curves II and III) at the horizons 100 and 300 m respectively. The curves were obtained using the A. M. Il'in scheme (dashed curve). It follows from numerical experiments that the A. M. Il'in scheme becomes unstable with values of the coefficients of turbulent exchange greater than 10^7 cm^2/sec. With respect to the weighted mean scheme, it ensures stability even with a value of the coefficient of turbulent viscosity $\mu = 10^5$ cm^2/sec. Now we will discuss the results of computations obtained using the weighted mean scheme proposed by Fiadeiro and Veronis.

Prognostic computations have indicated that with a decrease in μ the cyclonic centers, filled with dense water masses, are broadened and deepened. With an increase in the coefficient of turbulent exchange the prognostic fields become smoother, that is, physical viscosity dominates over the remaining effects. By comparing the zonal sections of the computed density fields at different horizons it can be demonstrated that the convexity of the isopycnic line decreases appreciably with an increase in μ. It is easy to see zones of upwelling of deep waters coinciding with the centers of the principal cyclonic rings.

It should be noted that the variation of the coefficient of lateral mixing exerts a greater influence on the results of prognostic computations. In order to confirm what we have said, we will examine Fig. 3 (section along the parallel 43°N), in which we have presented curves corresponding to the level surface isolines for different values of the coefficients A_l and μ. The deviation of the 1' curve (no allowance for the coefficient of horizontal exchange A_l; prognostic computations) from curve 1 ($A_l = 10^6$ cm^2/sec; prognostic computations) is much greater than the deviations of the 2' curves (A_l not taken into account; diagnostic computations) and the 2 curves ($A_l = 10^6$ cm^2/sec; diagnostic computations) from one another. In analyzing the figure we note that the dashed curve corresponding to the case of neglecting of the turbulent-topogenic effect (in the vorticity equation only the emphasized term is taken into account), in places of the maximum deviation of the 1 and 1' curves is held closer to the 1' curve. This fact indicates that the nature of the deviations to a high degree is related to the joint influence of relief and turbulence on the dynamics of currents.

Model computations indicated that a further increase in the coefficient of horizontal turbulent exchange A_l leads to an increase in the role of the turbulent-topogenic factor. But, as is easy to see, the value $A_l = 10^6$ cm^2/sec is the most reasonable for the Black Sea from the physical point of view. In actuality, using the Joseph-Zendner formula we have

$$A_l = \frac{Ph}{2} \approx 10^6 \text{ cm}^2/\text{sec},$$

where P = 1.5 cm/sec is the mean "diffusion rate."

As an illustration of the influence of the integral effect of lateral exchange on dynamics of the sea we computed the mean (for each horizon) kinetic energy E_k and the ratio of the kinetic energies E_k^i/E_k^j (the superscript i corresponds to the case $A_l = 0$, the superscript j corresponds to allowance for A_l) for different values of the coefficients A_l and μ (Fig. 4). It can be seen that the kinetic energy in prognostic computations (curve 1) attenuates more rapidly with depth than in diagnostic computations (curve 2). The ratio of energies E_k^i/E_k^j increases (curves 1', 2'), which indicates that the turbulent effects are most significant in the lower layers of the sea. In general, an increase in the coefficient of turbulent mixing leads to a general decrease in the energy of mean motion for the entire sea, which is evidently related to the assumption of a dissipative nature of turbulent exchange.

The model (mathematical) time for stabilization of the process of reckoning of the prognostic problem in dependence on the variant varied in the range 7-9 months. All the computations were made on an IBM-370 computer at the Central Institute of Scientific and Technical Information (Sofia, Bulgaria).

In conclusion the authors express sincere appreciation to A. S. Sarkisyan for attention to the work and a number of valuable comments.

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AERIAL GAMMA SURVEY OF THE SNOW COVER IN MOUNTAINS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 77-83

[Article by Candidate of Physical and Mathematical Sciences M. V. Nikiforov, Institute of Experimental Meteorology, submitted for publication 4 September 1978]

Abstract: A study was made of the potential possibility of broadening the range of snow reserves measured in a gamma survey by means of carrying out surveys over terrain with a great nonuniformity in depth of the snow cover and the use of a priori information on the shape of the distribution curve for the snow reserves over an area. In the case of a gamma distribution it is shown that the mean snow reserves for a territory accessible to measurements with an accuracy of 10-15% can attain 2000-5000 mm of water equivalent with values of the variation coefficient 0.7-1.0. The practical use of the considered survey procedure is discussed.

[Text] A gamma survey of the snow cover is based on the effect of the attenuation of the gamma radiation of natural radioactive elements present in the soil by snow [3-5]. The existing method for aircraft snow surveys [5, 6] is based on a physical-mathematical model of a homogeneous emitting-absorbing half-space covered by a uniform layer of an inactive absorber -- snow. The dependence of the intensity of gamma radiation of soils (I) at the height h above the snow is expressed well by the formula

$$I = \frac{I_{00}}{1 + \gamma w} \exp[-\alpha (S + h)], \quad (1)$$

where w is the moisture content of the soil by weight, S is the water reserve in the snow cover, I_{00} is the I value with S = h = w = 0; γ is a coefficient taking into account the difference in the absorbing properties of the soil (rock) skeleton and water (γ = 1.11); α is the coefficient of attenuation of gamma radiation by water, dependent on the energy of the measured radiation.

The height h in formula (1) is expressed in units of thickness of a water layer equivalent in absorbing capacity to the air layer between the earth's surface and the observation point. In Soviet-produced apparatus the detector of gamma radiation is a scintillation counter with a monocrystal of sodium iodide with a diameter of 150 mm and a height of 100 mm. The total intensity of gamma radiation is measured in the energy range from 0.02 to 3.0 MeV. In this case the coefficient α is 0.0065 mm^{-1} and is not dependent on the composition of the natural radioactive elements in the soils (rocks).

The water reserves in the snow cover are computed using the results of two measurements, the first of which is carried out before the falling of snow and the second at the time of a snow survey. The changes in soil moisture content during the time between measurements are not taken into account.

$$S = \frac{1}{\alpha} \ln \frac{I_1}{I_2} + h_1 - h_2, \qquad (2)$$

where the subscripts 1 and 2 apply to the first and second measurements respectively.

The fundamental limitation on the magnitude of the snow reserves which can be measured with a stipulated accuracy is governed by the statistical nature of radioactive decay and the registry of nuclear radiations. In actual practice this limit is not achieved due to instrumental-methodological errors and deviations of actual survey conditions from the initial physical-mathematical model.

On the lowlands measurements from an aircraft are made from an altitude of 50-100 m. The depth (height) is measured by an aircraft radioaltimeter with an accuracy of 2-3 m, which leads to errors in determining snow reserves of 2-3 mm. The total error (taking into account all the interfering factors) does not exceed 10 mm if the measured snow reserves fall in the range 300-400 mm.

In the mountains flights with a low altitude over the terrain are possible only in a helicopter over relatively "even" sectors of macro- and meso-relief: along slopes, over terraces, valleys, plateaus, flattish water divides, etc. For such sectors the gamma field created in the surface layer of the atmosphere by natural radioactive soil (rock) elements can be considered at the flight altitude to coincide with the field of an infinite emitting-absorbing half-space. We will limit ourselves to an examination of only such surveys and the only difference in our physical-mathematical model from a "lowland" survey is in allowance for the nonuniformity of snow depth in the terrain.

It is obvious that the gamma radiation of soils in the segment of the route where the liquid water content of the snow is in the interval from S to S + dS will be attenuated by snow by a factor of e^S. If F(S) is the distribution function for snow reserves along the route the contribution of the radiation of all such segments of the route to the mean intensity over it at the height h will be

$$dI = I_0 \exp(-\alpha h - \alpha S) F(S) dS \tag{3}$$

and the intensity averaged along the entire route will be

$$I = I_0 \exp(-\alpha h) \int_0^\infty F(S) \exp(-\alpha S)\, dS;$$
$$I_0 = \frac{I_{\text{гп}}}{1 + \gamma w}. \tag{4}$$

It can be shown that with a nonuniform depth of the snow and any form of the distribution function F(S) the gamma radiation of soils (rocks) is attenuated by the snow on the average over the area (along the route) to a lesser degree than in the case of its uniform depth and accordingly, the intensity of radiation over the nonuniformly deposited snow can be measured more precisely. Therefore, there is basis for assuming that under definite conditions the upper boundary of the snow reserves measurable by a gamma survey in the case of a nonuniform depth of the snow can be greater than is the case on the plain.

Now in expression (4) we will proceed from the distribution function for snow reserves F(S) to the distribution function for the modular coefficient of snow reserves P(k) ($k = S/\overline{S}$, where $\overline{S}$ is the mean snow reserve along the route). Taking into account that F(S)dS = P(k)dk, we obtain

$$I = I_0 \exp(-\alpha h) \int_0^\infty P(k) \exp(-\alpha \overline{S} k)\, dk. \tag{5}$$

The solution (5) for $\overline{S}$ gives a "working formula" for computing the mean water reserve in the snow cover along the route through the results of measurements of the intensity of gamma radiation. It is understandable that the form of the working formula is dependent on the shape of the distribution curve of the modular coefficient of snow reserves. Accordingly, in order to carry out aerial gamma surveys of the snow cover over terrain with nonuniformly deposited snow it is necessary to have a priori information on the shape of the distribution curve.

Now we will consider a specific example. We will assume that the shape of the distribution curve for snow reserves is known in advance and coincides with the gamma distribution [See Note], that is

$$P(k) = \frac{\beta^\beta}{\Gamma(\beta)} k^{\beta-1} \exp(-\beta k), \tag{6}$$

where $\beta = 1/C_v^2$, C_v is the coefficient of variations of snow reserves; $\Gamma(\beta)$ is the gamma function.

[Note] For the open terrain of a plain such a coincidence is a firmly established fact [7-10]. In the sectors of mountainous terrain which we considered the mechanism of formation of nonuniformities of the snow cover is similar to that in the lowlands: this is a redistribution of the snow by the wind in microrelief details. But the wind in the mountains is stronger and the terrain is more dissected. Therefore the variability of the snow is greater.]

Substituting (6) into (5) and carrying out integration, we obtain

$$I = I_0 \frac{\exp(-\alpha h)}{(\alpha \bar{S} c_v^2 + 1)^{\frac{1}{c_v^2}}}. \tag{7}$$

The formula for computing the mean snow reserve $\bar{S}$ along the route in this case will have the form

$$\bar{S} = \frac{1}{\alpha C_v^2}\left[\left(\frac{I_1}{I_2}\right)^{C_v^2} \exp\{\alpha C_v^2 (h_1 - h_2)\} - 1\right], \tag{8}$$

where the subscripts 1 and 2 have the same value as in (2).

Since we consider the coefficients α and C_v to be precisely known, the expression for the relative mean square error in the snow surveys, according to the theory of errors, will have the form

$$\delta_{\bar{S}} = \frac{1}{\alpha \bar{S}} (\alpha \bar{S} C_v^2 + 1) \sqrt{\delta_{I_1}^2 + \delta_{I_2}^2 + 2\alpha^2 \sigma_h^2}, \tag{9}$$

where δ_{I_1}, δ_{I_2} are the relative mean square errors in measuring I_1 and I_2, σ_h is the error in measuring height for a single flight, whose value is not dependent on the snow reserves, the intensity of gamma radiation and under the considered conditions is the same as in the plains, 2-3 mm water equivalent.

Under real conditions of aerial snow surveys the dispersion of the measurement for the intensity of gamma radiation for the most part consists of three terms, each of which in its own way is dependent on the measured intensity, and accordingly, on the snow reserves:

$$\delta_I^2 = \delta_0^2 + \delta_{\text{ст}}^2 + \delta_{\Phi}^2, \tag{10}$$

where δ_0 = const is a value not dependent on the level of the gamma radiation. The δ_0 value arises due to fluctuations of instrument response, flight velocity, deviations from the route and is 2-3% [6]; δ_{st} is the error caused by the statistical nature of the radioactive decay and the interaction of gamma quanta with matter. The δ_{st} value is dependent on the measured intensity I, the noise level I_{back} and the measurement time

$$\delta_{\text{ст}} = \frac{1}{I}\sqrt{\frac{I + I_\Phi}{t} + \frac{I_\Phi}{t_\Phi}}, \tag{11}$$

where t is measurement time over the route; t_{back} is the time for measuring I_{back}. (Over the route the useful radiation is measured as a total with the interfering radiation, and the interfering radiation is measured separately -- either over the water body or at an adequate distance from the earth. As a result, the intensity of the useful radiation is obtained as the difference in the results of these two measurements). Finally, the last term on the right-hand side of expression (10) -- δ^2_{back} -- is governed by methodological errors in determining the level of interfering radiations I_{back}:

$$\delta_{\Phi} = \frac{\sigma_{\Phi}}{I}, \tag{12}$$

[Φ = back]

where σ_{back} is dependent on I. As a result we have

$$\delta^2_I = \delta^2_0 + \frac{1}{It} + \frac{I_{\Phi}}{I^2}\left(\frac{t + t_{\Phi}}{tt_{\Phi}}\right) + \frac{\sigma^2_{\Phi}}{I^2}. \tag{13}$$

Table 1

Relative Mean Square Errors in Snow Surveys (in Percent) for Different Degrees of Nonuniformity of Snow Depth

C_v, %	1 Средний снегозапас, мм								
	200	500	700	1000	1500	2000	3000	4000	6000
0	3,3	2.6	7,0	34	>100	>100	>100		
25	3,6	2,8	4,9	20	>100	>100	>100		
50	4,3	3,1	3,9	6,5	16,6	45	>100	>100	>100
70	5,3	4,0	4,0	5,0	7,4	10,6	21	37	>100
100	7,3	5,8	5,6	5,6	6,2	7,0	8,6	10,4	14,4

KEY:

1) Mean snow reserve, mm

It is clear from (13) that the values of the second and third terms can be regulated by a change in the observation time t and t_{back}, whereas the value of the last (σ^2_{back}/I) is not subject to such regulation. Its decrease is possible only by an increase in instrument response to useful radiation (I) and an increase in noise immunity, that is, a decrease in σ_{back}. Table 1 gives the results of estimates of the errors in snow surveys for values of the coefficient of variations of snow reserves from 0 to 1.0. The estimates were made for the case of carrying out of gamma surveys using Soviet aerial snow-measuring apparatus, characterized by the following parameters: I_0 = 1000 quanta/sec, I_{back} = 100 quanta/sec, δ_0 = 2%, σ_{back} = 3 quanta/sec, σ_h = 3 mm. The measurement times over the route and the background were 100 sec each.

Table 1 shows that with C_v values from 0 to 50% (lowland conditions) with an accuracy to 10-15% snow reserves to 800-1100 mm are measured, whereas for values C_v = 0.7 and 1.0 (mountainous conditions [1, 2]) reserves up to 2000-5000 mm are measured. It should be noted that the principal contribution to the errors in snow surveys is from errors in taking interfering radiations (σ_{back}) into account.

The cited estimates must be regarded as an illustration of the fundamental possibility of broadening the field of use of aerial gamma surveys of the snow cover since they were obtained for a case when the form of the distribution curve for the modular coefficient of snow reserves is precisely known. In actual practice it can be known only approximately, which causes additional errors not taken into account in Table 1.

Now we will estimate the error arising as a result of the inaccuracy in knowledge of the distribution curve for the modular coefficient of snow reserves. As before, we will assume that the shape of the curve coincides with the gamma distribution, but the coefficient of variations is known inexactly. We will use $\delta_{\bar{S}_{C_v}}$ to denote the relative mean square error in measuring the mean snow reserves $\bar{S}$ along the route as a result of the error in knowledge of C_v, equal to δC_v. In accordance with the theory of errors, from (8) we obtain

$$\delta_{\bar{S}_{C_v}} = \frac{2}{\alpha \bar{S} C_v^2} \{1 + (\alpha \bar{S} C_v^2 + 1)[\ln(\alpha \bar{S} C_v^2 + 1) - 1]\} \delta_{C_v}. \quad (14)$$

The value of the total error will be equal to

$$\delta_n = \sqrt{\delta_{\bar{S}}^2 + \delta_{\bar{S}_{C_v}}^2}. \quad (15)$$

It is clear from (15) that with values of the ratio $\delta_{\bar{S}_{C_v}} / \delta_{\bar{S}}$ less than unity the decisive role in the value of the total error is played by $\delta_{\bar{S}}$ and the errors in the knowledge of C_v can be neglected. When

$$\delta_{\bar{S}_{C_v}} / \delta_{\bar{S}} > 1$$

the picture is reversed. In order for the total measurement error for snow reserves along the route not to exceed 10-15%, the coefficients of variations must be known with the accuracy cited in Table 2.

Table 2 shows that:

1. The requirements on the accuracy of knowledge of C_v increase with an increase in the snow reserves.

2. The differences in the requirements on the accuracy of a knowledge of C_V for $C_V = 70\%$ and $C_V = 100\%$ are small.

Table 2

Results of Evaluations of Admissible Errors in A Priori Knowledge of the C_V Value in Gamma Surveys of the Snow Cover With a 10-15% Accuracy Using Real Soviet-Produced Instrumentation. Errors are Expressed in the Same Units as the Coefficients of Variations

C_v%	1 Средний снегозапас, мм							
	200	500	700	1000	1500	2000	3000	4000
70	9—15	4—7	4—6	3—4	2—3	1—2	—	—
100	9—17	5—8	4—6	3—5	2—4	2—4	1—3	1—2

KEY:
1. Mean snow reserve, mm

What is the real accuracy of a priori knowledge of the value of the coefficient of variations for snow reserves on aerial snow-measuring routes and what are the prospects for increasing this accuracy? The coefficient of spatial variations for snow reserves in the mountains, according to data in the literature [1, 2], does not exceed 100% and the variability of the snow reserves in the mountains is greater than in the lowland, where $C_V \geqslant 50\%$. If it is assumed that in mountainous terrain any C_V value in the range from 50 to 100% can be encountered with an equal probability, and if the mean value (75%) is used in the computations, the mean square scatter of the actual values about this mean, that is, the very error in a priori knowledge of C_V, which is given in Table 2, is 14.4%. The classification of terrain on the basis of the characteristics of variability of the snow cover evidently already at the present time makes it possible to break down the entire mentioned range of C_V changes into 2-3 sectors, each with a width of 15-20%, which reduces the errors in a priori knowledge to 4-5% respectively, and accordingly, makes it possible to determine the snow reserves to 1000-2000 mm with an accuracy to 10-15%. Further studies in the direction of classification of mountainous territories on the basis of the nature of variability of the snow cover will evidently raise the limit of the snow reserves which can be investigated by gamma surveys to 2500-3000 mm.

It must be remembered that the errors in knowledge of the C_V value can be of two types:

1) Inexact knowledge of the mean C_V value for the defined terrain class. Such errors lead to systematic understatements or exaggerations of the mean values of the snow reserves even with averaging for a large number of routes.

2) Random deviation of the C_v value on the route from the mean value as a result of the finite length of the route. Such statistical errors will be reduced with increasing length of the route or with averaging of snow reserves along a great number of routes.

Now we will discuss ways in which to realize the considered survey procedure in actual practice. Obviously, first it is necessary to investigate the curves of distribution of snow reserves for those types of mountainous territories mentioned in this article. In our opinion, the available observational data and theoretical analyses are adequate for solving this problem in the first approximation. At the same time it is necessary to improve instrumentation in the direction of an increase in its response and noise immunity. The means for bringing about this improvement is clear -- a single-crystal scintillation counter must be replaced by a multicrystal spectrometric counter and change over to the registry of the spectral line of thorium with an energy of 2.62 MeV, for which the background is only cosmic radiation -- weak and stable. The estimates of the errors for such instrumentation, similar to those cited in Table 1, show that with a 10-15% accuracy it is possible to measure the mean snow reserves to 6000-8000 mm when $C_v = 0.7$-1.0.

We feel that theoretical analyses must be carried out in two directions: in gamma field theory change over to a model which takes meso- and macrorelief into account; estimate errors in surveys resulting from statistical variations in the shape of the distribution curve for individual routes and the dependence of the value of these errors on the length of the route.

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STRUCTURE AND REFORMATION OF SAND RIDGES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 84-87

[Article by Candidate of Geographical Sciences A. A. Levashov, Leningrad Hydrometeorological Institute, submitted for publication 2 July 1979]

Abstract: The article gives the results of field investigations of the structure and reformation of ridges.

[Text] An extensive literature has been devoted to the problems involved in the origin, movement, study of parameters and classification of sand ridges of different scales [1-8]. Nevertheless, attempts at application of the results to practical computations, as noted by B. F. Snishchenko and Z. D. Kopaliani [8], convincingly demonstrate that these investigations must be continued. In the hydrological literature little attention has been given to the internal structure of the ridges and the mechanism of their reformation. A true idea concerning this structure can be obtained with the development of field observations. Data from numerous laboratory experiments cannot reveal the full physical picture of structure of the ridges because when carrying out the experiments as a rule no allowance has been made for numerous natural factors. At the same time these exert an effect on the process of ridge formation and leave a trace in their internal structure.

During the years 1976-1978, on the Nadym and Pur Rivers, the author carried out special field observations of the structure of dry sand ridges remaining on the surfaces of exposed bars. For example, on a bar with a length of 920 m and a width of 130 m we studied 77 ridges with an average length of 12.0 m and a height of 0.24 m. In order to investigate the internal structure of the ridges for several of them (Figures 1, 2) we obtained sections (small ditches were dug) from the base of the preceding ridge to the crest of the next. The depth of the ditch reached the surface of the bar existing prior to high water and in a number of cases considerably below this surface. The following was observed in the sections of ridges not subjected to wind erosion. The ridges formed by the last high water differed greatly from the lower layers of the bar in the stratification of deposits. The inclination of the layers of the downstream slope of the ridge (steepness of dropoff), according to numerous measurements (Figures 2, 3), is 36°, which is equal to the natural slope of sand. This is evidence that the sand particles tumble from the crest into the trough under the influence of their own weight. At the base of the ridges it is easy to trace a dark, relatively horizontal band (Fig. 1, 2) with a thickness of 1-1.5 cm. It was represented by silty particles and fine woody residue usually observed in the troughs among ridges which are buried by a creeping ridge. The band passes through the entire ridge, that is, from the bottom of one to the bottom of the other.

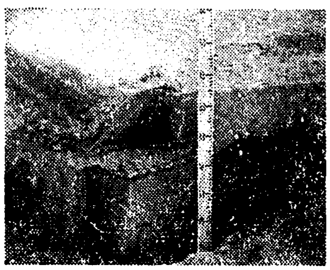

Fig. 1. In the longitudinal and transverse sections of the ridge below the upper stratified layer it is possible to see wind-reworked annual layers formed by the preceding high waters. They are separated by dark horizontal intercalations with a thickness 1.0-1.5 cm.

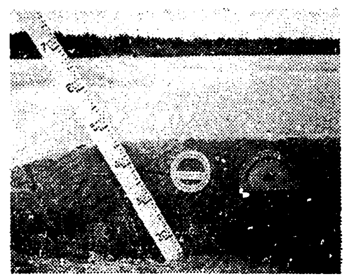

Fig. 2. The inclined bands in a longitudinal section of a ridge form an angle of 36° with the horizontal base (dark band).

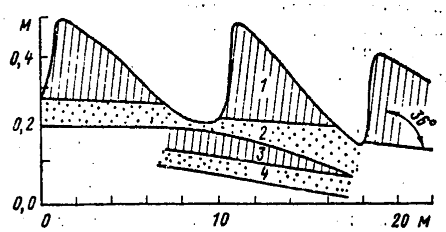

Fig. 3. Longitudinal section of sand ridges on bar in Pur River, 185 km from mouth. 1, 3) stratified deposits. Angle of inclination of layers with surface of below-lying nonstratified strata 2, 4 -- 36°.

Below the darkish silty band (Figures 1, 3) there are strata of unstratified light-colored deposits also separated by distinct dark silty bands. The lower-lying strata along their length have an approximately identical thickness and are not similar in form to ridges. This is attributable to the fact that the ridges (a surface with a sawtooth profile) formed by the preceding high waters during the period of drying-out are deflated and leveled by the wind. The thickness of these layers quite reliably makes it possible to decipher the annual increment of bars and mid-channel islands with respect to height.

On the basis of the detected strata it was possible to determine the annual increment of the investigated bar in the Pur River (with respect to its height); during the years 1975-1978 its increments were 8, 6, 8, 10 cm respectively.

A wearing-away of the surface was observed on some bars. During the high water of 1978 it constituted about 12 cm. The measurement was made using observations of 3 1/2-m metal rods driven into the bar and leveled in the autumn of 1977. Some strata of 1976 also retained a stratification which indicates their incomplete reworking by the wind. It is impossible to confuse this layer with the double creeping of ridges in 1978 because there is a wind-reworked unstratified stratum of 1977 between the layered strata of 1978 and 1976 (Fig. 3).

Now we will consider in greater detail the upper layered stratum of the last high water not subjected to wind reworking and represented by a ridge with a height of 0.24 m and a length of 12 m. The ridge retained traces of its formation. In its section oblique intercalations were represented by light- and dark-colored bands. The thickness of the lighter-colored sand bands (1.5 cm) was 3-4 times greater than the darkish bands. The presence of the intercalations is evidence of the discreteness of their formation. The slipping downward of the sand particles toward the bottom of the slope does not occur continuously, but periodically. Smaller formations -- ripples with a mean height of 0.02 m and a length of 0.20 m -- move on the surface of the investigated ridge.

The dimensions of the ripples along the length of the "pressure" slope of a 12-m ridge were not all the same: near the bottom of the slope they were approximately two times greater than at the crest [6]. The slippage of the ripples along the rear slope also is responsible for the banded oblique stratification of the ridge. However, simple calculations show that the area of one intercalation is 3-4 times greater than the area of the longitudinal profile of one average ripple. It must be postulated that there is washing away of not one, but several ripples.

With the leveling of the ridge crest it was found that in a sector with a length of about 1/15-th part of the ridge it is relatively horizontal and ripples are virtually absent here. The washing-away of sand particles from this sector of the ridge, equal in length to four lengths of the average ripples, occurs in rippleless form. This sector resembles the steps of a subway escalator, near whose "exit" 3-4 steps are rapidly transformed into a treadmill. A rippleless sector of approximately the same length is observed at the foot of the ridge. It differs from the rippleless crest sector by having some concavity and it can be compared with the "entrance" sector of the escalator.

A fluid eddy is localized at times behind the rear slope. The ascending jets of the rotating eddy catch up the silty particles downslope and they are thrust onto the rear slope. This forms the dark band. This band and part of the silty deposits at the base, adjacent to the slope, is filled with sand particles descending from the crest, forming a light-colored band. An increment of the ridge from the direction of the rear slope takes place by the "shuttle" principle -- the casting of silt particles onto the slope is replaced by their "sprinkling" with coarser fractions. The washing-away of sand from the crest is accomplished by converging and degenerating eddies forming in the preceding "trough" or the transient part of the flow if the eddies degenerate prior to approach to the crest. The casting of the silt particles onto the slope is accomplished by an eddy newly forming in the trough. The bottom layers of the flow over the crest, upon departure from it, form a sort of loop, forming a rotating eddy.

The formation of dark oblique bands of silty deposits in the trough is also demonstrated by the fact that in the bottom sediments, according to data from numerous samples, silty particles constitute only about 3% of the total weight of sediment samples entering a bathometer. However, their percentage in the body of the ridge, determined from the dimensions of the bands (Fig. 1), is 25-30%. This mechanism creates the distinct banded pattern in the body of the ridge detected in the sections.

The eddy generated by the crest and trough cannot remain in one place and must be transported ("rolled") by the oncoming transient part of the flow due to the vertical velocity gradient. A new eddy is formed (generated) in its place. The crest and trough of the ridge constitute a "factory" where the eddies are destroyed (degenerate) and are reformed. The apparent localization of the eddy [2] in the trough is not confirmed by observations. At the water surface it is easy to observe emerging eddies in the form of bulging and outspreading formations. The transient upper part of the flow, ensuring rotational movement of the eddies forming in the trough, itself moves along it, like on hydraulic rollers. The dimensions of the eddies and the ridge do not remain constant with time and are dependent on the heights of the high-water waves. The presence of eddies with a relatively horizontal axis of rotation behind the ridge crest, like behind any obstacle generating them in the presence of a current, is unquestionable. They are observed under both laboratory and natural conditions.

The dimensions of the eddy, with the ridges existing in the flow, is determined by their height. We will attempt to apply this hypothesis to practical computations. With a height of the investigated ridge 0.24 m and a length 12.0 m about 50 eddies with a mean diameter of 0.24 m would fit into its extent. We note in passing that in a detailed investigation of the ridge surface about 60 ripples with a mean length of about 0.20 m and a height of 0.02 m were discovered on it. Since the dimensions of the ripples decrease toward the crest [6], the eddy following its path from the trough to the crest is obviously deformed.

If it is postulated that the eddy moves at a velocity close to the bottom velocity of the flow (in this case 0.4 m/sec was used for the high water), it overcomes a ridge 12 m in length in 30 sec. The departure of the eddy from the crest evidently causes the formation of the observed oblique intercalation. For the investigated conditions the intercalation was 1.5-2.0 cm. With such a rate of formation of the intercalation the investigated ridge travels about 50 m in 24 hours. The result agrees with computed data on the rate of movement of ridges in this range on the Nadym River [6]. Computations of the rates of movement of ridges on the Nadym River were carried out by another method, with the use of the morphometric formula and the magnitude of discharge of bottom sediments.

When the bar is covered by a layer of water of a meter or more in about 20 days (the determination of this duration was made using a standard graph of water level fluctuations) ridges with the mentioned dimensions and with a computed mean rate of movement of 50 m/day cantravel a distance of about 1000 m at the time of high water. During this time 83 12-m ridges should tumble into the bar trough, resulting in its increment by 72 m. The average increment of bars, determined from surveys made at different times on the Nadym and Pur Rivers, is 60-80 m/year, in individual years attaining 150-200 m.

Thus, small ripples, slipping into the trough of a ridge of a higher, second order, form its elementary increment in the form of a band. This increment, causing movement of a second-order ridge, leads to its tumbling into the trough of a mesoform (ridges of a still higher, third order), causing its increment or movement downstream.

During formation of a mesoform there is also a lateral increment due to the ridges slipping along the channel and transverse circulation, but this is not considered in this paper.

This investigation of the structure of ridges will help in judging not only the mechanism of their formation and rate of movement, making it possible to predict bottom deformations and compute the discharge of sediments, and also the structure of turbulent flow in the bottom region over the ridges.

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INFLUENCE OF AGROMETEOROLOGICAL CONDITIONS ON THE WINTERING OF PERENNIAL LEGUMINOUS GRASSES IN THE CENTRAL REGIONS OF THE EUROPEAN USSR

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 88-94

[Article by A. I. Strashnaya, USSR Hydrometeorological Scientific Research Center, submitted for publication 19 September 1979]

Abstract: A study was made of the influence of agrometeorological conditions of the autumn-winter period on the formation of frost-resistance of leguminous grasses. The equations proposed here make it possible, under the conditions prevailing in the central oblasts of the European territory of the USSR, to calculate the critical temperature of winter killing of clover, the minimum temperature of the soil at the depth of 3 cm in fields with sown grasses and to predict the state of perennial sown grasses in spring both in a specific field and over the territory of an oblast.

[Text] The basis for a stable fodder base for livestock in the central oblasts of the European territory of the USSR is perennial leguminous grasses. During winter the main reason for damage to leguminous grasses is winter killing [2, 7, 10]. Due to the death of grasses and the lower productivity of damaged sown areas the gross yield of grasses during years with severe wintering conditions is reduced. Accordingly, methods for a quantitative evaluation of the conditions for the wintering of perennial grasses and prediction of their condition in spring are assuming timely importance.

The methods for quantitative evaluation of the conditions for the wintering of winter grain crops have been developed and are being successfully used at the USSR Hydrometeorological Center in the preparation of predictions of their state in spring [4]. No one has derived quantitative dependences making it possible, on the basis of agrometeorological conditions, to calculate the anticipated state of perennial sown grasses in spring.

At the present time agrometeorological observations of the wintering of perennial sown grasses are carried out by a limited number of hydrometeorological stations, which does not always make it possible to evaluate the conditions for their wintering.

Investigations which have been made [2, 10] have demonstrated the possibility of using observational data on the wintering of winter grain crops for evaluating the conditions for the wintering of sown grasses. Equations have also been proposed for computing the minimum soil temperature at the depth of the root neck of grasses in dependence on the minimum temperature of the air [2]. However, they were derived for only three gradations of depths of the snow cover and depths of freezing of the soil more than 30 cm.

Our investigations indicated that the principal elements of agrometeorological conditions, determining the cooling of the soil at a depth of 3 cm in fields with perennial grasses, as in the case of winter grain crops [4, 12], are minimum air temperature (η = 0.55-0.61), depth of snow cover (η = 0.58-0.69) and depth of soil freezing (η = 0.61-0.68). The snow cover considerably weakens the closeness of the relationship between the minimum air temperature and soil at a depth of 3 cm, especially in the nonchernozem zone, where its depth as a rule is greater than in the chernozem zone. In the absence of a snow cover the closeness of the correlation between these indices increases considerably (η = 0.86-0.93).

The influence of density of stand and the height of plants during the autumn period on soil temperature in fields with perennial sown grasses is poorly expressed (η = 0.16-0.41). This is attributable to the great diversity of methods, times for the sowing and harvesting of grasses, frequent and nonuniform grazing of sown areas in autumn.

The derived prognostic equations of the dependence of minimum soil temperature at the depth of the root neck of grasses on minimum air temperature and depth of the snow cover have the following form.

For the central nonchernozem oblasts:

$$t_3 = 0{,}534\,t - 0{,}018\,H + 0{,}129\,h - 0{,}005\,h^2 - 0{,}012\,th + 1{,}57;$$
$$R = 0{,}91; \quad S_{t_3} = \pm 1{,}4; \quad n = 140.$$

For the central chernozem oblasts:

$$t_3 = 0{,}588\,t - 0{,}026\,H + 0{,}169\,h - 0{,}004\,h^2 - 0{,}010\,th + 3{,}09;$$
$$R = 0{,}93; \quad S_{t_3} = \pm 1{,}6; \quad n = 116,$$

Here t_{depth} (t_3) is the minimum soil temperature at the depth of the root neck of grasses (°C); t is the minimum air temperature (°C); h is the depth of the snow cover (cm).

These correlations are represented in Fig. 1. The soil temperature at a depth of 3 cm increases sharply with an increase in the depth of the snow cover to 30 cm. With its further increase the increase in soil temperature occurs more slowly. At the same time, all other conditions being equal, the soil temperature at a depth of 3 cm in the chernozem zone is always lower than in the nonchernozem zone.

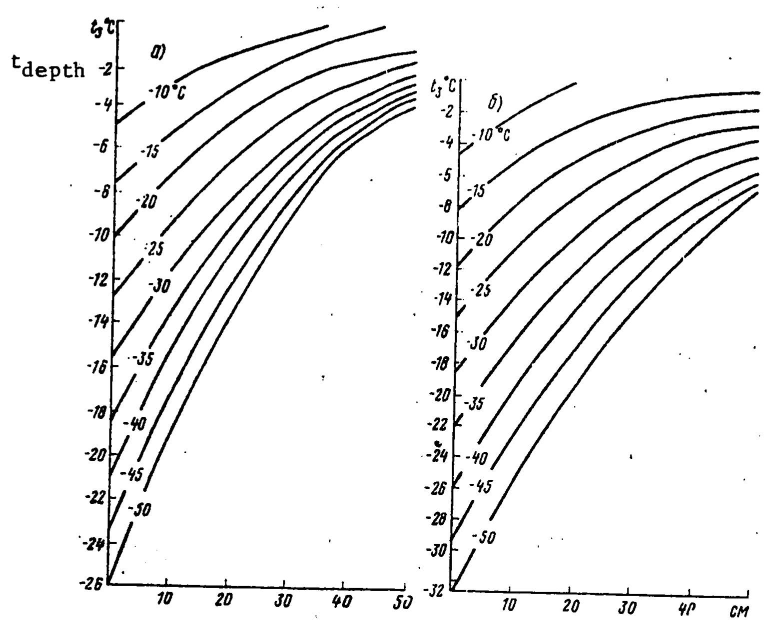

Fig. 1. Dependence of minimum soil temperature at a depth of 3 cm in fields with perennial sown grasses on minimum air temperature and depth of the snow cover. a) nonchernozem zone, b) clayey loam chernozem.

In the chernozem zone a snow cover with a depth of 10 cm does not safeguard the sown grasses against winter killing when the air temperature is -30°C because the soil temperature in this case at the depth of the root neck is reduced to -13.5, -14.0°C, which is already dangerous for clovers in years with average hardening conditions. In the nonchernozem zone, with these same conditions the soil temperature at the depth of the root neck of grasses is reduced to -11.5°C. At such a temperature the clover is usually not damaged. Prognostic equations have also been derived for the dependence of minimum soil temperature at the depth of the root neck of grasses on the minimum air temperature, depth of the snow cover and depth of soil freezing.

For the central nonchernozem oblasts:

$$t_3=0{,}484\,t+0{,}207\,h-0{,}004\,h^2-0{,}008\,th-0{,}75;$$
$$R=0{,}85;\quad S_{t_3}=\pm1{,}9;\quad n=140.$$

For the central chernozem oblasts:

$$t_3=0{,}701\,t+0{,}224\,h-0{,}005\,h^2-0{,}011\,th+2{,}41;$$
$$R=0{,}90;\quad S_{t_3}=\pm1{,}8;\quad n=116.$$

where t_{depth} is the minimum soil temperature at the depth of the root neck of grasses, °C; t is the minimum air temperature, °C; h is the depth of the snow cover, cm; H is the depth of soil freezing, cm.

The limits of application of the equations are: t -- from -10 to -45°C; h -- from 0 to 50 cm; H -- from 10 to 150 cm. The guaranteed probability of the equations is 77-90%.

Using these equations it is possible not only to compute the minimum soil temperature at the depth of the root neck of the grasses, but also to determine at what minimum air temperature and snow cover depth the minimum soil temperature will be reduced to critical values.

However, investigations of the critical temperatures of winter killing of grasses, and especially clover, occupying the leading place among the perennial leguminous grasses in the considered territory, have been extremely few in number [11] and have frequently been contradictory. The results of field and laboratory experiments of the author, carried out at the All-Union Fodder Institute imeni V. R. Vil'yams for determining the frost resistance of clover, indicated not only the influence of the age and type (species) of plants on the formation of their resistance to frosts [10], but also a considerable influence of agrometeorological conditions. The collected data, and also the results of determination of the critical temperatures of winter killing of clover by the agrometeorological stations Glukhov and Poltava, made possible a quantitative evaluation of the influence of agrometeorological factors on the critical temperature of winter killing of clover of the two-crop type during the first year of life.

Our correlation analysis indicated that there is a weak direct correlation ($r = 0.342$, $\eta = 0.489$) between the critical temperature of winter killing of clover and mean air temperature during the period of stable transition through +15°C to its transition through +5°C (t_{15-5}). This is attributable to the fact that with normal times of harvesting of the cover crop clover usually has succeeded in developing well in all years since the heat supply during this period in the zone of cultivation of two-crop clover in the considered area almost always is adequate. A closer correlation of the critical temperature of winter killing of clover exists with the precipitation total (R_{VIII-X}) during August-October ($r = -0.105$; $\eta = 0.631$) and the mean (for September-October) reserves of productive moisture (W_{IX-X}) in the cultivated soil layer ($r = -0.034$; $\eta = 0.617$).

The great difference between the correlation ratios and the correlation coefficients characterizing the relationships between the critical temperature of winter killing of clovers and the precipitation sum and the mean reserves of productive moisture indicates a nonlinear nature of the correlation. The nonlinearity of the correlation can be judged from the curvilinearity test t_k [5].

The t_k value, equal to 2.0 (for P = 0.01) or a large number, indicates a nonlinearity of the correlation. In our case the correlation between the critical temperature of winter killing of clover and the precipitation sum for August-October and the mean reserves of productive moisture in the cultivated soil layer for September-October is nonlinear; t_k is equal to 2.1 and 2.8 respectively.

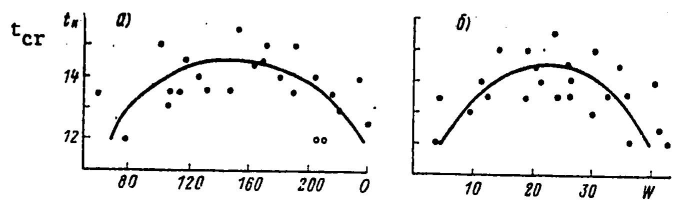

Fig. 2. Dependence of critical temperature of winter killing of two-crop clover on precipitation sum for August-October (a) and mean reserves of productive moisture for September-October in soil layer 0-20 cm (b).

A high frost resistance of the clover plant develops with a quantity of precipitation during this period from 100 to 200 mm; a decrease or increase in the precipitation sum from these limits usually leads to a decrease in the frost resistance of clover (Fig. 2a).

There is also a decrease in the frost resistance of plants when the mean reserves of productive moisture in the cultivated soil layer are less than 15 mm and more than 30 mm and the most favorable conditions are created with mean moisture reserves of 20-25 mm (Fig. 2b).

It is pointed out in [1] that the moisture shortage in the soil exerts an influence on the viscosity of plants. With an increase in the viscosity of the protoplasm there is an increase in the resistance to heat and a decrease in the tolerance of plants to the cold. At the same time, an over-moistening of the soil in autumn is also dangerous because it favors the winter killing of plants [6]. When there is optimum moistening of the soil [13] prior to cessation of the growing season, clovers form more leaves and the bushiness factor for them is greater and well-developed plants are more frost-resistant.

A factor of great importance in the formation of frost resistance of plants is the temperature conditions for their hardening [8, 9]. The considerable quantity of sugars in autumn in winter crops is accumulated in winter crops because on warm sunny days photosynthesis processes transpire intensively in the plants, whereas in the evening, nighttime and morning hours with a decrease in temperature their respiration and growth processes are slowed, which leads to an accumulation of sugars. According to the data in [3], the greatest quantity of sugars in the tillering nodes of winter wheat was accumulated under conditions when the air temperature at 1300-1600 hours attained 15-20°C, but at nighttime dropped to 2-3°C. An intensive accumulation of sugars in winter crops is usually observed [4] when the mean daily air temperature is stably reduced to 8-5°C with a high daily temperature amplitude (first phase of hardening). However, in itself the content of sugars is still not an index of plant frost resistance [9]. This is only one of the many factors necessary for the development of a high frost resistance. Another factor of importance is a change in properties of the protoplast, which occurs during a period of prolonged cooling of plants at negative temperatures in the second phase of hardening.

As a result of the processes transpiring in plants, described above, it is possible to note the closest correlation between the critical temperature of winter killing of clover and the mean amplitude of air temperature (A_{10-5}) during the period from the transition of the mean daily air temperature through +10°C to its transition through +5°C ($r = 0.518$; $\eta = 0.686$). The highest frost resistance (critical temperature of winter killing from -14.5 to -15.5°C) of two-crop clover is noted with a mean amplitude of air temperature 12-14°C; with a mean amplitude of 8°C or below the critical temperature increases to -12°C.

During the period from transition of air temperature through +5°C to transition through 0°C there was a somewhat poorer correlation ($r = 0.421$; $\eta = 0.513$) between the frost resistance of clover and the mean amplitude of air temperature during this period (A_{5-0}). There is not a significant correlation between the critical temperature of winter killing of clover and the duration of the period from the transition of air temperature through +10°C to its transition through +5°C; an extremely weak correlation ($r = 0.331$; $\eta = 0.389$) is also noted for the duration of the period of transition of air temperature through +5 and 0°C.

A successful passage through the second hardening phase is dependent on the level and duration of cooling of plants in the hardening process [8, 9]. There is a relatively close direct correlation ($r = 0.423$; $\eta = 0.546$) between the critical temperature of winter killing of two-crop clover at the end of December - beginning of January and the sum of negative mean daily air temperatures during the period from a stable transition of temperature through 0°C to the end of December ($\sum t_{0-XII}$). A weak frost resistance of clover plants is formed with a temperature sum during this period equal to -80°C or less (the critical temperature in this case will be from -12.0° to -12.5°C); with a temperature sum of -150°C or more the critical temperature is from -14.0° to -15.5°C.

The correlation between the critical temperature of winter killing of clover and the sum of mean daily air temperatures during the period from a stable transition through 0°C to its transition through -5°C is expressed more weakly ($r = 0.317$; $\eta = 0.428$).

Our investigations indicated that the critical temperature of winter killing of clover is influenced to the greatest degree by the precipitation sum for the period August-October, the mean reserves of productive moisture in the cultivated soil layer during September-October, the mean amplitude of air temperature during the period from its transition through 10°C to its transition through 5°C, and also the sum of mean daily air temperatures from transition through 0°C to the end of December.

We derived equations for the dependence of the critical temperature of winter killing of two-crop clover for the first year of life at the end of December - beginning of January on the factors enumerated above:

$$t_{кр} = -0{,}012 \sum R_{VIII-X} + 0{,}4 \cdot 10^{-4} \sum R^2_{VIII-X} + 0{,}246\, A_{10-5} - 0{,}008 \sum t_{0-XII} + 10{,}77;$$

$$R = 0{,}77, \quad S_{t_{кр}} = \pm 0{,}5.$$

[kp = cr]

With the availability of data on the reserves of productive moisture in the soil in fields with sown clover for similar computations it is possible to use the following equation:

$$t_{кр} = 0{,}078\, W_{IX-X} - 0{,}002\, W^2_{IX-X} + 0{,}228\, A_{10-5} - 0{,}006 \sum t_{0-XII} + 9{,}78;$$

$$R = 0{,}79; \quad S_{t_{кр}} = \pm 0{,}48.$$

[kp = cr]

The limits of applicability of the equations are: $\sum R$ from 40 to 240 mm; W_{IX-X} from 5 to 40 mm; A_{10-5} from 6 to 14°C; $\sum t_{0-XII}$ from -40 to -280°C.

The critical temperature of winter killing of clover in the course of the winter does not remain constant. In order to determine its values in January and February it is possible to use the averaged critical temperatures cited in [7].

The outcome of wintering of grasses is dependent on the minimum soil temperature at a depth of 3 cm [10]. It was established by V. A. Moiseychik in [4] that the absolute minimum soil temperature at a depth of 3 cm in 97% of the winters is noted to 20 February; therefore it is possible to predict the degree of spareness of grasses in spring only after this time. The authors of [10] give equations for the dependence of the degree of spareness of grasses by the beginning of spring for specific fields. However, further investigations have shown that the dependence of the spareness of grasses in spring on minimum soil temperature to 20 February is more precisely described by a logistic function [5], whose course adheres to a regularity expressed by the Ferchulst equation.

For a single-crop type of clover this function has the form

$$y = \frac{100}{1 + 10^{3,8960+0,2334\,x}}.$$

For a two-crop type of clover

$$y = \frac{100}{1 + 10^{3,2672+0,2184\,x}},$$

where y is the spareness of the clover (%) according to data from a spring inspection of the sown fields; x is the minimum soil temperature at a depth of 3 cm up to 20 February, °C.

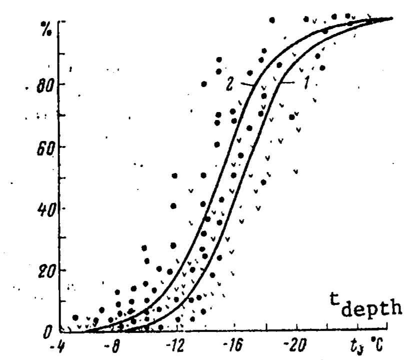

Fig. 3. Dependence of degree of spareness (%) of clover fields on minimum soil temperature at the depth of the tillering node for winter crops. 1) single-crop type of clover; 2) two-crop type of clover.

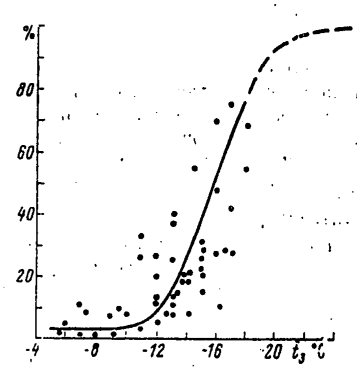

Fig. 4. Dependence of area (%) of dead sowings of grasses on mean (for oblast) observed soil temperature at a depth of 3 cm up to 20 February.

In order to accelerate the computations we constructed theoretical curves of these dependences (Fig. 3). Using these it is easy to compute the anticipated spareness of the clovers in spring two months in advance, provided one knows the minimum soil temperature at the depth of the root neck. The correlation coefficients between the actual spareness of clovers in spring and that computed using these equations are 0.84 and 0.87 respectively for single- and two-crop clovers.

The dependence of the area of death of perennial sown grasses in spring (S_{spr}) over the territory of an oblast on the oblast mean minimum soil temperature ($\bar{x}$) at a depth of 3 cm up to 20 February (Fig. 4) is described by an equation in the form

$$S_{spr} = \frac{98}{1 + 10^{4,1930+0,2650\,\bar{x}}} + 2.$$

The derived equations, cited in this article, can be used for computations of soil temperature at the depth of the root neck of grasses and the critical temperature of winter killing of clovers in evaluating the conditions for their wintering, and also for predicting the state of perennial sown grasses in spring both in a specific field and over the territory of an oblast.

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DETERMINING THE MEAN SLOPE OF A DRAINAGE BASIN

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[Article by Candidate of Technical Sciences D. M. Kudritskiy, submitted for publication 5 October 1978]

Abstract: In this article the author demonstrates the untenability of the formula recommended by A. P. Kopylov for determining the mean slopes of drainage basins and proposes a method for its improvement.

[Text] We are surprised and regretful that A. P. Kopylov [3], in criticizing weighted mean (an inapt expression!) slope methods, did not note that the method which he recommends as "the soundest and most objective" for determining the mean slopes of drainage basins, proceeding on the basis of the formula

$$\mathrm{tg}\,\alpha = \frac{\Sigma lh}{F}, \tag{1}$$

where Σ is the sum of the length of the contours, h is the contour interval, F is the area of the drainage basin, is not such, and moreover, in some cases it can be worse than the methods which Kopylov has discarded.

Formula (1), known in cartometry as the S. Finsterwalder formula, is rigorous for even slopes whose contours are depicted as parallel straight lines, that is, without distortions, and on horizontal and vertical projections of this slope. In this, as in other cases as well, the α angle is not at all an average, but simply the angle of the considered slope. In the case of an even slope the numerator in formula (1) represents the sum of projections of elementary areas of the slope in the vertical plane, whereas the denominator represents the area of the slope projected onto the horizontal plane (as seen from above).

Slope steepness can be expressed as desired, both in degrees and in abstract measure (the latter is preferable [4, 5]), making it possible to characterize the relief of the studied terrain sector differently (reference is to dissection, roughness, etc.) in the case of proper application of formula (1) and correct understanding of its possibilities.

Unfortunately, one frequently forgets (and possibly intentionally!) that on the slopes of uneven, dissected intermittent watercourses ("corrugated") the contour lines become sinuous and that they can be depicted only in a horizontal projection without distortions.

It therefore follows that the angles of inclination or slopes determined using formula (1) in all cases will be exaggerated, and, as was demonstrated in [5], rather significantly.

This circumstance was virtually simultaneously pointed out in the studies of different authors [1, 5] who proposed their own methods, if for not eliminating, at least reducing the errors introduced in applying formula (1). For example, in [5] it was proposed that the sums of the lengths be reduced (it would be difficult to reduce the length of each contour) by use of standard samples of sinuosity of the contours by means of multiplication by reduction coefficients, computed for each sample using the formula

$$K_i = \frac{\Sigma l_0}{\Sigma l_i}, \qquad (2)$$

in which Σl_0 is the sum of the lengths of the linear contours, the initial contours for constructing the samples; Σl_i is the sum of the length of the contours in the i-th sample.

The recommended method in its general features is similar to the method for determining the sinuosity of rivers when measuring them by the Shokal'skiy method and has its basic shortcoming -- subjectivity. However, for the time being it stands out as the only method for overcoming the incorrectness accompanying a noncritical use of the Finsterwalder formula.

Special investigations and experimental studies must be carried out in order to judge how great the admissible errors are in this case and whether they are comparable to the errors introduced when applying the formulas discarded by Kopylov and whether they are of great importance in hydrological investigations.

In formula (1) replacing $tg\alpha$ by the value of the I-th slope and denoting it by i for an f-th elementary area (part of the drainage basin), the mean value for the drainage basin slope is obtained using the known formula

$$I_{mean} = \frac{f_1 i_1 + f_2 i_2 + \ldots + f_n i_n}{\sum_1^n f}, \qquad (3)$$

employed in geomorphology. It seems to us that in hydrology a determination of the slopes must be accompanied by determination of the characteristics of the parts of the drainage basin, taking into account the exposure of its slopes, dissection, vegetation, etc. The use of the recommended formulas substantially facilitates obtaining them.

However, it should be noted that in addition to the already mentioned subjectivity, the method for reduction of the sums of contours is characterized by some complexity in comparison with the G. A. Alekseyev method [1, 2], is simple and graphic, although it is not without subjectivity. Therefore we feel that it is necessary to return once again to the proposals of G. A. Alekseyev and attempt to seek corrections for those cases which can be regarded as extremal in determining the slope of a drainage basin.

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UDC 535.343.4

ATTENUATION OF RADIATION IN THE WINDOW OF RELATIVE ATMOSPHERIC TRANSPARENCY 8-13 μ m

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[Article by Candidate of Physical and Mathematical Sciences V. N. Aref'yev, Institute of Experimental Meteorology, submitted for publication 22 May 1979]

Abstract: This is a review of data in the literature on the attenuation of radiation in the window of relative atmospheric transparency 8-13 μ m. The author analyzes the possible mechanisms of the absorption of radiation at 8-13 μ m by water vapor. The results of laboratory and field measurements are compared. The possibility of using approximate formulas derived in laboratory experiments for computations of the attenuation of radiation at 8-13 μ m by the atmosphere is demonstrated.

[Text] Introduction. The window of relative atmospheric transparency at 8-13 μ m plays an important role in the heat balance of our planet. The characteristics of attenuation of radiation in this window are used in many aspects of remote sensing of the atmosphere from artificial earth satellites and in a number of technical problems related to the propagation of radiation with wavelengths 8-13 μ m and, in particular, laser radiation in the air. In this review we examine data in the literature on the mechanisms and numerical values of radiation attenuation in the window 8-13 μ m.

It has now been established that the attenuation of radiation in the transparency window 8-13 μ m is determined by the following principal factors: continuous attenuation by water vapor, the so-called water vapor continuum; aerosol absorption and scattering; selective absorption by the weak lines of water vapor and some other small atmospheric components. The latter factor, described in detail in a number of studies (for example, see [15, 16, 31, 76]) is not considered in this review because its influence is small in narrow microwindows of transparency in the region 8-13 μ m, free of absorption lines.

A close correlation between continuous attenuation of radiation at 8-13μm in the atmosphere and water vapor content was established long ago. For a long time it was assumed that the continuum is formed by the far wings of strong lines and absorption bands of water vapor with their centers at about 6.3 and then 14μm [67]. According to this hypothesis, the optical thickness τ_λ = ln T_λ (T -- is the transmission of atmospheric radiation with the wavelength λ) can be stipulated in the very simple linear form

$$\tau_\lambda = \tau_{0\lambda} + K_\lambda \omega, \tag{1}$$

where $\tau_{0\lambda}$ is the optical thickness, most frequently aerosol, unrelated to water vapor; K_λ is the specific absorption coefficient; ω is the total moisture content, equal to the product of the water vapor concentration along the path.

The specific absorption coefficient was assumed to be an approximately linear function of the total air pressure P with a weak positive dependence on temperature θ (increase in K_λ with an increase in θ). In the analysis of data from field measurements on horizontal paths [59] such an idea led to a paradoxical result -- confirmation of the existence of negative $\tau_{0\lambda}$ values with small ω. Then the authors of [59] expressed the idea of a nonlinear dependence of τ_λ on ω. This idea later found confirmation in both laboratory and field experiments.

Laboratory experiments. The authors of [77] described laboratory measurements of the transmission of radiation of a CO_2 laser (λ = 10.6 and 9.4 μm) by pure water vapor and its mixture with air (80% N_2, 20% O_2 and less than 10^{-4}% CO_2) at a temperature θ = 298 K in a cell on a path with the length L = 980 m, and also under field conditions on a path with the length L = 1.95 km. The authors obtained the following dependences of the attenuation coefficient $K_{10.6}$ on the "broadening" P and partial p pressures for pure water vapor and moist air respectively:

$$K_{10,6} = \tau_{10,6}/L = 8,39\cdot10^{-4}\,p^2, \tag{2}$$

$$K_{10,6} = 4,32\cdot10^{-6}\,p\,(P + 193\,p), \tag{3}$$

where $K_{10.6}$ is in km^{-1} and P and p are in mm Hg.

The authors of [77] assumed that this result does not contradict the hypothesis [67] because the absorption coefficient at this wavelength is the sum of the coefficients of individual absorption lines in the region far from the centers and then with a Lorenz contour of the lines the optical thickness must be proportional to the square of pressure of the absorbing gas. The difference between computations using (3) and data from the author's own field measurements was attributable to the inexact value of the coefficient of resonance absorption of radiation at 10.6μm by carbon dioxide. However, at the present time it is clear that this difference is associated, in particular, with the failure to take the temperature dependence into account in (3).

In [60], on an optical path with a length of 500 m (cell with a base of 15.5 m) at temperatures 294-318 K a study was made of the transmission of radiation at 11-21 μm by artificially moistened air. In addition to a nonlinear dependence of τ on humidity, in [60] there was found to be a negative temperature dependence (decrease in absorption with an increase in temperature). These results led the author of [60] to the thought of the possibility of the existence of two mechanisms forming the water vapor continuum. One of them is governed by the broadening of the far lines and absorption bands of water vapor as a result of collisions with air molecules. The attenuation in such a case is proportional to the total pressure P and increases by approximately 0.5% with an increase in temperature by 1° (positive temperature dependence). The other, dependent on the partial pressure of water vapor and having a relatively strong (2% per degree) negative temperature dependence, is governed by the absorption of radiation by the water vapor dimer $(H_2O)_2$. Thus, the specific absorption coefficient at this temperature θ consists of two terms:

$$K_\lambda = K_1(\theta)P + K_2(\theta)p, \qquad (4)$$

where P and p are in atmospheres.

The K_2 value for the window 8-13 μm varies in the range from 10 to 17 $g^{-1} \cdot cm^2 \cdot atm^{-1}$ with θ = 303 K. The coefficient K_1, unfortunately, was not clearly determined in [60]. In the text K_1 = 0, in Fig. 8 [60] K_1 = 0.1 $g^{-1} \cdot cm^2 \cdot atm^{-1}$.

Here it is fitting to note that the dimer hypothesis, ascribed at the present time to the author of [60], was somewhat earlier (1967-1968) expressed by the authors of [80, 92] in the form of the idea of the role of water vapor associates and the energy of the hydrogen bond in the absorption of radiation in the continuum. It was demonstrated in [92], where the continuum was investigated in the water vapor layers 30-48 and 2 cm at vapor pressures of 2 and 10 atm respectively and at different temperatures in the interval 400-500 K that the energy of the bond is ΔE = 5 Cal/mol (p = 2 atm, θ = 400 and 450 K) and ΔE = 3 Cal/mol (p = 10 atm, θ = 460 and 500 K) if it is assumed that the absorption is proportional to the number of associates with the hydrogen bond, that is, is proportional to $\exp(\Delta E/R\theta)$, where R is the universal gas constant.

In [62] the spectral region 800-1250 cm^{-1} was studied at three temperatures: 296, 358 and 388 K. It was found in [62] that the absorption coefficient should be written similarly to (3), but at the same time it was shown that $K_1 \leqslant 0.005 K_2$ and it can be neglected. The form of the temperature dependence of the coefficient K_2 is represented as follows [92]:

$$K_2(\theta) = K_2(\theta = 296\,K) \times \exp[\theta_0(1/\theta - 1/296)], \qquad (5)$$

where θ = 1745 K is the temperature parameter.

It is easy to find from (5) that ΔE = 3.49 Cal/mol. The value $K_2(\Theta = 296$ K), for example, for a frequency $\nu = 945\ cm^{-1}$ is 8.25 $g^{-1}\cdot cm^2\cdot atm^{-1}$. The K_2 value for other frequencies can be found in [62, 75].

Thus, in a series of laboratory experiments in 1967-1970, it would seem, reliable confirmation was obtained of the idea of nonlinearity of the dependence of τ on ω and it was possible to demonstrate the existence of a negative temperature dependence in the transparency window 8-13μm. However, somewhat later (1972-1973) these facts were refuted in studies [41, 91].

It is pointed out in [91] that a study of the attenuation of radiation of a CO_2 laser ($\lambda = 10.6\mu$m) under laboratory conditions revealed a small positive temperature dependence (approximately 0.5% per degree) and that expression (3), derived earlier, gives somewhat understated K_λ values. A discussion of these results is not possible because [91] consists of brief summaries of a report and there was no subsequent publication of experimental data.

Source [41] gives measurements of the transmission of radiation of a CO_2 laser ($\lambda = 10.6\mu$m) by pure water vapor at $280 < \Theta \leqslant 600$ K in a multipass cell with a base of 1 m. The authors obtained an increase in the coefficient of continuous attenuation by an order of magnitude with an increase in temperature by 320 K; $K(\Theta = 280) = 0.08\ g^{-1}\cdot cm^2$ and $K(\Theta = 600) = 0.9\ g^{-1}\cdot cm^2$. Taking the linear dependence of τ on ω without any checking against their own experimental data, the authors found that with Θ = 300 K

$$\tau_{10.6} = 0{,}09\,\omega. \qquad (6)$$

In the case of the hypothesis in [60] on the two mechanisms obtained in [41], the increase in attenuation with temperature can be explained. The data in [41], which were obtained with $\Theta < 400$ K, probably cannot be taken into account because the experiments were carried out in a cell with a base of 1 m and a path length not exceeding 60-70 m. In this case with $\Theta < 400$ K the measurement error can be very great because it is necessary to measure the transmission values $T_\lambda \simeq 0.95$-0.97. Then, with a measurement error in determining T_λ of only 2-3% the error for $\tau_\lambda = -\ln T_\lambda$ can attain tens of percent and in actuality it is impossible to detect a temperature dependence. The results obtained at high temperatures ($\Theta > 400$-450 K), when dimers must be virtually absent, correspond to the concept of an increase in absorption with temperature, characteristic for the far wings of monomer water vapor molecules. This fact, incidentally, was mentioned in [78] and [92].

In [78], using a diode laser (PbSnTe) at temperatures 393-473 K, a study was made of the continuum near 1200 cm^{-1}. With $\Theta < 398$ K the results in [78] agree well with the dimer concept. At higher temperatures the attenuation is almost not dependent on Θ, but in degree is substantially higher than that computed using the empirical formula (5).

The contradictory nature of the data [60, 62, 77, 80, 92] and [41, 91] served as a basis for carrying out a cycle of experiments [2, 5-12, 22-25]. The experiments were carried out in an optical multipass cell with a base of 50 m. The transmission of the radiation of a CO_2 laser (λ = 10.6 μm) by pure water vapor, its mixture with nitrogen and moist air at different temperatures in the range 284-353 K was measured on paths up to 3 km in length. The results were compared with field data. It is shown that the main contribution to the mechanism of formation of the continuum at about 1000 cm^{-1} is from molecular (including dimer) absorption of radiation by water vapor. The results of the laboratory measurements are approximated by the expression

$$\tau_{10,6}=L[K_1a(1+\alpha P)+ +K_2a^2\exp(-\Delta H/R\Theta)], \quad (7)$$

where L is path length, km; a is absolute humidity, g/m^3; P is broadening pressure, mb; R is the universal gas constant; K_1, K_2, α, ΔH are adjusting parameters. For $\lambda = 10.6\mu m$, $K_1 = 1.76\cdot10^{-3}\ g^{-1}\cdot m^3\cdot km^{-1}$, $K_2 = 0.42\cdot10^{-6}\ g^{-2}\cdot m^6\cdot km^{-1}$, $\alpha = 1.78\cdot10^{-3}\ mb^{-1}$, ΔH = -4546 cal/mol [10, 11].

In 1976 the results of a laboratory investigation of the continuum using an adjustable CO_2 laser [89] were published in full. With a constant temperature Θ = 300 K the optical-acoustic spectral background was used in measuring the absorption coefficients of emission of 49 lines for a CO_2 laser adjustable from 9.2 to 10.7 μm for moist air from which the carbon dioxide had been removed. In the opinion of the authors of [89], their results, obtained with three values of the partial pressure of water vapor (5, 10, 15 mm Hb), are lower than the data in [62] by 5-10%. For emission with a wavelength of 10.6 μm the following approximation expression was proposed in [89]:

$$\tau_{10,6}=7{,}0\cdot10^{-6}\,p(P+88\,p)L, \quad (8)$$

where P and p are in mm Hg, L is in km.

Sources [79, 81] give the results of laboratory measurements by two methods (multipass cell L = 1.5 km and spectral background) of the attenuation of 27 lines of an adjustable CO_2 laser with a H_2O-N_2 mixture at P = 1 atm and a constant temperature. Both methods gave good agreement. The absorption coefficient $K_\nu = \tau_\nu/L$ is approximated by the expression

$$K_\nu(\Theta)=K_0(\nu,\Theta)\omega[p+\gamma(P-p)], \quad (9)$$

where ω is the number of H_2O molecules in one cm^3, P is the total pressure of the mixture in atmospheres, p is the partial pressure of H_2O in atmospheres, γ is the parameter of the relative role of broadening and self-broadening.

The authors of [79, 81] assume that their K(λ = 10.6 μm) agree well with the data in [77], are 5-10% lower than [62], and are approximately 18% lower than [11]. In [81] the γ parameter is different for different wavelengths, in contrast to [62], where $\gamma = 0.005$, and from [87], where $\gamma = 0.001$ for all wavelengths.

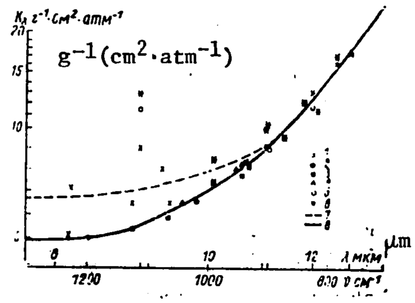

Fig. 1. Frequency dependence of coefficient of continuous absorption according to data from different authors. 1) data from [62]; 2) new data from author of [62], cited in [87]; 3) [10]; 4) [77]; 5) [60] with θ = 296 K; 6) [60] with θ = 303 K; 7) averaged data [62]; 8) computations using (10).

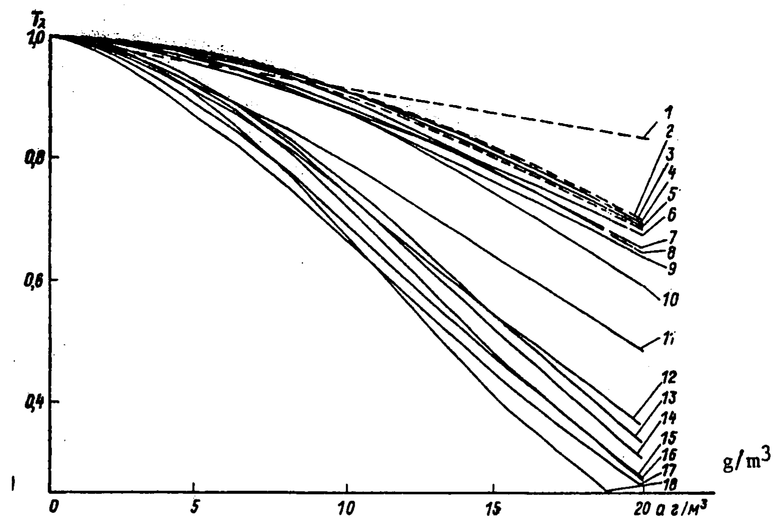

Fig. 2. Comparison of results of laboratory experiments of different authors. 1) [41]; 2, 7, 15 [77]; 3, 13) [81]; 5, 9, 17) [11]; 4, 14) [87], 6, 12) [89]; 8, 16) [62]; 10, 18) [60] with $K_1 = 0$; 11) [60] with $K_1 = 0.1$. The dashed curves are for pure water vapor, the remaining curves are for moist air or a mixture of nitrogen with water vapor. 1-11 for L = 1 km, 12-18 for L = 3 km.

Source [1] gives the value of the attenuation coefficient (0.35±0.07 db/km), obtained in a cell with a base of 5 m with θ = 298 K, P = 760 mm Hg, a = 7.5 g/m^3 for λ = 10.6 μ m, within the limits of accuracy agreeing with data in [77].

Comparison of results of laboratory experiments of different authors. Results of laboratory experiments of different authors are compared in [10, 11, 38, 46, 69, 87]. Figure 1, combining Fig. 2 from [69] and Fig. 1 from [87], for a constant temperature θ = 296 K, gives the frequency dependence of the absorption coefficient K_ν. Source [87] gives new "unpublished" data from the author of [62] (points in Fig. 1), which in the region shorter than 10 μm do not give the appreciable scatter observed in [62] (crosses), and in the region 10-12 μm completely coincide with the results in [62]. The authors of [87] relate the scatter to selective absorption and do not take them into account. For the same reason it is possible to neglect the data in [60] in the short-wave region. In such a case the results of laboratory measurements of different authors agree well and the frequency dependence can be approximated in the following way:

$$K_\nu(\Theta) = A + B\exp(-C\nu). \tag{10}$$

where with θ = 296 K, A = $1.25 \cdot 10^{-22}$, B = $1.67 \cdot 10^{-14}$, C = $7.87 \cdot 10^{-3}$ [87].

In [38] and [46], where the comparison of the absorption coefficients of different authors in the form (4) was made at a constant temperature and with different humidities, there was also a good agreement.

In [10, 11], which give a comparison of the transmission of radiation at 10.6 μm, computed using different approximation formulas, the conclusion is drawn that they are is satisfactory qualitative agreement. Here we will examine more carefully the results of the comparison of transmissions carried out in [10, 11], supplementing these with data from laboratory measurements not taken into account in these studies. For comparative purposes all the approximation formulas were reduced to a uniform form for λ = 10.6 μm at θ = 298 K and P = 1013 mb and the determined transmission values for L = 1 and 3 km are represented in Fig. 2. If we exclude the data in [41], not taking into account the nonlinearity of the dependence of τ on a, and due to the lack of clarity in determining K_1 the data in [60] with K_1 = 0.1, within the limits of experimental accuracy the results of the computations for the path L = 1 km are in agreement. However, for L = 3 km, in the case of great air humidities, there is a discrepancy attaining 25-30%. It was postulated in [38] that this discrepancy is associated with the presence of aerosols in the cells during the carrying out of laboratory experiments. In [12] it was confirmed that the cell contains aerosol and at the same time it was demonstrated that its contribution to the absorption of radiation at 10.6 μm does not exceed the experimental errors. Accordingly, the aerosol cannot explain the discrepancy in transmission values at L = 3 km observed in Fig. 2. This discrepancy can be attributed to some quantitative difference in the types of dependence on temperature and pressure in the formulas of different authors.

Table 1

Temperature Parameter According to Data from Different Authors

Θ_0 [87]	Литературный источник 1	ΔH [11]	Литературный источник 1
1810	[11]	4,55±0,37	[11]
—	—	3,60±0,30	[11]
1800	[60]	3,46—4,00	[75, 60]
1745	[62]	3,49	[75, 62]
1800	[63]	—	—
1750 *	[68]	—	—
2000	[92]	3,0—5,0	[92]
—	—	4,8 *	[18]
—	—	3,0—6,0 *	[61]
—	—	3,74±0,70 *	[75, 64]
—	—	5,76 *	[75, 66]

KEY:

1. Source in literature

Table 2

F Values According to Formulas of Different Authors Reduced to Form (7) (θ = 298 K and P = 1013 mb)

Литературный источник 1	2 a г/м³				3 Примечание
	0,1	1	10	20	
[11]	57,6	5,76	0,576	0,288	
[60]	76,5	7,65	0,765	0,382	4 При K_1=0,1
[62]	36,8	3,68	0,368	0,184	5 Верхний предел
[77]	38,1	3,81	0,381	0,191	5 Верхний предел
[87]	7,4	0,74	0,073	0,036	γ=0,001
[89]	79,7	7,97	0,797	0,398	6 λ=10,6 *мкм*
[81]	21,7	2,17	0,214	0,106	7 Спектрофон
[81]	26,9	2,69	0,266	0,131	8 Кювета λ=10,6 *мкм*
[42]	670—400	67—40	6,7—4,0	3,35—2,0	9 Для ν=1200÷735 *см*$^{-1}$ соответственно

KEY:

1. Source in literature
2. a g/m^3
3. Note
4. With K_1 = 0.1
5. Upper limit
6. ...μm
7. Spectral background
8. Cell...μm
9. For... respectively

An analysis of the temperature dependence was presented in [11] and [87]. The values of the temperature parameters (θ_0 [87] and ΔH [11] are presented in Table 1. The values marked with an asterisk were obtained from independent nonspectral measurements. It can be seen from Table 1 that the temperature parameters obtained in optical measurements agree with one another and with data from individual evaluations. The authors of [87] feel that the best evaluation is θ_0 = 1800 K.

In order to evaluate the role of broadening we will examine the data in Table 2, where for different absolute humidity a values we have presented the F values -- the ratios of the terms of the approximation formulas containing P, to the terms quadratic relative to humidity. It can be seen from Table 2 that F differs by several times according to the data of these authors, without taking into account the great exaggeration of F as given in [42], where use was made of the approximation of a linear dependence of τ on humidity. The reason for the differences in the dependence of τ on broadening pressure may be the errors in finding this dependence on the basis of two points: P = 0 and P = 1 atm. It was demonstrated in [47] that this difference can lead to an appreciable difference in the computed transmission values. In addition, it was noted in [79] that the H_2O-N_2 and H_2O-O_2 collision sections differ somewhat, but in most approximation formulas this is not taken into account.

Results of field measurements. An examination of the results of field investigations can be begun with [59]. References to earlier studies can be found in [69, 73]. Source [59] gives measurements of the attenuation of solar radiation by the atmosphere with a resolution of 0.3 cm^{-1} at a wavelength of 10μm. As already mentioned in the introduction, source [59] for the first time reported the existence of negative residual optical thicknesses and it was postulated that there is a nonlinear dependence of absorption in the window 8-13μm on water vapor content. Similar optical thicknesses were discovered in [58] in measurements of solar radiation in the neighborhood of Gelendzhik. In [3], a continuation of [58], no negative optical layers were discovered. The authors of [3] attribute this to the availability of a great number of measurements with low water vapor contents.

Source [90] gives data from measurements of the attenuation of solar radiation by the entire thickness of the atmosphere in the range of wavelengths 0.65-13.5μm. A separation of the results into five groups on the basis of the content of water vapor, temperature, water vapor elasticity and horizontal range of visibility made possible a reliable determination of the presence of a temperature dependence and two absorption components in the window 8-13μm, that is, a nonlinear dependence of absorption on humidity.

Sources [14, 46, 47] give data from systematic measurements of transmission of IR radiation by the atmosphere on horizontal and slant paths using an IKAU-1 apparatus [17]. The results of the measurements confirm the nonlinear dependence of absorption on absolute humidity. The authors of [14] proposed approximation formulas relating the molecular component of transmission of radiation at 8-13μm to the number of atmospheric masses and the quantity of precipitable water. These formulas do not take into account the dependence on temperature. In the considered studies there was a separation of the contribution of water vapor and aerosol to the attenuation of radiation in the window 8-13μm. In [47], after excluding continuous and selective absorption of water vapor, the authors obtained a virtually neutral spectral variation of aerosol attenuation for the entire region 8-13μm.

Source [35] gives the results of measurements of spectral transmission on a horizontal path L = 1.2 km with a meteorological range of visibility not less than 50 km. For the transparency microwindow 8.63, 9.2 and 10.1μm there was a linear dependence of the optical thickness τ on the water vapor content ω. For the microwindow 11.1 and 12.01μm, as in [52] for the microwindow in the spectral region beyond 15μm, there was a nonlinear dependence of τ on ω.

In [65, 72, 75, 82-86] the authors give the results of measurements with a radiometer from an aircraft or balloon at different latitudes of ascending and descending radiation and also counterradiation of the lower layer of the atmosphere. The authors of all these studies indicate a good agreement between field data and laboratory measurements [60 or 62].

Sources [70, 71], on the basis of measurements of atmospheric counterradiation in the window 8-13μm with a multichannel radiometer, set forth an analysis of the relative contribution to attenuation from aerosol, dimers and monomers of water vapor. The results are given in Table 3, from which it can be seen that during the warm season of the year (the measurements were made from March through November) in the low and middle latitudes there is a predominance of the contribution of dimer absorption.

It was shown in [29], in the processing of measurement data from the "Cosmos-149" satellite on outgoing radiation in the window 8-13μm for the purpose of determining the temperature of the underlying surface, that satisfactory results can be obtained when dimer absorption in the water vapor continuum is taken into account.

Thus, most of the field investigations carried out during the last decade confirm the existence of a nonlinear dependence of absorption of radiation at 8-13μm by the atmosphere on air humidity. Some of these investigations also reveal a negative temperature dependence.

Table 3

Relationship of Contribution of Different Factors to Attenuation of 11.5μm-Radiation [71]

Географический район 1	τ аэрозольное 2	τ димерное 3	τ мономерное 4	τ суммарное 5
6 Арктика	0,0202	0,0020	0,0012	0,0234
7 Умеренные широты	0,0202	0,1070	0,0097	0,1369
8 Тропики	0,0202	0,4510	0,0243	0,4955

KEY:
1. Geographic region
2. Aerosol
3. Dimer
4. Monomer
5. Total
6. Arctic
7. Temperate latitudes
8. Tropics

Table 4

Value of Factor $C_0(\lambda)$ for Microwindow of Transparency In Region 8-13μm According to Data from Different Authors

λ, мкм μm	[13]	[25]	[87]	λ, мкм μm	[25]	[87]
8,10	—	0,59	0,60	11,10	1,12	1,18
8,30	—	0,59	0,61	11,60	1,29	1,40
8,60	—	0,62	0,635	11,80	1,37	1,50
8,74	—	0,63	0,65	12,20	1,59	1,73
9,06	—	0,67	0,68	12,40	1,80	1,86
10,12	0,80	—	0,87	12,50	1,90	1,93
10,15	—	0,85	0,88	12,90	2,20	2,22
10,32	0,92	—	0,917	13,00	2,25	2,30
10,40	—	0,95	0,94	13,10	2,30	2,38
10,60	1,00	1,00	1,00			
10,72	1,035	—	1,038			

Comparison of results of field and laboratory experiments. A comparison of the results of field and laboratory measurements in many studies has a qualitative character when the data from field measurements are compared with laboratory data obtained with some fixed values of the parameters. An example is Fig. 3 from [69]. The agreement between the results of field measurements and laboratory data [62] for $\lambda > 10\mu m$ is satisfactory. In [69] the discrepancy at $\lambda < 10\mu m$ is attributed to the selective absorption by N_2, CH_4, CO_2.

In [8-11, 22-26] use was made of a quantitative approach to such a comparison, involving integration of the approximate expression (7) along the ray path with values of atmospheric parameters determined simultaneously with each spectral measurement. A comparison of field data for λ = 10.4

μm and computations (using (7)) indicated that in the computations it is necessary to take into account not the total water vapor content, but the real vertical profiles of humidity, temperature and total pressure [8, 9, 22, 23].

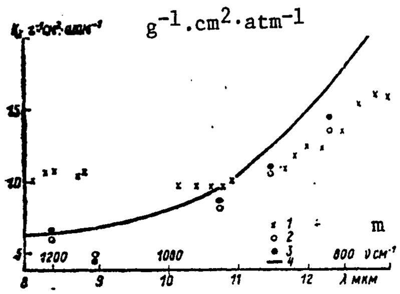

Fig. 3. Comparison of results of field measurements of different authors and data from laboratory experiments [62]. 1) [49]; 2) [70]; 3) [71]; 4) [62].

Expression (7) is directly applicable for computations of the attenuation of radiation only in a narrow spectral region near 10.6μm. Since, however, in all the microwindows of the region 8-13μm the temperature dependence is identical [23, 62, 90], the authors of [22-25] proposed that in the computations (7) be extrapolated into the entire region 8-13μm using such a $C_0(\lambda)$ factor that

$$\tau_p(\lambda) = C_0(\lambda)\, \tau_p(\lambda = 10{,}6 \text{ мкм}), \tag{11}$$

[p = comp] where τ_{comp} (λ= 10.6μm) is computed using (7).

Table 4 gives $C_0(\lambda)$ values from [25], computed using (10) [87] and found in laboratory experiments [13] with the lines of an adjustable CO_2 laser. The table shows that the $C_0(\lambda)$ values, obtained by different methods, are in good agreement.

In Fig. 4, on the basis of [25], we give the results of field measurements of the attenuation of solar radiation at 10-13μm τ_{ex} [see note] and computations on the basis of (7) τ_{comp} in the form of a regression function and in the form of the dependence of τ_{ex}/τ_{comp} on τ_{comp}. It can be seen from Fig. 4 that the ratio τ_{ex}/τ_{comp} with a dispersion 0.15 is equal to unity for all the considered wavelengths and all the observation points in the interval $0.05 < _{comp} \leqslant 1$, including for the conditions prevailing in the middle and low latitudes. [Note: Use was made only of those measurement results [3, 37, 58] which were obtained in a pure atmosphere (under stable anticyclonic conditions with a meteorological range of visibility $S_n > 20$ km.] Thus, in all these cases under natural conditions the principal reason for the attenuation of radiation at 10-13μm is molecular absorption, described by expression (7).

With $\tau_{comp} < 0.05$, which corresponds to winter conditions in the middle latitudes, the ratio τ_{ex}/τ_{comp} increases considerably, indicating the need for introducing the additional mechanism of attenuation of radiation, not taken into account by expression (7), in order to explain the "excess" optical thickness τ_{ex}. This mechanism can be related both to the scattering of radiation by large ice particles [24] and to the absorption of radiation by the finely disperse aerosol fraction [48].

Source [26] gives the results of measurements of the attenuation of solar radiation at 8-13μm under conditions of low humidity and low temperature in Antarctica. The measurements were accompanied by constant observations of the spectral transparency of the entire thickness of the atmosphere at a wavelength of 0.63μm. The value of the vertical optical thickness for $\lambda = 0.63\mu$m under the measurement conditions was $\tau_{0.63} = 0.048$-0.100 (mean value 0.064). In comparison with Rayleigh scattering ($\tau_{Ray} = 0.054$ and 0.059, according to [21] and [28] respectively), this value is only 10-20% greater, which makes it possible to consider the measurement conditions as corresponding to a virtually "aerosol-free" atmosphere. A comparison of the results of measurements in Antarctica with computations made using (7) is given in Fig. 5, from which it follows that for the microwindow in the interval 10.4-12.7 μm the computed τ_{comp} and measured τ_{ex} values of the optical thicknesses within the limits of measurement errors agree. The standard deviation from the straight line, corresponding to $\tau_{ex} = \tau_{comp}$ for the entire mass of data in this part of the spectrum, is 0.007. However, Fig. 5 shows that for the microwindows in the interval 8-10 μm $\tau_{ex} > \tau_{comp}$. Since in these measurements there is little aerosol attenuation, the additional absorption is evidently attributable to selective molecular absorption of small components in the atmosphere.

Source [87] gives the results of a comparison of data on measurements of transmission of radiation by the water vapor continuum along a horizontal path L = 20 km over the open water surface and computations on an electronic computer using the LOWTRAN program [88]. This program for computations in the region 8-13μm is based on data from laboratory experiments [62] and [77], but without taking into account the temperature dependence. The authors of [87] arrived at the following conclusions: 1) in the region 8 μm the computed data are 25% below the experimental data, 2) it is necessary to introduce a temperature dependence into the computation program; 3) there is uncertainty with respect to the value of the coefficient in the term containing the dependence on total pressure. Thus, the LOWTRAN program used for transmission computations in the window 8-13μm needs correction.

In [47] data from field measurements of transmission of solar radiation were compared with computations using (7) and (5), taking into account the corrections proposed in [87]. The results are presented in Fig. 6 [47]. Since aerosol introduces additional attenuation in this attenuation of radiation, it must be expected that the points in Fig. 6 must be near or above the straight line $\tau_{ex} = \tau_{comp}$. Proceeding on this basis, and taking into account that the points denoted by a cross were obtained in what is known to be a closed atmosphere, in Fig. 6 falling below the straight line $\tau_{ex} = \tau_{comp}$, the authors of [47] feel that their field data are in better agreement with computations using (5) than using (7).

Thus, a comparison of data from field measurements with computations using the approximation formulas, obtained in laboratory experiments, demonstrated that these formulas describe well the field data in the spectral region 10-13 μm at both positive and negative temperatures on the Celsius scale. During the warm season of the year in the pure atmosphere these formulas completely describe the attenuation of radiation at 10-13 μm by the atmosphere. In winter and in a turbid atmosphere the aerosol attenuation must also be taken into account.

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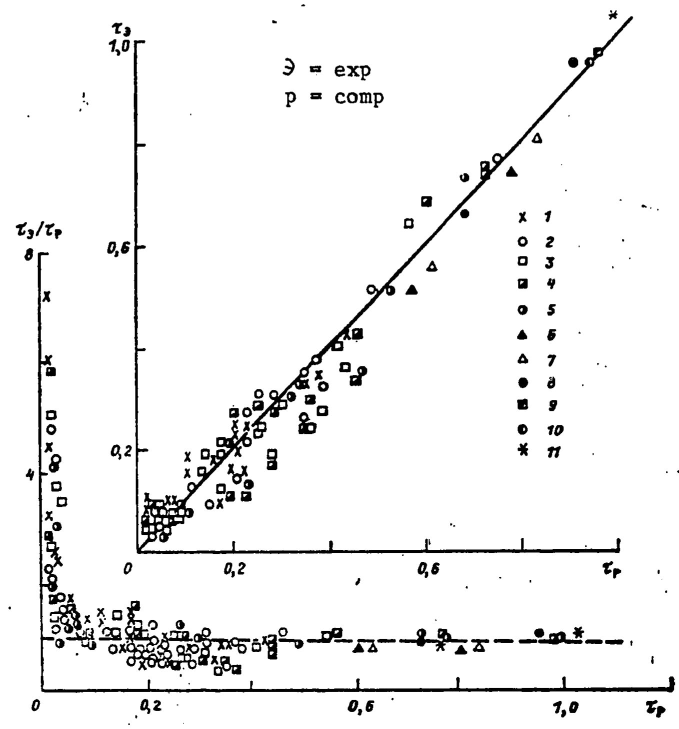

Fig. 4. Comparison of results of field measurements with computations using (7). 1) microwindow of transparency with center at 10.4 μm; 2) 11.08 μm; 3) 11.6 μm; 4) 11.8 μm; 5) 12.2 μm; 6) 12.4 μm; 7) 12.5 μm; 8) 12.8 μm; 9) 12.9 μm; 10) 13.0 μm; 11) 13.1 μm.

Role of aerosol in attenuation of radiation at 8-13 μm. The conclusions on the role of aerosol drawn in the preceding section on the basis of a comparison of field and laboratory measurements find confirmation in a cycle of special investigations of the contribution of an aqueous aerosol to the attenuation of radiation at 8-13 μm [36, 53]. In these investigations it was found that with a meteorological range of visibility $S_m < 20$ km aerosol attenuation can be extremely significant (Table 5). With $S_m >$ 20 km the coefficients of aerosol attenuation in the region 8-13 μm do not exceed 0.02 km^{-1} [36]. Computations of the volumetric scattering coefficients $\varkappa_{sc}$ and volumetric absorption coefficients $\varkappa_{abs}$ using the distribution functions for the sizes of aerosol particles measured using the "Kvant" aerosol spectrometer [54] indicated that in the considered range of wavelengths $\varkappa_{abs} \gg \varkappa_{sc}$ is almost always observed.

In a series of studies [20, 30, 37-40, 55-57], in contrast to all those considered above, the point of view is successively defended that aerosol attenuation plays a decisive role in forming the continuum 8-13 μm.

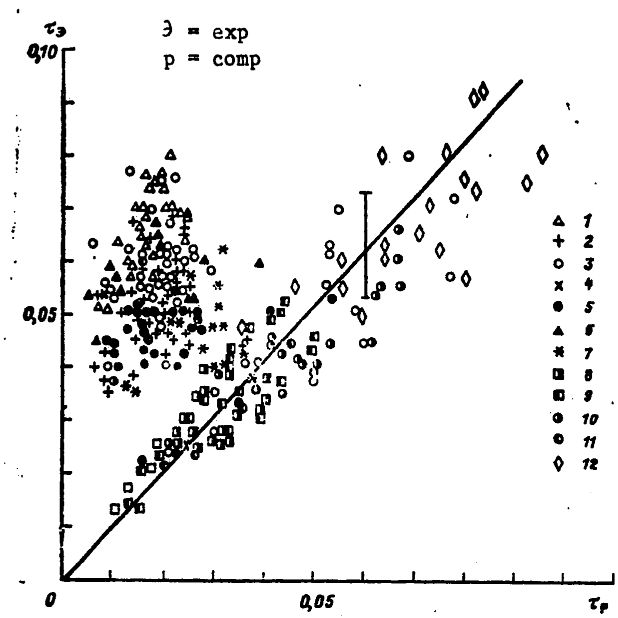

Fig. 5. Comparison of results of field measurements in Antarctica with computations using (7). 1) microwindow of transparency with center at 8.12 μm; 2) 8.33μm; 3) 8.63μm; 4) 8.74μm; 5) 9.06μm; 6) 9.34μm; 7) 10.15μm; 8) 10.40μm; 9) 11.08μm; 10) 11.60μm; 11) 12.20μm; 12) 12.80μm.

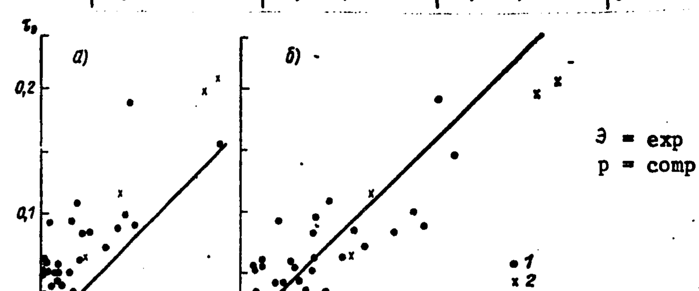

Fig. 6. Comparison of results of field measurements [47] with computations. a) according to (5); b) using (7). 1 and 2) experimental data [47] in pure and turbid atmospheres respectively.

The first part of the series [37, 38, 55, 57] gave the results of measurements of the transmission of atmospheric IR radiation in the window 8-13μm on horizontal and slant paths. The processing of experimental data was carried out in accordance with expression (1). The authors assume that aerosol is responsible for all points greater than the so-called minimum points (minimum within the limits of the scatter of points of the optical thickness value τ with different ω values, mandatorily including small ω values), caused by actual experimental errors.

Table 5

Coefficients of Aerosol Attenuation (km^{-1}) with Three Values $S_m < 20$ km [36]

S_m км	λ мкм m				
	0,55	10,0	11,0	12,0	13,0
10	0,39	0,014	0,029	0,043	0,074
5	0,78	0,027	0,058	0,086	0,148
1	3,9	0,140	0,290	0,430	0,740

The minimum τ_{min}, including not only the absorption of radiation by water vapor, but a small addition, attenuation due to dry aerosol, corresponds to the maximum value of the second term in expression (1). Qualitatively τ_{min} agree with laboratory data [60, 62, 77] without taking into account the dependence on temperature, but a nonlinear dependence of τ_{min} on water vapor content is not reliably detected [38, 55]. The temperature dependence τ_{min} does not exceed the limits of measurement errors [55], and under natural conditions it is impossible to use the temperature dependence obtained in laboratory experiments [57].

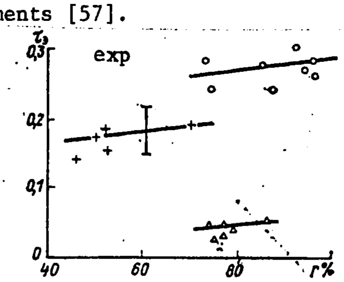

Fig. 7. Dependence of τ (λ = 10.2 μm) on relative humidity in case of a constant water vapor content.

In [55] the conclusion that there is a dependence on temperature was drawn in an analysis of a mass of experimental data obtained with a change in θ from 9 to 22°C, which on the basis of laboratory experiments should lead to a change in τ by approximately 25%. Absolute humidity was 11-13 mb, which with allowance for the nonlinear dependence should lead to a change in τ by this same value (25-30%). It is therefore clear that with such a choice of the mass of experimental data it is impossible to detect any temperature dependence. This same study [55] gives the dependence of τ_{min} on relative humidity r (Fig. 7) with a constant water vapor content and the conclusion is drawn that there is a weak dependence of τ on r. However, if it is taken into account that with ω= const an increase in r is related to a decrease in θ, Fig. 7 can be interpreted as an increase in τ with a decrease in temperature.

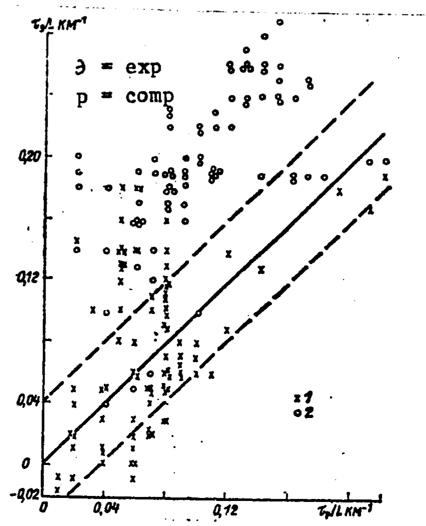

Fig. 8. Comparison of field measurements on horizontal path [55] with computations using (7). 1) $S_m > 20$ km, 2) $S_m < 20$ km; broken straight lines -- limit of measurement errors.

Accordingly, the concept of minimum points, as a single-parameter dependence, is clearly inadequate for an analysis of field measurements. This is also shown in [46]. In [25] many data [37, 38, 57] for slant paths and conditions in the pure atmosphere could be interpreted using expression (7) without taking aerosol into account (see Fig. 4). Moreover, it can seen from Fig. 8, where the author of [55] compares the results of his measurements on a horizontal path with computations using (7), that with $S_m > 20$ km in most cases it is possible to get by without taking aerosol into account.

In the second part of the series [20, 39, 40, 48, 56] a statistical analysis of experimental data was made for spectral transparency of the atmosphere on horizontal and slant paths obtained in the first part of the series. If it is taken into account that in the forming of a set of experimental data for statistical analysis there was no exclusion of the contribution of water vapor and also that there is possibly of some dual interpretation of some results of the analysis (a series of results, together with an "aerosol" interpretation, allows a "water vapor" interpretation), the idea of a role of absorption of radiation at 8-13 μm in the submicron fraction of some types of aerosol merits attention. The good promise of this idea is also strengthened by the fact that in the spectrum of some types of aerosol there is a strong absorption band in the region 8-13 μm [34].

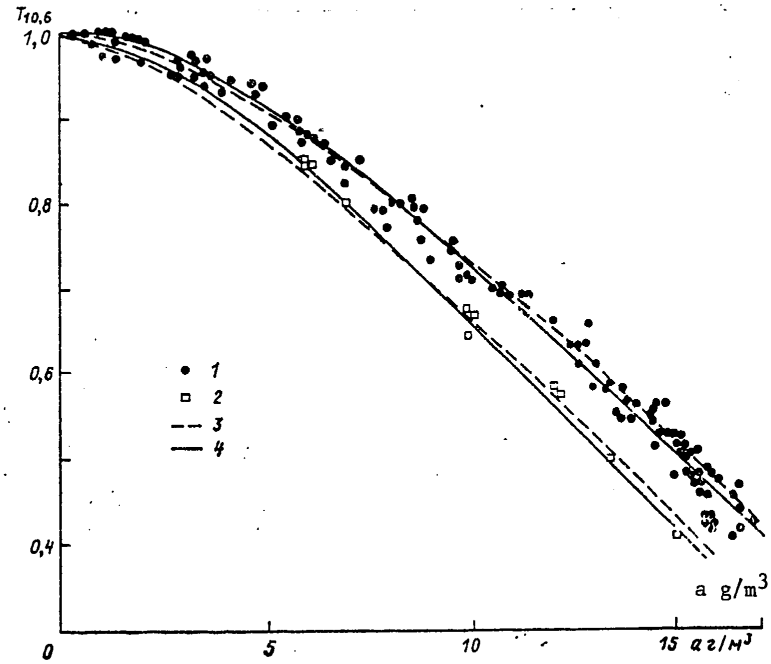

Fig. 9. Comparison of data from laboratory experiments [9] and computations using (12). 1) pure water vapor, 2) mixture of water vapor and nitrogen, 3) computations using (7), 4) using (12).

Thus, under conditions of a turbid atmosphere and in some cases under conditions of a pure atmosphere, in winter, for example, allowance for aerosol is necessary. In order to clarify the mechanism, the role of different types of aerosol and creation of engineering methods for taking aerosol attenuation into account it is necessary to carry out comprehensive investigations which must include measurements of spectral transparency and the properties (chemical composition, distribution functions, etc.) of aerosol.

Theoretical investigations of the continuum. Attempts to describe continuous absorption in the region 8-13μm have been made repeatedly. However, comparisons of theoretical computations in which different forms of contours of spectral lines (complete and simple dispersion, Van Vleck-Weisskopf, Zhevakin-Naumov, and others) with experiments were unsatisfactory (for example, see [4, 27, 30, 37, 59, 65]).

In theoretical investigations [32, 33, 43-45, 49-51] a fundamentally new approach was used for the description of absorption in the windows of relative transparency. A quantum mechanical scheme for computations includes interaction of pairs of near-colliding identical absorbing molecules and as a result leads to a nonlinear dependence of absorption on the concentration of absorbent. The absorption coefficient was obtained in the form of the sum of two terms:

$$K = \varkappa_1 P_1 + \varkappa_2 P_2, \tag{12}$$

where P_1 and P_2 are the partial pressure of the broadening gas and the absorbent; $\varkappa_1$ and $\varkappa_2$ are absorption coefficients related to broadening and self-broadening.

The authors obtained a good agreement between computations using (12) for different absorbents and in different parts of the spectrum with experimental data. In Fig. 9, from [44, 45] we give a comparison of the experiments [9] at a wavelength 10.6μm with computations using (12) with $\varkappa_1$ = 0.034 $g^{-1} \cdot cm^2 \cdot atm^{-1}$ and $\varkappa_2$ = 8.75 $g^{-1} \cdot cm^2 \cdot atm^{-1}$. As indicated in Fig. 9, there is a good agreement between experimental and computed results, although it must be noted that in the computations, as noted by the authors of [45] themselves, no allowance is made for the influence of the 6.3μm vibrational-rotational water vapor absorption band, and in addition, the computations were made with θ = 300 K, whereas the experiments in [9] were made with θ = 293 K. Agreement of computations using (12) with experiments is also noted in [30]. [In the caption to the figure in [44] and [45] there was a misprint in the notation of the solid and dashed curves which has been corrected in Fig. 9.]

Taking into account that in solution of the Schrödinger equation use is made of the unperturbed wave functions of colliding molecules, the authors relate the considered spectrum to the far wing of the purely rotational absorption band of a monomer water vapor molecule [45]. However, first, allowance for weak perturbations of the wave functions scarcely leads to radical qualitative changes in the results, and second, this spectrum is formed due to close collisions of two absorbing molecules, which at the time of interaction with radiation can be regarded as a weakly bound short-lived dimer [10]. At the present time it is difficult to assign the considered absorption to a specific energy transition due to the imperfection of theoretical analysis of the water vapor dimer, although in [18] absorption in the region 8-13μm is hypothetically related to translational-vibrational, whereas in [93] it is related to the vibrational $(H_2O)_2$ transition.

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FORMULATION OF THE PRINCIPLES OF HYDROLOGY AND THE DYNAMICS OF CHANNEL FLOWS IN THE PUBLICATIONS OF M. A. VELIKANOV

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[Article by Candidate of Technical Sciences M. M. Arkhangel'skiy, Professor D. I. Grinval'd, Candidate of Physical and Mathematical Sciences N. A. Mikhaylova and N. S. Sharashkina, Moscow State University and Odessa Hydrometeorological Institute, submitted for publication 6 July 1979]

Abstract: This paper gives a review of the publications of Corresponding Member USSR Academy of Sciences M. A. Velikanov, who created the theoretical basis of the method for computing runoff and who developed practical methods for predicting snow-produced and rain-induced high water. He made a substantial contribution to the probabilistic theory of movement of bottom sediments and for the first time directed attention to the necessity for making allowance for the energy expended by the flow on the transport of suspended sediments. M. A. Velikanov formulated the fundamental principles for study of the channel process, which served as a basis for the modeling of this process in erodable models.

[Text] The content of hydrology of the land as the science of water bodies was determined in the publications of Soviet hydrologists in 1920-1925. Among the founders of hydrology one of the leading places is occupied by M. A. Velikanov. At the present time it is difficult to visualize the science of the earth's hydrosphere without a number of its important aspects whose scientific principles were laid in the publications of M. A. Velikanov. He published about 200 studies.

M. A. Velikanov was born on 22 January 1879 in Kazan'. In 1903 he graduated from the Peterburg Institute of Transportation Engineers, after which he worked as an engineer on the waterways of Siberia. In 1925 M. A. Velikanov was selected as senior hydrologist of the State Hydrological

Institute, established by a decree of V. I. Lenin. In 1930 he was called to the Department of Hydrology of the Land at the Moscow Hydrometeorological Institute. In 1935 he headed the planning and outfitting of the Laboratory of Physical Hydrodynamics of the Power Institute USSR Academy of Sciences and then headed the investigations of this laboratory. Later the laboratory was renamed the Laboratory of Channel Processes and was transferred to the Institute of Geography USSR Academy of Sciences. In 1939 M. A. Velikanov was elected Corresponding Member USSR Academy of Sciences. In 1945 he was invited to head the newly organized Department of the Physics of Channel Flows of the Physics Faculty Moscow State University. M. A. Velikanov died on 30 April 1964.

M. A. Velikanov made a fundamental contribution to the theory of surface runoff, the water balance of river basins, the movement of sediments and channel processes. His book GIDROLOGIYA SUSHI (Hydrology of the Land) was published in 1925. This was the first generalizing work in this field of science. The book systematically discussed a broad range of hydrological problems, gave an original interpretation of a number of important problems and contained principles constituting a scientific basis for the development of a number of directions in hydrology of the land. This book was then reissued in four editions.

M. A. Velikanov gave great importance to different ways to solve hydrological problems. At the same time, adhering to his scientific points of view, he emphasized the physical aspects of hydrology. He defined hydrology as the science of the distribution and activity of water flows and water bodies on the surface of the earth's land. M. A. Velikanov pointed out that in hydrology many "such areas, whose thorough study mandatorily requires the use of physical experimentation and the use of a well-developed physicomathematical approach" had appeared.

M. A. Velikanov gave primary importance to problems relating to the water balance and runoff in his investigations. His studies created the theoretical basis for methods for computing river runoff. Emphasizing the need for a physical approach to study of hydrological processes, he, however, did not forget that these processes develop in a definite geographical environment. He regarded the study of the water balance "as the principal and most important part of the entire hydrology of the land."

M. A. Velikanov devoted great attention to the principles of organization of experimental investigations at hydrological runoff stations. In essence, he established the first of these near Moscow. This was the Kuchinskaya Hydrological Station, constituting the prototype for organizing investigations of runoff at many stations in the USSR and abroad. M. A. Velikanov devoted a series of studies to the problem of experimental study of runoff in small basins. The productiveness of the approach which he defined was confirmed by all subsequent development of hydrology. These studies also favored the creation of a global network of experimental and representative basins with the active participation of Soviet hydrologists within the framework of the International Hydrological Decade (1965-1974) and International Hydrological Program (since 1975).

In connection with the problem of study of the runoff of small rivers the problem arose of refining the water balance equation and the possibilities of its use for rivers of different size. M. A. Velikanov especially emphasized the role of water exchange with the surrounding basin. Relying on the water balance method, he theoretically demonstrated the legitimacy of the hydroclimatic approach to solution of problems in computing long-term runoff and proposed practical ways for their solution in this way. In collaboration with D. L. Sokolovskiy he carried out an investigation of the dependence between runoff, precipitation and the dew-point spread for a number of river basins in the USSR and obtained a corresponding formula for practical computations of the mean long-term runoff of rivers. It is also necessary to mention the quasiconstants method, developed by M. A. Velikanov for the purpose of giving engineers a reliable method for refining the computed values of the hydrological characteristics read from approximate maps in isolines.

Giving great importance to the development of the method for forecasting and computing rain-induced high waters and snow-induced high waters, M. A. Velikanov not only formulated the problem of studying these processes, but also gave general dependences describing the mechanism of formation of high waters. He developed a theory of slope runoff. By examining the surface of a slope watered by a shower with an intensity variable in time and area, and writing a water balance equation for an elementary area, he derived general differential equations for the runoff of water from a slope. Converting to a river basin and adopting one or another concept of runoff losses in infiltration, M. A. Velikanov introduced the concept of isochronal lines of the travel time of water to a river. Without limiting himself to computations of slope runoff, he examined the movement of a high-water wave in a channel. He developed a method for computations in the case of summing the high waters of merging rivers. The results of all these investigations constituted an important landmark on the path of development of the general theory of runoff and the development of practical methods for computing and predicting snow-induced high water and rain-induced high water. M. A. Velikanov for the first time gave a thorough analysis of the processes of formation of snow- and rain-induced high waters. The fundamental ideas of M. A. Velikanov in the field of study of the water balance of basins and computations of river runoff were set forth by him in his monograph VODNYY BALANS SUSHI (Water Balance of the Land), published in 1940.

M. A. Velikanov gave great importance to the use of the methods of mathematical statistics in hydrology, especially in study of long-term fluctuations of river runoff. Being not only a theoretician, but also an experimenter, he devoted a very great amount of attention as well to the rigor of evaluations of the empirical material and the method for the processing of observations. His great experience in this problem important for research was set forth in the monograph OSHIBKI IZMERENIYA I EMPIRICHESKIYE ZAVISIMOSTI. MATEMATICHESKAYA OBRABOTKA NABLYUDENIY (Measurement Errors and Empirical Dependences. Mathematical Processing of Observations), published in 1962.

Among the scientific studies of M. A. Velikanov the problems of movement of sediments and channel processes unquestionably occupy an important place. The extensive complex of investigations of these problems, which was carried out by him and his students, was the basis for discriminating from the hydrology of the land an independent scientific discipline which M. A. Velikanov named "dynamics of channel flows." At its basis is a determination of the channel process as interaction between the flow and channel. The first edition of the monograph DINAMIKA RUSLOVYKH POTOKOV (Dynamics of Channel Flows) appeared in 1936 and the last in 1954-1955. The outstanding intuition of the natural scientist and his great work experience on investigating and improvement of river channels enabled M. A. Velikanov from the beginning of his scientific activity to define the problems through whose investigation lies the path to solution of the final problem -- computation and prediction of the channel process. These problems include the theory of channel turbulence, the theory of movement of sediments and the theory of the channel process.

M. A. Velikanov was the first to evaluate the role of turbulence as a determining factor in the complex of phenomena making up the channel process. Beginning with his early studies, he points out the need for investigation of turbulence in flows with a free surface and deformable boundaries on the basis of a synthesis of a rigorous physical experiment with the mathematical approach of theoretical hydrodynamics. A major cycle of studies carried out by M. A. Velikanov and under his direction was devoted to an experimental study of channel turbulence. There is a surprising systematic approach in the advance toward solution of this complex problem, which is characteristic for all his scientific activity, his capacity for defining ways to obtain the necessary experimental data, and also his ability for gathering empirical data accumulated by practical workers as a result of observations of sediments and channel deformations in rivers for corresponding theoretical generalizations.

The experimental studies of M. A. Velikanov were distinguished by a clear formulation of problems which must be solved by different experiments and a rigorous evaluation of the correspondence between the experimental apparatus and measurement method and the investigated problem. He demonstrated experimentally that the instantaneous values of the velocity of turbulent channel flow satisfy the law of normal distribution of random values. He brought attention to the fact that the idea of an instantaneous value of velocity in the form of the sum of its mean value and the deviation from this value arose as a result of the fact that turbulence was interpreted as a disruption of stability of laminar movement. A number of facts indicated that such a representation of the velocity field of a channel flow is incomplete. The dependence of the characteristics of velocity fluctuations on coordinates and the constancy of the Strouhal number in the entire flow section made it possible to postulate that in the flow there are quasiperiodic disturbances occupying its entire depth. On the basis of the spectral theory of turbulence developed in the studies of A. N. Kolmogorov, L. D. Landau, A. M. Obukhov and others it was established that in the frequency spectrum of velocity fluctuations the greatest part of the energy is concentrated in the low-frequency range. In a channel flow these low-frequency fluctuations must be related to the dimensions and form of its boundary surfaces. M. A. Velikanov had every basis for assuming that specifically these large-scale eddies play the principal role in the structure of the flow and in the mechanism of its effect on a deformable channel.

M. A. Velikanov postulated that in the instantaneous velocity of channel flow it is possible to discriminate a term expressing the structural deviation of velocity from its mean value. Such a representation of the velocity field afforded new possibilities for studying channel phenomena; then the center of gravity in the study was shifted from purely random deviations of velocity and the diffusion processes caused by them to the processes of formation of a channel and all its deformations, being a sort of "replica" of the structural deviations, which, in turn, were regularly related to channel macroformations. Thus, randomness of another nature was introduced -- randomness in the deviation of the parameters of structural disturbances from their mean values and randomness in the phenomena associated with the instability of large-scale forms. Structural formations with linear scales of the order of the depth of flow were discovered by the colleagues of M. A. Velikanov Ye. M. Minskiy and B. A. Fidman. Then M. A. Velikanov, in collaboration with N. A. Mikhaylova, demonstrated that turbidity fluctuations have a character close to periodic. They were associated with large-scale disturbances of the velocity field. The investigations of M. A. Velikanov and his associates made it possible to assert that it is precisely large-scale structural disturbances which are the mechanism determining the processes transpiring in a channel, and specifically the link through which there is a feedback between the flow and channel.

An entire cycle of studies by M. A. Velikanov is related to the development of a theory of movement of bottom and suspended sediments. His concepts concerning the mechanism of movement in a flow of solid particles was refined with the accumulation of experimental data. It should be noted that already by the time of the Second All-Union Hydrological Congress (1928) he had outlined the principles for solution of this problem. These were, first of all, the probabilistic nature of the phenomenon, since the flow transport of solid particles is associated with turbulence, and second, the unity of the entire process of transport of solid particles by a turbulent flow.

M. A. Velikanov and his associates in this field carried out a number of experimental investigations for a thorough clarification of the details of the mechanism of flow transport of solid particles: falling of heavy particles in a viscous fluid, movement of sediments in the bottom region of the flow, forming of sand waves, distribution of suspended sediments in the depth of the flow, statistical relationship between fluctuations of velocity and turbidity. M. A. Velikanov performed great services in the development of the theory of movement of bottom sediments. His studies made it possible to proceed from purely empirical relationships, at the basis of which there were only general hydraulic schemes, to the formulation of a theory making use of modern attainments in the turbulence field. M. A. Velikanov in his consideration of the phenomenon no longer limited himself to average characteristics of the flow, but drew upon the statistical characteristics of the field of velocity of turbulent flow. As a result he developed a probabilistic theory of jumplike movement of bottom sediments. This theory, in the sense of a probabilistic approach, is a development of the Kh. A. Eynshteyn theory. The probabilistic approach to consideration of the detachment of solid particles from the bottom and their further movement in the flow layer is the most promising from the point of view of further development of the theory of movement not only of bottom, but also suspended sediments. This approach already merits attention because it makes it possible to obtain not the relative value of the concentration of heavy particles (as this is done in different theories of the movement of sediments), but its absolute value.

In his investigations M. A. Velikanov devoted much attention to sand waves. He created a theory of the development of sand waves. The basis of the theory is the hypothesis that the movement of sand particles on the bottom arises under the influence of a pulsating flow velocity. Using the equation for the balance of solid matter for a plane, uniform, steady flow, M. A. Velikanov derived an expression for the probability of change in bottom readings with time, from which follows a periodicity of increases and decreases in bottom readings, that is, transformation of a plane sandy bottom into a wavy bottom under the influence of turbulent flow. He carried out a convincing and thorough analysis of the diffusion theory of movement of sediments and proposed a variant of this theory closest to the nature of the phenomenon. As a result of careful experimental checking of the theory N. A. Velikanov demonstrated that in application to flows with a low concentration of solid particles of a small hydraulic granularity the theory gives results extremely close to reality.

An important achievement in the field of movement of sediments was the gravitational theory proposed by M. A. Velikanov; he developed this in a series of studies. This theory, whose principles were published as early as 1943, in 1946, 1951-1952 was subjected to a critical discussion on the pages of IZVESTIYA AN SSSR (News of the USSR Academy of Sciences). As a result of the discussion a number of comments on the theory were taken into account by the author. At the present time it can be represented as a theoretical generalization and a final conclusion from numerous attempts of different authors to give a computation formula for the solid discharge of a river (discharge of sediments). After analyzing the empirical dependences, M. A. Velikanov concluded that in all the formulas there is a proportionality of the solid discharge to the fourth power of velocity and of the mean discharge concentration of sediments -- to the third power, which corresponds to the essence of the gravitational theory. Using the dimensionality method, the author draws two important conclusions: that the mean concentration of sediments, both suspended and bottom, is a function only of the roughness parameter and the criterion of the transporting capacity of the flow, which are basic in the gravitational theory. The inverse proportionality between the concentration and the criterion of transporting capacity relates to both suspended and bottom sediments. Thus, in the case of suspended and bottom sediments one and the same work of weighting appears, evidently being the principal factor in the channel process. Despite a number of inadequacies, the gravitational theory is a fundamentally new stage in study of the movement of sediments.

Probably with the deepening of our knowledge concerning the kinematics and dynamics of suspension-carrying flows new and more perfect theories will appear. But without question in one form or another they will retain an allowance for the "work of weighting" the need for which was first pointed out by M. A. Velikanov. He retains his priority also in developing kinematic dependences of suspension-carrying flows, in creating the first theory of the movement of suspended sediments taking into account the effect exerted on them by gravity and free from restriction on the smallness of the hydraulic fraction and the smallness of the disturbances transporting the solid particles.

An independent cycle of studies by M. A. Velikanov is represented by his investigations of the channel process as a whole. In developing the channel process science he generalized the qualitative observations and practical conclusions drawn by research engineers of the last century V. M. Lokhtin, N. S. Lelyavskiy, L. Farg and others.

In connection with the construction of major hydroelectric power plants on the Dnepr and Volga for the first time the problem arose of predicting and regulating the erosional activity of water flows -- prediction of the channel process. Due to the studies of M. A. Velikanov our knowledge of the channel process was substantially broadened and deepened. He laid the basis for solution of the channel process problem as an integrated whole. An analysis and generalization of the enormous material from field investigations and laboratory experiments enabled M. A. Velikanov to give three principles which he considered to be guiding in study of the channel process.

The first principle is interaction between the flow and channel, expressing the dialectic unity of mutually operative elements: flow of fluid and constantly deformable solid channel boundaries, consisting of individual discrete particles, periodically penetrating into the thickness of the flow, transported by the latter and then again returning to the bottom. The essence of this principle is that the channel constantly controls the flow and the flow controls the channel. M. A. Velikanov regarded the channel and flow to be "one organically joined complex in which the channel reflects the form of the flow and the flow reflects the form of the channel." In connection with this postulate he formulated the problem of study of the specifics of turbulence, associated with elements of channel form (shallows, rapids, etc.). The channel stability problem is closely related to this principle. M. A. Velikanov proposed an integral evaluation of channel stability using the modified Lokhtin stability test. He proposed a classification of rivers by degree of channel stability.

The second principle is a restriction on the number of natural complexes. In natural channel flows, as a result of prolonged interaction between the flow and channel, it was possible to define a special type of dependence determining the most probable combinations of morphometric characteristics of the channel and sediments (width, depth, curvature, size of particles) and the hydrodynamic parameters of the flow (slope, discharge). With stipulated geological and soil-botanical conditions there is a restriction on the number of possible types of channel flows; this facilitates the forecast and regulation of the channel process. The considered principle lies at the basis of writing of so-called morphometric dependences. By developing the work of V. M. Lokhtin and using the dimensionality method, M. A. Velikanov obtained morphometric dependences -- channel criteria. Such dependences were earlier derived empirically by S. I. Rybkin.

The third principle is the minimum of energy dissipation (channel of least resistance). The physical sense of this principle applicable to the dynamics of channel flows is that in the formation of the river channel (primary or disrupted by the effect of a structure) the totality of channel deformations transpires in the direction of such a combination of channel forms and currents with which the total resistance of the flow becomes minimum. The reciprocal control of the flow and channel after a number of successive deformations leads precisely to this: among the hydrodynamically possible and geologically realizable forms there is a form for which the velocity structure of the flow gives a minimum level of energy dissipation. In the opinion of M. A. Velikanov, this principle must be especially productive in solving some special engineering problems when the flow is subjected to some external effect by the erection of some structure in the flow which disrupts its usual form.

To be sure, the formulation of the three enumerated principles is only the beginning of development or even an indication of the direction in which channel process theory can develop. But even only the formulation of the mentioned postulates provided a key to the development of the principles for rational effect on rivers for the purpose of improving navigational conditions, in the construction of bridges, water intakes, etc.

M. A. Velikanov devoted much attention to the problem of laboratory modeling of channel processes. The first models of microrivers in the USSR were created by N. S. Sharashkina under his direct direction. The problems involved in the transformation of river channels were studied in the laboratory in models with natural ground at the end of the 1940's-early 1950's. The experiments were initially carried out in small areas but later were transferred to open models in the floodplain of the Moskva River. The results made it possible to clarify the pattern of development of a river, ranging from primary furrows to complex branching channels with a well-expressed valley. They demonstrated that the morphometric characteristics of natural rivers -- the terraces, shallows, rapids, islands, ox-bow lakes -- are qualitatively completely reproduced in the model.

The principles formulated by M. A. Velikanov serve as a basis for the modeling of channel processes in erodable models. Relying on the principle of limitation of natural complexes and the derived morphometric dependences, which hypothetically were correct for both large and small rivers, M. A. Velikanov succeeded in establishing and validating the scales which a model must satisfy so that processes in it would transpire similar to natural processes. It is known that in the modeling of the channel process for hydraulic or other purposes it is necessary to retain distortion of the vertical scale in the model of turbulent flow. M. A. Velikanov for the first time was able to give computations of such a distortion, which made it possible to determine, with conversion from nature to the model, the criterion of transporting capacity, concentration of sediments and duration of the channel process.

The broad front of work carried out in the Soviet Union on the regulation and improvement of rivers, the creation of large-scale irrigation systems and the development of projects for the redistribution of water resources over an area, associated with the creation of canals which in their size and water-carrying capacity are commensurable with major rivers, require the development of increasingly more rigorous methods for predicting and computing channel phenomena, deformation of the canal channels and determination of the final stable result of transformation of artificial watercourses. The theoretical principles for engineering solution of such problems were laid in the studies of M. A. Velikanov.

The following questions lie within the framework of development of the ideas of M. A. Velikanov: investigation of the peculiarities of channel turbulence, development of hypotheses of closing of the averaged equations of hydrodynamics, taking into account the specifics of channel turbulence; investigation of secondary currents and their role in the development of structural macroscale formations in the flow; development for this purpose of reliable devices for measuring velocity for actual flows; seeking of the relationship between the characteristics of structural disturbances and channel formations; development and refining of the theory of individual kinds of movement of sediments and developing a unified theory, including all types of movement of sediments. Finally, the formulation of a quantitative theory of channel processes: development of a theory of nonstationary phenomena in a channel, the writing of morphometric dependences.

By virtue of historically developing factors Soviet hydrodynamics of river flow developed along its original path both in the field of experimentation and in the field of theory. And M. A. Velikanov played a leading role in the fact that it developed as an important component of the scientific-theoretical principles of enormous hydroengineering construction work, work on the regulation and improvement of rivers. The ideas of this outstanding scientist were fundamental in the development of the Soviet science of channel processes.

M. A. Velikanov devoted much attention to pedagogic work. Over a number of years he presented a number of courses to students at the Moscow Hydrometeorological Institute and, in particular, a course on hydrology of the land, and then a course on the water balance. At the Physics Faculty of Moscow State University M. A. Velikanov presented a course on the dynamics of channel flows.

REVIEW OF MONOGRAPH BY A. I. FAL'KOVICH: DINAMIKA I ENERGETIKA VNUTRITROPICHESKOY ZONY KONVERGENTSII (DYNAMICS AND ENERGY OF THE INTERTROPICAL CONVERGENCE ZONE), LENINGRAD, GIDROMETEOIZDAT, 1979

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 118-119

[Review by M. A. Petrosyants, Chairman of the GATE Working Group of the Soviet GARP Commission]

[Text] In world meteorology during the last decade ever-increasing attention has been devoted to the problems in tropical meteorology. The number of articles on tropical meteorology now is already in the hundreds. It is also not without reason that the first experiment carried out under the GARP program was the Atlantic Tropical Experiment (GATE). This is attributable to the fact that the tropics, occupying almost half the earth's area, is the principal supplier of heat and moisture to the atmosphere. Despite such broad attention to the problems of tropical meteorology, this region for the time being remains the least studied and the number of Soviet books on tropical meteorology obviously lags behind foreign publications.

In this sense the reviewed book is a highly interesting study. First of all, the very subject of the investigation is interesting: the intertropical convergence zone. Together with the Trades, it is the most important feature of circulation in the tropics. Second, it makes extensive use of observational data from the Atlantic Tropical Experiment, unquestionably the most grandiose and well-prepared experiment during the entire time of existence of international meteorology.

The book by A. I. Fal'kovich is an extremely literate critical review of modern points of view concerning the ICZ, its structure, atmospheric stability in it, development of wave disturbances, modeling of the ICZ. In this connection it gives a thorough analysis of the theory of "conditional instability of the second kind" which is used widely in the west and explains its true value and degree of applicability for describing the behavior of the tropical atmosphere. Taking into account that a great number of studies in this field have recently appeared in the English language and as yet are quite unfamiliar to a broad circle of Soviet meteorologists, such a compressed and literate analysis of the results attained by science is undoubtedly extremely valuable.

The book has been divided into three parts: "Intertropical Convergence Zone and the Tropical Atmosphere," "Basic Principles of Processing and Interpretation of Experimental Data in the Tropics" and "Dynamics and Energy of the ICZ on the Basis of Mean (by Class or Time) GATE Data."

The first part introduces the reader into a variety of concepts concerning the tropical atmosphere and about the place occupied by the ICZ in it. The author considers in detail the principal properties of the ICZ, the conditional atmospheric instability of the first and second kind, and also convective instability. Here a contribution of the author is a literate physical analysis of the actual value of the theory of conditional instability of the second kind and also the paradoxical position expressed recently by American specialists concerning the great potential instability of the cloudless tropical atmosphere in comparison with a center of cloud concentration. The author convincingly demonstrates that the latter position is incorrect.

This same part gives a concise exposition of the hypothesis of "hot towers" and the hypothesis of critical latitudes. The author also dwells on modern concepts concerning waves in the tropical atmosphere and their relationship to the ICZ. This part is completed by a critical review of existing numerical models of the ICZ.

The second part sets forth the principles of processing and methods for computing dynamic, thermodynamic and energy characteristics in the tropics and also the peculiarities of interpretation of experimental data applicable to the tropical atmosphere. This problem is of independent importance, taking into account the properties of the atmosphere in the tropics and their difference from the temperate latitudes. The computation methods proposed by the author can be applied in group shipboard expeditionary investigations of the oceans.

The third part gives a detailed analysis of the dynamics of atmospheric movements on the basis of sections of the tropical atmosphere constructed on the basis of GATE data for its stable and unstable states. An objective classification of atmospheric processes in the tropics is proposed with use of the methods of discriminant analysis. This part of the study is original and is based for the most part on the studies of Soviet scientists, including the author of the book.

The central place in the book is undoubtedly occupied by the chapter devoted to ICZ energy. In this chapter, on the basis of GATE observations, a study is made of the heat balance of the ocean surface, water temperature and energy balance in the ICZ. The quantity of precipitation is computed, the role of the ICZ in the energy of the tropical atmosphere is evaluated and the interaction of movements of different scales is analyzed. Here a number of very important and interesting conclusions are drawn, for example, that the ICZ, with a width of 3°, yields moisture collected from an area of the sea surface with a width of 30° in the Trades zone, that is, that the ICZ is not an exotic strip of well-developed cloud cover, but a very important circulation feature of the atmosphere, exerting an influence on moisture transfer from the tropics into the temperate latitudes.

The book is written in good language, at a high theoretical level and at the same time is not overburdened by mathematical formulas. Being a master of the mathematical approach describing the dynamics of the tropical atmosphere, the author was able to make his conclusions physically obvious without having recourse to excessively complex mathematical calculations.

The publication of this book DYNAMICS AND ENERGY OF THE INTERTROPICAL CONVERGENCE ZONE by A. I. Fal'kovich represents a definite stage in the scientific assimilation of observations of the Atlantic Tropical Experiment and indisputably will serve as a significant contribution of Soviet meteorologists to the international scientific literature on this subject.

CONCERNING THE REVIEW OF S. G. RUSTAMOV AND S. M. FLEYSHMAN OF THE COLLECTION OF ARTICLES "SELEVYYE POTOKI" ("MUDFLOWS") (MOSCOW, GIDROMETEOIZDAT, 1976)

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 119-121

[Article by Candidate of Geographical Sciences V. P. Mochalov, T. L. Kirenskaya, Candidate of Technical Sciences B. S. Stepanov and A. E. Zems, submitted for publication 20 April 1979]

[Text] METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 5, 1978, contained a review by S. G. Rustamov and S. M. Fleyshman of collection of articles No 1, published by the Kazakh Scientific Research Hydrometeorological Institute, SELEVYYE POTOKI (Mudflows).

While admitting the high scientific value of large-scale experiments with artificial mudflows in the basin of the Chemolgan River (the results of which are published in the collection of articles), the reviewers nevertheless accuse the authors of the articles of "invalid generalization and extremely arbitrary interpretation of observational data," and the editor of the collection, Yu. V. Vinogradov, "of assertions contradicting the nature of formation of mudflows, and in individual places even the actual data from the reported experiments."

The principal objections of the reviewers relate to two conclusions drawn in the article by Yu. B. Vinogradov: on the characteristic density of mudflows and on the role of obstructions in the wave regime of movement of mudflows.

A discussion of these problems, of importance in the scientific study of mudflows (determining the legitimacy of different computation methods) could only be welcomed if it had a more justifiable character. Unfortunately, the review is not sound. For example, the reviewers "refute" the first of the mentioned conclusions of Yu. B. Vinogradov only by the following assertion: "Tens and hundreds of field measurements and computations of the density of mudflows, widely known from the literature, make it entirely obvious that the density of mudflows varies in dependence on a whole series of factors from 1.2-1.3 to 1.9-2.1 tons/m^3, but the author counters this with only two computations for artificial mudflows, whose methodological legitimacy is also doubtful, so that the upper limit of mudflow density will indicate a common pattern for all mudflows. This assertion by Yu. B. Vinogradov can cause nothing but perplexity among mudflow specialists."

But the fact is that the "hundreds" of estimates of mudflow density known from the literature (for example, all estimates of the density of Alma-Ata mudflows during the years 1921-1950) were not based on instrumental measurements, but on purely speculative conclusions drawn in an inspection of mudflow "tracks." The instrumental measurement of mudflow density involves great difficulties and attempts in this direction have not been in the "hundreds" or even the "dozens;" there have only been a few. I. I. Kherkheulidze, specially concerned with the collection of data from field measurements of mudflows, cites a list [3], from which it can be seen that the density of natural mudflows has been measured in only two basins: on the Durudzhi River and on the Kok-Cheka River; the measurements pertained to nine mudflows. The mean density of the mudflow samples in all cases was close to 2 tons/m^3; rare deviations were in the range 1.8-2.25 tons/m^3. The imperfection of the measurement method (digging with a pail) inevitably led to an understatement of density since the samples were taken from the most fluid part of the flow and did not include large fractions.

The experiments in the Chemolganskiy polygon for the first time afforded a possibility for a more reliable determination of mudflow density. With use of the most rigorous (balance) method, and also other procedures (pressure sensors, magnetometric apparatus), the mean density of mudflows was found to be 2.04 and 2.07 tons/m^3, and during individual intervals -- 2.3 tons/m^3. During this time data were obtained on the density of the Maloalmatinskiy mudflow of 1973 (2.39 tons/m^3), which was completely held back by the Medeo dam, and the Issykskiy mudflow of 1963 (2.4 tons/m^3). The enumerated values indicate that the conclusion drawn by Yu. B. Vinogradov about the formation of mud-rock flows with a high density is by no means based on two "doubtful" calculations. Also not corresponding to reality is the assertion of the reviewers that Yu. B. Vinogradov has extended this conclusion to all mudflows. In the article by Vinogradov published in this collection of articles, entitled "The Erosional-Displacement Mudflow Process," he gives a classification scheme, according to which a high density is characteristic only of mud and rock flows, in contrast to alluvium-water mudflows, forming under other conditions, and whose density is determined only by the transporting capacity of the water flow.

Thus, the attempt of the reviewers to refute the first conclusion in the article by Yu. B. Vinogradov is unsuccessful.

The second, "especially serious error" and "dangerous delusion" of Yu. B. Vinogradov, in the opinion of the reviewers, "...is the denial of the obstructional nature of the movement of mudflows." Along these lines, it is necessary to note the following.

The suddenness of passage of mudflows and absence of measurements were the reasons why over the course of a prolonged period the development of scientific concepts concerning mudflows occurred under conditions of an almost total absence of observational data. Mudflows were studied exclusively from the tracks of their movement, as a result of which the lacking information on the mechanism of formation and dynamics of movement of mudflows was supplied by pure speculation. However, one of the most viable hypotheses of this kind was the concept of the obstruction nature of mudflows.

Meeting with descriptions of a wall-like movement of mudflows and observing the traces of marked fluctuation of the maximum levels of flow along the length of a valley, researchers attributed this phenomenon to the bursting of obstructions, by which is meant "barriers encumbering the channel," caused by landslides and talus from the sides, gravitational rock concentrations and other natural embankments. The hypothesis of "bursting of obstructions" was widely disseminated and as a direct consequence there were engineering recommendations on the straightening and cleaning up of mudflow channels, "preventive removal of obstructions, encumbrances, gravitational rock concentrations in channels" [2] facilitating the formation of barriers in the path of mudflows.

However, with a more thorough study of mudflow phenomena it became clear that the mentioned hypothesis is not confirmed. The first serious investigations of this matter, carried out by G. V. Ivanov in 1952-1956 by means of artificial reproduction of mud-rock flows, led to the conclusion that the formation of waves in a mudflow does not occur due to the bursting of obstructions, but in the process of mudflow movement [1]. The stationary observations in mudflow basins of the Durudzhi and Kok-Cheka Rivers carried out in subsequent years also confirmed this conclusion. During this same period a more thorough study was made of the catastrophic mudflows occurring in the territory of the country. However, up to the present time we do not know of a single example of observation of obstructions in the interpretation of the reviewers (despite the assertion of the latter "... an extensive literature is devoted to obstructions, which Yu. B. Vinogradov ignores..."). Experiments at Chemolgan, making it possible to observe the entire mechanism of mudflow formation in a typical mudflow "focus," made possible a final refutation of outdated concepts. Observation of the wave regime of movement of mudflows at Chemolgan in the absence of any obstructions is an obvious demonstration that obstructions are not a necessary condition for the appearance of a wave regime and that the reasons for the latter are related to the internal dynamics of the process (similar to the phenomenon of travelling waves in rapid flows). The reviewers do not agree with this and insist that the "formation of obstructions in mountain channels is one of the most important peculiarities characteristic of mudflows," that although "obstructions are not the only factor responsible for the wave character of mudflow movement, their role is exceptionally great." In demonstration that they are correct they cite a not entirely convincing example of the bursting of the Mynzhilki dam on the M. Almatinka River in 1973 as a result of a catastrophic glacial-breakthrough flood. "It is generally known," write the reviewers, "that the unstable Mynzkilki gabion dam... was burst, that there was a marked increase in the magnitude and power of the flow." In order to back up their arguments they cite an article from the collection devoted to the mudflow of 1973 and assert that the authors of this article, allegedly, "confirm the fact that the gabion dam was reached, for the most part, by a water flow, which after bursting of the dam was transformed into a mudflow, and that the magnitude of the flow after bursting of the obstruction (the dam) increased."

A natural question arises: what relationship to the role of obstructions in the wave dynamics of mudflows can have the example of bursting of a water-holding dam? What do the reviewers wish to convince the readers of: that the bursting of dams results in a catastrophic increase in water discharge and erosion? There is no need to demonstrate this.

In the article on the mudflow of 1973, published in the collection of articles, to which the reviewers refer, the opposite is demonstrated: there was no increase in the volume or maximum discharge of a mudflow-forming high water as a result of the bursting of the Mynzhilki dam. Observational data and computations are cited refuting the earlier published version of S. M. Fleyshman that this dam supposedly played a fateful role in forming the mudflow. Passing over these proofs in silence, the reviewers extract from the article only a reference to the effect that in the absence of the dam the first (but not the maximum) mudflow wave could be lower, and the fact that above the dam there was no mud and rock mudflow, whereas below it did develop. It must be emphasized that exaggeration of the role of obstructions in the movement of mudflows in actuality is a dangerous error. From such a point of view the maximum discharges of mudflows are not subject to any prediction at all because it is unthinkable to calculate the parameters of bursting of hypothetical obstructions capable of forming in the channel due to talus and the collapse of slopes. The assertion that it is specifically such obstructions which explain the absence of a direct proportionality between the area of a basin and the parameters of mudflows diverts one's attention from study of the real reason for this discrepancy, involving, most frequently, the possibility of the appearance of a purely displacement mechanism of mudflow formation. Relying on the point of view of the reviewers, the engineer and planner, in the field encountering a linear mudflow channel (mudflow "focus"), are forced to assume that the form of the mudflow hydrograph will be dependent, for the most part, on the intensity of the runoff from the drainage basin of the mudflow "focus" or the mudflow drainage basin. In actuality, the linearity of the mudflow channel does not exclude, in our opinion, but on the contrary, causes a marked exceeding of the extremal values of the discharges of the mudflow over the average.

The questions raised in the review are of considerable practical importance and this is why we have published these comments.

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SIXTIETH BIRTHDAY OF MIKHAIL IVANOVICH BUDYKO

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 p 122

[Unsigned article]

[Text] Corresponding Member USSR Academy of Sciences Mikhail Ivanovich Budyko marks his 60th birthday on 20 January 1980. An outstanding Soviet scientist, he is one of the creators of modern physical climatology, the scientific director of investigations on the problem of changes in climate, a member of the editorial board of the journal METEOROLOGIYA I GIDROLOGIYA, an honorary member of the American Meteorological Society, a winner of the Lenin Prize.

The board of the USSR State Committee on Hydrometeorology and Environmental Monitoring congratulates Budyko on this noteworthy date and wishes him good health and new creative successes.

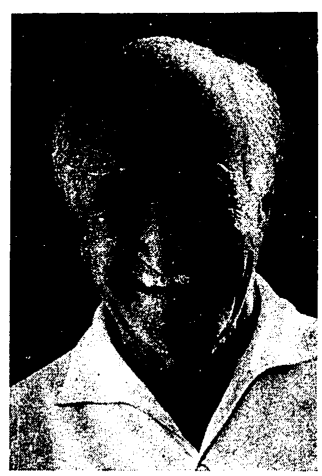
SEVENTIETH BIRTHDAY OF ALEKSANDR KHRISTOFOROVICH KHRGIAN

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 123-124

[Article by students, friends and colleagues]

[Text] Aleksandr Khristoforovich Khrgian marks his seventieth birthday on 21 January.

The name of A. Kh. Khrgian is inseparably associated with Soviet meteorology. He began his work activity 50 years ago in the Hydrometeorological Committee of the Council of People's Commissars of the RSFSR (the predecessor of the State Committee on Hydrometeorology), where he was assigned after graduation from the physics and mathematics faculty of Moscow State University. Productive scientific work, active teaching and scientific-public activity earned him universal respect and broad fame not only in the USSR, but also abroad.



In 1936 Aleksandr Khristoforovich defended his Candidate's dissertation on the climatology of precipitation. At the same time he carried out interesting investigations of blizzards and snow drifts.

Having good theoretical preparation, Aleksandr Khristoforovich ably combined a general physical theoretical approach to investigated phenomena and experimental observations. This remarkable quality was already clearly manifested in his earliest studies on investigation of permafrost in which laboratory and expeditionary investigations were closely combined.

Since the late 1930's Aleksandr Khristoforovich has actively studied different aspects of mountain meteorology. Without questión, his involvement with this subject matter was closely associated with the particular love of Aleksandr Khristoforovich for the mountains and his passionate devotion to alpinism. Together with his students, A. Kh. Khrgian carried out a number of expeditions in which he collected unique material on the peculiarities of air circulation in mountain regions, on clouds and the temperature and humidity fields.

Since 1945 A. Kh. Khrgian has been a specialist at the Central Aerological Observatory. There, under his direction, was born the Moscow school of experimental cloud physics which was later developed by his students A. M. Borovikov and others. Studies in this field were generalized in the widely known monograph FIZIKA OBLAKOV (Cloud Physics), published in 1961. Aleksandr Khristoforovich was the initiator of its writing, was the editor and one of its authors. The first Soviet Cloud Atlas, a highly important aid for meteorological observers, was prepared under his direction in 1957.

Beginning in the 1960's, Aleksandr Khristoforovich has devoted his main attention to the physics of atmospheric ozone. In 1973 he published a generalizing monograph on this problem, and in 1979 he published the book entitled SOVREMENNYYE PROBLEMY ATMOSFERNOGO OZONA (Present-Day Problems of Atmospheric Ozone), which he wrote in collaboration with his student S. P. Perov.

The teaching activity of A. Kh. Khrgian has been extensive and diversified. Beginning in 1934 he was a docent at the Moscow Hydrometeorological Institute. In the stormy years of the Great Fatherland War, he, working at the Higher Military Hydrometeorological Institute, devoted many efforts to the training of military meteorologists. There he completed and in 1943 defended his doctoral dissertation, becoming a major capital work on the history of meteorology, which he published in 1948 and republished in 1959. This publication, like the books on cloud physics and atmospheric ozone, was translated into English and published abroad. Love for the history of science was one of the vital interests of Aleksandr Khristoforovich. There are numerous well-known articles, reviews and sketches devoted to the biographies of leading Soviet and foreign meteorologists which were written with love by the pen of A. Kh. Khrgian.

Beginning in 1947 Aleksandr Khristoforovich was a professor in the Physics Faculty of Moscow State University. There he established the first course on physics of the atmosphere in our country. Over a period of 25 years the monograph of A. Kh. Khrgian entitled FIZIKA ATMOSFERY (Atmospheric Physics) went through four editions. It is difficult to find a meteorologist in our country who has not attended the lectures of Aleksandr Khristoforovich or has not studied from his textbooks.

The enormous scientific and teaching activity of Aleksandr Khristoforovich is combined with tireless public-scientific activity. For many years he was a member of the executive committee of the International Association of Meteorology and Atmospheric Physics, as well as international commissions on cloud physics and atmospheric ozone. He is still working on the ozone commission. He is deputy chairman of the section on meteorology and atmospheric physics of the Interdepartmental Geophysical Committee of the Presidium USSR Academy of Sciences, a member of the editorial board of the Hydrometeorological Publishing House and the editorial board of the journal IZVESTIYA AKADEMII NAUK SSSR, FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), a member of the scientific councils of a number of institutes, etc.

A. Kh. Khrgian has prepared more than 30 Candidates of Sciences. Tens of Doctors of Sciences regard themselves as his students. And today, as before, he is at the forefront of science. During the last 10 years alone he has published more than 30 scientific articles and 4 monographs, not counting numerous reviews, book reviews and sketches. He is filled with energy and creative plans.

In congratulating Aleksandr Khristoforovich on his memorable anniversary we want to wish him good health and the same active and productive work for long years.

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EIGHTIETH BIRTHDAY OF ALEKSANDR BOLESLAVOVICH KALINOVSKIY

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 124-125

[Article by P. I. Smirnov, G. G. Tarakanov, D. L. Laykhtman, L. G. Kachurin, N. Z. Pinus and M. A. German]

[Text] On 15 November 1979 Aleksandr Boleslavovich Kalinovskiy, the well-known Soviet professional aerologist, marked his 80th birthday.

Aleksandr Boleslavovich was born in Kaluga. During the years 1919 through 1922 he served in the Red Army and participated in the Civil War.

Kalinovskiy began his scientific activity in 1926, when he, still being a student at Moscow State University, proceeded to work as a scientific specialist (second level) at the aerological observatory in the State Scientific Research Geophysical Institute (Moscow). In 1927 Kalinovskiy graduated from the Physics-Mathematics Faculty of Moscow State University in the "geophysics" field of specialization. In 1930 he was named to the post of senior scientific specialist in the aerological observatory. After reorganization of the Geophysical Institute (1934) Kalinovskiy became head of the aerological observatory of the Central Institute of Experimental Hydrology and Meteorology. In a subsequent reorganization he was transferred to the post of senior scientific specialist in the aerology division of the Central Institute of Forecasts, where he worked until 1938.

The professional meteorologists V. I. Vitkevich and V. A. Khanevskiy, well known at that time, exerted a great influence on the formation of Kalinovskiy as a scientist. On the advice of Khanevskiy Kalinovskiy turned to the scientific problems of physics of the free atmosphere and methods for aerological investigations, which for many years later determined the direction of his scientific activity. Kalinovskiy made a whole series of important scientific-practical studies on aerosynoptic analysis of cyclones and on the use of aerological sounding data in weather analysis and forecasting (in collaboration with V. P. Nekrasov and Kh. P. Pogosyan).

Kalinovskiy defended his Candidate's dissertation in 1941 on the basis of the results of long-term investigations of the temperature-wind regime of the free atmosphere over Moscow.

The scientific activity of Aleksandr Boleslavovich was not limited to within the walls of the aerological observatory. Over a period of many years he was an active participant in different scientific expeditions: for study of the structure of breeze winds on the Kerch Peninsula (1924), for investigating the density of sea ice and evaporation in the Barents Sea aboard the scientific research ship "Persey" (1926) and on use of luminescent trails for studying stratospheric winds (1936).

In 1933 Kalinovskiy worked at the Institute of Dirigible Construction, where, in particular, he investigated problems relating to the meteorological support of flights of high-altitude balloons and dirigibles. In 1934 he participated in the meteorological support of launching of the stratospheric balloon "Osoviakhim-1" (Professor A. P. Molchanov was responsible for the meteorological support).

Kalinovskiy combined all his scientific activity successfully with teaching work. During 1932-1933 he served as an instructor at the technical school of the Civil Air Fleet, during 1934-1945 he was a docent at the Moscow Hydrometeorological Institute and since the year 1946 he has been a docent in the department of aerology and dynamic meteorology at the Leningrad Hydrometeorological Institute. All of the post-war scientific-pedagogic activity of Kalinovskiy is associated with the Leningrad Hydrometeorological Institute. Under his direction the students at the Leningrad Hydrometeorological Institute defended tens of course and diploma projects.

A result of the many years of scientific-pedagogic activity of Kalinovskiy was the academic aid written by him in collaboration with N. Z. Pinus entitled AEROLOGIYA (Aerology) (1951). In 1961 these same authors published a textbook on aerology for students at hydrometeorological colleges and universities.

At the present time Kalinovskiy is on a merited rest. We wish Aleksandr Boleslavovich good health and many more years of life.

NOTES ON ACTIVITIES AT THE USSR STATE COMMITTEE ON HYDROMETEOROLOGY AND ENVIRONMENTAL MONITORING

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 p 125

[Article by A. V. Kolokol'chikov]

[Text] The State Committee on Hydrometeorology has planned the holding of a number of all-union scientific and scientific-technical meetings, conferences and symposia for 1980.

In the plan much attention is being devoted to the problems involved in the study of climate, prediction of its changes and variations, and also preservation of the environment.

For example, in February at the Main Geophysical Observatory at Leningrad there will be a conference on the subject "Modeling of Climate, its Changes and Variations." Among the conference participants will be organizations of the State Committee on Hydrometeorology, USSR Academy of Sciences and the USSR Ministry of Higher Institutions of Education.

During March, at the State Hydrological Institute, there will be an All-Union Conference on the Problem of Anthropogenic Change of Climate. Specialists of the State Committee on Hydrometeorology and the USSR Academy of Sciences will participate in the work of the conference.

In March plans call for holding a conference on "Means and Prospects for the Development of Work on Monitoring Environmental Contamination," to be held at the All-Union Exhibition of Achievements in the National Economy. The Institute of Applied Geophysics will be responsible for carrying out this conference. Among the participants will be institutes of the State Committee on Hydrometeorology, Ministry of the Chemical Industry, Ministry of the Petroleum Industry and Ministry of Power.

In October, at Obninsk, the Institute of Experimental Meteorology will carry out a symposium on "Influence of Environmental Contamination on Change of the Earth's Climate." In addition to institutes of the State Committee on Hydrometeorology, the symposium will be attended by representatives of institutes of the USSR Academy of Sciences and the USSR Ministry of Higher Institutions of Education.

During this same month at the All-Union Exhibition of Achievements in the National Economy the All-Union Scientific Research Institute of Agricultural Meteorology will hold a conference on "Hydrometeorological Support of Agriculture in the Nonchernozem Zone of the RSFSR." In addition to the State Committee on Hydrometeorology, the institutes of the USSR Ministry of Agriculture will participate in the conference.

In November, also at the All-Union Exhibition of Achievements in the National Economy, the Institute of Applied Geophysics will hold a conference on "Status and Prospects of Development of Prediction of Changes in the Environment Under the Influence of Construction and Operation of the Second Stage of the Kansko-Achinskiy Fuel-Power Complex." The conference will be attended by scientists and specialists of the State Committee on Hydrometeorology, USSR Power Ministry, USSR Coal Industry Ministry and the Siberian Department USSR Academy of Sciences.

In November plans also call for carrying out a conference on "The Nature of the Arctic Under Conditions of Interzonal Redistribution of Water Resources" at the Arctic and Antarctic Scientific Research Institute. The participants will include specialists from the State Committee on Hydrometeorology, USSR Academy of Sciences, USSR Ministry of the Merchant Marine, Ministry of Higher Institutions of Education RSFSR, USSR Ministry of Higher Institutes of Education, USSR Water Management Ministry, USSR Fishing Ministry, Ministry of the River Fleet RSFSR.

Finally, in November, at Obninsk, the Central Design Bureau of Hydrometeorological Instrument Making will hold a conference on "Technical Means for the State System of Observations and Monitoring of the Environment and Climate."

CONFERENCES, MEETINGS AND SEMINARS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 125-127

[Article by L. G. Zastavenko and L. A. Chubukov]

[Text] An all-union symposium on "Physical Principles of Change in Modern Climate" was held in Moscow during the period 23-25 April 1979. The objective of the symposium was an exchange of knowledge on changes and fluctuations of modern climate and study of their physical causes.

The symposium was attended by 202 specialists from the USSR Academy of Sciences and the union republics, the State Committee on Hydrometeorology and Environmental Monitoring, universities and other colleges and organizations arriving in Moscow from 29 cities of the Soviet Union.

At the symposium opening addresses were presented by the Chairman of the Moscow Affiliate of the Geographical Society I. D. Papanin and the First Deputy Chairman of the State Committee on Hydrometeorology and Environmental Monitoring Yu. S. Sedunov.

In the 52 reports heard at the symposium, presented by representatives of 34 institutes of different departments, information was given on three different themes: factors in the formation and change of climate, natural changes in climate, anthropogenic changes in climate, including changes in urban climate.

Among the numerous geophysical factors which can determine modern changes and fluctuations in climate, at the symposium attention was given to changes in the solar constant, transformation of properties of the underlying surface, changes in the gas and aerosol composition of the atmosphere (K. Ya. Kondrat'yev, M. I. Budyko, K. Ya. Vinnikov, O. A. Drozdov, N. A. Yefimova); fluctuations in the short-wave part of UV radiation and X-radiation, solar corpuscular radiation, solar wind, changes in the sign of the interplanetary magnetic field (IMF) caused by solar and solar-planetary cycles (L. R. Rakipova, V. F. Loginov and B. I. Sazonov; V. A. Belinskiy, A. A. Dmitriyev, V. D. Reshetov, B. A. Sleptsov-Shevelevich, V. N. Chepurnoy, V. N. Plakhotnyuk); fluctuations of the earth's rotational regime (F. I. Rudyayev); autooscillatory processes in the atmosphere (O. A. Drozdov, V. D. Reshetov, O. P. Chizhov); disruption of the codimensionality between heat and moisture resources (V. S. Mezentsev); changes in atmospheric circulation (M. Kh. Baydal and A. I. Neushkin).

K. Ya. Kondrat'yev drew attention to what for the time being is inadequate understanding of the physical, chemical and biological processes exerting an influence on climatic changes and the necessity for clarification of the contribution of anthropogenic factors. M. I. Budyko, et al. demonstrated that with conformity of the increase in the atmospheric content of CO_2 in dependence on fuel combustion to an exponential law of maintenance of the present-day rates of energy development there should be an appreciable increase in the observable air temperature in individual regions, already by the end of the current century. In the opinion of Ye. P. Borisenkov, the existing models of an increase in CO_2 content are in need of improvement and therefore at the present time science is capable of giving only qualitative estimates of the trends of the effect of different anthropogenic factors on climate.

A. A. Dmitriyev investigated the mechanisms of the relationship between the changes in the density of the flux of X-radiation and also sectors of the IMF and the climatic characteristics of state of the atmosphere. The model computations of L. R. Rakipova, V. F. Loginov and B. I. Sazonov give a quantitative evaluation of the effectiveness of the ozone mechanism. According to B. A. Sleptsov-Shevelevich, the fluctuations of solar activity are transmitted to the earth and its atmosphere through the magnetosphere, in the final analysis causing 7- and 22-year variations of the pressure field and the instantaneous axis of the earth's rotation.

According to V. D. Reshetov, the propagation of the solar wind is influenced by the positioning of the planets. V. N. Plakhotnyuk detected frequency-dependent unstable matched fluctuations of different parameters of solar activity with fluctuations of geomagnetic activity, climatic characteristics of the state of the atmosphere and mutual displacement displacements of the axes of rotation of the core, mantle and atmosphere.

It is noted that UV radiation shorter than 315 nm (UV-B) changes far more under the influence of anthropogenic effects than the integral flux. F. I. Rudyayev gave a theoretical validation of the relationship of the change in the earth's rotational regime and change in the pressure field. N. M. Svatkov proposed a new method for determining the earth's planetary albedo, based on use of the method of orthographic cartographic projections together with satellite data on the coverage of the earth with clouds and the radiation influx at the upper boundary of the atmosphere.

The reports of L. I. Sverlova, V. M. Zhukov, V. S. Mezentsev and A. A. Velichko presented the paleogeographic direction in investigations of climatic changes. In particular, A. A. Velichko, on the basis of a comparison of the conditions prevailing in the last glaciation and the optimum of the last interglacial period with the conditions of today drew the conclusion that the optimum of the present interglacial period has already passed.

The reports in the second direction discussed the tendencies in changes in temperature and moistening on global (O. A. Drozdov, L. G. Spirina and L. Ye. Borisova, M. Kh. Baydal and A. I. Neushkin, Kh. P. Pogosyan and A. A. Pavlovskaya) and regional scales (L. G. Polozova, L. V. Klimenko, G. N. Vitvitskiy, I. S. Glukh and N. K. Kononova, V. V. Kupriyanov, and others) in dependence on atmospheric circulation or independently.

It was noted by O. A. Drozdov that the correlation between changes in temperature and precipitation can be direct and inverse, in dependence on the physical nature of the formation of precipitation. L. G. Spirina and L. Ye. Borisova found that against a background of general cooling in the northern hemisphere there are periodic fluctuations of temperature with considerable amplitudes, shifting from east to west. According to M. Kh. Baydal and A. I. Neushkin, the tendency to a change in the relationship of the number of months with latitudinal (W) and meridional forms of circulation (E and C) in the northern hemisphere noted since 1930-1940 is accompanied by an increase in the frequency of occurrence of large anomalies and shifts in the structure of climate against a background of insignificant changes in mean annual temperatures.

Kh. P. Pogosyan and A. A. Pavlovskaya formulated a hypothesis according to which the cyclic alternation of westerly and easterly winds in the equatorial stratosphere is governed by the cyclicity of atmospheric processes in the earth's extratropical latitudes.

On the basis of use of spectral analysis A. I. Voskresenskiy, L. S. Petrov and A. N. Lyubarskiy detected cyclic fluctuations of mean monthly temperature of different duration in the Arctic; V. K. Astok and A. Kh. Tarand -- in the Gulf of Finland; I. Ya. Alikina, S. I. Kulikova, M. V. Muraveyskaya and M. S. Akhmetov -- in the Urals; L. P. Sorokina, V. F. Durnev and M. F. Semenchuk -- in Eastern Siberia; V. P. Gagua and L. K. Papinashvili -- in fluctuations of temperature and precipitation in Transcaucasia. In all of the regions in one of the first places there is a cycle which in duration approaches the 11- or 22-year cycles of solar activity.

A. A. Nagaytsev demonstrated that the fluctuations and trends in the winter regime are clarified clearly by the methods of complex climatology. G. Ye. Grishankov and T. A. Novikova, I. A. Bashalkhanov and G. N. Mart'yanova, using different principles, developed a classification of moistening regime years for the Crimea and Steppe Transbaykalia for the purpose of using them for climatic prediction.

Reports in the third direction dealt with the problems involved in changes in carbon dioxide in the atmosphere (S. P. Gorshkov); the influence of thermal industrial effluent (Ye. P. Borisenkov, V. P. Meleshko, V. N. Priyemov and B. Ye. Shneyerov), reservoirs (A. I. D'yakonov and L. G. Strelochnykh) and hydroelectric power stations (Ye. B. Bravaya), shifting of waters of Western Siberian rivers into arid regions of the USSR (L. S. Potapova), reducing the forest coverage of areas (O. B. Soromotina; A. D.

Eyyubov, Kh. Sh. Rakhimov and N. D. Ulkhanov; G. N. Grigor'yev, I. Ye. Trofimova and V. A. Afanas'yev), hydromelioration of swamps (V. F. Shebeko and P. A. Kovrigo) and the development of livestock raising (L. A. Mikhaylova) on changes in different climatic characteristics of the atmosphere and also the preservation of the environment in the Lake Baykal area (N. P. Ladeyshchikov and Ye. N. Ladeyshchikova) and the method for quantitative evaluation of purposeful climatic, soil and landscape changes (A. R. Konstantinov).

As a supplement to the program G. N. Nikol'skiy told about the effect of nuclear shots which have been set off on fluctuations of air temperature in the stratosphere.

In the reports relating to changes in urban climate the speakers proposed for discussion problems relating to change in atmospheric transparency, the thermal effect of a city on the atmosphere, the moistening regime and the relationship of tendencies in the change of climatic characteristics, atmospheric circulation and solar activity.

It was noted in these reports that a city, without changing the general tendency in change of the turbidity factor, exerts a considerable influence on the degree of atmospheric turbidity (G. M. Abakumova, T. V. Yevnevich, N. P. Nikol'skaya; Ts. A. Afanas'yeva, L. K. Bondarenko, T. S. Rustamova; V. N. Gorbacheva). A warming influence of a city, which is most conspicuous during winter, is confirmed (A. A. Gerburt-Geybovich, A. A. Bagdasaryan, V. N. Gorbacheva), increases with the development of a city and is considerably greater in transpolar cities, all other conditions being equal.

The participants in the symposium unanimously noted the timeliness and usefulness of holding of the symposium. In a resolution adopted by the symposium there is a listing of the directions whose development is of fundamental importance for solution of the problem. For the purpose of coordinating the work it was deemed necessary that conferences on individual parts of the problem be organized regularly.

NOTES FROM ABROAD

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 80 pp 127-128

[Article by B. I. Silkin]

[Text] As reported in NATURE, Vol 277, No 5692, p 121, 1979, one of the sections of the international program "Dynamics of Climate" provides for study of chemistry of the atmosphere as a climatological factor. Accordingly, during 1977-1978 the meteorologists of the United States carried out the collection of air samples for its isotopic analysis and determination of the content of carbon dioxide. In order to obtain a planetary pattern the experiment was carried out at points extremely remote from one another: at La Jolla (California, 33°N, 117°W), on Fanning atoll (Central Polynesian Archipelago, 4°N, 159°W) and at the Antarctic polar station Amundsen-Scott (South Pole). In order to ensure comparability of the data, the collection of air samples and their analysis were carried out with use of the same method as in the course of a similar experiment carried out during 1955-1956.

It has been established that the content of carbon dioxide in dry air is closely related to the ratio of the carbon isotopes C^{13} and C^{12} in it. For example, in the earth's northern hemisphere, with a CO_2 concentration of $33.42 \cdot 10^{-3}$ %, this ratio is 7.55%. However, in the southern hemisphere, with a content of carbon dioxide attaining only $33.26 \cdot 10^{-3}$%, this value is 7.59%.

Applying the extrapolation method, scientists determined that during the last 22 years (1956-1978) the content of carbon dioxide in the earth's atmosphere decreased from $34.11 \cdot 10^{-3}$% to $33.42 \cdot 10^{-3}$%. In agreement with this, the ratio of the C^{13} isotope to the C^{12} isotope regularly increased from 6.69% to 7.24%. The difference in this ratio (although small) between the two planetary hemispheres is traced during the extent of the entire studied period.

The participants in the investigations postulate that the reduction in the content of carbon dioxide in the earth's air envelope which they discovered can be attributed to the decrease in coal combustion in most of the inustrially developed countries which has come about during recent decades.

The conclusions drawn in the course of these studies are applicable, in particular, in an independent checking of paleoclimatological investigations based on measurement of the carbon content in tree rings.

As reported in SCIENCE NEWS, Vol 115, No 2, p 22, 1979, the glaciologist J. D. Hays, speaking at an annual conference of the American Association for the Advancement of Science in Houston (Texas) in January 1979, reported on the results of study of the influence which the sea ice of the southern hemisphere exerts on climate at a global scale.

Until now the opinion has prevailed that glaciation in the northern hemisphere sets in earlier than in the southern hemisphere and is a factor in the chain of events resulting in the latter. The data collected by Hays, in his opinion indicate the opposite. The southern hemisphere as a whole, and Antarctica in particular, have a greater sensitivity to climatic changes and are entering into the epoch of the next glaciation several thousands of years earlier than the northern hemisphere.

In a study carried out earlier by Hays and his associates it was demonstrated that insignificant and predictable changes in the geometry of the earth's orbit (distance of the planet from the sun, inclination of the earth's axis, shape of orbit) lead to regularly repeating transitions from the glacial periods to warmer conditions. However, such changes in themselves are very small and the mechanism which in the long run leads to such large-scale shifts has remained unknown.

Now Hays has come to the conclusion that the principal factor in these processes is the quantity of sea ice in the waters washing the shores of Antarctica.

Such a conclusion was drawn on the basis of an analysis of a number of columns of bottom material taken by scientific research ships in the extreme southern regions of the Indian Ocean. It is well known that in cases when sedimentary rocks on the sea floor contain a great quantity of fossil residue of diatomaceous microorganisms, the corresponding regions in the past constituted an open basin, not locked by ice. On the other hand, if the sediments contain an abundant amount of clays, the sea in the past was covered with ice, impeding the growth of diatoms.

The advance and retreat of the ice, associated with change in season of the year, are ascertained by determining the position and time of formation of the line serving as the boundary between sediments containing clay and those rich in diatomaceous sediments. Thus, on the basis of such a delimiting line, traced in layers of a bottom core with an age of 20,000 years, researchers have established that during this time, during the period of the south polar summer, about 20 million square kilometers of ocean surface were covered with ice. This is approximately an order of magnitude greater than in the corresponding season at present.

In winter during this period there were up to 40 million square kilometers under the ice, that is, twice as much as at the present time.

An exceedingly important conclusion is that the transition from conditions with an extensive glacier cover to the present situation was extremely abrupt: it occupied only about 300 years. Similarly, the increase in glaciation at the beginning of the last glacial period also occupied several centuries, although in the northern hemisphere this happened several thousand years later.

In the bottom cores subjected to analysis it is possible to trace the following events. An insignificant change in the earth's orbit led to changes in the distribution of solar radiation on a global scale, and thereby to shifts in seasons. After 3,000-8,000 years the mass of sea ice in subantarctic waters increased sharply and the temperature of the sea surface was reduced.

An advance of glaciation also began in the northern hemisphere after several thousand years. It is extremely probable that the albedo associated with this increased over the entire earth, thereby reducing the quantity of solar energy absorbed by the planet. As a result, the climate became colder.

The lower temperatures of the sea surface could lead to changes in circulation of ocean waters. The contact of the cold waters of southern seas with the warmer waters in the Central Atlantic increased evaporation, which was followed by an increase in the condensation of precipitation and an intensification of snowfalls, which also led to glaciation in the northern hemisphere.

J. D. Hays, on this basis, has formulated a long-range climatological forecast in which he predicts that a new glacial period can be expected in the course of the next several thousand years. He proposed the organization of observations of the dynamics of sea ice using satellites because precisely these processes can serve all humanity as a means for "timely warning" of the approaching glaciation. However, at the present time there are no indications of an increase in the mass of sea ice, although the temperature of the surface of subantarctic waters has decreased slowly during the last 9,000 years.

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