ME

6 OCTOBER 1980 NO. 7, JULY 1980 1 OF 2

JPRS L/9334

6 October 1980

USSR Report

METEOROLOGY AND HYDROLOGY

No. 7, July 1980

FBIS FOREIGN BROADCAST INFORMATION SERVICE

NOTE

JPRS publications contain information primarily from foreign newspapers, periodicals and books, but also from news agency transmissions and broadcasts. Materials from foreign-language sources are translated; those from English-language sources are transcribed or reprinted, with the original phrasing and other characteristics retained.

Headlines, editorial reports, and material enclosed in brackets [] are supplied by JPRS. Processing indicators such as [Text] or [Excerpt] in the first line of each item, or following the last line of a brief, indicate how the original information was processed. Where no processing indicator is given, the information was summarized or extracted.

Unfamiliar names rendered phonetically or transliterated are enclosed in parentheses. Words or names preceded by a question mark and enclosed in parentheses were not clear in the original but have been supplied as appropriate in context. Other unattributed parenthetical notes within the body of an item originate with the source. Times within items are as given by source.

The contents of this publication in no way represent the policies, views or attitudes of the U.S. Government.

For further information on report content call (703) 351-2779 (political, sociological, military); 351-2780 (economic, science and technology).

JPRS L/9334

6 October 1980

USSR REPORT

METEOROLOGY AND HYDROLOGY

No. 7, July 1980

Translation of the Russian-language monthly journal METEOROLOGIYA I GIDROLOGIYA published in Moscow by Gidrometeoizdat.

CONTENTS

UDC 551.557(215-17)

EMPIRICAL MODELS OF WIND VELOCITY DISTRIBUTION IN THE STRATOSPHERE AND MESOSPHERE OF THE NORTHERN HEMISPHERE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 5-10

[Article by Professor S. S. Gaygerov, M. Ya. Kalikhman and V. V. Fedorov, Central Aerological Observatory, submitted for publication 15 January 1980]

[Text] Abstract: The article describes construction of empirical models of the wind (including zonal and meridional components) for altitudes greater than 30 km on the basis of joint use of data from rocket and satellite sounding.

The atmospheric layer in the range 30-80 km remains poorly investigated due to a lack of data. This layer is situated over the upper level of systematic ascents of radiosondes and considerably lower than the layer studied by orbital artificial earth satellites. Accordingly, the climatological description of the stratomesosphere for the time being is in the stage of empirical modeling on the basis of data from infrequent rocket launchings and thermal sounding from satellites.

The term "empirical models of the atmosphere" usually means the typical (spatial and temporal) distributions of its parameters, obtained on the basis of statistical processing and analysis of experimental data.

Great difficulties arise when developing empirical models of zonal and meridional components of wind velocity in connection with the need for taking into account the longitudinal differences which are especially significant during winter. It is found that for the levels of the upper atmosphere the greatest longitudinal differences are characteristic of the meridional wind components [5-7]. Due to the lack of data on the global distribution of wind in the COSPAR International Reference Atmosphere (CIRA-1972) data are given on the zonal components of wind velocity [3]. Even relatively recent studies are limited to a synoptic and statistical analysis of the zonal components of the wind, for the most part along the meridian 80°W [1, 2].

Due to practical requirements the need arises for obtaining empirical models of the wind based on global data, with the models extended to the levels of the upper mesosphere. The possibility of constructing such models is becoming increasingly realistic in connection with the appearance during recent years of high-level charts based on a joint analysis of rocket and satellite data, and also with the development of wind observations in the high layers of the atmosphere by indirect methods (radar tracking of the drift of meteor trails, observations of ionospheric inhomogeneities, etc.).

In this article we give an example of the development of empirical models of the distribution of the zonal and meridional components of wind velocity in the altitude range 30-80 km. These models are based on an analysis of the actual wind and geostrophic wind values, computed from pressure pattern charts constructed on the basis of joint use of satellite and rocket data.

Data sources and methodological problems. The principal result of construction of wind models for the stratosphere and mesosphere of the northern hemisphere was maps of the mean zonal and meridional components of the wind and their standard deviations for the levels 30, 35, 40, 45, 50, 55, 60, 70 and 80 km.

In compiling these maps use was made of data from the international network of rocket sounding stations and USSR scientific research ships for the period from 1962 to 1977. The mentioned data were taken from the bulletins of results of rocket sounding published in the USSR and in the United States.

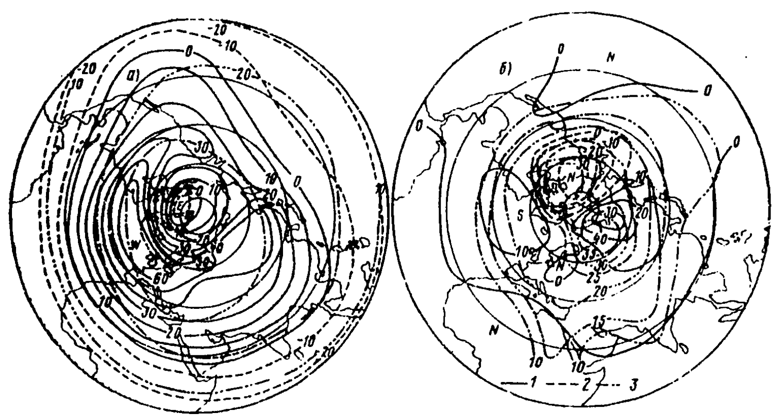

Fig. 1. Maps of mean monthly zonal (a) and meridional (b) wind components and their standard deviations. January, level 40 km. 1) lines of equal values of westerly and southerly components; 2) easterly and northerly components; 3) lines of equal values of standard deviations.

On the maps for the levels indicated above we plotted the monthly values of the zonal and meridional components of wind velocity and their standard deviations, which were corrected by means of construction and analsis of high-level time sections.

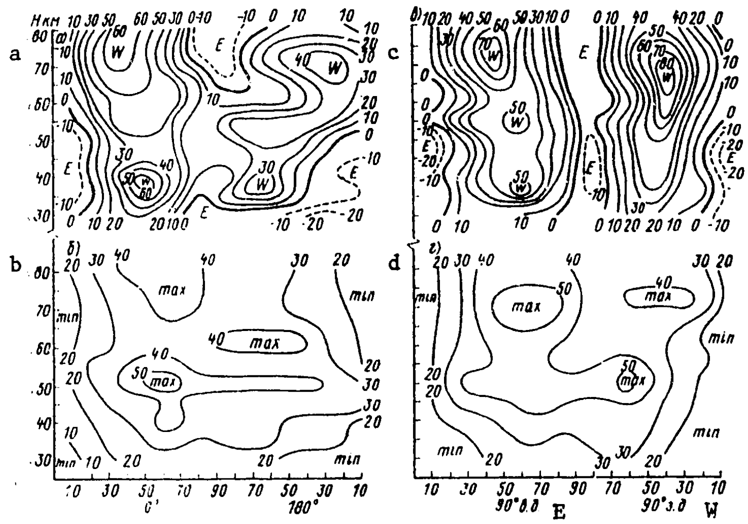

Fig. 2. Time sections of mean values of the zonal wind along the meridians 0-180° (a) and 90°E - 90°W (c) and standard deviations 0-180° (b) and 90° E - 90°W (d). January.

Taking into account that the observational data for some rocket stations did not ensure the necessary spatial resolution for representation of the planetary wind fields, the results of thermal sounding from satellites were also used in constructing the wind models. Weekly pressure pattern charts for 5, 2 and 0.4 mb for each Wednesday in January and July for a five-year period were compiled using rocket and satellite data for computing the values of the geostrophic wind. For the years 1972 and 1973 the charts were taken from [8]. Since 1975 such charts have been compiled at the Central Aerological Observatory on the basis of data from satellite sounding (VTPR radiometer) received from the United States by way of exchange of scientific information. The charts gave rocket data and the geopotential heights on the basis of data from thermal sounding from a satellite. The rocket and satellite data were in very satisfactory agreement.

The values of the geostrophic wind were computed using weekly pressure pattern charts at 48 points in the hemisphere (each 20° of latitude and 30° of longitude). Then we computed the mean monthly values of components of the geostrophic wind, which together with the corresponding values obtained on the basis of rocket observations were plotted on charts for the levels 35, 40 and 55 km (which approximately correspond to the levels 5, 2 and 0.4 mb) and analyzed. Examples of construction of such maps are given in Fig. 1. Maps of wind at the remaining levels were analyzed primarily on the basis of the results of rocket sounding.

The next stage in the development of wind models was the construction (on the basis of the charts) of vertical sections of the mean monthly values of the zonal and meridional wind components and their standard deviations along four meridians (0, 90°E., 180, 90°W) from 10 to 80°N. The choice of the indicated meridians was determined from the following considerations. The section 0-180° gives the latitudinal distribution of the wind over the oceans of the northern hemisphere, whereas the section 90°E - 90°W represents the wind distribution over the continents.

Description of models. As an example we will examine the distribution of the zonal wind along the above-mentioned meridians in January. Figure 2a, c shows that westerly transfer predominates along both meridians. Differences in winter circulation in the upper atmosphere in its western and eastern parts are clearly expressed in the section 0-180° longitude. Over the Atlantic Ocean there are two maxima of velocity of the westerly wind. One of these is situated at an altitude of 40 km near 50°N; the second is present in the upper mesosphere at 40°N. Over the Pacific Ocean there are three weaker maxima: the first is situated in the middle stratosphere at latitude 60°N, the second is at approximately this same latitude in the region of the stratopause and the third, which is situated in the middle mesosphere, is displaced into the tropics (Fig. 2a). It should be noted that the velocity of westerly transfer over the Pacific Ocean in the entire thickness of the stratosphere and mesosphere is approximately half the wind velocity over the Atlantic, which is attributable to the development of the Aleutian High. The easterly winds in the stratosphere and mesosphere over the polar regions are associated with frequent movements of the center of the circumpolar cyclone into the Canadian and European sectors of the Arctic.

Figure 2c shows the latitudinal distribution of the zonal wind along the meridian 90°E - 90°W. It is possible to discriminate two characteristic peculiarities: stronger westerly transfer in the middle mesosphere and the presence of only one maximum of velocity of the westerly wind over North America.

Figure 2b,d, which shows the variability of the zonal wind along the mentioned meridians, makes clear that the highest standard deviations of the zonal wind component are situated in the neighborhood of the stratopause and in the middle mesosphere. This agrees with the great variations of temperature and geopotential of the isobaric surfaces transpiring during the winter and having a quasiperiodic nature.

Results of Comparison of Represented Empirical Model of Wind Distribution in the Northern Hemisphere Upper Atmosphere (EVM-1978) and International Reference Atmosphere COSPAR (CIRA-1972). Zonal Component

H км	1 ЭВМ-1978 — CIRA-1972							
	10	20	30	40	50	60	70	80° с. ш. 6
				2 Январь, 90° з. д.				
60	—16	—6	—12	—15	42	8	50	45
55	—7	4	—10	—21	43	—11	45	65
50	—5	9	—8	—35	27	—12	25	72
45	—2	3	—6	—40	19	—6	0	66
40	—9	0	—1	—38	23	1	—20	52
35	—11	—1	1	—28	41	2	—27	35
30	—12	—7	1	—7	39	—5	—22	16
				3 Июль, 90° з. д.				
60	5	1	12	3	14	7	9	6
55	—9	—9	2	—3	7	—2	1	4
50	—9	—8	—5	—2	2	—2	4	4
45	—4	—4	—8	—3	—3	0	5	4
40	—2	—2	—8	—2	—3	0	6	2
35	0	0	—8	—3	—1	2	5	—1
30	0	—1	—7	—4	2	3	5	1
				4 Январь, 90° в. д.				
60	—22	—4	—5	—34	—8	—8	15	11
55	—12	0	3	—32	—18	—23	10	28
50	4	13	13	—48	—34	—25	—3	33
45	—2	3	11	—57	—45	—22	—5	35
40	—8	—3	12	—59	—40	—20	—2	31
35	—10	—1	12	—52	—31	—24	—1	18
30	—8	0	8	—40	—30	—53	—10	—4
				5 Июль, 90° в. д.				
60	—3	—18	13	11	22	12	12	4
55	—2	—2	8	8	17	3	4	3
50	—3	6	3	5	7	1	4	3
45	—3	1	0	3	—1	2	5	4
40	—5	3	—2	2	—1	1	6	2
35	—1	7	0	4	4	5	6	0
30	6	9	4	5	8	7	8	1

KEY:
1. EVM-1978 - CIRA-1972
2. January, 90°W
3. July, 90°W
4. January, 90°E
5. July, 90°E
6. N

In summer, by virtue of the symmetry of circulation in the stratosphere and lower mesosphere, the latitudinal distribution of the zonal wind in all the sections has an approximately identical character. As indicated above, the principal form of representation of the developed models is maps of the distribution of the zonal and meridional components of the wind in the stratosphere and mesosphere in the northern hemisphere. Figure 1a shows that at an altitude of 40 km over the northern part of the Atlantic Ocean and adjacent regions of the continents there are maximum velocities of westerly transfer, which is considerably reduced over the Pacific Ocean. The axis of the westerly jet stream with maximum winds of more than 60 m/sec is oriented approximately along 50°N. Weak easterly flows predominate over the western part of the Pacific Ocean, bounded by the Arctic region, in connection with development of the Aleutian High.

The interpretation of meridional circulation is most complex because of poor study. In winter the distribution of the meridional components in the polar and subpolar regions is most characteristic. Figure 1b shows that there is a predominance of air transfer from Eurasia to North America in the stratosphere, which is attributable to development of the Aleutian High and frequent displacement of the center of the circumpolar vortex into the Canadian and European sectors of the Arctic. Judging from the maps constructed for the higher levels, in the mesosphere there is a compensating flow in the opposite direction. The latter can be attributed to the fact that the Aleutian High cannot be traced in the upper mesosphere, whereas the center of the cyclone is displaced into the eastern sector of the Arctic.

Comparison with models developed earlier. Up to the present time different authors have repeatedly undertaken to construct global empirical models of wind distribution in the high layers of the atmosphere. The most modern of these is the International Reference Atmosphere COSPAR (CIRA) [3]. It is therefore of interest to carry out a comparison of the mean values of the zonal wind on the basis of the data in the model described here with the values in the CIRA-1972 model. The results of the comparison are presented in the table, in which we give the differences in the mean values of the zonal wind to an altitude 60 km along two meridians for winter and summer. The limitation of altitude is attributable to the fact that in the CIRA-1972 model above 60 km the longitudinal differences are not taken into account.

It is easy to see a considerable discrepancy in the compared data in January. For example, at latitude 80°N over North America they attain 70 m/sec, and in the middle latitudes approximately 40 m/sec. In the eastern hemisphere these differences are also considerable and in the middle latitudes approach 60 m/sec.

Such discrepancies can be attributed to the fact that, first of all, for the development of the represented model we also made use of a considerable number of observations carried out in 1971-1977 using an improved method. Second, in the considered model a more precise allowance has been made for the longitudinal differences in the characteristics of large-scale circulatory processes observed in winter in the northern hemisphere.

As is well known, during summer circulation in the upper atmosphere has an ordered stable character which varies extremely insignificantly from year to year. From this point of view the small differences in the mean values of the zonal wind in July which are discovered in a comparison of the two mentioned models are entirely explicable. At the same time this can serve as additional confirmation of the reality of the model described here.

Summary. Thus, on the basis of observations of the wind made using rocket probes and data on the geostrophic wind, computed with the additional use of the results of thermal sounding from satellites, we obtained a model distribution of the values of the zonal and meridional wind components and their variability over the northern hemisphere in the layer 30-80 km. It appears that the characteristics described above really reflect the wind distribution in the stratosphere and mesosphere in the northern hemisphere.

It should be noted that the proposed models of latitudinal distribution of the wind give some idea concerning the mean long-term conditions of circulation in the stratosphere and mesosphere in the northern hemisphere. The models of zonal and meridional wind components for July completely reflect the characteristics of atmospheric circulation in summer because the year-to-year variability at this time of the year is insignificant. This cannot be said of the winter wind distribution (January) due to the development of stratomesospheric warmings and the circulatory restructurings associated with them. It is well known that mid-winter warmings in the upper atmosphere in the polar regions of the northern hemisphere are observed every year. Nevertheless, the circulatory restructurings associated with them are not observed every winter and according to the data in some studies have a quasi-two-year periodicity [4]. Accordingly, the routine use of long-term mean wind values in the high latitudes is not always justified because in the process of averaging over a series of years there is a considerable smoothing of the data. Thus, the considerable year-to-year variability of wind distribution in the polar regions of the northern hemisphere in winter is not reflected in averaged models. As is well known, in some reference and standard atmospheres winter temperature models for the stratomesopause were developed separately for winters with and without warmings. In the future it is evidently necessary to consider similar approaches in constructing wind models; it is necessary to develop models for winters without circulatory restructurings (with a predominance of westerly flow over the polar regions) and winters with restructurings (with easterly winds over the Arctic).

BIBLIOGRAPHY

1. Gaygerov, S. S., Zaychikov, B. P., Kalikhman, M. Ya., Fedorov, V. V., VOZDUSHNYYE TECHENIYA V MEZOSFERE ANTARKTIKI (Air Currents in the Antarctic Mesosphere), Leningrad, Gidrometeoizdat, 1975.

2. Belmont, A. D., Dartt, D. G., Nastrom, G. D., "Variations of Stratospheric Zonal Winds, 20-65 km 1961-1971," J. APPL. METEOROL., Vol 14, No 4, 1976.

3. COSPAR INTERNATIONAL REFERENCE ATMOSPHERE 1972-CIRA, 1972, Academic Verlag, Berlin, 1972.

4. Gaigerov, S. S., "On Stratospheric Warmings in the Antarctic and Arctic," POLAR METEOROLOGY. WMO TECHN. NOTE, No 87, 1967.

5. Groves, G. V., "Atmospheric Structure and its Variation in the Region from 25 to 120 km," AFCRL ENVIRON. RES. PAPER No 368, 1971.

6. Murgatroyd, R. J., "Winds in the Mesosphere and Lower Thermosphere," PROC. ROY. SOC., Vol 288, No 1415, 1965.

7. Kantor, A. J., Cole, A. E., "Zonal and Meridional Wind to 120 km," JGR, Vol 69, No 24, 1964.

8. Staff, Upper Air Branch NOAA. National Weather Service, National Meteorological Centre, "Synoptic Analyses, 5-, 2- and 0.4-Millibar Surface for January 1972 Through June 1973, NASA SP-3091, Washington, D. C., 1975.

UDC 551.551.2

NONLOCAL PARAMETERIZATION OF TURBULENT FLUXES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 11-19

[Article by Doctor of Physical and Mathematical Sciences V. M. Voloshchuk and Candidate of Physical and Mathematical Sciences P. N. Svirkunov, Institute of Experimental Meteorology, submitted for publication 20 February 1980]

[Text]

Abstract: The authors examine a general scheme for nonlocal parameterization of turbulent fluxes based on a formal solution of the continuity equation. The kernal of the integral expression, relating the turbulent flux to the gradient of the mean concentration of an impurity, is expressed in terms of the statistical characteristics of the turbulent medium. The article indicates the possibility of a nonlocal relationship between the global meridional heat flux and the gradient of the mean temperature profile, following from the well-known Budyko climatic model.

The problem of parameterization of turbulent fluxes is one of the most complex aspects of the theory of turbulent diffusion which has not been completely investigated. In most of the existing theories there is predominance of a semiempirical approach, the basis of which is a local parameterization [5]:

$$\vec{q} = -K \vec{\nabla} \langle c \rangle ,$$

where $\vec{q}$ is the turbulent flux, $\langle c \rangle$ is the mean (determined from turbulence records) concentration of the impurity, K is the so-called coefficient of turbulent exchange (in a general case a second-order tensor).

In order to ascertain K we make use of a variety of physical assumptions, as well as empirical data. However, in a general case it is impossible to make an unambiguous comparison of the exchange coefficient with the statistical characteristics of the turbulent medium. This is indicated, in particular, by the circumstance that in the field of stationary turbulence for some conditions, for example, with boundary conditions not dependent on time, the K coefficient can be represented conveniently by some function of coordinates, whereas for other conditions, such as diffusion from an instantaneous point source, there are data indicating that K is a function of time. A detailed analysis of this problem is given in [9]. The main reason for such a circumstance is that the transport of an impurity in a turbulent flow is determined by the entire spectrum of velocity fluctuations, up to fluctuations of external scales. The external scale of the turbulent flux as a rule is comparable with the characteristic scale of change in the mean concentration profile [4, 5]. As a result, the turbulent flux can be dependent not only on the gradient of the mean concentration, but also on higher-order derivatives. The presence of this sort of "latent parameters" does not make it possible to relate the exchange coefficient unambiguously to the statistical properties of the turbulent medium.

It appears probable that such difficulties can be overcome by using a nonlocal parameterization in which the turbulent flux is related by a linear integral expression with the mean concentration gradient. This hypotheses was expressed for the first time in [2], in which, proceeding on the basis of physical considerations, the author proposed some integral expression relating the turbulent flux in the surface layer to the impurity concentration gradient.

In this article we propose a general scheme of nonlocal parameterization based on a formal solution of the continuity equation.

Such an approach makes it possible, on the one hand, to confirm the hypothesis of a nonlocal parameterization by formal reasonings, and on the other hand, makes it possible to express the parameterization nucleus in terms of the statistical (generally speaking, Lagrangian) characteristics of the turbulent flux. The latter can be useful in refining the different approximate approaches in the theory of turbulent diffusion.

1. As the basis for the examination we will use the continuity equation for a passive scalar substance

$$\frac{\partial c}{\partial t} + \operatorname{div} c\vec{v} = Q(\vec{r}, t), \tag{1}$$

where c is the concentration, $Q(\vec{r}, t)$ are the sources, $\vec{v}(\vec{r}, t)$ is the velocity field.

In order to simplify the reasonings we will consider the medium to be incompressible: $\operatorname{div} \vec{v} = 0$.

Applying the usual averaging procedure, which we will represent by the symbolization $\langle ... \rangle$, from (1) we obtain

$$\frac{\partial \langle c \rangle}{\partial t} + \operatorname{div} (\langle \vec{v} \rangle \langle c \rangle + \langle \vec{v}' c' \rangle) = Q(\vec{r}, t), \tag{2}$$

$$\frac{\partial c'}{\partial t} + \operatorname{div} \vec{v} c' = \operatorname{div} \langle \vec{v}' c' \rangle - \operatorname{div} \vec{v}' \langle c' \rangle . \tag{3}$$

Here $c' = c - \langle c \rangle$, $\vec{v}' = \vec{v} - \langle \vec{v} \rangle$. The fundamental problem is to relate the turbulent flux $\langle \vec{v}' c' \rangle$ to the characteristics of the mean concentration field $\langle c \rangle$.

We will introduce the function of the point source of equation (1), which we will denote by $G(\vec{r}, t \mid \vec{r}', t')$. This function satisfied the equation

$$\frac{\partial G}{\partial t} + \operatorname{div} \vec{v} G = \delta (\vec{r} - \vec{r}') \, \delta (t - t') \tag{4}$$

and the causality condition $G(\vec{r}, t \mid \vec{r}', t') = 0$ with $t' > t$. In explicit form

$$G(\vec{r}, t \mid \vec{r}', t') = \delta(\vec{r} - \vec{R}(t \mid \vec{r}', t'),$$

where $\vec{R}(t \mid \vec{r}', t')$ is a Lagrangian trajectory passing through the point $\vec{r}$ at the time t'.

In order not to complicate the examination with boundary conditions, we will solve the problem in all space. Using the function $G(\vec{r}, t \mid \vec{r}', t)$, the solution of equation (3) will be written in the form

$$c' = \int_{-T}^{t} dt' \int d\vec{r}' \, G(\vec{r}, t \mid \vec{r}', t') (\operatorname{div} \langle \vec{v}' c' \rangle - \operatorname{div} \vec{v}' \langle c \rangle) \Big|_{\vec{r}', t'} + \int d\vec{r}' \, G(\vec{r}, t \mid \vec{r}', -T) \, c'(\vec{r}', -T). \tag{5}$$

Here T is the initial moment in time.

If the fluctuation $c'(\vec{r}, -T)$ has been "prepared" by the turbulent flux, then, writing for $c'(\vec{r}, -T)$ an expression which is similar to (5), and using the group property

$$\int d\vec{r}' \, G(\vec{r}, t \mid \vec{r}', t') \, G(\vec{r}', t' \mid \vec{r}_0, t_0) = G(\vec{r}, t \mid \vec{r}_0, t_0); \quad t > t' > t_0,$$

it is possible to shift the initial condition in the solution (5) to a later moment in time, in particular, to the moment when the concentration is for the first time stipulated arbitrarily, in a form not correlated with the velocity field. This circumstance makes it possible to neglect the influence of the initial value of the fluctuation. Henceforth as a simplification we will assume that $T = \infty$.

Multiplying expression (5) by $\vec{v}'(\vec{r}, t)$ and averaging, for the components of the turbulent flux $\vec{q} = \langle \vec{v}' c' \rangle$ we obtain

$$q_i = \int_{-\infty}^{t} dt' \int d\vec{r}' \, (\langle v_i' (\vec{r}, t) \, G(\vec{r}, t \mid \vec{r}', t') \rangle \operatorname{div} \vec{q} (\vec{r}', t') - \tag{6}$$

$$- \langle v_i' (\vec{r}, t) \, G(\vec{r}, t \mid \vec{r}', t') \, v_k' (\vec{r}', t') \rangle \nabla_k \langle c (r', t) \rangle .$$

Using repeating indices, as usual, we represent summation from 1 to 3. The expression for div $\vec{q}$ is found by averaging expression (5) directly. It follows from this that

$$\operatorname{div} \vec{q} = \int dt_1\, d\vec{r}_1\, dt_2\, d\vec{r}_2 < G(\vec{r}, t | \vec{r}_1 t_1) >^{-1} \times$$
$$< G(\vec{r}_1, t_1 | \vec{r}_2, t_2)\, v'_k(\vec{r}_2, t_2) > \nabla_k < c(\vec{r}_2, t_2) >. \quad (7)$$

Here $< G(\vec{r}, t | \vec{r}_1, t_1) >^{-1}$ is the kernel of the integral operator, the reverse of $< G(\vec{r}, t | \vec{r}_1, t_1) >$.

Substituting (7) into expression (6), we obtain a final expression for the flux:

$$q_i = \int dt_1\, d\vec{r}_1\, dt_2\, d\vec{r}_2\, dt_3\, d\vec{r}_3 < v'_i(\vec{r}, t)\; G(\vec{r}, t | \vec{r}_1, t_1) > \times$$
$$\times < G(\vec{r}_1, t_1 | \vec{r}_2, t_2) >^{-1} < G(\vec{r}_2, t_2 | \vec{r}_3, t_3)\, v'_k(\vec{r}_3, t_3) > \nabla_k < c(r_3, t_3) > - \quad (8)$$
$$- \int dt_1\, d\vec{r}_1 < v'_i(\vec{r}, t)\; G(\vec{r}, t | \vec{r}_1, t_1)\; v'_k(\vec{r}_1, t_1) > \nabla_k < c(\vec{r}_1, t_1) >.$$

Symbolically, in shortened operator form, expression (8) can be written in the form

$$\vec{q} = (<\vec{v}'G><G>^{-1}<G\vec{v}> - <\vec{v}'G\vec{v}'>)\,\vec{\nabla}<c>. \quad (8a)$$

Thus, expressions (8), (8a) show that the turbulent flux is expressed, generally speaking, nonlocally in the form of a space and time integral of the mean concentration gradient.

The accuracy of the approximation of local parameterization will be dependent on how close the kernel of expression (8) is in its properties to the delta function $\delta(\vec{r} - \vec{r}')\,\delta(t - t')$.

In the subsequent examination we will limit ourselves to the case when $\langle\vec{v}\rangle = 0$, that is, we will examine diffusion in a coordinate system moving with a mean velocity. As indicated by (8), (8a), the turbulent flux is determined by two terms which correspond to the two terms on the right-hand side of equation (3). It appears that the second term is related to the temporal nonuniformity of $<c>$. In actuality, integrating it by parts and taking into account that $G(\vec{r}, t | \vec{r}', t')$ in the coordinates $\vec{r}'$, t' satisfies the equation

$$\frac{\partial G}{\partial t'} + v_i(\vec{r}', t')\,\nabla'_i G = 0, \quad (9)$$

we will have

$$-\int dt'\, d\vec{r}' < v'_i(\vec{r}, t)\; G(\vec{r}, t | \vec{r}_1', t')\; v'_k(\vec{r}', t') > \nabla'_k < c(\vec{r}', t') > =$$
$$= \int dt'\, d\vec{r}' < v'_i(\vec{r}, t)\; v'_k(\vec{r}', t')\; \nabla'_k G(\vec{r}, t | \vec{r}', t') > < c(\vec{r}', t') > = \quad (10)$$
$$= -\int d\vec{r}' \int_{-\infty}^{t} dt' < v'_i(\vec{r}, t) \frac{\partial G}{\partial t'} > < c(\vec{r}', t') > =$$

$$= \int d\vec{r}' \int_{-\infty}^{t} dt' \langle v_i' (\vec{r}, t) \; G(\vec{r}, t | r', t') \rangle \frac{\partial \langle c \rangle}{\partial t'}, \tag{10}$$

which had to be demonstrated. In particular, in a stationary case it is equal to zero. It is important to bear this circumstance in mind when developing different approximate approaches when it is necessary to evaluate the role of different terms in the expression for a turbulent flux.

Expression (8) can be considerably simplified by integrating the first term by parts. It appears that it contains a contribution compensating (10) and the final expression for q_i is rewritten in the following equivalent form:

$$\begin{gathered} q_i = \int dt_1 d\vec{r}_1 dt_2 d\vec{r}_2 \langle v_i (\vec{r}, t) \; G(\vec{r}, t | \vec{r}_1, t_1) \rangle \times \\ \times \langle G(\vec{r}_1, t_1 | \vec{r}_2, t_2) \rangle^{-1} \langle c(\vec{r}_2, t_2) \rangle. \end{gathered} \tag{11}$$

We will emphasize that the representations (8) and (11) in essence are formal identities; in order to put them in specific form it is necessary to invoke additional physical considerations. Assume, for example, that the velocity field is a Gaussian random function. Then, using the known Frunts-Novikov formula [3], we will have

$$\begin{gathered} \langle v_i (\vec{r}, t) \; G(\vec{r}, t | \vec{r}_1, t_1) \rangle = - \int_{t_1}^{t} dt' \int d\vec{r}' B_{ik} (\vec{r}, t | \vec{r}', t') \times \\ \times \langle G(\vec{r}', t' | \vec{r}_1, t_1) \; \nabla_k' G(\vec{r}, t | \vec{r}', t') \rangle, \end{gathered} \tag{12}$$

where $B_{ik}(\vec{r}t | \vec{r}'t')$ is the velocity field correlation function.

Then it is possible to act in the spirit of the Kreychan approximation and the mean of the product in expression (12) is approximately replaced by the product of the means. As a result we obtain

$$q_i = \int d\vec{r}' \int_{-\infty}^{t} dt' B_{ik} (\vec{r}, t | \vec{r}', t') \langle \nabla_k' G(\vec{r}, t | \vec{r}', t') \rangle \langle c(\vec{r}', t') \rangle. \tag{13}$$

It is evident that this approach incorporates the shortcomings of the Kreychan approximation [6], although it can be expected that it will be more precise than local parameterization.

2. Now we will examine the case of homogeneous isotropic turbulence in greater detail. It is convenient to make the examination in terms of a time-and-space Fourier transform because as a result of homogeneity the integral expressions (8), (11) represent convolutions in space and time, and it is known [7] that the Fourier transform of convolutions of the functions is equal simply to the product of the Fourier components. The latter is determined by the expressions

$$<G(\vec{r}, t|\vec{r}', t')> = \int \frac{d\vec{k}\,d\omega}{(2\pi)^4} \exp(i\vec{k}(\vec{r}-\vec{r}') - i\omega(t-t'))\, G(\vec{k}, \omega),$$
$$<c(\vec{r}, t)> = \int \frac{d\vec{k}\,d\omega}{(2\pi)^4} \exp(i\vec{k}\vec{r} - i\omega t)\, c(\vec{k}, \omega). \tag{14}$$

It follows from the isotropicity condition that $<\vec{v}G>(\vec{k}, \omega) \sim \vec{k}$. Taking this circumstance into account, after simple transformations from (11) we obtain

$$\vec{q}(\vec{k}, \omega) = -\frac{i\vec{k}}{k^2}(G^{-1}(\vec{k}, \omega) + i\omega)\, c(\vec{k}, \omega). \tag{15}$$

Inverting (15), for the flux we obtain the expression

$$\vec{q} = -\int d\vec{r}' \int_{-\infty}^{t} dt'\, A(\vec{r}-\vec{r}', t-t')\, \vec{\nabla}<c(\vec{r}', t')>, \tag{16}$$

in which the kernel $A(\vec{r}, t)$ was determined by the expression

$$A(\vec{r}, t) = \int \frac{d\vec{k}\,d\omega}{(2\pi)^4} k^{-2} \exp(i\vec{k}\vec{r} - i\omega t)(G^{-1}(\vec{k}, \omega) + i\omega). \tag{17}$$

On the basis of expressions (15)-(17) we will examine some limiting situations. First we will discuss the diffusion approximation and the corrections to it. For the function $<G(\vec{r}, t\,|\,\vec{r}_1 t_1)>$ we use a Gaussian approximation

$$<G(\vec{r}, t|\vec{r}'t')> = (2\pi\sigma^2(t-t'))^{-3/2} \exp\left[-\frac{(\vec{r}-\vec{r}')^2}{2\sigma^2(t-t')}\right], \tag{18}$$

where $\sigma^2(t - t')$ is the dispersion of the displacements of a particle of the impurity during the time $t - t'$.

Accordingly $G(\vec{k}, \omega)$ will be determined by the integral:

$$G(\vec{k}, \omega) = \int_0^\infty dt \exp\left(i\omega t - \frac{1}{2}k^2\sigma^2(t)\right). \tag{19}$$

For $\sigma^2(t)$ we will use the Taylor formula and the asymptotic expansion following from it:

$$\sigma^2(t) = 2\,\overline{v^2}\,t \int_0^t dt' \left(1 - \frac{t'}{t}\right) R(t') = 2\varkappa(t - t_l + O(t_l^2)), \tag{20}$$

where R(t) is the Lagrangian correlation function, normalized to unity

$$\varkappa = \int_0^\infty dt\, R(t)\, \overline{v^2}$$

is the turbulent diffusion coefficient,

$$t_l = \left(\int_0^\infty dt\, tR(t)\right)\left(\int_0^\infty dt\, R(t)\right)^{-1}$$

is a value close in sense to the Lagrangian time scale. For the time $t < t_L$ $\sigma^2(t) = 2\overline{v}^2 t^2 + O(t^3)$.

Using these expansions in an evaluation of the integral (19), for $\vec{q}(\vec{k},\omega)$ we obtain an asymptotic expansion:

$$\vec{q}(\vec{k},\omega) = -i\vec{k}(\varkappa - \varkappa^2 k^2 t_l + i\varkappa\omega t_l + O(t_l^2))\, c(\vec{k}\omega), \tag{21}$$

and for the turbulent flux $\vec{q}(\vec{r}, t)$ we accordingly will have

$$\vec{q}(\vec{r}, t) = -\varkappa\left(\vec{\nabla}<c> - \varkappa t_l \Delta\vec{\nabla}<c> - t_l \frac{\partial}{\partial t}\vec{\nabla}<c> + O(t_l^2)\right). \tag{22}$$

The first term in (22) corresponds to a local parameterization; the others correspond to corrections to it.

If we denote by L and T the characteristic spatial and temporal scales of the mean concentration field, it can be seen from (22) that the expansion is carried out in powers of the parameters $\varkappa t_L L^{-2}$ and $t_L T^{-1}$. For problems with sufficiently slowly changing external conditions the parameter $t_L T^{-1}$ can be small. However, the value $\varkappa t_L L^{-2}$ for typical conditions of turbulent currents in order of magnitude is equal to unity. This circumstance is also the reason for the shortcomings in local parameterization. In order to make them clearer, we will examine in greater detail the situation $\varkappa t_L L^{-2} \gg 1$. For the function $<G(\vec{r}, t\,|\,\vec{r}', t')>$ we will use a self-similar representation which follows from dimensionality and isotropicity considerations [5],

$$<G(\vec{r}, t\,|\,\vec{r}', t')> = z^{-3}(t-t')\, g\left(\frac{|\vec{r}-\vec{r}'|}{z(t-t')}\right). \tag{23}$$

In this case the normalization condition must be satisfied

$$4\pi\int_0^\infty dx\, x^2 g(x) = 1. \tag{24}$$

The function z is related to the dispersion σ by the expression

$$\sigma^2(t) = 4\pi\int_0^\infty dx\, x^4 g(x)\, z^2(t). \tag{25}$$

Under the condition $\varkappa t_L L^{-2} \gg 1$, as is well known [5], $z(t) \sim t$.

We will examine the more general dependence

$$z(t) = \gamma t^\alpha. \tag{26}$$

We will limit ourselves to the stationary case $\partial<c>/\partial t = 0$. In this case the flux will be determined by the component G(k, 0), which can be computed without putting g(x) into specific form:

$$G(\vec{k}, 0) = \left(1 + \frac{1}{\alpha}\right) \frac{4\pi}{(\gamma k)^{1/\alpha}} \sin\frac{\pi}{2}\left(\frac{1}{\alpha} - 1\right) \Gamma\left(\frac{1}{\alpha} - 1\right) \times$$
$$\times \int_0^\infty dx\, x^{2-\frac{1}{\alpha}} g(x). \tag{27}$$

Substituting (27) into (15) and inverting the Fourier transform, we obtain the sought-for relationship

$$\vec{q} = -\int d\vec{r}'\, A(\vec{r} - \vec{r}')\, \vec{\nabla} \langle c(\vec{r}') \rangle, \tag{28}$$

where $A(\vec{r})$ is determined by the expression

$$A(\vec{r}) = \frac{\gamma^{1/\alpha}}{(2\pi)^3} \frac{1-\alpha}{1+\alpha} \frac{\left(\int_0^\infty dx\, x^{2-\frac{1}{\alpha}} g(x)\right)^{-1}}{\sin\frac{\pi}{2}\left(\frac{1}{\alpha} - 1\right)} \cdot \frac{\sin\frac{\pi}{2\alpha}}{r^{1+\frac{1}{\alpha}}}. \tag{29}$$

In particular, in the inertial interval ($z \sim t$) we have an essentially nonlocal relationship

$$\vec{q} \sim -\int d\vec{r}'\, \frac{1}{|\vec{r} - \vec{r}'|^2} \vec{\nabla} \langle c(\vec{r}') \rangle. \tag{30}$$

With $\alpha \to 1/2$ (diffusion approximation) it can be shown that $A(\vec{r}) \sim \delta(\vec{r})$ and we will return to a local parameterization.

3. A nonlocal parameterization can find use in simple models of climate. As is well known, their basis [8] is the energy balance equation integrated in latitude

$$2\pi R^2 \cos\varphi\, (Q - I(T)) = \frac{\partial}{\partial\varphi} H(\varphi). \tag{31}$$

On the left-hand side we have the difference between the heat receipts due to solar energy and losses due to thermal radiation. On the right-hand side we have the derivative of the total meridional heat flux. In formulating the model an important consideration is the parameterization of the flux $H(\varphi)$. It can be assumed that $H(\varphi)$ in a general case is nonlocally related to the mean meridional temperature profile:

$$H(\varphi) = -\int_0^{\frac{\pi}{2}} d\varphi'\, A(\varphi, \varphi') \frac{\partial T}{\partial\varphi'}. \tag{32}$$

The $A(\varphi, \varphi')$ kernel in its sense should characterize the distribution of sizes of the vortices or disturbances of zonal circulation, making the principal contribution to meridional transfer. Unfortunately, at the present time there is not sufficient information for drawing definite conclusions concerning the form of $A(\varphi, \varphi')$. Nevertheless, the representation (32) is convenient because it makes it possible to examine different approximate approaches used in simple models of climate from a common point of view. For example, the North model [10] corresponds to the local parameterization $A(\varphi, \varphi') \sim \cos\varphi\,\delta(\varphi-\varphi')$

$$Q - I = D \frac{1}{\cos\varphi} \frac{\partial}{\partial\varphi} \cos\varphi \frac{\partial}{\partial\varphi} T. \qquad (33)$$

Such an approach corresponds to the assumption that the characteristic dimensions of the vortices responsible for meridional transfer are much less than the earth's radius. We will examine another dependence $A(\varphi, \varphi')$, modeling the nonlocal parameterization and allowing simple examination,

$$A(\varphi, \varphi') = B - F\,|\mu - \mu'|. \qquad (34)$$

Here B and F are some constants, $\mu = \sin\varphi$.

Substituting expression (34) into (32) and (31), we obtain

$$Q - I = \frac{D}{\pi R^2} (T - T_A), \qquad (35)$$

where T_A is the mean arithmetical temperature between the temperature of the pole and equator.

Expression (35) is very similar to the known Budyko model [1]. The difference is that in the Budyko model in place of the mean arithmetical temperature T_A there is a global mean temperature

$$\overline{T} = \int_0^{\frac{\pi}{2}} d\varphi \cos\varphi\, T.$$

However, since the really existing dependence $T(\varphi)$ is quite smooth, this difference is small. It is possible to select more precisely a dependence $A(\varphi, \varphi')$ which would correspond directly to the Budyko model. For this purpose we will examine the relationship between the divergence of the heat flux and the temperature characteristic for the particular model:

$$-\frac{1}{2\pi R^2 \cos\varphi} \frac{\partial}{\partial\varphi} H(\varphi) = \beta \left(T - \int_0^{\frac{\pi}{2}} d\varphi \cos\varphi\, T \right). \qquad (36)$$

Hence, using the variable $\mu = \sin\varphi$, we obtain

$$H(\mu) = 2\pi R^2 \beta \int_0^1 d\mu'\, T(\mu')(\mu - \Theta(\mu - \mu')). \qquad (37)$$

Integrating by parts, it is possible to reduce expression (37) to a more symmetric form

$$H(\mu) = -2\pi R^2 \beta \int_0^1 d\mu' (\mu\mu' - \mu'\Theta(\mu - \mu') - \mu\Theta(\mu' - \mu)) \frac{\partial T}{\partial \mu'}. \qquad (38)$$

which determines the corresponding expression for $A(\varphi, \varphi')$. Since the Budyko model rather successfully describes the fundamental characteristics of the earth's climatic system, this can serve as an indirect indication that a nonlocal parameterization of the heat flux corresponds more closely to meridional transfer.

BIBLIOGRAPHY

1. Budyko, M. I., IZMENENIYA KLIMATA (Climatic Changes), Leningrad, Gidrometeoizdat, 1974.

2. Voloshchuk, V. M., "Nonlocal Parameterization of Vertical Turbulent Exchange in the Surface Layer," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 6, 1976.

3. Klyatskin, V. I., STATISTICHESKOYE OPISANIYE DINAMICHESKIKH SISTEM S FLUKTUIRUYUSHCHIMI PARAMETRAMI (Statistical Description of the Dynamic Systems With Fluctuating Parameters), Moscow, Nauka, 1975.

4. Landau, L. D., Lifshits, Ye. M., MEKHANIKA SPLOSHNYKH SRED (Mechanics of Continuous Media), Moscow, Gostekhizdat, 1953.

5. Monin, A. S., Yaglom, A. M., STATISTICHESKAYA GIDROMEKHANIKA (Statistical Hydromechanics), Part I, Moscow, Nauka, 1965.

6. Monin, A. S., Yaglom, A. M., STATISTICHESKAYA GIDROMEKHANIKA, Part 2, Moscow, Nauka, 1967.

7. Shilov, G. Ye., MATEMATICHESKIY ANALIZ. VTOROY SPETSIAL'NYY KURS (Mathematical Analysis. Second-Special Course), Moscow, Nauka, 1965.

8. FIZICHESKIYE OSNOVY TEORII KLIMATA I YEGO MODELIROVANIYA (Physical Principles of the Theory of Climate and its Modeling), edited by Corresponding Member A. S. Monin, Leningrad, Gidrometeoizdat, 1977.

9. Yaglom, A. M., "Equations With a Time-Dependent Coefficient Describing Diffusion in the Stationary Surface Air Layer," IZV. AN SSSR, FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), No 11, 1975.

10. North, G. R., "Theory of Energy Balance Climate Model," J. ATMOS. SCI., Vol 32, No 11, 1975.

UDC 551.509.323:62-5:519.24

CHOICE OF PARAMETERS FOR FORMULATION OF REGRESSIONAL MODELS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 20-26

[Article by M. S. Kogan and Candidate of Physical and Mathematical Sciences L. N. Romanov, West Siberian Regional Scientific Research Institute, submitted for publication 23 January 1980]

[Text]

Abstract: The article examines the procedures for arrangement of parameters and obtaining evaluations of regressional models with the use of algebraic methods for supplementation and limitation. Constructive algorithms are proposed for the choice of an optimum set of parameters using continuous checking, making it possible to formulate regressional models on the basis of experimental data represented by great masses of information. Experiments in formulating temperature forecasting models are described.

Introduction. The choice of a set of parameters (predictors) is the fundamental problem in formulating statistical forecasting models, the solution of which to a considerable degree determines their effectiveness. The fundamental possibility of obtaining a regression function in all possible combinations of initial parameters still does not make it possible to solve the problem of choosing an optimum set because this also requires effective methods for evaluating the mean risk (mean error using independent material). On the other hand, a complete sorting-out of all the possibilities cannot be considered satisfactory from the point of view of solution of practical problems. These two aspects of the problem give rise to two stages into which it is natural to break down its solution. The first stage involves the arrangement of the parameters from a stipulated set; in the second stage some optimum number of the first parameters is selected from the arranged set of parameters. The solution of the first stage involves purely algorithmic difficulties, whereas the second stage, in addition to algorithmic difficulties, also involves fundamental difficulties associated with evaluation of the models under conditions of restrictions on the sample of situations.

At the present time studies on the formulation of statistical models devote considerable attention to problems relating to the arrangement of parameters [3, 6, 12]. Nevertheless, existing solutions of these problems cannot be considered final due to their limited possibilities in the solution of practical problems. The arrangement models used in formulating statistical models are either inadequately effective in the sense of the "quality" of arrangement or are considerably limited with respect to the possibilities for the processing of data represented by great masses of information.

The problem of choosing spatial dimensionality is related to the creation of effective methods for the evaluation of models. Existing evaluation methods, using theoretical premises, are usually based on assumptions limiting universality. The striving for universality of the method usually leads to a loss of accuracy.

The most universal method for evaluating statistical models under conditions of limitation of the initial sample is the "continuous checking" method, first proposed in [1] for evaluating linear alternative models. This method, involving one-by-one exclusion of situations from the sample and the evaluation in these situations of the quality of operation of a model formulated in the remaining situations is, with its direct use, the most time-consuming with respect to the computations required. The authors of [11] proposed a simplified algorithm for carrying out continuous checking for the evaluation of alternative models formulated on the assumption of normality of the distribution of multidimensional situations. In monograph [2], in evaluating linear alternative models obtained using the generalized portraits model, the author formulated an algorithm making possible a substantial reduction in the time required for carrying out continuous checking. The method for constructing a separating hyperplane, based on an ordered sorting-out of situations [8], also serves for continuous checking; the time required for this procedure does not exceed the time for constructing the hyperplane itself. An unbiased nature of the evaluation of continuous checking was demonstrated in [5].

In [9] continuous checking was used in evaluating regressional models. In order to shorten the time required for calculations on the basis of the teaching material, in each case a group of situations of a prestipulated scale is employed, not a single situation.

The use of continuous checking for the choice of parameters became realistic only due to the use for these purposes of algorithms based on the synthesis of algebraic methods for limiting and supplementation [10]. In this study we examine constructive algorithms for the arrangement of parameters for the purpose of formulating regressional models; we also solve the problem of choosing the optimum set of parameters. This article represents a continuation of investigations [4, 7, 8], related to the use of supplementation and limiting methods in the formulation of statistical models, and also includes a description of experiments in formulating models with the use of real data.

Formulation of problem. Assume that there is a stipulated set of N situations $x_1,\ldots,x_N$, each of which consists of n parameters, and assume that there is a stipulated series of numbers $y_1,\ldots,y_N$, representing the values of the function to be restored at the points $x_1,\ldots,x_N$. Using the stipulated information, from the set of parameters $x^1,\ldots,x^n$ it is necessary to discriminate some subset $x^{k_1},\ldots,x^{k_l}$ and from this subset obtain the linear regression function

$$y = f(x) = \alpha_1 x^{k_1} + \ldots + \alpha_l x^{k_l} + c, \qquad (1)$$

having some optimum properties. It is natural to require that the subset of predictors be selected in such a way that the function (1), obtained on the basis of the condition

$$\sum_{i=1}^{N} (y_i - f(x_i))^2 = \min_{\alpha,\, c}, \qquad (2)$$

will minimize the mean square error

$$\overline{(y - f(x))^2} \qquad (3)$$

on the basis of independent material.

Expressing the regression function in matrix form, we will have

$$y = x'A^{-1}MY,$$

where A is a matrix of covariations of the order $l \times l$, M is the matrix of l-dimensional situations $l \times N$, Y is the vector of the known values of the function.

In order to find the function (1) the initial parameters are arranged by minimizing the empirical risk (mean error on the basis of dependent material). However, we will determine dimensionality proceeding on the basis of the mean risk minimum (3) obtained using continuous checking.

Arrangement of parameters. In order to discriminate a subset from the initial set of parameters $x^1,\ldots,x^N$ we will arrange it:

$$x^{k_1}, \ldots, x^{k_n},$$

so that in a system of "embedded" subsets

$$S^{k_1} \subset S^{k_2} \subset \ldots \subset S^{k_n}, \qquad (4)$$

where $S^{k_i} = \{x^{k_1},\ldots,x^{k_i}\}$, there is assurance of the greatest rate of decrease in the nonclosure of the solution of the corresponding systems of normal equations. Such an order of the system of "embedded" sets (4) evidently corresponds to the maximum rate of increase in the multiple correlation coefficient.

We will arrange the set of initial parameters by their successive inclusion. For this purpose as the set S^{k1} we will select a parameter imparting a maximum value to the special correlation coefficient. In order to obtain the set S^{k2} we include in S^{k1} a parameter which together with that already selected imparts a maximum to the multiple correlation coefficient with the predicted parameter. In order to obtain S^{k3} we include in S^{k2} a parameter which together with the first two imparts a maximum to the multiple correlation coefficient with respect to the variables x^{k1}, x^{k2}, x^{k3}, etc.

We will assume that l parameters have already been ordered. This means that for the covariation matrix

$$A_l = \begin{Vmatrix} a_{11} & \ldots & a_{1l} \\ \cdot & & \\ \cdot & & \\ \cdot & & \\ a_{l1} & \ldots & a_{ll} \end{Vmatrix}$$

of these parameters its inversion A_l^{-1} is known. In order to include the $l+1$-st parameter we will examine the matrix

$$A_{l+1} = \begin{Vmatrix} A_l & b_{l+1} \\ b'_{l+1} & a_{l+1,\,l+1} \end{Vmatrix},$$

in which A_l is the main minor of the matrix A_{l+1} of the order $l \times l$,

$$b_{l+1} = (a_{1,\,l+1}, \ldots, a_{l,\,l+1}), \quad b'_{l+1} = (a_{l+1,\,1}, \ldots, a_{l,\,l+1}).$$

Representing the matrix which is the inverse of A_{l+1} in the form

$$A_{l+1}^{-1} = \begin{Vmatrix} B_l & \beta_{l+1} \\ \beta'_{l+1} & \alpha_{l+1,\,l+1} \end{Vmatrix},$$

it is possible to obtain the expressions

$$\begin{aligned} \beta_{l+1} &= -\alpha_{l+1,\,l+1}^{-1} A_l b_{l+1}, \\ \alpha_{l+1,\,l+1} &= \alpha_{l+1,\,l+1} - b'_{l+1} A_l^{-1} b_{l+1}, \\ B_l &= A_l^{-1} + \alpha_{l+1,\,l+1}^{-1} (A_l^{-1} b_{l+1} b'_{l+1} A_l^{-1}), \end{aligned} \tag{5}$$

by means of which it is easy to evaluate the matrix A_{l+1}^{-1}. The sorting-out of the $n - l$ remaining parameters, which is necessary for selecting the $l+1$-st parameter, does not require the operation of matrix inversion, but requires only an $(n - l)$-th correction of its elements.

Thus, using expressions (5) for the matrices

$$\| a_{11} \|, \quad \begin{Vmatrix} a_{11} & a_{12} \\ a_{21} & a_{22} \end{Vmatrix}, \quad \begin{Vmatrix} a_{11} & a_{12} & a_{13} \\ a_{21} & a_{22} & a_{23} \\ a_{31} & a_{32} & a_{33} \end{Vmatrix}, \ldots,$$

from which each next is obtained from the preceding by means of limitation, it is possible to obtain a sequence of inverse matrices and thus in each interval avoid unwieldy matrix inversion operations. If the covariation matrix formed on the basis of all parameters is nondegenerate, the process of arranging the parameters will be ended after computation of the matrix A_n^{-1}.

In arrangement by means of exclusion (reverse procedure) in each interval we determine the parameter whose exclusion imparts a minimum increase to the nonclosure of solution of the system of normal equations or the minimum decrease to the corresponding multiple correlation coefficient. Such an arrangement involves a great volume of computations and for its realization requires multiple solution of systems of linear equations. However, using the limitation method, the number of operations when arrangement is by the exclusion procedure can be reduced by an order of magnitude. For this we use the expression

$$A_l^{-1} = B_l - \alpha_{l+1,\,l+1}(\beta_{l+1}\,\beta'_{l+1}), \tag{6}$$

which can be derived easily from (5). Thus, if the matrix A_{l+1}^{-1} is known, using formula (6) it is possible to compute a matrix of the l-th order $l + 1$ times using this same matrix and thereby determine the parameter whose elimination imparts the minimum decrease to the multiple correlation coefficient. By repeating the considered procedure $n - l$ times, we obtain a new series of parameters the order in which corresponds to the order of their exclusion.

It is evident that the order of the parameters obtained by means of inclusion with whatever considerable dimensionality cannot coincide with the order obtained by means of exclusion. However, the use of direct and reverse procedures in arrangement frequently makes it possible to obtain a more precise solution.

Choice of set of parameters. After the parameters are arranged, the problem of choosing the optimum set is reduced to determination of the number of first parameters with which the evaluation of the mean risk attains a minimum value. The mean risk will be computed using the formula

$$\delta_l = \frac{1}{N} \sum_{i=1}^{N} \left(\frac{y_i - f(x_i)}{1 - x_i' A^{-1} x_i} \right)_l^2, \tag{7}$$

which follows (see [7]) from the matrix identity (see [13])

$$B^{-1} = A^{-1} + \frac{B^{-1}\,uv'\,B^{-1}}{1 + v'\,B^{-1}\,u},$$

correct for any nondegenerate matrices A and $B = A - uv'$. Here u and v are arbitrary vectors of the dimensionality coinciding with the dimensionality of the A matrix. By computing δ_l for all $1 \leqslant l \leqslant h$, it is possible to determine the dimensionality k for which

$$\min_l \delta_l = \delta_k.$$

Introducing the notations in formula (7) for the denominator of the expression and the regression function with the corresponding dimensionality l

$$\gamma_l(x_l) = (1 - x_l' A^{-1} x_l)_l,$$
$$f_l(x_l) = (x_l' A^{-1} MY)_l, \tag{8}$$

by means of expressions (5) it is easy to obtain the expressions

$$\gamma_l(x_l) = \gamma_{l+1}(x_l) - \alpha_{l+1,\, l+1}(x_l,\ \beta_{l+1})^2,$$
$$f_l(x_l) = f_{l+1}(x_l) - \alpha_{l+1,\, l+1}(x_l,\ \beta_{l+1})(\beta_{l+1}\ MY)_{l+1}. \tag{9}$$

Using the recurrent expressions (9), the volume of computations in determining the optimum dimensionality can be reduced considerably by use of the results of computations (8) obtained in the preceding interval.

An evaluation of the mean risk by use of (9) can be made after arrangement by both inclusion and exclusion.

Experiments. Using the method described above we carried out experiments for predicting temperature for different short times in advance. The purpose of the experiments was primarily a clarification of the general laws of behavior of the mean error in dependence on the dimensionality of the ordered set of parameters, and also an investigation of the possibilities of the algorithms in the sense of the effective processing of data represented by great masses of information.

In order to predict temperature at Novosibirsk 33 hours in advance we selected data recorded at radiosonde and surface observation stations located in a radius of 1,000 km. The sample consisted of 868 June-July situations taken in chronclogical order 1957-1970. Each situation was represented by 60 parameters. The initial information in the form of a matrix (868 x 60) was used in formulating models and continuous checking: the matrix columns are arranged by inclusion of parameters and procedures for evaluation of the mean error and the choice of dimensionality are used.

Then the parameters were rearranged by means of exclusion using expression (6) and the dimensionality with which the mean error attains a minimum was again determined.

Figure 1 shows two curves representing the dependence of the mean error on the dimensionality of the ordered space, the difference in which is caused by forming of the regression using different sets of parameters. Curve 1 corresponds to an arrangement by means of inclusion of parameters; curve 2 corresponds to an arrangement by means of their exclusion. The figure shows that the minimum for curve 1 is attained with a number of parameters equal to 27, whereas arrangement by means of exclusion leads to a minimum with a number of parameters equal to 30. The closeness of the local minima with the two types of arrangement indicates a stability of the evaluations of dimensionality by means of continuous checking and at the same time makes it possible to hope for a closeness of the resulting minima on a plane to the global minimum in multidimensional space. The smooth behavior of the curves is evidently the result not only of the stability of evaluations of the mean error, but also the result of effective arrangement of the parameters.

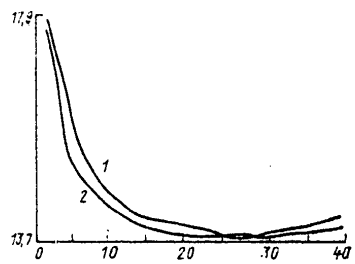

Fig. 1. Curves of the dependence of the mean error computed using continuous checking on the number of arranged parameters. Curve 1 corresponds to arrangement by inclusion, curve 2 -- exclusion.

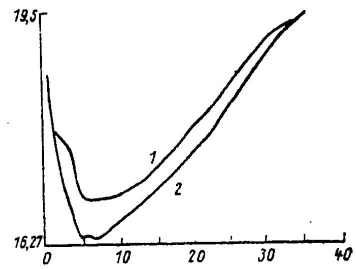

Fig. 2. Curves of the dependence of the upper evaluation of mean risk on the number of arranged parameters. Curve 1 corresponds to arrangement by inclusion, curve 2 -- exclusion.

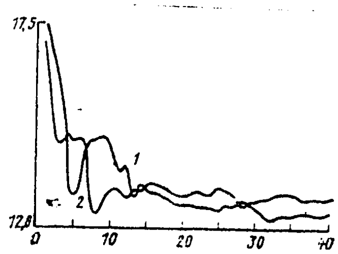

Fig. 3. Curves of dependence of mean error computed using 427 independent situations on the number of arranged parameters. Curve 1 corresponds to arrangement by inclusion, curve 2 -- exclusion.

Figure 2 shows similar curves constructed using the same initial information, but using the ordered minimization method [4]. The figure shows that the behavior of the curves differs in having a lesser smoothness and the corresponding minima differ more in their value.

Figure 3 shows curves of the dependence of the mean error on the dimensionality of the space, constructed on the basis of 427 situations not used in constructing the models (independent material). The figure shows that the curves are characterized by a lesser smoothness than the curves in Fig. 1.

However, the global minimum shown in Fig. 3 is quite close to the minimum obtained using continuous checking (see Fig. 1). It is interesting to note that the point of the minimum of the upper evaluation of the mean risk (see Fig. 2) is rather close to the point of the local minimum with a dimensionality 8 obtained on the basis of independent material.

The probable success of a temperature forecast for 33 hours in advance by means of a model using 27 parameters is 90.3%. The forecast is considered unsuccessful if its error exceeds 5°C.

Processing of other masses of meteorological information by the described method confirms the nature of behavior of the curves shown in Fig. 1. This is expressed in the fact that they all have a U shape, and if some fluctuations sometimes appear, impairing the uniqueness of the minimum, as a rule they are usually insignificant. It therefore follows that the time required for calculating the optimum dimensionality can be substantially reduced. In actuality, after the mean error begins to increase stably the computation process can cease without danger of the appearance of a deeper minimum with higher dimensionalities.

All the computations were made using a BESM-6 electronic computer; the time for arrangement by inclusion and the determination of dimensionality was about five minutes. Arrangement by the exclusion of parameters required approximately the same time.

In this article the choice of the set of parameters was accomplished for the purpose of optimizing linear models. However, the importance of creating constructive arrangement algorithms and choice of the parameters for forming a linear regression is dictated not only by the need for restoration of a linear dependence. Using the mathematical approach of restoration of the linear dependence, it is possible to formulate a more complex piecewise-linear model. Such a model as a part includes linear models formulated on the basis of corresponding subsamples. The restoration of a polynomial dependence also involves the use of the approach of restoration of a linear regression. Effective models for rapid arrangement and obtaining evaluations of models make it possible not only to determine the degree of the approximating polynomial, but also to discriminate power terms, and also mixed products of power terms exerting a positive influence on the quality of the approximation.

BIBLIOGRAPHY

1. Bongard, M. M., "Modeling of the Recognition Process on a Digital Computer," BIOFIZIKA (Biophysics), No 2, 1961.

2. Vapnik, V. N., Chervonenkis, A. Ya., TEORIYA RASPOZNAVANIYA OBRAZOV (Theory of Image Recognition), Moscow, Nauka, 1974.

3. Gruza, G. V., Kaznacheyeva, V. D., "Statistical Structure of the Pressure Field in the Northern Hemisphere," TRUDY SANIGMI (Transactions of the Central Asian Scientific Research Hydrometeorological Institute), No 38(53), 1968.

4. Kogan, M. S., Romanov, L. N., "On the Arrangement of Parameters in the Formulation of Statistical Models," TRUDY ZAP.-SIB. RNIGMI (Transactions of the Western Siberian Regional Scientific Research Hydrometeorological Institute), No 41, 1978.

5. Lunts, A. L., Brailovskiy, V. L., "Evaluation of Criteria Obtained in Statistical Recognition Procedures," IZV. AN SSSR, TEKHNICHESKAYA KIBERNETIKA (News of the USSR Academy of Sciences, Technical Cybernetics), No 3, 1967.

6. Marchenko, A. S., "Attainable Accuracy in Statistical Forecasting and Optimum Dimensionality of a Predictor," IZV. AN SSSR, FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), No 9, 1969.

7. Polyakov, G. G., Romanov, L. N., "Choice of Parameters in Regressional Models Using Continuous Checking," TRUDY ZAP.-SIB. RNIGMI, No 46, 1979.

8. Polyakov, G. G., Romanov, L. N., "Separation of Classes of Situations Using Ordered Sorting," TRUDY ZAP.-SIB. RNIGMI, No 41, 1978.

9. Romanov, L. N., Kogan, M. S., "Nonlinear Regressional Models for Predicting Weather Elements," TRUDY II VSESOYUZNOGO SIMPOZIUMA PO PRIMENENIYU STATISTICHESKIKH METODOV V METEOROLOGII (Transactions of the Second All-Union Symposium on Use of Statistical Methods in Meteorology), Leningrad, Gidrometeoizdat, 1977.

10. Fadeyev, D. K., Fadeyeva, V. N., VYCHISLITEL'NYYE METODY LINEYNOY ALGEBRY (Computational Methods of Linear Algebra), Moscow, Fizmatgiz, 1963.

11. Fukunaga, K., Kessell, D. L., "Estimation of Classification Error," IEEE TRANS. COMP., pp 1521-1527, c-20, December 1971.

12. Glahn, R. H., "Numerical Statistical Forecasting in the U.S. National Weather Service," NOTES FROM A COLLOQUIUM: WEATHER FORECASTING AND WEATHER FORECASTS, Vol 2, 1976, National Center for Atmospheric Research, Boulder, Colorado.

13. Waugh, F. V., Dwyer, P. S., "Compact Computation of the Inverse of a Matrix," ANN. MATH. STATISTICS, Vol 16, No 3, 1945.

UDC 551.509.313

OPTIMIZATION OF A METHOD FOR SOLVING THE BALANCE EQUATION IN A SPHERICAL COORDINATE SYSTEM

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 27-34

[Article by Candidate of Physical and Mathematical Sciences G. S. Rivin and Z. K. Urazalina, Computation Center Siberian Department USSR Academy of Sciences, submitted for publication 15 October 1979]

[Text]

Abstract: A second parameter α is introduced for the single-parameter (parameter η) Miyakoda-Shuman method for solving the nonlinear balance equation in the case of a spherical coordinate system. The region of optimum values of these parameters is found by means of numerical experiments. It is shown that the choice of optimum α and η values will make possible a substantial decrease in the number of iterations.

Introduction. Recently in the construction and use of numerical models of the atmosphere based on the full equations of hydrothermodynamics ever-increasing attention is being given to stipulation of the initial fields, matched with one another in such a way that the results of computations with a numerical model are not distorted by "meteorological noise" [10].

Already the first investigations [13, 14, 21] revealed the possibility of such matching. Thus, in a linear case the use of the geopotential and wind fields, related to one another by geostrophic expressions, as the initial data, has the result that the amplitudes of the undesirable noise become less than the amplitudes of movement with a low frequency [14]. It is possible to obtain a further decrease of undesirable noise if the initial fields mentioned above satisfy the balance equation [13]. The use of such fields as the initial fields improves the quality of the forecast [6], although it does not give a good suppression of the undesirable noise [21].

The balance equation relates two functions with one another -- the stream function and the pressure function (if altitude is used as the vertical coordinate) or geopotential (if pressure is used as the vertical coordinate).

If geopotential (pressure) is considered to be known in the balance equation, whereas the stream function is considered unknown, from the mathematical point of view it is a nonlinear differential equation of the Monge-Ampere type. With stipulation of the real geopotential (pressure) values this equation is an equation of a mixed type, that is, there are regions in which it is elliptical, and regions in which it is not elliptical. Attention was immediately given to this circumstance in the very first studies [11], but experiments indicated that the region of nonellipticity is considerably smaller than the region of ellipticity. A rather complete investigation of the position, size and frequency of appearance of regions of nonellipticity for different levels was presented in [19] on the basis of data for one year, beginning with June 1969, for the extratropical zone of the northern hemisphere.

A small change in the values of the geopotential (pressure) field makes it possible to arrive at a geopotential (pressure) field for which the balance equation becomes elliptical. The geopotential (pressure) field was reduced to an "elliptical" form prior to solution of the balance equation, beginning from the first studies [12, 22], and as noted by many authors and is confirmed by our experiment, this requires small changes in the field at an insignificant number of points of intersection.

A number of studies [1, 4, 8, 9, 11, 12, 17, 22] have been devoted to the formulation of iteration methods for solving the balance equation and corresponding difference schemes. The iteration method, proposed independently by Miyakoda [17] and Shuman [22], has come into the widest use; it is based on a preliminary representation of the balance equation by means of the Petterssen transform [20] in the form of a quadratic equation relative to $\Delta\psi$ -- the Laplacian of the stream function. Then expressing $\Delta\psi$ as a solution of a quadratic equation through the stream function and the known geopotential (pressure)

$$\Delta\psi = F(\psi),$$

where F is the corresponding nonlinear function, whose form will be indicated below, we arrive at the following Miyakoda-Shuman iteration method

$$\Delta\psi^{(j+1)} = F(\psi^{(j)}), \tag{1}$$

where j is the number of the iteration and $\psi^{(0)}$ is stipulated.

A rather careful investigation of the choice of the difference scheme and the iteration process was made by Miyakoda [4], who proposed that the iteration process be changed, to wit:

$$\begin{gathered} \Delta\psi^{(j+1/2)} = F(\psi^{(j)}), \\ \psi^{(j+1)} = 0{,}5\ \psi^{(j+1/2)} + 0{,}5\ \psi^{(j)}. \end{gathered} \tag{2}$$

This process can be modified if the above-mentioned quadratic equation is obtained by simple addition of the function $\eta(\Delta\psi)^2$ to both sides of the balance equation, where η is some parameter; then we arrive at the following iteration process:

$$\Delta \psi^{(l+1/2)} = F(\psi^{(l)}, \eta), \qquad (3)$$
$$\psi^{(l+1)} = 0{,}5\ \psi^{(l+1/2)} + 0{,}5\ \psi^{(l)}.$$

We note that with η = 0.5 the iteration process (3) coincides with (2). Then selecting the optimum value of the η parameter [7], it is possible to achieve a further acceleration of convergence of the Miyakoda-Shuman process.

All the studies mentioned above were made in a Cartesian coordinate system for a rectangular region or for a Cartesian coordinate system in the plane of a stereographic projection and were used in a number of numerical models.

A changeover to global models requires solutions of the balance equation in a spherical coordinate system. The corresponding ellipticity condition was derived in [15] and a generalization of the Miyakoda-Shuman method (2) for a spherical case was used in [16]. A further increase in the rate of the iteration process (2) was obtained by Paegle and Tomlinson [18] in the following way. They described an iteration process (2) in the form

$$\Delta \psi^{(l+1/2)} = F(\psi^{(l)}), \qquad (4)$$
$$\psi^{(l+1)} = \alpha\psi^{(l+1/2)} + (1-\alpha)\ \psi^{(l)}.$$

and it was demonstrated that α = 0.75 is optimum. We note that in this article we have actually introduced the parameter γ, but it is related to α, as is easily confirmed, in the following way:

$$\alpha = 1/(1+\gamma).$$

In addition, in contrast to all the preceding studies, in each iteration they used a direct inversion of the Laplace operator. The choice of the optimum α value was obtained by them in the zone $9° \leqslant \theta \leqslant 69°$, where θ is the complement to latitude, for the 500 mb surface and a uniform latitude-longitude grid with the interval $\Delta\theta = \Delta\lambda = 3°$. In addition, prior to reduction to an elliptical form we carried out triple smoothing of the geopotential field by the Shuman method [23] with ν = 0.5, the Coriolis parameter and β were regarded as constant.

The objective of our article is a study of the behavior of the rate of convergence of the iteration process

$$\Delta\psi^{(l+1/2)} = F(\psi^{(l)}, \eta),$$
$$\psi^{(l+1)} = \alpha\psi^{(l+1/2)} + (1-\alpha)\ \psi^{(l)}$$

with different values of the parameters α and η, in the choice of their optimum values at different levels in the zone $10° < \theta < 70°$ with $\Delta\theta = 5°$, $\Delta\lambda = 10°$, variable values of the Coriolis parameter and β with use of a direct method for inversion of the difference analogue of the Laplace operator.

Formulation of problem and solution method. In a spherical coordinate system (λ, θ, z) for finding the stream function ψ from the known field of pressure p in the D region with the boundary ∂D the balance equation has the form

$$2J(u, v) - f\Delta\psi - \frac{\beta}{a}\frac{\partial\psi}{\partial\theta} + \frac{1}{a^2}\left[1 - \operatorname{ctg}\theta\frac{\partial}{\partial\theta}\right](u^2 + v^2) = \\ = -\frac{1}{\rho}\Delta p. \tag{5}$$

As the boundary condition we use the following:

$$\psi(\lambda, \theta) = G(\lambda, \theta), \ (\lambda, \theta) \in \partial D. \tag{6}$$

The following notations are used in these expressions:

$$J(u, v) = \frac{1}{a^2\sin\theta}\left[\frac{\partial v}{\partial\lambda}\frac{\partial u}{\partial\theta} - \frac{\partial v}{\partial\theta}\frac{\partial u}{\partial\lambda}\right],$$

$$u = \frac{1}{a}\frac{\partial\psi}{\partial\theta}, \quad v = -\frac{1}{a\sin\theta}\frac{\partial\psi}{\partial\lambda},$$

$f = 2\Omega\cos\theta$ is the Coriolis parameter, Ω is the angular velocity of the earth's rotation,

$$\Delta = \frac{1}{a^2\sin^2\theta}\left[\frac{\partial^2}{\partial\lambda^2} + \sin\theta\frac{\partial}{\partial\theta}\left(\sin\theta\frac{\partial}{\partial\theta}\right)\right]$$

is the Laplace operator.

$$\beta = \frac{1}{a}\frac{\partial f}{\partial\theta},$$

a is the earth's radius, ρ is air density, θ is the complement to latitude, λ is longitude, G(λ, θ) is a function stipulated at the boundary ∂D of the D region.

The balance equation (5) is a nonlinear equation of the Monge-Ampere type and has no more than two solutions if the ellipticity condition is satisfied

$$\frac{f^2}{2} + \frac{1}{\rho}\Delta p - \frac{\beta}{a}\frac{\partial\psi}{\partial\theta} + \frac{1}{a^2}(u^2 + v^2) > 0. \tag{7}$$

As in [6], to both sides of equation (5) we add the value $\eta(\Delta\psi)^2$, where η is some nonzero parameter, then

$$\eta(\Delta\psi)^2 + f\Delta\psi - F = 0. \tag{8}$$

where

$$F = \frac{1}{\rho}\Delta p + 2J(u, v) - \frac{\beta}{a}\frac{\partial\psi}{\partial\theta} + \\ + \frac{1}{a^2}\left[1 - \operatorname{ctg}\theta\frac{\partial}{\partial\theta}\right](u^2 + v^2) + \eta(\Delta\psi)^2.$$

Solving (8) as a quadratic equation relative to $\Delta\psi$, we obtain

$$(\Delta\psi)_{1,2} = \frac{1}{2\eta}\left(-f \pm \sqrt{f^2 + 4\eta F}\right). \tag{9}$$

For finding the stream function in the northern hemisphere we use the following iteration process:

$$\begin{gathered}
\Delta\psi^{(j+1/2)} = \frac{1}{2\eta}\left(-f + \sqrt{f^2 + 4\eta F^{(j)}}\right),\\
\psi^{(j+1)} = \alpha\psi^{(j+1/2)} + (1-\alpha)\,\psi^{(j)},\\
\psi^{(0)} = p(\lambda, \Theta)/(\rho f),\ (\lambda, \Theta) \in D,\\
G(\lambda, \Theta) = p(\lambda, \Theta)/(\rho f),\ \ (\lambda, \Theta) \in \partial D.
\end{gathered} \tag{10}$$

An inversion of the difference analogue of the Laplace operator was carried out by a direct method [3]. This method uses a fast Fourier transform for expansion of the sought-for solution and the known right-hand side along the parallel into series of

$$\left\{\cos\frac{\pi li}{N},\ \sin\frac{\pi li}{N}\right\}.$$

The equations then derived for the coefficients were solved by the elimination method along the meridian. The determined coefficients were used in determining the value of the sought-for function at the points of grid intersection once again using the fast Fourier transform [3, 5].

Numerical experiments. The regions of the optimum α and η values were determined by numerical experiments due to the nonlinearity of the iteration process.

In evaluating the rate of convergence of the iteration process (10) it is important to choose the norm for which this evaluation is made because we are interested not so much in the convergence $\psi^{(j)}$ as in the convergence of the derivatives of this function. As in [7], we traced the behavior of the value

$$\|U\| - \sum_N \sqrt{u^2 + v^2}/N,$$

where $U = (u, v)'$ is the wind velocity vector, N is the number of internal points in the D region.

The convergence of the iteration process (10) was evaluated by a comparison of $U^{(j)}$ with the precise solution $U^{(\infty)}$. For each level as $\|U^{(\infty)}\|$ we selected the $\|U^{(j_0)}\|$ value, which did not change even one of the first ten significant decimal places in the subsequent iteration.

Henceforth we will say that $U^{(j)}$ has n true places, if for all $i \geqslant j$ we have the inequality

$$\big|\, \| U^{(i)} \| - \| U^{(l)} \| \,\big| < 0{,}5_{10}^{m-n},$$

where m is the decimal order $\| U^{\infty} \|$.

Table 1

Number of Iterations Required for Obtaining n True Places of Solution With M-Parameters, (I), With PT-Parameters (II) and With RU-Parameters (III)

	3 апреля ~~3 April~~			22 августа ~~22 August~~		
	$z=2900$	$z=5460$	$z=9120$	$z=2900$	$z=5460$	$z=9120$
			$n=8$			
I	28	28	41	22	24	40
II	17	18	25	12	18	24
III	8	9	9	9	10	10
			$n=7$			
I	22	25	33	19	22	33
II	13	12	21	9	13	20
III	8	8	8	8	8	8
			$n=6$			
I	12	18	23	16	18	26
II	10	10	15	8	10	20
III	6	6	6	5	7	7

In [7] we carried out an adequate number of numerical experiments for clarifying the optimum η values with $\alpha = 0.5$ with data in a rectangular region covering all seasons of the year. In this connection numerical experiments were carried out using data analyzed by Professor I. V. But' for 3 April and 22 August 1965 at six levels. As the levels we used z = 0, 1400, 2900, 5460, 9120, 11 870 m. These values correspond to the altitudes of the standard atmosphere with p = 1000, 850, 700, 500, 300, 200 mb. The finding of the p values with the above-mentioned z values on the basis of data for the isobaric surfaces was accomplished using the equation of statics, as was described in [2].

Hereafter as a convenience we will use the following notations: we will call the pair $\alpha = 0.5$ and $\eta = 0.5$, used by Miyakoda, the M-parameters, the pair $\alpha = 0.75$ and $\eta = 0.5$ from [18] -- the PT-parameters, and the values $\alpha = 1.0$ and $\eta = 0.15$ -- the RU-parameters.

The table gives the number of iterations necessary for obtaining 8, 7 and 6 true places of the solution with M-parameters in case I, with PT-parameters in case II, with RU-parameters in case III. This table shows that the choice $\alpha = 0.75$ gives a gain in the number of iterations in comparison with $\alpha = 0.5$, as in the study by Paegle and Tomlinson. At the same time, with this value $\alpha = 0.75$ a choice $\eta = 0.15$ instead of $\eta = 0.5$ leads to a further acceleration of the rate of convergence of the iteration process (10), as can be seen clearly in Fig. 1. This figure shows curves of the number of iterations necessary for obtaining n true places of the solution with z = 0, and z = 11 870 m respectively. We note that with n = 8 as a rule the nonclosure of the initial equation becomes a value of about 10^{-16} sec^{-2}.

Now we will examine these results in greater detail. In order to obtain 8 true places of the solution at the level 1000 mb for 22 August with use of the M-parameters it is necessary to have 38 iterations; with the PT-parameters it is necessary to have 23 iterations; using the RU-parameters it is necessary to have only 9 iterations. With these same values of the parameters for the data for 3 April at this same level it is necessary to have 30, 18 and 8 iterations respectively.

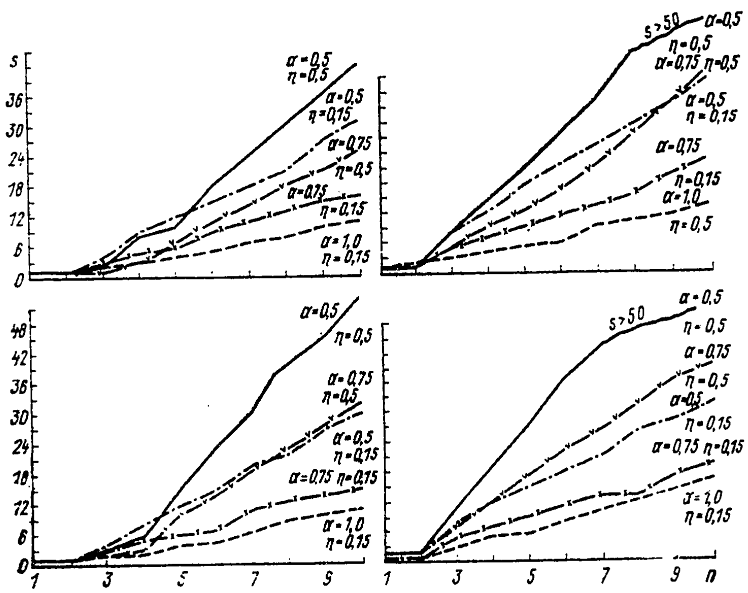

Fig. 1. Dependence of number of iterations s required for obtaining n true places of solution on values of α and η parameters at the level z = 0 (at left) and z = 11 830 m (at right) for altitudes of isobaric surfaces 1000 mb (at left) and 200 mb (at right) on 3 April (at top) and on 22 August 1965 (at bottom).

Similar curves were also constructed for other levels. The behavior of the curves is very similar, which makes it possible to speak of an optimum value of the RU-parameters both in comparison with the M-parameters and in comparison with the PT-parameters.

In order to clarify the optimum α and η intervals, Fig. 2 shows curves of the number of iterations necessary for obtaining 8 true places of the solution with z = 5460 m for different α and η values (as a convenience a nonuniform scale has been used along the x-axis).

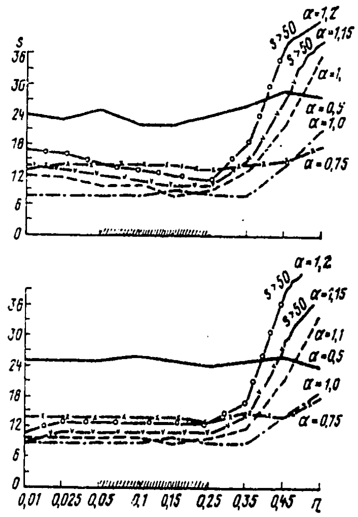

Fig. 2. Curves of the number of iterations s required for obtaining 8 true places of solution at level z = 5460 m for altitudes of isobaric surface 500 mb for 3 April (at top) and 22 August 1965 (at bottom).

With each fixed α value the preferable interval for is the segment [0.05; 0.25]. The advantage of choice of the η values from this interval in comparison with $\eta = 0.5$ is obvious. We note that with a fixed α value for different η the number of required iterations from the mentioned interval virtually does not change. Such a rather broad region of optimum η values, common for all α , makes it possible to hope that the choice of one of the η values from the indicated region will not give a marked change in the behavior of the rate of convergence of the iteration process even with other factual data.

On the other hand, with stipulated η from the indicated interval it can be seen that with $\alpha = 0.5$ the iteration process (10) converges considerably more slowly than with all remaining α values. For example, for $\eta = 0.15$ it is necessary to have 14 iterations with $\alpha = 0.75$; 10 iterations with $\alpha = 1.0$; 9 iterations with $\alpha = 1.1$; 11 iterations with $\alpha = 1.15$, but 25 iterations with $\alpha = 0.5$ (for data for 22 August). For the data for 3 April the behavior of the relationship in the number of iterations with these same values of the α parameter is very similar. It should be noted that in both cases with $\alpha = 0.75$ it is necessary to have more iterations than with

$\alpha = 1.0$ and $\alpha = 1.1$. Thus, for α the interval (0.75; 1.15) can be considered acceptable.

Summary. Similar curves were also constructed for the other levels. The behavior of these curves also leads us to the conclusion that the region of optimum η values is the segment [0.05; 0.25], but with η values from this region α from the interval (0.75; 1.15) are optimum. The results of numerical experiments show that the use of optimum α and η values, without changing the total number of arithmetical and logical operations in one iteration, decreases the number of required iterations on the average by a factor of 3 in comparison with the M-parameters and by a factor of 2 in comparison with the PT-parameters.

Although the region of optimum values for each level was found only for two dates, our experience [7] with computations for a Cartesian coordinate system with different values of the η parameter with a fixed α value shows that the region of optimum η values is virtually independent of the initial information. Thus, the above-mentioned optimum values of the α and η parameters can be recommended in practical computations for any factual data.

BIBLIOGRAPHY

1. Dzhabar-Zade, R. M., "One Algorithm for Solution of the Balance Equation," IZVESTIYA AN SSSR, FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), Vol 5, No 3, 1969.

2. Kalenkovich, Ye. Ye., "Numerical Scheme for Predicting the Fields of Meteorological Elements for the Northern Hemisphere," Dissertation for Award of the Academic Degree of Candidate of Physical and Mathematical Sciences, Manuscript, Novosibirsk, 1970.

3. Marchuk, G. I., METODY VYCHISLITEL'NOY MATEMATIKI (Methods of Computational Mathematics), Moscow, Nauka, 1977.

4. Miyakoda, K., "Introduction to Numerical Weather Forecasting," CHISLENNYYE METODY RESHENIYA ZADACH DINAMIKI ATMOSFERY I OKEANA (Numerical Methods for Solving Problems in the Dynamics of the Atmosphere and Ocean), Leningrad, Gidrometeoizdat, 1968.

5. Obraztsov, N. N., "Solution of the Helmholtz Equation on a Sphere," ALGORITMY I PROGRAMMY (Algorithms and Programs), No 2, 1975.

6. Rivin, G. S., Urazalina, Z. K., "Determination of the Initial Wind Field for a Weather Forecasting Scheme," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 12, 1977.

7. Rivin, G. S., Urazalina, Z. K., "Optimization of a Method for Solving the Balance Equation," METEOROLOGIYA I GIDROLOGIYA, No 9, 1978.

8. Arnason, G. A., "A Convergent Method for Solving the Balance Equation," J. METEOROL., Vol 15, No 2, 1958.

9. Asselin, R., "The Operational Solution of the Balance Equation," TELLUS, Vol 19, No 1, 1967.

10. Bengtsson, L., "Four-Dimensional Assimilation of Meteorological Observations WMO/ICSU," GARP PUBL. SER., No 15, 1975.

11. Bolin, B., "Numerical Forecasting With the Barotropic Model," TELLUS, Vol 7, No 1, 1955.

12. Bolin, B., "An Improved Barotropic Model and Some Aspects of Using the Balance Equation for Three-Dimensional Flow," TELLUS, Vol 8, No 1, 1956.

13. Charney, J., "The Use of the Primitive Equations of Motion in Numerical Prediction," TELLUS, Vol 7, No 1, 1955.

14. Hinkelmann, K., "Der Mechanismus des Meteorologischen Larmes," TELLUS, Vol 3, No 4, 1951.

15. Houghton, D. D., "Derivation of the Elliptic Condition for the Balance Equation in Spherical Coordinates," J. ATMOS. SCI., Vol 25, No 1, 1968.

16. Houghton, D. D., Washington, W., "On Global Initialization of the Primitive Equations: Part I," J. APPL. METEOROL., Vol 8, No 5, 1969.

17. Miyakoda, K., "On a Method of Solving the Balance Equation," J. METEOROL. SOC. JAPAN, Vol 34, No 6, 1956.

18. Paegle, J., Tomlinson, E. M., "Solution of the Balance Equation by Fourier Transform and Gauss Elimination," MON. WEATHER REV., Vol 103, No 6, 1975.

19. Paegle, J., Paegle, J. N., "On Geopotential Data and Ellipticity of the Balance Equation," MON. WEATHER REV., Vol 104, No 3, 1976.

20. Petterssen, S., "On the Relation Between Vorticity, Deformation and Divergence and the Configuration of the Pressure Field," TELLUS, Vol 5, 1953.

21. Phillips, N. A., "On the Problem of Initial Data for the Primitive Equations," TELLUS, Vol 12, No 2, 1960.

22. Shuman, F. G., "Numerical Methods in Weather Prediction: I. The Balance Equation," MON. WEATHER REV., Vol 85, No 10, 1957.

23. Shuman, F. G., "Numerical Methods in Weather Prediction: II. Smoothing and Filtering," MON. WEATHER REV., Vol 85, No 11, 1957.

UDC 551.513

SPATIAL STRUCTURE OF CIRCUMPOLAR VORTICES OF THE ATMOSPHERE AND CIRCULATION IN THE EQUATORIAL ZONE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 35-41

[Article by Candidate of Geographical Sciences Ts. A. Kanter, Saratov State University, submitted for publication 11 November 1979]

[Text]

Abstract: In the 60-km layer of the atmosphere there was found to be a three-level spatial structure of circumpolar motion with strongly expressed broadenings in the region of the tropopause and stratopause. It is shown that in the layer 20-40 km, where the circumpolar vortex of the winter hemisphere has a minimum diameter, the subtropical zone of high pressure adjacent to it migrates along the meridian and causes a change in westerly and easterly circulation in the equatorial stratosphere. It is postulated that a quasi-two-year cycle is the characteristic period of interaction of processes in the northern and southern hemispheres. A difference in the structure of circumpolar vortices in both hemispheres was established.

A knowledge of the patterns of general circulation of the atmosphere is a highly important link in creating a theory of climate and improvement in long-range weather forecasts. Accordingly, at the present time work has considerably broadened on the investigation and modeling of large-scale atmospheric processes.

However, the fact should be noted that the largest features of planetary circulation caused by the temperature difference between the equator and the poles, circumpolar vortices (CPV) of the northern and southern hemispheres, for the time being have been poorly investigated. However, the circumpolar motion is a grandiose unified three-dimensional formation. It takes in the extratropical latitudes, at least the troposphere, stratosphere and mesosphere; it is easily detected on pressure pattern charts for all levels above 1000 mb. Now we will turn to these materials.

APPROVED FOR RELEASE: 2007/02/08: CIA-RDP82-00850R000300040014-9

Mean long-term pressure pattern charts have now been published for each month for the northern [2] and southern [3, 4] hemispheres. They were prepared for nine isobaric surfaces: 1000, 850, 700, 500, 300, 200, 100, 50 and 30 mb, that is, to an altitude of 24 km. In addition to long-term maps, daily charts are published for the northern hemisphere which characterize the pressure field to an altitude of 30 km (10 mb). Above 30 km mean weekly pressure charts for six levels (35, 40, 45, 50, 55, 60 km) are published for the northern hemisphere on the basis of data from rocket sounding of the atmosphere [1].

We used the materials enumerated above, characterizing the 60-km layer of the atmosphere in the northern hemisphere and a 24-km layer in the southern hemisphere.

Figure 1 shows long-term mean monthly pressure pattern charts for the northern hemisphere (January) for the isobaric surfaces 850 and 300 mb and Fig. 2 shows pressure charts for the levels 40 and 60 km (January 1977). An analysis of this material, first of all, does not leave doubt that in the winter (at least to an altitude of 60 km) there is a well-expressed cyclonic circumpolar vortex; it is the largest feature of semiglobal circulation. Second, in a comparison of the diameters of the CPV vertically there is found to be a clear pattern -- first the vortex expands considerably (Fig. 1), then it is narrowed (Fig. 2a) and thereafter again expands (Fig. 2b). This pattern is confirmed by all cartographic materials, both long-term mean monthly and daily, to wit: topography of the isobaric surfaces 700, 500, 200, 100, 50, 30, 10 mb; pressure charts at the levels 35, 45, 50, 55 km. The CPV is very well expressed in the southern hemisphere, where its contours are more concentric with the circles of latitude than in the northern hemisphere and the pattern described above is manifested still more clearly.

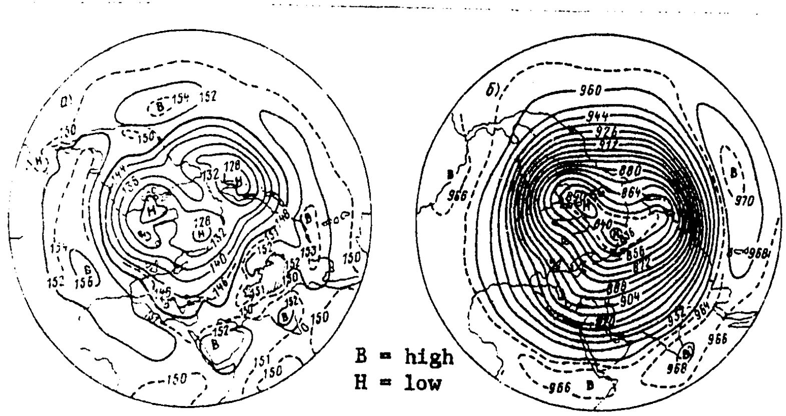

Fig. 1. Mean long-term maps AT_{850} and AT_{300} (a, b). January, northern hemisphere.

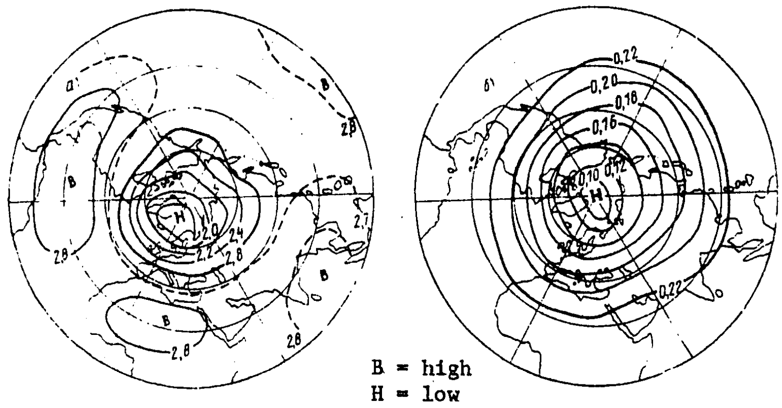

Fig. 2. Pressure maps at the levels 40 and 60 km (a, b). January 1977, northern hemisphere.

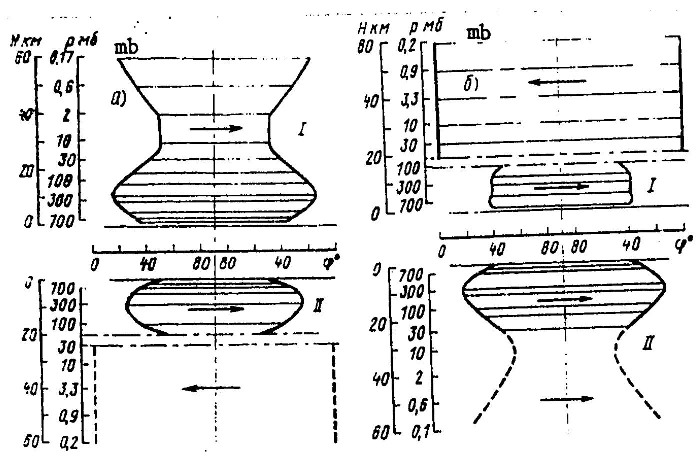

Fig. 3. Models of planetary circumpolar movement of atmosphere. a) January; b) July; I) northern hemisphere, II) southern hemisphere.

In order to compare the magnitude of the circumpolar vortices at different altitudes in different hemispheres and seasons it is necessary somehow to make a quantitative evaluation of the space occupied by them. The area bounded by the last closed isohypse of the CPV can serve as such an evaluation. In the proposed investigation the area was determined using a special overlay proposed in [5] and later improved.

Figures 1-2 show that the isohypses of the circumpolar vortices do not differ too much from the circles of latitude; therefore, for the purposes of clarity the position of the last closed isohypse should be expressed in degrees of latitude of the parallel which limits the area identical to it. The described method was used in determining the boundaries of the CPV for January and July at all isobaric surfaces in both hemispheres and on rocket sounding maps for the northern hemisphere. This made it possible to construct models of planetary circumpolar motion, represented in Fig. 3, where for January and July we successively plotted (vertically) the diameters of the circumpolar vortices at the isobaric surfaces 1000, 850, 700, 500, 300, 200, 100, 50, 30 mb, at the levels 40, 50, 60 km, and envelopes were drawn through their ends.

Table 1

Boundaries of CPV (in Degrees of Latitude) in Troposphere and in Lower Stratosphere in the Winter

1 Изобарические поверхности, мб								
1000	850	700	500	300	200	100	50	30
2 Северное полушарие, январь								
нет ЦПВ 4	37	26	24	15	15	23	29	29
3 Южное полушарие, июль								
42	38	27	19	15	16	24	30	36

KEY:
1. Isobaric surfaces, mb
2. Northern hemisphere, January
3. Southern hemisphere, July
4. no CPV

Since the vertical structure of the CPV was specific, we will examine its most important peculiarities in greater detail.

Table 1 gives the limits of the vortex at all isobaric surfaces according to the mean monthly long-term pressure pattern charts in the layer from 1000 to 30 mb.

An analysis of this table and Fig. 3 clearly shows that in winter in each of the hemispheres in the troposphere and lower stratosphere the circumpolar movement has a highly expanded part which is situated in the layer 300-200 mb. The area of the vortex in the region of the expansion on the average is twice as great as outside it (850 and 30 mb). It must be noted that the expansion of the CPV coincides with such atmospheric features as the tropopause and jet streams. These three phenomena have a high probability of an interrelationship and there is possibly an intercausality.

A comparison of the data in Table 1 shows that the winter circumpolar movement in the layer from 1000 to 30 mb in both hemispheres has virtually identical limits at all levels. The difference between them is that the southern hemisphere vortex has a greater depth than that in the northern hemisphere (Table 2).

Table 2

Geopotential Height (dam) in the Central Part of the CPV in the Northern and Southern Hemispheres at Standard Isobaric Surfaces According to Mean Long-Term Data

1 Полушарие	2 Изобарические поверхности, мб								
	1000	850	700	500	300	200	100	50	30
Северное 3	нет ЦПВ 5	128	270	504	836	1088	1516	1832	2240
Южное 4	—0,12	112	252	484	824	1056	1452	1832	2112

KEY:
1. Hemisphere
2. Isobaric surface, mb
3. Northern
4. Southern
5. No CPV

Table 3

Limits (in Degrees of Latitude) of Circumpolar Vortex in Troposphere and Lower Stratosphere According to Data for 15 January for Five Years in the Northern Hemisphere

Изобарическая поверхность, мб 1	1971	1974	1975	1976	1977	Среднее 2
850	38	46	41	36	37	40
700	29	31	34	27	29	30
500	24	25	27	25	25	25
300	22	22	20	20	20	21
200	23	23	21	23	19	21
100	24	26	25	26	21	24
50	40	47	35	39	43	40
30	54	53	41	45	47	48
10	59	47	52	47	42	49

KEY:
1. Isobaric surface, mb
2. Mean

Now we will examine to what degree the limits of circumpolar movement in the troposphere and lower stratosphere are stable. In the first approximation this can be done by examining the vertical profile of the CPV on individual days. As an example we will take the 15 of January for different years (1971, 1974, 1975, 1976, 1977). The boundaries of the vortex are given in Table 3.

Here we can note two facts. The first is a stable regularity involving a broadening of the CPV from the lower levels to the isobaric surfaces 300 and 200 mb and subsequent narrowing of the vortex with altitude. This can be seen from the data in each column. The second is the great stability of the horizontal dimensions of the CPV in the layer from 500 to 200 mb, whereas above and below it the limits of the vortex vary substantially and in individual cases the area of the expanded part is three times greater than outside it.

Table 4

Limits of CPV in Atmosphere (in Degrees of Latitude) in Layer 40-60 km in January in Two Years in the Northern Hemisphere

Уровень, км (1)	1976				1977			
	Числа (2)							
	3—9	10—16	17—23	24—30	31—7	8—14	15—21	22—28
	1	2	3	4	5	6	7	8
40	54	49	46	40	50	44	57	50
50	53	40	45	38	50	48	40	33
60	50	32	20	29	53	54	35	22

KEY:
1. Level, km
2. Dates

Thus, the circumpolar vortex has the most stable limits in the region of the tropopausal expansion. The variation of its boundaries in the lower troposphere is evidently associated with the distorting influence of macroturbulent vortices, cyclones and anticyclones, whereas the instability of the boundaries in the stratosphere is attributable to the fact that its temperature regime in winter is unstable (the depth of the CPV and its diameter are sharply reduced in the layer where stratospheric warming occurs).

Next we will turn to the vertical profile of the winter circumpolar vortex in the layer 40-60 km on the basis of data from rocket sounding of the atmosphere [1]. Since these data for the time being are not being reduced to long-term values but are being published, as mentioned above, in the form of mean weekly pressure charts, we will cite the limits of the CPV in this layer for four weeks in each of the Januaries of 1976 and 1977 (Table 4).

Here in columns 2, 3, 4, 7 and 8 it is easy to trace the broadening of the vertex from altitude to altitude. In column 1 this expansion is expressed weakly, whereas in columns 5 and 6 it can be seen that the limits of the vortex are even narrowed with altitude.

An attentive study of all the materials for the mentioned time intervals and also similar materials for other winter months indicated that usually from 40 km and above the CPV expands and it is narrowed only in the case and in the layer where winter stratospheric warming occurs. This narrowing is the greater the stronger the warming.

After summarizing the peculiarities of the vertical profile of the circumpolar vortex in winter it can be said that in the 60-km layer of the atmosphere it is characterized by a three-level structure with two expanded parts in the regions of the tropopause and stratopause (we will call them tropopausal and stratopausal expansions), and there is a third narrow part between them (we will call it the "saddle" of the CPV).

Now we will examine the relationship between the vertical profile of the winter CPV and the peculiarities of circulation in the tropical and equatorial zones. Figures 1-2 clearly show that the circumpolar vortex borders on the subtropical region of high pressure. Naturally, circulation in the tropical and equatorial zones is determined by the position of the axial line of this region.

The vertical profile of the subtropical zone of high pressure, like the circumpolar vortex, has a three-level structure. At the levels of the tropopause and stratopause, where the CPV is considerably expanded, the subtropical anticyclonic zone is narrow and is "pressed" toward the equator. Its axial line reaches latitude 10° (Fig. 1b). In the region of the saddle of the CPV the high pressure zone is "floating." For example, in the northern hemisphere in January 1976 its axial line at the isobaric surface 30 mb ran approximately along the parallel 28° and in January 1977 -- 7-9° to the south.

Migration of subtropical anticyclones along the meridian has a very important series of consequences. It leads to a change in the direction and intensity of transfer in the tropics and in the equatorial zone. In the equatorial stratosphere the change in circulation has a quasi-two-year cyclicity. This phenomenon, discovered about 20 years ago, is now being extensively studied [6-9]. It is exceedingly noteworthy that the layer in which an alternation of westerly and easterly circulations is observed completely coincides with the region of narrowing of the circumpolar vortex, that is, with the region in which the anticyclonic zone can first come close to the equator and cause an easterly flow in its zone, then withdraw from it, giving place to a small-cellular circulation with westerly, less stable winds. We recall that this peculiarity of the pressure field is characteristic of the winter hemisphere. However, on the planet one of the hemispheres is always a winter hemisphere. It can be assumed that in the course of one year they alternately ensure easterly transfer in the equatorial stratosphere, and in the course of the next -- westerly, manifesting itself as an indicator of a unified circulation mechanism. In this case the quasi-two-year cycle is the characteristic period of interaction of processes in the two hemispheres. This assumption requires further checking. It becomes possible with an increase in the quantity and quality of aerological and rocket observations in the equatorial and tropical zones.

Now we will proceed to an examination of summer circumpolar movement of the atmosphere in both hemispheres (Table 5).

Table 5

Limits of CPV (in Degrees of Latitude) in Troposphere and Lower Stratosphere in Summer

1 Изобарические поверхности								
1000	850	700	500	300	200	100	50	30
2 Северное полушарие, июль								
нет 3 ЦПВ	38	37	39	40	41	53	антициклон 4	антициклон 4
5 Южное полушарие, январь								
47	41	36	26	24	24	34	55	антициклон 4

KEY:
1. Isobaric surfaces
2. Northern hemisphere, July
3. No CPV
4. Anticyclone
5. Southern hemisphere, January

As indicated in the cited tables, and also in Fig. 3, in summer, as in winter, the circumpolar vortex has a three-level structure. The lower part is the cyclonic vortex, the upper part is the anticyclonic vortex, at each of the considered levels taking in all the hemispheres, whereas the third is a thin layer between them in which the circumpolar movement is disorganized (in the northern hemisphere this is the layer between 100 and 50 mb, and in the southern hemisphere -- 50-30 mb).

The difference in the vertical profiles of summer CPV in both hemispheres is substantial and involves the following. First, its lower part -- the cyclonic vortex -- differs with respect to vertical extent by approximately 4-5 km (in the northern hemisphere it is shorter). Second, in the southern hemisphere there is a clearly expressed tropopausal expansion of the CPV (Fig. 3a), whereas in the northern hemisphere it does not exist (Fig. 3b). Third, in summer the cyclonic circumpolar vortices of the southern hemisphere have a greater depth than in the northern hemisphere.

Thus, the difference in the underlying surfaces in both hemispheres finds total expression in the difference in spatial structure of the largest features of planetary circulation.

Summary

In a study of circumpolar vortices in the 60-km layer of the atmosphere it was discovered that they have not only a strongly expressed three-level structure, but in accordance with this, the subtropical high-pressure field has a similar structure, that is, in essence, the entire planetary pressure field. A direct result of this is a multilevel structure of the wind over the equator.

It was found that in the expanded parts of the CPV there are such atmospheric phenomena as the tropopause, stratopause and jet streams, whereas in the layer of narrowing of the vortex subtropical anticyclones migrate along the meridian, causing a change in the westerly and easterly circulations over the equator.

A specific structure of the pressure field probably could also be detected in a study of layers of the earth's atmosphere above 60 km, as well as in the atmospheres of other planets. This would make it possible to clarify specifically how physicochemical and astronomical factors exert an influence on planetary circulation.

The fact that the difference in the underlying surfaces of the northern and southern hemispheres is manifested totally in the difference of the vertical structure of their circumpolar vortices can become a key to construction of models of the "underlying surface - three-dimensional planetary circulation" system. Such models are necessary both for reconstructing the pattern of general circulation of the atmosphere in climates of the past and for predicting the nature of future circulation as a result of natural or inadvertent change of the underlying surface.

BIBLIOGRAPHY

1. "Atlas of High-Altitude Charts for the Layer 35-60 km," PRILOZHENIYE BYULLETENYU "REZUL'TATY RAKETNOGO ZONDIROVANIYA ATMOSFERY" (Appendix to the Bulletin "Results of Rocket Sounding of the Atmosphere"), No 9, 11, TsAO, GUGMS, 1978.

2. ATLAS KLIMATICHESKIKH KHARAKTERISTIK TEMPERATURY, PLOTNOSTI I DAVLENIYA VOZDUKHA, VETRA I GEOPOTENTSIALA V TROPOSFERE I NIZHNEY STRATOSFERE SEVERNOGO POLUSHARIYA, VYP IV (Atlas of Climatic Characteristics of Temperature, Density and Air Pressure, Wind and Geopotential in the Troposphere and Lower Stratosphere of the Northern Hemisphere) No IV, edited by D. I. Stekhnovskiy and B. S. Chuchkalov, Moscow, Gidrometeoizdat, 1974.

3. Zastavenko, L. G., ATLAS KLIMATICHESKIKH KART BARICHESKOY TOPOGRAFII NIZHNEY STRATOSFERY YUZHNOGO POLUSHARIYA (Atlas of Climatic Pressure Pattern Charts of the Lower Stratosphere in the Southern Hemisphere), Moscow, Gidrometeoizdat, 1975.

4. Zastavenko, L. G., Zakharova, I. B., Olintseva-Nebrat, G. G., ATLAS KLIMATICHESKIKH KART BARICHESKOY TOPOGRAFII YUZHNOGO POLUSHARIYA (1000-10 mb)(Atlas of Climatic Pressure Charts of the Southern Hemisphere (1000-10 mb)), Moscow, 1972.

5. Kanter, Ts. A., ISSLEDOVANIYE NEKOTORYKH SVOYSTV BARICHESKOGO POLYA SVOBODNOY ATMOSFERY METODOM PLOSHCHADEY I TSIRKUMPOLYARNOGO MODELIROVANIYA (Investigation of Some Properties of the Pressure Field in the Free Atmosphere by the Areas Method and by Circumpolar Modeling), Saratov, Izd-vo Saratovskogo Un-ta, 1975.

6. Kats, A. L., TSIRKULYATSIYA V STRATOSFERE I MEZOSFERE (Circulation in the Stratosphere and Mesosphere), Leningrad, Gidrometeoizdat, 1968.

7. Nemirovskaya, L. G., "Wind Structure in the Stratosphere of the Equatorial Zone," TRUDY GIDROMETTSENTRA SSSR (Transactions of the USSR Hydrometeorological Center), No 128, 1974.

8. Pavlovskaya, A. A., "Structure of Global Geopotential Fields in Relation to Quasi-Two-Year Cyclicity in the Equatorial Stratosphere," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 7, 1973.

9. Pogosyan, Kh. P., "Some Characteristics of Wind Cyclicity in the Equatorial Stratosphere," METEOROLOGIYA I GIDROLOGIYA, No 9, 1973.

10. SINOPTICHESKIY BYULLETEN'. SEVERNOYE POLUSHARIYE (Synoptic Bulletin. Northern Hemisphere), Moscow, Gidromettsentr SSSR, 1971-1975.

UDC 551.510.42

MODELING OF TRANSBOUNDARY TRANSPORT OF SULFUR DIOXIDE WITH ALLOWANCE FOR VERTICAL MOVEMENTS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 42-49

[Article by Candidate of Physical and Mathematical Sciences N. S. Vel'tishcheva, USSR Hydrometeorological Scientific Research Center, submitted for publication 25 January 1980]

[Text]

Abstract: This paper presents the results of work on improvement of the model for evaluating the long-range transport of sulfur dioxide by means of improvement in the approximation of boundary conditions at the earth's surface, introduction of a nonuniform vertical interval, and also allowance for vertical movements. The author gives the results of computation of the flux of sulfur dioxide through elements of the boundary and its precipitation onto the underlying surface.

A study of the transport of contaminating substances for distances of 1,500-2,000 km is important for determining the contribution of different countries to contamination of the air basin. Experimental methods for determining the fluxes of impurity through boundaries are costly. Accordingly, during recent years there has been development of a combined approach matching the use of different models and observation systems [8].

The development of models for evaluating the transport of impurity pursues two objectives. First, using them, compute the quantity of matter transported across the boundaries of different countries. The second objective is a determination of the quantity of substance precipitated onto the underlying surface. The collection of this information is necessary for evaluating the degree of the effect of contaminating substances on the environment.

A three-dimensional model for computing the concentration of sulfur dioxide during its propagation over a distance of 1,500-2,000 km has been formulated at the USSR Hydrometeorological Center (1976) [2]. The choice of a difference scheme and an investigation of its stability for solution of the three-dimensional equation for the transport of mass of impurity were described in [3]. In this investigation emphasis is on improvement of the model [2] by an improvement in the approximation of boundary conditions at the earth's surface, introduction of a nonuniform vertical interval, and also development of a variant of a model with vertical movements taken into account. In addition, we give an algorithm for computing the flux of contaminating substances across the elements of a boundary and their precipitation onto the underlying surface.

The propagation of an impurity in the atmosphere will be described by the three-dimensional equation

$$\frac{\partial q}{\partial t}+u\frac{\partial q}{\partial x}+v\frac{\partial q}{\partial y}+w\frac{\partial q}{\partial z}=K_s\left(\frac{\partial^2 q}{\partial x^2}+\frac{\partial^2 q}{\partial y^2}\right)+ \tag{1}$$

$$+\frac{\partial}{\partial z}\left(K_z\frac{\partial q}{\partial z}\right)-R-S+F,$$

where q is the volumetric concentration of SO_2; u, v, w are the components of wind velocity; K_s, K_z are the coefficients of horizontal and vertical diffusion; R and S are the SO_2 losses as a result of chemical reactions and washing out by precipitation, F is the source.

In order to solve equation (1) we selected the following boundary and initial conditions:

$$\begin{aligned} &K_z\frac{\partial q}{\partial z}-\beta q=0 && \text{at } z=0,\\ &q=0 && \text{at } z=H, \end{aligned} \tag{2}$$

where β is a parameter determining the interaction between the impurity and the underlying surface; H is the height of the mixing layer, which was assumed to be constant for the entire region and equal to 2 km for the computed cases. A zero concentration was stipulated at the lateral boundaries.

$$q(x, y, z, 0)=0 \tag{3}$$

for model computations and

$$q(x, y, z, 0)=q_0(x, y, z) \tag{4}$$

for comparison of the results of computations and observational data, where $q_0(x, y, z)$ is the mean daily SO_2 concentration during the preceding 24-hour period.

Before proceeding to an exposition of the results of numerical experiments we will discuss two matters related to representation of equation (1) and checking of the conservative character of the scheme.

The equation for the transport of a mass of impurity was reduced to dimensionless form using the following expressions for its coefficients:

$$x = x_6 L;\quad z = z_6 \frac{L}{m};\quad u = u_6 V;\quad v = v_6 V;\quad w = w_6 \frac{V}{m};$$

$$t = t_6 \frac{L}{V};\quad K_s = K_{s,6} LV;\quad K_z = K_{z,6} \frac{LV}{m^2};\quad \beta = \frac{\beta_6 K_{z,6} V}{m}. \tag{5}$$

[6 = dimensionless] where L is the horizontal scale, which was stipulated at 10^6 m; V is the characteristic velocity (10 m/sec); m is a scale factor determining the relationship between the horizontal and vertical scales (5×10^2); the subscript "6" indicates dimensionless values.

In making the numerical experiments the difference scheme was checked for the conservation of mass. The balance of mass was computed in the entire volume in each time interval on the assumption that the impurity is introduced by a single source situated at an adequate distance from the lateral boundaries in order to exclude their influence and also with $\beta = 0$.

$$\sum_{i=1}^{N} F_i \Delta t = \sum_{i=1}^{N} (q_i^{n+1} - q_i^n) + \sum_{i=1}^{N} (R_i + S_i) \Delta t - \sum_G \left(K_z \frac{\partial^2 q}{\partial z^2}\right)_i \Delta t + \sum_G \frac{\partial}{\partial z} (wq)_i \Delta t, \tag{6}$$

where N is the number of points in the considered redgion; the last two terms on the right-hand side of expression (6) designate the loss of mass through the upper boundary (G) as a result of diffusion and vertical movements. For a more precise evaluation of the loss of impurity through the boundary the derivatives were approximated with a second order of accuracy.

$$\frac{\partial^2 q}{\partial z^2} \approx \frac{1}{\Delta z^2} (2 q_n - 5 q_{n-1} + 4 q_{n-2} - q_{n-3}),$$

$$\frac{\partial}{\partial z} (wq) \approx \frac{1}{\Delta z} \left(- 2 (wq)_{n-1} + \frac{3}{2} (wq)_n + \frac{1}{2} (wq)_{n-2}\right). \tag{7}$$

The results of the following numerical experiments relate to solution of equation (1) without allowance for vertical movements.

Model computations, using equation (1) and its dimensionless analogue, obtained using expressions (5), indicated that the mass deficit -- the difference between the left and right sides of formula (6), is reduced by a factor of 3 when using the dimensionless form of equation (1). These results can be attributed to the fact that with a changeover to a dimensionless analogue of equation (1) there is a decrease in the relationship between the horizontal and vertical intervals, and also between the terms of the difference equation.

Now we will examine the problem of the approximation of boundary conditions at the ground surface. In a preceding study [2] $\partial q/\partial z$ in equation (2) was approximated with the first order of accuracy by means of introduction of a fictitious level, in accordance with [4]. In this case the value of the first elimination coefficient

$$a_0 = \frac{K_z/\Delta z - \beta/2}{K_z/\Delta z + \beta/2}$$

to a high degree was dependent on the relationship between K_z and β. With $a_0 \to 1$ the solution became little stable. In order to improve the approximation use was made of a principle proposed by A. A. Samarskiy [7] and based on expansion of $\partial q/\partial z$ into a Taylor series at a boundary point and its determination through the solved equation. Then the first elimination coefficients assume the following form:

$$a_0 = \frac{K_z}{K_z + \beta\Delta z + 2\Delta z^2/\Delta t},$$

$$b_0 = \frac{(K_z q_1 - K_z q_0 - q_0\beta\Delta z + q_0 2\Delta z^2/\Delta t)}{K_z + \beta\Delta z + 2\Delta z^2/\Delta t}, \qquad (8)$$

where Δz and Δt are time and height intervals and the subscripts on q indicate the level at which this value was selected. In such an approximation method the question arises: what is the value of the sought-for function at the level adjacent to the boundary which should be used in computing a_0 and b_0? We tested two variants: 1) at $z = 1$ -- use the value $q^{n+2/3}$ with a correction for scattering by diffusion; 2) at $z = 1$ -- take the value q^n. Model computations indicated that the use of q^n at $z = 1$ gives a better value of the deficit of mass of impurity (3%) in comparison with $q^{n+2/3}$, at which the mass deficit is 11%.

Due to the fact that most of the sources are situated in the lower part of the boundary layer, a more detailed allowance for the structure of this layer is desirable. A solution of this problem was carried out in two directions:

-- the initial algorithm, using a uniform vertical interval, was modified by the introduction of an additional level (150 m) in the lower 300 m;

-- the change in the coefficient of vertical diffusion with altitude was taken into account.

The introduction of an additional level considerably reduces the deficit of impurity mass: whereas the use of a large interval in the lower part of the atmosphere gave an excess of the remaining mass by 17% in comparison with that introduced, when the grid was made finer in a downward direction the mass deficit was reduced to 2-3%. The sharp decrease in mass imbalance of the impurity observed with the introduction of the additional level is attributable to the fact that this level is introduced in a layer of a considerable concentration gradient (the source in our case was stipulated at a height of 300 m). As was demonstrated in [3], the closer the initial dispersion is to the selected grid interval, the lesser is the value of the residual term with the replacement of the differential equation by a difference equation. The concentration levels in the entire layer in the case of a nonuniform interval were somewhat lower and above the source its sharper decrease was observed.

The inclusion in the model of a variable with height K_z (in our case K_z increases to 300 m and decreases upward) leads to a systematic excess (up to 9%) in the mass of matter remaining in the considered volume in comparison with the mass introduced. The reason for this becomes understandable if we represent the term describing vertical diffusion as

$$\frac{\partial}{\partial z}\left(K_z \frac{\partial q}{\partial z}\right) = \frac{\partial K_z}{\partial z}\frac{\partial q}{\partial z} + K_z \frac{\partial^2 q}{\partial z^2}.$$

When K_z changes with height, the vertical transport value is determined by both terms, and in the last analysis is dependent on the K_z and q profiles. It has been established theoretically and experimentally [6, 8] that both functions decrease with height, except for the lower 200-300 m, and therefore the contribution

will almost always be positive. In other words, the introduction of the variable K_z into the equation for the transport of mass will lead to a nonstationary solution. With respect to the concentration profile, we note that with a constant K_z there will be slower mixing in the layer where the source is situated and therefore a smoother change in q with height than with variable K_z.

Now we will proceed to the formulation of a difference scheme which takes vertical movements into account. The inclusion of w in the model for the transport of contaminating substances involves two difficulties. One of these is determined by the complexity of computation of the vertical velocity components. The second is associated with the introduction of the term $\partial q/\partial z$ into the model. The considerable anisotropicity of the process in horizontal and vertical directions requires the creation of a stable numerical scheme which would be economical.

We computed the transport of impurity for a distance of 1,500-2,000 km on the basis of the real wind field at the standard isobaric levels. Accordingly, as a first approximation it was decided that vertical movements would be computed from the continuity equation, using the same information. In order to increase accuracy and obtain a greater smoothness the derivatives were approximated with a second order of accuracy using a six-point scheme within the region

$$\frac{\partial}{\partial x} \approx \frac{1}{2h}\left[(\varphi_{j,i+1} - \varphi_{j,i-1}) - \frac{1}{4}(\varphi_{j-1,i+1} - \varphi_{j-1,i-1}) - \frac{1}{4}(\varphi_{j+1,i+1} - \varphi_{j+1,i-1})\right], \tag{9}$$

and using a three-point scheme at the boundary. For example, for i = 0 (the x coordinate) the derivative was computed using the formula

$$\frac{\partial}{\partial x} \approx \frac{1}{h}\left(2\varphi_{j,1} - \frac{3}{2}\varphi_{j,0} - \frac{1}{2}\varphi_{j,2}\right). \tag{10}$$

Vertical movements were determined using layers with a thickness of 500 m. A qualitative evaluation demonstrated a satisfactory agreement of the computed fields of vertical movements with the nature of the pressure field.

A numerical scheme for the solution of equation (1) with the conditions (2)-(3) or (4) belongs to the splitting method. The principle for its construction is that the operators for each direction of coordinates are reduced to the upper time level and are represented in the form of a product. At the lower time level we select an expression satisfying the approximations of the differential equation and stability. An important difference in our scheme is in the method for obtaining the products of the operators. Adhering to the idea expressed by Ye. G. D'yakonov in [5], the product of the operators present on the right side of the difference scheme is

$$\prod_{s=1}^{3}\left(E+\frac{t}{2}\Lambda_s\right)\varphi^{n+1}=\prod_{s=1}^{3}\left(E-\frac{t}{2}\Lambda_s\right)\varphi^{n}+ \qquad (11)$$
$$+t\sum_{s=1}^{3}\overline{\Lambda}_s\varphi^{n}+t(F-R-S),$$

where $\Lambda_s\approx V_s\frac{\partial q}{\partial x_s};\ \overline{\Lambda}_s\approx K_s\frac{\partial^2 q}{\partial x_s^2};$

s = 1, 2, 3; E is a unit operator, realized in the first time interval. However, taking into account the great difference in the order of the advective transport terms and the scattering by diffusion [2], the operators describing advection were solved alternately in all the intermediate intervals, and diffusion, the gains and losses were included only in unit time intervals. Taking what has been said into account, the scheme for realizing (11) can be written in the form

$$\left(E+\frac{t}{2}\Lambda_1\right)\varphi^{n+1/3}=\sum_{s=1}^{3}\left(E-\frac{t}{2}\Lambda_s\right)\varphi^{n}+t\sum_{s=1}^{3}\overline{\Lambda}_s\varphi^{n}+t(F-R-S),$$
$$\left(E+\frac{t}{2}\Lambda_2\right)\varphi^{n+2/3}=\varphi^{n+1/3}, \qquad (12)$$
$$\left(E+\frac{t}{2}\Lambda_3\right)\varphi^{n+1}=\varphi^{n+2/3}.$$

A solution of system (12) was obtained by the elimination method. On the basis of the results described above, we introduced an additional level in the lower layer and $w\,\partial q/\partial z$ was approximated using the expression

$$w\frac{\partial q}{\partial z}\approx\frac{2}{z_{i+1}-z_{i-1}}\left[\frac{w_i-|w_i|}{2}\,\frac{z_{i+1}-z_i}{z_i-z_{i-1}}(q_{i-1}-q_i)+\right. \qquad (13)$$
$$\left.+\frac{w_i+|w_i|}{2}\,\frac{z_i-z_{i-1}}{z_{i+1}-z_i}(q_i-q_{i-1})\right].$$

The same as the scheme without vertical movements, conservancy was determined by expression (6).

Numerical experiments carried out with a different number of levels selected at different heights indicated that the use of seven levels in comparison with five considerably improves the mass deficit (1%), whereas with five computation levels the deficit is 12%.

It is interesting to evaluate the influence of vertical movements on the vertical distribution of the concentration. It was found that with the introduction of vertical movements there is a more uniform distribution of the impurity than without them. In addition, the introduction of vertical velocities reduces the time for stabilizing the solution, that is, with the introduction of w the solution becomes stationary 36 hours after the initial moment, whereas when they are absent a complete stabilization of the solution does not occur even after 48 hours.

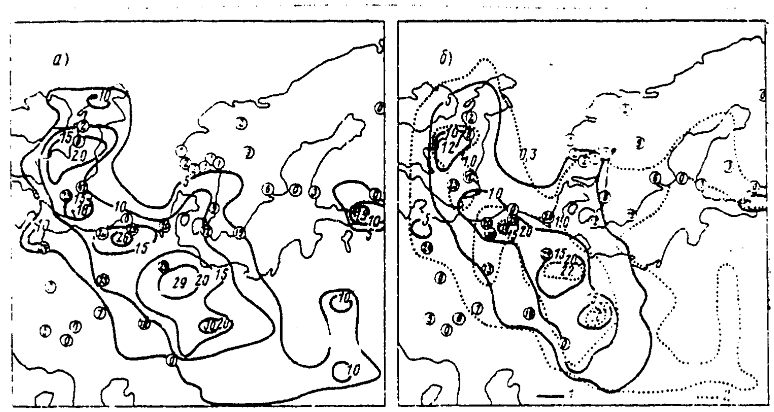

Fig. 1. Computed mean daily concentration of sulfur dioxide ($\mu g/m^3$) at ground surface (1) with allowance for vertical movements (a) and without allowance for vertical movements (b) and precipitation of sulfur dioxide onto surface ($g/(m^2 \cdot day)$) (2) on 15 September 1974. The observed concentration values are indicated in circles.

Figure 1 gives the computed concentration fields with and without allowance for vertical movements. As can be seen from a comparison with the measured values, a model with w better describes the concentration field than without w. Appreciable differences are observed in southern Scandinavia, in northern France and in southern England.

In solution of the problem of propagation of an impurity an important question is the form of representation of SO_2 effluent. In further developing the model we used data on SO_2 effluent obtained as a result of implementation of the "Joint Program for Evaluating the Distant Transport of Air Contaminants in Europe" [8] and representing the SO_2 flux (tons/(km^2·year)) from an area of 127 x 127 km^2. In the case of one-dimensional models the conversion from the flux to the influx of SO_2 (μg/(m^3·hour) is accomplished by normalization to the height of the considered layer. This assumption is not correct from the point of view of experimental reliability [8]. The entry of SO_2 into the atmosphere occurs for the most part through stacks whose height is from 40 to 150 m and the effective height of the effluent [1] is in the range 100-600 m.

We made the assumption that the ejection of SO_2 from a definite area corresponds to the total influx to the layer into which the contaminating substances enter. For conversion from the flux of impurity to the influx we introduced the function $F = -\partial Q/\partial z$, which was computed by layers. Such a representation can be used in a more realistic description of the vertical nonuniformity of effluent.

The influence of different stipulation of the SO_2 discharge on the distribution and degree of the surface concentration was evaluated using a variant of the model with the introduction of real effluent and with stipulation of the initial concentration at the earth in the territory of Eurasia. The concentration fields were computed using the real wind for a 48-hour period and were compared with the measured SO_2 levels in the European network of stations. Three experiments were computed: 1) the SO_2 mass was introduced into a 1,000-m layer and uniformly distributed vertically; 2) the SO_2 influx to a layer with a thickness of 400 m was stipulated in the form $\partial Q/\partial z$ and was introduced at the level 500 m; 3) an experiment similar to the second, but the SO_2 mass was nonuniformly distributed in the 625-m layer (in 250 m -- 20%, in 500 m -- 80%). An analysis of the results indicated that the distribution of centers of high concentration, like its value, coincide more with the real data when the sources are described by the third method (Fig. 1).

The precipitation of SO_2 onto the underlying surface was determined by time integration of the flux at the ground, determined from condition (2). The quantity of SO_2 in mg/m^2 in 48 hours is shown in Fig. 1b. We note the fact that although the concentration levels in regions remote from large industrial centers (for example, Scandinavia) are small, the precipitation of SO_2 here is close to the quantities observed over central Europe.

Finally, the SO_2 flux across elements of the boundary was computed using the formula

$$Q = \overline{q}_{n.\,z}\,\overline{V}_{n.\,z}\cos(\alpha), \tag{14}$$

where $\bar{q}_{n,z}$ and $\bar{V}_{n,z}$ are the concentration and velocity at the center of an area with sides equal to the length of an element of the boundary and the vertical interval, obtained by means of interpolation from the nearest points of grid intersection; α is the angle between the direction of the normal to an element of the boundary and the extension of the vector determining the wind direction.

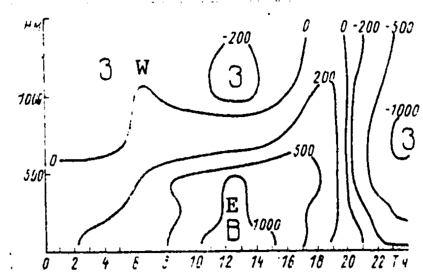

Fig. 2. Vertical time section of fluxes (W -- westerly, E -- easterly) of sulfur dioxide through segment of western boundary of USSR observed on 13-14 September 1974.

Figure 2 shows a vertical time section of impurity fluxes (in kg/hour) across a segment of the western boundary of the USSR (from Kaliningrad to the Gulf of Riga) as observed on 13-14 September 1974. We should note the considerable variability of the fluxes both vertically and with time. The latter circumstance is particularly important in computing the quantity of matter transported across the boundaries of individual countries.

BIBLIOGRAPHY

1. Berlyand, M. Ye., SOVREMENNYYE PROBLEMY ATMOSFERNOY DIFFUZII I ZAGRYAZNENIYA ATMOSFERY (Modern Problems in Atmospheric Diffusion and Atmospheric Contamination), Leningrad, Gidrometeoizdat, 1975.

2. Vel'tishcheva, N. S., "Numerical Model of Distant Transport of Sulfur Dioxide," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 9, 1977.

3. Vel'tishcheva, N. S., "Numerical Solution of the Turbulent Diffusion Equation in the Field of a Variable Wind," TRUDY GIDROMETTSENTRA SSSR (Transactions of the USSR Hydrometeorological Center), No 139, 1974.

4. Godunov, S. K., Ryaben'kiy, V. S., VVEDENIYE V TEORIYU RAZNOSTNYKH SKHEM (Introduction to the Theory of Difference Schemes), Moscow, Fizmatgiz, 1962.

5. D'yakonov, Ye. G., "Difference Schemes With a Splitting Operator for General Second-Degree Parabolic Equations With Variable Coefficients," ZHURNAL VYCHISLITEL'NOY MATEMATIKI I MATEMATICHESKOY FIZIKI (Journal of Computational Mathematics and Mathematical Physics), Vol 4, No 2, 1964.

6. Zilitinkevich, S. S., DINAMIKA POGRANICHNOGO SLOYA ATMOSFERY (Dynamics of the Atmospheric Boundary Layer), Leningrad, Gidrometeoizdat, 1974.

7. Samarskiy, A. A., VVEDENIYE V TEORIYU RAZNOSTNYKH SKHEM (Introduction into the Theory of Difference Schemes), Moscow, Fizmatgiz, 1971.

8. THE OECD PROGRAMME ON LONG-RANGE TRANSPORT OF AIR POLLUTANTS, OECD, Paris, 1977.

UDC 551.(583+521.14+524.34)

EFFECT OF CHANGE IN ALBEDO OF THE EARTH'S SURFACE ON THE EARTH'S THERMAL REGIME

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 50-56

[Article by Doctor of Geographical Sciences N. A. Yefimova, State Hydrological Institute, submitted for publication 15 January 1980]

[Text]

Abstract: The author evaluates the influence of the feedback between changes in mean global air temperature and albedo of the earth's surface (including changes in albedo with a change in vegetation cover) for conditions of warming of climate in the example of the epoch of the Early Pliocene and for conditions of cooling during the last glaciation.

Introduction. In a study of climatic changes with the use of semi-empirical models it is necessary to take into account the feedbacks between the thermal regime of the atmosphere and outgoing radiation, air humidity and albedo of the earth's surface. It must be remembered that the influence of the first two factors on the thermal regime is manifested in all cases, whereas the third is manifested when there are more or less prolonged climatic changes.

Changes in the albedo of the earth's surface when there are variations in the thermal regime occur as a result of an increase or a decrease in the area of the polar ice, snow cover, and also due to changes in the types of vegetation on the continents.

An allowance for feedback between the thermal regime and the snow-ice cover was introduced in the studies of M. I. Budyko (1968) and Sellers (1969).

In the studies of Manabe and Wetherald (1975, 1979) the feedback between the the thermal regime and changes in snow cover on the continents and sea ice was taken into account separately.

Cess (1978) brought attention to the need for allowance for the feedback between the thermal regime and the change in the albedo of the vegetation cover on the continents. An evaluation of the change in albedo of the vegetation cover and its influence on air temperature was made by Cess in the example of the epoch of the last glaciation on the basis of data from CLIMAP (1976) and Gates (1976).

In this study an attempt is made to evaluate the overall effect of all the components of the feedback between change in the mean global air temperature and albedo of the earth's surface. These evaluations were made for conditions of climatic warming in the example of the epoch of the Early Pliocene (about 6 million years ago) and for the conditions of cooling during the last glaciation (about 18,000-20,000 years ago). The results are compared with paleoclimatic data and the results of computations using a climatic model.

Warm climate, Early Pliocene. As the initial data for determining albedo of the surface of the continents under the conditions of the warm climate of the Early Pliocene we used the maps prepared by V. M. Sinitsyn (1965, 1967) showing the distribution of the vegetation cover, air temperatures of the warmest and coldest months, and the annual sums of precipitation over the territory of Eurasia. On the basis of these data, with modern climatic analogues and types of vegetation into account, it was possible to reconstruct the annual variation of temperature and the characteristics of the snow cover in the high latitudes. Then, using a known method (Budyko, 1971; Yefimova, 1977), we determined the mean monthly and annual albedo values for the territory of Eurasia and the northern half of Africa. The mean annual albedo values obtained in this way for the latitudinal zones of the mentioned continents were deemed characteristic for the latitudinal zones of the continents in the northern hemisphere as a whole by analogy with the circumstance that in the modern epoch the mean albedo values for the latitudinal zones of Eurasia are extremely close to those for the northern hemisphere.

The mean annual albedo values for the continents in the present epoch were obtained on the basis of materials used in constructing maps of the heat balance components (Budyko, et al., 1978; Mukhenberg, 1967). When determining the albedo of the continents for the modern epoch an allowance was made for the real areas of forested expanses, meadows and agricultural fields.

According to paleoclimatic data, the climate of the Early Pliocene was warmer and moister in comparison with the modern epoch. In the middle and high latitudes of Eurasia and North America the vegetation cover was characterized by a richer species composition and the forests occupied extensive territories extending to the northern shores of the continents and to the south of the boundary of the present-day forested zone.

The snow cover was situated considerably to the north and was briefer than at the present time. As is well known, the albedo of the forests both during the growing season and in the presence of a snow cover is less than the albedo of tundra, grassy vegetation and thin forests. Accordingly, over a great part of the surface of the continents the albedo was less than is the case at the present time.

The mean albedo of the land surface in the northern hemisphere in the Pliocene was 0.060 less than the present-day value (the difference in albedo of the land in the northern hemisphere between the present-day epoch and the Pliocene was obtained as the mean weighted value of the differences in the albedos of the latitudinal zones, taking into account the solar radiation levels incident on the earth's surface).

Taking into account the relationship of the areas of the land and ocean, the mean difference in albedos between the present epoch and the Pliocene for the entire surface of the northern hemisphere was 0.024.

In order to evaluate the influence of change in albedo of the surface of the continents on the thermal regime we determined the difference in albedo of the earth - atmosphere system in the considered epochs ($\Delta\alpha_s$). It was obtained by summing the differences in albedo of the earth's surface of the latitudinal zones ($\Delta\alpha^1, \ldots, \Delta\alpha^n$), taking into account the influence exerted on planetary albedo by Rayleigh scattering and cloud cover (N) in the form proposed by Cess (1978):

$$\Delta\alpha'_s = \Delta\alpha' (0,69 - 0,52\ N),$$

where $\Delta\alpha'_s$ is the difference in zonal albedo of the earth - atmosphere system.

The mean zonal differences in albedo are given in the table. The mean difference in albedos of the earth - atmosphere system for the entire northern hemisphere between the present epoch and the Early Pliocene was 0.0104 (in the averaging allowance was made for the dependence of receipts of solar radiation at the upper boundary of the atmosphere on latitude). This value was governed by the change in albedo of the continents as a result of change in the types of vegetation and snow cover. Some of the difference in albedo caused by differences in the vegetation cover was obtained from a comparison of present-day data on albedo of the surface of the continents during the summer months with the albedo in the Pliocene under conditions of absence of a snow cover over the entire considered territory. This part of the difference in albedos was equal to 0.0040; the remaining part -- 0.0064 -- was related to differences in the distribution and duration of presence of snow cover during these epochs.

It is known from paleogeographic investigations that at the end of the Tertiary the area of the sea ice in the arctic basin was considerably less than today. In accordance with the data of V. M. Sinitsyn on the thermal regime of the cold season in the northern part of Eurasia it can be assumed that in the Early Pliocene the boundary of the sea polar ice ran approximately 15° of latitude to the north in comparison with the present time.

An evaluation of change in albedo of the earth - atmosphere system due to this contraction of the area of sea ice indicated that it decreased by 0.0035. Thus, the total decrease in albedo of the earth - atmosphere system during the Pliocene was 0.0139, including 0.0104 as a result of the change in the albedo of the continents and 0.0035 due to a decrease in the area of the sea ice. The contribution to the total change in albedo from the change in albedo of the vegetation cover was substantial, constituting about 30% of the total change in albedo of the earth - atmosphere system. We note that this evaluation applies to the northern hemisphere. For the earth as a whole the general change in the albedo of the earth - atmosphere increased somewhat due to the great change in area of the polar ice and decreased due to the lesser area of the continents in the southern hemisphere; the corresponding albedo change was 0.0102 (0.0050 due to the decrease in the area of sea ice, 0.0052 as a result of decrease in the snow cover and change in vegetation). The contribution of the difference in the state of the vegetation cover to the change in albedo for the earth as a whole was less than the evaluation cited above, constituting about 20%.

Cold climate. Epoch of the last glaciation of the Pleistocene. In determining the albedo of the continents in the epoch of the last glaciation of the Pleistocene (about 18,000-20,000 years ago) we used CLIMAP data (1976), the paleogeographic generalizations of A. A. Velichko (1973), I. T. Avenarius, M. V. Muratova, et al. (1978) and others. In these investigations it was established that under conditions of a cold and in many regions of more arid climate of the glacial period the forest vegetation on the continents was replaced in the high latitudes by tundras, in the temperate latitudes by tundra steppes and thin forest, in the lower latitudes by savannas and steppes. There was a considerable broadening of zones of dry steppes, semideserts and deserts. In accordance with this change in the vegetation cover, and also in connection with the occurrence of continental glaciation in the high and temperate latitudes there was an increase in the albedo of the surface of the continents.

Taking these changes into account, we determined the mean annual albedo values and the difference in albedo of the surface of the continents in the glacial period and the modern epoch for the earth's latitudinal zones. The mean difference in the albedo of the land surface during the period of glaciation and the present epoch (obtained by averaging the differences in the albedos of the land by 10° latitudinal zones, with allowance for their areas and the distribution of solar radiation) was 0.083.

Taking into account the ratio of the areas of the land and ocean, we computed the mean zonal differences and then the total difference in albedo of the earth's surface caused by the change in albedo of the continents during the glacial period in comparison with the present epoch, which was 0.024.

Taking into account the already considered dependence of planetary albedo on cloud cover and the receipts of solar radiation we computed the mean zonal (see table) and total mean difference in the albedos of the earth - atmosphere system during the glacial period, caused by an increase in the albedo of the continents as a result of change in the vegetation cover and the presence of continental glaciation. This value was found to be equal

Mean Zonal Albedo Differences

	1 Широтные			
	90—80	80—70	70—60	60—50
	2 Современная эпоха —			
4 Северное полушарие	0,0575	0,1416	0,0799	0,0288
	3 Современная эпоха —			
4 Северное полушарие	0	—0,0958	—0,1343	—0,0794
5 Южное полушарие	0	0	—0,1416	—0,1469

to 0.0105. It was found that the contribution of the increase in continental glaciation to the change in albedo of the earth - atmosphere system is 0.0041, whereas the remaining part — 0.0064 — is attributable to the change in albedo due to a change in snow cover and vegetation (0.0030 and 0.0034 respectively).

An evaluation of the influence of an increase in the area of polar sea ice on change in albedo during the glaciation period indicated that because of this the albedo of the earth - atmosphere system increased by 0.0110 in comparison with the present epoch. Most of this albedo difference (0.0090) is attributable to the extensive occurrence of a zone of sea ice in the southern hemisphere. Thus, the total increase in albedo of the earth - atmosphere system in the glacial period in comparison with the present epoch was 0.0215; the contribution of change in the albedo of vegetation (0.0034) to the total change in the planetary albedo was about 15%.

In a study by Cess (1978) the change in the albedo of the vegetation cover in the glacial period is about 0.01 or 40% of the total change in the planetary albedo, which also was somewhat greater (0.025) than that obtained in our study. These differences are probably partially attributable to the fact that Cess determined the differences in albedo for July, whereas here we have made computations of the mean annual values. Another reason for the indicated difference is the noncoincidence of the data used on the albedo of the surface of the continents. We feel that the data of Posey and Clapp (1964) on the albedo of the continents in July for the present epoch, which Gates and Sess used in the computations, are considerably too low. For example, in large territories occupied by different types of vegetation — from forests and thin forests to steppes, praries and savannas, they adopted an albedo of 0.07-0.15, whereas using the observational data which we employed in constructing the maps the albedo of these types of vegetation is from 0.12 to 0.22.

Effect of feedback of albedo changes on thermal regime. The changes in albedo of the earth - atmosphere system in the Early Pliccene and during the last glaciation make it possible to evaluate the changes in mean global air

Earth - Atmosphere System ($\Delta\alpha'_s$)

зоны, град				
50—40	40—30	30—20	20—10	10—0
6 млн. лет назад				
0,0155	0,0068	0,0036	0,0021	0,0004
18—20 тыс. лет назад				
—0,0451	—0,0033	—0,0008	—0,0021	—0,0044
—0,0047	—0,0058	—0,0030	—0,0030	—0,0026

KEY:

1. Latitudinal zones, degrees
2. Modern epoch -- 6 million years ago
3. Modern epoch -- 18,000-20,000 years ago
4. Northern hemisphere
5. Southern hemisphere

temperature at the earth's surface caused by this factor. Assuming that with a change in planetary albedo by 0.01 the global temperature changed by 2.1°C (Budyko, 1974, 1979) we find that in the Pliocene the change in the mean annual air temperature at the earth's surface as a result of change in the albedo of the earth - atmosphere system was about 2.1°C, including 0.4°C due to a change in the albedo of the vegetation cover, 0.7°C -- due to changes in the snow cover and 1° due to change in the area of sea ice. Taking into account that the approximately doubled content of carbon dioxide in the atmosphere in this epoch leads to a temperature increase by approximately 2.5°C (Budyko, 1972, 1977), we find that the total increase in air temperature in the Early Pliocene in comparison with the modern epoch is about 4.6°C.

The increase in air temperature in the northern hemisphere (4.6°C) obtained for the Early Pliocene is very close to the similar value determined from the paleoclimatic maps prepared by V. M. Sinitsyn (4.8°C).

During the period of the last glaciation the decrease in air temperature as a result of the increase in planetary albedo, in accordance with the data cited above, was 4.5°C, including by 0.7°C due to the change in the albedo of vegetation, by 0.6°C -- due to change in the snow cover, by 0.9°C -- due to an increase in continental glaciation and by 2.3°C -- due to an increase in the area of polar sea ice, especially in the southern hemisphere. The difference in the mean global annual temperatures during the glacial period and in the modern epoch (4.5°C) found here is close to the values obtained for July by Gates (1976) -- 4.9°C and by Cess (1978) — about 5°C.

It is obvious from the materials cited here that in the study of climatic changes transpiring over long time intervals there is a need to take into account all the principal factors exerting an influence on the albedo of the earth's surface, including the state of the vegetation cover.

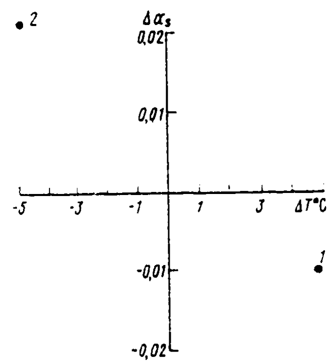

Fig. 1. Dependence of change in albedo of the earth - atmosphere system $\Delta\alpha_s$ on change in global temperature ΔT. 1) Pliocene - modern epoch, 2) last glaciation - modern epoch.

Influence of changes in air temperature on planetary albedo. Figure 1 illustrates the dependence of planetary albedo on changes in mean global air temperature at the earth's surface. In this figure $\Delta\alpha_s$ denotes the change in albedo of the earth - atmosphere system in comparison with the modern value, ΔT is the difference in mean air temperature relative to the modern epoch, determined independently on the basis of paleoclimatic data and computations using climatic models.

We note that the accuracy of the data used in constructing this graph is limited, in particular, due to the failure to take into account the influence exerted on planetary albedo by changes in the cloud cover.

It is evident that the feedback between air temperature and albedo of the earth's surface considerably intensifies the sensitivity of the thermal regime to variations in the heat influx. Taking into account the dependence of temperature of the lower air layer on albedo of the earth - atmosphere system cited above, it can be concluded that the relationship represented in Fig. 1 doubles the sensitivity of the thermal regime to variations of the heat influx with an increase in the concentration of carbon dioxide in the atmosphere in the Pliocene.

A still greater role is played by the change in albedo in development of cooling during the last glaciation. Assuming that this glaciation was caused by the redistribution of the solar energy arriving during the warm and cold seasons of the year at different latitudes, we can conclude that change in the albedo of the earth's surface was of decisive importance for the mentioned cooling.

Although the influence of changes in the albedo of the earth's surface is fully manifested for prolonged climatic variations occurring during time intervals during which continental ice covers are formed or are destroyed, it is also of importance for relatively short climatic changes transpiring over the course of decades and centuries which lead to a change in the area of sea ice, snow cover and vegetation on the continents.

BIBLIOGRAPHY

1. Avenarius, I. G., Muratova, M. V., Spasskaya, I. I., PALEOGEOGRAFIYA SEVERNOY YEVRAZII V POZDNEM PLEYSTOTSENE, GOLOTSENE I GEOGRAFICHESKIY PROGNOZ (Paleogeography of Northern Eurasia and the Late Pleistocene, Holocene and Geographic Prediction), Moscow, Nauka, 1978.

2. Budyko, M. I., "Origin of the Glacial Epochs," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 11, 1968.

3. Budyko, M. I., KLIMAT I ZHIZN' (Climate and Life), Leningrad, Gidrometeoizdat, 1971.

4. Budyko, M. I., VLIYANIYE CHELOVEKA NA KLIMAT (Man's Influence on Climate), Leningrad, Gidrometeoizdat, 1972.

5. Budyko, M. I., IZMENENIYE KLIMATA (Climatic Change), Leningrad, Gidrometeoizdat, 1974.

6. Budyko, M. I. SOVREMENNOYE IZMENENIYE KLIMATA (Modern Change in Climate), Leningrad, Gidrometeoizdat, 1977.

7. Budyko, M. I., PROBLEMA UGLEKISLOGO GAZA (The Carbon Dioxide Problem), Leningrad, Gidrometeoizdat, 1979.

8. Budyko, M. I., et al., TEPLOVOY BALANS ZEMLI (The Earth's Heat Balance), Leningrad, Gidrometeoizdat, 1978.

9. Velichko, A. A., PRIRODNYY PROTSESS V PLEYSTOTSENE (Natural Process in the Pleistocene), Moscow, Nauka, 1973.

10. Yefimova, N. A., RADIATSIONNYYE FAKTORY PRODUKTIVNOSTI RASTITEL'NOGO POKROVA (Radiation Factors in Productivity of the Vegetation Cover), Leningrad, Gidrometeoizdat, 1977.

11. Mukhenberg, V. V., "Albedo of the Earth's Solid Surface," TRUDY GGO (Transactions of the Main Geophysical Observatory), No 193, 1967.

12. Sinitsyn, V. M., DREVNIYE KLIMATY YEVRAZII, CH. 1. PALEOGEN I NEOGEN (Ancient Climates of Eurasia, Part 1. Paleogene and Neogene), Leningrad, Izd-vo LGU, 1965.

13. Sinitsyn, V. M., VVEDENIYE V PALEOKLIMATOLOGIYU (Introduction to Paleoclimatology), Leningrad, Nauka, 1967.

14. "CLIMAP Project Members. The Surface of the Ice-Age Earth," SCIENCE, Vol 191, 1976.

15. Cess, R. D., "Biosphere-Albedo Feedback and Climate Modelling," J. ATMOS. SCI., Vol 35, 1976.

16. Gates, W. L., "A Numerical Simulation of Ice-Age Climate With a Global General Circulation Model," J. ATMOS. SCI., Vol 33, 1976.

17. Manabe, S., Wetherald, W. T., "The Effect of Doubling the CO_2 Concentration on the Climate of a General Circulation Model," J. ATMOS. SCI., Vol 32, No 2, 1975.

18. Manabe, S., Wetherald, W. T., "On the Horizontal Distribution of Climate Change Resulting from an Increase in CO_2 Content of the Atmosphere," Preprint, 1979.

19. Posey, J. W., Clapp, P. F., "Global Distribution of Normal Surface Albedo," GEOPHYS. INTERN., Vol 4, 1964.

20. Sellers, W. D. A., "A Global Climatic Model Based on the Energy Balance of the Earth-Atmosphere System," J. APPL. METEOROL., Vol 8, No 3, 1969.

UDC 551.(583+521.14+524)(268/269)

DEPENDENCE OF THE ALBEDO OF POLAR ICE ON AIR TEMPERATURE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 57-60

[Article by Candidate of Geographical Sciences L. A. Strokina, State Hydrological Institute, submitted for publication 15 January 1980]

[Text] Abstract: A study was made of the dependence of the albedo of the snow-ice surface on air temperature and the angle of incidence of solar rays in regions of Arctic and Antarctic ice cover for more precise determination of the feedback between the thermal regime and the area of polar ice in models of the theory of climate.

Beginning with the studies of M. I. Budyko [1] and W. Sellers [12], in most modern models of the theory of climate an allowance is made for the feedback between air temperature and the area of the polar ice. Since with an increase in the ice area the albedo of the earth's surface increases, this leads to a decrease of absorbed radiation and a decrease in air temperature. The inverse process occurs with a decrease in the ice area.

Since the considered feedback intensifies the fluctuations of the thermal regime, in accordance with the adopted terminology it is considered positive.

For a correct allowance for the dependence between the thermal regime and the area of the polar ice in models of the theory of climate it is necessary to study the influence of air temperature and the angle of incidence of the solar rays on albedo of the ice cover.

In one of the first studies devoted to this problem it was postulated that with an increase in the area occupied by the ice cover the mean albedo of the earth-atmosphere system in the ice cover zone changes little since a decrease in albedo due to an increase in mean solar altitude in the first approximation is compensated by a change in climatic conditions in the ice zone, increasing the mean albedo [2]. Later Lian and Cess [13] made quantitative computations of the dependence of albedo of the earth-atmosphere system on latitude, with allowance for the influence exerted on albedo by changes in solar altitude. However, they did not take into account the effect exerted on reflectivity of ice by changes in the complex of climatic conditions in connection with spreading of the ice cover. From these computations it follows that there is a decrease in the mean albedo of the ice zone with an increase in its extent.

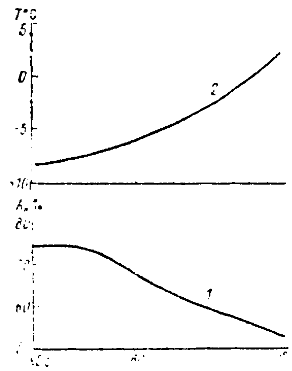

Fig. 1. Latitudinal variation of the albedo of sea ice (1) and air temperature (2) in the Arctic during the April-September period.

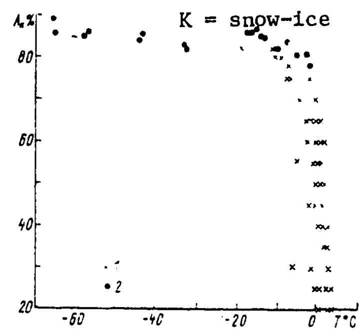

Fig. 2. Dependence of albedo of the snow-ice surface in the Arctic (1) and Antarctica (2) on air temperature in the course of the polar day.

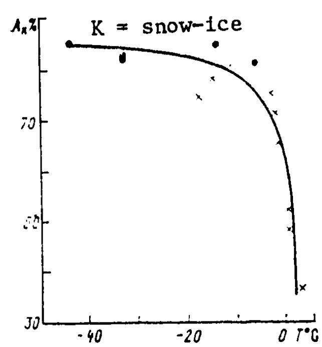

Fig. 3. Dependence of albedo of snow-ice surface in Arctic and Antarctica on air temperature for solar zenith angles from 55 to 65°.

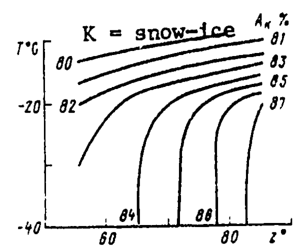

Fig. 4. Change in albedo of snow-ice surface in Arctic and Antarctica in dependence on air temperature and solar zenith angle.

This conclusion evidently does not correspond to the conditions of a real climate. It can be noted that with the existing differences in the extent of the Arctic and Antarctic ice covers the mean albedo of the earth-atmosphere system, according to satellite data, is greater over Antarctic ice, which in area exceeds the zone of arctic ice. This fact is probably attributable to a considerable degree to the above-mentioned effect of a change in climatic conditions, which are associated, in particular, with the appearance of stable regions of high pressure over extensive ice covers. Under such conditions cloud cover is usually limited and the albedo of the earth-atmosphere system is closer to the albedo of the underlying snow surface than in the case of an overcast state of the sky [3].

In order to understand the dependence between the temperature regime and the area of the polar ice it is first of all interesting to clarify the relationship between the albedo of the ice surface and air temperature on the basis of empirical materials. For this purpose we used climatic handbooks and monographs on the meteorological and radiation regimes of the polar region in which most of the data are presented in the form of charts of the mean long-term monthly air temperature and the albedo of sea ice [4, 5, 7-9, 11]. From the mentioned maps of the Arctic Ocean we read the air temperatures and albedo of the sea ice during the period of the polar day (April-September) in the polar region and at the circles of latitude corresponding to 65, 70, 75, 80 and 85°N with a longitude interval of 5°. The mean latitudinal temperature and albedo values determined in this way for the Arctic are presented in Fig. 1 as a function of latitude.

It can be seen that the albedo of the surface of arctic ice changes considerably with latitude, decreasing from the pole to the southern edge of the floating ice by 23% (curve 1). This change in albedo corresponds to an increase in the mean air temperature during the period of the polar day by 11°C (curve 2). It is necessary to clarify to what degree the discovered albedo change is dependent on the thermal regime and to what extent -- on the angle of incidence of the solar rays.

The presence of a dependence between air temperature and ice reflectivity can be postulated from data presented on the graph (Fig. 2). This graph shows the mean monthly data for albedo of the ice cover surface in the Arctic and also data on the albedo of the snow surface in Antarctica, obtained from long-term observations at the stations Mirnyy (66°33'S, 93°01'E) and Vostok (78°28'S, 106°48'E), in dependence on air temperature.

In an examination of the data in Fig. 2 it must be taken into account that the mean monthly albedo of the ice changes not only due to the differences in air temperature, but also as a result of changes in solar altitude, decreasing with an increase in these altitudes.

For a separate analysis of both dependences we found the empirical relationship between the mean monthly albedo values for the polar ice and air temperature for several 10° intervals of the values of the minimum monthly solar zenith angle (z), falling in the range of zenith angles 40-90° during the period of the polar day. Figure 3 shows one such graph, which corresponds to the minimum monthly values of solar zenith angle from 55 to 65°.

In this figure, as in Fig. 2, one can clearly see the correlation between the albedo of the ice surface and the air temperature, which intensifies considerably when the temperatures are above -10°C.

The materials in Fig. 3 and graphs similar to it for other zenith angle intervals made it possible to evaluate the dependence between the mean monthly albedo values and the minimum zenith angle for air temperatures below -10°C (Fig. 4). As might be expected, it was found that with one and the same air temperature with an increase in the solar zenith angle the ice albedo values increase. However, this change in albedo is not great. It does not exceed 4-5% at air temperatures -20°C and below. In the temperature region above -20°C this dependence is still weaker. For higher air temperatures, close to 0°C, it is difficult to establish the dependence of ice albedo on zenith angle on the basis of empirical data because in this case small temperature changes result in considerable variations in the albedo values, which greatly exceed its changes caused by differences in solar altitude.

The slight dependence of the albedo of a snow-ice surface on solar altitude was also pointed out by A. A. Timerev [10]. As a result of an analysis of measurements of albedo in the polar regions the author drew the conclusion that within the limits of observational accuracy the change in reflectivity of the snow-ice surface is caused by a change in the state of the underlying surface. A definite role in the state of snow and ice in the change of albedo was also noted in the investigations of M. S. Marshunova and N. T. Chernigovskiy [4, 11]. It can be seen from the data cited above that in the Arctic with a decrease in air temperature from 4.0 to -6.0°C, when there is a marked change in the state of the ice cover surface, the albedo increases by approximately 50% (Figures 2, 3).

Although the nature of the dependence of albedo of the ice cover on air temperature in the regions of both polar caps is identical, we note some difference in the albedo values in the Arctic and Antarctica with one and the same air temperature values and with identical zenith angle values. The albedo values in Antarctica are usually several percent higher than in the Arctic, which is probably attributable to the greater dryness of the snow, the lesser contamination and more even surface of the snow and ice cover in comparison with the ice cover of the Arctic basin [6, 7].

The mean albedo of the surface of the snow and ice cover in the Arctic Ocean, determined from the above-mentioned data, is 65%. It is difficult to determine the similar value for Antarctica due to the lesser volume of observations made there. However, without question, the mean albedo of the surface of the Antarctic ice cover is considerably greater than the value cited above. Thus, in particular, the albedo value at Mirnyy station, situated at the periphery of the Antarctic glacier, during the period from October through March is 81.4%. On the glacier plateau of Central Antarctica the albedo is still greater.

In this connection it can be surmised that with an increase in the polar ice covers the mean albedo of their surface increases. This favors an intensification of the positive feedback between air temperature and the area of the polar ice.

BIBLIOGRAPHY

1. Budyko, M. I., "On the Origin of the Glacial Epochs," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 11, 1968.

2. Budyko, M. I., IZMENENIYE KLIMATA (Climatic Change), Leningrad, Gidrometeoizdat, 1969.

3. Budyko, M. I., "Semi-Empirical Model of the Thermal Regime of the Atmosphere and Real Climate," METEOROLOGIYA I GIDROLOGIYA, No 4, 1979.

4. Marshunova, M. S., Chernigovskiy, N. T., RADIATSIONNYY REZHIM ZARUBEZHNOY ARKTIKI (Radiation Regime of the Foreign Arctic), Leningrad, Gidrometeoizdat, 1971.

5. METEOROLOGICHESKIY REZHIM ZARUBEZHNOY ARKTIKI (Meteorological Regime of the Foreign Arctic), edited by I. M. Dolgin, Leningrad, Gidrometeoizdat, 1971.

6. Romanov, A. A., "Ice Conditions for Navigation in Antarctic Waters," TRUDY AANII (Transactions of the Arctic and Antarctic Scientific Research Institute), Vol 335, 1976.

7. Rusin, N. P., METEOROLOGICHESKIY I RADIATSIONNYY REZHIM ANTARKTIDY (Meteorological and Radiation Regime of Antarctica), Leningrad, Gidrometeoizdat, 1961.

8. SPRAVOCHNIK PO KLIMATU ANTARKTIDY. T I. SOLNECHNAYA RADIATSIYA, RADIATSIONNYY BALANS, SOLNECHNOYE SIYANIYE (Handbook of the Climate of Antarctica. Vol I. Solar Radiation, Radiation Balance, Sunshine), Leningrad, Gidrometeoizdat, 1976.

9. SPRAVOCHNIK PO KLIMATU ANTARKTIDY. T II. TEMPERATURA VOZDUKHA, ATMOSFERNOYE DAVLENIYE, VETER, VLAZHNOST' VOZDUKHA, OBLACHNOST', OSADKI, ATMOSFERNYYE YAVLENIYA, VIDIMOST' (Handbook on the Climate of Antarctica, Pressure, Wind, Air Humidity, Cloud Cover, Precipitation, Atmospheric Phenomena, Visibility), Leningrad, Gidrometeoizdat, 1977.

10. Timerev, A. A., "Reflective Properties of the Underlying Surface of the Polar Regions," TRUDY AANII (Transactions of the Arctic and Antarctic Scientific Research Institute), Vol 328, 1976.

11. Chernigovskiy, N. T., Marshunova, M. S., KLIMAT SOVETSKOY ARKTIKI (RADIATSIONNYY REZHIM) (Climate of the Soviet Arctic (Radiation Regime)), Leningrad, Gidrometeoizdat, 1965.

12. Sellers, W. D., "A Global Climatic Model Based on the Energy Balance of the Earth-Atmosphere System," J. APPL. METEOROL., Vol 8, No 3, 1969.

13. Lian, M. S., Cess, R. D., "Energy Balance Climate Models; a Reappraisal of Ice-Albedo Feedback," J. ATMOS. SCI., Vol 34, No 7, 1977.

UDC 551.509.314

SMOOTHING OF EMPIRICAL HYDROMETEOROLOGICAL RELATIONSHIPS BY A CUBIC SPLINE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 61-67

[Article by Professor A. R. Konstantinov and N. M. Khimin, Leningrad Hydrometeorological Institute, submitted for publication 28 February 1980]

[Text]

Abstract: The authors examine problems in the theory of splines applicable to an analysis of hydrometeorological processes. The article formulates the problem of nonlinear multiple regression and existing solution methods are evaluated. Splines are regarded as a universal tool for constructing nonlinear relationships between statistically linked variables. Recommendations on the use of splines can also be useful in an analysis of this class of problems in other scientific fields, but the effectiveness of their use is demonstrated in statistical problems of a hydrometeorological nature.

Most hydrometeorological relationships have a nonlinear character. In these cases the use of the multiple linear regression approach leads to substantial errors. Accordingly, researchers have been forced to seek new analytical methods. Among such methods is the "residual method" of statistical analysis, suitable for any form of relationships, including nonlinear relationships. Initially the relationship between the selected function and the first determining argument is taken into account; the residual value of the function is related to the second argument, etc. This method has found extensive application abroad [4, 10, 11, 13]. During recent years it has also been used in Soviet investigations [5].

In order to realize this method it is necessary to construct graphic regressions of the sought-for relationships. Such a construction sometimes to one degree or another has a subjective nature, although it is accomplished using the condition of maximizing of the closeness of the correlation between the values of the function for one and the same arguments, determined by experimental points and regression curves [5, 6]. Some problems in the mathematical approximation of the graphic regression curves were examined in [6].

The method proposed in this article makes it possible to construct the mentioned graphic relationships quite rigorously and objectively. It can also have broader application for constructing any one-dimensional relationships. In developing the method it is assumed a priori that the relationship between the determining factor and the phenomenon is expressed by a smooth function, whereas the experimental data can have a considerable scatter. The construction of the smoothing function is an approximation problem. In this study we give an algorithm for solution of this problem; its realization in some hydrometeorological examples is demonstrated.

Assume that in the segment [a, b] the grid

$$w_n = \{a = x_0 < x_1 < \ldots < x_n = b\}, \tag{1}$$

is stipulated and the values y_i, $i = 0, 1, \ldots, n$ are stipulated at the points x_i. The y_i values can be regarded as quantitative characteristics of some process at the times x_i. As an adequately general mathematical expression we use a model from [2] in the form

$$y_i = \psi(x_i) + \Delta_i. \tag{2}$$

In this model the observed series is regarded as the sum of the determined sequence $\{\psi(x_i)\}$ and the random sequence $\{\Delta_i\}$. These components are usually computed theoretical values.

It is assumed that at least theoretically it is possible to repeat the experiment fully as many times as desired, obtaining new sets of observations. With such repetition the function $\psi(x)$, called the trend, should remain one and the same, but the random components would be different as different realizations of a random process.

The problem of smoothing of experimental data includes the forming of the function $f(x)$, in some sense being the best approximation to the trend $\psi(x)$ in [a, b].

Splines have recently come into wide use for smoothing purposes. We will recall the definition of a polynomial spline. The breakdown (1) is stipulated in the segment [a,b]. The function $S_m(x) = S_m(x, w_n)$ is called a polynomial spline in the breakdown w_n if:

1) $S_m(x) \in P_m$, $x \in [x_i, x_{i+1}]$, $i = 0, 1, \ldots, n-1$;

2) $S_m(x) \in C^{(r-1)}[a, b]$,

where P_m is a set of polynomials of a degree not higher than m, $m > 0$, but $C^{(k)}[a,b]$ is a set of functions in [a,b] having a continuous k-th derivative.

There is a series of approaches to solution of the smoothing problem leading to splines (see review [12]). A distinguishing characteristic of all these approaches is that in actuality for constructing a smoothing function it is necessary to know the value of the standard error in measuring the y_i value. We will examine one of these problems, the algorithm of whose solution is given in [14]; we will use it in the text which follows.

It is necessary to find the function f(x) by which is attained

$$\min I(g) = \int_a^b g''(x)^2\,dx \tag{3}$$

amongst all $g(x) \in C^{(2)}$ [a,b] and satisfying the inequality

$$\sum_{i=0}^{n} \left(\frac{g(x) - y_i}{\delta y_i} \right)^2 \leq S. \tag{4}$$

Here $\delta y_i > 0$ and $S \geqslant 0$ are stipulated numbers.

In other words, among all the functions for which the deviation from the measured values with the weights δy_i^{-2} does not exceed a stipulated value we seek a function which is smoothest in the sense (3). The integral in (3) gives a good approximation for the integral of the square of curvature of the curve y = g(x). If an evaluation of the standard error of the ordinate y_i is used as δy_i, S should fall within the interval $(N - (2N)^{1/2}, N + (2N)^{1/2})$, N = n + 1. A solution of this problem is a cubic spline which we will represent in the form

$$f(x) = \sum_{j=0}^{3} a_{ij} (x - x_i)^j, \quad x_i \leq x \leq x_{i+1}. \tag{5}$$

If y_i is the result of a laboratory experiment, the Δ_i values in (2) for the most part are governed by the errors introduced by the measuring apparatus. These errors can be considered normally distributed random values with zero mean values and dispersions which are easy to evaluate if information on the accuracy of the used instruments is taken into account. In this case the δy_i values are known and function (5), being a solution of problem (3), (4), serves as a good approximation to the trend $\psi(x)$.

We have a different situation in an experiment under natural conditions. In this case the Δ_i values are governed by both instrumental inaccuracy and by the influence exerted on the y_i values by external factors, not taken into account, which can have a random character. In this case the δy_i values are unknown and the evaluation of the parameters of the distribution of the Δ_i random values in this case is an extremely complex problem having independent importance.

Now we will proceed to the problem of smoothing from points of view somewhat different than in the studies cited above. Regardless of the nature of the Δ_i values, henceforth we will call them errors. We will assume that the Δ_i errors are uncorrelated and have a zero mean value. This means that Δ_i form a purely random process, but the determined dependence of y on x is completely determined by the trend $\Psi(x)$. In this case, according to [3], the autocorrelation function $\rho(k)$ of the Δ_i process is equal to 0 for all whole positive k.

Assume that r(k) is a sample autocorrelation function determined by the expression

$$r(k)=\frac{\sum_{i=0}^{n-k}(\Delta_{i+k}-\overline{\Delta})(\Delta_i-\overline{\Delta})}{\sum_{i=0}^{n}(\Delta_i-\overline{\Delta})^2}, \tag{6}$$

where $\overline{\Delta}$ is the sample mean value of the sequence $\{\Delta_i\}$. In [3] it was demonstrated that when the number of terms in the series is sufficiently great it is admissible to assume that r(k) is distributed in conformity to a normal law with a zero mean value and a dispersion equal to n^{-1}. This means that in order for the sequence $\{\Delta_i\}$ to be considered a sample of a set of purely random numbers, r(k) must satisfy the expression

$$|r(k)|\leqslant\frac{u_q}{\sqrt{n}},$$

where q is the selected significance level and u_q can be obtained from a normal distribution law table. In most practical cases the function $|r(k)|$ is decreasing. Accordingly, hereafter we will limit ourselves to an examination of the r(1) value, being an evaluation of the first-order autocorrelation coefficient for the series $\{\Delta_i\}$, which we will denote by r_1. As the randomness test for the $\{\Delta_i\}$ sequence we use the expression

$$|r_1|\leqslant\frac{u_q}{\sqrt{n}}=r_{q.\,n}. \tag{7}$$

Assume that

$$\delta_i=y_i-g(x),\quad g(x)\in C^{(2)}[a,\,b]. \tag{8}$$

Assume that G is a set of such g(x) for which the sequence $\{\delta_i\}$ satisfies the randomness test (7), that is

$$|r_1|=\left|\frac{\sum_{i=0}^{n-1}(\delta_{i+1}-\overline{\delta})(\delta_i-\overline{\delta})}{\sum_{i=0}^{n}(\delta_i-\overline{\delta})^2}\right|<r_{q.\,n}, \tag{9}$$

where $\overline{\delta}$ is the sample mean value of the series $\{\delta_i\}$.

We will call the f(x) function best with a $C^{(2)}$ [a,b] approximation to the trend $\Psi(x)$ of the observed y_i series if f(x) minimizes the functional (3) amongst all $g(x)\in G$. Now the problem of smoothing of experimental data is formulated in the following way.

It is necessary to find the function f(x) with which the minimum of the functional (3) among all $g(x) \in C^{(2)}$ [a,b] is attained, for which the conditions (8), (9) are satisfied. In other words, among all the twice continuously differentiable functions for which the series of residues (8) can be considered a sample from a set of purely random numbers it is necessary to find the smoothest in the sense of the integral (3).

For solution of this problem we will use an algorithm for solution of problem (3), (4). In (4) we will assume that $\delta y_i = \delta y$, i = 0, 1,...,n and we will introduce the parameter $Q = (\delta y)^2 S$. The inequality (4) will be written in the form

$$\sum_{i=0}^{n} [g(x_i) - y_i]^2 \leq Q. \qquad (10)$$

In accordance with the "existence and uniqueness theorem" [12], each Q corresponds to a unique function f_Q, being a solution of problem (3), (10). We will use X_Q to denote a set of such values Q for which the functions f_Q satisfy condition (9). The set of functions f_Q corresponding to X_Q will be denoted F_Q.

According to the definition of the best approximation to the trend $\psi(x)$, given above, the sought-for function $f_0(x)$ minimizes the functional (3) in the set F_Q.

Since a unique $f_Q \in F_Q$ corresponds to each $Q \in X_Q$, we can determine the unambiguous function q(Q):

$$q(Q) = \int_a^b f_Q''(x)^2\, dx, \qquad (11)$$

being, evidently, a nonincreasing function.

It was demonstrated in [14] that there is such $Q' > 0$ that for all $Q > Q'$ the solution of problem (3), (10) is a straight line $l(x)$, constructed by the least squares method on the basis of observed values (x_i, y_i).

We will denote by r'_1 the evaluation of the autocorrelation coefficient, computed using formula (6) with k = 1, for the sequence

$$\Delta_i = y_i - l(x_i).$$

It is easy to show that r'_1 with a probability not less than (1 - q) satisfies the condition $r'_1 >= r_{q,n}$.

It therefore follows that either $r'_1 \leq r_{q,n}$ and then the straight line $l(x)$ is the sought-for approximation to the trend $\psi(x)$ or $r'_1 > r_{q,n}$. In the latter case X_Q is a limited set.

We denote $Q_0 = \sup \{X_Q\}$. It is evident that $Q_0 \leq Q'$. Since q(Q), determined by expression (11), is rigorously decreasing when $Q \leq Q_0$, then the function $f_0(x)$, imparting a minimum to the functional (3) in the set F_Q is a solution of the problem (3), (10) with $Q = Q_0$.

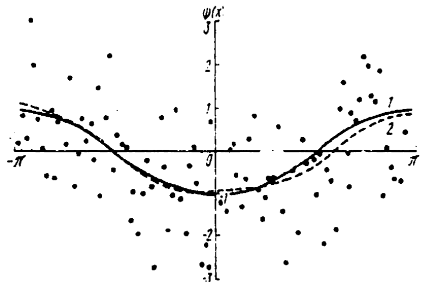

Fig. 1. Comparison of trend $\psi(x)$ (2), constructed from values $y_i = -\cos x_i + \Delta_i$, and curve of the function $y = -\cos x$ (1).

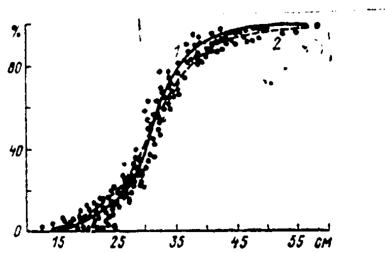

Fig. 2. Correlation between area of field S% and depth of snow cover $h > 30$ cm for different h values, cm. 1) curve constructed by author [9], 2) trend $\psi(x)$.

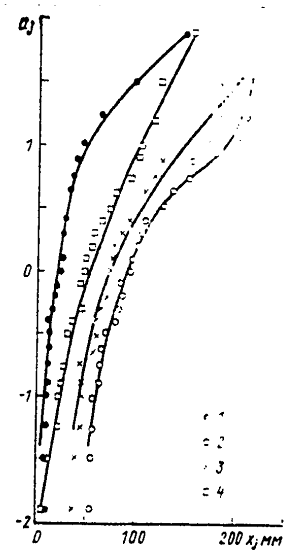

Fig. 3. Curves of correlation $u_j = u_j(x_j)$ between initial (x_j) and normalized (u_j) variables ($j = 0, 1, 2, 3$).

1) $u_0(x_0)$, (x_{0i}, u_{0i}); 2) $u_1(x_1)$; (x_{1i}, u_{1i}); 3) $u_2(x_2)$, (x_{2i}, u_{2i}); 4) $u_3(x_3)$, (x_{3i}, u_{3i}).

Thus, the solution of the formulated problem is either a straight line, constructed by the least squares method, or a cubic spline (5), being a solution of problem (3), (10) with $Q = Q_0$. In the latter case the solution of the problem is reduced to a determination of Q_0 and further application of the algorithm represented in [14]. On a practical basis it is necessary to find Q belonging to a quite small neighborhood of the Q_0 value.

We wrote the program in ALGOL-60 language, formalized in the form of a TREND procedure for realizing solution of this particular problem.

In order to check the quality of operation of the TREND procedure we carried out a great number of numerical experiments with different masses of values $y_i (i = 1,\ldots,n)$ which were stipulated by expression (1), where as $\Psi(x)$ we tested different smooth functions of the type x^2, $x^{1/2}$, sin x, cos x, e^x, and Δ_i were a realization of a random value having a zero mean value and distributed either uniformly in [-1, 1] or normally with a dispersion equal to 1. The Δ_i were generated by a computer, x_i were a uniform breakdown of the segment [a, b], the a and b values were different for different $\Psi(x)$. The experiments were carried out for n = 50, 100, 250. The $f_0(x)$ function, obtained as a result of use of the TREND procedure, in each case was very close to the true trend $\Psi(x)$.

Figure 1 shows the result of one of the experiments. The values $\Psi(x) = -\cos x$ in $[-\pi, \pi]$, n = 100, Δ_i had a normal distribution. The dots in the figure show the y_i values and the solid curve represents the approximation $f_0(x)$ to the trend $\Psi(x)$, obtained as a result of application of the TREND procedure.

We deliberately selected as an illustration of use of the algorithm a case when the characteristic change in the values of the trend $\Psi(x)$ in [a,b] is comparable to the mean square error in measuring the $\Psi(x)$ values. This case is very characteristic for hydrometeorological processes, where quite often it is necessary to seek the relationship between poorly correlated variables.

Figure 2 illustrates the application of the algorithm described above in agrometeorological practice. The example was taken from [9].

As the next example we will demonstrate the possibility of using approximation splines in the method of nonlinear multiple regression proposed by G. A. Alekseyev [1]. The method is based on the transformation of initial variables x_j into normalized variables u_j. In [1] this transformation is accomplished using tabulated values $\{x_{ji}, u_{ji}\}$ by means of the procedure of smoothing "by hand." For closing the Alekseyev algorithm it is necessary to carry out this procedure analytically. In our opinion, splines are an applicable tool for solving this problem primarily because the method for constructing them is completely "blind," that is, is based only on initial information, without relying on any a priori information on the characteristic form of the dependence $u_j(x_j)$. The latter circumstance is decisive for complete automation of the Alekseyev algorithm.

Figure 3 shows the curves for the correlation $u_j(x_j)$ for the example, published in [1], of determination of the correlation between runoff during December and precipitation during December, November and October for the

the White Hollow River basin. The constructed curves represent curves of approximation cubic splines.

As indicated at the beginning of the article, the residual deviations method is an extremely promising multidimensional statistical analysis method. At the present time we have developed an algorithm realizing this method in analytical form [7]. The central place in the algorithm is occupied by the smoothing procedure described in this article.

BIBLIOGRAPHY

1. Alekseyev, G. A., OB"YEKTIVNYYE METODY VYRAVNIVANIYA I NORMALIZATSII KORRELYATSIONNYKH SVYAZEY (Objective Methods for Smoothing and Normalizing Correlations), Leningrad, Gidrometeoizdat, 1971.

2. Anderson, G., STATISTICHESKIY ANALIZ VREMENNYKH RYADOV (Statistical Analysis of Time Series), Moscow, Mir, 1976.

3. Jenkins, H., Watts, D., SPEKTRAL'NYY ANALIZ I YEGO PRILOZHENIYA (Spectral Analysis and its Applications), Vol 1, Moscow, Mir, 1971.

4. Yezekiyel, M., Foks, K. A., METODY ANALIZA KORRELYATSII I REGRESSIY LINEYNYKH I NELINEYNYKH (Methods of Analysis of Correlations and Regressions, Direct and Indirect), Translated from English, Moscow, Statistika, 1966.

5. Konstantinov, A. R., Golitsina, Ye. F., "Analysis of Results of Laboratory Investigations of Runoff Losses into Thawed and Frozen Soil," TRUDY GGI (Transactions of the State Hydrological Institute), No 250, 1977.

6. Konstantinov, A. R., Serikova, V. V., "Mathematical Approximation of Graphic Regression Curves for the Correlation Between Runoff Losses and Yield and Determining Factors," TRUDY LGMI (Transactions of the Leningrad Hydrometeorological Institute), No 42, 1977.

7. Konstantinov, A. R., Khimin, N. M., "Use of the Residual Statistical Analysis Method for Investigating Hydrometeorological Processes," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 2, 1980.

8. Laykhtman, D. L., FIZIKA POGRANICHNOGO SLOYA ATMOSFERY (Physics of the Atmospheric Boundary Layer), Leningrad, Gidrometeoizdat, 1970.

9. Moiseychik, V. A., AGROMETEOROLOGICHESKIYE USLOVIYA I PEREZIMOVKA OZIMYKH KUL'TUR (Agrometeorological Conditions and the Wintering of Winter Crops), Leningrad, Gidrometeoizdat, 1975.

10. Panovskiy, G. A., Brayer, G. V., STATISTICHESKIYE METODY V METEOROLOGII (Statistical Methods in Meteorology), Translated from English, Leningrad, Gidrometeoizdat, 1967.

11. Solomon, S., "Statistical Correlations Between Hydrological Variables," STATISTICHESKIYE METODY V GIDROLOGII (Statistical Methods in Hydrology), Leningrad, Gidrometeoizdat, 1970.

12. Stechkin, S. B., Subbotin, Yu. N., SPLAYNY V VYCHISLITEL'NOY MATEMATIKE (Splines in Computational Mathematics), Moscow, Nauka, 1976.

13. Brandon, D. B., DEVELOPING MATHEMATICAL MODELS FOR COMPUTER CONTROL, GSA, Vol 7, 1954.

14. Reinsch, H., "Smoothing by Spline Functions," N. M., Vol 10, 1967.

UDC 556.16"32"

DETERMINING RUNOFF DURING WINTER AND TRANSITIONAL PERIODS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 68-77

[Article by V. S. Ryazanov, Upper Volga Territorial Administration of Hydrometeorology and Environmental Monitoring, submitted for publication 30 October 1979]

[Text]

Abstract: The possibility of using correlation-hydraulic models based on the Chezy-Manning formula for taking winter runoff into account in operational work is examined. In the example of individual hydraulic stations, situated on the rivers of the basin of the upper Volga and having a different character of flow pattern under winter conditions it is shown that the multiple regression equations, computed on the basis of measurements from preceding years, are an objective basis for routine (operational) determination of winter water runoff.

At the present time there are no objective methods for routine determination of river runoff in the absence of an unambiguous dependence between water discharges and levels, in particular when ice formations are present and ice has set in. In actual practice use is made of intuitive procedures for the extrapolation of water discharges up to the next measurement when it is possible to correct them. Such procedures lead to considerable errors in computing daily water discharges, considerably exceeding the admissible accuracy of ±10%. Recently hydraulic models have been proposed for the hydrometric determination of runoff [1]. Their checking and experimental introduction have been carried out at a number of hydraulic stations on rivers in the Volga basin. The article gives the results relating to models of hydrometric determination of winter runoff.

The appearance of ice formations on rivers leads to an increase in hydraulic resistances to the movement of flow. For this reason water discharge in the presence of ice formations and setting-in of the ice (Q_{win}) differs from the water discharge when there is a free state of the channel (Q_0) with one and the same river level (H). This difference in the Q_{win} and Q_0 values is evaluated using the conversion factor

$$K_{win} = Q_{win}/Q_0 = \omega_{win}\, v_{win}/\omega_0\, v_0, \tag{1}$$

where ω_0 and ω_{win} are the cross-sectional areas of the open channel and the river channel under the ice in m^2, v_0 and v_{win} are the mean current velocities of the water in the cross section of the open flow and under the ice in m/sec.

When using the Chezy and Manning formulas expression (1) for determining K_{win} is written in the form of the known formula

$$K_{зим} = m_1 \left(1 - \frac{\omega_л}{\omega_0}\right)^{5/3} \frac{n_0}{n_{зим}} \left(\frac{I_{зим}}{I_0}\right)^{0.5}, \tag{2}$$

[ЗИМ = win; Л = ice] where ω_{ice} is the area of the submerged ice in m^2; ω_0 and n_0 are the cross-sectional area in m^2 and the open channel roughness coefficient respectively; n_{win} is the generalized coefficient of roughness of the channel under the ice; I_0 and I_{win} are the hydraulic slopes of the open and ice-covered river channels; m_1 is a coefficient taking into account the relationship between the hydraulic radius and the geometrical characteristics of the flow cross section.

For the period of continuous ice cover, when the wetted perimeter (P_{win}) is equal to doubled the wetted perimeter of the channel when in an open state (P_0) and the hydraulic radius is $R_{win} = 0.5\ R_0$, the coefficient $m_1 = 0.63$. During the transitional periods of freezing and opening-up of the river, when P_{win} varies in dependence on the degree of coverage of the water surface with ice formations and is difficult to determine; the coefficient m_1 is also virtually impossible to determine.

Many authors, including P. N. Belokon', V. N. Goncharov and S. I. Kolupaylo, in investigations of the possibility of applying formulas of the type (2) for determining K_{win}, made the assumption of an equality between I_0 and I_{win} [3]. The same assumption was made by I. F. Karasev [1] in the multiple correlation equation which he proposed as a computation model for determining K_{win}

$$K_{зим} = m_1 - m_1 m_2 \frac{T}{T_л} e^{-5\frac{T}{T_л}} - \frac{5}{3} m_1 \frac{\omega_л}{\omega_0} + \frac{5}{3} m_1 m_2 \frac{\omega_л}{\omega_0} \frac{T}{T_л} e^{-5\frac{T}{T_л}} + \\ + \frac{5}{9} m_1 \frac{\omega_л^2}{\omega_0^2} - \frac{5}{9} m_1 m_2 \frac{\omega_л^2}{\omega_0^2} \frac{T}{T_л} e^{-5\frac{T}{T_л}}, \tag{3}$$

where T_{ice} is the duration of ice coverage of the river channel in days, varying from $T = 0$ to $T = T_{ice}$, m_2 is an empirical coefficient.

Equation (3) in general form describes the change in the conversion factor K_{win} during the entire period of the winter low water (from $K_{win} = 1.00$ at the beginning of the appearance of ice formations in autumn to $K_{win} = 1.00$ with the clearing of the river from ice at the beginning of the spring high water). The terms of this equation take into account the influence of the two principal factors on which the carrying capacity of the channel is dependent:

-- the changing roughness of the lower ice surface, represented in dependence on the time T from the time of setting-in of the ice,

-- degree of restriction of the river channel.

The terms on the right side of expression (3), except for m_1, can be regarded as variables entering into a regression equation in the form

$$y = a_0 + \sum_{i=1}^{i=N} a_i x_i, \tag{4}$$

where

$$x_1 = \frac{T}{T_{\text{л}}} e^{-5\frac{T}{T_{\text{л}}}}; \quad x_2 = \frac{\omega_{\text{л}}}{\omega_0}; \quad x_3 = \frac{\omega_{\text{л}}}{\omega_0} \frac{T}{T_{\text{л}}} e^{-5\frac{T}{T_{\text{л}}}};$$

$$x_4 = \frac{\omega_{\text{л}}^2}{\omega_0^2}; \quad x_5 = \frac{\omega_{\text{л}}^2}{\omega_0^2} \frac{T}{T_{\text{л}}} e^{-5\frac{T}{T_{\text{л}}}};$$

a_0, a_1,...,a_5 are the parameters of the regression equation.

The cross-sectional area of the open channel (ω_0), entering into equation (4), on any date is determined from the dependence $\omega_0 = f(H)$. The area of the submerged ice (ω_{ice}) is determined using data on the ice thickness (h_{ice}) in a hole:

$$\omega_{ice} = m_4 h_{ice} B_{ice}.$$

Here $B_{ice} \approx B_0$ is the river channel width at the hydraulic station in m; m_4 is a coefficient taking into account the noncoincidence of the mean thickness of the ice in the width of the channel and its thickness measured in the hole.

Since daily measurements of ice thickness are not made at hydrological posts, we also tested a model which takes into account the known relationship between the increase in ice thickness and the sum of negative air temperatures. According to investigations of different authors, for example, F. I. Bydin, $h_{ice}\sqrt{|\sum t|}$. According to observations made on the rivers of the upper Volga basin during the period from 1946 through 1965, Z. S. Surina [4] obtained the dependence

$$h_{ice} = 0{,}97\,(|\Sigma t|)^{0{,}59}.$$

Accordingly, the area of the submerged ice (ω_{ice}) in equation (3) can be completely represented by the expression

$$\omega_{ice} = m_5 B_{ice} \sqrt{|\sum t|}, \tag{5}$$

where m_5 is a proportionality factor.

After substituting expressions (5) in place of ω_{ice} the regression equation (4) as one of the determining factors will take the temperature factor into account.

Models based on the multiple regression equation (4) and using data on the ice thickness or on the sum of negative air temperatures are valid only for conditions of a stable ice cover without significant ice jam - water-under-snow phenomena during the period of winter low water. Precisely such a regime is characteristic for most of the rivers in the upper Volga basin. On the basis of the nature of river freezing in the considered basin, in accordance with the classification proposed by R. A. Nezhikhovskiy, types I and II can be distinguished. A fixed ice cover on them is established for the most part by a gradual expansion and closing-in of ice forming along the shore (type I) or by simultaneous formation of "bridges" in a number of places where the ice-transporting capacity of the flow is reduced, with the subsequent filling of the spaces between the ice bridges with floating floes (type II).

The entire difficulty in operational determination of runoff is that the daily water discharges are computed under conditions when there is still no complete set of measurements of discharges for a particular year. In this case as a base for the computations it is possible to use regression equations of type (4), derived using the results of preceding measurements either during periods with winters close in severity (WCS) or during the long-term period as a whole (LTP). Such a type of regression equation with the use of data on the ice thickness (WCS-I, LTP-I) and on the sum of negative air temperatures (WCS-II, LTP-II) was computed for the period 1959 through 1975 for the hydraulic stations Oka River (Gorbatov), Oka River (Murom), Oka River (Belev), Vetluga River (Vetluzhskiy) and Mera River (Malo-Berezovo), having basin areas from 224,000 to 820 km^2.

The parameters of the multiple regression equations were determined at the Computation Center Verkhne-Volzhskoye Administration of the Hydrometeorological Service by the least squares method using a "Minsk-32" electronic computer with a standard program developed at the State Hydrological Institute. The mean air temperature during the time T_{ice} was adopted as the index of winter severity. Winters for which $|\sum t|/T_{ice} < 7.0$ were classified as mild, with $|\sum t|/T_{ice} = 7.0\text{-}9.0$ -- as normal, with $|\sum t|/T_{ice} > 9.0$ -- as severe.

The free term (a_0) in the derived regression equations was always close to 1.00, which is natural, because with $T = 0$ $\omega_{ice} = 0$ and $K_{win} = 1.00$. The remaining parameters of the regression equations ($a_1, \ldots, a_0$) vary in a rather broad range and reflect the physical conditions of processes of change in K_{win}.

The derived regression equations in most cases have rather high multiple correlation coefficients ($R_0 = 0.75\text{-}0.95$). An exception is the regression equation for the station on the Mera River, Malo-Berezovo, R_0 for which is only 0.56-0.73. The multiple correlation coefficients as a rule are higher for models using data on ice thickness in comparison with models taking into account the temperature factor and this is logical. The highest R_0 correspond to the regression equations for severe winters when a stable ice cover is formed on rivers rapidly and is not impaired by thaws to the very beginning of the spring high water. The minimum mean square errors (Δ_Q), which directly characterize the accuracy of the values computed from the regression equations, correspond to these same winters. The R_0 values are usually lower for gentle unstable winters when the formation of the ice cover lasts a long time and is disrupted by thaws leading to the temporary clearing of ice from the river.

In actual hydrological computations it is necessary to compute the daily water discharges during years when the measurements of discharges during the winter low-water period are not made at all. The need for this kind of computations arises, for example, in cases of restoration of winter runoff on the basis of measurements of preceding or subsequent years. In such cases it is usually recommended that winter runoff be computed from the K_{win} values for the closest years or the means for a number of years. However, such averaged $K_{win}(T)$ curves usually do not make it possible to compute the daily water discharges with sufficient reliability. For this reason the results of such computations of winter runoff during past years in most cases were deemed unreliable when preparing the handbook SURFACE WATER RESOURCES IN THE USSR.

Comparison of the K_{win} values on the basis of measurements and computations using WCS models, on the assumption that measurements were not made, was carried out for the posts Oka River (Gorbatov) and Oka River (Murom) in the years 1958-1959 and 1975-1976, that is, during winters not entering into the period adopted for determining the regression equations. Taking into account that the mean air temperature for the winter of 1958/1959 for Gorbatov was -4.4°C and for Murom was -5.2°C, the K_{win} values were computed using the following type of regression equations for gentle winters:

1) Oka River -- Gorbatov

$$\text{WCS-I } K_{win} = 0{,}89-1{,}63x_1-3{,}38x_2-10{,}2x_3+8{,}37x_4+37{,}7x_5,$$
$$\text{WCS-II } K_{win} = 1{,}01+1{,}69x_1-0{,}03x_2-2{,}05x_3-0{,}0001x_4+0{,}14x_5;$$

2) Oka River -- Murom

$$\text{WCS-I } K_{win} = 0{,}99-4{,}62x_1-2{,}72x_2-14{,}7x_3+1{,}94x_4+205x_5,$$
$$\text{WCS-II } K_{win} = 1{,}08-3{,}01x_1-0{,}05x_2-1{,}19x_3+0{,}0003x_4+0{,}13x_5.$$

For the winter of 1975/1976 the mean air temperature at Gorbatov was -7.7°C and at Murom was -8.6°C. Proceeding on the basis of this value, K_{win} was computed using regression equations for normal winters in the form:

1) Oka River -- Gorbatov

WCS-I $K_{win} = 0,94-7,99x_1-6,14x_2+117x_3+20,4x_4-343x_5,$
WCS-II $K_{win} = 1,00-7,45x_1-0,04x_2+0,72x_3+0,001x_4-0,05x_5;$

2) Oka River -- Murom

WCS-I $K_{win} = 0,95-2,13x_1-2,57x_2-35,8x_3+1,34x_4+266x_5,$
WCS-II $K_{win} = 0,99+0,74x_1+0,0002x_2-1,64x_3-0,002x_4+0,10x_5.$

Table 1

Results of Computations of Water Discharges (WCS Model) for Cases of Absence of Measurements

Период 1	Количество измерений 2	3 Отклонения вычисленных значений Q от измеренных, %		
		$\sqrt{\Delta Q^2}$	$\Delta Q_{предельное}$ 4	дата 5
6 *Модель УРП, учитывающая данные о толщине льда*				
8 р. Ока — г. Горбатов				
10 Зима 1958/59	11	8,6	23,4	1 XII 1958
Зима 1975/76	14	10,4	22,3	17 XII 1975
9 р. Ока — г. Муром				
Зима 1958/59	10	15,1	—28,7	11 XII 1958
Зима 1975/76	15	9,4	24,7	19 XI 1975
7 *Модель УРП, учитывающая температурный фактор*				
8 р. Ока — г. Горбатов				
Зима 1958/59	11	13,6	45,6	1 XII 1958
Зима 1975/76	14	9,0	20,6	17 XII 1975
9 р. Ока — г. Муром				
Зима 1958/59	10	14,0	—33,4	22 XII 1958
Зима 1975/76	15	12,2	47,3	19 XI 1975

KEY:
1. Period
2. Number of measurements
3. Deviation of computed Q values from measured values, %
4. Limit
5. Date
6. WCS model taking into account data on ice thickness
7. WCS model taking into account temperature factor
8. Oka River -- Gorbatov
9. Oka River -- Murom
10. Winter...

The results of comparison of the computed and measured K_{win} values are given in Table 1.

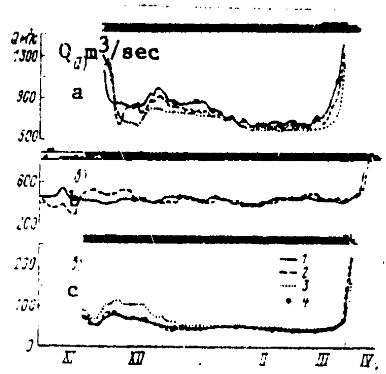

Fig. 1. Matching of runoff hydrographs. a) Oka River -- Gorbatov. Winter 1977/1978; b) Oka River -- Gorbatov. Winter 1975/1976; c) Vetluga River -- Vetluzhskiy post. Winter 1974/1975. 1) According to hydrological yearbook; 2) According to WCS model; 3) According to routine data; 4) Measured water discharges.

The cited data make it possible to draw the preliminary conclusion that models of the WCS type are adequately effective, especially models taking into account data on the thickness of the ice, for the purposes of reconstructing the winter daily water discharges during years when no measurements are made. However, very high deviations (ΔQ) when using models of the WCS type in individual cases are considerable and usually correspond to the characteristic points at the beginning of winter when there is a minimum K_{win} value determining the general form of the $K_{win}(T)$ function.

Computed models of the WCS type, when reliable long-term weather forecasts (of the nature of winter) are available, could also serve as a basis for routine determination of water runoff. But, unfortunately, hydrologists for the time being do not have such forecasts. For this reason for operational purposes at the present time it is possible to use only models based on the WCS, obtained on the basis of long-term data for the preceding years of observations.

Numerical experiments with the use of such models for the purposes of routine computation of daily water discharges were carried out for the posts Oka River -- Gorbatov and Vetluga River -- Vetluzhskiy post. Gorbatov post on the Oka River was of the greatest interest from the point of view of the possibilities of use of correlation-hydraulic models for operational determination of runoff: it is one of the two principal stations for

Table 2

Comparative Results of Operational Computation of Daily Water Discharges Using LTP Model With Allowance for Temperature Factor

1 Период	2 Характеристика стока	3 Отклонения расходов от опубликованных в гидрологическом ежегоднике, %					
		4 по модели УРМ			5 по оперативным данным		
		$\Delta \bar{Q}$	$\sqrt{\Delta \bar{Q}^2}$	$\Delta Q_{пред.}$ 6	$\Delta \bar{Q}$	$\sqrt{\Delta \bar{Q}^2}$	$\Delta Q_{пред.}$
		8 р. Ока — г. Горбатов					
7 Зима 1975/76	10 суточный	8,1	10,5	−31,0	—	—	—
	11 декадный	7,7	9,8	−25,0	—	—	—
	12 месячный	3,6	4,7	−6,4	—	—	—
Зима 1977/78	суточный	6,0	7,8	27,1	11,9	16,1	34,5
	декадный	4,5	5,7	−8,0	11,2	14,3	−21,8
	месячный	3,8	4,8	−5,0	9,6	11,9	−15,1
		9 р. Ветлуга — р. п. Ветлужский					
Зима 1973/74	суточный	8,6	11,8	34,5	19,9	21,8	44,4
	декадный	8,4	10,5	21,6	18,4	21,0	42,0
	месячный	6,1	7,6	11,8	20,4	23,0	31,2
Зима 1974/75	суточный	4,7	9,9	−26.6	19,4	23,7	51,5
	декадный	2,8	3,9	−19.3	18,4	22,5	44,7
	месячный	2,6	3,1	5,3	16,8	20,5	40,2
Зима 1975/76	суточный	6,2	7,1	−35,6	8,8	11,9	−44,7
	декадный	4,2	6,4	−15,2	8,7	11,5	−31,7
	месячный	3,3	3,8	4,5	8,5	10,9	20,4

KEY:

1. Period
2. Runoff characteristic
3. Deviation of discharges from data published in hydrological yearbook, %
4. Using LTP model
5. Using operational data
6. ΔQ_{lim}
7. Winter
8. Oka River -- Gorbatov
9. Vetluga River -- Vetluzhskiy post
10. 24 hours
11. 10-day period
12. month

determining the annual discharges of Volga water at Gor'kiy below the confluence of the Oka. Upstream is the post at the Gor'kovskaya Hydroelectric Power Station, at which routine determination of runoff is accomplished quite reliably. In the experiments we used the following regression equations for a long-term period in which the degree of channel restriction was taken into account through the sum of negative air temperatures (LTP-II):

1) Oka River -- Gorbatov

$K_{win} = 1{,}01-5{,}38\,x_1-0{,}04\,x_2+0{,}23\,x_3+0{,}001\,x_4-0{,}02\,x_5;$

2) Vetluga River -- Vetluzhskiy post

$K_{win} = 1{,}01-4{,}78x_1-0{,}03x_2-0{,}10x_3+0{,}000003x_4+0{,}02x_5.$

The values of the mean daily air temperatures entering into these equations were used for the post Gorbatov on the Oka River in accordance with observations at Gor'kiy (Stri..no) meteorological station and for the post Vetluzhskiy on the Vetluga River on the basis of observations at Krasnyye Baki meteorological station.

Taking into account that in the routine computation of the daily water discharges the actual duration of ice phenomena and presence of ice (T_{ice}) is an unknown value, as the computed T_{ice} values it is desirable to use the mean for the long-term period of preceding observations. Such an assumption does not exert a significant influence on the final results of computations.

In order to tie in the LTP parameters employed in routine computation of annual water discharges to the peculiarities of the winter for the computation year we carried out correction of the free term of the equation (Q_0) on the basis of factual K_{win} values for the date of each measurement of water discharge (4) written relative to (a_0),

[ЗИМ = win] $$a_{0\,i} = (K_{зим})_i - a_1\,(x_1)_i - a_2\,(x_2)_i - a_3\,(x_3)_i - a_4\,(x_4)_i - a_5\,(x_5)_i. \quad (6)$$

A comparison of the data from the hydrological yearbook and the results of determination of the annual water discharges by existing methods and on the basis of the LTP model is given in Table 2 and in the figure.

The deviations (ΔQ) show that the values of the daily water discharges, computed using the LTP model, for the most part coincide with those published in the hydrological yearbooks and are considerably more exact than the daily water discharges computed using schematic procedures for the extrapolation of winter water discharges used for routine purposes at the present time. However, during the initial period of river freezing, and in individual cases also during the opening-up period, the deviations of the daily water discharges, computed using the LTP model, on individual days attain 30% or more. Particularly significant errors are observed in the initial period of river freezing when a considerable time elapses from the date K_{win} = 1.00 to the date of the first measurement of water discharge. An example is the winter of 1975/1976 for Gorbatov post, when the possibility of correction of the free term (a_0) appeared only after 1 1/2 months from the onset of setting-in of the ice on the Oka River

(Fig. 1b). This once again confirms the correctness of the requirement for the need of measuring water discharges during the period of river freezing as frequently as is practical, taking into account safety conditions in performing the work.

Table 3

Parameters of Regression Equation Based on Logarithmic Model. Mera River -- Malo-Berezovo Station, F = 820 km^2

Вид уравнения 1	Количество измерений 2	a_0	a_1	a_2	a_3	R_0
3 УРП для мягких зим	68	−0.128	0,537	0,841	0,006	0.81
4 УРП для нормальных зим	103	2.372	−0.895	2.410	0,007	0.89
5 УРП для суровых зим	81	0.734	0.126	1.732	0.007	0.90
6 УРМ за период 1960—75 гг.	252	0.841	0,064	1.661	0.006	0.87

KEY:

1. Type of equation
2. Number of measurements
3. MCS for gentle winters
4. MCS for normal winters
5. MCS for severe winters
6. LTP model for period 1960-1975

In cases when additional resistances to flow, caused by ice formations and an ice cover, are aggravated by any other additional factors, the use of models of the type (4) loses physical sense and leads to considerable errors in computation of daily water discharges. One of these additional factors, distorting the general form of the function $K_{win}(T)$, is water vegetation, which, dying out in the autumn, in many small rivers in the Volga River basin is not carried away by the current, but continues to exert an additional influence on the relationship between water discharge and level. An example of this type of river is the Mera River at Malo-Berezovo station, adopted as one of the analogue rivers in the scheme for operational computation and prediction of the daily lateral inflow into Gor'kovskoye Reservoir.

The daily water discharges during the period of winter low water on the Mera River at Malo-Berezovo, like on most such rivers, are computed using the coefficient $K_{win,\ pre}(T)$. Therefore, the WCS and LTP data computed for a particular post using a model of the form (4) have low multiple correlation coefficients (R_0 = 0.56-0.73). The use of these models for operational computation of water discharges leads to inadmissibly high errors (on the average 25-45% with values of the limiting errors exceeding 100%).

In such cases when the K_{win} conversion factor loses physical sense, an attempt was made to apply a model for determining water runoff which makes direct use of the Chezy-Manning formula

$$Q = \frac{\omega_{зим} R_{зим}^{2/3}}{n_{зим}} \sqrt{I_{зим}}, \quad (7)$$

[ЗИМ = win; Л = ice]

transformed by I. F. Karasev [1] into the expression

$$Q = \frac{1.41\, \omega_{зим} R_{зим}^{2/3}}{n_0 \sqrt{1 + e^{1.7\, h_л}}} \sqrt{I_{зим}}. \quad (8)$$

The reduction of expression (8) to logarithmic form makes it possible to obtain a multiple correlation equation in the form (4) in which

[ЗИМ = win; Л = ice]

$$y = \lg Q; \quad x_1 = \lg \omega_{зим}; \quad x_2 = \lg R_{зим};$$
$$x_3 = \lg (n_0) \sqrt{1 + e^{1.7\, h_л}}.$$

Table 4

Comparative Table of Deviations of Computed Discharges From Those Published in Hydrological Yearbook for Mera River - Malo-Berezovo Station

Период 1	Характеристика стока 2	По логарифмической модели УРМ 3: $\Delta\overline{Q}$	По логарифмической модели УРМ 3: $\sqrt{\Delta\overline{Q}^2}$	По оперативным данным 4: $\Delta\overline{Q}$	По оперативным данным 4: $\sqrt{\Delta\overline{Q}^2}$
5 Зима 1975/76	6 суточный	7,3	13,2	8,0	10,8
	7 декадный	6,1	8,9	6,0	6,6
	8 месячный	3,3	5,0	3,9	5,2
Зима 1976/77	суточный	9,3	12,1	12,2	15,8
	декадный	5,2	6,8	9,8	12,3
	месячный	3,7	4,7	8,4	10,7

KEY:

1. Period
2. Runoff characteristic
3. According to LPT logarithmic model
4. According to operational data
5. Winter
6. 24 hours
7. Ten-day
8. Month

The a_0, a_1, a_2, a_3 values statistically generalize the parameters of equation (7), including the constant slope I_{win}. The factor x_3 for simplification of the computations is adopted in the form of the sum of the moduli of negative air temperatures ($|\sum t|$), which integrally take into account the degree of channel restriction and roughness.

The equations derived for the Malo-Berezovo post on the Mera River (WSC and LTP) with the use of a logarithmic model are given in Table 3.

ME , _OGY

6 OCTOBER 1980 NO. 7, JULY 1980 2 OF 2

Checking of the possibility of use of a multiple regression equation (LTP) for the purpose of operational computation of the daily water discharge was carried out during the winter of 1975/1976 and 1976/1977. Data on air temperature were taken from observations at Ostrovskoye meteorological station, the closest to Malo-Berezovo station.

The correction of the free term of equation (a_0) for the purpose of tie-in of the LTP parameters to the peculiarities of winter of a specific year was carried out applicable to equation (8) using the expression

$$a_{0\,i} = (\lg Q)_i - a_1 (x_1)_i - a_2 (x_2)_i - a_3 (x_3)_i. \tag{9}$$

The comparative results of the computations are given in Table 4. The discrepancy of the discharges ΔQ characterizes only the degree of coincidence of the data in the hydrological yearbook and the results of computations on the basis of models and does not at all serve as an evaluation of the accuracy of the latter. The deviations ΔQ are related primarily to the inadequate soundness of the procedures for computing the daily water discharges contained in the yearbooks. According to evaluations made by the Hydrometry Section at the State Hydrological Institute, correlation-hydraulic models ensure a decrease in the errors in operational determination of runoff by a factor of 1.2-1.5 in comparison with the methods used.

The results of numerical experiments presented in the article show that equations of the LTP type, taking the temperature factor into account, are an entirely objective basis for operational determination of river runoff during winter on rivers where an increase in hydraulic resistances to the movement of the flow is caused only by ice formations and the ice cover. In the case of rivers on which additional resistances, caused by ice formations or an ice cover, are complicated by the influence of water vegetation remaining in the channel, it is desirable for operational purposes to use a logarithmic model for taking winter runoff into account.

BIBLIOGRAPHY

1. Karasev, I. F., "Mathematical Models for Hydrometric Determination of River Runoff," TRUDY GGI (Transactions of the State Hydrological Institute), No 256, 1978.

2. Nezhikhovskiy, R. A., "Types of River Freezing and Types of Ice Jams," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 2, 1974.

3. Polyakov, B. V., GIDROLOGICHESKIYE ISSLEDOVANIYA NIZHNEY VOLGI (Hydrological Investigations of the Lower Volga), Moscow, Gosstroyizdat, 1938.

4. Surina, Z. S., "Investigation of Growth and Computation of Ice Thickness on Rivers of the Upper Volga Basin," SB. RABOT GOR'KOVSKOY, VOLZHSKOY I RYBINSKOY GMO (Collection of Papers of the Gor'kiy, Volga and Rybinsk Hydrometeorological Observatories), Leningrad, Gidrometeoizdat, 1973.

UDC 556.536

METHOD FOR MAKING OBSERVATIONS OF THE WATER SURFACE SLOPE OF RIVER FLOWS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 78-83

[Article by Candidate of Technical Sciences V. V. Kovalenko, Leningrad Hydrometeorological Institute, submitted for publication 21 January 1981]

[Text]

Abstract: The author defines a quantitative criterion which makes it possible to select the location of a hydrometric station and the length of the base for making observations of the water surface slope of river flows. The restrictions on base length which follow from the hydraulic conditions of the water flow current are pointed out.

The slope of the free surface of a river flow is one of the most important hydraulic characteristics [5, 6, 16]. However, the recommendations on the organization of slope observations given in [13, 14] sometimes have a purely qualitative character, which sometimes leads to subjectivity in siting a hydrometric station and the choice of the slope base length.

In this article an attempt is made to define objective quantitative criteria making it possible to regularize slope observations.

The equations of hydromechanics relate the field of velocities and the pressure field. Within the framework of one-dimensional hydraulic idealization this has the following result: in principle, on the basis of the measured pressure differential (or piezometric slope) it is possible to judge the discharge passing in the section. In the well-developed countries most of the industrial flow meters (in Great Britain, for example, 90%) are based precisely on measurement of the pressure differential (slope).

The principal norm-setting documents in the field of hydrometry of slopes [13, 14] define the purpose of observations of the longitudinal slope as follows:

1) for estimating the capacity of the channel, determined by the water discharge;

2) for ascertaining the value of the Chezy coefficient C, taking into

account the hydraulic resistances.

Taking into account that the water movement in the river flow has a nonuniform and nonstationary character, it is natural to generalize the Chezy formula and obtain a general mathematical model which can be used for the above purposes and more graphically reflect the role played by the slope in the solution, possibly, of other hydrometric problems. In order to obtain such a "hydrometric model" it is natural to use hydraulic idealization equations, for example, in the following approximation [8]:

$$i - \frac{\partial h}{\partial_* x} = j + \frac{1}{g_*}\frac{\partial U}{\partial t} + \frac{\alpha U}{g_*}\frac{\partial U}{\partial x} - \frac{\alpha - 1}{g_*}\frac{U}{F}\frac{\partial F}{\partial t} +$$

$$+ \frac{1}{g_* F}\left\{\frac{\partial^2}{\partial t\,\partial x}\left(\beta_1' FRU\frac{\partial h}{\partial x} + \beta_2 FR\frac{\partial h}{\partial t}\right) + \frac{\partial^2}{\partial x^2}\left(\beta_3 FRU^2\frac{\partial h}{\partial x} + \right.\right. \quad (1)$$

$$\left.\left. + \beta_4 FRU\frac{\partial h}{\partial t}\right)\right\}, \quad (2)$$

$$\frac{\partial UF}{\partial x} + \frac{\partial F}{\partial t} = 0,$$

where j is a dissipative term; α, β'_1, β_2, β_3, β_4 are coefficients dependent on the velocity distribution in the cross section; the remaining notations are those in general use in hydrometry.

For a case of practical interest, when information is known on the level $H = f(x, t)$ and the morphometry of a fixed hydrometric location with the coordinate x_0, after simple transformations of equations (1) and (2) it is possible to derive an equation in full derivatives [10]

$$\frac{dU}{dt} + f_1(x_0, t)\,U^2 + f_2(x_0, t)\,U + f_1'(x_0, t)\,j = f_3(x_0, t). \quad (3)$$

In (3) the functions f_1, f'_1, f_2, f_3, with a known morphometry of the hydrometric location, are dependent only on the water level and its derivatives in coordinates and time.

If it follows from the Chezy formula for uniform movement that $Q = f(H)$, then from (3) for smoothly changing movement

$$Q = f\left(H, \frac{\partial H}{\partial x}, \frac{\partial H}{\partial t}\right),$$

whereas for a general case $Q = f\left(H, \frac{\partial H}{\partial x}, \frac{\partial H}{\partial t}, \frac{\partial^2 H}{\partial x^2}, \frac{\partial^2 H}{\partial t^2}, \ldots\right)$,

that is, in the most general case for determining the discharge it is necessary to have information not only on the level and slope, but also on the curvature and derivatives of curvature of the free surface.

Information on the water surface slope is necessary for the solution of at least four fundamental problems in hydrometry which in general form can be solved using equation (3), to wit:
a) indirect measurement (determination) of water discharge,
b) extrapolation of discharge,
c) restoration of the hydrograph;
d) determination of hydraulic resistances (Chezy coefficient).

An investigation of the sensitivity of equation (3), that is, for discharge or current velocity, to changes in the parameters entering into it [9], revealed that the free surface slope is the most important.

We note that in the determination of hydraulic resistances from equation (3) (with measured slope and current velocity) the measurement of convective acceleration in actuality is not accomplished from the difference in velocities in two sections (which in the case of nonstationary movement leads to considerable errors), but from the slope and derivative of the level in time (more detail concerning this is given in [2]). In addition to the piezometric level, which has been discussed until now, a certain role is played by the inertial and hydraulic (dissipative term in equation (3)) slopes. For example, in determining the additional slope caused by the difference in dissipation in a uniform flow from that in the case of nonstationary movement, the piezometric slope is used again [3].

The following points can be noted in connection with the method for making slope observations [13, 14]:
a) choice of site for a hydrometric post for slope observations;
b) choice of length of segment (base) for measuring slope.

The instructions give quite clear and specific recommendations only on the choice of base length, with accuracy characteristics taken into account; otherwise the instructions are limited to qualitative indications, for example, on the unidirectional change in the areas of the cross sections in the sector. In order to obtain quantitative criteria for the purpose of designating the site of the hydrometric post and base length it is natural to select a criterion which at the most "advantageous" hydrometric post would assume an extremal value. In obtaining such a criterion it is necessary to take into account the purpose of the slope observations, that is, on the one hand, at the optimum hydrometric post the measured slope must react well to a change in discharge and hydraulic resistances, and on the other hand, with the further use of the slope in computations the discharges and resistances must be adequately sensitive to slope. As such a criterion we can use the extremum (in the longitudinal coordinate) of the matrix norm of the corresponding sensitivity functions S_i in some functional space whose metrics is determined by the purpose of the observations (stationary long-term observations, high-water observations, observations of minimum runoff, etc.).

The sensitivity functions are derived from equations (1) and (2) by known methods. In particular, the equation for the function of sensitivity of discharge to slope will be

$$\dot{S}+\left[\frac{-F^2C^2R\,\frac{\partial Q}{\partial t}-I\alpha BQ^2C^2R+gQ^2F-\alpha Q^3C^2R\left(\frac{\partial F}{\partial x}\right)_*}{(\alpha+1)\,Q^2FC^2R}+\right.$$
$$\left.+\frac{I\alpha BQ^2C^2R+IgF^3C^2R}{(\alpha+1)\,Q^2FC^2R}\right]S=\frac{gF^3QC^2R-\alpha BQ^3C^2R}{(\alpha+1)\,C^2RQ^2F}. \tag{4}$$

The solution of equation (4) with zero "initial" conditions and "freezing-in" [11] of the coefficients and free term, corresponding to their values at the hydrometric post, will be

$$S=(gF^3QC^2R-\alpha BQ^3C^2R)\left(-F^2C^2R\,\frac{\partial Q}{\partial t}-I\alpha BQ^2C^2R+gQ^2F-\right.$$
$$\left.-\alpha Q^2C^2R\left(\frac{\partial F}{\partial x}\right)_*+I\alpha BQ^2C^2R+IgF^3C^2R\right)^{-1}\left\{1-\exp\left[-((\alpha+\right.\right.$$
$$+1)\,C^2RQ^2F)^{-1}\left(-F^2C^2R\frac{\partial Q}{\partial t}-I\alpha BQ^2C^2R+gQ^2F-\right.$$
$$\left.\left.\left.-\alpha Q^2C^2R\left(\frac{\partial F}{\partial x}\right)_*+I\alpha BQ^2C^2R+IgF^3C^2R\right)\cdot x\right]\right\}. \tag{5}$$

In a general case the components of the matrix of sensitivity functions are dependent on the coordinate of the hydrometric post, the extent of the channel reach (slope base), time and specific realization of the hydrological year or observation interval. The location of the optimum hydrometric post and the slope base length are not independent of one another, have a stochastic nature and are determined by the extremum of the norm of the matrix of sensitivity functions. After designating the location of the hydrometric post it is necessary to select the length of the base on which the slope will be determined. If the equations of hydraulics are used as a point of departure, we have $\frac{\partial H}{\partial x}=\lim_{\Delta x\to 0}\frac{\Delta H}{\Delta x}$.

From the point of view of hydraulics, it would seem that the best base is a zero base (we note that there are methods for measuring a slope at a point [1, 4]). In actuality, however, this is not the case.

The longitudinal coordinate x explicitly enters the sensitivity function S. Theoretically S attains its maximum value when $x = \infty$. On a practical basis, as has been done in an investigation of dynamic systems [9], it is reasonable to limit ourselves to those $x = L$ values for which the exponential term in S attains the level 0.63. Using (5), taking into account the principal hydraulic characteristics and channel morphometry, we derive a formula for base length

$$L = \frac{h^{4/3} F}{gn^2 F - 0{,}5\, h^{4/3} \left(\frac{\partial F}{\partial x}\right)_*}. \qquad (6)$$

The restrictions on base length, following from formula (6), are related to the inertial effect in water flow movement. However, there is still another restriction associated with the correctness of the one-dimensional hydraulic idealization itself. The H level, figuring in the equations of hydraulics, naturally differs from that level z which in actuality is observed in a real flow and with which the relief of the free surface is associated. The difference between the averaged (hydraulic) and actual levels also dictates the choice of the slope measurement base from the condition that the drop of the averaged slope be at least an order of magnitude greater than the level drop as a result of its random variations, that is

$$\int_L \frac{\partial \|f\|}{\partial x}\, dx \ll \int_L \frac{\partial H}{\partial x}\, dx, \qquad (7)$$

where f is the norm of level variations not taken into account by the one-dimensional idealization. At the present time only estimates of the systematic level deviations in width from a horizontal position as a result of rounding-off of the flow, Coriolis force of the earth's rotation and relief of the channel cross section are real. At the present time nothing is known about the dependence of this same systematic deviation on the phases for the hydrological year, and also about the distribution of water volume in the long-term section. It is reasonable to adopt the following hypothesis:

$$\|f\| = \max_{y \in [-B/2,\, B/2]} |H - z|.$$

The slanting of the free surface due to the enumerated factors can be determined using the known formulas [7].

Using the mean value theorem, from (7) we obtain an estimate of the slope measurement base length

$$\frac{10\,\|f\|}{\bar{I}} \lesssim L, \qquad (8)$$

where the bottom slope can be taken approximately as $\bar{I}$.

The instructions [13, 14] recommend a formula for designating the base length, using as a point of departure the accuracy in measuring water level and the leveling of posts, that is, the determination of L is related only to the possibilities for technical outfitting of the hydrological network. However, the restrictions corresponding to formulas (6) and (8) have a fundamental character and cannot be eliminated by any technical means. Taking into account that when making slope observations in the network use is made of a standard program and technical means, the errors in leveling and level measurement can be considered identical. Then the base length, according to [13, 14], will be determined only by the flow slope.

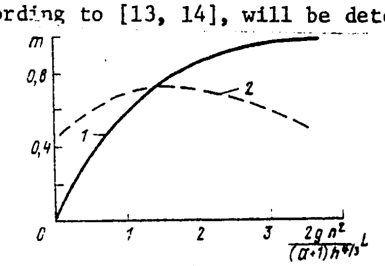

Fig. 1. Results of analysis of field data. 1) theoretical curve, 2) data from yearbooks.

On the other hand, the bilinear symmetric functional corresponding to the scalar product of the sensitivity vector in measurements and the sensitivity vector in computations and relating the measured and computed parameters, is dependent not only on the slope, but also (in the simplest case of a quasisteady regime) on channel roughness, depth of filling and the value $(\partial F/\partial x)_*$.

However, the instructions do not fix the base length in dependence on channel filling and roughness, but only in inverse dependence on slope, which should lead to a decrease in the correlation of the measured slopes and discharges with an increase in roughness and a decrease in filling. In order to check this assumption we made an analysis of 1,500 measured discharges (on the basis of data from the yearbooks) at 78 hydrometric posts located on different rivers in the USSR. The results of computations are shown in Fig. 1 (significance level 5%). Taking into account the measurement errors and the fact that data from standard observations cited in yearbooks cannot be used in calculating the above-mentioned functional at each point in the observation interval, with allowance for sensitivity to the Chezy coefficient, along the y-ordinate we plotted the product of the discharge-to-slope and slope-to-discharge regression coefficients m. An evaluation of statistical homogeneity was made using the Fisher test [15]. The figure shows that the representativeness of the observations is less than is theoretically possible. The dropoff of the theoretical curve from their point of intersection is attributable to the fact that with a zero base a zero sensitivity is adopted.

Summary

1. The official norm-setting documents dealing with slope observations have a qualitative character, which can reduce the representativeness of measurements.

2. The site for a hydrometric post and slope base length are not mutually independent.

3. An objective method for designating the hydrometric post coordinate involves finding the extremum of the norm of the matrix of sensitivity functions within the limits of the channel reach in which the carrying out of hydrometric observations is planned.

4. The slope measured at a point, that is, on a zero base, may or may not characterize the discharge acually passing through the hydrometric post. D Due to the explicit dependence of the sensitivity function on the longitudinal coordinate there is a finite length of the base in which the measured slope will determine discharge. This base is dependent for the most part on channel filling, roughness and morphometry of the channel reach.

5. The base length should be such that the drop of the averaged (hydraulic) level not only substantially exceeds the error in determining slope, but also is an order of magnitude greater than the level drop as a result of its random variations (wind waves, etc.) or the systematic deviations associated with incorrectness of the one-dimensional hydraulic idealization (water surface slanting in the case of channel curvature or in the case of a nonuniform distribution of discharge in the cross section).

6. At the 5% significance level there are systematic deviations of field data from the theoretical data, which indicates a nonrepresentativeness of the measured slopes at a number of posts in the network and the desirability of taking into account the results cited above when organizing slope observations.

BIBLIOGRAPHY

1. Arbuzov, I. A., Kovalenko, V. V., "Device for Measuring the Longitudinal Slope of the Free Surface of a Watercourse," Author's Certificate USSR 535456, OTKRYTIYA, IZOBRETENIYA, PROMYSHLENNYYE OBRAZTSY, TOVARNYYE ZNAKI (Discoveries, Inventions, Industrial Models, Trademarks), No 42, 1976.

2. Kovalenko, V. V., "Method for Determining Convective Acceleration in Open Channels," Author's Certificate USSR 606135, OTKRYTIYA, IZOBRETENIYA, PROMYSHLENNYYE OBRAZTSY, TOVARNYYE ZNAKI, No 17, 1978.

3. Kovalenko, V. V., "Method for Determining Additional Hydraulic Slope of a Nonsteady Open Flow," Author's Certificate USSR 605085, OTKRYTIYA, IZOBRETENIYA, PROMYSHLENNYYE OBRAZTSY, TOVARNYYE ZNAKI, No 16, 1978.

4. Kovalenko, V. V., "Method for Determining the Longitudinal Slope of a Water Surface," OTKRYTIYA, IZOBRETENIYA, PROMYSHLENNYYE OBRAZTSY, TOVARNYYE ZNAKI, No 19, 1978.

5. Zheleznyakov, G. V., Danilevich, B. V., TOCHNOST' GIDROLOGICHESKIKH IZMERENIY I RASCHETOV (Accuracy of Hydrological Measurements and Computations), Leningrad, Gidrometeoizdat, 1966.

6. Karasev, I. F., Chizhov, A. N., "Criteria in Measuring the Accuracy of Water Levels on Rivers and Reservoirs," TRUDY GGI (Transactions of the State Hydrological Institute), No 150, 1968.

7. Karaushev, A. V., RECHNAYA GIDRAVLIKA (Fluvial Hydraulics), Leningrad, Gidrometeoizdat, 1969.

8. Kartvelishvili, N. A., NEUSTANOVIVSHIYESYA OTKRITYYE POTOKI (Unsteady Open Flows), Leningrad, Gidrometeoizdat, 1968.

9. Katys, G. P., ELEMENTY SISTEM AVTOMATICHESKOGO KONTROLYA NESTATSIONARNYKH POTOKOV (Elements of Systems for Automatic Monitoring of Nonstationary Flows), Moscow, Izd-vo AN SSSR, 1959.

10. Kovalenko, V. V., "Measurement of the Hydraulic Characteristics of Nonsteady Open Flows," SBORNIK RABOT PO GIDROLOGII (Collection of Papers on Hydrology), No 12, 1977.

11. Kovalenko, V. V., Varyshnikov, N. B., "On the Problem of the Extrapolation of the Discharges Curve in a Case of Unsteady Movement of an Open Flow," MEZHVUZOVSKIY SBORNIK (Intercollege Collection of Articles), No 63, 1977.

12. Kovalenko, V. V., Stolyar, S. Ye., "Some Problems in Hydrometry of Unsteady Open Flows," PROBLEMY ARKTIKI I ANTARKTIKI (Problems of the Arctic and Antarctica), No 53, 1978.

13. METODICHESKIYE UKAZANIYA UGMS No 81. ORGANIZATSIYA NABLYUDENIY NAD PRODOL'NYM UKLONOM VODNOY POVERKHNOSTI (Systematic Instructions of the Main Administration of the Hydrometeorological Service No 81. Organization of Observations of the Longitudinal Slope of a Water Surface), Leningrad, Gidrometeoizdat, 1971.

14. NASTAVLENIYE GIDROMETEOROLOGICHESKIM STANTSIYAM I POSTAM, VYP 6, CH I (Instructions to Hydrometeorological Stations and Posts, No 6, Part I, Leningrad, Gidrometeoizdat, 1978.

15. Rozhdestvenskiy, A. V., Chebotarev, A. I., STATISTICHESKIYE METODY V GIDROLOGII (Statistical Methods in Hydrology), Leningrad, Gidrometeoizdat, 1974.

16. Shestakova, R. A., "Choice of the Reach for Measuring the Slope of a Water Surface in Channels Without Floodplains," TRUDY GGI, No 164, 1968.

UDC 556.(536+537)

STUDY OF KINEMATIC STRUCTURE OF FLOW IN RIVER MOUTH REACH MODEL

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 84-89

[Article by Candidates of Physical and Mathematical Sciences N. A. Mikhaylova and V. P. Petrov, O. P. Petrosyan, Moscow State University, submitted for publication 25 October 1979]

[Text]

Abstract: Data from laboratory investigations are compared with the energy spectra of turbulence in the main channel and in the mouth reach of a river. It is shown that the dimensions of the eddies in the thickness of the flow coincide with the dimensions of the macroscale channel formations at the bottom of the flow.

Despite the fact that the problem of formation of the mouth reaches of rivers has long attracted the attention of researchers, until now there has been no clear understanding of the mechanism of this process and its quantitative evaluation. The mouths of rivers are characterized by all the peculiarities of channel flow determined by the principle of interaction between channel and flow formulated by M. A. Velikanov. But at the same time there are additional difficulties associated primarily with considerable horizontal broadening of the channel, including with depth. All the mentioned circumstances determine the specific structure of flow in mouth reaches, which as of yet has not been adequately studied. In analyzing the considered flow it is evidently desirable to apply the method already developed for the investigation of channel flows [3, 4]. To be sure, it is necessary to take into account the peculiarities of movement of sand waves and bars under conditions of nonuniform movement of the flow [2].

The investigations described in this article were carried out under laboratory conditions in the hydrophysics laboratory of the Department of Physics of the Sea and Waters of the Land of the Physics Faculty at Moscow State University. The model of a microriver with a mouth reach was formed in a channel flume 21 m in length, 4 m wide and 1 m high using sand 0.2 mm in diameter. The initial groove was linear longitudinally and in cross section had a trapezoidal configuration with the following river parameters: width at bottom 30 cm, side slope coefficient 1:2, longitudinal bottom slope 0.001.

Flow Characteristics

	10 ч 1		80 ч	
	I	II	I	II
h см	11,6	17,8	10,5	11,4
2 V см/с	41,1	37,1	37,6	38,7
Fr	0,15	0,09	0,13	0,13
Re	$0,48 \cdot 10^5$	$0,67 \cdot 10^5$	$0,38 \cdot 10^5$	$0,44 \cdot 10^5$

KEY:
1. hours
2. cm/sec

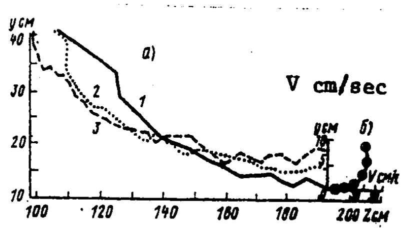

Fig. 1. Transverse sections of channel (up to its midpoint along the right bank) at post I 10 (curve 1), 50 (curve 2) and 80 (curve 3) hours after onset of experiment (a) and distribution of mean velocity in depth of flow at post I 10 hours after onset of experiment (b).

In the mouth reach the horizontal configuration persisted but the bottom slope changed sharply and was equal to 0.01. The banks of the model, which were horizontal in the river reach, had a slope 0.015. The length of the microriver was 16.5 m; the remaining part of the model was accounted for by the mouth reach of the river and sea. The microriver was formed with a water discharge Q = 44 liters/sec. The experiment lasted 82 hours. During this time the channel of the microriver and its mouth reach were reformed; the intensity of the process gradually attenuated with transition from the river to the mouth, where there was a marked broadening of the flow.

In order to judge the temporal change of characteristics of the channel and flow along the length of the model we selected characteristic control points at distances of 14 m, 17.5 m and 18 m from the entry into the flume. Point ("post") I was in the channel reach of the model where there was no influence from the broadening and deepening of the flow, points II and III were in the mouth reach. Table 1 gives the flow characteristics at points I and II 10 and 80 hours after onset of the experiment. All the characteristics relate to the vertical on the flow axis. When carrying out the experiment observations were made of formation of bottom relief in the channel and mouth. At the indicated moment in time the mean length of the sand waves in the channel was 12 cm and their rate of movement was 5 cm/sec. Measurements of the bottom profile were made systematically at the control points. Figure 1a shows cross sections at point I 10 (curve 1), 50 (curve 2) and 80 (curve 3) hours after onset of the experiment. The sections were run to the midpoint of the channel along the right bank. The cited curves show that with the course of time the shore slope becomes more gentle and the channel is widened. In this process the bottom readings in the middle of the channel increase, whereas on the slope they decrease.

As a characteristic of the degree of formation of the channel we used the flow width B at point I, which was measured each 5-6 hours. With the course of time the B value first increased rapidly and after attaining some value during the time t = 55 hours then remained essentially constant. By this time the bottom relief in the channel (point I) had already formed. In the mouth reach (points II, III) the formation of a bar was observed.

Two series of measurements of the mean and fluctuating characteristics of velocity at points I and II on the vertical in the middle of the channel were carried out for a comparison of the kinematic and energy characteristics of the flow in the channel and mouth in the stage of formation and in a formed state. The first series of measurements was made after 10 hours and the second was made 80 hours after onset of the experiment. At these same moments in time we measured the bottom profile along the axis of channel symmetry (see Fig. 2c,f). Measurements of the mean velocities were made with a vane. As an example, Fig. 1b shows the velocity curve at point I 10 hours after channel formation. The origin of the coordinate system was matched with a point on the bottom through which the control vertical passed, where measurements were made. Both in the initial stage of channel formation and in the case of a forming channel it was found that the velocity gradients in the bottom region at point I were greater than at point II. However, the elevation of the bottom region, where the mean velocity varies substantially, at point II is greater; this is attributable to broadening and deepening of the flow.

Fluctuations of the vertical and horizontal velocity components were registered using a thermohydrometer with a compensating resistor [5]. Registry was with an N-327 automatic recorder; a UT-401B amplifier with a microcircuit was used. At each point registry was for 1 min. Since the water discharge did not change in the course of the experiment, and channel deformation was insignificant during the time of registry, during this time interval the flow movement can be considered steady. In processing these records we applied a theory developed for stationary random processes [1]. Each record was represented in the form of a series of 600 points; the discreteness interval was $\Delta t = 0.1$ sec and the corresponding Nyquist frequency was

$$f_N = 1/2\Delta t = 5 \text{ Hz},$$

which is less than the limiting frequency transmitted by the recording instrument. The duration of the record (1 min) imposes a limitation on the minimum frequency

$$f_{\min} = \frac{10}{T} = 0{,}16 \quad \mathrm{Hz},$$

where T is the length of the record.

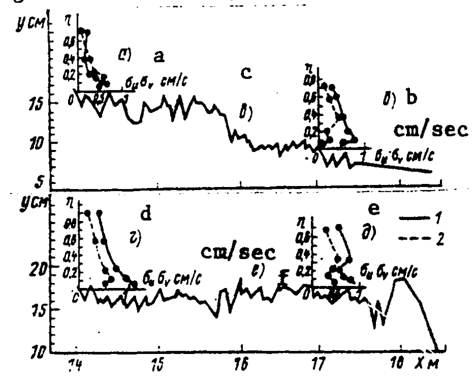

Fig. 2. Vertical distribution of standard deviations of horizontal (1) and vertical (2) velocity components at posts (points) I (a, d) and II (b, e) after 10 and 80 hours and also longitudinal section of channel after 10 and 80 hours (c, f) from onset of experiment.

Thus the range of investigated frequencies falls in the limits

$$0.16 \text{ Hz} \leqslant f \leqslant 5 \text{ Hz}.$$

On the basis of the results of statistical processing we obtained data on the distribution (in the depth of the flow) of the standard deviations of fluctuations of the horizontal σ_v and vertical σ_v velocity components at points I (Fig. 2a,d) and II (Fig. 2b,e) 10 and 80 hours after onset of the experiment. At point I the σ_u and σ_v values in both cases have a maximum at the bottom, where the main zone of production of turbulent energy is situated. Then, beginning with a distance of 0.4 H from the bottom, the intensity of the turbulence remains constant, which agrees with the investigations in [4].

At point II, with an unformed mouth (10 hours after onset of the experiment) σ_u has a maximum at the bottom and the elevation of the region with an increased σ_v value is greatly increased. However, when the mouth is formed, the σ_u and σ_v distribution curves, together with the maxima at the bottom, have a tendency to an increase in the values of the mentioned characteristics at the midpoint of the flow as well.

In order to evaluate the contribution of velocity fluctuations of different frequency to the energy of turbulent fluctuations we examined the correlation functions and the spectral density functions. These were computed from numerical series, processed in accordance with a program prepared by V. R. Nikitin applicable to the problems described in [3]. In order to reduce the record to a stationary form we employed the "moving average with a Tukey cosine-kernel" operation. In this case the "moving" mathematical expectation for the i-th term of the series was determined by the expression

$$\check{U}_i = \frac{1}{2l-1} \sum_{k=-l}^{l} U_{i+k}\left(1 + \cos\frac{2\pi k}{l}\right).$$

In order to select the optimum value of the filter parameter l we carried out a methodological study similar to that described in [3]. In all the investigated cases for both velocity components the spectral density functions S(f) are multimodal and the main energy of the turbulent fluctuations is concentrated in the region of low frequencies. As a sample, Figure 3 shows the energy spectra of fluctuations of the horizontal (a) and vertical (b) velocity components at point II 10 hours after onset of the experiment at a distance of 0.05 H from the bottom. The maximum of the S(f) functions, obtained for point I 10 hours after onset of channel formation, falls at one and the same frequency for the vertical and horizontal velocity components. At points situated at a distance of 0.05 H and 0.15 H from the bottom the spectral functions have a clearly expressed maximum at a frequency of 0.3 Hz. With increasing distance from the bottom the intensity of this maximum decreases for both the vertical and horizontal components, remaining greater for the horizontal component. However, at the surface, at a distance 0.8 H from the bottom, once again there is a marked increase in the intensity of the first maximum, evidently caused by the presence of an interface. With increasing distance from the bottom of the flow the spectra become more broad-banded.

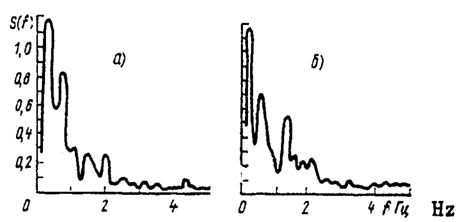

Fig. 3. Energy spectra of fluctuations of horizontal (a) and vertical (b) velocity components at point II 10 hours after onset of experiment at distance 0.05 H from bottom.

In order to estimate the degree of concentration of energy at a particular frequency we used the characteristic $f_0/\Delta f$, similar to the quality of the vibrational system $\omega_0/2\Delta\omega$ in radiophysics. Here Δf is the width of the spectral curve at the 0.7-level from the maximum value, f_0 is the frequency which corresponds to the maximum. An increase in the $f_0/\Delta f$ characteristic with an identical f_0 value is evidence of a clearer definition of structural formations in the flow. The maximum values of the $f_0/\Delta f$ parameter are observed at the bottom and at the surface of the flow and are equal to 3 and 2.5 respectively.

On the assumption of correctness of the hypothesis of "frozen-in" turbulence we made an evaluation of the size of an eddy using the expression

$$L = \overline{U}T,$$

where $\overline{U}$ is the mean velocity at a point, T is the characteristic period, obtained using the spectral function.

The characteristic size of the eddies, corresponding to a frequency f = 0.3 Hz, is 1.1-1.4 m, which is 13 H.

At this same vertical with forming bottom relief 80 hours after onset of the experiment the main energy of the turbulent fluctuations is also concentrated in the low-frequency region. However, in this case the spectra become wider-banded. For both velocity components the spectral density functions have maxima at one and the same frequencies and the first maximum falls at a frequency of 0.2 Hz. In contrast to the initial stage in channel formation, the first maximum for the vertical velocity component is everywhere more intensive than for the horizontal component. The characteristic size of the eddies corresponding to a frequency of 0.2 Hz is 1.4-1.9 m, which is 20 H and is equal to the size of macroscale formations at the bottom of the flow. The maximum values of the $f_0/\Delta f$ parameter, the same as in the initial stage of formation, are observed at the bottom and at the surface of the flow.

In spectra obtained 10 hours after the onset of formation of the mouth at the bottom at distances 0.05 H and 0.15 H from the bottom, the first maximum falls at a frequency of 0.4 Hz. The $f_0/\Delta f$ value attains a maximum value at a distance 0.05 H; 0.4 H; 0.98 H from the bottom. Both the spectral components have maxima at the same frequencies. The main energy of the turbulent fluctuations is concentrated in the low-frequency region. The characteristic dimensions of the eddies present in the bottom region of the mouth and corresponding to a frequency of 0.4 Hz are equal to 0.14-0.27 m, which corresponds to a dimension of the order of the depth of flow at this vertical. The formation of such eddies is evidently associated with a marked deepening of the flow at the investigated point. In addition to these eddies, in the upper part of the flow there are eddies measuring about 6 H. Thus, the eddies arising during the passage of the flow in the channel also persist in the mouth reach.

The spectra obtained at this same vertical after 80 hours, when the mouth has been formed, and the bar has already been formed, are wider-banded. The main energy of the turbulent fluctuations is also concentrated in the low-frequency region. The maximum values of the $f_0/\Delta f$ parameter are attained at the bottom and at the surface. The characteristic sizes of the eddies corresponding to a frequency of 0.3 Hz, at which the first S(f) maximum falls, are equal to 1.1-1.3 m, which corresponds to 12 H, that is, we again observe eddies existing also at point I (in the channel) with a formed bottom.

Summary

1. The kinematic structure of the flow in the mouth reach was governed by the broadening and deepening of the channel and differs from the structure of the flow in the channel. The bottom region, where there is a substantial change in velocity in the depth of the flow in the mouth reach is greater than in the main channel.

2. Energy spectra in the main channel and in the mouth reach are multimodal for different moments of channel formation. The main part of the energy in all cases is concentrated in the region of low frequencies. However, if the main frequency in the channel persists in the entire depth, in the mouth reach it becomes lower with an increase in elevation.

3. In the channel reach there are eddies of about 10-13 H which also persist in the upper part of the flow in the mouth reach. The indicated sizes of eddies coincide with the sizes of the macroscale channel formations on the bottom of the flow.

4. The standard deviations of the velocity fluctuations have maxima at the bottom, which indicates the existence of a region with an increased turbulence intensity. In the mouth reach the distribution of the standard deviations has a tendency to an increase in the middle of the flow.

BIBLIOGRAPHY

1. Jenkins, H., Watts, D., SPEKTRAL'NYY ANALIZ I YEGO PRILOZHENIYA (Spectral Analysis and its Applications), Vol 1, Moscow, Mir, 1972.

2. Mikhaylov, V. N., DINAMIKA POTOKA I RUSLA V NEPRILIVNYKH UST'YAKH REK (Flow and Channel Dynamics in Nontidal River Mouths), Moscow, Gidrometeoizdat, 1971.

3. Mikhaylova, N. A., Shevchenko, O. B., "Laboratory Investigations of the Characteristics of Channel Turbulence and Channel Deformation in the Coastal Region," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 4, 1976.

4. Mikhaylova, N. A., Kharchenko, I. P., "Laboratory and Field Investigations of Turbulence in Channel Flows in the Low-Frequency Part of the Spectrum," VESTNIK MGU. FIZIKA, ASTRONOMIYA (Herald of Moscow State University. Physics, Astronomy), No 6, 1976.

5. Petrov, V. P., "Instrumentation, Method and Results of Investigation of Turbulent Fluctuations of Concentration and Velocity in a Sediment-Transporting Channel Flow," Author's Summary of Dissertation for the Award of the Degree of Candidate of Physical and Mathematical Sciences, Moscow, Moscow State University, Physics Faculty, 1971.

UDC 551.5:633.14

EFFECT OF WINTERING ON THE YIELD AND GROSS HARVEST OF WINTER RYE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 90-96

[Article by V. A. Shavkunova, USSR Hydrometeorological Scientific Research Institute, submitted for publication 27 February 1980]

[Text] Abstract: The author cites the results of investigations of the effect of wintering conditions on the yield and gross harvest of winter rye. The change in the yield of winter rye during the last 20 years is investigated. Equations reflecting the trend of this change with time are derived for oblasts, republics and economic regions in which plantings of winter rye are for the most part concentrated.

The principal grain crop in the Nonchernozem zone is winter rye. About 80% of the entire area of winter rye in the USSR is located in this zone. Due to the climatic characteristics in the Nonchernozem zone its yields are more stable than in a number of other agricultural regions in the country. However, even here there are variations in yield by years. For example, during the period from 1958 through 1978 the maximum yield of winter rye in a number of regions was 22-25 centners/hectare, whereas the minimum yield was 5-9 centners/hectare (Table 1).

Many researchers have dealt with the problems relating to the influence of agrometeorological conditions on the yield of winter crops in our country [1, 3, 4, 6]. They determined the influence of wintering on the yield of winter crops on the basis of the state of sown crops in spring (using the number of plants and stems per 1 m^2 or using the thinness of plantings after wintering).

The investigations of V. A. Moiseychik [3] have demonstrated that with a decrease in the number of stems in spring in comparison with their number in autumn as a result of damage to the sown areas during the wintering period the yield is reduced with one and the same state of the winter crops in autumn. The correlation between the yield and the percentage of intact stems by spring is linear. However, its dependence on the degree of bushiness of plants in autumn has a nonlinear character. For winter rye of the Vyatka and Vyatka 2 varieties the dependence of yield on the state of the sown crops in autumn and spring is expressed by the equation

$$y=0{,}063\,\overline{K}P;\quad r=0{,}84\pm0{,}04,$$

where y is the yield, centners/hectare, $\overline{K}$ is the mean bushiness of the sown crops in autumn, P is the percentage of intact stems by the time of renewal of the spring growing season of plants.

Table 1

Yield of Winter Rye (Centners/Hectare) During Period 1958-1978

Территория 1	Средняя 2	Максимальная 3	Год 4	Минимальная 5	Год 6	Колебание за период 7
8 Ленинградская обл	15,4	24,4	1973	7,2	1961	17,2
9 Московская обл.	14,3	23,9	1973	8,2	1964	15,7
10 Воронежская обл.	14,9	23,0	1973	6,6	1963	18,4
11 Куйбышевская обл.	12,9	24,5	1978	7,0	1967	17,5
12 Саратовская обл.	12,0	22,4	1978	5,4	1963	17,0
13 Башкирская АССР	13,8	22,3	1978	9,3	1964	13,0
14 Оренбургская обл.	11,8	25,7	1978	4,6	1975	21,1
15 Белорусская ССР	14,6	26,7	1978	6,8	1958	19,9
16 Латвийская ССР	14,9	25,1	1976	7,8	1962	17,3

KEY:

1. Territory
2. Mean
3. Maximum
4. Year
5. Minimum
6. Year
7. Variation during period
8. Leningradskaya Oblast
9. Moskovskaya Oblast
10. Voronezhskaya Oblast
11. Kuybyshevskaya Oblast
12. Saratovskaya Oblast
13. Bashkirskaya ASSR
14. Orenburgskaya Oblast
15. Belorussian SSR
16. Latvian SSR

As indicated by an analysis of the data for the last 20 years which we made, the yield of winter rye, despite the increased intensification of agriculture during recent years, increases more slowly than the yield of winter wheat. In order to clarify the trend in the increase in the yield of winter rye we constructed curves of the dynamics of its yield from 1958 through 1978 and derived equations for the trend lines for 37 oblasts, 5 economic regions, the Baltic region and Belorussia, where the sowings of winter rye for the most part are concentrated. Table 2 gives these equations for economic regions.

The change in the yield of winter rye during the last 20 years, without allowance for the influence of weather during individual years, can be traced from the position of the trend line at the beginning (1958) and at the end of the period (1978). The greatest increase in the yield of winter rye as a result of an increase in the level of agricultural techniques and the introduction of new varieties into production, judging from the trend lines, was in the Belorussian SSR, Lithuanian SSR and Estonian SSR (16.5-18.5 centners/hectare). In the Latvian SSR the yield of winter rye increased by 11.9 centners/hectare, in the Northwestern and Central Regions -- by 7 centners/hectare. Its minimum increase was in the Volgo-Vyatskiy, Tsentral'nochernozemnyy and Povolzhskiy Regions (3-4 centners/hectare). However, in individual years favorable for wintering (1973, 1978), the yield of winter rye in the Tsentral'nochernozemnyy Region attained 22 centners/hectare, and in Povolzhskiy Region -- 19 centners/hectare.

Table 2

Equations for Trend Lines for Yield of Winter Rye by Economic Regions for the Period 1958 to 1978

Территория 1	Уравнения линий трендов 2	Коэффициент корреляции 3	Урожайность по тренду, 4 ц/га		
			на начало периода 5	на конец периода 6	увеличение за период 7
8 Северо-Западный район	$y=0.327T+6.066$	0,836	6,4	13,0	6,6
9 Центральный район	$y=0.343T+6.198$	0,850	6,5	13,4	6,9
10 Волго-Вятский район	$y=0.151T+7.614$	0,492	7,7	10,8	3,1
11 Центральночерноземный район	$y=0.193T+11.874$	0,313	12,0	16,4	4,4
12 Поволжский район	$y=0.151T+9.914$	0,379	10,0	13,1	3,1
13 Белорусская ССР	$y=1.174T+7.395$	0,771	7,3	23,8	16,5
14 Литовская ССР	$y=0.891T+5.444$	0,889	6,3	24,2	17,9
15 Латвийская ССР	$y=0.697T+7.481$	0,822	8,1	22,0	11,9
16 Эстонская ССР	$y=0.917T+8.382$	0,849	9,2	27,7	18,5

17 Примечание. y — урожайность по тренду, T — порядковый номер года, считая с 1958 г., номер которого взят за единицу (для БССР — 1963 г.).

KEY:

1. Territory
2. Trend line equations
3. Correlation coefficient
4. Yield according to trend, centners/hectare
5. At onset of period
6. At end of period
7. Increase during period
8. Severo-Zapadnyy Region
9. Tsentral'nyy Region

[Severo-Zapadnyy = Northwestern; Tsentral'nyy = Central]

10. Volgo-Vyatskiy Region
11. Tsentral'nochennozemnyy Region
12. Povolzhskiy Region
13. Belorusskaya SSR
14. Lithuanian SSR
15. Latvian SSR
16. Estonian SSR
17. Note. y is the yield according to the trend, T is the sequence number of the year, reckoned from 1958, the number of which was used as unity (for the Belorussian SSR -- 1963)

Table 3

Deviation Δy of Yield of Winter Rye from its Mean Value Computed With Allowance for the Trend of Increase (Trend Line) in Dependence on Area With Dead Plantings S_{dead}

	Область 1	$\sigma_{s_в}$ 2	Показатели, % 3	1965/66	1966/67	1967/68	1968/69	1969/70	1970/71	1971/72	1972/73	1973/74	1974/75	1975/76	1976/77	1977/78
4	Вологодская	7,08	Δy $S_в$ S_{dead}	−27 25,6	+33 4,8	+12 3,8	+15 10,0	+19 8,0	+23 10,8	−7 6,7	0 16,4	−22 12,4	+13 4,5	−2 7,1	−12 11,6	−30 26,7
5	Псковская	8,54	Δy $S_в$	−24 23,4	−10 4,7	−28 13,6	−7 16,0	+1 9,4	+15 5,3	0 2,8	+23 2,1	+16 6,1	+2 9,9	+16 6,2	−24 35,3	−6 12,6
6	Воронежская	13,06	Δy $S_в$	+6 4,6	−26 20,3	−6 14,1	−23 45,0	+28 0,8	+5 1,9	−14 12,4	+39 3,1	+33 6,7	−29 9,2	+19 5,2	−19 24,4	+14 4,0
7	Липецкая	18,56	Δy $S_в$	−9 3,6	−8 3,3	−6 4,7	+10 27,4	+28 12,3	+1 4,4	−7 24,2	+67 10,6	+23 15,5	−48 77,2	+6 13,3	−23 36,5	+17 13,6
8	Тамбовская	15,96	Δy $S_в$	−2 1,0	−4 2,0	−2 2,7	+1 34,5	+26 0,6	−8 2,8	−4 28,9	−30 7,0	−36 3,8	−56 57,7	+26 4,0	−17 23,4	+16 13,0
9	Куйбышевская	10,50	Δy $S_в$	−6 21,2	−44 26,7	+6 16,9	+16 27,0	+11 14,6	+18 6,8	−17 18,5	+40 16,9	+45 2,1	−41 30,9	+8 24,8	−23 36,8	+78 7,8
10	Саратовская	11,58	Δy $S_в$	−13 29,1	−12 20,2	+21 6,2	−33 41,1	+14 5,6	+12 2,7	−6 12,4	+71 11,2	+19 3,4	−48 17,4	+28 20,9	−29 36,4	+65 4,7

KEY TO TABLE 3

1. Oblast
2. σS_{dead}
3. Indices, %
4. Vologodskaya
5. Pskovskaya
6. Voronezhskaya
7. Lipetskaya
8. Tambovskaya
9. Kuybyshevskaya
10. Saratovskaya

In the trend line equations the regression coefficients characterize the mean year-by-year tendency of increase in the yield of winter rye. Their analysis by individual oblasts and republics of the European USSR indicated that the highest trend in increase in the yield of winter rye was observed in Leningradskaya (0.747 centner/hectare) and Moskovskaya (0.811 centner/hectare) Oblasts, Belorussian SSR (1.174 centner/hectare), Lithuanian SSR (0.891 centner/hectare), Latvian SSR (0.697 centner/hectare) and Estonian SSR (0.917 centner/hectare).

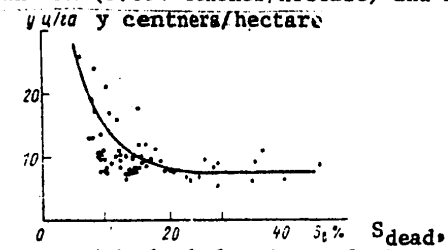

Fig. 1. Dependence of yield on area with dead plantings of winter rye in the Northwestern Economic Region.

The deviations of yield from the trend line, which characterize the degree of stability of yields, during the course of the entire analyzed period by oblasts, regions and republics consistently fell in a large range (±15-30%), but in a number of regions during the last decade even increased to ±50-60% of the yield according to the trend. In Leningradskaya, Novgorodskaya and Pskovskaya Oblasts in the Northwestern Region, in all oblasts of the Central Region, in Gor'kovskaya, Voronezhskaya, Kurskaya and Ul'yanovskaya Oblasts and Chuvashskaya ASSR, in the year 1972/73, which was favorable for the wintering and growth of crops, the positive deviations of yield from the trend were 25-50%. In years with unfavorable wintering conditions, when the plantings of winter rye in the course of 120-160 days were under a thick snow cover with weak freezing of the soil and soil temperature at the depth of the tillering node of about 0°C (1965/66, 1977/78) or were subjected to the harmful influence of strong freezes or a thin snow cover (1968/69, 1971/72), the deviations of yield from the trend attained -30 and -50%.

In a comparison of the deviation of yield and the extent of the area with plantings of winter rye killed during the autumn-winter period it was established that there is a nonlinear inverse relationship between them (Fig. 1). As a rule a large area with dead plantings of winter rye corresponds to a minimum value of its yield. We also discovered similar dependences of the yield of winter rye on the area with plantings which had perished by spring for other economic regions. This pattern was violated only in individual years with unfavorable wintering conditions, but very good conditions for yield formation during the spring-summer period. Table 3 shows that in 1968/69 winter rye perished over great areas as a result of winterkilling and was resown with spring crops; its yield in Kuybyshevskaya and Lipetskaya Oblasts was somewhat higher than the yield according to the trend. A positive deviation from the trend was observed in Kuybyshevskaya and Saratovskaya Oblasts in 1975/76 as well, when due to an autumn drought a considerable part of the area of winter rye was resown in spring.

As indicated by an analysis of data for the last 17 years, a decrease in the yield and a decrease in the harvested area of winter rye also lead to a decrease in the gross grain harvest.

Figure 2 shows the change in the gross yield of winter rye and the area of its perishing by years for the period from 1961/62 through 1976/77 for Novgorodskaya, Tul'skaya and Penzenskaya Oblasts. For the comparability of data for different years the gross yield of winter crops was reduced to a unit planted area (mean for the period) and was given in percent of its mean value. Figure 2 shows that the gross yield decreases considerably with an increase in the area of perished winter rye. In years with poor wintering conditions (Novgorodskaya Oblast -- 1965/66, 1976/77, Tul'skaya Oblast -- 1963/64, 1974/75, 1976/77, Penzenskaya Oblast -- 1962/63, 1968/69, 1974/75, 1976/77), with the perishing of crops over an area equal to 20-50% of the total area of sown winter rye, it decreased to 50-70%. In years with favorable weather conditions for the autumn growing season and the wintering of crops (Novgorodskaya Oblast -- 1966/67, 1970/71, 1972/73, Tul'skaya -- 1969/70, 1972/73, 1975/76, and Penzenskaya -- 1969/70, 1973/74, 1975/76), when the death of winter rye was not great, the gross yield was 110-170% of the mean.

In the Nonchernozem zone one of the principal reasons for the decrease in the yield of winter rye is a weakening of the crops and the death of a great number of stems as a result of their prolonged presence under a thick snow cover with only a slight freezing of the soil. The area with the dead plantings may be small, but in the remaining part of the plantings of winter rye the plants are damaged. In winter the apical cone grows without becoming differentiated, as a result of which in spring some of the sprouts completely die off or the ears are shortened, with a lesser number of grains.

We determined the dependence of the gross yield of winter rye for the territory of different oblasts on the area with dead crops in winter. The method cited in [5] was used in the computations.

For the oblasts of the Northwestern economic region the dependence is expressed by the equation

$$W = 235.5\, e^{-0{,}08\, S_{\text{в}}} + 32{,}4; \qquad (1)$$

[B = dead]

$$\eta = 0{,}671 \pm 0{,}082; \quad E_W = \pm 19{,}9\%;$$

for the oblasts of the Tsentral'nochernozemnyy (Central Chernozem) economic region

$$W = 171{,}8\, e^{-0.038\, S_{в}} + 3{,}5;$$
$$\eta = 0{,}781 \pm 0{,}058; \quad E_W = \pm\, 17{,}9\%; \tag{2}$$

[B = dead]

for the oblasts for the Volgo-Vyatskiy economic region

$$W = 168{,}7\, e^{-0.055\, S_{в}} + 29{,}5;$$
$$\eta = 0{,}762 \pm 0{,}102; \quad E_W = \pm\, 17{,}7\%; \tag{3}$$

for the oblasts of the Ural'skiy economic region

$$W = 150{,}7\, e^{-0.069\, S_{в}} + 20{,}0;$$
$$\eta = 0{,}717 \pm 0{,}062; \quad E_W = \pm\, 22{,}1\%; \tag{4}$$

for the oblasts of the Tsentral'nyy (Central) economic region

$$W = 222{,}8\, e^{-0.063\, S_{в}} + 21{,}1;$$
$$\eta = 0{,}547 \pm 0{,}091; \quad E_W = \pm\, 23{,}4\%. \tag{5}$$

Here W is the gross yield of winter rye grain over the territory of the oblasts, reduced to a unit planted area in percent of its mean value; S_{dead} is the area with dead plantings of winter rye; η is the correlation ratio; E_W is the error of the equation.

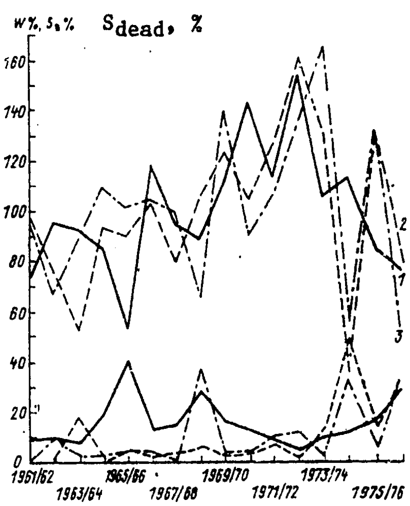

Fig. 2. Change in gross yield W and area with dead plantings of winter rye S_{dead} in Novgorodskaya (1), Tul'skaya (2) and Penzenskaya (3) Oblasts.

The correlation between the gross yield of winter rye and the area with dead plantings is nonlinear. It is expressed with particular clarity in the case of an area with dead plantings in spring equal to 15% of its sown area or more. In years with conditions favorable for wintering, when the area of death of the plantings is less than 10% of the sown area, the gross yield of grain as a rule is determined by the conditions of the spring-summer period and the level of agricultural engineering. An exception is the years when a considerable percentage of the stems of winter rye in the Nonchernozem zone perished as a result of rotting (1966/67, 1967/68). The correlation between the gross yield of winter rye and the area of destruction of plantings during the winter period, as can be seen from the cited equations, for the Volgo-Vyatskiy, Ural'skiy and Tsentral'nochernozemnyy econonomic regions was closer (η = 0.72-0.78) than in the Central and Northwestern regions (η = 0.55-0.67). This can be attributed, evidently, to the more favorable conditions for the wintering of winter rye in the Central and Northwestern regions in comparison with the Volgo-Vyatskiy region, where rye is frequently subjected to rotting, and in the Ural'skiy and Tsentral'nochernozemnyy regions where it frequently freezes out.

V. A. Moiseychik [3] earlier obtained a quantitative dependence of the gross yield on the area with dead plantings as a whole for winter crops for 1950-1970. It revealed that each percent of area with dead plantings of winter crops in spring reduces the gross yield as a whole for the USSR on the average by 1.7%.

The introduction of more winter-resistant varieties, an increase in the level of agricultural techniques for the cultivation of plants and the timely adoption of measures for the care of plantings during the early spring period in many regions to a definite degree can reduce the yield losses due to unfavorable agrometeorological conditions during the winter period.

BIBLIOGRAPHY

1. Kulik, M. S., METODICHESKIYE UKAZANIYA PO SOSTAVLENIYU DOLGOSROCHNYKH AGROMETEOROLOGICHESKIKH PROGNOZOV SREDNEY OBLASTNOY UROZHAYNOSTI OZIMYKH ZERNOVYKH V NECHERNOZEMNOY ZONE (Systematic Instructions on Preparing Long-Range Agrometeorological Forecasts of the Mean Oblast Yield of Winter Grain Crops in the Nonchernozem Zone), Moscow, Gidrometeoizdat, 1976.

2. Kuperman, F. M., Moiseychik, V. A., VYPREVANIYE OZIMYKH KUL'TUR (Rotting of Winter Crops), Leningrad, Gidrometeoizdat, 1977.

3. Moiseychik, V. A., AGROMETEOROLOGICHESKIYE USLOVIYA I PEREZIMOVKA OZIMYKH KUL'TUR (Agrometeorological Conditions and Wintering of Winter Crops), Leningrad, Gidrometeoizdat, 1975.

4. Obukhov, V. M., UROZHAYNOST' I METEOROLOGICHESKIYE FAKTORY (Yield and Meteorological Factors), Moscow, Gosplanizdat, 1949.

5. Semendyayev, K. A., EMPIRICHESKIYE FORMULY (Empirical Formulas), Moscow-Leningrad, Tekhteoretizdat, 1933.

6. Ulanova, Ye. S., AGROMETEOROLOGICHESKIYE USLOVIYA I UROZHAYNOST' OZIMOY PSHENITSY (Agrometeorological Conditions and Yield of Winter Wheat), Leningrad, Gidrometeoizdat, 1975.

UDC 551.(576.1+509.33+513)

INVESTIGATION OF A CLOUD ENSEMBLE MODEL ON THE BASIS OF GATE DATA

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 97-108

[Article by Candidate of Physical and Mathematical Sciences A. I. Fal'kovich, USSR Hydrometeorological Scientific Research Center, submitted for publication 15 January 1980]

[Text]

Abstract: The author has formulated a cloud ensemble model for the purpose of parameterization of the moist convection processes in problems of general circulation of the atmosphere and long-range weather forecasting. The investigation is carried out on the basis of observation phase III in GATE polygon A/B. A study is made of the spectrum of the cloud ensemble, the distribution of entrainment and expulsion of mass, heat and moisture balance in the cloud ensemble, heating function as a result of condensation. A new principle for the parameterization of moist convection is proposed.

In the numerical integration of the equations of hydrothermodynamics a part of the spectrum of movements (so-called subgrid processes) is cut off. Due to the nonlinearity of the equations its influence on the remaining part must be parameterized, that is, must be expressed quantitatively through parameters described by the grid, the redistribution of energy and momentum among these parts of the spectrum. Here it should not be thought that if we integrated the weather forecasting equations with a very small interval it would be possible to describe the life cycle of each cloud separately and that the parameterization is governed only by the technical possibilities of modern computers. This is not so. First, for the time being there is still no satisfactory cloud model, and second, even if it was, at the initial moment in time we will never know with the required accuracy the necessary characteristics of the cloud. The model of an individual cloud is described by a complex system of equations in hydrothermodynamics. A very important role in its development is played by microphysical processes, whose description in weather forecasting problems is impossible.

In this connection clouds in weather forecasting must be regarded statistically as a cloud ensemble, investigating its total influence on the change on the macroscale characteristics of the atmosphere. In this case it is necessary to make abstraction from the nature of an individual cloud, introducing some stylization of this cloud, and no longer model a cloud, but the entire cloud ensemble corresponding to some macroscale disturbance.

There are several different approaches to solution of this problem. For example, Fraser [6] in 1968 attempted to describe the mechanism of the life cycle of a cloud. He regarded the cloud as a "white box," consisting of the cloud proper, a homogeneous sphere, occupied by an ascending current, and the cloud "jacket," where the temperature can not only be below the temperature of the cloud, but also below the ambient temperature. The temperature minimum is attained at the visible boundary of the cloud (where the liquid water content becomes equal to zero). Writing the equation for the first law in thermodynamics for the mixture of cloud air and ambient air, Fraser finds the temperature distribution with increasing distance from the core of the cloud. As the horizontal coordinate here use is made of the ratio of the mass of cloud air to the mass of the mixture. The particles, entering into the cloud "jacket," have a negative buoyancy force relative to the ambient air. They descend first in conformity to the moist adiabat, for the time being without evaporation of all the liquid water, and then in conformity to the dry adiabat, and for the time being their temperature is not comparable to the ambient temperature. Computing the flow of mass in the settling region, it is possible to relate it to the upward transport in the cloud core and thereby compute the total vertical transport of mass generated by the "white box" at any level. It is true that for this it is somehow necessary to parameterize precipitation. Many unclear points remain in this cloud stylization. For example, it is not clear how to stipulate the cloud boundaries vertically. In this connection it is unclear whether the continuity equation is satisfactory here. The third equation of dynamics is used only qualitatively: it is assumed that the particle descends (rises) if its temperature differs from the ambient temperature to the point where these temperatures are comparable.

Another approach is based on the use of the theory of a turbulent nonisothermic jet for the modeling of a cloud [1].

Here the cloud is a jet with characteristics homogeneous along the section and is entrained into homogeneous surroundings. In the model of a jet use is made of the third equation of motion, in which an allowance is made for the buoyancy force, the weight of the liquid water in the jet, entrainment of ambient air into the jet and the resistance of the jet to the external flow. Use is also made of the equations for conservation of energy, specific liquid water content and ice content. Precipitation does not fall and ice content is parameterized by the introduction of the radius of the cloud particles. It is assumed that the pressure in the jet and in the surrounding atmosphere is identical. The entrainment is stipulated inverse proportional to the radius of the jet, which can change with altitude. The expulsion of mass and radiation cooling at the tops of the clouds are neglected. The model is stationary. The cloud top is situated where the vertical velocity becomes equal to zero. We note that it follows from the condition of stipulation of entrainment that the vertical velocity can become equal to zero only where the jet under the influence of the external flow turns and becomes completely horizontal. The slope of the jet is determined by the effect of aerodynamic drag forces; difficulties arise in the choice of the aerodynamic drag coefficient.

This cloud model was used in [2] for computing some properties of the ensemble of convective clouds on the basis of aerological sounding data. At the level of the cloud base it was assumed that the overheating of the cloud relative to the surrounding air was 0.1°C and W = 1 m/sec for all clouds. The distribution function for clouds was stipulated on the basis of horizontal dimensions, following Plank [8]:

$$N(D) = N_0 e^{-\alpha D}. \tag{1}$$

Here $0 < D \leqslant D_{max}$ is the diameter of the cloud base,

$$N_0 = f(S, D_{max}),$$

S is the part of the area occupied by clouds, α is a parameter.

The vertical structure of the cloud was determined from the expression

$$\frac{\Delta H}{D} = \gamma \left(\frac{D}{D_{max}}\right)^{\beta}, \tag{2}$$

also proposed by Plank. Here ΔH is the vertical thickness of the clouds, γ and β are parameters. We note that expressions (1) and (2) were obtained by Plank by the processing of experimental data. It is assumed that the limitation $\Delta H/D \geqslant 1$ is imposed on the cloud geometry. Using it, by means of successive approximations it is possible to determine the range of changes in the horizontal and vertical dimensions of the clouds and their internal properties with a stipulated state of the atmosphere.

The state of the atmosphere is characterized by the vertical distribution of temperature and humidity. Wind shear data are necessary for computing the slope of the jet. Large-scale divergence and vertical velocity are not used in this model and this is its basic shortcoming (the geometry of the cloud ensemble is determined only from stratification).

The most general approach to modeling of a cloud ensemble was formulated in [9]. There Yanai, et al. generalized experience acquired up to 1973 from modeling of an individual cloud and a group of clouds.

Arakawa and Shubert [5] in 1974 proposed a new theory of parameterization of moist convection based on a spectral representation of the cloud ensemble. This theory makes it possible to determine not only the general properties of the cloud ensemble, but also the properties of individual types of clouds. Then Nitta [7] investigated the properties of the cloud ensemble on the basis of data from aerological sounding in the Trades zone, obtained during the BOMEX expedition.

Yanai, et al. [10] used a spectral model for processing observation data in a polygon in the Marshall Islands.

This study is a continuation of the investigations of Yanai, Nitta, Arakawa and Shubert. In our article [4] we already dealt with the problems involved in deriving a system of equations for a model of a cloud ensemble, its numerical solution. Accordingly, here we will not discuss them and will only cite the basic hypotheses on which the model is based and we will copy some equations of the model necessary for further discussion.

As already mentioned above, in order to formulate a model of a cloud ensemble it is somehow necessary to stylize the cloud and make a number of simplifying assumptions concerning its internal and external structure. Following Yanai, Arakawa and Shubert, we will assume that:

1) active clouds occupy only an insignificant part of the cloud ensemble. These are vertical columns of a constant cross section occupied by an ascending flow and embedded in homogeneous surroundings;
2) all the liquid water which is ejected from the cloud is also evaporated here;
3) a state of saturation is attained in the cloud;
4) each cloud produces an ejection of mass in a very thin layer where the cloud loses its buoyancy, that is, where $h_{st} = \overline{h}^*$ (h_{st} is the static energy of the moist air in the cloud).

It is also assumed that in the cloud the laws of conservation of mass, energy, water vapor and liquid water are satisfied. It must be noted at once that satisfaction of the equations of motion in the cloud is not assumed, that is, the cloud dynamics is neglected. (The transport of momentum by subgrid processes must be parameterized separately.) It is true that the cloud can entrain mass from the surroundings, but the change in the rate of entrainment with altitude (this will be discussed below) will be stipulated. The vertical flow of mass in the cloud is fully determined by the continuity equation and not by buoyancy forces.

Writing the equations of conservation of mass, energy and water vapor in each cloud, summing them for the entire cloud ensemble and using the assumptions made above, the equations for the conservation of energy and water vapor can be written in the form

$$Q_1 - Q_R = -M_c \frac{\partial \overline{S}}{\partial p} - \frac{L}{c_p} \delta \hat{l}, \tag{3}$$

$$Q_2 = \frac{L}{c_p} M_c \frac{\partial \overline{q}}{\partial p} - \frac{L}{c_p} \delta (\overline{q}^* - \overline{q} + \hat{l}). \tag{4}$$

Here the line at top denotes averaging in a horizontal region which should be sufficiently great in order to contain a cloud ensemble but sufficiently small in order to be only a part of the large-scale disturbance. As a convenience in investigating the cloud ensemble we introduced the parameters Q_1 (apparent heating due to macroscale movements) and Q_2 (apparent moisture losses due to macroscale movements):

$$Q_1 \equiv \frac{\partial \overline{S}}{\partial t} + \overline{\nabla \cdot S\vec{v}} + \frac{\partial \overline{S}\,\overline{\omega}}{\partial p}, \qquad (5)$$

$$Q_2 \equiv -\frac{L}{c_p}\left(\frac{\partial \overline{q}}{\partial t} + \overline{\nabla \cdot q\vec{V}} + \frac{\partial \overline{q}\,\overline{\omega}}{\partial p}\right). \qquad (6)$$

Here $S = 1/c_p\,(c_pT + gz)$ is the static energy of dry air -- an analogue of potential temperature (measured in degrees). Equations (3)-(4) contain three parameters of the cloud ensemble: M_c -- the mass flow in clouds at the level p, its dimensionality coincides with the dimensionality ω; δ is the rate of ejection of mass from the clouds for a unit pressure interval and $\hat{l}$ is the specific liquid-water content of the cloud at the ejection level p. On the left-hand sides of the equations in addition to radiation cooling Q_R we have the parameters Q_1 and Q_2, which are determined by macroscale (averaged over a quite large area) atmospheric characteristics. We will consider them to be stipulated and determine them from experimental data.

Thus, we have two equations with three unknowns: M_c, δ and $\hat{l}$. For closing the problem it is necessary to consider still another equation giving the relationship between these parameters. The equation of conservation of liquid water in a cloud is such an equation. But this equation includes precipitation, which must somehow be parameterized. This was the approach used by Yanai, et al. in [9]. They postulated that the quantity of precipitation is proportional to the quantity of liquid water in the cloud; the proportionality factor was selected empirically from the condition of convergence of the computation scheme. It was found that it increases proportionally with altitude.

There is another way to solve this problem. We will exclude $\hat{l}$ from equations (3) and (4). We obtain the equation

$$Q_1 - Q_2 - Q_R = -M_c\,\frac{\partial \overline{h}}{\partial p} + \delta\,(\overline{h}^* - \overline{h}). \qquad (7)$$

Once again for closing the problem it is necessary to have the relationship between M_c and δ. Arakawa and Shubert [5] proposed a spectral representation of the cloud ensemble which in the last analysis makes it possible to ascertain the relationship among these parameters.

They postulated that the single positive parameter λ can completely characterize the type of cloud. As this parameter we will select the "fractional" entrainment rate. The total mass flow in the clouds M_c can be expressed as follows:

$$M_c(z) = \int\limits_0^{\lambda_D(z)} m(z, \lambda)\, d\lambda, \tag{8}$$

where $m(z, \lambda)\, d\lambda = \sum\limits_{\lambda_k \in (\lambda, \lambda + d\lambda)} M_k(z)$

is the mass flow in the "subensemble of clouds" for which λ_k falls in the interval $(\lambda, \lambda + d\lambda)$. The ejection of mass at the level z will be

$$\delta(z) = -m(z, \lambda_D(z)) \frac{d\lambda_D(z)}{dz}. \tag{9}$$

Here $\lambda_D(z)$ is the λ of clouds which give an ejection of mass at the z level. Expressions (8)-(9) also give us the sought-for additional relationship between the parameters M_c and δ. It is convenient to normalize $m(z, \lambda)$, introducing $m_B(\lambda)$, the density of mass flow at the base of the clouds: $m(z, \lambda) = m_B(\lambda)\eta(z, \lambda)$.

Now the problem becomes closed and with stipulated values of a macroscale disturbance Q_1, Q_2, $\bar{S}$, $\bar{q}$, $\bar{h}$, $\bar{h}^*$ it is possible to compute the parameters of the cloud ensemble corresponding to it if the Q_R value is known.

Source [4] gives the results of computations for the first phase of observations in GATE polygon A/B. It was GATE phase III, indeed, which was most interesting for investigations of cloud clusters; during that time cloud clusters very frequently covered a considerable part of the polygon A/3. Accordingly, here we will discuss the results obtained in GATE phase III.

Before proceeding to numerical experiments for computing the parameters of a cloud ensemble it is necessary to understand what conditions the initial data must satisfy. For example, the averaging area must be only a part of the macroscale disturbance. Unfortunately, all the polygons in which the complex investigations were made do not satisfy this property because they are very extensive. GATE polygon A/B, whose data we will use, was a hexagon with a 3.5° side (about 400 km); the polygon in the Marshall Islands in which Yanai made his investigations is still larger. The macroscale disturbances in these polygons, and this, as a rule, is the ICZ convective cloud zone, are appreciably narrower than the diameter of the polygon. Accordingly, the macroscale (averaged over the area of the polygon) atmospheric characteristics are smoothed. As was demonstrated in [3], if the dynamic characteristics (divergence and vertical velocity) in this case are computed unhurriedly the approximation of temperature and humidity averaged over the area of the polygon is frequently lost. Only a very small variability of mean temperature and humidity in this region is saved (the center of the polygon is situated at 8.5°N).

It is also bad to reduce the dimensions of the polygon because in this case the influence of the measurement errors increases. With the present-day accuracy in computations they are great even for these polygons. In order to reduce their influence we will not make computations for individual sounding times, but for the mean class of the disturbance (standard disturbance).

This is so-called class B [3], which includes all those times when a thick cloud accumulation was observed over the polygon. Since direct measurements of Q_R were made in GATE only in the nighttime hours, for the computations we took the Q_R climatic profile obtained by Dopplik and already used in our own study [4] and in those of Yanai [9, 10]. Despite the fact that the Q_R values are small in comparison with the Q_1 and Q_2 values, they play a very important role in the development of convection in the tropics. Accordingly, the lack of a true Q_R profile corresponding to the cloud accumulation over the polygon A/B can appreciably distort the results of the computations. But for the time being these data are not available and we are forced to use the Q_R climatic profile. The profiles of static energy of dry, moist and saturated air, obtained during GATE phase III, were very close to the profiles for phase I [4] and therefore we will not cite them here.

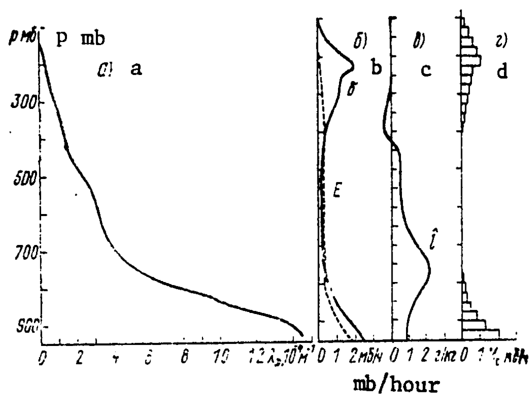

Fig. 1. Dependence of the entrainment parameter λ_D (a), entrainment E and ejection of mass δ (b), liquid water content of tops of clouds (c) and density of mass flow at base of clouds (d) on level of ejection of mass in clouds.

Figure 1a shows the dependence $\lambda_D(p)$; $\lambda_D(p)$ varies from 1 km^{-1} to 0. The maximum value $\lambda_D \sim 1.4$ km^{-1}. An investigation of the asymptotic behavior of $\lambda_D(p)$ with $p \rightarrow p_B$ shows that $\lambda_D(p) \rightarrow 0$. This is also confirmed by computations with a smaller interval. We recall that in formulating the spectral theory it was assumed that $\lambda_D(p)$ has a monotonic behavior. To be sure, the asymptotic behavior of $\lambda_D(p)$ is dependent on the upper and lower boundary conditions which are set in computing $\lambda_D(p)$. Thus, if we stipulate a small superheating downward or passage through the level of loss of buoyancy in an upward direction, then $\lambda_D(p)$ no longer will tend to zero but will tend to infinity. In this case at the point p_B there will be a singularity in any case. Its origin is attributable to the fact that low clouds are known to have roots in the mixing layer (below the condensation level), whereas according to the assumptions of the model the level of the cloud base is the condensation level and the lowest clouds begin ejection at the level of the base. We note that the $\lambda_D(p)$ values are entirely determined by stratification, that is, by the vertical distribution of temperature and humidity, and the monotonic character of $\lambda_D(p)$ is also dependent on the stratification.

Figure 1b shows the vertical distribution of entrainment and ejection of mass, computed by the spectral method described above. Since different types of clouds exist at each level, the total entrainment and ejection of mass for the cloud ensemble are observed at all altitudes. Entrainment has a maximum at the bottom, gradually decreasing with altitude. Ejection, however, has two maxima. The lower maximum is attributable to the great number of small clouds, which give ejection almost immediately over the cloud base. The upper maximum is situated near the 200-mb surface, where the maximum macroscale divergence values are observed. The dependence of the distribution of mass flow at the level of the cloud base on the ejection level has a similar character (Fig. 1d). These curves show that there is a predominance of very low and very high clouds. There are virtually no clouds giving an ejection of mass in the middle troposphere from 750 to 400 mb. Thus, the distribution in the cloud ensemble spectrum is bimodal. Similar results were obtained in the study by Yanai [9], where, evidently, attention was first given to the bimodality of the distribution. A bimodal distribution was also obtained in the processing of data in GATE phase I in polygon A/B [4]. Only in the study of Nitta [7], in which a study was made of the Trades zone, was the distribution of the mass flow at the level of the cloud base found to be unimodal. In actuality, in the cloud ensemble do only very thick and very low clouds coexist? Is this fact observed in nature? We do not know of its experimental confirmation. The Plank distribution (1), obtained using observational data, does not confirm this.

Figure 1c shows the distribution of liquid-water content at the level of cloud ejection. The quantity of liquid water at the tops of low clouds is about 1 g/kg; it gradually increases, attaining 2 g/kg for clouds having ejection near the 700-mb surface; then it rapidly decreases and becomes virtually equal to zero at the 300-mb surface. At a level of about 400 mb $\hat{l}$ even assumes small negative values. It is evident that such parameters as $\hat{l}$, δ , E, M_B should be positive. The appearance of negative values can be attributed to three factors: defects in the model, approximate nature of the initial data, errors in the numerical record. The initial data have already been discussed above; they are, of course, very approximate. But for the time being there are no better data. There are ways making possible their more precise processing and correction, but we will not discuss this here. We cannot always separate the influence of these sources of error. Only by varying the properties of the model and initial data is it possible to obtain some idea concerning their approximate character. For example, computations with $\lambda = 0$ and these same initial data gave large negative $\hat{l}$ values, beginning with 750 mb or more. By varying the initial data, for example, by stipulating radiation cooling a little high (adding to it a value increasing linearly with pressure from 0 at the level 960 mb to -1°C/day at the level 100 mb) we obtained $\hat{l}$ everywhere positive, but the same as here, the liquid-water content of the cloud tops decreases rapidly with an increase in cloud thickness. Figure 2a represents the heat balance of the cloud ensemble. Apparent heating as a result of macroscale movements without radiation cooling Q_1-Q_R is attributable to heating, which gives the term $- M_c \, \partial \bar{S}/\partial p$, and cooling due to the evaporation of liquid particles ejected from the cloud. Cooling due to evaporation is maximum at the bottom. The figure shows that fictitious heating due to errors in computing $\hat{l}$ near 400 mb (there $\hat{l} \leq 0$) is small and falls in the range of computation accuracy. The term $-M_c \, \partial \bar{S}/\partial p$ corresponds to adiabatic heating of the space surrounding the clouds if it settled at the rate M_c. From this the conclusion is frequently drawn [9] that "clouds act through the compensatory settling induced by them, heating space."

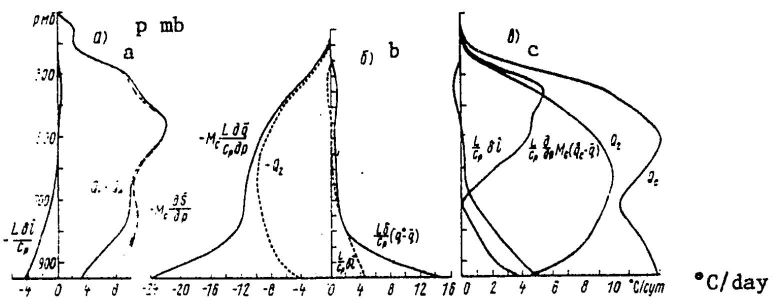

Fig. 2. Balance of heat (a), moisture (b) and heating as a result of condensation (c) in cloud ensemble.

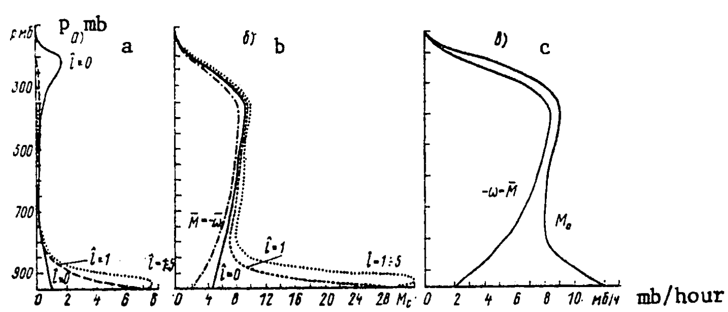

Fig. 3. Dependence of vertical profile of ejection of mass (a) and total mass flow (b, c) on quantity of liquid water at cloud tops.

Figure 2b gives the moisture balance in a cloud ensemble in heating units. Here also the apparent loss of moisture Q_2 as a result of macroscale movements for the most part is attributable to compensatory settling in the surrounding space, which dries out the space. The ejection of liquid water and water vapor at the cloud tops maintains the vapor balance in the surrounding space. Since the ejection is maximum at the bottom, it can be concluded that low clouds not yielding precipitation may be necessary for the growth of thick towers of cumulonimbus clouds; indeed, they moisten the surrounding space, ejecting water vapor and liquid water from their tops.

The atmospheric heating function as a result of water vapor condensation in clouds is of great interest. Its profile is shown in Fig. 2c. Using the equations of the cloud ensemble model it is possible to obtain

$$Q_c = Q_2 + \frac{L}{c_p} \delta \hat{l} - \frac{L}{c_p} \frac{\partial}{\partial p} \overline{q' \omega'}, \qquad (10)$$

where

$$\overline{q'\omega'} = M_c(\overline{\overline{q}}_c - \overline{q})$$

($\overline{\overline{q}}_c$ is the mean specific humidity, weighted for the cloud ensemble).

It can be seen from expression (10) that in clouds there should be condensation of sufficient water vapor for compensating the apparent loss of moisture due to macroscale movements, the evaporation of liquid particles ejected from the cloud tops into surrounding space and vertical divergence of the vertical turbulent flux of water vapor. The latter parameter has maxima at the base of the clouds and at the 350-mb level, where it even exceeds the Q_2 values. Evaporation decreases as a result of the ejection of liquid particles from the cloud tops and its contribution above the 700-mb level is already insignificant.

Thus, after stipulating the relationship between M_c and δ by use of the Arakawa-Shubert hypothesis on the spectral structure of the cloud ensemble we saw that the quantity of the liquid phase at the cloud tops decreases rapidly with an increase in cloud thickness. In addition, bimodality was observed in the cloud ensemble spectrum. We note that Yanai [9], stipulating precipitation proportional to the quantity of liquid water in the cloud and assuming that the proportionality factor increases exponentially with altitude, also obtained a bimodality, as well as a rapid decrease in $\hat{l}$ with altitude. We will dispense with the Arakawa-Shubert hypothesis and we will assume that we know the distribution of $\hat{l}$ with altitude. Now we will consider how δ and M_c behave.

Figure 3a shows the δ profiles with different stipulation of $\hat{l}$. When $\hat{l}$ = 0 a bimodality is obtained. We note that the δ maximum will always be observed in the upward direction when $\hat{l}$ there is small. With $\hat{l}$ = 1 g/kg there is no bimodality. There is also no bimodality when $\hat{l}$ linearly increases with altitude. Figure 3b shows the M_c profiles with different liquid water contents of the cloud tops. The figure shows that the M_c values change greatly in dependence on the liquid-water content in the lower troposphere; above the 850-mb level M_c is virtually not dependent on the liquid water content of the cloud tops. As a comparison Fig. 3c shows the M_c profile which is obtained when using the Arakawa-Shubert hypothesis, that is, when the dependence of $\hat{l}$ on the level of ejection has the shape of the curve shown in Fig. 1c.

We note in conclusion that here a study was made of a standard steady disturbance (the time derivatives in expressions (5)-(6) were assumed to be equal to zero). Due to the errors in measurements and approximation it is virtually impossible to compute these derivatives using measurement data from the investigated polygons. In forecasting problems the time derivatives are sought-for values and still another additional condition is required for closing the system of equations. For this purpose Arakawa and Shubert [5] propose that the assumption be made that the cloud ensemble is in a state of quasiequilibrium with the macroscale disturbance. Specifically, they assume that the time derivative of the operation of buoyancy forces in each subensemble becomes equal to zero. Indeed, at the present time this is the sole sufficiently clearly formulated principle for the parameterization of moist convection. It is true that in formulating the model of the cloud ensemble the third equation of motion was not used and buoyancy forces did not play any role in computing the mass flow in the cloud. Accordingly, it remains unclear how close this hypothesis is to reality and to what extent its realization is logically noncontradictory.

The processing of experimental data leads to the conclusion that it would be logical to use a different principle as a point of departure, specifically that the parameterization of moist convection be divided into two stages: the adaptation process and the forecast proper. We will rewrite equations (3)-(4) in the form

$$\frac{\partial \bar{S}}{\partial t} = -Q_1^c + Q_R - M_c \frac{\partial \bar{S}}{\partial p} - \frac{L}{c_p} \delta \hat{l}, \tag{11}$$

$$\frac{L}{c_p} \frac{\partial \bar{q}}{\partial p} = Q_2^c - \frac{L}{c_p} M_c \frac{\partial \bar{q}}{\partial p} + \frac{L}{c_p} \delta (\bar{q}^* - \bar{q} + \hat{l}). \tag{12}$$

Here $$Q_1^c = \overline{\nabla \cdot S\vec{v}} + \frac{\partial \bar{S}\bar{\omega}}{\partial p},\quad Q_2^c = -\frac{L}{c_p}\left(\overline{\nabla \cdot q\vec{v}} + \frac{\partial \bar{q}\bar{\omega}}{\partial p}\right).$$

They coincide with the parameters Q_1 and Q_2 in a stationary case. The variability of macroscale fields in the tropics is very small. Even with the passage of a tropical disturbance $\bar{S}$ and $\bar{q}$ do not vary greatly. In any case, the derivatives $\partial \bar{S}/\partial t$ and $L/c_p\ \partial \bar{q}/\partial t$ are appreciably less than the Q_1^c and Q_2^c values. For example, in the ICZ, in the region of polygon A/B, $\partial \bar{S}/\partial t \sim 1°C$/day, whereas $Q_1^c \sim 10°C$/day. The parameters $L/c_p\ \partial \bar{q}/\partial t$ and Q_2^c have similar orders of magnitude. Therefore, if the macroscale movements are left "alone" with the atmosphere, that is, the parameterization of moist convection is not included (in our case, in equations (11)-(12) it is assumed that $M_C = \delta = 0$), then the macroscale movements greatly "distort" atmospheric stratification, they "drive" a great quantity of water vapor into the atmosphere and reduce the quantity of sensible heat. The derivatives $\partial \overline{S}/\partial t$ and $\partial \overline{q}/\partial t$ will exceed the real values by an order of magnitude. Accordingly, moist convection in the first approximation should compensate the effect of macroscale movements, to restore the initial stratification, that is, the parameters of the cloud ensemble in the first approximation must be computed under the condition $\partial \overline{S}/\partial t = \partial \overline{q}/\partial t = 0$. This corresponds to the process of adaptation (adjustment) of the parameters of moist convection to the macroscale effect. Thus, in the first stage, using the known Q_1^c and Q_2^c values, we compute the m_B^c values which determine the main part of the density of mass flow in the spectrum of the cloud ensemble.

In order to detect the tendencies ($\partial \overline{S}/\partial t$ and $\partial \overline{q}/\partial t$) it is necessary to find the small deviations from the compensatory values of the parameters of the cloud ensemble. In our opinion, a second approximation must be based on the "splicing" of the turbulent flux of energy from the ocean F_0, which, as a rule, is computed on the basis of the semiempirical theory [3], with the turbulent energy flux in the cloud ensemble [9]

$$F(p) = \int_{p_T}^{p} (Q_1 - Q_2 - Q_R) \frac{dp}{g}. \tag{13}$$

Here p_T is the level where the turbulent energy flux is equal to zero. In actuality, F(p) characterizes the intensity of moist convection, which should be intensified with an increase in the energy flux from the ocean and weaken with its decrease. Since the parameters of the cloud ensemble were determined only with $p_T \leqslant p \leqslant p_B$, the splicing of the flows must be done at the level p_B. The boundary layer of the atmosphere, to be precise, the mixing layer, transports the energy received from the ocean upward into the interaction layer to the level of the cloud base. Small-scale turbulence participates in this energy transfer, as does cellular convection. Therefore, energy is not accumulated in the mixing layer and almost all of it is transported to the level p_B. We will denote the magnitude of this flow at the level p_B by $F_0(p_B)$ and we will assume that it is known to us. The magnitude of the turbulent energy flux in the cloud ensemble at the level p_B is determined from (13). In the first approximation the turbulent energy flux at the level p_B will be

$$F^c(p_B) = \int_{p_T}^{p_B} (Q_1^c - Q_2^c - Q_R) \frac{dp}{g}.$$

In order to satisfy the splicing condition $F(p_B) = F_0(p_B)$ it is necessary to somewhat change the parameters of the cloud ensemble obtained in the first approximation (in the adaptation stage). As a second approximation we will seek the density of the mass flow in the cloud ensemble in the form

$$m_B(\lambda) = m_B^c(\lambda) + \varepsilon\, m_B^c(\lambda), \tag{14}$$

where ε is a small parameter.

Using expressions (11)-(12) it is easy to obtain

$$\frac{\partial \bar{h}}{\partial t} = -(Q_1^c - Q_2^c - Q_R) - M_c\,\frac{\partial \bar{h}}{\partial p} + (\bar{h}^* - \bar{h})\,\delta. \tag{15}$$

Substituting (14) into (8) and (9), equation (15) can be reduced to the form

$$\frac{\partial \bar{h}}{\partial t} = \varepsilon\,(Q_1^c - Q_2^c - Q_R). \tag{16}$$

Now it is easy to find the ε parameter from the splicing condition:

$$\varepsilon = \frac{F_0\,(p_B)}{F^c\,(p_B)} - 1. \tag{17}$$

If it is assumed that the changes in $\hat{l}$ are similar to $m_B(\lambda)$ with transition from the first to the second approximation, then

$$\frac{\partial \bar{S}}{\partial t} = \varepsilon\,(Q_1^c - Q_R) - \varepsilon\,\frac{L}{c_p}\,\delta^c\,\hat{l}^c, \tag{18}$$

$$\frac{L}{c_p}\,\frac{\partial \bar{q}}{\partial t} = -\varepsilon\,Q_2^c + \varepsilon\,\frac{L}{c_p}\,\delta^c\,\hat{l}^c. \tag{19}$$

It remains to say a few words about computation of the flux $F_0(p_B)$, which is dependent on the method for parameterization of the boundary layer in the model. We will mention one of the possible methods for its computation. We will integrate the equation of conservation of static energy of moist air for the mixing layer from p_0 -- sea level to p_B, and we obtain

$$F_0\,(p_B) = F_0 - \int_{p_B}^{p_0} (Q_1^c - Q_2^c - Q_R)\,\frac{dp}{g} - \int_{p_B}^{p_0} \frac{\partial \bar{h}}{\partial t}\,\frac{dp}{g}. \tag{20}$$

Judging from the experimental data, the second term in (20) is considerably smaller than the first. For example, with a standard disturbance (class B) in the ICZ [3] $F_0 \sim 270$ cal/(cm^2·day)

$$\int_{p_B}^{p_0} (Q_1^c - Q_2^c)\,\frac{dp}{g} \sim 20 \text{ cal/(cm}^2\cdot\text{day)}, \quad \int_{p_B}^{p_0} Q_R\,\frac{dp}{g}$$

also does not exceed 20-30 cal/(cm^2·day). The first two terms in (20) are easily computed. Difficulties arise in determination of the third term because expression (16), generally speaking, is correct only under the condition $p \leqslant p_B$ (in the mixing layer the parameters of the cloud ensemble were not determined). But it is natural to assume that it is also correct when $p > p_B$, since the mass flow at the base of the clouds should be related to the mass flow in the mixing layer. Substituting (16) into (20) and using the splicing condition, we obtain the final expression for ε

$$(1+\varepsilon)\int\limits_{p_T}^{p_0}(Q_1^c-Q_2^c-Q_R)\,\frac{dp}{g}=F_0. \qquad (21)$$

Expression (21) means that we actually "lowered" the splicing condition to sea level.

BIBLIOGRAPHY

1. Bekryayev, V. I., Vorob'yev, B. M., "Jet Model of Cloud Convection. A Numerical Experiment," IZV. AN SSSR. FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), Vol VIII, No 9, 1972.

2. Bekryayev, V. I., Dovgalyuk, Yu. A., Zinchenko, A. V., "Determination of Some Properties of an Ensemble of Convective Clouds Using Data from Aerological Sounding," TRUDY GGO (Transactions of the Main Geophysical Observatory), No 405, 1979.

3. Fal'kovich, A. I., DINAMIKA I ENERGETIKA VNUTRITROPICHESKOY ZONY KONVERGENTSII (Dynamics and Energy of the ICZ), Leningrad, Gidrometeoizdat, 1979.

4. Fal'kovich, A. I., "On the Problem of Modeling of a Cloud Ensemble," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 8, 1979.

5. Arakawa, A., Shubert, W., "Interaction of Cumulus Cloud Ensemble With the Large Scale Environment," P. I. - J. ATMOS. SCI., Vol 31, No 3, 1974.

6. Fraser, A., "The White Box: The Mean Mechanics of the Cumulus Cycle," QUART. J. ROY. METEOROL. SOC., Vol 94, No 399, 1968.

7. Nitta, T., "Observational Determination of Cloud Mass Flux Distributions," J. ATMOS. SCI., Vol 32, No 1, 1975.

8. Plank, V., "The Size Distribution of Cumulus Clouds in Representative Florida Populations," J. APPL. METEOROL., Vol 8, No 1, 1969.

9. Yanai, M., Esbensen, S., Chu, J., "Determination of Bulk Properties of Tropical Cloud Clusters from Large-Scale Heat and Moisture Budgets," J. ATMOS. SCI., Vol 30, 1973.

10. Yanai, M., Chu, J., Stark, T., Nitta, T., "Response of Deep and Shallow Tropical Maritime Cumuli to Large-Scale Processes," TROPICAL METEOROLOGY PAPER No 10, 1975, Department of Meteorology, University of California, Los Angeles.

INVESTIGATION OF SPECTRA OF VARIABILITY OF METEOROLOGICAL ELEMENTS AND REQUIREMENTS ON METEOROLOGICAL MEASUREMENTS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 109-112

[Article by Candidate of Physical and Mathematical Sciences A. S. Krantsberg, Candidates of Technical Sciences K. N. Manuylov and N. A. Rubichev, V. G. Sorokin, Scientific Research Institute of Instrument Making, submitted for publication 15 January 1980]

[Text]

Abstract: This article presents some results of experimental investigation of the temporal variability of a number of meteorological elements carried out by the Scientific Research Institute of Instrument Making in collaboration with the Central Design Bureau of Hydrometeorological Instrument Making and the Administration of the Hydrometeorological Service Belorussian SSR. On the basis of these results an attempt is made to determine and validate the requirements on the desirable accuracy of the network meteorological measurement apparatus, the temporal discreteness of measurements and also the requirements on the filtering of the measurement results.

As is well known, hydrometeorological elements are functions of space and time. Their fluctuations are not regular and are not subject to a deterministic description. The value of a meteorological element at the measurement point can be represented in the form

$$\xi(t) = Z(t) + \lambda(t),$$

where Z(t) is the information component of the meteorological process of interest to the user; $\lambda(t)$ are interfering local disturbances caused by the turbulent atmosphere.

Analytically Z(t) is stipulated in the form of some mathematical model reflecting the physics of the meteorological process.

One of the widely used models of the information component of the meteorological process is its representation in the form of a power-law or harmonic polynomial whose coefficients are evaluated on the basis of measurement results.

Another mathematical model of the meteorological process is expressed by a function whose spectrum of changes is limited by some upper frequency f_{up}.

The interfering disturbances $\lambda(t)$ can be represented in the form of a multiplicative nonstationary process:

$$\lambda(t)=\psi(t)\nu(t),$$

where $\nu(t)$ is a stationary process with a higher frequency than Z(t), $\psi(t)$ is a random or determined function changing more slowly than Z(t) and possibly dependent on Z(t) or on other meteorological elements at the measurement point.

Such a representation reduces the interfering disturbances to a stationary random process. At the same time it reflects the nonstationary nature of the interfering disturbances, which, in particular, is noted in [3].

Thus, the changes in the meteorological element at the measurement point

$$\xi(t)=Z(t)+\psi(t)\nu(t)$$

are represented in the form of an additive-multiplicative model of a nonstationary random process [5].

At the present time extensive use is made of statistical methods for describing the variability of meteorological elements using covariation functions or, which is the same, using the spectral density (spectrum), that is, the distribution of the dispersion of frequency fluctuations. Accordingly, the representation of different processes and the identification of models should be carried out within the framework of the correlation theory of random processes.

The identification of models which can be reduced to stationary form essentially involves checking the stationary state of some process. For example, in the identification of an additive-multiplicative model it is necessary to check the stationary character of centered and normalized processes. The checking of the stationary state within the framework of the correlation theory can be accomplished by checking the agreement of the spectra of different segments of one record of the process. Different criteria for the agreement of spectra have been examined, for example, in [1, 6].

The reliability of the information received as a result of meteorological measurements is dependent on the correctness of choice of the mathematical model, that is, on its correspondence to a real physical process, on the dispersion value and the form of the spectrum of interfering disturbances

$\lambda(t)$, and also on the error of measurement instruments. Unfortunately, data on the degree of correspondence of the selected models $Z(t)$ to the real process are represented extremely inadequately in the literature.

Requirements on the desirable accuracy of the measurement apparatus can be determined in the first approximation using as a point of departure the problem of restoring the $Z(t)$ process on the basis of the results of network measurements made at time intervals ΔT. According to the well-known Kotel'nikov theorem, using these readings it is possible to restore a process whose spectrum does not exceed the limits of the frequency band $[0, 1/2\Delta T]$. If one uses this "band" model of the meteorological process $Z(t)$, all the components of the process registered by the instrument with frequencies above $f_{upper} = 1/2\Delta T$ must be related to $\lambda(t)$ and they must be regarded as interference distorting the measurement results, and accordingly, the process restored from them. A measure of the uncertainty arising in this case is the dispersion $\sigma^2_{low\lambda}$ of all the components of the process $\xi(t)$ with frequencies above f_{upper}. It can be determined by knowing the spectrum of changes in the hydrometeorological elements and as ΔT using the existing interval between the measurements (for most network meteorological measurements $\Delta T = 3$ hours).

However, the use of the mentioned "band" model of the process is justified if an abstraction is made from an analysis of its correspondence to a real meteorological process and as a point of departure one uses only the organizational possibilities of constructing a system for the collection of primary meteorological information. In a general case it is necessary to take into account the degree of the mentioned correspondence. Therefore, in the dispersion of the uncertainty σ^2_{low} in addition to the component $\sigma^2_{low\lambda}$, related to the distorting effect of the disturbances $\lambda(t)$, it is necessary to include the component $\sigma^2_{low\,Z}$ attributable to the inadequacy of the $Z(t)$ model to the real meteorological process.

It is desirable that the admissible value of the dispersion σ^2_{ad} of the random component of error of the measuring instrument be stipulated as a fraction of σ^2_{low}:

$$\sigma^2_{ad} = \alpha\sigma^2_{low}.$$

However, in this case it may be found that the measurement error

$$\sigma_{meas} = \sqrt{\sigma^2_{ad} + \sigma^2_{low}} = \sigma_{low}\sqrt{1+\alpha} > \sigma_{low}$$

obtained as a result does not satisfy the requirements of users of the information, following from the essence of the problems to be solved using the collected data. In this case in order to ensure the required accuracy in solving higher-level problems it is insufficient to increase the accuracy of measurement instruments. It is evident that with $\alpha < 0.3$-0.5 a decrease in instrument errors exerts virtually no influence on σ_{meas}, leading only to an unjustifiable complication and increased cost of the measurement instrument.

A substantial increase in the representativeness of meteorological information can be attained by an improvement in the measurement process in the following directions:

-- refinements in the adopted model of the information component of the meteorological process Z(t) and a corresponding decrease in the $\sigma_{low\ Z}$ value. This direction in improvement, involving the carrying out of serious physical investigations necessary for formulating the requirements on the mathematical model Z(t), should lead to a change in the algorithms for the processing of primary information;

-- improvement in algorithms for collecting and processing data from meteorological measurements for the purpose of decreasing the $\sigma^2_{low\lambda}$ value and more precise restoration of Z(t) as a result of filtering of the high-frequency components of the process;

-- decreases in definite limits of the interval ΔT between measurements, which also leads to a decrease in the dispersion $\sigma^2_{low\lambda}$, and a more precise restoration of Z(t) from the results of the primary measurements.

It goes without saying that in addition to the mentioned directions in improvement of the measurement process there is also a possibility of increasing the reliability of the meteorological information as a result of an increase in the accuracy of network measurement instruments. However, the realization of measures for increasing the accuracy of modern measurement instruments as a rule involves considerable costs and is feasible only under the condition of a matched decrease in $\sigma_{low\lambda}$. This requires having information on the stochastic characteristics of the temporal variability of meteorological elements.

An experimental investigation of the spectra of meteorological elements has been made by a number of researchers. For example, the spectra of wind velocity were obtained by Van der Hoven [7], and then by N. L. Byzova, V. N. Ivanov and S. A. Morozov [2], the spectra of air temperature -- by A. S. Monin and V. N. Kolesnikova [4]. On the basis of these investigations some conclusions were drawn concerning the nature of the temporal variability of meteorological elements. In particular, it was discovered that their spectra have a so-called "mesoscale minimum," that is, a considerable decrease in the intensity of the process in the mesoscale frequency region.

However, due to the inadequate breadth of the investigations, the relatively small volume of collected experimental data, and also due to the noncorrespondence of the experimental conditions to the conditions of standard surface meteorological measurements, the results of the preceding investigations cannot serve as a basis for formulating requirements on network measurement systems.

Taking this into account, the Scientific Research Institute of Hydrometeorological Instrument Making, in collaboration with the Central Design Bureau and the Administration of the Hydrometeorological Service of the Belorussian SSR, has carried out extensive experimental investigations of the temporal variability of four meteorological elements: air temperature, dew point, wind velocity and atmospheric pressure. Investigations were organized on the basis of M 106-M 106 M stations at different sites in the Belorussian SSR and fixed for a frequent schedule of measurements. The automation of the measurements made it possible to obtain considerable volumes of experimental data under different weather conditions and synoptic situations at different times of the year at a number of points in the territory of the Belorussian SSR. The primary research data (punched tapes) were subjected to processing with a "Minsk-32" electronic computer. Taking into account that the spectral theory of random processes has now been developed only for stationary processes, the registered series of measurements were checked for the nondependence of the mathematical expectation and dispersion on time. It was demonstrated that the initial process can be described using an additive-multiplicative model.

Values $\sigma_{\text{low}\,\lambda}$

Метеоэлемент 1	Месяц 2	3 Средние				4 Предельные	
		T=3 ч hours		T=1 ч		T=3 ч	T=1 ч
		без фильтрации 5	20-мин. осреднение 6	без фильтрации 5	20-мин. осреднение 6	без фильтрации 5	20-мин. осреднение 6
7 Температура воздуха, °C	декабрь 11	0,3	0,25	0,15	0,1	0,70	0,35
	май 12	0,55	0,5	0,3	0,2	0,95	0,45
8 Точка росы, °C	декабрь	0,45	0,35	0,25	0,15	1,2	0,5
	май	0,7	0,6	0,5	0,35	1,7	1,5
9 Скорость ветра (средняя за 2 мин), м/с	декабрь	0,7	0,5	0,5	0,35	0,95	0,7
	май	0,85	0,65	0,6	0,40	2,0	1,0
10 Давление, мб	декабрь	0,2	0,18	0,15	0,1	0,35	0,2
	май	0,2	0,018	0,[illegible]	0,05	0,40	0,25

KEY:

1. Meteorological element
2. Month
3. Means
4. Limiting
5. Without filtering
6. 20-minute averaging
7. Air temperature, °C
8. Dew point, °C
9. Wind speed (averaged for 2 minutes), m/sec
10. Pressure, mb
11. December
12. May

Then using a special program for centered records we computed the covariation functions and by means of the Fourier transform -- sample spectral densities. In order to increase the reliability of the results and decrease the dispersion of the spectral evaluations, in the processing of data we used the "Bartlett correlation window" [1] and carried out additional smoothing of the spectral evaluations by frequency bands.

For checking purposes, some records were additionally processed on an electronic computer using a fast Fourier transform. The resulting spectra virtually coincided with the spectra obtained as a result of the already described data processing.

The results of the investigations were analyzed and compared with data obtained by earlier researchers. It was confirmed in the analysis that in the mesoscale frequency region the intensities of the spectra in the overwhelming majority of cases decrease. It was discovered that the dispersions of the fluctuations in this region still are relatively great and that their values cannot be neglected; the decrease in the intensity of the spectrum with a transition from the low frequencies to the mesoscale region, as a rule, is displaced from the frequency $f_{high} = 1/2\Delta T = 1/6$ hour in the direction of the higher frequencies. We determined the values of the dispersion of uncertainty $\sigma^2_{low\lambda}$ in restoring the process of change in meteorological elements on the basis of measurements made at time intervals $\Delta T = 3$ hours and $\Delta T = 1$ hour.

The table gives the values of the mean square error in the uncertainty $\sigma_{low\lambda}$. Taking into account that the use of filtering of the high-frequency components in the spectrum decreases the uncertainty of the processes to be restored, the table gives the $\sigma_{low\lambda}$ values both for the traditional measurements, during which filtering is usually not used (except for smoothing as a result of instrument inertia), and for measurements with the use of special filtering of the high-frequency components in the form of 20-minute averaging of data. Taking into account that the variability of the meteorological elements is dependent on the season, the sample spectra were averaged separately for the winter and spring-summer seasons. Accordingly, the table contains the $\sigma_{low\lambda}$ values for both seasons.

The following basic conclusions can be drawn from the results of the investigations:

1. An additive-multiplicative model is applicable for describing the processes of temporal variability of the meteorological elements.

2. The determined values of the dispersion of uncertainty $\sigma^2_{low\lambda}$ can serve as a basis for determining and validating the requirements on the errors of the meteorological measurement instruments.

3. In order to reduce $\sigma^2_{low\lambda}$ it is desirable:
a) to use filtering of the high-frequency components, for which it is possible to use, for example, a combination of an analog filter in the form of an inertial sensor with 10-20-minute averaging of the discrete measurement results.
b) to decrease the periods of measurement and output of information to one hour.

The results can serve as a basis for further investigations whose purpose is to increase the reliability and accuracy of meteorological information supplied to the user.

BIBLIOGRAPHY

1. Bendat, Dzh., Pirsol, A., IZMERENIYE I ANALIZ SLUCHAYNYKH PROTSESSOV (Measurement and Analysis of Random Processes), Moscow, Mir, 1971.

2. Byzova, N. L., Ivanov, V. N., Morozov, S. A., "Turbulent Characteristics of Wind Speed and Temperature in the Atmospheric Boundary Layer," ATMOSFERNAYA TURBULENTNOST' I RASPROSTRANENIYE RADIOVOLN (Atmospheric Turbulence and Radio Wave Propagation), Moscow, 1967.

3. Lepekhina, N. A., Fedorchenko, Ye. I., "Diurnal Variation of Air Temperature Dispersion," TRUDY GGO (Transactions of the Main Geophysical Observatory), No 308, 1973.

4. Monin, A. S., Kolesnikova, V. N., "Spectra of Fluctuations of Meteorological Fields," IZV. AN SSSR, FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), No 1, 1965.

5. Romanenko, A. F., Sergeyev, G. A., VOPROSY PRIKLADNOGO ANALIZA SLUCHAYNYKH PROTSESSOV (Problems in Applied Analysis of Random Processes), Moscow, Sovetskoye Radio, 1968.

6. Rubichev, N. A., "Criterion of Matching the Spectra of Records of Random Processes," RADIOTEKHNIKA I ELEKTRONIKA (Radio Engineering and Electronics), No 2, 1978.

7. Van der Hoven, "Power Spectrum of Horizontal Wind Speed in the Frequency Range from 0.0007 to 900 Cycles per Hour," J. METEOROL., Vol 14, 1957.

UDC 551.(501.7+508.822)

EVALUATION OF THE INFORMATION CONTENT OF SUCCESSIVE RADIOSONDE MEASUREMENTS OF METEOROLOGICAL PARAMETERS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 112-116

[Article by Candidate of Physical and Mathematical Sciences A. F. Kuzenkov, Central Aerological Observatory, submitted for publication 30 October 1979]

[Text]

Abstract: The author gives computations of the mean information present in one measurement and the excess of the measurements of temperature and geopotential in making radiosonde observations with the "Meteorit-RKZ" apparatus. It is shown that under optimum conditions the information content of radiosonde observations in the case of measurements of the variability of meteorological elements can be increased by approximately 50%.

In the construction of measurement systems and also in the processing and storing of data it is important to determine the parameters characterizing the information content of the data obtained by the adopted measurement method, specifically, the entropy and excess in the communication. We will examine a communication whose elements are the results of successive radiosonde measurements of meteorological parameters at a definite height, carried out at an aerological station at standard times using the "Meteorit-RKZ" system [2].

Using an electronic computer we made computations of the mean information falling in one measurement of temperature and geopotential in the course of each season in the European territory of the USSR. The mean information falling in a communication element is usually expressed through entropy [1], which in a general case is dependent on the multidimensional distribution of probabilities of elements in the communication. If a communication consists of a finite number N of elements with different and independent probabilities P_j of their appearance, the entropy H and the total quantity of information C in a communication are determined by the quite simple expression

$$H = -\sum_{j=1}^{N} P_j \log P_j, \quad C = NH. \tag{1}$$

The decorrelation of elements in the communication can be accomplished by the prediction method, when the preceding value is selected as the predicted value and the difference between the actual and predicted values corresponds to the j element:

$$\varepsilon_j = T_j - T_{j-1}. \tag{2}$$

Since in forming the first differences the correlation of the elements is not completely eliminated, it is possible to take the second differences and obtain more precise evaluations of entropy. For example, the author of [10] used this method in computing the information content of telemetry in the course of radiosonde ascent.

In the course of computation of entropy the series of temperature and entropy values were subjected to transformation in the form (2). In such cases P_j, substituted into (1), corresponds to the probability of appearance of a j value in the measured process in some interval σ. The σ value characterizes the uncertainty of the j value and is numerically equal to the error in measuring the meteorological parameter. The N value then employed corresponds to the quantity (whole number) of σ intervals which fit in the range of change in the j values. In the presence of noise the entropy is reduced by the value

$$H(i) = -\sum_{j=1}^{N} P_j \sum_{i=1}^{I} P_{ij} \log P_{ij}, \tag{3}$$

where P_{ij} is the probability that the reading i will be registered when measuring the f value. This entropy H(i) is a measure of the uncertainty i in the case of known f. In our case with an a priori known j value of the measured parameter the i value obtained at the system output will fall in the limits of the measurement error σ. When making measurements at one altitude in the course of some time period the scatter of i readings is slightly dependent on the measured value of the j parameter, that is, it can be assumed that

$$H(i) = -\sum_{i=1}^{I} P_i \log P_i \left(\sum_{j=1}^{N} P_j \right) = -\sum_{i=1}^{N} P_i \log P_i \approx \log I. \tag{4}$$

In order to obtain the numerical value of the quantity of information in binary digits reduction to logarithmic form in expression (4) must be accomplished with a base of 2.

In determining the probabilities of the readings P_i the choice of the gradations G was approximately 0.2° with respect to temperature and 0.5 dam with respect to geopotential, which corresponds to the accuracy of sample measurements in radiosonde measurements [7]. The entropy value H(i) was determined on the assumption that the value of the i reading has the indicated gradations and is equally probable in the σ interval, determined by the real measurement error. The standard deviations σ were stipulated at the level of values characteristic for systems used in making radiosonde measurements of the atmosphere [5]. In this case $I = \sigma/G$ corresponds to the number of gradations of the registered reading.

Entropy H_{max} attains a maximum value in a case when the measurements of the meteorological parameters are precise ($\sigma = G$), independent and the j values of the communication elements are equally probable.

$$P_j = \frac{1}{N}; \quad H_{max} = \log N. \tag{5}$$

Using expression (4) we determined the values

$$h = H - H(I), \tag{6}$$

characterizing the information content of the communication, whose elements are the T values for temperature (or geopotential), made successively at time intervals $\Delta\tau$ = 8 hours during different seasons of the year. As a characteristic of saturation of the measurement system we selected the parameter

$$\beta = 1 - \frac{h}{H_{max}}, \tag{7}$$

called the excess [1].

For the possible use of the method of evaluations of information content in similar cases we will cite a program for computing h values written in FORTRAN language.

PROGRAM FOR COMPUTING ENTROPY OF COMMUNICATION, INCLUDING N VALUES OF METEOROLOGICAL ELEMENT

```
НБЭНТРП000000000000000000000000
DIMENSION T(300), B(300), A(300)
READ (BK1,9)N, (T(I),I=1,N)
9 FORMAT (I3/(16F5.1))
WRITE(ПЧ1, 523)N, (T(I), I=1,N)
523 FORMAT (10X, I3/(2X, 16F6.1))
L=N-1
DO1I=1,L
1 B(I)=T(I+1)-T(I)
M=L-1
DO2I=1,M
2 A(I)=B(I+1)-B(I)
CALL MINMAX(A, M, AMN, AMX)
DO7J=1, 10
G=0.2
U=G*J
D=(AMX-AMN)/U
I1=D+0.5
Z=(AMX-AMN)/(I1)
WRITE(ПЧ1, 524)Z
524 FORMAT(2X, F5.1)
Э=0.
DO6K=1,I1
S=AMN+K*Z
X=AMN+(K-1)*Z
SAM=0.
DO5I=1,M
IF(S.LT.A(I).AND.A(I).LF.X)
SAM=SAM+1.
5 CONT INUE
IF(SUM.LT.0.5)GO TO 6
P=SUM/M
Q=ALOG(P)/0.693
Э=Э-P*Q
6 CONT INUE
WRITE(П41, 525)Э
525 FORMAT (F4.2)
7 CONTINUE
END
КБЭНТРП000000000000000
```

Mean Information Content in One Measurement of Temperature and Geopotential in System for Radiosonde Measurements in Atmosphere

z км		σ °C			
		0,4	0,6	0,8	1,6
1	h β	4,48 0,36	3,98 0,44	3,50 0,50	2,50 0,60
5	h β	4,29 0,36	3,69 0,45	3,29 0,52	2,29 0,66
10	h β	4,43 0,30	3,83 0,40	3,43 0,46	2,43 0,62
20	h β	4,21 0,31	3,61 0,41	3,21 0,48	2,21 0,64

P мб mb		σ дам dam					
		1,0	2,0	3,0	4,0	6,0	8,0
900	h β	2,96 0,45	1,96 0,64	1,36 0,75	0,96 0,82	0,36 0,93	— —
500	h β	3,50 0,37	2,55 0,54	1,95 0,65	1,55 0,72	0,95 0,83	0,55 0,90
50	h β	4,76 0,24	3,76 0,40	3,16 0,50	2,76 0,56	2,16 0,65	1,76 0,72

The program operates with a mass of data including a series of T elements in the communication, the number n of elements, the G value and the number I of gradations. The printout gives the entropy value h (in the program $\ni$) corresponding to the stipulated measurement error σ (in the program Z). The table gives the values characterizing the information content of communications on the temporal variability of temperature and geopotential in the atmosphere obtained using the results of radiosonde measurements with the "Meteorit-RKZ" system. It was found that the h and β values are slightly dependent on the season, and therefore the table gives the values characteristic for summer observations.

The excess β indicates the value by which the communication is lengthened in a particular case in comparison with the minimum information on the variability of length necessary for transmission. For example, it can be seen from the data in the table that in the case of sounding at synoptic times and a mean square error in measuring temperature $\sigma = 0.6°C$ there is an approximately 50% ($\beta = 0.46$-0.51) lengthening in the series of observations in comparison with the minimum possible in the case of optimum sounding. It naturally follows from the table that with an increase in the measurement errors it is necessary to increase the series of observations, that is, the required excess β of the series increases.

In measuring geopotential (see Table) the information content of the reading increases with an increase in the altitude of the isobaric surfaces and accordingly a decrease in pressure at the isobaric surface (assuming that measurement accuracy is maintained). The communication excess decreases; this indicates that high-altitude geopotential measurements are very informative. However, in real situations the measurement accuracy decreases with a decrease in pressure at the isobaric surfaces. The information contained in one geopotential reading in this case for all the levels does not exceed h = 3.0 binary digits, whereas the communication excess falls at the level 50%. (As a comparison we point out that the information corresponding to one letter in the English language corresponds to 4.15 binary digits [1]).

The assumption of a "frozen-in" atmosphere makes it possible to apply the resulting evaluations of information content to the case of equidistant measurement of the spatial changes in meteorological parameters with intervals between the measurement points $L = \overline{V}\Delta\tau = 250$ km (with a mean velocity of the steering current $\overline{V} \approx 30$ km/hour in the transport of atmospheric inhomogeneities).

The excess of information in the communication on the variability of the meteorological elements does not at all mean that sounding must be reduced in the corresponding proportion. Sounding should be carried out in such a volume as to give an adequately complete idea concerning the weather process. In accordance with the method for processing information from a continuous source proposed by A. N. Kolmogorov [3, 4], it is necessary to form a communication which contains a set of discrete poorly correlated elements and the entropy value differs from the entropy of the initial communication by the small value ε. The evaluations which we made only indicate that there can be an average increase at the level of approxately 50% in the information content of each measurement in the case of optimum organization of radiosonde measurements in comparison with the existing "Meteorit-RKZ" apparatus.

The very presence of excess in measurements is completely natural in measurements of nonstationary atmospheric processes by means of readings which are regular in time (or space). For example, it was demonstrated in [8] that in individual regions it is common to observe weather systems close to stationary which it is sometimes possible to observe over the course of several weeks (for example, at the 500-mb surface). A radiosonde measurement made at a standard time adds virtually nothing new relative to the preceding time. In such a period in order to increase the meaningful measurement load it is desirable to increase the intervals between radiosonde launchings in comparison with the standard times. In the case of weather changes, on the other hand, it is necessary to sound the atmosphere more frequently.

We note that the maximum information content H_{max} of a reading is ensured in measurements carried out with a high accuracy and with maximum intervals between readings allowed in this process. In order to compress information the readings need not necessarily be equidistant. The practical realization of such measurements requires a priori additional information concerning the measured process and also modern radiosonde apparatus.

The problem of improvement in sounding means at the level of the sensing elements, the radiosonde itself, radar apparatus, equipment for the processing and transmission of data, is constantly at the center of attention of instrument aerology [2, 9]. The possibilities for improvement of apparatus are determined both by the modern level of technology and by economic considerations [6]. There are definite possibilities for the formation of an optimum sequence of radiosonde measurements. Optimum measurements, for example, with regulation of the place and time for radiosonde measurements, can be organized with the use of additional information on the nature of the process subjected to measurement. Such information can be obtained by using sufficiently investigated and extremely operational means for indirect sounding of the atmosphere from the surface of the land and satellites and also considerations on the predictability of measurement results.

The combination of highly precise direct remote and indirect methods for atmospheric sounding forms a highly informative adaptive system of aerological measurements with a minimum excess and in this sense is economically effective.

BIBLIOGRAPHY

1. Goldman, S., TEORIYA INFORMATSII (Theory of Information), Moscow, IL, 1957.

2. Yermakov, V. I., Kuzenkov, A. F., Yurmanov, V. A., SISTEMY ZONDIROVANIYA ATMOSFERY (Systems for Atmospheric Sounding), Leningrad, Gidrometeoizdat, 1977.

3. Kolmogorov, A. N., "Three Approaches to Determining 'Quantity of Information'," PROBLEMY PEREDACHI INFORMATSII (Problems of Data Transmission), Vol I, No 1, 1965.

4. Kolmogorov, A. N., Tikhomirov, V. M., "ε-Entropy and ε-Capacity of Sets in Functional Spaces," USPEKHI MATEMATICHESKIKH NAUK (Successes in the Mathematical Sciences), Vol 14, No 2(86), 1959.

5. Kuzenkov, A. F., Arbuzova, V. N., "Evaluation of Errors in Measuring Temperature and Geopotential by the 'Meteorit-2 - RKZ-5' Apparatus," TRUDY TsAO (Transactions of the Central Aerological Observatory), No 118, 1976.

6. Kuzenkov, A. F., Sakhnev, B. I., Lyubanskaya, V. D., "Economic Criterion of Effectiveness of an Automated Radiosonde System," GIDROMETEOROLOGIYA I NARODNOYE KHOZYAYSTVO (Hydrometeorology and the National Economy), Moscow, Gidrometeoizdat, 1976.

7. Kuzenkov, A. F., Shlyakhov, V. I., "Results of International Comparisons of Aerological Radiosondes of a Sample Type," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 4, 1973.

8. NAUCHNO-TEKHNICHESKIY OTCHET PO TEME: "RAZRABOTAT' METOD REGULIROVANIYA SROKOV RADIOZONDIROVANIYA V ZAVISIMOSTI OT PROGNOZIRUYEMYKH IZMENENIY METEOROLOGICHESKIKH PARAMETROV ATMOSFERY" (Scientific-Technical Report on the Theme: "Development of a Method for Regulating Radiosonding Times in Dependence on Predicted Changes in Meteorological Parameters of the Atmosphere), prepared under the direction of Doctor of Geographical Sciences V. D. Reshetov, Dolgoprudnyy, 1973, 50 pages.

9. Trifonov, G. P., RADIOZONDIROVANIYE ATMOSFERY I PERSPEKTIVY YEGO RAZVITIYA: OBZOR VNIIGMI-MTsD (Radiosonde Measurements of the Atmosphere and Prospects of Their Development: Review of the All-Union Scientific Research Institute of Hydrometeorological Information-World Data Center), Obninsk, 1979.

10. Trifonov, G. P., "Automatic Processing of Data from Radiosonde Measurements in the Atmosphere," Dissertation for Award of the Academic Degree of Candidate of Technical Sciences, Leningrad, Main Geophysical Observatory, 1969.

UDC 551.508.77

ON THE PROBLEM OF THE HEIGHT OF INSTALLATION OF A FIELD RAIN GAUGE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 116-118

[Article by N. N. Podgayskiy, Sverdlovsk Hydrometeorological Observatory, submitted for publication 26 November 1979]

[Text]

Abstract: The article presents the results of comparative observations of the quantity of precipitation measured by the Tret'yakov precipitation gauge and by the Davitaya rain gauge installed at different heights.

Precipitation is the principal and in essence the only source of accumulation of soil moisture and the water supply of agricultural crops under natural conditions.

A quantitative determination of the layer of falling precipitation, averaged over an area, is complicated to a considerable degree by its extremely non-uniform distribution over a territory even within the boundaries of a single farm, especially during the summer, during the growing season. The need therefore arises for measuring the quantity of precipitation not only directly in the neighborhood of a kolkhoz or sovkhoz enterprise or experimental sector, but also in agricultural fields. The installation of a great number of Tret'yakov precipitation gauges is not always possible from the economic point of view.

For carrying out the simplest observations of the quantity of precipitation during the growing season directly in fields for the time being there is only a single instrument, which is extremely cheap, the F. F. Davitaya rain gauge. The field rain gauge, despite the simplicity of its design, is convenient for observations and determines the quantity of falling liquid precipitation quite precisely.

The author, during the period May-September 1973-1975, organized and carried out comparative observations for clarifying the reliability of determination of the quantity of precipitation by the field rain gauge with total measurements once each day and month and with daily observations using the

Tret'yakov precipitation gauge. The observations were made at the meteorological station site of the observation division of the Sverdlovsk Hydrometeorological Observatory (at Verkhneye Dubrovo village).

The results of comparative observations indicated that the accuracy in determining the quantity of precipitation with the Davitaya field rain gauge evidently makes its use possible not only in daily, but also in total measurements of precipitation once each 10-day period. The preliminary conclusions from the comparative observations of 1973-1975 were published in [1, 2].

During 1978-1979 additional investigations were made for determining the influence of the height of installation of the field rain gauge on determination of the quantity of falling precipitation in comparison with the Tret'yakov precipitation gauge.

The comparative observations were made during the period May-September at the meteorological observation site at Sverdlovsk-Istok station, which is 22 km to the ESE of Sverdlovsk. The meteorological site, of the open type, was located amidst cultivated fields in an extensive forest glade with forest podzolized soil (heavy clay with an admixture of tiny pebbles). The plant cover was regularly mown grass.

There we installed two pairs of field rain gauges at a height of 1.5 m in accordance with the MANUAL [3] and at a height of 0.35 m (instrument height) respectively at a distance of 1.2 m from one another and 4 m from the site of installation of a standard Tret'yakov precipitation gauge. The observations were made by the technicians at the agrometeorological station using a Tret'yakov precipitation gauge each day at the usual observation times, and using the field rain gauges -- totally (once in 10 days, at 1800 hours Moscow time on the last day of the 10-day period).

Table 1 shows that the deviations of the quantity of precipitation determined using the rain gauge at a height of 1.5 m during the observation seasons 1978-1979 varied in the range 0.9-2.3%, whereas with installation at a height of 0.35 m the corresponding figures were 7.5-6.1% in comparison with the quantity of precipitation according to the Tret'yakov precipitation gauge.

In addition, during the growing seasons 1977-1979 observations were made using field rain gauges installed at a height of 1.5 and 0.35 m in a sector of the protected type -- in a collective garden 7 km from Sverdlovsk. Here, in a radius of 0.75 m from the instrument, installed at a height of 0.35 m, there was no vegetation of significant height, and beyond there were strawberry bushes with a height of 15-17 cm. According to observations made by the author, during the falling of shower precipitation there were no traces of splattering on the outer side of the rain gauge wall above half its height.

Table 1

Instrument and observation times	1978 (May-September)		1979 (May September)	
	Precipitation (mm)	Deviation in % from Tret'yakov precipitation gauge	Precipitation (mm)	Deviation in % from Tret'yakov precipitation gauge
Tret'yakov precipitation gauge. Ordinary observations.	468	--	261	--
Davitaya field rain gauge. Receiving part at height of 1.5 m. Observations once in 10 days.	472	0.9	267	2.3
Davitaya field rain gauge. Receiving part at height of 0.35 m. Observations once in 10 days.	503	7.5	277	6.1

Table 2

Place of observation	Sverdlovsk-Istok agrometeorological station		Collective garden at Sverdlovsk		
Period of observations	1978 (V-IX)	1979 (V-IX)	1977 (V-IX)	1978 (V-IX)	1979 (V-IX)
Quantity of precipitation (mm) measured by field rain gauges at height of 1.5 m	472	267	220	496	280
Quantity of precipitation (mm) measured by field rain gauge at height of 0.35 m	503	277	236	531	303
Deviation in % from quantity measured by field rain gauge at height of 1.5 m	6.6	3.8	7.3	7.0	8.2

Table 2 shows that the quantity of precipitation determined in a season with a field rain gauge installed at a height of 0.35 m, that is, directly at the ground surface, in both places of comparative observations, exceeded the amount determined by a similar instrument at a height of 1.5 m from 3.8 to 8.2%.

It should be noted that the seasons of comparative observations in 1977 and 1979 were arid and the shortage of precipitation was 40 and 24% respectively; 1978 was overmoist, with the quantity of falling precipitation exceeding the norm by 32%.

After summarizing the results of the comparative observations cited above the preliminary conclusion can be drawn that a field rain gauge installed at a height of 0.35 m more completely determines the quantity of falling precipitation than one installed at the standard height 1.5 m.

It is also interesting to note that when using a field rain gauge in the surface variant of the apparatus (0.35 m) there is a greater degree of concealment and accordingly it is more secure, which is a condition of more than a little importance when it is employed under field conditions.

Such investigations of the Davitaya field rain gauge, carried out in different regions of the country, make it possible to draw final conclusions concerning the feasibility of its installation in the surface variant.

BIBLIOGRAPHY

1. Podgayskiy, N. N., "Use of the Davitaya Field Rain Gauge as a Total Instrument," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 2, 1975.

2. Podgayskiy, N. N., "Total Measurement of Precipitation in Rotated Fields," METEOROLOGIYA I GIDROLOGIYA, No 10, 1977.

3. RUKOVODSTVO DLYA AGROMETEOROLOGICHESKIKH POSTOV, MTS, KOLKHOZOV I SOVKHOZOV (Manual for Agrometeorological Posts, Machine Tractor Stations, Kolkhoz and Sovkhoz Enterprises), Leningrad, Gidrometeoizdat, 1955.

REVIEW OF MONOGRAPH BY I. D. KOPANEV: SNEZHNYY POKROV NA TERRITORII SSSR (SNOW COVER OVER THE TERRITORY OF THE USSR), LENINGRAD, GIDROMETEOIZDAT, 1978, 180 PAGES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 119-120

[Review by Candidates of Geographical Sciences A. I. Voskresenskiy and N. N. Bryazgin]

[Text] The author of this newly published book, I. D. Kopanev, is one of the outstanding specialists in our country in the field of study of the snow cover. The new monograph which he has written contains a scientific generalization of an enormous volume of long-term observations of the snow cover by more than 600 hydrometeorological stations and posts in the USSR.

The reviewed book is distinguished by its practical nature. It is intended for a broad circle of readers, not only professional climatologists and geographers, but also workers in agriculture and transportation, and also those at planning and construction agencies. The monograph includes extensive, well-selected reference material and also the results of analysis necessary for solving scientific problems related to the climatology of the snow cover.

Much attention is devoted to the stochastic analysis method, by means of which it is possible to take into account the entire statistical totality of initial information. This enabled the author to obtain some characteristics of the snow cover with a stipulated guaranteed probability for most of the stations in the USSR (excluding mountain and arctic stations).

The book gives a concise history of the method for making snow-measuring observations since 1892 and its changes during subsequent years. The author feels that at the present time the method for carrying out snow-measuring observations in general has been considerably improved. However, it should be noted that the last review of the method for snow-measuring observations, made in 1965 and directed to a decrease in the volume of observations (a changeover from snow surveys every 10 days to monthly surveys only) reduced the information yield of data on the snow cover. This applies, in particular, to the northern regions, where as a result of frequent blizzards there is a considerable horizontal redistribution of the snow cover.

In the monograph considerable attention is devoted to an investigation of the accuracy of standard snow-measuring observations. Using the mathematical statistics approach, the author defined the most advantageous solutions of the evaluation of errors in snow-measuring observations and a method for computing the temporal characteristics of the snow cover. The conclusion is drawn that the maximum admissible error in snow-measuring observations (5-10%) in different regions of the USSR requires a different choice of the number of measurement points, the time interval, length and form of the route. In connection with the change in the method for making snow-measuring observations the author draws attention to the necessary evaluations of homogeneity of series of snow-measuring observations. One of the sections of the monograph is devoted to this problem. Here the author has established that the inhomogeneity in the series of snow-measuring observations at most hydrometeorological stations in the USSR is insignificant and is within the limits of measurement error.

In a number of sections in the monograph the author also gives the climatic aspects of the snow cover. For example, the duration of the snow cover, the dates of formation and destruction of the snow cover, are presented in the form of maps and tables giving the characteristics of their spatial and temporal variability over the territory of the USSR. These results will be useful in solving a number of practical and methodological problems.

In the monograph the information on the depth of the snow cover is given in the greatest detail. Its mean characteristics are given for the territories of the administrations of the Hydrometeorological Service, separately for the Baykal-Amur Railroad and the entire Soviet Union in the form of monthly maps and also maps of the mean maximum and extremal depths of the snow cover on the basis of data from snow-measuring surveys. As new information the author gives the variability of the depth of the snow cover in the course of winter and also evaluations of changes in its depth in field and forest sectors in comparison with observational data obtained using permanent rods. However, it must be admitted that the problem of the nonrepresentativeness of observations with permanent rods at meteorological stations in arctic and probably subarctic regions remains timely. In such areas these observations possibly should be abolished since they do not reflect the real distribution of the snow cover and accordingly cannot be used for practical and scientific purposes.

The computations of the probability of distribution of the snow cover by regions and individual stations made by the author make it possible to satisfy more completely and on a modern level the requirements of many branches of the national economy. Using the tables and nomograms presented in the monograph practical workers can independently obtain the necessary information. The same purpose is solved by the material, presented for the first time, on the spatial and temporal variability of depth of the snow cover. The author established a considerable year-to-year variability of depth of the snow cover (on the basis of data for stations with long series of observations), dependent on the combinations of conditions of circulation of the atmosphere and local physiographic conditions.

In the monograph considerably less attention is devoted to two other characteristics of the snow cover -- density of the snow cover and water reserve in the snow cover. However, the author has been able to supplement considerably the materials contained in handbooks on the climate of the USSR. In addition to variability of density of the snow cover, the author also examined the characteristics of blizzard activity, and also the correlation between data on water reserves in the snow cover and precipitation, etc. The results of investigations described in this section can be used for solving different practical problems, in particular, in computations of snow transport during blizzards and computations of the water balance.

The appearance of the new publication must be regarded as an important contribution of hydrometeorology to supporting the needs of the national economy of the country.

REVIEW OF MONOGRAPH BY A. P. FEDOSEYEV: AGROTEKHNIKA I POGODA (AGRICULTURAL TECHNIQUES AND THE WEATHER), LENINGRAD, GIDROMETEOIZDAT, 1979, 240 PAGES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 120-121

[Review by Candidate of Geographical Sciences V. N. Strashnyy and G. Z. Goloverdyuk]

[Text] In this book by A. P. Fedoseyev, which was edited by Academician I. S. Shatilov of the All-Union Academy of Agricultural Sciences, agricultural specialists will find information on the influence of agrometeorological conditions on agricultural production, and also on different agricultural engineering methods which make possible more effective use of the potential natural resources of a territory and favor obtaining higher and more stable yields.

The author demonstrates the results of investigations of the influence of differentiated agricultural techniques for the cultivation of agricultural crops on their yield in relation to climatic and agrometeorological conditions. The author sets forth the essence of recommendations on taking into account existing and anticipated weather conditions in determining the desirable structure of sown areas, methods and times for working the soil, determining the optimum times and doses of application of mineral fertilizers, times for the sowing of grain crops, and also for the care for sown areas and carrying out the harvest.

It was established as a result of the investigations that with deviation of the times for the sowing of spring and winter grain crops from the optimum times a noncorrespondence arises between the biological needs of plants and the prevailing meteorological conditions and this leads to a decrease in crop yield. The possibility of selecting the optimum times for sowing and the optimum seeding norms for grain crops in each specific year in dependence on meteorological conditions and the climatic probability of different gradations of precipitation in July is made clear.

It is demonstrated on the basis of extensive experimental data that with allowance for the state of grain crop sprouts and prevailing agrometeorological conditions it is possible to select the corresponding agricultural techniques for caring for sown areas favoring an increase in crop yield.

The agrometeorological criteria of desirability of packing the soil and harrowing sown areas are given.

The influence of unfavorable agrometeorological conditions during the cold season of the year on the state of winter grain crops and methods for protecting sown areas from their effect is considered.

The book gives the agrometeorological indices for the beating down of the principal regionalized varieties of barley and a method for predicting the beating down of crops and the effectiveness of their processing with TUR preparation are examined in dependence on weather conditions.

The author attaches much importance to the choice of the optimum techniques of harvesting work, taking into account the influence of agrometeorological conditions on the magnitude of the grain losses.

In the book special attention is devoted to the effectiveness of mineral fertilizers in dependence on weather conditions. Among the agrometeorological factors determining the effectiveness of fertilizers the author includes the level of the exposure of plants to light, temperature, air and soil moisture content. In the climatic aspect, a decrease in the annual quantity of precipitation from the northern to the southern agricultural regions of the European USSR by 100 mm causes a decrease in the effectiveness of moderate doses of fertilizers on the average by 1.1 centner/hectare of grain for grain crops as a whole and by 1.9 centner/hectare for winter crops. A decrease in the reserves of productive moisture in the soil during the growing season for grain crops by 10 mm results in a decrease in the effectiveness of fertilizers on the average by 0.1-0.2 centner/hectare of grain.

It is shown that the content of nitrates in the soil in spring is dependent on the quantity of precipitation during the winter period and the nature of snow melting. With precipitation of 190-200 mm or more the content of nitrates decreases sharply, which predetermines the high effectiveness of nitrogen fertilizers.

The author has proposed equations for computing evaluations of the effectiveness of fertilizers as a function of meteorological factors and individual agrochemical properties of the soil.

Maps of the average effectiveness of fertilizers for grain crops in dependence on agroclimatic conditions are given. Methods are proposed for determining the precipitation for the autumn-winter period in ascertaining the optimum doses of nitrogen fertilizers for grain crops in each specific year. Agrometeorological recommendations are given for optimizing the doses of nitrogen topdressing for grain crops in dependence on meteorological conditions. It is shown that allowance for the quantity of falling precipitation or soil moisture content during the autumn period makes it possible to make a correct decision concerning the advantage of sowings of winter or spring grain crops for the purpose of obtaining the maximum yield in each specific year.

In the author's opinion, allowance for agrometeorological conditions when using differentiated agricultural techniques for the cultivation of agricultural crops and the making of optimum agricultural engineering decisions constitutes a significant, but for the time being a poorly used reserve for the increasing of crop yields.

The book also gives examples of computations of the economic effectiveness of different agrometeorological recommendations.

The book has individual shortcomings. For example, there is inadequate discussion of the matters of using two- and three-day weather forecasts and also predictions of air temperature and precipitation anomalies for 5 and 10 days for the making of economic decisions in agricultural production. References are made to long-range predictions of the time of onset of the summer precipitation maximum, climatic stochastic forecasts, stochastic synoptic forecasts of moistening conditions despite the fact that such forecasts are not made in the system operated by the State Committee on Hydrometeorology.

In general the book merits a high evaluation because a study of this type has appeared for the first time and its publication must be regarded as a significant contribution to solution of the problem of introduction of hydrometeorological information into agricultural production. It will be a valuable aid for a wide range of specialists in agriculture and agrometeorologists. Without question, it will be read with interest and profit.

SIXTIETH BIRTHDAY OF SAMUIL MOISEYEVICH SHUL'MAN

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 122-123

[Article by members of the Board of the USSR State Committee on Hydrometeorology and Environmental Monitoring]

[Text] Samuil Moiseyevich Shul'man, Director of the West Siberian Regional Scientific Research Institute and Chief of the West Siberian Territorial Administration of Hydrometeorology and Environmental Monitoring, marked his 60th birthday on 20 July 1980.

Samuil Moiseyevich began his work activity in the Hydrometeorological Service in 1944 after graduation from the Higher Military Hydrometeorological Institute in the post of meteorological engineer at the polar station Amderma. During the period 1945-1947 he worked as a meteorological engineer at the aviation meteorological station Sofia in Bulgaria. In May 1947, after demobilization from the ranks of the Soviet Army, Samuil Moiseyevich was sent to Austria, where until November 1949 he worked as chief of the foreign aviation meteorological station in Vienna. The next five years he headed prognostic subdivisions at the Administration of the Hydrometeorological Service of the Karelian-Finnish SSR.

Beginning in 1954 and to the present time the work activity of S. M. Shul'man has been associated with the West Siberian Territorial Administration of Hydrometeorology and Environmental Monitoring, which he has headed since 1959. Samuil Moiseyevich is devoting much effort to the development and improvement of a system of prognostic agencies and a network of stations in the administration. He is devoting great attention to improvement in the forms and methods for hydrometeorological support of Party and soviet agencies in Western Siberia, which has led to an increase in the effectiveness of operation of the service in the national economy.

The broad development of scientific research in the subdivisions of the administration led to the organization in 1968 of the Novosibirsk Affiliate of the USSR Hydrometeorological Center and its subsequent transformation in 1970 into the West Siberian Regional Scientific Research Institute.

While combining the tasks of chief of the administration and director of the institute, S. M. Shul'man exhibits concern about the development of a broad complex of scientific investigations in the field of the hydrometeorology of Siberia, an increase in the scientific potential of the institute and the creation of a progressive scientific atmosphere there.

The activity of S. M. Shul'man, directed to the universal development and strengthening of creative contacts with the key institutes of the Siberian Division USSR Academy of Sciences and the Siberian Division of the All-Union Agricultural Academy, has been of great importance in establishing the institute and ensuring a high level of investigations by its workers. Samuil Moiseyevich made a definite contribution to the creation and successful operation of the West Siberian Regional Computation Center, the Novosibirsk Service of the Automated System for Data Transmission and the Center for the Reception and Processing of Satellite Information. Under his direction and with his direct participation work is being successfully done on the organization of a national service of observations and monitoring of environmental contamination. During the time of his work at the Administration of the Hydrometeorological Service he has done much work on the introduction of modern technical equipment in the network of stations in the administration, on the development of an automated system for the processing of hydrometeorological information at the West Siberian Regional Hydrometeorological Center, and in compiling and issuing regime and reference materials.

Samuil Moiseyevich is carrying out much work for strengthening the operational and observation agencies of the West Siberian Territorial Administration of Hydrometeorology and Environmental Monitoring.

In addition to his great routine productive activity and organizational work, S. M. Shul'man is actively participating in public life. The businesslike and personal qualities of Samuil Moiseyevich have won him merited authority in the organizations of the State Committee on Hydrometeorology, among subordinates, and also in the organizations of other departments.

The services of S. M. Shul'man in the development of the Hydrometeorological Service, as well as active participation in public life, have been recognized by government awards -- the "Emblem of Honor" and medals.

In warmly congratulating the veteran of the Hydrometeorological Service on his noteworthy anniversary, we wish him long and productive years of life, strong health and new work successes.

AT THE USSR STATE COMMITTEE ON HYDROMETEOROLOGY AND ENVIRONMENTAL MONITORING

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 p 123

[Article by V. N. Zakharov]

[Text] During the period 25-27 March 1980 the State Hydrological Institute held a coordination conference on the draft of a plan for scientific-research and experimental design work for 1981-1985 in the field of hydrology of the land.

A number of scientific reports were presented prior to discussion of the draft of this plan. The most important scientific and technical problems in the field of hydrology of the land in the Eleventh Five-Year Plan were discussed by the conference chairman, the director of the State Hydrological Institute A. A. Sokolov. Reports were presented on the present status and tasks of hydrological investigations in the Ukraine (A. V. Shcherbak), in Central Asia (Yu. N. Ivanov), in the Far East (V. N. Glubokov), in Kazakhstan (V. V. Golubtsov), in Transcaucasia (V. Sh. Tsomaya), on the prospects for the development of investigations for study of the mouth reaches of rivers (M. M. Rogov) and on the status and prospects for development of investigations for study of the quality of water resources (L. V. Brazhnikova).

Communications of the scientific directors of the corresponding sections were presented in relation to the draft plan presented for consideration. The draft plan was approved after taking into account the comments and additions expressed or proposed in the course of the discussion.

The plan dealt with such themes as changing the water balance, hydrological regime and surface water resources, hydrometeorological basis for the territorial redistribution of water resources, water balance, water regime and hydrological computations, channel processes, erosion and sediments, and also on some other sections.

In addition to the scientists and specialists of the State Hydrological Institute, the conference was attended by representatives of the State Oceanographic and Hydrochemical Institutes, Far Eastern Scientific Research Institute, Transcaucasian Scientific Research Institute, Kazakh Scientific Research Institute, Central Asian Scientific Research Institute, Ukrainian Scientific Research Institute, Institute of Water Problems USSR Academy of Sciences.

AT THE ALL-UNION EXHIBITION OF ACHIEVEMENTS IN THE NATIONAL ECONOMY

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 123-125

[Article by S. B. Iokhel'son, I. A. Koloskov and M. M. Novikov]

[Text] The problem of preservation of the environment occupies one of the first places among the vitally important problems exciting mankind today. In our country the preservation of the environment has been elevated to a constitutional issue and it is being devoted great attention on an everyday basis.

A special exhibit entitled "The Environment — Reliable Monitoring" was opened on 10 January 1980 in the pavilion "USSR Hydrometeorological Service" at the All-Union Exhibition of Achievements in the National Economy. This exhibit reflected the achievements brought about in this field. The exhibit consisted of five main sections: State System for Observing and Monitoring the State of the Environment; Instruments and Methods for Studying and Monitoring the Environment; Methods for Routine and Long-Range Forecasting; Scientific Research Work; International Activity of the Institutions of the State Committee on Hydrometeorology in the Field of Preservation of the Environment.

In the Soviet Union a National Service for Observing and Monitoring the Level of Environmental Contamination has been established with the direct participation of the key scientific research institutes of the State Committee on Hydrometeorology (Institute of Applied Geophysics, Main Geophysical Observatory, State Hydrochemical and Oceanographic Institutes, and Institute of Experimental Meteorology). Its activity can be characterized by the following data: the state of the air basin is monitored in 350 cities in our country, including at stationary observation points in 250 cities; each year laboratories make more than 3 million analyses of aerosols and gases in the atmosphere; monitoring of contamination of waters of the land is carried out in 1,900 rivers, lakes and reservoirs and in 14 seas; observations in fresh-water bodies are carried out at 4,000 points, and in sea waters — at 1,800 stations; the total number of analyses of water samples annually exceeds 2 million. The network for the monitoring of soil contamination, organized four years ago, even now monitors the content of pesticides and metals at 1,700 points in the Soviet Union.

Water quality is monitored with respect to hydrobiological indices in 170 rivers, lakes and seas.

The effectiveness of measures for preventing the contamination of the environment is dependent to a high degree on the instruments and equipment used in monitoring. The wide variety of instrumentation shown at the exhibit shows that the main direction in developments in the field of analytical instrument making is directed to ensuring the reliability, operability and effectiveness of monitoring of the state of the environment. The exhibit familiarizes visitors with new instruments, standard-produced by industry, introduced in the system of the State Committee on Hydrometeorology and other ministries and departments, and also new instrument developments in this field.

During recent years automatic systems for observation, collection and processing of information have been successfully introduced into the monitoring network. The Central Design Bureau of Hydrometeorological Instruments, together with other organizations, is demonstrating an automatic station for the monitoring of atmospheric contamination -- ASKZA - avtomaticheskaya stantsiya kontrolya zagryazneniya atmosfery. It makes possible monitoring and transmission of data on contamination of atmospheric air with CO and SO_2.

The Hydrochemical Institute is exhibiting an automatic station for the monitoring of surface waters of the land -- ASKPV -- avtomaticheskaya stantsiya kontrolya poverkhnostnykh vod sushi. The station makes possible the simultaneous collection of information on seven parameters (oxygen content, turbidity, pH, and others).

Among the most interesting instruments shown at the exhibit we can mention the following:
-- the "Atmosfera II" truck laboratory, intended for carrying out expeditionary work and routine monitoring of air contamination (Safonovskiy Plant);
-- the "Komponent" sampling apparatus, making it possible to take 32 gas samples from the air in accordance with a stipulated program (Leningrad Special Design Bureau of Thermophysical Instrument Making);
-- portable sensitive instruments for continuous measurement of the atmospheric dust content (Leningrad Instrument of Aviation Instrument Making);
-- a complex of laboratory equipment, including the S 112 atomic absorption spectrophotometer, the APV-102 automatic photocolorimetric analyzer and the S603 spectrophotometer for determining organic and inorganic contaminants in objects in the environment (Tbilisi Scientific-Production Combine "Analitpribor"). The S 112 instrument has a large range of lamps, attachments and auxiliary devices and in its sensitivity is equal in every way to similar foreign spectrophotometers;

-- experimental model of the "Khrustal' 4001" two-channel gas chromatograph; the instrument is supplied with four detectors for the registry of a wide range of organic compounds (Hydrochemical Institute and Moscow Special Design Bureau of Gas Chromatography).

In addition, at the exhibit there was demonstration of standard samples of soil and water intended for the calibration of instruments fabricated at the Moscow Soil Institute imeni V. V. Dokuchayev and the Odessa Physicochemical Institute. In the soil samples it was possible to certify the presence of 34 elements, in the water samples -- 17 elements, including mercury, lead, cadmium, antimony and other toxic elements.

The experimental operational system for the collection and processing of daily information arriving from the atmospheric monitoring network, developed at the Institute of Applied Geophysics, is of great practical importance. The information analyzed and generalized on the "Minsk-32" electronic computer is transmitted on the very same day to interested organizations for the adoption of the corresponding decisions.

In one of the sections of the exhibit information was given on the USSR State Standards, prepared with the direct participation of the Main Geophysical Observatory, intended for safeguarding the purity of the air basin of our country:
GOST 17.2.1.04-77 -- "Meteorological Aspects of Contamination and Industrial Wastes. Principal Terms and Definitions";
GOST 17.2.3.01-77 -- "Rules for Monitoring the Quality of Air of Populated Places";
GOST 17.2.3.02-78 -- "Rules for Setting the Admissible Wastes of Harmful Substances by Industrial Enterprises."

The Central High-Elevation Hydrometeorological Observatory presented a method for computing the admissible load of waste waters in watercourses and regulation of their discharge. The introduction of this method in the purification structures of the Kurovskiy Melange Combine and the Orekhovo-Zuyevskiy Plant "Karbolit" considerably improved the quality of water in the rivers of the Moscow region.

A section of particular interest was that telling of the scientific research work carried out in the field of monitoring of contamination of the environment. At the Institute of Applied Geophysics specialists have developed a mathematical model for routine computation of the transport of contaminating substances across USSR national boundaries. Data on the fluxes of transported substances are regularly transmitted to international organizations. Using a flying laboratory created at the Institute of Applied Geophysics, a study was made of the distant transport of sulfur compounds, nitrogen oxides, mercury vapors, hydrocarbons, pesticides and metals in the atmosphere. The collected data made it possible to formulate models of behavior of mercury in the biosphere.

The Institute of Applied Geophysics is presenting a method which makes it possible to use standard network snow-measuring surveys for determining contamination of the snow cover by the fallout of heavy metals, sulfur compounds, pesticides, benzapyrene, etc. This makes it possible, without significant additional expenditures, to determine the wastes expelled by individual enterprises and to study the distant transport of contaminating substances.

Aircraft investigations of aerosol effluent are made by the Institute of Applied Geophysics. Interesting data have been obtained on the fractional composition of smoke discharged at the Shchekinskaya State Regional Electric Power Station.

The Northwestern Administration of the Hydrometeorological Service, in collaboration with the Botanical Institute imeni Komarov USSR Academy of Sciences, has developed an experimental method for the joint monitoring of atmospheric contamination in Leningrad on the basis of chemical and biological indices. Several types of ferns have been used in carrying out bioindication monitoring, since these have increased sensitivity to contaminations.

Specialists at the Institute of Applied Geophysics have created a remote apparatus on the basis of a CO_2 laser for highly sensitive monitoring of the atmospheric content of ozone, ammonia, ethylene and other contaminating gases. An operational model of the apparatus is being demonstrated. At the Institute of Applied Geophysics specialists have also developed a method for determining the mass concentration of aerosol on the basis of the results of multifrequency laser sounding which is undergoing testing.

Specialists at the Main Geophysical Observatory have developed methods for numerical modeling for studying the patterns of dispersal of impurities and establishing the admissible discharge into the atmosphere. These methods served as a basis for the INSTRUCTIONS ON COMPUTING THE ATMOS_HERIC SCATTERING OF HARMFUL SUBSTANCES PRESENT IN THE EFFLUENT OF ENTERPRISES SN369-74.

One of the sections at the exhibit was devoted to methods for predicting the levels of environmental contamination.

A method for long-range prediction of the effect of economic activity on the state of the environment has been created under the direction of scientists at the Institute of Applied Geophysics. Methods for meteorological prediction of high levels of atmospheric contamination have been developed at the Main Geophysical Observatory. In 1979 warnings concerning an anticipated high atmospheric contamination were prepared for 103 cities.

The Hydrochemical Institute and the Institute of Applied Geophysics for the first time have developed a method for the routine prediction of contamination of river water. The method developed by the State Hydrochemical

Institute has undergone practical testing at a number of administrations of the Hydrometeorological Service. The results of the checking were entirely satisfactory.

During recent years the Northern Administration of the Hydrometeorological Service has begun to predict the oxygen regime of rivers. The measures taken on the basis of the forecast prevented the freezing-in of fish in the rivers of the Severnaya Dvina basin during the severe winter cold of 1977 and 1979.

In conclusion we should note the exhibit devoted to international cooperation in the field of monitoring the state of the environment. The Soviet Union is taking an active part in all the principal international programs. Particularly important is realization of a global system for monitoring the environment. The member countries of the Socialist Economic Bloc are participating in solution of this problem, together with other countries. The first joint expeditionary experiment of the member countries of the Socialist Economic Bloc on the problem of the global system for monitoring the environment was carried out in the autumn of 1979 in the territory of the Hungarian People's Republic. Specialists of the Laboratory for Monitoring the Environment and Climate of the State Committee on Hydrometeorology and the USSR Academy of Sciences participated in the experiment from the USSR. The program of the experiment (observations under the background monitoring program, intercalibration of methods and instruments, mutual training of specialists) was completely carried out. As a result of implementation of the program it was possible to obtain new and interesting data on the background state of the environment in the central European region.

The special exhibit entitled "Environment - Reliable Monitoring" objectively reflects the activity of the State Committee on Hydrometeorology in the field of preservation of the environment. The successes attained in the Soviet Union in this field were possible due to the everyday attention given to these problems by the Party and the government.

CONFERENCES, MEETINGS AND SEMINARS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 125-127

[Article by I. A. Yankovskiy]

[Text] Scientific-methodological seminars were held in 1979 for examining ways to increase the effectiveness and quality of work on monitoring and predicting air contamination, the inventorying of harmful effluent.

Field seminars were held at the Central Asian Scientific Research Institute (Tashkent) during the period 28 May through 2 June, at the Verkhne-Volzhskoye Territorial Administration of the Hydrometeorological Service (Gor'kiy) during the period 1 through 5 October, at the Far Eastern Scientific Research Institute (Vladivostok) during the period 15 through 20 October 1979. These seminars were attended by representatives of 36 administrations of the Hydrometeorological Service and scientific research institutes, a number of organizations of the USSR Health Ministry, planning, industrial and public service organizations, the type of whose activity is related to preservation of the environment. Altogether more than 360 representatives of 160 organizations and departments participated in the seminars.

In opening the seminars, N. N. Aksarin, director of the Central Asian Scientific Research Institute, V. S. Ryazanov, chief of the Verkhne-Volzhskoye Administration of the Hydrometeorological Service and Yu. P. Kovtanyuk, deputy director of the Far Eastern Scientific Research Institute, presented reports on the tasks of the National Service for Observing and Monitoring the Environment (SKZA -- Sluzhba Nablyudeniy i Kontrolya Okruzhayushchey Sredy) and on the tasks of the seminar. The speakers told about the status of work on investigation of contamination of atmospheric air, stated some shortcomings in the activity of network subdivisions of the SKZA and brought attention to the need for further development and improvement in operation of the network for monitoring contamination of the air medium. They demonstrated that strengthening of the relationships between scientific institutes and practical workers and creative cooperation of the personnel of the administrations of the Hydrometeorological Service and scientific research centers is, under present-day conditions, a highly important factor in increasing the quality and ensuring the maximum effectiveness of operation of network subdivisions in the service.

The reports and communications of specialists in the Administration for Observing and Monitoring the Environment (V. V. Chelyukanov), Main Geophysical Observatory (I. A. Yankovskiy, L. R. Son'kin, I. A. Solomatina, E. Yu. Bezuglaya, N. S. Vol'berg, Ye. A. Shaykova, Ya. S. Kanchan, M. N. Zashikhin), USSR Hydrometeorological Center (L. M. Neronova and I. A. Tikhomirova) were devoted to:
-- analysis of the status of definite types of work for studying atmospheric contamination;
-- prospects for the development of means and methods for analysis of contamination of atmospheric air, taking into account the official new "Manual on Monitoring Atmospheric Contamination;"
-- automatic methods for analysis of atmospheric contamination;
-- analysis and generalization of data on the discharge of harmful substances into the atmosphere;
-- organization of work on the inventorying of harmful discharges;
-- examination of norms for the maximum admissible discharges into the atmosphere and implementation of the plan for measures for the introduction of GOST [State Standard] 17.2.3.02-78, a method for evaluating the quality and effectiveness of operation of the network for observing and monitoring atmospheric contamination;
-- status of work for predicting air contamination;
-- introduction of new methods for analysis of air contamination;
-- examination of ways to improve operation of new technical means, such as automatic gas analyzers, computations of the norms for maximum admissible discharges;
-- coordination of schemes for siting and plans for the construction of industrial facilities;
-- computation methods for determining the harmful substances expelled into the atmosphere by industrial sources, proposals on the sequence for monitoring sources of contamination of the air medium.

V. V. Chelyukanov, a specialist of the Administration for Monitoring Contaminants of the State Committee on Hydrometeorology, discussed the problems facing republic and territorial administrations of the Hydrometeorological Service. He noted that the local subdivisions of the SKZA have done much for the development and improvement of monitoring of atmospheric contamination, but can do far more. The most important task of each subdivision is intensifying attention to improvement in the quality of work and increasing its effectiveness. He also emphasized the need for the speediest possible introduction of new gas analysis apparatus and new methods for analysis of air contamination, rapid mastery of new manuals and State Standards. In conclusion the speaker mentioned the need for improving the routine servicing of Party and state agencies and organizations dealing with the national economy.

The communications of specialists of the republic and territorial administrations of the Hydrometeorological Service and scientific research institutes noted the positive results obtained in the course of investigations of atmospheric contamination. These investigations are being made by specialists under the direction of M. Ye. Berlyand with the broad participation of specialists in network subdivisions of the service. A high evaluation was given to the "Method for Evaluating the Quality and Effectiveness of Network Operation," developed on the basis of an analysis of the activity of the existing network and allowance for the requirements set forth in the "Manual on Montitoring of Atmospheric Contamination," as well as GOST [State Standard] 17.2.3.01-77 -- "Preservation of the Environment. Atmosphere. Rules for Monitoring Air Quality of Populated Places" and including use of data from an experimental evaluation of the quality of work performed by specialists of republic (territorial) administrations of the Hydrometeorological Service and scientific research institutes. It is assumed that the use of this method will favor a further improvement in the effectiveness of servicing of the national economy.

An equally important effect can be obtained from the introduction of the "Methodological Instructions on the Prediction of Air Contamination in Cities," also developed at the Main Geophysical Observatory, which were published and disseminated to all administrations of the Hydrometeorological Service and scientific research institutes. This has created the prerequisites for the organization of work on protecting the atmosphere against contamination during periods of dangerous meteorological conditions. It was noted in a report by L. R. Son'kin that at the present time predictions of air contamination are prepared for more than 100 cities in the country. Warnings are transmitted to several hundred industrial, power and transport organizations. It was noted that a good effect was obtained by the specialists of the Central Asian Scientific Research Institute, West Siberian Scientific Research Institute, Azerbaydzhan, Irkutsk, Kazakh, Kirgiz, Krasnoyarsk (Noril'sk), Northern, Northwestern and Northern Caucasus Administrations of the Hydrometeorological Service. The greatest number of cities was supplied with prognostic information on atmospheric contamination by the Ural, Volga Region, Upper Volga, Northern Caucasus Administrations of the Hydrometeorological Service and the Central Asian Scientific Research Institute.

In the seminars an important place was devoted to a discussion of the problems involved in the sequence of development and examination of the norms for maximum admissible levels in the atmosphere and implementation of the plan for measures for introducing GOST [State Standard] 17.2.3.02-78 entitled "Preservation of Nature. Atmosphere. Rules for Setting Admissible Discharges of Harmful Substances by Industrial Enterprises." Revealing the essence of these important steps in the field of improvement of work in the field of scientific expertise, the speakers M. N. Zashikhin and Ya. S. Kanchan emphasized the necessity for both coordination of the schemes for distribution and the plans for construction of industrial facilities and also for computing the norms for maximum admissible discharges.

As indicated in the reports of N. S. Vol'berg and Ye. A. Shaykov, great possibilities for a substantial improvement in the quality of collection of informative data are being afforded by the introduction of the "Manual on Monitoring of Atmospheric Contamination," the new methods and technical means for analysis of the state of the air set forth in this manual. The objective is to make fuller allowance for the peculiarities of analysis of atmospheric air for its content of harmful substances by manual and automatic methods and make more extensive use of already available automatic methods recommended by the Administration for Monitoring Contamination of the State Committee on Hydrometeorology. However, at some administrations of the Hydrometeorological Service and scientific research institutes automatic gas analyzers are being introduced into the practical work of the SKZA very slowly.

Exceptional importance has been given to the organization and development of forms of operational servicing of Party and soviet agencies and organizations servicing the national economy. A good basis for increasing the effectiveness of operational servicing of users is the use of information accumulated in the experience of specialists of a number of administrations of the Hydrometeorological Service in close collaboration with the prognostic agencies of the service (report of I. A. Yankovskiy).

A particular place was devoted to an analysis and generalization of data on the discharge of harmful substances into the atmosphere and also to an examination of the organization of inventorying work. These and other aspects of this problem were covered in the reports of I. I. Solomatina, which were read in the plenary and section sessions.

It was demonstrated in a report by E. Yu. Bezuglaya entitled "Study of Climatic Conditions of Scattering of Impurities in the Atmosphere" that this matter is closely related to the quality and effectiveness of servicing of interested organizations.

D. V. Vinokurova told about the results of work on complex themes related to the study of the climatic conditions of transport and scattering of impurities in the atmosphere and to the development of a method for evaluating the quality of operation of the network for the monitoring of atmospheric contamination. There has been considerable work on the plan for scientific-methodological direction of the network. Among the many forms of information on the state of atmospheric air a leading place at the administrations of the Hydrometeorological Service is occupied by graphic information. Maps, photographs, graphs and figures have been produced and beautifully finalized; these reflect the nature and tendency of atmospheric contamination and illustrate the principles and methods for analysis of the state of the atmosphere.

The resolutions adopted by participants in the seminars formulated specific proposals on the further improvement of operation of network subdivisions of the SKZA and improvement in methodological leadership on the part of the central agencies of the service. Particular attention was devoted to the need for a highly speedy introduction of new technical apparatus, preparation of methodological aids for carrying out expert examination of projects, preparation of a plan for the section "Preservation of the Air Basin" and determination of the discharge of harmful substances into the atmosphere and computation of the maximum admissible discharge levels.

NOTES FROM ABROAD

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 7, Jul 80 pp 127-128

[Article by B. I. Silkin]

[Text] As reported in SCIENCE NEWS, Vol 115, No 24, the American meteorologist W. A. Lyons has carried out an analysis of daytime space photographs taken in June 1979 from aboard meteorological satellites situated in an equatorial orbit at an altitude of about 46,000 km above the earth's surface. On a series of photographs it is possible to distinguish a thick zone of haze which extends over a considerable part of the United States -- from Kentucky to Maryland -- and which then extends 1,300 km to the east of the shore, reaching the central part of the Atlantic Ocean.

The analysis revealed that this haze is a dense air mass saturated with sulfates released into the atmosphere as a result of combustion of great masses of coal at electric power stations in industrial regions of the basin of the Ohio River and New England (the northeastern states in the United States situated along the Atlantic coast of the country). The sulfates are condensation nuclei for the moisture droplets which form the haze and usually reduce visibility by more than half. Such haze can persist in one place for several weeks. The rain "washes away" the haze, but absorbing sulfuric acid, the precipitation acquires toxic and corrosive properties.

The investigations of W. A. Lyons also indicated that within the limits of the haze there is a high ozone concentration. Within this air mass it attains 91 parts per billion, whereas outside this mass it is only 69 parts per billion. As demonstrated by recent studies in the field of agrometeorology, such an ozone concentration is extremely harmful for plantings of soy beans and legumes.

As reported in NATURE, 31 May 1979, and in SCIENCE NEWS, Vol 116, No 1, 1979, the American chemist and meteorologist J. O. Nriagu carried out an investigation of the intensity at which contamination of the earth's air envelope with metals is transpiring at the present time.

According to his conclusions, during the last decade approximately 74 million kilograms of cadmium, 585 million kilograms of copper, 4.3 billion kg of lead, 4.5 million kilograms of nickel and 3.3 billion kilograms of zinc have entered the atmosphere.

The estimate of the quantity of metals entering into air space was made separately for natural sources (eruptions of volcanoes, forest fires, erosion, rising dust particles and salts expelled from the surface of seas and oceans) and for sources associated with man's activity (mining industry, metallurgy, combustion of coal in different fields of production, combustion of firewood and wastes, blowing away of fertilizers, etc.).

As might be expected, human activity was a much more important source of agents contaminating air than natural processes. For example, in 1975 the combustion of liquid fossil type of fuel (including gasoline) led to the entry of 273 million kilograms of lead into the atmosphere. At the same time, eruptions of all the earth's volcanoes, being highly important natural sources of lead, introduced only about 6.4 million kilograms of this metal into the atmosphere.

The high lead content in air space is a relatively new phenomenon. In the decade 1910-1920 its total quantity in the atmosphere did not exceed 493 million kilograms. During the next decade alone it attained 1.1 billion kilograms. J. O. Nriagu attributes such a rapid increase to the large-scale use of automotive vehicles and the appearance of gasoline with lead additives in 1923.

As reported in CHEMICAL WEEK, 21 February 1979, and in THE SCIENCES, Vol 19, No 16, 1979, on the request of the US Environmental Protection Agency a group of TVA specialists, headed by the geochemist and soil scientist G. S. Noggle, carried out an investigation, lasting two years, of the influence exerted on vegetation by sulfur present in the air.

The studies were made in greenhouses into which no air could penetrate from the outside. The inside atmosphere was purified from the sulfur which it usually contains by means of charcoal filters. In addition, in order to determine the quantity of sulfur absorbed by vegetation from the soil and from the air, the radioactive tracer S^{35} was used under the open sky.

It was established that with a reduction in the content of sulfur in the soil the plant increases its absorption from the air medium. Cotton, hay and other plants cultivated in the alkaline soils of the southeastern United States satisfy a considerable part of their needs for sulfur from the air.

With an increase in the combustion of fossil fuel, increasing the sulfur content in the atmosphere, there was an approximate coincidence in time with the decrease in the quantity of fertilizers containing sulfur which is applied to the soil. Sulfur increases the content of chlorophyll in plants and thereby facilitates the processes of photosynthesis and growth. Thus, the dependence of vegetation on chemical substances in the atmosphere has recently increased.

In the opinion of G. S. Noggle, if the present-day efforts for air purification are crowned with success, in the territory of the Tennessee River valley, covering seven states in the United States, there will be a decrease in crop yields by approximately 10%. In order to make up for the loss of sulfur from the air here it will be necessary to apply fertilizer at a cost up to 7.6 million dollars.

Although it still has not been possible to establish a difference between the sulfur "naturally" present in the atmosphere and the sulfur ejected in the course of industrial activity, it is nevertheless clear that the contamination of air by sulfur has some unquestionable positive aspects in addition to negative aspects.

[11-5303]

5303 -END-
CSO: 1864