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SOME PROBLEMS IN THE THEORY OF CLOUD CONDENSATION NUCLEI

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 5-12

[Article by Doctor of Physical and Mathematical Sciences I. P. Mazin, Central Aerological Observatory, submitted for publication 18 Feb 80]

[Text]

Abstract: Some problems in the theory of cloud condensation nuclei (CCN) are defined. It is demonstrated that the effectiveness of CCN is fully determined by their soluble part, which is unambiguously described by the introduced effective radius of the condensation nuclei. The experimental data indicate unsuitability of a power law approximation for describing the real spectrum of CCN. This imposes restrictions on the applicability of the well-known Twomey formula relating the maximum supersaturation and concentration of cloud droplets and the parameters of CCN.

Introduction. A physically clear and graphic exposition of the theory of cloud condensation nuclei (CCN) can be found in a study by L. M. Levin and Yu. S. Sedunov [2, 4]. We will briefly repeat the fundamental points in the theory, introducing some refinements in passing. In contrast to the authors of [2], we will use a natural definition of supersaturation

$$S_0 = \frac{e - E_0}{E_0}. \tag{1}$$

In [2, 4] it is assumed that $S_0 = \frac{e - E_0}{e}, \quad S = \frac{e - E_s}{e}.$

Here e is the actual elasticity of water vapor, E_0 is the saturating elasticity of water vapor over a plane surface.

As in [2], we introduce the value $x = 1/r$. Then

$$\frac{E_s}{E_0} = 1 + Bx - Cx^3, \tag{2}$$

where E_s is the saturating elasticity of vapor over a droplet of solution.

[Here and in the text which follows use is made of the following notations: σ -- coefficient of surface tension (75.6 g/sec^2), ρ_w -- 1 g/cm^3 is water density, $R = 8.314 \cdot 10^7$ (erg/(mol·degree) is the universal gas constant, T is absolute temperature (0°C = 273.15 K).] The B coefficient characterizes the increase in the saturating elasticity of vapor over a droplet with a decrease in its radius.

$$B = \frac{2\sigma M_w}{\rho_w RT} \approx 1.1985 \cdot 10^{-3}\ \mu\text{m}.$$

(with T = 273.15 K).

The coefficient C, in [2] called nucleus activity, characterizes the decrease in the saturating elasticity of vapor over the solution. We will call the parameter

$$r_n = C^{1/3} \tag{3}$$

the effective radius of the condensation nucleus. (Usually r_n differs from the radius of the dry hygroscopic nucleus r_{dry} by not more than tens of percent).

Supersaturation over droplets of solution $S = e - E_s/E_s$ can be related to S_0, and with an accuracy to the second order of magnitude relative to S_0 and ν we find

$$S = \frac{1 + S_0}{1 - (Bx - r_n^3 x^3)} - 1 \approx S_0 - (1 + \nu)(Bx - r_n^3 x^3), \tag{4}$$

where $\nu = \nu(x) = S_0 - (Bx - r_n^3\ x^3)$. In clouds $\nu \ll 1$. [If S is determined as $e - E_s/E_0$, then $\nu = 0$.]

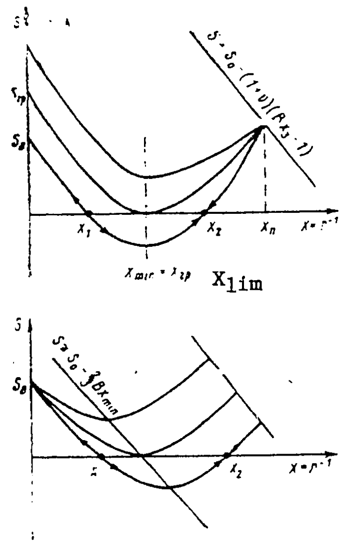

Fig. 1.

Now we will examine the family of curves S(x) for different S_0 and r_n (see Fig. 1). We will assume that equation (4) is correct in the entire range of sizes $x \ll x_3 = r_n^{-1}$. When the radius of the CCN becomes equal to the effective radius ($x = r_n^{-1}$), equation (4) is transformed into a straight-line equation (5), limiting the S(x) curves to the right,

$$S = S_0 - (1+\nu)(Bx_3 - 1). \tag{5}$$

With $S_0 > 0$ there is such a limiting value $r_n = r_{nlim}(S_0)$ at which the S(x) curve touches the x-axis at some point $x_{1im} = r^{-1}_{1im}$. All the droplets forming on CCN, whose effective radius is greater than r_{nlim}, are in air which is supersaturated relative to them ($S(x) > 0$ with any $x \leqslant x_3$) and grow without limit. When $r_n < r_{nlim}$ the S(x) curve twice intersects the x-axis. At the points x_1 and x_2 $S(x) = 0$ and the droplets are in equilibrium with the medium. However, at the point $x_1 = r_1^{-1}$ the equilibrium is unstable, whereas at the point $x_2 = r_2^{-1}$ the equilibrium is stable. This means that with $S_0 > 0$ the droplets forming on the CCN either increase in size without limit ($r > r_1$) or are in a stable state, having the radius $r = r_2$. If the supersaturation S_0 increases, then the points r_1, r_{1im}, r_2 approach one another and are shifted to the right, whereas r_{nlim} decreases. The r_1, r_{1im}, r_2, r_{nlim} values are dependent only on S_0. And this means that with a known supersaturation S_0 the stable water-enveloped nuclei have the radius $r < r_{1im}$ and they were formed on CCN whose effective radius is $r_n \leqslant r_{nlim}$. This made it possible to draw a distinct boundary between cloud droplets and water-enveloped condensation nuclei. Droplets whose radius is $r < r_{1im}$ with a given S_0 must be classified as water-enveloped condensation nuclei, whereas those with $r > r_{1im}$ must be classified as cloud droplets.

With $S_0 < 0$ with any r_n there is only one point of intersection of the S(x) curve with the x-axis corresponding to stable equilibrium. With $S_0 > 0$ the two points (x_1, x_2) appear, which with an increase in S_0 begin to be drawn closer to one another. Finally, with a definite S_0 value for stipulated r_n these points merge into one, that is, $x_1 = x_2 = x_{1im} = r^{-1}_{1im}$.

Thus, water-enveloped nuclei are formed on condensation nuclei whose effective radius is $r_n < r_{nlim}$. However, nuclei with $r_n > r_{nlim}$ are activated and give rise to cloud droplets.

Correlation Between Characteristic Radii of Nuclei and Supersaturation

The correlation between r_{nlim} and r_{1im} with S_0 is found from the conditions

[гр = lim]

$$\begin{cases} \left.\dfrac{dS(x)}{dx}\right|_{x = x_{\text{гр}}} = 0 \\ S(x_{\text{гр}}) = 0. \end{cases} \tag{6}$$

Rejecting small terms of the order of ν, from (4) and (6) we find

$$r_{\text{гр}} = x^{-1}_{\text{гр}} = \left(\frac{3\, r^3_{n_{\text{гр}}}}{B}\right)^{1/2}, \tag{7}$$

$$r_{n_{\text{гр}}} = \frac{B}{3}\left(\frac{2}{S_0}\right)^{2/3} \approx 6{,}352 \cdot 10^{-4}\, S_0^{-2/3} \approx 0{,}013685\, S_{\#}, \tag{8}$$

[The leaving of terms of the order ν in similar expressions in [2, 4] is not legitimate. The relationship $r_{1im} = 2B/3S_0$ was obtained earlier by Mason [3].]

$$r_{\text{гр}} = \frac{2\, B}{3\, S_0} = r_{n_{\text{гр}}}\left(\frac{2}{S_0}\right)^{1/3} = 1{,}2599\, r_{n_{\text{гр}}}\, S_0^{-1/3} \approx 5{,}8479\, r_{n_{\text{гр}}}\, S^{-1/3}. \tag{9}$$

where r is in μm, S_* is supersaturation in percent and S_0 is in fractions of unity. For example, with $S_* = 1\%$ $r_{n\,lim} = 1.37\cdot10^{-2}\,\mu$m, and $r_{lim} \approx 5.85 r_{n\,lim} \approx 0.08$ μm. Since with the rising (cooling) of the air the increase in S_0 takes place gradually, in the first approximation it can be assumed that at any moment in time the condensation nuclei are in a water-enveloped equilibrium state. And this means that all droplets with the radius $r < r_{lim}$ are water-enveloped nuclei and all droplets with $r > r_{lim}$ are growing cloud droplets [2,4]. As a result, in clouds a minimum in the curve of the size distribution of cloud droplets is formed in the region $r \approx r_{lim}$.

Still another useful expression can be derived from equation (4). It was proposed by A. G. Laktionov and used as a basis in his development of an isothermal chamber -- a promising instrument for measuring the distribution of atmospheric condensation nuclei by supersaturations. In these chambers $S_0 = 0$ and the equilibrium radii r_2 are measured. It follows from (4) with $S = S_0 = 0$ that

$$r_2^2 = \frac{r_n^3}{B}. \tag{10}$$

Taking (8) into account, we find that an equilibrium droplet of the radius r_2 grew on a nucleus r_n, active with a supersaturation

$$S_* = 100\,S_0 = \frac{200\,B}{3\sqrt{3}\,r_2} = 4{,}6\cdot10^{-2}\,r_2^{-1}, \tag{11}$$

where r_2 is in μm. Expression (11) somewhat refines the formula used by Laktionov, for which $S_* = 4\cdot10^{-2}\,r_2^{-1}$ [6].

In concluding this section we will compare the critical radius of an insoluble nucleus r_{cr}, active at a given supersaturation, with the critical radius of a soluble nucleus $r_{n\,cr}$, active at this same supersaturation.

$$r_{\text{кр}} = \frac{B}{\ln(eE)} \approx \frac{B}{S_0} = \frac{100\,B}{S_*}. \tag{12}$$

It follows from (12) and (8) that

[kp = cr]

$$\varepsilon = \frac{r_{\text{кр}}}{r_{n_{\text{кр}}}} = \frac{3}{4^{1/3}}\,S_0^{-1/3} \approx 1{,}890\,S_0^{-1/3} \approx 8{,}757\,S_*^{-1/3}. \tag{13}$$

With $S_* = 0.1\%$ $\varepsilon \approx 19$, whereas with $S_* = 1\%$ $\varepsilon \approx 9$, that is, in order for the activity of insoluble nuclei to be comparable to the activity of hygroscopic nuclei their radius must be greater by an order of magnitude (and the mass by three orders of magnitude). Thus, if in an aerosol particle the soluble part is more than 0.1%, it is more important for the formation of droplets than the remaining part. According to data from a number of authors, the mass of hygroscopic particles in the atmosphere is several percent. (From 5 to 30%, according to [1]). Therefore, with assurance we can ~~say of~~ cloud condensation nuclei that they constitute soluble or mixed nuclei and that a factor of real importance is the distribution of CCN by supersaturations, or what is the same, by their effective radii, which are close to the radius of the soluble part of the nuclei. However, the formation of droplets on insoluble (both wettable and nonwettable) particles plays virtually no significant role in the physics of droplet clouds. [At negative temperatures this process can be important from the point of view of formation of ice crystals.]

Spectrum of Distribution of Cloud Condensation Nuclei by Supersaturations

Using (8), we will find the relationship between the distribution of CCN by supersaturations f(S) and the distribution by effective radii $\varphi(r_n)$:

$$\varphi(r_n) = f(S)\left|\frac{dS}{dr_n}\right| = f(1{,}601 \cdot 10^{-3}\, r_n^{-3/2}) \cdot 2{,}4015 \cdot 10^{-3}\, r_n^{-5/2}, \tag{14}$$

where r is in μm, φ is in $cm^{-3} \cdot \mu m^{-1}$. [Here and in the text which follows we will dispense with the subscript * on S_* and by the letter S we will denote supersaturation over a plane surface of pure water in %.]

Now we will turn to the experimental data. It is known that aerosol particles (AP) with radii R from 0.1 to tens of μm are distributed in the atmosphere, in general, in conformity to the Junge law:

$$\Psi(R) = \frac{A}{R^{\alpha}}. \tag{15}$$

If the radius of the soluble part r_s is related to the radius R of the particle by the expression [4]

$$r_s = bR^{\beta}, \tag{16}$$

then, since the fraction of the soluble fraction in the atmosphere decreases with an increase in R, $\beta < 1$. Using (15) and (16), we find that the distribution of AP by the radii of the soluble part will have the form

[c = s]

$$\Psi^*(r_c) = \Psi\left[\left(\frac{r_c}{b}\right)^{1/\beta}\right]\frac{dR}{dr_c} \sim r_c^{-\varkappa}, \tag{17}$$

where

$$-\varkappa = -\frac{\alpha}{\beta} + \left(\frac{1}{\beta} - 1\right) = -\alpha\left[1 + \frac{\alpha - 1}{\alpha}\left(\frac{1}{\beta} - 1\right)\right].$$

Since $\alpha > 1$ (usually $\alpha \approx 4$) and $1/\beta - 1 > 0$, then $\varkappa > \alpha \approx 4$. Thus, with a distribution of atmospheric aerosols in conformity to the Junge law and correctness of the correlation between r_s and R of the type (16) the AP are distributed by the radii of their soluble particles in conformity to a power law with the exponent $\varkappa > 4$.

Now we will turn to the distribution of aerosol particles by supersaturations. According to numerous data [8],

$$f(S) = ckS^{k-1}, \tag{18}$$

where k is usually less than 1. According to (14),

$$\varphi(r_n) \sim r_n^{-\frac{3}{2}k - 1} = r_n^{-(\lambda+1)}. \tag{19}$$

However, the integral distribution has the form

$$N(r_n) = N_1\left(\frac{0{,}1}{r_n}\right)^{\lambda}, \tag{20}$$

where

$$\lambda = \frac{3}{2}k, \quad N_1 = 0{,}05063^k\, c. \tag{21}$$

Table 1

Characteristic Distributions of CCN by Masses of Soluble Part and Effective Condensation Radii*

$m(r)$	r_c мкм r_s µm	r_n мкм µm	N см$^{-3}$	
			background	continent.
$1,00\cdot10^{-16}$	$2,38\cdot10^{-2}$	$1.90\cdot10^{-2}$	42.0	1246
2.37	3.17	2,98	40,7	996
4.63	3.97	3,16	38,5	831
6.00	4.76	3,79	36,3	701
$1.90\cdot10^{-15}$	6.35	5.06	33,4	501
3.70	7.93	6,32	28,6	365
6.40	9.52	7,59	21,9	270
$1.25\cdot10^{-14}$	$1.19\cdot10^{-1}$	9,48	12.4	173
2.96	1.59	$1.26\cdot10^{-1}$	4.78	85,8
5.79	1.98	1,58	2,78	42,1
$2.00\cdot10^{-13}$	3.00	2,39	1.68	22,3
3.18	3.60	2.79	1,06	11,6
4.74	4.00	3.19	0,710	5.82
9.26	5.00	3,98	0,370	1,67
$1,60\cdot10^{-12}$	6.00	4,78	0,200	0,63
2.54	7,00	5,58	0,113	0,29
3,79	8.00	6,37	0,071	0,14
5,40	9,00	7.17	0,046	0,077
7,41	$1.00\cdot10^{0}$	7,96	0,029	0,039
$1,28\cdot10^{11}$	1,20	9,56	0,011	0,011
2,03	1.40	$1,11\cdot10^{0}$	0.0035	0.0032

*Data on the concentration N were taken from [7]; the r_s and r_n columns were computed from the conditions $r_n = 0.7967\ r_s$, $\rho = 1.77$ g/cm^3, $r_s = (3m/4\pi\rho)^{1/3} = 0.512837\ m^{1/3}$. The background distribution is recommended for the modeling of sea clouds.

Taking into account that usually $k<1$, we have $\lambda + 1 < 2.5$, which, it would seem, clearly contradicts the earlier determined value $\chi > 4$. (We recall that $r_n \approx r_s$). However, this contradiction is only apparent. The fact is that formula (15), from which (17) follows, describes aerosols with a radius $R > 0.1\mu$m; however, formula (20) is correct with $S > 0.1\%$, that is, $r_n < 0.064\mu$m. The conclusion can therefore be drawn that either in particles with $R > 0.1\mu$m there is a soluble part whose radius r_n emerges beyond this limit, that is, greater than 0.064μm, and then formula (17) describes another part of the spectrum, or in the considered region the relationship between r_n and R is not described by a dependence of the type (16). It is possible that both to some degree are correct and thus the real spectrum of effective condensation radii in the range from 0.01 to 1μm and above is evidently not described by a power law.

Up to the present time the overwhelming majority of the measurements of the distribution of CCN by supersaturations have been carried out using diffusion chambers in the region $S > 0.1\%$, that is, in the region $r_n < 0.06\mu$m. In this field the measurements were satisfactorily described by empirical expression (18). However, the very first experiments of Laktionov [6], who carried measurements into the region of low supersaturations (to $S \approx 0.025\%$), in actuality revealed that formula (18) with k = const no longer is suitable for describing the spectrum of CCN in the entire measured range. According to the data in [4], in the region $0.025 \leqslant S \leqslant 0.16$ k on the average is approximately equal to 1.6, that is, twice as great as the k value for the range $0.16 \leqslant S \leqslant 1\%$.

Meszaros [7], who was greatly involved in measurements of the concentration of CCN in the atmosphere, generalized the accumulated experimental data and cited a table of the mean distribution of CCN with respect to the mass of their hygroscopic part, assuming that the solid part of the nucleus is ammonium sulfate $(NH)_2SO_4$. In reproducing the Meszaros table, we added columns for r_s and r_n, from which it can be seen that the distribution which he proposed covers the range r_n from 0.019 to 1.1 μm and that it is essentially not a power law distribution. The value d lg N/d lg r_n changes in this distribution by almost an order of magnitude with a change in r_n from 0.02 to 1μm.

Applicability of Power Law Approximation of Spectrum of CCN With Respect to Supersaturations

A representation of the CCN spectrum in the form

$$N = cS^k, \tag{22}$$

where N is the total concentration of nuclei active with the supersaturation S, has come into wide use in cloud physics. The well-known expressions derived by Twomey [9] for determining the maximum supersaturation in a portion of air rising with the stipulated velocity u are based specifically on this distribution. And knowing S_{max}, using (22) it is also possible to determine the concentration of forming droplets $N_k = cS_{max}$. Taking into account the importance of expression (22), Braham [5] proposed that data on the CCN spectra be represented in (c, k)-space and, using the Twomey expressions, a diagram be constructed making it possible to find S_{max} and N_k easily from known c and k.

However, we should note that expression (22), as emphasized above, approximates the experimental data in a rather narrow range of sizes. If it is taken into account that in clouds $S < 1$-2%, this range of change in S from 0.1% to approximately 2% is equivalent to a change in r_n from 0.01 to 0.06μm. At the same time the Twomey formulas and the Braham diagram were constructed on the assumption that (22) is correct in the entire range of sizes, and what is especially important, in the field of smaller S, up to S = 0, that is, to $r_n = \infty$. The question arises: possibly can such a substantial increase in the concentration of large nuclei with $r_n > 0.06\mu$m given by extrapolation of formula (22) into regions of small S distort the results of determination of S_{max} and N? Are the Twomey expressions and the Braham diagram applicable?

It is evident that if the flux of water vapor onto droplets growing on nuclei with $r_n > 0.06\mu$m is small in comparison with the flux of vapor onto droplets forming on smaller nuclei the Twomey expressions are applicable. If the flux is not small, the expressions derived in such a way must, as a minimum, be examined attentively.

In order to solve this problem we recall that the equilibrium radius r_{equil} of the droplets at 100% humidity is related to r_n by the expression

$$r_{equil} = B^{-1/2} r_n^{3/2}. \quad (23)$$

And if initially the CCN were distributed in conformity to the law $\varphi(r_n) \sim r_n^{-3/2k-1}$, then, upon attaining a 100% humidity, the water-enveloped nuclei will be distributed in conformity to the law

$$[p = equil] \qquad \varphi^*(r_p) \sim r_p^{-\left(\frac{3}{2}k+1\right)\frac{2}{3}} \frac{dr_n}{dr_p} \sim r_p^{-k-1}. \quad (24)$$

In the first approximation it can be assumed that the vapor flux on a growing droplet is proportional to its radius. Thus, in the estimates it can be assumed that the total flux of vapor on droplets of a radius greater than r_0 is proportional to

$$[p = equil] \qquad \int_{r_0}^{r_{max}} r_p \varphi^*(r_p)\, dr_p \sim r_p^{-k+1} \Big|_{r_0}^{r_{max}}. \quad (25)$$

According to experimental data, k varies from 0.3 to 1.2 with a mean value of about 0.5 [5]. Accordingly, with a distribution of droplets by radii of the effective condensation nuclei described by (19) the flux of vapor, in any case during the initial period of cloud formation, proceeds primarily on large droplets. It can be postulated that if the distribution (19), obtained from the distribution (22), substantially exceeds the concentration of large condensation nuclei, this can lead to an appreciable exaggeration of the computed fluxes of vapor and the understating of S_{max} and N_k associated with this. With an increase in the size of the droplets (due to rapid growth of the small droplets) the noted effect will be weakened. How should one regard the estimates of S_{max} and N_k made using the Twomey formulas or the Braham diagram? It appears that the problem is not that difficult. A change in k in a relatively broad range, such as from 0.3 to 1.6, changes the N_k concentration by a factor not greater than 1.5-2 and S_{max} even less. Since usually k does not vary so greatly, the real deviations of S_{max} and N_k from the estimates obtained using the Twomey formula must not exceed tens of percent. Thus, despite the sseming unsatisfactory approximation of the CCN distribution by formula (22), the estimates of S_{max} and N_k based on it are entirely acceptable.

It appears that (N_1, λ)-space, transformation to which from (c, k)-space is readily accomplished using formulas (21), can become equally convenient for a practical comparison of the CCN spectra with one another and can be even more graphic. It seems that the N_1 value, equal to the concentration of CCN, whose effective radius is greater than 0.1μm, is more graphic in comparison with the c value.

Summary

The numerical modeling of the processes transpiring in clouds is becoming an increasingly effective means for investigating cloud physics. The results in some cases can be essentially dependent on the initial spectrum of CCN introduced into the model. In order to be able to understand correctly the reasons for the discrepancy in the results obtained in different models (for example, one-, two-, three-dimensional, etc.) it is necessary that each model be tested for a standard, generally accepted distribution of CCN. The distributions of CCN proposed by E. Meszaros and cited in the table for sea and continental clouds are entirely acceptable for this purpose.

Taking advantage of the opportunity, I express appreciation to the participants in the seminar on cloud physics at the Central Aerological Observatory for useful discussions.

BIBLIOGRAPHY

1. Aleksandrov, E. L., Yasevich, N. P., "Determination of the Weight Concentration of Atmospheric Aerosols and Their Content of Soluble Substances," PROCEEDINGS OF THE VIII INTERN. CONF. ON NUCLEATION, Gidrometeoizdat, Moscow, 1975.

2. Levin, L. M., Sedunov, Yu. S., "Some Problems in the Theory of Atmospheric Condensation Nuclei," DOKLADY AN SSSR (Reports of the USSR Academy of Sciences), Vol 170, No 1, 1966.

3. Mason, B. J., FIZIKA OBLAKOV (Cloud Physics), translated from English, Leningrad, Gidrometeoizdat, 1961.

4. Sedunov, Yu. S., FIZIKA OBRAZOVANIYA ZHIDKOKAPEL'NOY FAZY V ATMOSFERE (Physics of Formation of the Liquid-Drop Phase in the Atmosphere), Leningrad, Gidrometeoizdat, 1972.

5. Braham, R. R., "CCN Spectra in c-k Space," J. ATMOS. SCI., Vol 33, 1976.

6. Laktionov, A.G., "Spectra of Cloud Condensation Nuclei in the Supersaturation Range 0.02-1%," PROC. OF THE VIII INTERN. CONFER. ON NUCLEATION, Gidrometeoizdat, Moscow, 1975.

7. Meszaros, E., "Present Status of Our Knowledge on the Atmospheric Condensation Nuclei," VOPROSY FIZIKI OBLAKOV (Problems in Cloud Physics), Leningrad, Gidrometeoizdat, 1978.

8. Pruppacher, H. R., Klett, J. D., MICROPHYSICS OF CLOUDS AND PRECIPITATION, D. Rudel Publishing Company, Dordrecht, Holland, 1978.

9. Twomey, S., "The Nuclei cf Natural Cloud Formation. Part 1. The Chemical Diffusion Method and its Application to Atmospheric Nuclei," GEOFIS. PURA APPL., 43, 1959.

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COMPUTATION OF WAVE FLUCTUATIONS OF ATMOSPHERIC PRESSURE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 13-19

[Article by V. V. Simonov, Main Geophysical Observatory, submitted for publication 4 Feb 80]

[Text]

Abstract: The author clarifies the possibility of computing pressure on an undulating boundary directly from the first and third equations of motion. An analysis of numerical experiments is given for the purpose of finding a preferable formula.

Sources [2,3] give a formulation of the problem and some results of computations of the structure of a two-dimensional stratified turbulent flow of fluid over a chain of monochromatic waves with the amplitude a, length L, velocity c_{f1} and relief $\zeta(x, t)$. A fourth-degree equation for the stream function ψ and a second-degree equation for turbulent energy b are considered. The system is closed by the generalized Karman hypothesis for the turbulence scale. The problem is solved in curvilinear coordinates ξ, η moving with the velocity c_{f1}, related to $\zeta(x)$ and the Cartesian coordinates x, z by the expressions

$$\xi = x; \quad \eta = \frac{z - \zeta(x)}{1 - \zeta(x)}. \tag{1}$$

Here and in the text which follows we use a dimensionless form of writing for which, for the sake of simplicity, no special notations are introduced. As the velocity scale we used u_*^0 -- some characteristic dynamic velocity equal in this case to the dynamic velocity at the upper boundary of the wave sublayer h; the horizontal coordinate is normalized to L; in all the remaining cases the length scale is h.

The transformation

$$\eta = z - \zeta(x), \tag{2}$$

is simpler and more convenient than (1), used, for example, in [1]. However, the transformation (1) more completely reflects the considered formulation of the problem. The thickness of the wave sublayer, at whose upper boundary the induced oscillations attenuate, is assumed to be limited and does not exceed the limits of the air surface layer. The stipulation of boundary conditions at a relatively low height does not make the use of (2) very constructive. In "straightening" the lower boundary, this replacement "curves" the upper boundary near which the vertical gradients of different characteristics are still quite great. The transformation (2) is effective in analytical methods for solution and formulation of boundary conditions with $z \to \infty$ or in an examination of the entire atmospheric boundary layer, in whose upper part there is attenuation of all the perturbations caused by the underlying surface. In computing the pressure distribution in the wave, which is closely associated with vorticity and its gradients, in the formulation of the problem there is a need for greater accuracy and a minimum of different kinds of simplifications. Otherwise such a characteristic important in the problem of wave generation as the averaged flux of momentum at the discontinuity F_{dis} caused by pressure forces will be computed with a substantial error. Even small changes in the relationship of phases between pressure and the wave surface can lead to a change in the sign of F_{dis}, not to mention the quantitative results. But the physically obvious and simply realized boundary conditions are more easily formulated for ψ than for ω. Accordingly, the formulation of the problem does not contain equations for vorticity, but only the stream function is considered, despite the fourth degree of the corresponding equation.

The system of finite-difference equations is solved by the method of successive Gauss-Seidel displacements. The formula for successive displacements for computing the stream function contains a whole series of coefficients at the internal points of grid intersection which are computed in advance in [2, 3] in the form of two-dimensional masses of data. For example, one of the intermediate factors in this formula has the form

$$M=2\,\psi_{i-1,\,j}-\psi_{i-2,\,j}+a_j\,(b_1\,V^1_{i+1,\,j}+b_2\,V^2_{i-1,\,j}).$$

An error was introduced in [2, 3], namely the coefficients V^1 and V^2 were taken at a point of grid intersection with the indices i, j.

In contrast to [2, 3] we changed the grid in the upper part of the wave sublayer, which in the last five intervals now became uniform with the interval 0.1. Some approximation expressions were refined. In particular, the derivatives $\partial\psi/\partial\eta$ and $\partial^2\psi/\partial\xi\partial\eta$, necessary for computing the transformation term in the equation for the balance of turbulent energy, are now determined using formulas with a second degree of accuracy, and not the first, as in [2, 3].

It is not the distribution of total pressure which is of fundamental interest, but its deviation from values at some boundary or from some point. We will examine the deviation of pressure from the vertical $\xi = 0$, that is, $\Delta p(\xi,\eta) = p(\xi,\eta) - p(0,\eta)$. In particular, with $\eta = 0$, which will be indicated by the subscript "π," we will be concerned with the values

$$\Delta p(\xi,\,0)\equiv\Delta p_\pi=p\,(\xi,\,0)-p\,(0,\,0). \tag{3}$$

Henceforth, a similar notation Δf_π will also be used for other functions.

By definition $F_{dis} = D\langle p_\pi\xi'\rangle = D\langle\Delta p_\pi\zeta'\rangle$; in our case Δp includes both the change in external pressure in the wave and its wave oscillations. The symbols "$\langle\ \rangle$" denote averaging along the length of the wave. Beginning, evidently, with the studies of O. Phillips, including in [2, 3], the pressure fluctuations with $\eta = 0$ are computed using the formula

$$\Delta p_{\pi} = p(\xi,\ 0) - \langle p(\xi,\ 0)\rangle. \tag{4}$$

Since the problem is solved through the stream function, the external pressure field is automatically taken into account. In principle, when making computations using (4), an error is allowed which is associated with the failure to take into account that at a periodic boundary total pressure is not a periodic function. However, a transformation to (3) was made not only for this reason, but also so as to obtain additional information on the accuracy of pressure computations made using the first and third equations of motion.

We introduce the notation $Q = Duw - \tau$ and we write the first and third equations of motion in the form

$$\frac{\partial p}{\partial \xi} = n_0 \frac{\partial}{\partial \eta} (p + u^2 + \sigma_1 b) - \frac{\partial}{\partial \xi} (u^2 + \sigma_1 b) - \frac{1}{\gamma D} \frac{\partial Q}{\partial \eta}; \tag{5}$$

$$\frac{\partial p}{\partial \eta} = -\frac{\gamma}{\mathrm{Fr}} - \gamma D \frac{\partial Q}{\partial \xi} + \gamma D n_0 \frac{\partial Q}{\partial \eta} - \frac{\partial}{\partial \eta} (D^2 w^2 + \sigma_3 b). \tag{6}$$

Here
$$\mathrm{Fr} = \frac{(u_*^0)^2}{gh}$$
is the Froude number; σ_1 and σ_3 are the components of the tensor relating the normal Reynolds stresses and turbulent energy.

We obtain Δp_{π} from the first equation of motion, for which we substitute (6) into (5) and integrate the derived expression for ξ. Recalling that in our formulation $\zeta' = 0$ with $\xi = 0$, for the case of isotropic turbulence ($\sigma_1 = \sigma_3$) and the level $\eta = 0$, after some transformations we obtain

$$\Delta p_{\pi}^{1} = -\Delta u_{\pi}^{2} - \sigma_1 \Delta b_{\pi} - \frac{\Delta \zeta}{\mathrm{Fr}} - D\zeta' Q_{\pi} + D \int_0^{\xi} Q \zeta' \, d\xi + \int_0^{\xi} \frac{1}{\gamma} \frac{\partial \varphi_1}{\partial \eta} d\xi + \frac{1}{D} \int_0^{\xi} \frac{f_1}{\gamma} \frac{\partial \tau}{\partial \eta} d\xi; \tag{7}$$

$$\varphi_1 = (u\zeta' - w)(u + D^2 w \zeta');\ f_1 = 1 - (D\zeta')^2.$$

The superscript 1 on Δp_{π} shows that in computing pressure the first equation of motion was used as the fundamental equation.

Numerical experiments have shown that the contribution of the term containing $\partial \varphi_1 / \partial \eta$ to Δp_{π}^{1} is small. However, these small changes can exert an effect on the F_{dis}^{1} value. The values of the parameter are cited below

$$F_{dis\,\varphi}^{1} = D \langle \zeta' \int_0^{\xi} \frac{1}{\gamma} \frac{\partial \varphi_1}{\partial \eta} d\xi \rangle$$

and its relative influence R in percent. It can be seen that the neglecting of this term is justified only in definite situations.

c_Φ	0	0	0	G		$-G$
δ	0,04	0,08	0,12		0,08	
F^1_p	0,22	0,22	0,37	−0,65		9,04
F^3_p	0,21	0,90	2,38	−0,08		2,09
F^a_p	0,46	2,20	6,90	−0,24		9,50
$F^1_{p\varphi}$	−0,03	−0,14	−0,31	−0,02		−0,10
R	16	63	83	4		1

[Φ = f1; p = dis]

Using the third equation of motion it is also possible to obtain the pressure distribution in the wave Δp^3_π. First from (6) we find

$$p^3(\xi, \eta) - p^3(\xi, 0) = \varphi(\xi, \eta) \equiv -\frac{\gamma}{\mathrm{Fr}}\eta - D\frac{\partial}{\partial\xi}\int_0^\eta \gamma Q d\eta + D\zeta'[(1-\eta)Q(\xi, \eta) - Q(\xi, 0)] - D^2[w^2(\xi, \eta) - w^2(\xi, 0)] - \sigma_0[b(\xi, \eta) - b(\xi, 0)]. \tag{8}$$

Then from (8) we determine $\varphi(\xi, 1)$, $\varphi(0, 1)$, after which we obtain

$$\Delta p^3_\pi = \Delta p^3(\xi, 1) - \Delta\varphi(\xi, 1). \tag{9}$$

The $\Delta\varphi(\xi, 1)$ value in (9) is computed from (8). Different methods can be used in an approach to determination of the change in external pressure $\Delta p^3(\xi, 1) \equiv \Delta p_\infty$. If, however, we limit ourselves only to the wave sublayer and consider the equation

$$\frac{\partial p_\infty}{\partial x} = \frac{1}{D}\frac{\partial \tau}{\partial z}, \tag{10}$$

to be correct for the undisturbed flow over it, then Δp_∞ can be evaluated using additional considerations on the possible changes in shearing stress within the limits of the wave sublayer.

An attempt was made to determine Δp_∞ not from (10), but from equation (5), written for the level $\eta = 1$,

$$\frac{\partial p}{\partial \xi} = -\frac{1}{\gamma D}\frac{\partial Q}{\partial \eta}. \tag{11}$$

Using (11) we find

$$\Delta p_\infty = -\frac{1}{D}\int_0^\xi \frac{1}{\gamma}\frac{\partial Q}{\partial \eta} d\xi. \tag{12}$$

Computations on the basis of (12) give for Δp_∞ values which are substantially different from those which could be expected, although with δ = 0 (δ is wave steepness) entirely reasonable results are obtained. For example, with δ = 0.08

and $c_{f1} = 0$ we obtained $\Delta p_\infty (\xi = 1) = -3.16$, whereas with $\delta = 0$ computations using (12) gave -0.078. Since the real Δp_∞ values are small in comparison with other terms in formula (9), except in the neighborhood of the point $\xi = 1$, where $\Delta p^3_\pi = \Delta p_\infty$, computations of pressure using the third equation of motion are carried out without allowance for Δp_∞. We feel that this causes a lesser error than allowance for Δp_∞ with the use of formula (12).

Using the same simplifications as in the derivation of (7), from (8)-(9) we find

$$\Delta p^3_\pi = -\frac{\Delta \zeta}{\mathrm{Fr}} + \Delta N - D\zeta' \tau_\pi - c_3 \Delta b_\pi; \qquad (13)$$

$$N(\xi) = D \frac{\partial}{\partial \xi} \int_0^1 \tau Q \, d\eta.$$

It is not very clear for what reason there was an appreciable deterioration of convergence of the iteration process. Approximately 20% of the total time expenditures are on computation of the term

$$2n_0 \frac{\partial^2 \tau}{\partial \xi \, \partial \eta}$$

in the equation for ψ. But allowance for it, together with the other above-mentioned refinements and corrections, introduces significant changes into the behavior of many characteristics. As an illustration of this we can use the results of computations of shearing stress with $\eta = 0$ shown in Fig. 1 for the case $\delta = 0.08$ and $c_{f1} = 0$ obtained in [3] and in this study. At the same time, the nature of pressure distribution in the wave remained as before, although the quantitative changes are extremely significant. These results are also shown in Fig. 1. As a comparison, p 13 gives a line denoted F^a_{dis} which contains the values of the mean flux of momentum to the waves obtained in [3] from the third equation of motion.

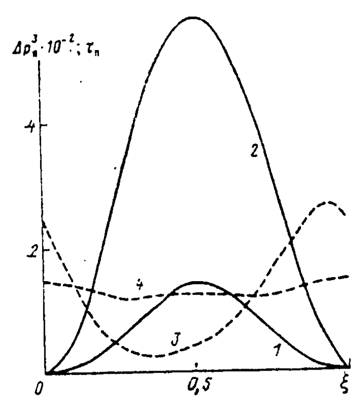

Fig. 1. Distribution of Δp^3_π (1, 2) and τ_π (3, 4) according to the results in this study (1, 3) and from [3] (2, 4). 0.08; $c_{f1} = 0$.

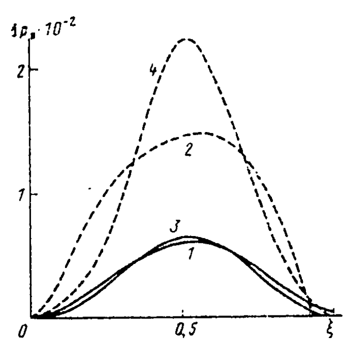

Fig. 2. Distribution of Δp^1_π (1, 2) and Δp^3_π (3, 4) with $c_{f1} = 0$ and $\delta = 0.04$ (1, 3) and $\delta = 0.12$ (2, 4).

Expressions (7) and (13) are formally equally justified. Figure 2 shows Δp_{π}^{1} and Δp_{π}^{3} for $c_{f1} = 0$ and $\delta = 0.04$ and 0.12, and on p 13 — the corresponding F_{dis} values. Available experimental data do not make it possible without reservation to give preference to any of the formulas. The solution of this problem is related to a great extent to the accuracy of computation of the fundamental terms in these formulas. The last term in (7) is decisive, to be more precise, the integral of $\partial\tau/\partial\eta$ with $\eta = 0$, since $\gamma \simeq f_1 \simeq 1$. But the results of computation of the derivative of the flux of momentum with $\eta = 0$ are highly dependent on the method for computing the flux itself at this boundary. The table, for $\delta = 0.08$ and $c_{f1} = 0$, gives the values τ_{π}^{for}, computed using a 4-point formula, and τ_{π}^{ex}, found by extrapolation. The table also gives the values of uniformly obtained $\partial\tau/\partial\eta$ with $\eta = 0$ with the use of τ_{π}^{for} and τ_{π}^{ex}. It can be seen that not only the value, but also the sign of the derivative are dependent on the method for computing τ_{π}. Accordingly, both the pressure distribution in the wave (fifth and sixth lines in the table) and the resistance will be substantially different.

Our numerical experiments indicated that the values of the derivatives at the boundary, obtained by use of the formula and by extrapolation, differ little from one another. The choice of the turbulence coefficient in the approximation formula for computing the fluxes is important. It is common to use the turbulence coefficient obtained as the sum $c_1k_1 + c_2k_2$, where the subscript on k indicates the number of the grid point of intersection in the vertical direction. Assigning c_i different values from 0 to 1, the value of the flux can change substantially. The τ_{π}^{for} values were determined with $c_1 = 1$ and $c_2 = 0$. On the basis of a comparison of the results of computations, and not only those cited in the table, in this study all the turbulent fluxes at the boundaries were determined by extrapolation using the three grid points of intersection adjacent to the boundary.

In formula (13) the main term is ΔN, which is considerably less sensitive to the numerical values τ_{π}. The values Δp_{π}^{3} computed from (13) with τ_{π}^{for} and τ_{π}^{ex} coincided with an accuracy to the third place. Therefore, the table gives only one line Δp_{π}^{3}.

The computed value of the pressure decrease in the wave using formula (7), with $\delta = 0.08$ and $c_{f1} = 0$, was equal to -9.82. After comparing the figures -9.82, -3.16 and -0.078, it must be concluded that different types of approximation and computation errors lead to greater errors in computing pressure on the basis of the first equation of motion than when Δp_{∞} in formula (13) is not taken into account. The sources and magnitude of the computation errors remain the very same if formula (4) instead of (3) is taken as a basis.

Still another advantage of formula (13) is that it was obtained without use of the first equation of motion, if the term Δp_{∞} is not taken into account, which in an extreme case can be taken into account, assuming $\Delta p_{\infty} = -q\,\xi$, where q is evaluated from the thickness of the wave sublayer and has a value of about 0.1. The only thing which does not support (13) is the presence in the main term ΔN of a vertical integral whose upper limit, that is, the thickness of the wave sublayer, is stipulated quite subjectively.

The following circumstance is of more than a little importance. In order to obtain reliable Δp_{π}^{3} values it is necessary to have a lesser number of iterations than for Δp_{π}^{1}. This is particularly noticeable with respect to the F_{dis} values. For example, for the case $\delta = 0.12$ and $c_{f1} = 0$ after 5,000, 10,000 and 15,000 iterations the F_{dis}^{1} values were equal to -2.96, -0.33, 0.37 respectively. At the same time the similar F_{dis}^{3} were 2.13, 2.35, 2.38.

Taking into account everything stated above, the following conclusions can be drawn. First, with the inclusion of pressure among the sought-for characteristics it is desirable to use one fourth-degree equation for the stream function instead of a system of two equations in the variables ψ, ω and especially a system of three equations in the variables u, p. Second, when it is necessary to compute the pressure distribution the use of the third equation of motion is more justified.

	ε										
	0	0,1	0,2	0,3	0,4	0,5	0,6	0,7	0,8	0,9	1,0
τ_n^{Φ}	0,88	0,52	0,23	0,12	0,10	0,14	0,24	0,40	0,65	0,91	0,88
$\tau_n^{Э}$	2,40	1,38	0,58	0,29	0,27	0,40	0,69	1,17	1,88	2,56	2,40
$\frac{\partial \tau^{\Phi}}{\partial r_i} \cdot 10^{-3}$	8,25	4,97	2,24	1,16	1,03	1,46	2,40	3,95	6,30	8,63	8,25
$\frac{\partial \tau^{Э}}{\partial r_i} \cdot 10^{-3}$	−0,04	0,27	0,32	0,22	0,13	0,05	−0,06	−0,24	−0,40	−0,36	−0,04
$\Delta p_n^{1\Phi} \cdot 10^{-2}$	0	6,93	10,4	11,9	12,9	14,1	15,9	18,8	23,6	31,0	39,9
$\Delta p_n^{1Э} \cdot 10^{-2}$	0	0,18	0,52	0,80	0,98	1,07	1,05	0,88	0,54	0,11	−0,10
$\Delta p_n^{3} \cdot 10^{-2}$	0	0,09	0,34	0,74	1,18	1,42	1,30	0,90	0,46	0,13	0

[$\pi = \pi$; $\tilde{\Phi}$ = for(mula); $Э$ = extrap(olation)]

BIBLIOGRAPHY

1. Makin, V. K., Chalikov, D. V., "Numerical Modeling of Wind Waves," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 10, 1979.

2. Simonov, V. V., "Turbulent Flow Over an Undulating Boundary," TRUDY GGO (Transactions of the Main Geophysical Observatory), No 423, 1979.

3. Simonov, V. V., "Some Results of Computation of the Structure of a Turbulent Flow Over an Undulating Boundary," TRUDY GGO, No 423, 1979.

4. Phillips, O. M., DINAMIKA VERKHNEGO SLOYA OKEANA (Dynamics of the Upper Layer of the Ocean), translated from English, Moscow, Mir, 1969.

5. Gosmen, A. D., Pan, V. M., et al., CHISLENNYYE METODY ISSLEDOVANIYA TECHENIY VYAZKOY ZHIDKOSTI (Numerical Methods for Investigation of Flows of a Viscous Fluid), Moscow, Mir, 1972.

6. Banner, M. L., Phillips, O. M., "On the Incipent Breaking of Small-Scale Waves," J. FLUID MECH., Vol 65, Part 4, 1974.

7. Barnett, T. P., Kenyon, K. E., "Recent Advances in the Study of Wind Waves," REPORTS ON PROGRESS IN PHYSICS, Vol 38, No 6, 1975.

8. Benjamin, T. B., "Shearing Flow Over a Wavy Boundary," J. FLUID MECH., Vol 6, Pt 2, 1959.

9. Gent, P. R., Taylor, P. A., "A Note on 'Separation' Over Short Wind Waves," BOUND.-LAYER METEOROL., Vol 11, No 1, 1977.

10. Gent, P. R., Taylor, P. A., "A Numerical Model of the Air Flow Above Water Waves," J. FLUID MECH., Vol 77, Pt 1, 1976.

11. Taylor, P. A., "Some Numerical Studies of Surface Boundary-Layer Flow Above Gentle Topography," BOUND.-LAYER METEOROL., Vol 11, No 4, 1977.

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SOME METHODS FOR EVALUATING THE PARAMETERS OF CORRELATION FUNCTIONS OF METEOROLOGICAL FIELDS

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[Article by Candidate of Technical Sciences V. L. Savanov and T. A. Yarygina, Moscow Power Institute, submitted for publication 18 Jan 80]

[Text]

Abstract: The article gives a validation of one of the possible approaches to solution of the problem of evaluating the parameters of the correlation function of a random meteorological field which essentially involves finding evaluations of the coefficients of a regression model when there are observation correlated errors. The authors propose several methods for evaluating the correlation function of a random (in a general case nonisotropic and nonuniform) field. The choice of the method is dependent on whether the correlation function is linear or nonlinear with respect to the parameters and on the availability of a priori information concerning the distribution law or on the central moment function of the fourth order of the investigated field. Expressions are derived making it possible to evaluate the accuracy of the proposed methods.

Introduction. A factor of great importance for solving problems in meteorology and oceanography is finding methods for describing the spatial-temporal variability of parameters characterizing the different properties of the atmosphere and ocean. These parameters form fields (temperature, pressure, humidity, etc.) which, as is well known, in a general case are nonstationary relative to the time coordinate, nonuniform and nonisotropic relative to the space coordinates. One of the principal problems arising in study of random meteorological and oceanographic fields is an experimental determination of their statistical characteristics. For a complete statistical description of spatial-temporal random fields it is necessary to know the multidimensional distribution functions, but in actual practice it is common to limit the examination to simpler characteristics of the random fields -- mathematical expectations and covariation or correlation functions.

A correlation analysis is one of the most important methods for the study of random fields. A knowledge of the covariation function (CF) makes it possible to form a general idea concerning the structure and properties of meteorological fields, solve many applied problems, including the objective analysis of the investigated fields [2, 5, 6], solve problems involved in rationalization of the network of measuring stations in space and the optimum discretization of measurements in time [3, 4, 10].

An experimental determination of the CF of the fields of meteorological elements usually essentially involves computations of evaluations of the parameters of a CF of a stipulated form from the readings of one or more independent field series. The methods for determining the CF of a random field used at the present time assume satisfaction of a series of conditions which in practical situations frequently are violated. Thus, in the stipulation of the readings of several field series it is usually required that these readings for all series correspond to one and the same set of field arguments [6]. There are methods applicable in cases when there are gaps in the series of observations [7], but even they can be employed effectively only when there are relatively few gaps and the arguments coincide for most of the readings of different series.

However, if the readings for only one field series are stipulated, the limitations become still more rigorous. In this case the developed methods are applicable only to isotropic random fields and are rigorously valid with stipulation of the readings at the points of intersection of a uniform grid of the values of the argument. Modifications of these methods with a nonuniform network essentially involving averaging of the initial readings within the limits of some regions (gradations) of the values of the argument [2, 6] have been inadequately formalized and lead to the appearance of systematic errors whose value usually cannot be evaluated. The studies of R. L. Kagan [8] are devoted to investigation of a number of problems involved in evaluation of the accuracy in computing the CF by the gradations method.

Below we present one approach to determination of the CF of a random field making it possible to propose a number of methods for its evaluation free of the abovementioned shortcomings. We will examine the problem of evaluating the CF with stipulation of the readings of only one series for the random field. Such a formulation of the problem corresponds well to the practical situations arising in an investigation of meteorological and oceanographic fields. In addition, the results are generalized for the case of several series [11].

Formulation of problem. Assume that $F(u)$, where $u = (u_1,\dots,u_k)$ is a k-dimensional argument, denotes the investigated scalar random field. For problems in meteorology and oceanography in a general case $u = (x, y, z, t)$, where x, y, z are the space coordinates, t is time. It is assumed that in a closed limited region of $U \in R^k$ values of the argument the mathematical expectation (ME) of the field is equal to zero, that is, $E[F(u)] = 0$ with $u \in U$ (this means that the determined field component is absent or first excluded), whereas the CF $K_F(u, s)$ for the field is known with an accuracy to m unknown parameters:

$$K_F(\mathbf{u}, \mathbf{s}) = E[F(\mathbf{u})\, F(\mathbf{s})] = k(\mathbf{u}, \mathbf{s}, \mathbf{b}), \quad \mathbf{u}, \mathbf{s} \in U. \qquad (1)$$

where $b = (b_1,\dots,b_m)^T$ is the vector of the unknown parameters ($b \in R^n$), $k(u,s,b)$ is a function of a known type.

Such a formulation is entirely justified since for most meteorological elements the type of CF is known approximately on the basis of data accumulated in the course of observations over a series of years. In a special case when the field is uniform k(u,s,b) is dependent on the difference of the arguments r = u, s, that is, k(u, s, b) = k(r, b). If, in addition, the field is isotropic, then the CF is dependent only on the modulus of the difference in the arguments: k(u, s, b) = k(r, b), where $r = |\mathbf{r}| = |\mathbf{u} - \mathbf{s}| = \sqrt{(u_1 - s_1)^2 + \ldots + (u_k - s_k)^2}$.

It should be noted that the methods for determining the CF used at the present time, although to some degree they can be modified for computing the characteristics of nonisotropic and nonuniform fields, nevertheless are oriented primarily on isotropic and uniform fields. However, the condition of uniformity and isotropicity of meteorological and oceanographic fields is very rigorous, and only for extremely limited regions is it possible to speak of its approximate satisfaction. Accordingly, henceforth we will examine the general case of a nonuniform and nonisotropic field.

Assume that as a result of the experiment, whose purpose is determination of the CF parameters, N values are obtained

$$f_k = f_k(\mathbf{u}_k), \quad \mathbf{u}_k \in U, \quad k = \overline{1, N}, \tag{2}$$

where f_k denotes the result of measurement (reading) of the series f(u) of the F(u) field at the point u_k.

Conversion to regression model. Using the available readings (2), we form L values

$$w_{kr} = f_k f_r = f(\mathbf{u}_k) f(\mathbf{u}_r),$$

$$k, r = \overline{1, N}, \quad k \leq r, \quad L = \frac{1}{2} N(N+1)$$

and introduce a vector of the dimensionality L

$$\mathbf{W} = (w_{11}, w_{12}, \ldots, w_{1N}, w_{22}, w_{23}, \ldots, w_{2N}, \ldots, w_{(N-1)N}, w_{NN})^T. \tag{3}$$

It is easy to see that the w_{kr} values can be interpreted as readings of the series w(u¦s) of the random field W(u¦s) = F(u) F(s) at the point u¦s = u_k¦u_r. Here u¦s = ($u_1 \ldots, u_k$, $s_1 \ldots, s_k$). For the mathematical expectation (ME) and CF of the W(u¦s) field we accordingly have

$$E[W(\mathbf{u}\,|\,\mathbf{s})] = E[F(\mathbf{u})\,F(\mathbf{s})] = k(\mathbf{u}, \mathbf{s}, \mathbf{b}), \tag{4}$$

$$K_w(\mathbf{u}\,|\,\mathbf{s}, \mathbf{p}\,|\,\mathbf{q}) = E\{[F(\mathbf{u})\,F(\mathbf{s}) - k(\mathbf{u}, \mathbf{s}, \mathbf{b})] \times \\ \times [F(\mathbf{p})\,F(\mathbf{q}) - k(\mathbf{p}, \mathbf{q}, \mathbf{b})]\}, \tag{5}$$

where p, q $\in$ U.

Therefore the field W(u¦s) can be represented in the form

$$W(\mathbf{u}\,|\,\mathbf{s}) = k(\mathbf{u}, \mathbf{s}, \mathbf{b}) + \Gamma(\mathbf{u}\,|\,\mathbf{s}),$$

where the field Γ (u¦s) has a zero ME and a CF K_Γ (u¦s, p¦q) = K_W(u¦s, p¦q) and can be considered as the field of errors in measuring the nonrandom function k (u,s,b).

By analogy with the W vector we introduce the vectors K and Γ with the components $k_{kr} = k(u_k, u_r, b)$ and $\gamma_{kr} = \gamma(u_k \vdots u_r)$ respectively under the condition $k \leqslant r$, where $\gamma(u_k \vdots u_r)$ denotes the reading of the field $\Gamma(u\ s)$ at the point $u \vdots s = u_k \vdots u_r$; we write

$$\mathbf{W} = \mathbf{K}(\mathbf{b}) + \mathbf{\Gamma}. \tag{6}$$

The W and Γ vectors can be interpreted as sample values of the vectoral random values for which there is satisfaction of the conditions $E(\Gamma) = 0$, $E(W) = K(b)$, and their covariation matrices $D(\Gamma) = D(W)$ have the order L.

Thus, the problem of finding the parameters b of the CF k(u, s, b) can be formulated as the problem of evaluating the parameters of a linear or nonlinear (depending on the type of function k(u, s, b)) regression model with correlated observation errors [13].

A sufficiently general approach to formulation of evaluations of the parameters of the model (6) is as follows. We will use $\hat{b}$ to denote some evaluation of the b vector and use $\hat{W} = K(\hat{b})$ to denote the ME evaluation of the W vector. We introduce the vector of the residues

$$\mathbf{V}(\hat{\mathbf{b}}) = \mathbf{W} - \hat{\mathbf{W}} = \mathbf{W} - \mathbf{K}(\hat{\mathbf{b}}). \tag{7}$$

Then a broad class of evaluations of the b vector can be obtained by minimizing some appropriately selected evaluation function $\rho[V(\tilde{b})]$, that is

$$\hat{\mathbf{b}} = \arg\min_{\tilde{\mathbf{b}}} \rho\,[\mathbf{V}(\hat{\mathbf{b}})] = \arg\min_{\tilde{\mathbf{b}}} \rho\,[\mathbf{W} - \mathbf{K}(\tilde{\mathbf{b}})]. \tag{8}$$

Henceforth we will limit ourselves to the special case of such an evaluation which corresponds to the use in (8) of the function

$$\rho\,[\mathbf{V}(\tilde{\mathbf{b}})] = \mathbf{V}^{\mathsf{T}}(\tilde{\mathbf{b}})\ C\,\mathbf{V}(\tilde{\mathbf{b}}),$$

where C is an appropriately selected positively determined matrix of the order L.

Then we come to the evaluation

$$\hat{\mathbf{b}} = \arg\min_{\tilde{\mathbf{b}}} \,[\mathbf{W} - \mathbf{K}(\tilde{\mathbf{b}})]^{\mathsf{T}}\ C\,[\mathbf{W} - \mathbf{K}(\tilde{\mathbf{b}})], \tag{9}$$

minimizing the generalized sum of the squares of the residues, the so-called evaluation of the generalized least squares method [13].

Evaluation of parameters of a linear model. In a linear parameterization the CF of the random field F(u) is stipulated in the form

$$k(\mathbf{u}, \mathbf{s}, \mathbf{b}) = \mathbf{b}^{\mathsf{T}} \varphi(\mathbf{u}, \mathbf{s}), \quad \mathbf{u}, \mathbf{s} \in U,$$

where the vector $\varphi(u, s) = (\varphi_1(u,s), \ldots, \varphi_m(u,s))^T$; $\varphi_i(u, s)$ are known functions. Then

$$\mathbf{W} = \Phi\,\mathbf{b} + \mathbf{\Gamma},$$

where Φ is a matrix (measuring L x m) of the form

$$\Phi = (\varphi_{11}, \varphi_{12}, \ldots, \varphi_{1N}, \varphi_{22}, \varphi_{23}, \ldots, \varphi_{2N}, \ldots, \varphi_{(N-1)N}, \varphi_{NN})^{\mathsf{T}}, \quad \varphi_{kr} = \varphi(\mathbf{u}_k, \mathbf{u}_r), \quad k \leqslant r.$$

In this case (9) will be equivalent to the expression

$$\hat{\mathbf{b}} = \hat{\mathbf{b}}^{[1]} = (\Phi^{\tau} C \Phi)^{-1} \Phi^{\tau} C \mathbf{W}, \tag{10}$$

in this case the evaluations are not biased ($E[\hat{b}^{[1]}] = b$), and for the covariation matrix of the evaluations

$$D(\hat{\mathbf{b}}^{[1]}) = E[(\hat{\mathbf{b}}^{[1]} - \mathbf{b})(\hat{\mathbf{b}}^{[1]} - \mathbf{b})^{\tau}]$$

we have

$$\hat{D}(\mathbf{b}^{[1]}) = (\Phi^{\tau} C \Phi)^{-1} \Phi^{\tau} C D(\mathbf{W}) C \Phi (\Phi^{\tau} C \Phi)^{-1}. \tag{11}$$

This formula makes it possible for any selected weighted matrix C and the known covariation matrix D(W) (for this it is sufficient to know the CF of the field W(u;s)) to evaluate the statistical accuracy (covariations) of the evaluations $\hat{b}^{[1]}$.

If D(W) is known, then for attaining the maximum accuracy of the evaluations as the weighted matrix it is necessary to select $C = D^{-1}(W)$ [1, 13]. In this case instead of (10) we obtain

$$\hat{\mathbf{b}}^{[2]} = (\Phi^{\tau} D^{-1}(\mathbf{W}) \Phi)^{-1} \Phi^{\tau} D^{-1}(\mathbf{W}) \mathbf{W}, \tag{12}$$

and the covariation matrix of the evaluations

$$D(\hat{\mathbf{b}}^{[2]}) = (\Phi^{\tau} D^{-1}(\mathbf{W}) \Phi)^{-1}. \tag{13}$$

We note that with a stipulated CF of the field W(u;s) the matrices $D(\hat{b}^{[1]})$ and $D(\hat{b}^{[2]})$, characterizing the statistical accuracy of the evaluations (10) and (12), are dependent on the position of the points u_k at which the $f(u_k)$ measurements were made since the elements of the matrices Φ and D(W) are dependent on the u_k values. Accordingly, in this case it is possible to formulate the problem of a priori formulation of optimum experimental plans, that is, rational distribution of the network of observation stations. However, if the D(W) matrix is unknown, as is usually the case in actual practice, as the weighted matrix C in (10) it is necessary (depending on the available a priori information) to take some symmetric positively determined matrix (for example, of the diagonal type). The simplest variant of such a matrix is a unit matrix. In this case we come to the usual evaluations of the least squares method.

Evaluation of parameters of nonlinear model. Now we will examine a case when k(u, s, b) is a function nonlinear relative to the evaluated parameters b, continuous with respect to b. If the function k(u, s, b) is smooth with respect to b in the neighborhood of the true values of the b parameters and if equation (9) has a unique solution (we will denote it $\hat{b}^{[3]}$), then with sufficiently great dimensions of the U region in which the measurements (2) were made and with a large number of measurements N it can be assumed approximately that $\hat{b}^{[3]}$ is an unbiased evaluation of the b vector and it can be assumed that

$$k(\mathbf{u}, \mathbf{s}, \hat{\mathbf{b}}^{[3]}) \approx k(\mathbf{u}, \mathbf{s}, \mathbf{b}) + (\hat{\mathbf{b}}^{[3]} - \mathbf{b})^{\tau} \varphi(\mathbf{u}, \mathbf{s}),$$

where in contrast to the linear case $\varphi(u, s)$ are dependent on the true values of the parameters

$$\varphi(\mathbf{u}, \mathbf{s}) = \varphi(\mathbf{u}, \mathbf{s}, \mathbf{b}) = \nabla k(\mathbf{u}, \mathbf{s}, \mathbf{b}), \tag{14}$$
$$\nabla^{T} = \left(\frac{\partial}{\partial b_1}, \ldots, \frac{\partial}{\partial b_m}\right).$$

Accordingly, the covariation matrix of the evaluations $\hat{b}^{[3]}$ is also dependent on b, although it is also determined, as for the linear case, by the expression

$$D(\hat{\mathbf{b}}^{[3]}) = D(\hat{\mathbf{b}}^{[3]}, \mathbf{b}) \approx (\Phi^{T} C \Phi)^{-1} \Phi^{T} CD(\mathbf{W})\, C\Phi (\Phi^{T} C \Phi)^{-1}. \tag{15}$$

However, here the Φ matrix is dependent on b, that is, $\Phi = \Phi(b)$, since in its forming use is made of elements formed by the values of the function (14). If the D(W) matrix is known, by selecting $C = D^{-1}(W)$ (as in the linear case), we obtain evaluations with the best statistical properties. The dispersion matrix of these evaluations (we will denote them $\hat{b}^{[4]}$) has the form

$$D(\hat{\mathbf{b}}^{[4]}) = D(\hat{\mathbf{b}}^{[4]}, \mathbf{b}) \approx (\Phi^{T} D^{-1}(\mathbf{W})\, \Phi)^{-1}. \tag{16}$$

With a CF quite smooth relative to b, instead of the values of the function (14) in computing the matrix $\Phi = \Phi(b)$ it is possible to use the values of the functions

$$\hat{\varphi}(\mathbf{u}, \mathbf{s}) = \varphi(\mathbf{u}, \mathbf{s}, \hat{\mathbf{b}}^{[\cdot]}) = \nabla k(\mathbf{u}, \mathbf{s}, \hat{\mathbf{b}}^{[\cdot]}),$$

therefore, it can be assumed approximately that $\Phi \approx \Phi(b^{[\cdot]})$. With a known CF of the field $W(u \mid s)$, as in a linear case, it is possible to formulate the problem of optimum planning of an experiment, but a priori planning here is already impossible although it is possible to use sequential planning, minimax or Bayes approaches [12].

Evaluation of CF of a Gaussian random field. Now we will examine an important special case of practical importance — evaluation of the parameters of the CF of a Gaussian (normal) field. As is known [9], for such a field F(u) the CF of the field $W(u \mid s) = F(u)\, F(s)$ is expressed in the following way through the CF of the initial field:

$$K_{w}(\mathbf{u} \mid \mathbf{s}, \mathbf{p} \mid \mathbf{q}) = k(\mathbf{u}, \mathbf{p}, \mathbf{b})\, k(\mathbf{s}, \mathbf{q}, \mathbf{b}) + k(\mathbf{u}, \mathbf{q}, \mathbf{b})\, k(\mathbf{s}, \mathbf{p}, \mathbf{b}).$$

Thus, the CF of the field $W(u \mid s)$, like the field F(u), and accordingly, also the covariation matrix D(W), is dependent in a known manner on the values of the evaluated b parameters, that is D(W) = D(W, b). In this case, regardless of the nature of the function k(u, s, b) (linear or nonlinear relative to b) it is possible to propose evaluations of three types.

Evaluations of the first type are computed in accordance with (9) (for the case of linear parameterization — in accordance with (10)). The covariation matrix of these evaluations, dependent (through D(W)) on the unknown b values, can be computed approximately using formula (11) or (15) by a replacement of b by their determined evaluations.

An evaluation of the second type differs from the first in that in order to increase the accuracy of the evaluation use is made of an iteration procedure

$$\hat{\mathbf{b}}_l = \arg\min_{\tilde{\mathbf{b}}} (\mathbf{W} - \mathbf{K}(\tilde{\mathbf{b}}))^\tau D^{-1}(\mathbf{W}, \hat{\mathbf{b}}_{l-1})(\mathbf{W} - \mathbf{K}(\tilde{\mathbf{b}})), \; l = 1, 2, \ldots \tag{17}$$

where $\hat{b}_l$ is the evaluation of the b parameters in the l-th step of the procedure; $\hat{b}_0$ is the appropriately selected initial evaluation of the parameters; $D(W, \hat{b}_0)$ is a matrix of a stipulated type or a matrix selected on the basis of a priori information. Together with the values of the evaluation of the $\hat{b}$ vector, in each step of the procedure the evaluation of the matrix D(W,b) is refined. Accordingly, such a procedure can also be used in the case of a virtually complete a priori uncertainty concerning the form D(W,b); in this case it is possible to stipulate $D(W, b_0)$, for example, in the form of a unit matrix. The covariation matrix of the evaluations $\hat{b}_l$ can be computed approximately using the formula

$$D(\mathbf{W}, \hat{\mathbf{b}}_l) \approx (\Phi^\tau D^{-1}(\mathbf{W}, \hat{\mathbf{b}}_{l-1})\,\Phi)^{-1}\,\Phi^\tau D^{-1}(\mathbf{W}, \hat{\mathbf{b}}_{l-1})\,D(\mathbf{W}, \hat{\mathbf{b}}_l) \times$$
$$\times D^{-1}(\mathbf{W}, \hat{\mathbf{b}}_{l-1})\,\Phi\,(\Phi^\tau D^{-1}(\mathbf{W}, \hat{\mathbf{b}}_{l-1})\,\Phi)^{-1}.$$

With adherence to the necessary conditions the procedure (17) converges

$$(\lim_{l\to\infty} \hat{\mathbf{b}}_l = \hat{\mathbf{b}}_\infty),$$

in this case

$$D(\hat{\mathbf{b}}_\infty) \approx (\Phi^\tau D^{-1}(\mathbf{W}, \hat{\mathbf{b}}_\infty)\,\Phi)^{-1}.$$

The procedure can be stopped under the condition of attaining a stipulated closeness of the evaluations $\hat{b}_l$ and $\hat{b}_{l-1}$ or their covariation matrices.

Finally, the third method for obtaining the evaluations is a direct minimizing of the generalized sum of the squares of the residues with use of a weighted matrix dependent on the evaluated parameters, that is, it essentially involves solution of the equation

$$\hat{\mathbf{b}} = \arg\min_{\tilde{\mathbf{b}}} (\mathbf{W} - \mathbf{K}(\tilde{\mathbf{b}}))^\tau D^{-1}(\mathbf{W}, \tilde{\mathbf{b}})(\mathbf{W} - \mathbf{K}(\tilde{\mathbf{b}})). \tag{18}$$

However, even with linear parameterization of the function k(u,s,b) computation of the evaluation (18) involves satisfaction of the nonlinear operations for the components of the W vector. It is extremely difficult to obtain an expression for $D(\hat{b})$ in this case but it is clear intuitively that the evaluations (18) and (17) must not differ significantly in their statistical properties. We note that with use of evaluations of the first two types considered above there is a possibility for the optimum planning of an experiment (both with the sequential and with the minimax or Bayes approaches).

Some models of the CF for a nonuniform field. At the present time in meteorological and oceanographic computations use is made primarily of models of an isotropic and uniform field, although in the literature there is constant mention of the limitations on their use for the description of macroscale processes. In part such a situation is evidently attributable to the lack of sufficiently effective algorithms for the processing of data for general cases of nonuniform and nonisotropic fields.

As one of the possible models of a CF of a nonisotropic and nonuniform field it is possible to propose a function in the form

$$K_F(\mathbf{u}, \mathbf{s}) = \Psi(\mathbf{u}, \mathbf{s}, \mathbf{b})\, \varphi(r, \mathbf{b}), \tag{19}$$

where $r = \sqrt{(u_1 - s_1)^2 + \ldots + (u_k - s_k)^2}$,

Ψ(u,s,b) and φ(r,b) are appropriately selected functions. In special cases only one of the functions Ψ and φ may be dependent on b.

We note that CF in the form

$$K_F(\mathbf{u}, \mathbf{s}) = \Psi_1(\mathbf{u}, \mathbf{b})\, \Psi_1(\mathbf{s}, \mathbf{b})\, \varphi(r, \mathbf{b}),$$

obtained from (19) with the choice $\Psi(u,s,b) = \Psi_1(u,b)\Psi_1(s,b)$, corresponds to the CF of a nonuniform and nonisotropic field represented in the form of the product of a uniform and isotropic field Q(u) and a CF of the form φ(r, b) and a determined function Ψ_1(u, b).

Example. We will examine computation of the CF parameter dependent on two arguments and being a special case of (19),

$$K_F(\mathbf{u}, \mathbf{s}) = b\Psi(\mathbf{u}, \mathbf{s})\, \varphi(r), \quad \mathbf{u}, \mathbf{s} \in U,$$

where U is a part of a plane, $r = \sqrt{(u_1 - s_1)^2 + (u_2 - s_2)^2}$,

φ(r) and Ψ(u,s) are known functions, b is the sought-for parameter.

Assume that in the U region there are 2 n readings of the centered Gaussian random field. With a limited Ψ (u, s) and with a function $|\varphi(r)|$ attenuating with an increase in r it is always possible to indicate such an r_0 for which $|K_F(u,s)| \leqslant \delta$, where δ is a prestipulated small parameter. Assume that the existing 2n readings fall at points forming n pairs of points u_i, v_i, $i = \overline{1,n}$ with the distance r_i

$$(r_i = \sqrt{(u_{i1} - v_{i1})^2 + (u_{i2} - v_{i2})^2})$$

between the points of the i-th pair (see Fig. 1); the correlation between the readings at the points of each pair is significant, whereas between the readings of adjacent pairs it is negligible (that is, the distance between the points of any different pairs is greater than r_0).

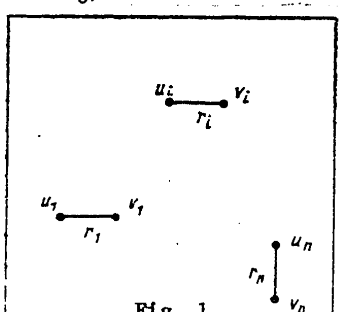

Fig. 1

As a simplification we will first examine one pair of points u_i and v_i, situated at the distance r_i. In this case, writing the vectors

$$\Phi_i = (\Psi(u_i, u_i)\,\varphi(0);\quad \Psi(u_i, v_i)\,\varphi(r_i);\quad \Psi(v_i, v_i)\,\varphi(0))^{\mathsf{T}},$$

$$W_i = (f^2(u_i);\quad f(u_i)\,f(v_i);\quad f^2(v_i)),$$

forming the $D(W_i)$ matrix and making computations in accordance with (12) and (13), we obtain

$$\hat{b}_i = (\Phi_i^{\mathsf{T}} D(W_i)\,\Phi_i)^{-1}\,\Phi_i^{\mathsf{T}} D^{-1}(W_i)\,W_i = \frac{1}{2}\,[\Psi(v_i, v_i)\,\varphi(0)\,f^2(u_i) -$$

$$-2\,\Psi(u_i, v_i)\,\varphi(r_i)\,f(u_i)\,f(v_i) + \Psi(u_i, u_i)\,\varphi(0)\,f^2(v_i)],\qquad D(\hat{b}_i) = b^2.$$

Thus, in this case the accuracy of the evaluation is dependent only on the value of the evaluated parameter b itself and is not dependent on the distance r_i between the points at which the measurements were made.

Proceeding now to the case of n pairs of points u_i, v_i, $i = \overline{1,n}$, forming the vectors Φ and W of the dimensionality 2n(2n+1)/2 and the D(W) matrix, making computations using formula (13), it is easy to show that $D(\hat{b}) = b^2/n$. It therefore follows that the accuracy of the evaluation is dependent for the particular example only on the value of the parameter to be evaluated and the number of pairs of points and is not dependent on the distances between them. It is easy to show that the evaluation of the b parameter, together with (13), can be computed using the formula

$$\hat{b} = \frac{1}{n}\sum_{i=1}^{n}\hat{b}_i.$$

This is attributable to the fact that the dimensionality of the Φ and W vectors in this case without a loss in accuracy can be reduced to 3n, excluding from consideration the noninformative elements of the Φ vector corresponding to a zero correlation between the readings at the points of different pairs and the corresponding components of the W vector, in this case also changing the D(W) matrix. (In this case the order of D(W) is equal to 3n).

Concluding remarks. The examined methods are quite universal and are applicable for evaluating the parameters of the CF of arbitrary random fields (including nonuniform and nonisotropic fields) with stipulation of one field series.

The results can also be generalized for the case of stipulation of the readings of several field series at arbitrary (not necessarily coinciding for all series) points of a finite set of values of the argument. In this case for each i-th series we will have a vector W_i which is represented in the form

$$W_i = K_i(b) + \Gamma_i,\quad i = \overline{1, n},$$

where the vectors W_i, K_i, Γ_i have the same sense as the corresponding W, K, Γ vectors in formula (6).

We introduce the vector $\tilde{W}$ $(W_1,\ldots,W_n)^{T}$, which can be represented in the form

$$\tilde{W} = \tilde{K}(b) + \tilde{\Gamma},$$

where $\tilde{K}(b) = (K_1(b), \ldots, K_n(b))^{\mathsf{T}}$ and $\Gamma = (\Gamma_1, \ldots, \Gamma_n)^{\mathsf{T}}$.

It is easy to see that the covariation matrices of the $\tilde{W}$ and $\tilde{\Gamma}$ vectors have a block structure

$$D(\tilde{\Gamma}) = D(\tilde{W}) = \begin{pmatrix} D_{11} & O_{12} & \ldots & O_{1n} \\ O_{21} & D_{22} & \ldots & O_{2n} \\ \cdot & \cdot & \cdot & \cdot \\ O_{n1} & O_{n2} & \ldots & D_{nn} \end{pmatrix},$$

where $D_{ii} = D(\Gamma_i) = D(W_i)$, and O_{ij} are zero matrices. The principal formulas for finding the b evaluations and their covariation matrices in this case do not change if W, K(b) in them is replaced by $\tilde{W}$ and $\tilde{K}(b)$.

However, in the case of stipulation of readings of field series at points coinciding for all the series it is necessary to average the readings corresponding to different series. Thus, there is a conversion to the readings of one series of another field differing from the initial field in that the values of its CF are n times less than the corresponding values of the CF of the initial field and further processing is carried out the same as for one series.

Finally, in a case when the initial field is not centered and its ME is a function of a known form, dependent on several unknown parameters, for application of the proposed methods for evaluating the CF it is first necessary to find evaluations of the ME parameters (for example, using the least squares method) and corresponding evaluations of the ME itself at the points of stipulation of the readings of series for the initial field and proceed to field readings which can be considered approximately as already centered. Since the accuracy in evaluating the field ME is usually greater than the accuracy in evaluating its CF, such a conversion does not lead to appreciable errors.

The proposed approach can be used not only for finding the covariation function of the field, but also for determining the parameters of the cross-variation function of the fields of two meteorological elements. Assume that $F_1(u)$ and $F_2(u)$ are random fields whose cross-covariation function k(u,s,b) is known with an accuracy to the unknown vector of the b parameters, and as a result of the experiment N_1 readings of the $F_1(u)$ field are obtained,

$$f_{1k} = f(\mathbf{u}_k), \quad \mathbf{u}_k \in U, \quad k = \overline{1, N_1}$$

and N_2 readings of the $F_2(u)$ field,

$$f_{2r} = f_2(\mathbf{u}_r), \quad \mathbf{u}_r \in U, \quad r = \overline{1, N_2}.$$

We will form L values

$$w_{kr} = f_{1k} f_{2r} = f_1(\mathbf{u}_k) f_2(\mathbf{u}_r), \quad k = \overline{1, N_1}, \quad r = \overline{1, N_2}, \quad L = N_1 N_2.$$

We introduce the dimensionality vector L:

$$\mathbf{W} = (w_{11}, w_{12}, \ldots, w_{1N_1}, w_{21}, w_{22}, \ldots, w_{2N_1}, \ldots,$$

$$w_{N_2 1}, w_{N_2 2}, \ldots, w_{N_2 N_1}), \quad L = N_1 N_2.$$

Then reasoning the same as in the case of the covariation function, it can be confirmed that all the formulas for computing the b vector will also be correct for this case.

It should be noted that the proposed procedures for computing the CF with their use with an electronic computer can require a greater computer memory volume in comparison with those used at the present time, but also have a number of advantages, the most important of which is their applicability for nonuniform and nonisotropic fields. These procedures are particularly effective in the case of a small number of initial data (thin network of measuring stations in the considered region and an infrequent reading of data at an individual station or research ship). However, if the masses of data are large and the direct realization of the proposed procedure is difficult due to limitations on the volume of the computer memory (this is dependent on the specific type of computer used for the processing) it is possible to propose a number of simplified, decomposition or iteration procedures for the processing of data.

BIBLIOGRAPHY

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

2. Belyayev, V. I., OBRABOTKA I TEORETICHESKIY ANALIZ OKEANOGRAFICHESKIKH NABLYUDENIY (Processing and Theoretical Analysis of Oceanographic Observations), Kiev, Naukova Dumka, 1973.

3. Brimkulov, U. N., Krug, G. K., Savanov, V. L., PLANIROVANIYE REGRESSIONNYKH EKSPERIMENTOV PRI ISSLEDOVANII SLUCHAYNYKH POLEY (Planning of Regression Experiments in an Investigation of Random Fields), Preprint USSR Academy of Sciences, Scientific Council on the Complex Problem "Cybernetics," Moscow, VINITI, 1978.

4. Brimkulov, U. N., Krug, G. K., Savanov, V. L., "Rationalization of the Measuring Network Using the Criterion of Accuracy of the Mathematical Description of the Field of Norms," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 7, 1978.

5. Gandin, L. S., OB"YEKTIVNYY ANALIZ METEOROLOGICHESKIKH POLEY (Objective Analysis of Meteorological Fields), Leningrad, Gidrometeoizdat, 1963.

6. Gandin, L. S., Kagan, R. L., STATISTICHESKIYE METODY INTERPRETATSII METEOROLOGICHESKIKH DANNYKH (Statistical Methods for the Interpretation of Meteorological Data), Leningrad, Gidrometeoizdat, 1976.

7. Zhuravleva, Ye. B., Kagan, R. L., Polyak, I. I., "Computation of Autocorrelation and Cross-Correlation Functions from Several Records of a Random Process," TRUDY GGO (Transactions of the Main Geophysical Observatory), No 289, 1972.

8. Kagan, R. L., "On the Accuracy in Computing Space Correlation Functions," TRUDY GGO, No 308, 1973; No 336, 1974.

9. Levin, B. R., TEORETICHESKIYE OSNOVY STATISTICHESKOY RADIOTEKHNIKI (Theoretical Principles of Statistical Radio Engineering), Moscow, Sovetskoye Radio, 1974.

10. Mashkovich, S. A., "Increasing the Quality of Objective Analysis of the Pressure Field Over Regions With a Thin Network of Aerological Stations," TRUDY MMTs (Transactions of the World Meteorological Center), No 10, 1965.

11. Savanov, V. L., Yarygina, T. A., "Some Methods for Determining the Correlation Function of a Random Field," TRUDY MEI (Transactions of the Moscow Power Institute), No 399, 1979.

12. Fedorov, V. V., TEORIYA OPTIMAL'NOGO EKSPERIMENTA (Theory of an Optimum Experiment), Moscow, 1971.

13. Khudson, D., STATISTIKA DLYA FIZIKOV (Statistics for Physicists), Moscow, Mir, 1970.

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MEAN ANNUAL ZONAL ATMOSPHERIC CIRCULATION

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[Article by Candidate of Physical and Mathematical Sciences N. S. Sidorenkov, USSR Hydrometeorological Scientific Research Center, submitted for publication 9 Jan 80]

[Text]

Abstract: A semi-empirical differential equation is derived for the distribution of the absolute moment of momentum. It is shown that the change in the moment of momentum in the atmosphere is determined by the balance of two fluxes of the moment of momentum: horizontal, arising as a result of macroturbulent exchange and vertical, associated with microturbulent viscosity. Its analytical solution is obtained in the case of a very simple model of change in the coefficients of turbulent exchange. It is shown that the mean annual zonal circulation arises as a result of the redistribution of the absolute moment of momentum of air between the low and high latitudes under the influence of macroturbulent mixing. Macroturbulent viscosity weakens zonal circulation. Its stationary state is attained only after accumulation of a definite positive moment of momentum in the atmosphere, which is transferred from the earth to the atmosphere.

It is well known that as an average for the year in the low latitudes there is a prevalence of winds having a velocity component from east to west, whereas in the temperate and high latitudes there is a prevalence of west-east transport. A change in the sign of the zonal wind velocity component is observed at the earth's surface in the so-called "horse latitudes" (near ±35°). Zones of calms are situated here. During recent decades it has become clear that variations in the intensity of zonal circulation of the atmosphere are associated with changes in the earth's rate of rotation [5, 6]. An intensification of zonal circulation of the atmosphere occurs as a result of influx of the moment of momentum from the earth, whereas weakening occurs as a result of its transfer toward the earth. It was also established that the moment of momentum of zonal winds is not equal to zero and as an average for the year is about $13 \cdot 10^{25}$ $kg \cdot m^2 \cdot sec^{-1}$ [6].

Modern numerical models of zonal circulation of the atmosphere, applied on an electronic computer, reproduce the pattern of zonal circulation of the atmosphere well [3]. However, they did not lead to a significant deepening of our ideas concerning the physics of the processes forming and supporting zonal circulation. The results of the numerical models are sets of numbers which differ little from the data from meteorological observations. We feel that the simplest qualitative theory makes possible a deeper understanding of the physical essence of the phenomena than even the most inventive of the existing numerical models. The purpose of this article is the proposal of such a theory.

We will select a fixed Cartesian coordinate system with its origin at the center of the earth's mass and with axes oriented relative to "fixed" stars. We will direct one of the axes along the earth's axis of rotation and the other two will be laid out in the equatorial plane.

We will write the equation of motion of a unit volume of air for this inertial reference system:

$$\rho \frac{d\mathbf{V}_a}{dt} = -\nabla P - \frac{fM_{\oplus}\rho}{R^3}\mathbf{R} + \mathbf{F}, \tag{1}$$

where $V_a = [\Omega \times R] + V_0$ is the velocity of absolute motion, V_0 is wind velocity, Ω is the angular velocity of the earth's rotation, R is the geocentric radius-vector of the considered volume, ρ is air density, f is the gravitational constant, $M_{\oplus}$ is the earth's mass, ∇P is the gradient of atmospheric pressure P, F is frictional force, related to a unit volume, t is time.

We will multiply each term in equation (1) vectorally from the left by the geocentric radius-vector R and we will take into account that in the free atmosphere the frictional force is negligible. Then we obtain an equation for the absolute moment of momentum of a unit volume of air in the atmosphere

$$\mathbf{R} \times \rho \frac{d\mathbf{V}_a}{dt} = \rho \frac{d}{dt}[\mathbf{R} \times \mathbf{V}_a] = [\nabla P \times \mathbf{R}]. \tag{2}$$

The moment of terrestrial gravitation is equal to zero since this force is directed along the radius R. Projecting the vector equation (2) onto the earth's axis of rotation, we obtain

$$\rho \frac{dl}{dt} = -\frac{\partial P}{\partial \lambda}, \tag{3}$$

where $l = \Omega R^2 \sin^2\theta + uR\sin\theta = (\Omega + \alpha)R^2\sin^2\theta$ is the projection of the absolute moment of momentum of a unit air mass onto the axis of the earth's rotation, $u = \alpha R\sin\theta$ is the velocity of the zonal wind (positive direction to the east), α is the angular velocity of zonal air motion relative to the earth's surface, θ is the polar angle or colatitude φ to $\pi/2$, λ is longitude, reckoned to the east.

We note that the projections of the absolute moment of momentum onto the axes situated in the equatorial plane are negligible in comparison with the l value since the deviation of the instantaneous vector Ω from the mean long-term value does not exceed 10^{-6} radian [7], and $V_0 \ll V_a$.

The left-hand side of equation (3) by means of the continuity equation can be represented in the form

$$\rho \frac{dl}{dt} = \frac{d\rho l}{dt} - l\frac{d\rho}{dt} = \frac{\partial \rho l}{\partial t} + \nabla \cdot \rho l \mathrm{V}_a - l\left(\frac{d\rho}{dt} + \right.$$
$$\left. + \rho \nabla \cdot \mathrm{V}_a\right) = \frac{\partial \rho l}{\partial t} + \nabla \cdot \rho l \mathrm{V}_a. \quad (4)$$

We will represent the divergence $\nabla \cdot \rho l \mathrm{V}_a$ of the flux of the projection of the absolute moment of momentum in a spherical coordinate system and substitute (4) into equation (3). Then we obtain

$$\frac{\partial \rho l}{\partial t} = -\frac{1}{R^2}\frac{\partial}{\partial R}(R^2 \rho l v_R) - \frac{1}{R\sin\theta}\frac{\partial}{\partial \theta}(\sin\theta \rho l v_\theta) - \quad (5)$$
$$-\frac{1}{R\sin\theta}\frac{\partial}{\partial \lambda}(\rho l v_\lambda) - \frac{\partial P}{\partial \lambda}.$$

Since we are interested in the zonal circulation of air in the atmosphere, we will average equation (5) in longitude — we will integrate all its terms for λ from 0 to 2π. We take into account that the integrals of the last two terms are equal to zero, since it is possible to neglect the jumps of pressure and velocity u at mountain ranges. Introducing the traditional notations

$$\overline{x} = \frac{1}{2\pi}\int_0^{2\pi} x\, d\lambda,$$

we obtain

$$\frac{\partial \overline{\rho l}}{\partial t} = -\frac{1}{R^2}\frac{\partial}{\partial R}\left[R^2(\overline{\rho l}\,\overline{v}_R + \overline{\rho l' v'_R})\right] - \quad (6)$$
$$-\frac{1}{R\sin\theta}\frac{\partial}{\partial \theta}[\sin\theta\,(\overline{\rho l}\,\overline{v}_\theta + \overline{\rho l' v'_\theta})].$$

In equation (6) terms of the type $\overline{\rho l}\,\overline{v_i}$ and $\rho \overline{l' v'_i}$ reflect transfer of the projection of the absolute moment of momentum by ordered circulation and eddies (turbulence) respectively. Investigations of the components of the balance of moment of momentum in the atmosphere, whose results are summarized in the monographs [4, 8], indicated that the transfer of the moment of momentum is accomplished for the most part by turbulence. The role of ordered circulation in the atmosphere is relatively small. On the basis of these empirical data terms of the type $\overline{\rho l}\,\overline{v_i}$ can be neglected in comparison with the terms $\rho \overline{l' v'_i}$.

As is well known, in physics extensive use is made of intuitive phenomenological expressions relating the fluxes of different physical parameters and the gradients of these parameters. In particular, in the semi-empirical theory of turbulence the fluxes of momentum described by Reynolds stresses are assumed to be proportional to the gradient of mean wind velocity. Adhering to these hypotheses, the fluxes of the moment of momentum in the vertical and meridional directions can be expressed by the following approximate equations:

$$\overline{\rho l' v'_R} \approx -A_R \frac{\partial \overline{l}}{\partial R} \text{ и } \overline{\rho l' v'_\theta} \approx -A_\theta \frac{\partial \overline{l}}{R\,\partial \theta}, \quad (7)$$

where A_R and A_θ are the coefficients of vertical and meridional exchange respectively.

Substituting (7) into (6), we obtain a second-degree differential equation in partial derivatives for the projection of the absolute moment of momentum onto the earth's axis of rotation

$$\frac{\partial \rho \bar{l}}{\partial t} = \frac{1}{R^2} \frac{\partial}{\partial R} \left(R^2 A_R \frac{\partial \bar{l}}{\partial R} \right) + \frac{1}{R^2 \sin \theta} \frac{\partial}{\partial \theta} \left(\sin \theta \, A_\theta \frac{\partial \bar{l}}{\partial \theta} \right). \tag{8}$$

Equation (8) shows that the rate of change in the projection of the absolute moment of momentum in a unit volume is determined by the sum of the convergences of the moment of momentum, or, what is the same, the moments of the forces arising due to microturbulent (first term) and macroturbulent (second term on the right-hand side) transfers of the moment of momentum. It makes it possible to compute the zonal circulation of the atmosphere because the distribution of the l value in an atmosphere rotating together with the earth is known.

Now we will attempt to find a solution of equation (8). For this we will select the boundary conditions and make simplifications. An obvious boundary condition is the condition of "attachment" of the air to the earth's surface

$$\alpha = 0, \quad l = \Omega R_0^2 \sin^2 \theta \quad \text{when } R = R_0. \tag{9}$$

Here R_0 is the earth's radius. A less obvious condition is the condition at the upper boundary of the atmosphere. We will assume that there (when $R = R_\infty$) the vertical flux of the moment of momentum tends to zero and therefore

$$\frac{\partial \bar{l}}{\partial R} \approx \frac{\partial \alpha}{\partial R} = 0 \text{ when } R = R_\infty. \tag{10}$$

We will examine a stationary (not dependent on time) zonal circulation of the atmosphere. In this case $\partial \rho \bar{l} / \partial t \approx 0$ and equation (8) can be written in the form of an equality of the moments of forces operative in a unit volume due to the vertical and meridional transfers of the moment of momentum in the atmosphere respectively:

$$\frac{\partial}{\partial R} \left(A_R R^2 \frac{\partial \bar{l}}{\partial R} \right) = - \frac{1}{\sin \theta} \frac{\partial}{\partial \theta} \left(A_\theta \sin \theta \frac{\partial \bar{l}}{\partial \theta} \right). \tag{11}$$

The coefficients of meridional and vertical exchange are essentially dependent on latitude and altitude. In the first very rough approximation it can be assumed that

$$A_\theta = \text{const}, \tag{12}$$

$$A_R = A_z \sin \theta, \tag{13}$$

where A_z = const. It is impossible to assume the coefficient A_R to be independent of latitude because, as indicated by the theory of the atmospheric boundary layer, it is proportional to wind velocity $u = \alpha R \sin \theta$.

Substituting into (11) the exchange coefficients A_θ and A_R from (12) and (13) and taking into account that

$$\frac{\partial \bar{l}}{\partial R} = \frac{\partial \bar{\alpha}}{\partial R} R^2 \sin^2 \theta + 2 (\Omega + \bar{\alpha}) R \sin^2 \theta \approx \frac{\partial \bar{\alpha}}{\partial R} R^2 \sin^2 \theta,$$

$$\frac{\partial \bar{l}}{\partial \theta} = (\Omega + \bar{\alpha}) R^2 \sin 2 \theta + \frac{\partial \bar{\alpha}}{\partial \theta} R^2 \sin^2 \theta \approx \Omega R^2 \sin 2 \theta,$$

we obtain

$$\frac{\partial}{\partial R} R^4 \frac{\partial \alpha}{\partial R} = -6\,\Omega\,\frac{A_\theta}{A_z}\,\frac{2/3-\sin^2\theta}{\sin^3\theta}\,R^2. \tag{14}$$

We will denote the constant value $2\,\Omega\,\frac{A_\theta}{A_z} \equiv G$,

and the latitude function $\frac{2/3-\sin^2\theta}{\sin^3\theta} \equiv \Phi(\theta)$.

Integrating equation (14), we have

$$\bar{\alpha} = -G\Phi(\theta)\ln R - C_1(\theta)\frac{1}{3R^3} + C_2(\theta). \tag{15}$$

The unknown functions of latitude $C_1(\theta)$ and $C_2(\theta)$ are easily found from conditions (12) and (13) respectively:

$$C_1(\theta) = G\Phi(\theta)R_\infty^3, \tag{16}$$

$$C_2(\theta) = G\,\Phi(\theta)\left(\ln R_0 + \frac{R_\infty^3}{3R_0^3}\right). \tag{17}$$

After the substitution of these functions into expression (15) we obtain a final expression for the relative angular velocity of rotation of the atmosphere

$$\bar{\alpha} = G\,\Phi(\theta)\,f(R) \equiv 2\,\Omega\frac{A_\theta}{A_z}\,\frac{2/3-\sin^2\theta}{\sin^3\theta}\left[\frac{R_\infty^3}{3}\left(\frac{1}{R_0^3}-\frac{1}{R^3}\right)-\ln\frac{R}{R_0}\right]. \tag{18}$$

Formula (18) shows that as an average for the year the relative angular velocity in both hemispheres is negative at latitudes below 35° and positive at latitudes above 35°. A change in sign occurs at latitudes 35°N and 35°S. With altitude the velocity increases in conformity to a complex law.

The negative α values indicate that the atmosphere lags in its rotation behind the earth, whereas positive α values indicate that it outpaces the earth. The earth rotates from west to east. Accordingly, in the first case ($\alpha < 0$) there is an easterly wind, whereas in the second ($\alpha > 0$) there is a westerly wind.

Thus, the resulting solution (18), despite the exceeding roughness of the assumptions (12)-(13), satisfactorily describes the latitudinal distribution of zonal winds in the atmospheric surface layer. The circumstance that the angular velocity α at the poles tends to infinity is without fundamental significance since there the $R\sin\theta$ value tends to zero.

Now we will check to see whether the derived solution (18) satisfies the observed fact that the moment of momentum of the zonal winds is not equal to zero but as an average for the year is about $13\cdot10^{25}$ kg·m²·sec⁻¹. For this purpose we will estimate the moment of momentum of the determined zonal circulation of the atmosphere (18). In the first approximation we will consider a homogeneous atmosphere with a constant density $\rho = \rho_0 = 1.29$ kg·m⁻³, so that its height is equal to $R_\infty - R_0 \approx 8$ km.

$$h = \int_W \varkappa R^3 \sin^2\Theta \rho \, dW = 2\pi G \rho_0 \int_{R_0}^{R_\infty} R^4 f(R) \, dR \int_0^\pi \Phi(\Theta) \sin^8\Theta \, d\Theta \approx$$

$$\approx \frac{8}{9} \pi^2 \Omega \frac{A_\theta}{A_z} \rho_0 R_0^2 (R_\infty - R_0)^3 = 17 \cdot 10^{21} \frac{A_\theta}{A_z} \ \text{kg} \cdot \text{m}^2 \cdot \text{sec}^{-1}.$$

Here it is assumed that $R = 6.37 \cdot 10^6$ m; $\Omega = 7.29 \cdot 10^{-5}$ sec^{-1}; W is the volume of the atmosphere. According to empirical data, the values of the coefficients of horizontal macroturbulent exchange fall in the range 10^6–10^7 kg·m^{-1}·sec^{-1}, whereas the values of the coefficients of microturbulent exchange fall in the range 1-10 kg·m^{-1}·sec^{-1} [1]. Therefore the ratio $A_\theta / A_z \approx 10^6$, and accordingly, the moment of momentum is $h \approx 17 \cdot 10^{27}$ kg·m^2·sec^{-1} or approximately 140 times greater than the observed value. Evidently, this noncorrespondence is obtained due to the fact that the values of the coefficients of microturbulent exchange A_z used in meteorology do not coincide with their values for the entire atmosphere (because they are computed, as a rule, on the basis of local, rather than planetary data). We note that a similar conclusion was drawn by N. Ye. Kochin and his students. In computing zonal circulation on the basis of the stipulated temperature distribution in the atmosphere they obtained plausible results only with exchange coefficients A_z of the order of magnitude 10^3 kg·m^{-1}·sec^{-1} [2].

Thus, the above-cited rough semi-empirical theory describes well the principal peculiarities of the mean annual zonal circulation of the atmosphere: presence of zones of easterly winds in the low latitudes, westerly winds in the temperate and high latitudes, zones of calms near latitudes ±35°, and presence in the atmosphere of a constant positive value of the moment of momentum of zonal winds. The physical essence of the processes transpiring in the atmosphere is also clear. Zonal circulation of the atmosphere arises and is maintained under the influence of meridional and vertical fluxes of the absolute moment of momentum, which are caused by macroturbulent and microturbulent mixing of the atmosphere respectively.

Macroturbulence, caused first and foremost by the equator-pole temperature contrast, transfers the moment of momentum in the direction of a decrease in the ℓ value, that is, from the earth's equator to the poles. As a result, in the low latitudes the moment of momentum decreases (easterly winds develop), whereas in the temperate latitudes the moment of momentum increases (westerly winds appear). At latitudes ±35° the ℓ value is equal to the average value for the entire atmosphere. Accordingly, here at the time of mixing the moment of momentum does not change and no wind arises.

With appearance of winds the forces of microturbulent friction of the air against the earth's surface arise, that is, vertical fluxes of the moment of momentum arise. In zones of easterly winds the moment of momentum flows from the earth to the atmosphere, whereas in zones of westerly winds, vice versa, from the atmosphere to the earth. The vertical flux of the moment of momentum weakens the wind. The greater the wind velocity, the greater is the vertical flux and the greater is the weakening of the wind. The weakening of wind velocity continues until the vertical microturbulent flux of the moment of momentum comes into equilibrium with the horizontal macroturbulent flux in the direction of the meridian. Only then does a stationary state set in and will the wind velocity not change with time. The geometry of the zones of influx and loss of the moment of momentum in the atmosphere is different. As a result, the stationary state of zonal circulation is attained only after accumulation of a definite positive moment of momentum in the atmosphere, which is transferred from the earth to the atmosphere.

BIBLIOGRAPHY

1. Gandin, L. S., Laykhtman, D. L., Matveyev, L. T., Yudin, M. I., OSNOVY DINAMICHESKOY METEOROLOGII (Principles of Dynamic Meteorology), Leningrad, Gidrometeoizdat, 1955.

2. DINAMICHESKAYA METEOROLOGIYA. CHAST' II (Dynamic Meteorology. Part II), edited by B. I. Izvekov and N. Ye. Kochin, Moscow-Leningrad, Gidrometeoizdat, 1937.

3. Dymnikov, V. P., Perov, V. L., Lykosov, V. N., "Hydrodynamic Zonal Model of General Circulation of the Atmosphere," IZV. AN SSSR. FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), Vol 15, No 5, 1979.

4. Lorentz, E. N., PRIRODA I TEORIYA OBSHCHEY TSIRKULYATSII ATMOSFERY (Nature and the Theory of General Circulation of the Atmosphere), Leningrad, Gidrometeoizdat, 1970.

5. Munk, W., MacDonald, G., VRASHCHENIYE ZEMLI (The Earth's Rotation), Moscow, Mir, 1964.

6. Sidorenkov, N. S., "Investigation of the Atmospheric Moment of Momentum," IZV. AN SSSR, FIZIKA ATMOSFERY I OKEANA, Vol 12, No 6, 1976.

7. Sidorenkov, N. S., "Tensor of Atmospheric Inertia, Annual Changes in its Components and Variations of the Earth's Rotation," IZV. AN SSSR, FIZIKA ATMOSFERY I OKEANA, Vol 9, No 4, 1973.

8. Starr, V., FIZIKA YAVLENIY S OTRITSATEL'NOY VYAZKOST'YU (Physics of Phenomena With Negative Viscosity), Moscow, Mir, 1971.

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DETERMINATION OF WIND VELOCITY AND DIRECTION AND TEMPERATURE IN THE ATMOSPHERIC SURFACE LAYER BY THE RADIOACOUSTIC SOUNDING METHOD

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 36-45

[Article by S. I. Babkin, G. N. Miloserdova, Candidate of Physical and Mathematical Sciences M. Yu. Orlov, Yu. I. Pakhomov, Candidate of Technical Sciences Ye. G. Proshkin, Yu. N. Ul'yanov and B. S. Yurchak, Institute of Experimental Meteorology, submitted for publication 4 Mar 80]

[Text]

Abstract: A study was made of the influence exerted on the accuracy of measurements of temperature and wind velocity in the atmospheric surface layer which are sensed by the radioacoustic sounding (RAS) method, as well as the accuracy of the speed of sound, angular coordinates and air humidity. A comparison is made of the results of measurements of the temperature and wind velocity profiles by the RAS method with similar data obtained simultaneously with sensors mounted on the high meteorological mast of the Institute of Experimental Meteorology. The results obtained by the two methods coincide within the limits of error. The possibility of using a priori information on atmospheric parameters in making measurements by the RAS method is investigated. It is demonstrated that the use of such information makes possible an appreciable improvement in the accuracy of the results. Recommendations are given on the choice of the directions for multiple sounding which are best from the point of view of the accuracy of the results.

Introduction. The radioacoustic sounding (RAS) method is based on measurement of the velocity of propagation of a sound pulse in the air by a Doppler radar [7].

The speed of sound in the air is

$$c = a\sqrt{T} + V_r, \tag{1}$$

where T is absolute temperature, V_r is the projection of the velocity of motion of air (wind) on the direction of sound propagation, a is a coefficient dependent on the gas composition. For dry air a = 20.0528 $m \cdot sec^{-1} \cdot K^{-1/2}$.

When measuring the speed of sound in the atmosphere in not less than four different directions by means of the RAS method there is a possibility of simultaneous determination of the profiles of temperature (T) and all wind components (V_x, V_y, V_z) [4, 5, 8].

The RAS method can be used in the following variants:
-- vertical sounding for measuring the temperature profile [7];
-- sounding in four directions for simultaneous determination of the profiles of temperature and the wind components [8];
-- sounding in several different directions ($n > 4$) for determination of the profiles of temperature and the wind components by the least squares method.

Vertical Sounding

In vertical sounding for determining the temperature profile the accuracy of the temperature value will be determined using the formula

$$\Delta T = 2\frac{T}{a}\sqrt{(\Delta a)^2 + \frac{(\Delta c)^2}{T} + \frac{(\Delta V_z)^2}{T} + \frac{(V_x^2 + \Delta V_x^2)}{T}(\Delta\delta)^2}. \quad (2)$$

Here Δa is the accuracy in determining the proportionality factor in the Laplace formula (see above); Δc is the accuracy in determining the speed of sound; Δv_z is the accuracy in determining the vertical wind component; V_x and ΔV_x are the horizontal wind velocity and the error in its determination.

The assumption that the vertical wind component is equal to zero is introduced in the case of vertical sounding. Thus, the uncertainty in ΔV_z is a measure of the accuracy with which this assumption is satisfied.

The fourth term in the radicand of formula (2) determines the contribution to the accuracy in measuring temperature from the accuracy in determining the horizontal component of wind velocity and the accuracy in determining δ -- the angle between the vertical and the direction of sounding. It will be assumed that $\sin \delta = \Delta\delta = \delta$.

The coefficient a for dry air is known with a very high accuracy. If it is assumed that relative humidity in the lower layer of the atmosphere is equal to (75±25)% (as is satisfied in most cases in the middle zone of the USSR), the a value can be assumed equal to 20.10±0.02 [4].

The accuracy in measuring the speed of sound with a Doppler radar is assumed to be equal to ±0.1 m/sec [8]. If the vertical wind component is known with an accuracy to ±0.3 m/sec, the accuracy in determining temperature in a vertical sounding regime will be ±0.8°C. If the uncertainty in the vertical wind component (according to data from independent measurements) is equal to zero, the accuracy in measuring temperature is ±0.65°C.

The presence of vertical or horizontal wind velocity at the time of sounding leads to a systematic error in estimating temperature. Computations using formula (1) show that with a deviation of the direction of sounding from the vertical by ±1°, a = 20.10, T = 273 K

$$\frac{\partial T}{\partial V_z} = -1.6\ K/(\text{m/sec}) \text{ and } \frac{\partial T}{\partial V_x} = -0.0287\ K/(\text{m/sec}),$$

that is, the presence of an ascending flow with a velocity of about 10 cm/sec or a moderate horizontal wind with a velocity 7 m/sec leads to an exaggeration of the measured temperature by approximately 0.2 K. The horizontal wind will exert an influence on the measured temperature value only in a case when the sounding direction differs from the vertical.

Strictly speaking, such a situation exists only when the radar antenna has a very narrow directional diagram. For real directional diagrams, even in the case of rigorously vertical sounding, the presence of a horizontal wind will exert an influence on the accuracy in measuring temperature. However, an examination of this factor, dependent on design of the apparatus, is beyond the scope of this article.

Thus, vertical sounding makes it possible to determine the temperature profile but requires the use of additional (a priori) information for increasing the accuracy of the results or for introducing corrections.

An improvement in the accuracy of the temperature value can be achieved, in particular, with a decrease in the uncertainty of the a coefficient by means of use of additional information on air humidity in the place where the temperature is measured. A decrease in the uncertainty of the a value makes possible a doubling of the accuracy of the temperature value (to ±0.45°C). Further improvement in accuracy requires an improvement in the accuracy in measuring the speed of sound and a decrease in the uncertainties of the δ and V_x values.

Sounding in Four Directions

Formulation of problem. In more detailed form formula (1) for the speed of sound propagation in the direction with the number i will read as follows:

$$c_i = a\sqrt{T} + u_i V_x + v_i V_y + w_i V_z. \qquad (1')$$

where u_i, v_i, w_i are the direction cosines of the vector of the direction i.

In carrying out sounding in directions indicated in Fig. 1 the system of four equations for determining the four unknowns ($x_1 = \sqrt{T}$; $x_2 = V_x$; $x_3 = V_y$; $x_4 = V_z$) in matrix form is written as follows:

$$\vec{C} = M\vec{X}. \qquad (3)$$

Here M is the matrix of the system of equations and

$$M = \begin{Vmatrix} a & \sin\delta_1\cos\alpha_1 & \sin\delta_1\sin\alpha_1 & \cos\delta_1 \\ a & \sin\delta_2\cos\alpha_2 & \sin\delta_2\sin\alpha_2 & \cos\delta_2 \\ a & \sin\delta_3\cos\alpha_3 & \sin\delta_3\sin\alpha_3 & \cos\delta_3 \\ a & \sin\delta_4\cos\alpha_4 & \sin\delta_4\sin\alpha_4 & \cos\delta_4 \end{Vmatrix}. \qquad (4)$$

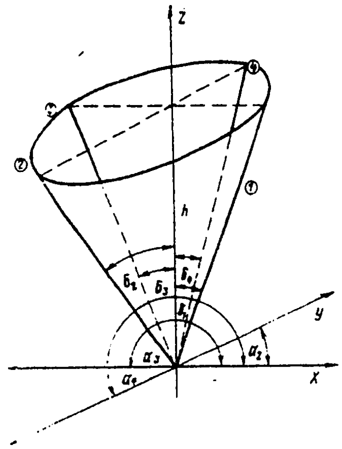

Fig. 1. Directions of sounding with simultaneous determination of temperature and wind components (four soundings).

The angles δ_i are the angles between the direction of sounding and the vertical (z-axis), the angles α_i are the angles between the projections of the sounding directions onto the (x, y) plane and the positive direction of the x-axis.

Similar expressions can be written as well for a case when the sound pulse irradiated by the radar is propagated vertically along the z-axis and the electromagnetic radiation reflected from it is received by four separate antennas with axes inclined at some angles relative to the z-axis. When making soundings in several directions the resulting values for temperature and wind direction are averaged for the area of the base of a cone with the radius H tg δ , and the measurements in individual directions are not made simultaneously. However, with the propagation of a sound pulse in a vertical direction and reception of the reflected radiation in four directions the measurements give the instantaneous values of temperature and the wind components at one point at a given altitude.

The parameters whose errors in determination will exert an influence on the accuracy of determination of temperature and the wind components will be

$$p_i = \begin{cases} c_j \ (i=j=1,2,3,4) \\ \delta_k \ (i=5,6,7,8;\ k=1,2,3,4) \\ \alpha_l \ (i=9,10,11,12,;\ l=1,2,3,4) \\ a \ (i=13) \end{cases}.$$

The solution of equations (3) is written in the form

$$\vec{X} = M^{-1}\vec{C}. \tag{5}$$

The accuracies of the values of components of the $\vec{X}$ vector can be determined from the diagonal terms of the covariation matrix

$$D(\vec{X}) = \frac{\partial \vec{X}}{\partial \vec{p}} D(\vec{p}) \frac{\widetilde{\partial \vec{X}}}{\partial \vec{p}}. \tag{6}$$

Here $\partial\vec{X}/\partial\vec{p}$ is the matrix of derivatives of the values of the components of the $\vec{X}$ vector for the $\vec{p}$ parameters; $D(\vec{p})$ is the covariation matrix of the $\vec{p}$ parameters; the symbol $\sim$ denotes transposition.

The matrix $\partial \vec{X}/\partial \vec{p}$ was obtained numerically, by solution of system (3) with the introduction, in turn, of small increments $\Delta p_i \ll p_i$ in the values of all the p_i parameters. In this case no assumption is made concerning the smallness of the changes in elements of the inverse matrix M^{-1} with small changes of the Δp_i parameters, such as adopted in obtaining the results in [4].

The errors in determining the values for temperature and the wind components can be computed using the formulas

$$\Delta T = 2x_1 \Delta x_1 = 2\sqrt{Td_{11}(\vec{X})},$$
$$\Delta V_x = \sqrt{d_{22}(\vec{X})};\quad \Delta V_y = \sqrt{d_{33}(\vec{X})};\quad \Delta V_z = \sqrt{d_{44}(\vec{X})}. \tag{7}$$

In a case when in the measurements it is necessary to obtain the wind modulus and its direction, other than the wind components,

$$|V| = \sqrt{x_2^2 + x_3^2};\quad \beta = \operatorname{arctg}\frac{x_2}{x_3}. \tag{8}$$

The errors in determining $|V|$ and β are easily computed from the known values of the errors in the wind components (with allowance for the nondiagonal terms of the $D(\vec{X})$ matrix).

Choice of values of parameters and their uncertainty. Computations of the values of the errors were made with some typical values of temperature and the wind components. It was assumed that the temperature at the sounding point was 273 K; the wind components were $V_x = V_y = 10$ m/sec and $V_z = 0.1$ m/sec.

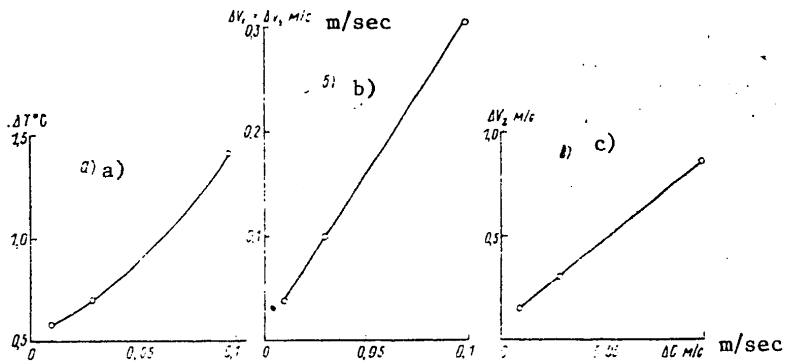

Fig. 2. Dependence of accuracy in determining temperature (a), horizontal wind components (b) and vertical wind component (c) on accuracy in measuring speed of sound (four soundings).

On the basis of available data on uncertainties of the parameters used in the measurements (coefficient, speed of sound, angles in the horizontal and vertical planes) the principal computations were made with the following values of the uncertainties: $\Delta a = \pm 0.02$ m·sec^{-1} K$^{-1/2}$ [4]; $\Delta\delta_i = \Delta\delta_i = \pm 0.001745$ (1°); $\Delta c = \pm 0.1$ m/sec.

[redacted]

The value of the uncertainty Δa was discussed above. The accuracy in determining the angular coordinates is ensured by modern instrumentation. The components of the uncertainty Δc are dependent on the instruments and apparatus used in the method and therefore a detailed analysis of their values is not given here. We will only point out that for a RAS system constructed at the Khar'kov Institute of Radioelectronics [2], the principal components of uncertainty in the measured speed of sound value are caused by the following phenomena or errors:
a) the error with which the velocity of propagation of electromagnetic waves in the atmosphere is known;
b) the short- and long-term frequency stability of the Doppler radar transmitter;
c) the error in measuring the Doppler frequency, which in turn is dependent on the frequency distortions of the echo signal in its processing by the receiver and tracking filter, the errors in measuring the Doppler frequency and the width of the Doppler spectrum of echo signal frequencies.

Results of computations and experiments. The contributions of individual components of errors to the total uncertainties of the measured values were evaluated.

The dependence of the total uncertainties of temperature and the wind components on the accuracy in measuring the speed of sound is given in Fig. 2. Our computations also made it possible to select the optimum (for obtaining the most precise results) combinations of sounding directions. The best accuracy for temperature ($\Delta T = \pm 1.4$ K) and the vertical wind component ($\Delta V_z = \pm 0.85$ m/sec) is obtained with with a combination of the angles $\delta_1 = \delta_3 = 30°$; $\delta_2 = \delta_4 = 10°$. The components V_x and V_y are determined with an accuracy to ± 0.3 m/sec. In this case the points for which the speeds of sound are determined at a particular altitude are situated on an ellipse with the ratio of the semimajor and semiminor axes equal to $a_1/a_2 = 2.8$; the semimajor axis of this ellipse is oriented along the vector of the horizontal wind.

With the combination of angles $\delta_1 = \delta_2 = \delta_3 = 30°$; $\delta_4 = 0°$, which in routine work is ensured far more simply, the error in determining temperature in this case increases by 6% (to ± 1.49 K) and the error ΔV_z -- by 10% (to ± 0.92 m/sec). There is also a 50% improvement in the accuracy in determining the horizontal wind components (to ± 0.2 m/sec).

Thus, it is possible to recommend the following sequence for carrying out simultaneous measurements by the RAS method when using four soundings ensuring the best accuracy of the results:

1. Preliminary measurements with an arbitrary orientation of the directions of sounding relative to the direction of the horizontal wind.

2. On the basis of the results of these measurements, the choice of the orientation of directions ensuring the best accuracy of the results. The principal measurements are also made with this orientation.

The dependence of the accuracy in determining temperature on the choice of sounding angles is shown in Fig. 3.

It should be noted that a case when the sounding directions coincide with the generatrices of a circular cone is exceedingly unfavorable with respect to obtaining a high accuracy of the results, specifically, when the axis of the cone is inclined by a small angle to the vertical.

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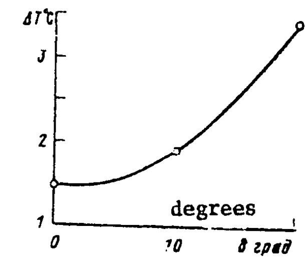

Fig. 3. Dependence of the accuracy in determining temperature on the value of the angle δ_4 in the vertical plane (Δ c = ±0.1 m/sec; $\delta_1 = \delta_2 = \delta_3 = 30°$; four soundings).

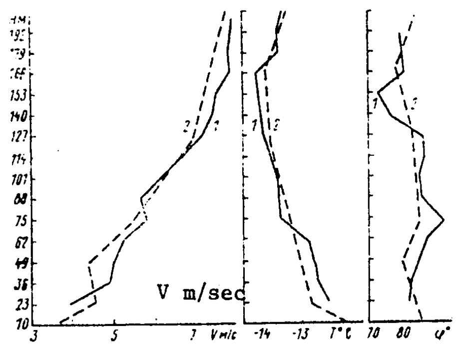

Fig. 4. Comparison of results of determination of temperature, modulus and direction of wind by the RAS method (1) and sensors on high meteorological mast (2).

The results of measurements of temperature and the horizontal wind components in the case of multiple sounding are given in Fig. 4. In this same figure, as a comparison, we show the profiles of temperature and wind obtained during measurements by the RAS method using the set of sensors on the high meteorological mast (HMM) of the Institute of Experimental Meteorology.

The measurements were made using radioacoustic sounding apparatus developed at the Khar'kov Institute of Radioelectronics. The operating principle for this instrumentation and the principal technical specifications were described in [2].

As already noted in [4], the presence of correlations of errors of the parameters exerts considerable influence on the results. An analysis of the structure of errors in measuring the speed of sound indicated that the correlation coefficients between the errors are not equal to zero because the individual components of errors for measurements in different directions are the same. A similar situation also exists for errors in determining angles in the vertical plane and the angles in the horizontal plane, that is

$$\overline{\Delta c_i\, \Delta c_j}\,|_{i\neq j} > 0;\quad \overline{\Delta\delta_i\, \Delta\delta_j}\,|_{i\neq j} > 0;\quad \overline{\Delta\alpha_i\, \Delta\alpha_j}\,|_{i\neq j} > 0;$$
$$\overline{\Delta c_i\, \Delta\delta_j} = \overline{\Delta c_i\, \Delta\alpha_j} = \overline{\Delta\delta_i\, \Delta\alpha_j} = 0,$$
$$\overline{\Delta c_i\, \Delta a} = \overline{\Delta\delta_i\, \Delta a} = \overline{\Delta\alpha_i\, \Delta a} = 0.$$

The evaluations show that with correlation coefficients $\rho(c_i c_j) = \rho(\delta_i \delta_j) = \rho(\alpha_i \alpha_j) = 0.5$ the accuracy of the results is appreciably better (by a factor of approximately 1.5).

Multiple Sounding

One of the possible modifications of the multiple sounding method is sounding in several ($n > 4$) directions. In this case the method for processing the results will be similar to that which is used in the Manning and Greenhow quasi-multiray methods [1] for determining the wind in the upper atmosphere from the movement of meteor trails.

In this case the matrix of the system of equations for multiple sounding has the form

$$M = \left\| \begin{array}{cccc} na^2 & a\sum_{i=1}^{n} u_i & a\sum_{i=1}^{n} v_i & a\sum_{i=1}^{n} w_i \\ a\sum_{i=1}^{n} u_i & \sum_{i=1}^{n} u_i^2 & \sum_{i=1}^{n} u_i v_i & \sum_{i=1}^{n} u_i w_i \\ a\sum_{i=1}^{n} v_i & \sum_{i=1}^{n} v_i u_i & \sum_{i=1}^{n} v_i^2 & \sum_{i=1}^{n} v_i w_i \\ a\sum_{i=1}^{n} w_i & \sum_{i=1}^{n} w_i u_i & \sum_{i=1}^{n} w_i v_i & \sum_{i=1}^{n} w_i^2 \end{array} \right\|, \qquad (8')$$

and the column of free terms will consist of the following values:

$$S = \left\| \begin{array}{c} a\sum_{i=1}^{n} c_i \\ \sum_{i=1}^{n} c_i u_i \\ \sum_{i=1}^{n} c_i v_i \\ \sum_{i=1}^{n} c_i w_i \end{array} \right\|. \qquad (8'')$$

Use of A Priori Information on Temperature and Wind Values

The examination made above was made on the assumption that when making measurements of temperature and wind there are no data concerning them, or at least, a priori information was taken into account in implicit form, as in the case of vertical sounding for determining temperature.

However, the experience accumulated in meteorology makes it possible, prior to the onset of measurements, to obtain some information concerning the temperature and wind at different altitudes (for example, see [3]).

Now we will examine the possibility of the use of a priori information in determining temperature and the wind components if the RAS method is used in measuring k values $c_k(T, \vec{V})$ (speed of sound). One of the possibilities of such use is associated with an approach based on the Bayes theorem and set forth in [6].

Assume that prior to the onset of measurements of temperature and wind velocity at some altitude we know their a priori values: $t_1 = \sqrt{T}$; t_2, t_3, t_4 -- the wind components, V_x, V_y, V_z; t_j (j = 5, 6, 7, 8) -- the angles of the directions of sounding relative to the vertical; t_l (l = 9, 10, 11, 12) -- the angles between the direction of sounding and the x-axis in the horizontal plane; t_{13} — the value of the a coefficient. We will also assume that the a priori covariation matrix R of these values is known. We will make measurements of k values of the speed of sound c_k and determine their covariation matrix V.

Then, using the Bayes theorem, the assumption of a Gaussian distribution of all the parameters and the measured values and a linear representation of changes in the c_k values in dependence on the parameters t_i in some region of change in the parameters

$$c_k(t_i) = \overline{c}(\overline{t_i}) + \sum_j g_{kj}(t_j - \overline{t_j}), \tag{9}$$

we will find the corrections to the a priori values of the parameters and the a posteriori matrix of errors of the parameters R' [6]:

$$(R^{-1} + \tilde{g}V^{-1}g)(\overline{T} - T) = \tilde{g}V^{-1}(C - \overline{C}). \tag{10}$$

$$R' = R - R\tilde{g}(gR\tilde{g} + V)^{-1}gR. \tag{11}$$

It follows from the positive determinancy of the R and V matrices that with satisfaction of procedure (11) the diagonal elements of the a posteriori covariation matrix R' (a posteriori dispersions t_i) will be less than the mentioned diagonal elements of the a priori covariation matrix (a priori dispersions t_i). Thus, there is a refinement of the a priori values of the t_i parameters.

The system of equations (10) makes it possible to determine the corrections to the a priori values of the parameters t_i, taking into account the information obtained in the experiment.

As an illustration of the possibilities of the RAS method with the use of a priori information, Fig. 5 shows the temperature profile determined by such a method in the case of vertical sounding (curve 3). As a comparison, in this same figure we show the temperature profiles obtained using HMM sensors (curve 1) and by means of the traditional RAS method (curve 2). The estimated uncertainties in temperature values are also cited.

In obtaining the temperature profile, with a priori information taken into account, it was assumed that prior to the onset of measurements we knew the relative humidity value at the ground level with an accuracy to ±10%, the temperature at the ground level with an accuracy to ±3°C and the vertical wind component at the ground level with an accuracy to ±0.3 m/sec. It was also assumed that in the lower 300-m layer of the atmosphere the relative humidity changes with altitude by not more than 10%, the vertical wind component does not change with altitude, the temperature decreases linearly with altitude and its gradients can be taken from the the data in [3] for the corresponding meteorological conditions. The accuracies in the a priori values of the a coefficient, temperature T at an altitude of 150 m and the vertical wind V_z are 0.005 $m \cdot sec^{-1}/K^{-1/2}$, 1°C and 0.3 m/sec.

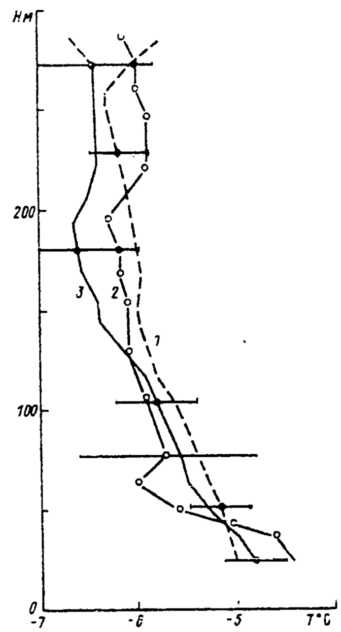

Fig. 5. Comparison of temperature profile measured with HMM [high meteorological mast] sensors (1), by the RAS method in vertical sounding (2) and by RAS method with the use of a priori information on humidity, temperature and vertical wind component (3).

The accuracy in measuring the speed of sound was assumed equal to 0.3 m/sec. The data cited in Fig. 5 show that the temperature values obtained using the sensors on the high meteorological mast, the RAS method and the RAS method with the use of a priori information coincide with one another in the limits of their estimated uncertainties. Computations on the basis of formula (2) on the assumption of absence of a priori information gives for the temperature values obtained by the RAS method an uncertainty of about ±0.9°C (see above). Thus, allowance for a priori information with an insignificant complicating of the processing procedure makes possible an appreciable improvement in the accuracy of the results without expending efforts on the improvement of instrumentation for improving the accuracy in measurement of the speed of sound.

BIBLIOGRAPHY

1. Babadzhanov, P. B., et al., RADIOMETEORNYYE ISSLEDOVANIYA TSIRKULYATSII VERKHNEY ATMOSFERY (Radiometeor Investigations of Circulation of the Upper Atmosphere), Dushanbe, Donish, 1974.

2. Babkin, S. I., et al., "System for Radioacoustic Sounding of the Atmosphere in the Centimeter Wavelength Range," TEZISY DOKLADOV NA V VSESOYUZNOM SIMPOZIUME PO LAZERNOMU I AKUSTICHESKOMU ZONDIROVANIYU ATMOSFERY (Summaries of Reports at the Fifth All-Union Symposium on Laser and Acoustic Sounding of the Atmosphere), Part 3, Tomsk, 1978.

3. Mashkova, G. B., "Peculiarities of Stratification of Air Temperature in the Lower 300 m of the Atmosphere," TRUDY IEM (Transactions of the Institute of Experimental Meteorology), No 10(53), 1975.

4. Orlov, M. Yu., Yurchak, B. S., "Accuracy in Measuring Temperature and the Wind Components by the Radioacoustic Method," TRUDY IEM, No 19(72), 1978.

5. Stepanenko, V. D., RADIOLOKATSIYA V METEOROLOGII (Radar in Meteorology), Leningrad, Gidrometeoizdat, 1973.

6. Dragt, J. B., "Statistical Considerations on Techniques for Adjustment of Differential Cross Sections With Measured Integral Parameters," TRUDY TREKHSTORONNEGO SOVETSKO-BEL'GIYSKO-GOLLANDSKOGO SIMPOZIUMA PO NEKOTORYM VOPROSAM FIZIKI BYSTRYKH REAKTOROV (Transactions of the Trilateral Soviet-Belgian-Dutch Symposium on Some Problems in the Physics of Fast Reactors), Report R-28, Melekess, Moscow, Atomizdat, 1970.

7. North, E. M., et al., "RASS, a Remote Sensing System for Measuring Low-Level Temperature Profile," BULL. AMER. METEOROL. SOC., Vol 54, No 9, 1973.

8. Marshall, J. M., et al., "Combined Radar-Acoustic Sounding System," APPL. OPTICS, Vol 11, 1972.

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ORGANOCHLORINE PESTICIDES IN PRECIPITATION

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 46-51

[Article by Candidate of Technical Sciences Ts. I. Bobovnikova and A. V. Dibtseva, Institute of Experimental Meteorology, submitted for publication 2 Jan 80]

[Text]

Abstract: Contamination of precipitation by organochlorine pesticides (DDT, DDE, DDD, α- and γ-BHC) is examined. The article gives the results of investigations carried out in the Moscow region in 1978-1979. It is shown that despite the ban or sharp restriction on the use of DDT in the well-developed countries in the early 1970's there has been no appreciable decrease of DDT in precipitation. The fallout of DDT with rain exceeds the fallout of DDT with snow by a factor of 5-6 and the fallout of γ-BHC in rain exceeds its fallout in snow by a factor of 10-20.

Considerable quantities of pesticides after their use for contending with agricultural pests and weeds enter the atmosphere. This is attributable, first of all, to the fact that the pressure of the vapors of poisonous chemicals is sufficiently great for their evaporation from the soil surface, water and leaves [6]. In addition, a rather high percentage of the pesticides does not reach the surface to be protected even in the process of their application. For example, according to data published by Taylor [13], in the application of dieldrin and heptachlor on fields covered with grass by means of a surface spraying apparatus 60% of the dieldrin and 42% of the heptachlor did not reach the surface but remained in the atmosphere. Pesticides can be present in the atmosphere in the form of vapor or aerosols [5]. Their fate thereafter is dependent on the size of the particles and meteorological conditions. Reference for the most part is to organochlorine pesticides and some herbicides which are rather resistant to photochemical reactions and which therefore can be present for a long time in the atmosphere. Most of the organophosphorus pesticides, as a result of their small persistence, decompose rapidly and it is difficult to trace their further behavior.

The atmosphere is therefore a potential reservoir for the transport of pesticides. Pesticides can be washed out from it by precipitation in the form of rain and snow and the soil and surface waters can be recontaminated. According to [16], the principal mechanism by which pesticides are washed from the atmosphere is rain.

Many scientists in different countries have determined the concentrations of pesticides in samples of air and rain water taken both at the places where they have been applied and in clean regions. The concentrations of pesticides in the air vary in a wide range — from $1 \cdot 10^{-15}$ to $1 \cdot 10^{-3}$ g/m^3, depending on the time and place of the sampling. Higher concentrations have been noted in the regions of application of the poisonous chemicals and low concentrations are associated with global residue [6]. The content of pesticides in precipitation was first measured in the 1960's, when their application, especially the use of DDT, attained the maximum scale [16].

In 1965 Weibel, et al. (United States) determined the content of organochlorine pesticides (DDT, DDE, chlordan, dieldrin, etc.) and the herbicide 2,4,5-T in samples of rain falling in a rural area 500 feet from the place of application of poisonous chemicals [14]. The DDT concentration in this rain was 0.6 μg/liter, DDE -- 0.2, chlordan -- 0.5, dieldrin — 0.003, 2,4,5-T -- 0.04 μg/liter.

In Great Britain studies of the contamination of precipitation by organochlorine pesticides (OCP) began in 1964. Wheatly and Hardman [15] in the spring of 1964 decided to clarify the possibility of detecting OCP in the atmosphere and therefore in rain water in order to explain their presence in uncultivated soils at Wellesbourne. The investigations were made at the aerometeorological station at Wellesbourne. A copper rain gage with a diameter of 8" was set out in a sector overgrown with grass. Rain water was collected in a glass container with a volume of 2.5 liters. Each month the container was changed. The observations were made from April 1964 through February 1965. Individual rains were analyzed sometimes. The determination was made by the methods of thin-layer and gas-fluid chromatography. The concentrations of aldrin, dieldrin, γ-BHC and DDT were at the level of tenths and hundredths of μg/liter, that is, at the limit of response in determining pesticides. On the basis of data on the use of OCP and the rate of their elimination from the soil the authors calculated the residues of aldrin, dieldrin, γ-BHC and DDT in the soil. These were 1, 8, 2 and 30 tons respectively, which probably were distributed in the upper 5 mm of all the occupied agricultural areas in England and Wales.

Seven stations for the collection of precipitation have functioned in England since 1966. Rain water was collected in a glass container which was painted yellow in order to decrease photochemical decomposition. Samples of rain water collected during August 1966–July 1967 were found to contain organochlorine pesticides (dieldrin, p, p'-DDT, o,p'-DDT, p, p'-DDE, α- and γ-BHC) in concentrations equal to tenths and hundredths of μg/liter. Great differences in the content of OCP in dependence on the place of sampling were not discovered.

Systematic measurements of pesticides in the air and in rain water at five background and two standard (for comparison) stations have been made in West Germany since 1970 [7, 9]. Precipitation was collected in a 10-liter glass container through a glazed funnel 30.5 cm in diameter (intake area 0.730 m^2). The samples were taken monthly and were analyzed by the gas-fluid chromatography method [8].

Aerosol particles of air were found to contain DDT, dieldrin and lindane in concentrations of about 1 ng/m^3; in rain water it averages 10 ng/g or more. Despite the banning of the use of DDT in West Germany since 1970 tnere has been no significant decrease in the insecticide in the air and precipitation. In 1971 the DDT content in rain was about 400 ng/liter and the content of γ-BHC was about 300 ng/liter [9].

In Japan special attention has been devoted to the study of hexachloran in the environment, especially in rain. This was attributed to the fact that at the end of the 1960's BHC was used in great quantities. The precipitation collector was situated on the roof of a building at Tokio University, in a city area with no nearby fields. The samples were taken in dark five-liter bottles, from December 1968 to November 1969 [10].

The concentration of α-BHC was from 45 to 830 ng/liter and the concentration of γ-BHC was from 29 to 398 ng/liter. Seasonal variations in the concentrations of α- and γ-BHC were observed. The maximum values were discovered in the summer months and the minimum values were observed in winter, which was associated with the use of pesticide in summer. However, the contamination of rain water with BHC was observed throughout the entire year. This fact gives basis for assuming that the principal portions of BHC entering into the atmosphere fall onto the earth in the course of a short time; the remaining quantity remains in the atmosphere and gradually falls out over the course of a prolonged period.

During 1973-1974 Gitsova in Bulgaria [3] investigated the content of organochlorine pesticides in natural waters, including in rain water. The precipitation samples were collected at the hygiene center in Sofia over a period of 13 months (October 1973 - November 1974). The average concentrations of γ-BHC were 35 ng/liter, and for p,p'-DDT — 68 ng/liter.

According to the data in [11], the quantity of DDT used on the earth was about 100,000 tons, of which up to 40% of the DDT was used for contending with malaria and the remaining 60% for contending with domestic insects and agricultural pests. DDT is used for the most part in the developing countries (India, Republic of Chad, and elsewhere) to which 90% of the DDT produced in the well-developed countries (United States, West Germany, Japan, France and others) is exported.

In 1974 specialists at the Institute of Experimental Meteorology began work on study of contamination of the soils and rivers by residual quantities of organochlorine pesticides. In earlier studies we demonstrated that at the present time organochlorine pesticides are discovered in all objects of the biosphere -- soil, surface waters, fish and precipitation [2]. The discovery of DDT, its metabolites, α- and γ-BHC in the soils of preserves and in rivers flowing far from agricultural regions led to the idea of initiating systematic observations of the levels of contamination of precipitation. For this purpose we used an automatic collector of dry and wet fallout developed at the Institute of Experimental Meteorology by G. G. Belov [1]. The instrument consists of a precipitation indicator, actuating mechanism and two enameled receiving containers. The sensitivity of the precipitation indicator makes it possible to activate the actuating mechanism with the entry of raindrops or snowflakes with a diameter of 200 μm. From the moment of onset of the falling of precipitation the receiver of dry fallout (atmospheric dust) is tightly covered by a lid. Upon the ending of the rain or snow the lid covers the tank with precipitation and the tank for the collection of dry fallout is opened. The precipitation for the month is poured into a dark glass bottle and analyzed. With the falling of more than 2 mm of precipitation each rain is analyzed.

The precipitation is doubly extracted with distilled n-hexane; after purification and concentration of the extract the analysis is made with a gas-fluid chromatograph. The analysis error is 15%.

Snow falling on a backing over a month period is analyzed. For comparative purposes the snow is sampled at several points around the backing. After the snow is thawed in an enameled container the volume of the sample is determined and then it is analyzed using the method described in [4]. In determining organochlorine pesticides in precipitation an effort is made to reduce to a minimum the contamination from reagents and other sources within the laboratory. The container is carefully washed and all the reagents are purified and checked for chromatographic purity. In addition, blank experiments are made in each case.

Table 1 gives the mean monthly concentrations of organochlorine pesticides and their fallout with precipitation during 1978 and the nine months of 1979. The table shows that DDT and γ-BHC are discovered in all seasons of the year. On the average, the observed concentrations are at the level of tenths of ng/liter, specifically: DDT -- from 15 to 120 ng/liter, γ-BHC -- from 13 to 100 ng/liter. The content of the metabolite DDE in precipitation exceeded the quantity of DDT or was equal to it. It is known that DDE is a product of the transformation of DDT caused to a considerable degree by photochemical processes transpiring in the atmosphere.

During summer the transformation of DDT into DDE is accelerated appreciably. A confirmation of this fact is comparative data on the ratio of the concentrations of DDE/DDT in separately collected rains and precipitation collected during the month. Such an experiment with the use of two separate precipitation collectors was carried out in June and July 1979. Individual rains were collected in one collector for analysis and the monthly precipitation was collected in the other. At the end of the month the precipitation was poured out and analyzed for its OCP content. In addition, dry fallout was collected in the dry fallout collector. At the end of the month this collector was washed out with acetone and carefully wiped with gauze wetted with acetone. The pesticides were extracted from the wash fluid and marl by hexane.

The results obtained in this experiment are given in Table 2. The DDE/DDT ratio for individual cases of precipitation varied near 1 and did not exceed 1.5; for precipitation collected during the month it was 3.2-3.5 (Table 2). The transformation of DDT into DDE was still more appreciable in the sample of dry fallout in July when there were few days with precipitation (four days). The remaining time the collector of dry fallout was open; the mean air temperature was 17.6°C, which favored the photochemical transformation of DDT in contrast to a closed collector. In July the quantity of precipitation was maximum (147 mm), the number of days with precipitation was 17, and despite such a large number of days with precipitation the DDE/DDT

Table 1

Organochlorine Pesticides in Precipitation in Moscow Region During 1978-1979

Month	Количество осадков 1	Среднемесячная температура, °C 2	p,p'-DDT а	p,p'-DDT б	o,p'-DDT а	o,p'-DDT б	p,p'-DDD а	p,p'-DDD б	p,p'-DDE а	p,p'-DDE б	γ-BHC а	γ-BHC б	α-BHC а	α-BHC б
1978 г.														
3 Январь—март	97,8	—	15	1,10	Сл.	—	Сл.	—	38	2,58	18	0,7	18	—
4 Апрель	10,0	5,6	30	0,30	10	0,10	Сл.	—	65	0,65	120	2,62	Сл	—
5 Май	80,4	11,4	22	1,65	10	0,75	Сл.	—	27	2,02	35	2,62	30	2,25
6 Июнь	97,2	14,5	30	2,46	Сл.	—	Сл.	—	80	6,58	13	1,06	15	1,23
7 Июль	55,4	16,4	25	1,25	33	1,65	25	1,25	86	4,38	28	1,4	28	1,4
8 Август	58,0	16,2	35	1,85	50	2,65	20	1,06	75	3,97	45	2,4	20	1,0
9 Сентябрь	57,3	10,0	77	3,40	50	2,55	14	0,7	18	1,10	22	1,12	40	2,0
10 Октябрь	63,3	3,8	28	1,85	20	1,04	14	0,73	11	0,57	20	1,04	50	2,6
11 Ноябрь	50,3	2,5	28	1,40	20	—	Сл.	—	10	0,50	25	1,25	17	0,85
12 Декабрь	28,1	—14,3	36	0,42	Сл. 16	—	Сл.	—	40	0,91	20	0,40	20	0,4
16 **1979 г.**														
3 Январь	57,2	—9,9	55	1,9	Сл.	—	Сл.	—	23	0,98	25	0,85	30	1,3
13 Февраль	22,6	—9,1	66	0,97	Нет None	—	Сл.	—	36	0,54	Сл	—	Сл	—
14 Март	44,9	—0,9	30	0,75	Сл.	—	20	0,5	60	1,12	15	0,38	14	0,35
15 Апрель—май	14,9 18,7	3,7 16,8	66	1,98	Сл.		40	1,2	100	3,00	90	2,7	90	2,7
6 Июнь	35,5	17,6	55	3,02	17	0,6	17	0,6	280	10,00	60	2,17	70	3,45
7 Июль	147,1	16,8	30	3,53	20	2,35	10	1,17	55	6,48	30	3,52	30	3,52
8 Август	69,5		80	5,2	30	1,95	20	1,35	30	1,95	30	1,95	35	2,3
9 Сентябрь	65,7		60	4,56	10	0,6	Сл.		30	2,28	30	2,28	45	3,4

Note: a) mean monthly concentration, ng/liter; b) fallout, g/km^2

KEY:
1. Quantity of precipitation
2. Mean monthly temperature, °C
3. January-March
4. April
5. May
6. June
7. July
8. August
9. September
10. October
11. November
12. December
13. February
14. March
15. April-May
16. Trace

Table 2

Content of DDT and DDE in Individual Falling Rains and in Monthly Fallcut (Wet and Dry), ng/liter

Date of falling of rain		p, p'-DDT	p,p'-DDE	DDE/DDT
June	10	433	640	1.5
	27	170	200	1.18
	28	55	64	1.16
	30	40	37	0.93
Mean monthly		85	280	3.2
Dry fallout		73	1250	17.2
July	5-6	18	18	1.0
	7-8	30	30	1.0
	9, 10, 11, 12	58	66	1.14
	13	21	23	1.1
	14, 15	30	33	1.65
	18, 20	114	95	0.83
	23, 29	162	137	0.84
Mean monthly		25	80	3.2
Dry fallout		95	700	7.4

Table 3

Fallout of Organochlorine Pesticides With Rain and Snow, g/km^2

Pesticide	1978			1979 (9 months)		
	with rain	with snow	rain/snow	with rain	with snow	rain/snow
p, p'-DDT	12.53	2.1	5.98	21.7	3.62	5.15
p, p'-DDE	14.5	2.6	5.6	22.0	2.62	8.3
γ-BHC	13.8	0.7	19.8	13.4	1.23	10.9

ratio for dry fallout was 7.4. These results show that it is possible to give a correct estimate of the fallout of DDT from the atmosphere during summer when individual rains are collected and analyzed and that the principal mechanism for the transformation of DDT in the atmosphere is probably in the chain DDT-DDE. Such a transformation is characteristic for summer. During other seasons of the year the formation of DDE is considerably slower (Table 1).

With respect to another product of DDT metabolism, DDD, there are few conditions for its formation in the atmosphere. It is known that the transformation of DDT into DDD occurs for the most part in biological systems, both living and dead. The process of transformation of DDT into DDD is favored by anaerobic conditions, porphyrins of bivalent iron, etc. [17]. Accordingly, the DDD concentrations in precipitation are lower than for DDT and DDE. During winter DDD is either not observed at all or is discovered in concentrations below the instrument response.

During 1978 the following fell with precipitation: DDT -- 14.6 g/km^2, DDE -- 17.1 g/km^2, γ-BHC -- 14.5 g/km^2. In 1979 the quantity of falling organochlorine pesticides increased. For example, over a nine-month period the following quantities fell: p,p'-DDT -- 25.3 g/km^2, p,p'-DDE -- 24.6 g/km^2, γ-BHC -- 14.6 g/km^2.

The greatest quantity of OCP falls with rain. Table 3 shows the fallout of DDT, DDE and γ-BHC with snow and rain. Five or six times more DDT falls with rain than with snow; ten to twenty more times γ-BHC falls with rain than with snow.

The results show that at the present time, despite a sharp restriction on the use of DDT in the well-developed countries, the possibility of atmospheric contamination has not decreased. This is possibly attributable to the fact that the use of this pesticide in the countries situated in the southern latitudes is leading to its more intensive evaporation and secondary contamination of environmental objects due to precipitation.

BIBLIOGRAPHY

1. Belov, G. G., Author's Certificate No 558239 (USSR), "Instrument for Separate Sampling of Precipitation and Dry Fallout," BYULL. IZOBR. (Bulletin of Inventions), No 18, 1977.

2. Bobovnikova, Ts. I., Virchenko, Ye. P., Malakhov, S. G., "On the Problem of the Global Contamination of Natural Media by Organochlorine Pesticides," OPYT I METODY EKOLOGICHESKOGO MONITORINGA (Experience and Methods for Ecological Monitoring), Pushchino, 1978.

3. Gitsova, S., "Content of Organochlorines in the Natural Waters of Bulgaria," HIGIENA I ZDRAVEOPAZVANE, 18, No 5, 1975.

4. METODICHESKIYE UKAZANIYA PO KONTROLYU ZAGRYAZNENIYA POCHV (Methodological Instructions on Monitoring Soil Contamination), edited by S. G. Malakhov, Moscow, Gidrometeoizdat, 1977.

5. Abbott, D. C., Harrison, R. B., Tatton, J. O. G., Thompson, J., "Organochlorine Pesticides in the Atmospheric Environment," NATURE, Vol 208, 1965.

6. Belf, G., Liptak, --, "Pesticides, Herbicides and Other Economic Poisons," ENVIRON. ENG. HANDBOOK, Vol 11, 1974.

7. Hellman, H., "Pesticide in Gewassern," DTSCH GEWASSERN MITT., H 17, No 14, 1973.

8. Herzel, F., "Schnellverfahren zur Spurenbestimmung von Chlorkohlenwasserstöff-Insektiziden im Wasser," ARCH. HYG. BAKT, Vol 154, 1970.

9. Lahmann, E., Herzel, F., "Pestizind-Bestimmung in Luft und Niederschlagen," GESUNDHEITS-INGENIEUR, Bd 92, 1971.

10. Masahiro, O., Takahisa, H., "Alpha and Gamma BHC in Tokyo Rainwater," ENVIRON. POLLUT., Vol 9, No 4, 1975.

11. Monod, M. I., "Les Residus de Pesticides dans l'Environment," TRAV. SOCPHARM. MONTPELIER, Vol 37, No 18, 1977.

12. Tarrant, K. R., Tatton, J. G., "Organochlorine Pesticides in Rainwater in the British Isles," NATURE, Vol 219, No 5155, 1968.

13. Taylor, A. W., Glotfelty, D. E., Turner, B. C., Silver, R. E., Freeman, H. P., Weiss, A., "Volatilization of Dieldrin and Heptachlor Residues from Field Vegetation," J. AGRIC. FOOD CHEM., Vol 25, No 3, 1977.

14. Weibel, S. R., Weidner, R. B., Cohen, J. M., Christianson, A. G., "Pesticides and Other Contaminants in Rainfall and Runoff," AWWA, Vol 58, No 8, 1966.

15. Wheatly, G. A., Hardman, J. A., "Indication of the Presence of Organochlorine Insecticides in Rainwater in Central England," NATURE, Vol 207, No 4996, 1965.

16. Woodwell, J. M., Craig, P. P., Johnson, H. A., "DDT in the Biosphere. Where Does it Go?" SCIENCE, Vol 174, No 4014, 1971.

17. Zoro, J. A., Hunter, J. M., Eglinton, G., Ware, G. C., "Degradation of p',p'-DDT in Reducing Environments," NATURE, Vol 247, 1974.

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NATURAL COMPONENTS OF THE FIELD OF TOTAL OZONE CONTENT IN THE NORTHERN HEMISPHERE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 52-57

[Article by A. Ye. Kaygorodtsev and Candidate of Physical and Mathematical Sciences O. M. Pokrovskiy, Main Geophysical Observatory, submitted for publication 18 Dec 79]

[Text]

Abstract: It is demonstrated that 15-20 empirical orthogonal functions ensure the effective description of a broad spectrum of fluctuations of the field of total ozone content observed at stations in the ozonometric network in the northern hemisphere in both winter and summer. The latitudinal and zonal structure of the most important natural components for each of two seasons is analyzed. The regions corresponding to the maximum statistically independent variations of ozone and the discriminated natural components are indicated. The correlation between individual spatial anomalies in variations of the ozone field and the peculiarities of atmospheric dynamics is traced.

In addition to numerical modeling methods, an important means for investigating atmospheric processes is an analysis of the results of observations of the principal physical parameters of the atmosphere. Statistical algorithms for the processing of the results of observations have now acquired great popularity; among these a leading place is occupied by the natural components method [2].

A study of modern climatic variations involves investigation of a number of physical mechanisms, among which is the possible change in radiation-dynamic-photochemical processes in the stratosphere as a result of fluctuations in the concentration of its optically active components. Atmospheric ozone is among the latter. Accordingly, monitoring of the evolution of the global distribution of ozone is attracting the attention of many researchers. The prospects for a possible change in the structure and mass of the ozonosphere caused by anthropogenic factors is one of the reasons for the increasing interest in this problem. The basis for such investigations should be an analysis of data from observations made in the ozonometric network. At the present time the most complete data are mean monthly ozone values in a regular grid for both hemispheres for 1957-1967, published in [4]. It is clear that the planning and development of new observation systems, as well as the evaluation of their information yield, should be based on the use of accumulated observational data.

It is known [2] that the use of empirical orthogonal functions (EOF) ensures the most compact representation of statistical information on the variability of the components of geophysical fields. The EOF basis can be used not only for optimum approximation of this field for the purpose of carrying out an objective analysis of the results of current observations, but also in realization of the method for evaluating the information yield of the global system of observations and its optimization [3].

This article is devoted to an analysis of the results of computations of EOF for the horizontal field of total ozone content in the northern hemisphere according to the data in [4]. The computations were made separately for the summer and winter seasons.

Characteristics of Effectiveness of Approximation

The characteristics of computation of the eigenvalues $\{\lambda_k\}$ and the eigenvectors $\{P_k\}$ for a case when the number n of grid intersections exceeds the volume r of the sample are presented in [1] and are not considered in this article.

We will discuss the characteristics of effectiveness of an approximation using an EOF basis. The measure of the effectiveness of an approximation containing k elements of the EOF base ($k < n$) is described by the relative accumulated dispersion

$$\delta_k^2 = \sum_{i=1}^{k} \lambda_i^2 \Big/ \sum_{i=1}^{n} \lambda_i^2.$$

The $1 - \delta_k^2$ value represents the relative residual dispersion of the analysis relative to the total base and is an integral measure of effectiveness of the approximation. The diagonal elements of the residual matrix of covariations in the grid

$$\hat{R}_k = R - \sum_{i=1}^{k} \lambda_i^2 P_i \cdot P_i^T$$

give some idea concerning the spatial distribution of the error in the approximation using k elements of the EOF base.

Two seasons must be distinguished in the investigation of ozone. The ozone distribution in the winter season is characterized for the most part by the appearance of peculiarities of atmospheric circulation. In the summer the anomalies of the ozone field are attributable to the influence of both dynamic and photochemical factors. In the computations use was made of data for 1957-1967, applying to the periods December-April and June-October. Each of the two samples consisted of 50 series of observations. The dimensionality of the problem was determined by the number of grid points of intersection and was 162. The regular grid had a 10° interval in the range 5-85°N. The interval in longitude was 20°.

First we will examine the integral characteristics of effectiveness of approximation with the use of EOF. The table gives the eigenvalues λ_k and the relative accumulated dispersions δ_k^2 corresponding to the first 18 elements of the EOF base for each of the two seasons. The λ_k value is the mean amplitude of the k-th harmonic from the EOF base and therefore characterizes its contribution to the description of the natural variability of the ozone distribution over the hemisphere. The mean value (from the grid) of the natural variations of ozone content is described by the value

$$V = \sqrt{\mathrm{tr}(\Lambda)} = \sqrt{\sum_i \lambda_i^2}$$

During the winter-spring period the mean variations are $82 \cdot 10^{-3}$ cm, whereas in the summer-autumn period they are $51 \cdot 10^{-3}$ cm. Thus, the natural variations of ozone during winter are approximately 1 1/2 times greater than in summer. It must be remembered that the data which we have cited on the structure of variations of the ozone field to a considerable degree reflect the nonuniform character of the distribution of stations in the ozonometric network (Fig. 1). In both seasons there is a rapid convergence of the δ_k^2 values to 1 (see Table 1). The first nine EOF account for 95% of the dispersion of the sample in winter and 91% in summer. The next nine functions ensure only an increase in the mentioned values by 3 and 6% respectively. The table gives the decrease in values of the mean amplitudes λ_k to values of about $3 \cdot 10^{-3}$ cm, which is approximately 1% of the mean total ozone content. Thus, in general, a small number of EOF can satisfactorily reproduce a broad spectrum of ozone variations observed in the northern hemisphere by the existing network of ozonometric stations.

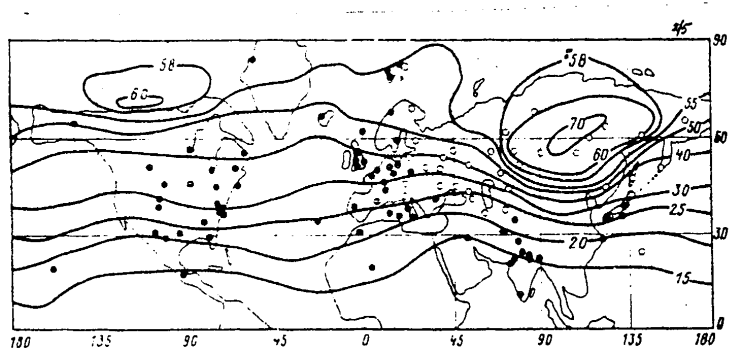

Fig. 1. Natural variations of field of total ozone content during winter season, in 10^{-3} cm. The dots and circles represent the location of stations in the ozonometric network.

Spatial Structure of Most Important EOF

First we will examine the winter-spring period. The distribution of standard deviations (see Fig. 1) is characterized by a quite uniform zonal structure. Exceptions are the Siberian and Canadian anomalies, corresponding to regions of stationary (standing) waves. It is possible to note some zonal nonuniformity in the region of the Northeast Atlantic, being an active synoptic zone, and also in the middle latitudes part of the Pacific Ocean coast of Asia, where there is an effect from an easterly jet stream. It can also be noted that the mentioned regions are supplied with data from the ozonometric network (see Fig. 1). Accordingly, it can be assumed that the detected anomalies reflect the real nature of variations of the ozone field.

Table 1

Characteristics of Approximation of Ozone Field by EOF

k	Зима—весна 1		Лето—осень 2		k	Зима—весна 1		Лето—осень 2	
	λ_k $(10^{-3}$ см$)$	δ_k^2	λ_k $(10^{-3}$ см$)$	δ_k^2		λ_k $(10^{-3}$ см$)$	δ_k^2	λ_k $(10^{-3}$ см$)$	δ_k^2
1	61,6	0,74	34,1	0,64	6	8,5	0,92	5,9	0,86
2	19,3	0,81	12,1	0,72	9	5,5	0,95	4,7	0,91
3	13,2	0,85	9,6	0,77	12	4,5	0,96	3,7	0,93
4	12,3	0,88	8,8	0,82	18	3,2	0,98	2,7	0,97
5	11,2	0,90	6,8	0,84					

KEY:
1. Winter-spring
2. Summer-autumn

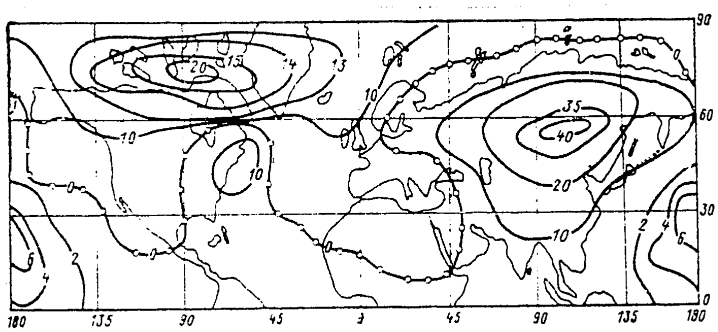

Fig. 2. Second natural component S_2 (winter season), 10^{-3} cm.

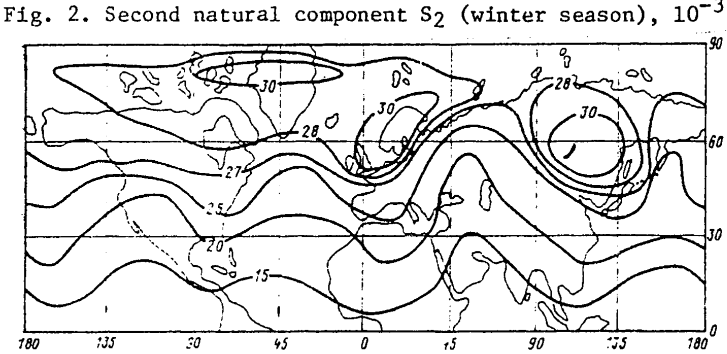

Fig. 3. Natural variations of field of total ozone content (summer), in 10^{-3} cm.

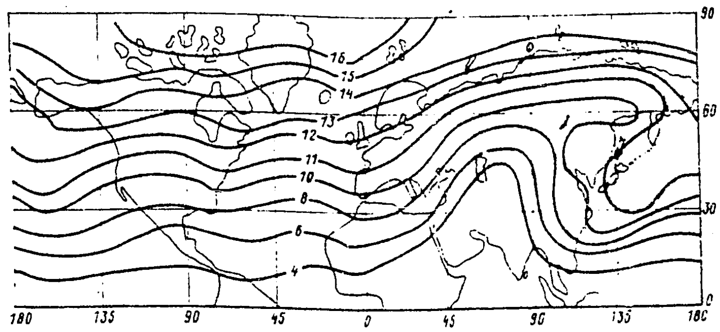

Fig. 4. First natural component S_1 (summer season), 10^{-3} cm.

Now we will discuss the structure of the statistical modes $S_i = \lambda_i P_i$. The nonnormalized EOF S_i are more graphic than the normalized functions P_i, since they give some idea how a priori uncertainty of the field described by the covariation matrix R is covered. The fact is that

$$R = \sum_{i=1}^{r} S_i \cdot S_i^T.$$

The S_1 mode describes the latitudinal structure of variations of the ozone field and some characteristics of the above-mentioned zonal nonuniformity. Our computations indicated that the meridional gradient of variations has a structure similar to the meridional gradient of the mean field. The mode S_2 represents a harmonic with two maxima in the regions of the Canadian and Siberian anomalies, having different signs (see Fig. 2, where the zero isoline is represented by circles). The maxima of the third mode S_3 are situated in the northern synoptically active zones of the Atlantic and Pacific Oceans, where there is the most significant heat transport from the ocean to the atmosphere in winter. In the low latitudes the zonal nonuniformities in ozone variations are manifested only in the fifth mode. The corresponding amplitudes are only about $3 \cdot 10^{-3}$ cm. Thus, the principal fluctuations of the ozone field in winter are in the high and middle latitudes, where there are regions of stable variations of the pressure field.

Now we will examine the summer-autumn period. The distribution of natural variations (Fig. 3) in this case differs from the corresponding winter distribution (compare with Fig. 1) in an appreciable increase in zonal nonuniformity, the appearance of an extensive maximum over Europe and the Northeastern Atlantic. We note that the latitudinal dependence of the distribution of natural variations, the same as the latitudinal dependence of the mean ozone field, decreases appreciably during the summer-autumn period. The maximum values of the variations in summer are only less than half the corresponding winter values. The mentioned characteristics are attributable to a considerable degree to the fact that in summer meridional circulation (and this means ozone transport as well) in the northern hemisphere attenuates considerably at all latitudes. The first mode S_1 (Fig. 4) has a well-expressed zonal nonuniformity over Eurasia in Northwestern Europe, Eastern Siberia, and also in the region of Southern and Southeast Asia. It can be assumed that the existence of the latter two anomalies is governed to a considerable degree by the influence of the Asiatic monsoon, which is characterized by a deep intrusion of air masses from the ocean onto the continent and also by an easterly jet stream in the western sector of the Pacific Ocean.

The second mode S_2 is a harmonic with two maxima of different signs in Eastern Siberia and on the Atlantic coast of North America. We note the stability of the Siberian anomaly throughout the year.

The mode S_3 demonstrates the diversity of characteristics of atmospheric dynamics in summer at different latitudes. The maxima of this mode are situated in the northern part of Canada, in the Northeast Atlantic, in the central and southern parts of Asia, and also in North Africa. The existence of the last of the mentioned zones is probably associated with the baroclinic mechanism of formation of disturbances at the center of the African continent [5]. The Asiatic anomaly, due to its elongation in the direction of the Indian Ocean, probably is of a monsoonal origin.

In summarizing, we can formulate some conclusions.

1. A small number of elements of the EOF base can quite effectively represent the broad spectrum of fluctuations of the field of total ozone content observed at stations in the ozonometric network. Sets of the first 15-20 EOF make it possible to describe 97-99% of the generalized dispersion of ozone variations.

2. The natural variations of the field of total ozone content are characterized by a latitudinal dependence during winter, and also a well-expressed zonal nonuniformity in summer.

3. The study of the first natural components makes it possible to define regions of statistically independent ozone fluctuations and trace their relationship to the characteristics of atmospheric dynamics.

4. The most stable region of anomalous ozone variations is the central region of Eastern Siberia. Among the other regions of this type are the northern synoptically active zones of the Atlantic and Pacific Oceans, the region of influence of the Asiatic monsoon and the zone of stationary waves in the northern part of Canada.

BIBLIOGRAPHY

1. Bagrov, A. N., "Natural Components of Small Samples With a Large Number of Parameters," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 12, 1978.

2. Obukhov, A. M., "Statistically Orthogonal Expanded Empirical Functions," IZV. AN SSSR, SERIYA GEOFIZICH. (News of the USSR Academy of Sciences, Geophysical Series), No 3, 1960.

3. Pokrovskiy, O. M., Ivanykin, Ye. Ye., Kaygorodtsev, A. Ye., "Numerical Modeling of Systems for Observation and Analysis of the Ozone Field," METEOROLOGIYA I GIDROLOGIYA, No 1, 1979.

4. London, J., Bojkov, R. D., Oltmans, S., Kelley, J. I., ATLAS OF GLOBAL DISTRIBUTION OF TOTAL OZONE July 1957-June 1967, NCAR TECHN. NOTE. NCAR/TN/113+STR. Boulder, Colorado, 1976.

5. Reed, R. J., Recker, E. E., "The Energetics of African Wave Disturbances," PROC. OF THE INTERN. SCI. CONF. ON THE ENERGETICS OF TROPICAL ATMOSPHERE, WMO, Geneva, Oct. 1978.

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PARAMETERIZATION OF HEAT AND MOISTURE EXCHANGE IN STORMS APPLICABLE TO PROBLEMS IN INTERACTION BETWEEN THE ATMOSPHERE AND OCEAN

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[Article by A. L. Kazakov and Candidate of Physical and Mathematical Sciences V. N. Lykosov, Computation Center Siberian Department USSR Academy of Sciences, submitted for publication 28 Jan 80]

[Text]

Abstract: On the basis of the Borisenkov-Kuznetsov model [3] the authors propose a simple and economical method for computing the fluxes of apparent and latent heat caused by the presence of spray clouds in the surface (near-water) layer of the atmosphere during storms. The derived parametric formulas can be used in numerical models of interaction between the atmosphere and ocean and also in the processing of experimental data.

Introduction. According to [2, 12], during storms, in addition to turbulent transfer of momentum, heat and moisture, an important role is played by heat and mass exchange caused by the presence of spray clouds in the near-water layer of the atmosphere. The principal mechanism determining the entry of droplets into the near-water layer is assumed to be the collapse of foam air bubbles. Different aspects of experimental investigations of this process are discussed in detail in [6, 9, 11-13].

In the development of a physical-mathematical model of interaction between a spray cloud and the surrounding medium, and also in the parameterization of these processes in problems of atmospheric dynamics, it is necessary to solve problems related to both the formation and development of a spray cloud (so-called "dynamic" problem) and determination of the heat regime of an individual droplet (so-called thermodynamic problem) [3]. Their joint solution meets with great difficulties and therefore in the first stage of the investigations it was desirable to solve these problems separately in order to analyze the solution of each of them in detail with stipulated "external" parameters. For example, in [4, 5] a solution of the heat balance equation for a droplet was obtained and analyzed on the assumption that: a) interaction of droplets with one another is not considered; b) during the time of "flight" of a droplet from the moment of its departure from the water to its falling back into the water ($\lesssim 1$ sec) the changes in its mass, size, salinity and heat capacity can be neglected; c) processes of radiation exchange of a droplet with the surrounding atmosphere are not taken into account; d) wind velocity, temperature and air humidity are assumed to be constant with height and equal to their values at the level 10 m; e) in order to stipulate the specific humidity at the droplet surface use is made of a linear approximation of the Magnus formula, correct only with small deviations of droplet temperature from air temperature.

Despite the fact that the evaluations of heat and moisture exchange made in [12] for stormy weather conditions by the use of this model had a preliminary character, the results of their comparison with real data were encouraging. With respect to the dynamics of a spray cloud, in [1, 3, 4] it is proposed that it be parameterized by stipulating the flux of spray droplets and the time of their flight, determined from solution of the equations of motion of a spherical droplet. In [3], being a development of [4, 5], the flux of droplets is expressed through the empirical droplet-size distribution function and the dependence of liquid water content on wind velocity. The use of the asymptotic Laplace method by these same authors for simplification of expressions including both dynamic and thermodynamic parts enabled them to derive parametric formulas for computing the fluxes of heat and moisture under stormy conditions. However, for the realization of the algorithm cited in [3] it is necessary to have considerable computer time (about 5 sec on a BESM-6 computer for computing one meteorological situation). In connection with this circumstance the purpose of this communication is the derivation, on the basis of the results presented in [3], of simple and economic parametric formulas for computing the fluxes of heat and moisture under stormy conditions for their use in numerical models of interaction between the atmosphere and ocean.

Initial Model

By definition [3], the vertical transport of apparent (H_{drop}) and latent (LE_{drop}) heat accomplished by droplets from 1 cm^2 of water surface is described by the following expressions. [Here and in the text which follows use is made of the cgs system of units and temperature is indicated in °C. The sought-for fluxes have the dimensionality cal/(cm^2·sec).]

$$H_\kappa = \int_0^\infty f w_0 H_{\kappa s}\, dD, \tag{1}$$

$$LE_\kappa = L \int_0^\infty f w_0 E_{\kappa s}\, dD, \tag{2}$$

[K = drop] where f is the droplet-size distribution function,

$$f = \frac{6\, k \mu^{\frac{n+4}{k}}\, \delta^*}{\Gamma\left(\frac{n+4}{k}\right) \pi \rho_w D_m^4} \left(\frac{D}{D_m}\right)^n \exp\left[-\mu \left(\frac{D}{D_m}\right)^k\right], \tag{3}$$

$$\mu = \frac{n}{k},$$

k, n are dimensionless constants, selected on the basis of observational data, D is droplet diameter, D_m is the so-called modal diameter of the droplet for which the f function attains a maximum value, $\Gamma(x)$ is the gamma function, ρ_w is water density, δ^* is the liquid water content of the spray cloud,

$$\delta^* = 0{,}22 \exp\left(-\frac{218\ u_б^2}{172 + u_б^3}\right),\quad u_б = \frac{u}{\tilde{u}},\quad \tilde{u} = 100\ см\ с, \qquad [\text{cm/sec}] \tag{4}$$

[$б$ = spray] u is wind velocity, w_0 is the initial vertical component of droplet velocity,

$$w_0 = \frac{D_m}{D}\, w_{0\ m}, \tag{5}$$

$w_{0\ m}$ is its value with $D = D_m$, $H_{drop\ s}$, $E_{drop\ s}$ are the total quantities of heat and moisture respectively released by droplets of the diameter D during the time (τ) of their presence in the air,

$$H_{к\ s} = 2\,\pi D \lambda k_H \left\{\tau (t_a - t_p) + \frac{t_w - t_p}{\alpha_t}\left[\exp(-\alpha_t \tau) - 1\right]\right\}, \tag{6}$$

$$E_{к\ s} = 2\,\pi D \rho\, d k_E \left\{(\alpha_1 t_p + \beta_1)\,\tau + \frac{\alpha_1 (t_w - t_p)}{\alpha_t}\left[1 - \exp(-\alpha_t \tau)\right]\right\}, \tag{7}$$

$$\tau = A_0 (D, w_{0\ m}) \exp\left[-\alpha_0 (D, w_{0\ m})\, u\right], \tag{8}$$

[p = equil]

$$A_0 (D, w_{0\ m}) \quad \text{and} \quad \alpha_0 (D, w_{0\ m})$$

are approximation coefficients,

$$\alpha_t = 12\, \frac{\alpha_1 d \rho k_E L + k_H \lambda}{c_w \rho_w D^2}, \tag{9}$$

ρ is air density, c_w is the heat capacity of water,

$$\alpha_1 = \frac{4250\ q_s}{[241{,}9 + 0{,}5\,(t_a + t_w)]^2}, \tag{10}$$

$$\beta_1 = Д - \alpha_1 t_a, \tag{11}$$

t_a is air temperature, t_w is the temperature of the water surface, q_s is saturating specific humidity,

$$q_s = \frac{3{,}8}{p} \exp\left(\frac{17{,}57\ t_a}{241{,}9 + t_a}\right), \tag{12}$$

$Д$ is the deficit of specific humidity in the near-water layer,

$$Д = q_s - q_a, \tag{13}$$

q_a is air specific humidity, p is pressure, t_{equil} is the so-called equilibrium temperature of the droplet,

$$t_{equil} = \frac{k_H \lambda t_a - \beta_1 d \rho k_E L}{k_H \lambda + \alpha_1 d \rho k_E L}, \tag{14}$$

k_H, k_E are the ventilation coefficients, taking into account the influence of wind velocity on the intensity of heat and moisture exchange between a droplet and the surrounding medium [10],

$$k_H = 1 + 0{,}23\ \mathrm{Pr}_M^{1/3}\,\mathrm{Re}^{1/2}, \quad k_E = 1 + 0{,}23\ \mathrm{Sc}_M^{1/3}\,\mathrm{Re}^{1/2}, \tag{15}$$

Re, Pr_M, Sc_M are the Reynolds, Prandtl and Schmidt numbers respectively,

$$\mathrm{Re} = \frac{Du\rho}{\eta}, \quad \mathrm{Pr}_M = \frac{\eta}{\rho\chi}, \quad \mathrm{Sc}_M = \frac{\eta}{\rho d}, \tag{16}$$

λ, χ, η are coefficients characterizing the molecular properties of the air: its thermal conductivity, thermal diffusivity and dynamic viscosity respectively, d is the coefficient of molecular diffusion of water vapor, L is the latent heat of evaporation. In accordance with [10],

$$\lambda \cdot 10^5 = 5.77 + 0{,}16\, t_a, \quad \eta \cdot 10^6 = 172 + 0{,}5\, t_a,$$
$$d = 0{,}22 + 0{,}0015\, t_a, \quad L = 597{,}26 - 0{,}647\, t_a. \tag{17}$$

However, it seems entirely natural that in the first approximation these values can be considered constant and equal to their values with $t_a = 10°C$. With respect to the values c_w, ρ_w, they also, without substantial error, can evidently be assumed to be as follows:

$$c_w = 1\ \mathrm{cal/(g \cdot °C)}, \quad \rho_w = 1\ \mathrm{g/cm^3}.$$

For the practical use of (3) and (5), the following values of the parameters are recommended in [3]:

$$D_m = 0.015\ \mathrm{cm},\ k = 1,\ n = 2,\ w_{0\,m} = 300\ \mathrm{cm/sec}. \tag{18}$$

In this connection it must be noted that qualitative and statistically supported experimental data on change in the function f and the parameters D_m, k, n associated with it, and also on the liquid water content during strong winds are lacking. What has been said also applies to $w_{0\,m}$. Accordingly, adopting nevertheless for further practical use the values for the parameters (18), it must be remembered that they can be replaced by others corresponding to more reliable measurements or numerical experiments. For this same reason we considered some values (as in [3]) with allowance for a possible range of changes $w_{0\,m}$ = 50–300 cm/sec.

Derivation of Economical Parametric Formulas

It follows from the form of formulas (3), (5)-(7) that after their substitution into (1), (2) the greatest difficulties for analytical integration are the expressions (6) and (7). Therefore, we will discuss in detail their transformation so as to facilitate the determination of (1), (2). An analysis of (6), (7) shows that the parameters α_i, t_{equil} and τ are functions of droplet diameter. Taking into account the relative smallness of the difference in the ventilation coefficients, for the heat and moisture exchange of a droplet with the surrounding air with the usually employed values $Pr_M = 0.71$ and $Sc_M = 0.62$, we introduce some mean ventilation coefficient

$$k = 1 + 0{,}2\sqrt{\mathrm{Re}} \tag{19}$$

and we will transform expressions (9) and (14) to the form

[эф = eff]

$$\alpha_t = 12\,\frac{c_{эф}\,\rho\,dk}{c_w\,\rho_w\,D^2}, \tag{20}$$

[p = equil]

$$t_p = t_a - \frac{L}{c_{эф}}\,Д, \tag{21}$$

where

$$c_{эф} = \frac{\lambda}{\rho\,d} + \alpha_1\,L. \tag{22}$$

In order to explain the nature of the dependence of droplet flight time on its diameter we used expression (8) and a table of the coefficients A_0 and α_0 cited in [3]. An analysis of the resulting graphic material indicated that there is a possibility for approximating this dependence by a formula in the form

$$\tau = \hat{\tau}\left(\frac{D}{D_m}\right)^{\gamma}, \tag{23}$$

where

$$\hat{\tau} = \tau_0 - \tau'_u\,(u - u_0). \tag{24}$$

In finding the proportionality factor $\hat{\tau}$ and the exponent γ we used as a point of departure the fact that the relative error in computing τ from (23) in comparison with its values computed from (8) must not exceed 10% and insofar as possible must be minimum near $D = D_m$. In addition, since the influence of spray in the general balance of heat and moisture exchange becomes appreciable at wind velocities $\gtrsim 15$ m/sec, we limited the possible range of u changes downward to the value $u_0 = 15$ m/sec. The values of the parameters τ_0, τ'_u and γ computed in accordance with the conditions stipulated above are given in Table 1. We note, however, that since expression (23) is an approximation of formula (8), which in turn is an approximate result of numerical experiments, the resulting values of the parameters are also approximate and in the future can be made more precise.

Table 1

Dependence of the γ, τ_0 and τ'_u Parameters on the Modal Value of the Vertical Velocity of Departure of Droplets from Water Surface

	w_{0m}, cm/sec			
	50	100	200	300
γ	0,89	0,74	0,68	0,68
$\tau_0 \cdot 10^2$	7,1397	9,3221	14,044	16,8518
$\tau'_u \cdot 10^3$	0,9831	1,0418	1,5239	1,829

We introduce the notation

$$y = \frac{1 - \exp(-\alpha_t\,\tau)}{\alpha_t\,\tau} \tag{25}$$

and we will transform expressions (6), (7) with the introduced simplifications (19)-(24) taken into account. Then we have

$$H_{\kappa s} = 2\pi D \lambda k \hat{\tau} \left(\frac{D}{D_m}\right)^{\gamma} \left[\text{Д} \frac{L}{c_{\text{эф}}} - \left(t_w - t_a + \text{Д} \frac{L}{c_{\text{эф}}}\right) y\right], \qquad (26)$$

$$E_{\kappa s} = 2\pi D \rho d k \hat{\tau} \left(\frac{D}{D_m}\right)^{\gamma} \left[\text{Д} \left(1 - \alpha_1 \frac{L}{c_{\text{эф}}}\right) + \alpha_1 y \left(t_w - t_a + \text{Д} \frac{L}{c_{\text{эф}}}\right)\right]. \qquad (27)$$

[K = drop, ЭФ = eff]

Table 2

Dependence of Integrals I_1, I_2 on Modal Value of Vertical Velocity of Departure of Droplets from Water Surface

	w_{0m} см/с cm/sec			
	50	100	200	300
I_1	0,35309	0,32714	0,31791	0,31791
I_2	0,47691	0,4327	0,41687	0,41687

Now substituting (3), (5), (26) and (27) into (1), (2), as a result of integration we obtain the following parametric formulas:

$$H_{\kappa} = C\lambda \left[\text{Д} \frac{L}{c_{\text{эф}}} (\varphi_1 - \varphi_2) + (t_a - t_w)\varphi_2\right], \qquad (28)$$

$$LE_{\kappa} = LCd\rho \left\{\text{Д} \left[\varphi_1 - \alpha_1 \frac{L}{c_{\text{эф}}} (\varphi_1 - \varphi_2)\right] + \alpha_1 (t_w - t_a)\varphi_2\right\}, \qquad (29)$$

where

$$C = 12 \frac{k\mu^{\frac{n+1}{k}} w_{0m} \delta^* \hat{\tau}}{\Gamma\left(\frac{n+1}{k}\right) \rho_w D_m^2}, \qquad (30)$$

$$\varphi_1 = I_1 + 0{,}2\, I_2 \sqrt{\mathrm{Re}_m}, \quad \varphi_2 = I_3 + 0{,}2\, I_4 \sqrt{\mathrm{Re}_m}, \qquad (31)$$

$$I_1 = \int_0^\infty x^{\gamma+n} \exp(-\mu x^k)\, dx, \quad I_2 = \int_0^\infty x^{0.5+\gamma+n} \exp(-\mu x^k)\, dx, \qquad (32)$$

$$I_3 = \int_0^\infty x^{\gamma+n}\, y \exp(-\mu x^k)\, dx, \quad I_4 = \int_0^\infty x^{0.5+\gamma+n}\, y \exp(-\mu x^k)\, dx. \qquad (33)$$

$$\mathrm{Re}_m = \frac{D_m u \rho}{\eta}, \qquad (34)$$

$$x = \frac{D}{D_m}. \qquad (35)$$

The values of the I_1, I_2 integrals were found analytically in [8] and are given in Table 2. With respect to computation of the integrals I_3, I_4, for their determination we used the Lobatto numerical method with automatic choice of the integration interval [7]. In this connection, for evaluating the error arising due to the

Table 3

Dependence of Integrals I_3, I_4 on Air Temperature t_a and Modal Value of Vertical Velocity of Droplet Departure from Water Surface

$w_{0\,m}$ см/с cm/sec		t_a °C							
		−5	0	5	10	15	20	25	30
50	I_3	0,24437	0,22566	0,20715	0,18701	0,16714	0,14745	0,12842	0,11053
	I_4	0,34088	0,3166	0,29161	0,26555	0,23883	0,212	0,18572	0,16071
100	I_3	0,20379	0,1855	0,16674	0,14846	0,13059	0,11348	0,0975	0,0829
	I_4	0,28307	0,25906	0,23509	0,21088	0,18685	0,16348	0,14133	0,12089
200	I_3	0,16426	0,14515	0,12602	0,11044	0,09477	0,08051	0,06781	0,05675
	I_4	0,2304	0,20536	0,18155	0,15866	0,13706	0,1171	0,0991	0,08325
300	I_3	0,1485	0,12949	0,11214	0,09613	0,0816	0,06867	0,05739	0,04775
	I_4	0,20978	0,18447	0,16097	0,13893	0,11863	0,10032	0,08417	0,07023

introduction of the finite droplet spectrum we first carried out computations using formulas (32) for comparison of the numerical and analytical integration methods. This comparison indicated that when using the selected droplet spectrum ($D = 10^3 - 2 \cdot 10^{-1}$ cm) the relative error in numerical integration was $\lesssim 10^{-6}$%. Table 3 gives the values of the integrals I_3, I_4 obtained by averaging of their extremal values with possible changes in pressure p = 970-1030 mb and wind velocity u = 15-40 m/sec for each stipulated air temperature value t_a.

Now we will note another important property of the derived parametric formulas (28) and (29). This is the relative influence of the temperature and humidity fields on the heat and moisture exchange of a spray cloud. Thus, the apparent heat flux is dependent not only on the diffusion of heat, but also on the moisture deficit in the near-water layer. And evaporation, in turn, is determined, in addition to the deficit of saturation in the surrounding air, by the conditions of thermal stratification in the near-water layer.

The algorithm for computing the fluxes of apparent and latent heat due to spray clouds during storms cited above was applied in the form of subprograms in FORTRAN for the BESM-6 electronic computer. The time for computing one variant with the input data t_a, t_w, q_a, p and u does not exceed 0.01 sec. At the present time this method is being tested in a model of general circulation of the atmosphere and ocean formulated by the Computation Center Siberian Department USSR Academy of Sciences.

In conclusion the authors consider it their pleasant duty to express their appreciation to Ye. P. Borisenkov, R. S. Bortkovskiy, M. A. Kuznetsov and A. G. Yantsen for useful discussion of the article.

BIBLIOGRAPHY

1. Ariel', N. Z., Bortkovskiy, R. S., "Refined Model of Energy and Mass Exchange of Spray Over the Ocean Surface During a Storm," TAYFUN-75 (Typhoon-75), Vol 2, Leningrad, Gidrometeoizdat, 1978.

2. Borisenko, Ye. P., "Some Mechanisms of Interaction Between the Atmosphere and Ocean During Stormy Weather Conditions," PROBLEMY ARKTIKI I ANTARKTIKI (Problems of the Arctic and Antarctic), No 43-44, 1974.

3. Borisenkov, Ye. P., Kuznetsov, M. A., "Parameterization of Interaction Between the Atmosphere and Ocean Under Stormy Weather Conditions Applicable to Models of General Circulation of the Atmosphere," IZV. AN SSSR, FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), Vol 14, No 5 [year not given].

4. Bortkovskiy, R. S., "Mechanism of Interaction Between the Ocean and the Atmosphere During a Storm," TRUDY GGO (Transactions of the Main Geophysical Observatory), No 282, 1972.

5. Bortkovskiy, R. S., "Refinement of Estimates of Heat and Moisture Exchange Between the Ocean and the Atmosphere," TRUDY GGO, No 326, 1975.

6. Bortkovskiy, R. S., Kuznetsov, M. A., "Some Results of Investigation of the State of the Sea Surface," TAYFUN-75, Vol 1, Leningrad, Gidrometeoizdat, 1977.

7. Brushlinskaya, O. V., Vasil'yeva, L. G., "Set of Standard Programs for Approximate Computation of Simple Integrals With Automatic Choice of Interval," CHISLENNYY ANALIZ NA FORTRANE (Numerical Analysis in FORTRAN), Moscow, MGU, No 8, 1974.

8. Gradshteyn, I. S., Ryzhik, N. M., TABLITSY INTEGRALOV, SUMM, RYADOV I PROIZVEDENIY (Tables of Integrals, Sums, Series and Products), Moscow, Nauka, 1971.

9. Kraus, Ye. B., VZAIMODEYSTVIYE ATMOSFERY I OKEANA (Interaction Between the Atmosphere and the Ocean), Leningrad, Gidrometeoizdat, 1976.

10. Matveyev, L. T., KURS OBSHCHEY METEOROLOGII (Course in General Meteorology), Leningrad, Gidrometeoizdat, 1976.

11. Preobrazhenskiy, L. Yu., "Evaluation of the Content of Spray Droplets in the Near-Water Layer of the Atmosphere," TRUDY GGO, No 282, 1972.

12. Bortkovskiy, R. S., Byutner, E. K., Malevskiy-Malevich, S. P., Preobrazhenskiy, L. Yu., PROTSESSY PERENOSA VBLIZI POVERKHNOSTI RAZDELA OKEAN-ATMOSFERA (Transport Processes Near the Ocean-Atmosphere Discontinuity), Leningrad, Gidrometeoizdat, 1974.

13. Wu, J., "Spray in the Atmospheric Surface Layer: a Review and Analysis of Laboratory and Oceanic Results," JGR, Vol 84, No C4, 1979.

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VARIABILITY OF ACTIVE LAYER CHARACTERISTICS IN THE NORTHWESTERN PACIFIC OCEAN DURING PASSAGE OF A STORM

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 65-68

[Article by K. A. Rogachev, Pacific Ocean Oceanological Institute, submitted for publication 20 Aug 79]

[Text]

Abstract: The interaction between the boundary layers of the atmosphere and the ocean is modeled at a synoptic time scale. The wind mixing of different thermal structures in the northwestern part of the Pacific Ocean is examined in a differential model of the active layer. A strong dependence of the change in active layer characteristics during a storm on the vertical thermal structure is demonstrated.

Among the most important disturbances occurring in the active layer (AL) of the ocean are synoptic and seasonal disturbances. Under the term "synoptic variability" we will understand processes with a characteristic time scale of several days, which corresponds to the synoptic maximum of atmospheric variability. Such a variability is associated with the passage of individual pressure systems, which account for 80% of the kinetic energy in the atmosphere in the middle latitudes [2]. Elsberry and Camp [5], who examined the characteristics of strong atmospheric disturbances, pointed out that their main part is concentrated in short periods of time with a duration of about two days associated with the passage of cyclones. Accordingly, in the study of thermal structure of the AL it is the reaction of the AL to the passage of powerful atmospheric formations which is of the greatest interest in study of the thermal structure of the AL.

We can discriminate three physical processes responsible for restructuring of the thermal structure of the AL during a storm: cooling of the ocean surface as a result of evaporation and heat exchange with the atmosphere, including nonturbulent mechanisms, wind mixing of the AL and vertical circulation of waters in the zone affected by the storm.

In this article we will examine changes in the structure of the AL as a result of evaporation and turbulent heat exchange and wind mixing. A one-dimensional differential model of the AL is used for this purpose. The role of one-dimensional processes in the reaction of the AL to strong atmospheric disturbances was examined

[redacted]

by Camp and Elsberry. In [4] they drew the conclusion that they play a decisive role at a synoptic time scale.

Using a one-dimensional model of a nonstationary Ekman friction layer for the AL, we will have the following system of equations:

$$\begin{cases} \frac{\partial u}{\partial t} - lv = \frac{\partial}{\partial z} K \frac{\partial u}{\partial z} \\ \frac{\partial v}{\partial t} + lu = \frac{\partial}{\partial z} K \frac{\partial v}{\partial z} \\ \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} K \frac{\partial T}{\partial z} \\ K = K_0 (u, v, T, z) \end{cases} \tag{1}$$

with the boundary conditions

$$\begin{gathered} K \frac{\partial u}{\partial z} = -\tau_x/\rho_0, \quad K \frac{\partial v}{\partial z} = -\tau_y/\rho_0, \quad K \frac{\partial T}{\partial z} = -Q, \quad z = 0, \\ u = v = 0, \quad \frac{\partial T}{\partial z} = 0, \quad z = H_0, \end{gathered} \tag{2}$$

where u, v are components of velocity of the drift current, l is the Coriolis parameter, t is time, z is the vertical coordinate. The origin of coordinates is placed at the ocean surface with an axis directed vertically downward, T(z, t) is water temperature, K is the coefficient of turbulent diffusion, assumed equal to the coefficient of turbulent viscosity along the vertical, τ_x, τ_y are the components of the shearing stress of the wind at the ocean surface, ρ_0 is water density, Q is the heat flux into the AL.

The function K_0 is determined from the balance equation for turbulent energy, which includes the following components: generation of turbulence by velocity shear, dissipation of turbulent energy and work against buoyancy forces. The solution of this equation has the form [1]

$$K_0 = (ch)^2 \sqrt{\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2 + g\beta \frac{\partial T}{\partial z}},$$

where h is the depth of the well-mixed layer, g, β are the acceleration of free falling and the coefficient of thermal expansion of water, c is some constant.

By the term "depth of the well-mixed layer" is meant the depth at which the expression under the radical sign becomes equal to zero. Deeper than the well-mixed layer the value of the K coefficient was assumed to be constant and equal to the minimum value 1 cm^2/sec. This value is two orders of magnitude less than the value of the coefficient in the well-mixed layer and two orders of magnitude greater than the molecular viscosity coefficient in water. Such a determination of the diffusion coefficient deeper than the mixed layer is associated with the paucity of ideas concerning mixing processes in the thermocline.

If one uses the scale [t] = 10^5 sec (day) and a scale of changes in depth of the mixed layer [H] = 20 m, then from (1) [K] = $[H]^2/[t]$ = 40 cm^2/sec. If [Q] = 800 W/m^2, then [T] = 1°. If the velocity scale [u] = 10 cm/sec, then [τ] = 0.2 dyne/cm^2.

The thermal balance of the ocean surface Q is dependent on many factors: wind velocity, water-air temperature difference, air humidity, cloud cover, intensity of short-wave radiation. In this study the heat flux was represented in the form

$$Q=\sum_{i=1}^{4} H_i.$$

Here the following components are considered:
$H_1 = \mu R$ is short-wave radiation, where R is the flux of short-wave radiation incident on the upper boundary of the atmosphere, μ is the attenuation coefficient, dependent on surface albedo, air humidity and atmospheric albedo.

$H_2 = c_1(T + 273°)^4$ is the long-wave radiation of the surface, where c_1 is some constant, $H_3 = -c_2(T - T_A)c_D$ w is turbulent heat exchange between the ocean surface and the atmosphere, where c_D is the friction coefficient, w is wind velocity, T_A is air temperature, c_2 is a constant,

$H_4 = BoH_3$ is heat exchange as a result of evaporation, where Bo is the Bowen ratio.

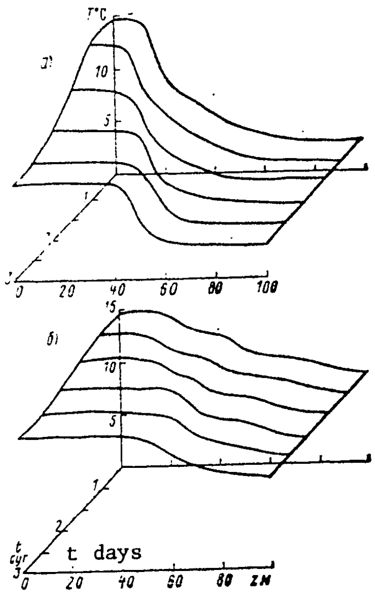

Fig. 1. Reaction of different thermal structures to passage of a storm. a) 41°N, 170° E; b) 46°N, 170°E.

System (1) with the boundary conditions (2) was solved numerically with stipulated T_A, $\tau_{x,y}$ values with different initial conditions for u(z,0), v(z,0), T(z,0), K(z, 0). The thermal conductivity equation was solved using an implicit triangle scheme by the factorization method and the equation of motion was solved by the transverse fractional steps method [3] with a time interval of $5 \cdot 10^3$ sec and with an interval along the vertical coordinate 2 m with H_0 = 100 m.

In order to investigate the reaction of the active layer to wind intensifications and changes in air temperature we selected the mean monthly long-term temperature profiles in the northwestern part of the Pacific Ocean. Data on water temperature were taken from the archives of the Pacific Ocean Oceanological Institute of the Far Eastern Scientific Center USSR Academy of Sciences, data on air temperature -- from the ATLAS OF THE OCEANS (Pacific Ocean), and the μ and c_i parameters -- from [6].

Thus, different vertical thermal structures were disturbed by an identical wind. We will examine the reaction of the AL to identical disturbances by the wind in the northwestern part of the Pacific Ocean. Figure 1 shows the change in the temperature profiles with time for August with an intensification of the wind in conformity to a linear law from 3 to 20 m/sec and attenuation to the initial value:

$$\tau_x = \begin{cases} 0{,}2 & t < 1 \\ 0{,}2 + 9{,}2\,(t-1) & 1 \leq t \leq 2 \\ 9{,}4 - 9{,}2\,(t-2) & 2 \leq t \leq 3 \\ 0{,}2 & t > 3, \end{cases}$$

where time is in days. The air temperature decreased by 5°C relative to the mean monthly value during wind intensification. The figure shows that despite identical disturbance by the wind, different thermal structures in the AL react differently. Thus, the change in surface temperature at the point 46°N, 170°E attains 6°C, and at the point 41°N, 170°E -- 2°C.

The value of the turbulence coefficient is different for different structures, but its variability is far more dependent on the wind stress. The values of the coefficient increase by a factor of 20 during the wind stress peak and attain 600 cm^2/sec. A comparison of the temperature gradient values in the AL with the changes in surface temperature indicates their full correspondence. Such a correspondence indicates that for an adequate description of changes in surface temperature during the passage of a storm it is necessary to use a differential model which makes it possible to describe the continuous thermal structure of the AL.

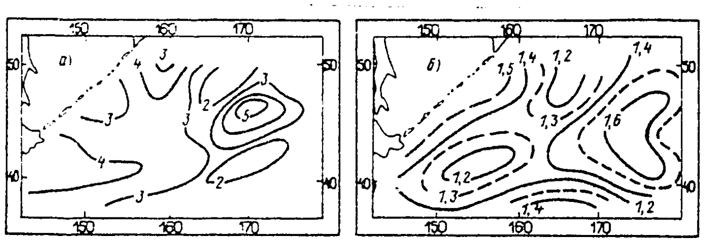

Fig. 2.

Figure 2a shows the decrease in surface temperature (°C) in the northwestern part of the Pacific Ocean, computed for 30 profiles for August and Fig. 2b shows the change in depth (x 20 m) of the well-mixed layer with the passage of a two-day storm. The considered region is characterized by a complex hydrological structure. This is associated primarily with the existence here of the subarctic front between subarctic and subtropical waters. The maximum change in temperature attains 6°C and the maximum increase in depth of the well-mixed layer is 30 m.

Computations of the variability of characteristics of the AL during passage of a model storm, carried out for the entire northern part of the Pacific Ocean, indicate that the maximum variability is observed in the northwestern part of the Pacific Ocean.

Thus, after examining the reaction of different thermal structures to the passage of a model storm it is possible to draw the following conclusion: the change in surface temperature is determined by the thermal structure of the AL. The maximum changes must be expected in an AL with large vertical temperature gradients. This indicates that capture of cold water from the deep layers is the dominant process determining the change in surface temperature.

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BIBLIOGRAPHY

1. Kalatskiy, V. I., MODELIROVANIYE VERTIKAL'NOY TERMICHESKOY STRUKTURY DEYATEL'-NOGO SLOYA OKEANA (Modeling of the Vertical Thermal Structure of the Active Layer in the Ocean), Leningrad, Gidrometeoizdat, 1978.

2. Lappo, S. S., SREDNEMASSHTABNYYE DINAMICHESKIYE PROTSESSY OKEANA, VOZBUZHDA-YEMYYE ATMOSFEROY (Mesoscale Dynamic Processes in the Ocean Excited by the Atmosphere), Moscow, Nauka, 1979.

3. Yanenko, N. N., METOD DROBNYKH SHAGOV RESHENIYA MONOGOMERNYKH ZADACH MATEMAT-ICHESKOY FIZIKI (Fractional Steps Method for Solution of Multidimensional Problems in Mathematical Physics), Moscow, Nauka, 1967.

4. Camp, N. T., Elsberry, R. L., "Oceanic Thermal Response to Strong Atmospheric Forcing: The Role of One-Dimensional Process," J. PHYS. OCEANOGR., Vol 8, No 3, 1978.

5. Elsberry, R. L., Camp, N. T., "Oceanic Thermal Response to Strong Atmospheric Forcing: Characteristics of Forcing Events," J. PHYS. OCEANOGR., Vol 8, No 3, 1978.

6. Saltzman, B., Ashe, S., "The Variance of Surface Temperature Due to Diurnal and Cyclone-Scale Forcing," TELLUS, Vol 28, No 4, 1976.

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MEANDERING AND EDDY FORMATION IN ZONAL OCEAN CURRENTS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 69-77

[Article by Candidate of Physical and Mathematical Sciences V. F. Kozlov and Ye. V. Ivanchenko, Far Eastern State University and Pacific Ocean Oceanological Institute, submitted for publication 28 Sep 79]

[Text]

Abstract: A study was made of the reaction of a stationary poorly stratified flow with a transverse velocity shear in an infinite channel whose width in some section changes in a jump. It was possible to discriminate a class of profiles of undisturbed flow leading to a linear equation of conservation of potential vorticity. The meandering and eddy formation conditions were investigated. The results of numerical computations of the streamlines are given for different symmetric profiles. The conclusion is drawn that the curvature of the velocity profile of the oncoming flow plays a decisive role.

The meandering and eddy formation phenomena in regions of strong currents of the Gulf Stream or Kuroshio type have become the subject of theoretical investigations only in recent decades [4, 8, 10, 12, 14, 18, 20, 22, 24-26]. As a model it is customary to examine limiting cases either of thin inertial jets [2, 6, 16, 19, 21], or, on the other hand, kinematically homogeneous (without a transverse velocity shear) flows [2, 3, 7, 13, 15, 17]. A qualitative explanation of the observed effects can be given using a fundamental relationship in geophysical hydrodynamics -- the law of conservation of potential vorticity, which in the simplest stationary barotropic case has the form

$$\frac{\zeta + \Omega}{H} = \Pi(\Psi).$$

Here ζ is relative vorticity, Ω is the Coriolis parameter, H is the total height of the column of fluid, Ψ is the integral stream function.

It can be seen from this expression that the relative vorticity closely associated with the curvature of the streamlines changes every time when along the trajectory there are changes in Ω (meridional displacements of the streamlines) and/or H (change in bottom relief). An extreme form of manifestation of bottom relief is the presence in the flow of lateral solid boundaries, including islands. It is of interest to study the influence of disturbance of the form of the lateral boundary on a flow with a transverse mean (vertically) velocity shear.

A problem of a similar type is examined in this article. Within the framework of an inertial quasigeostrophic model, on the assumption of weak stratification, a study is made of the behavior of a flow with a transverse velocity shear in a zonal channel whose width in some section changes in a jump. In the first approximation such a schematic representation can be used for the Kuroshio region to the south of Japan.

Now we will formulate the principal assumptions of the model. We will introduce the characteristic scales of a horizontal extent L^*, horizontal velocity U^*, depth H^*, Coriolis parameter Ω^*, rise of bottom relief h^*, buoyancy frequency N^*; assume that $\beta = d\Omega/dy$ is the constant Rossby parameter. The fluid is assumed to be incompressible and nonviscous and motion is assumed to be steady; in addition, it is assumed that

$$\varepsilon \equiv \frac{U^*}{L^* \Omega^*} \ll 1, \quad B \equiv \frac{H^* N^*}{\Omega^* L^*} \ll \varepsilon^{1/2}, \quad \sigma \equiv \frac{h^*}{\varepsilon H^*} = 0$$

which denote quasi-geostrophicity, weak stratification and weak changes in topography respectively. It is easy to show [11] that under the indicated conditions the anomaly of hydrodynamic pressure p and vertical velocity w in dimensionless variables satisfy the simplified system of vorticity equations

$$J(p, \Delta p + by) - w_z = 0 \tag{1}$$

and density advection

$$J(p, p_z) = 0 \tag{2}$$

with the vertical boundary conditions

$$w = 0, \qquad z = 0, \tag{3}$$

$$w + \sigma J(p, h) = 0, \qquad z = 1. \tag{4}$$

Here Δ is the horizontal Laplacian, J is the Jacobi operator, h(x,y) is the dimensionless rise of bottom relief, $b = \beta L^{*2}/U^*$ is a planetary parameter.

The general integral of equation (2) has the form

$$p = G(z, \varphi(x, y)), \tag{5}$$

where G and φ are arbitrary functions of its arguments. Under the conditions of a stationary problem the G function must be stipulated and for determining φ use is made of equation (1) under the conditions (3) and (4). Substituting (5) into (1) and integrating the derived expression vertically with allowance for the boundary conditions, we arrive at an equation which allows a first integral in the form

$$\Delta\varphi \int_0^1 G_\varphi^2 \, dz + \frac{1}{2} (\nabla\varphi)^2 \frac{\partial}{\partial\varphi} \int_0^1 G_\varphi^2 \, dz + by \int_0^1 G_\varphi \, dz + \tag{6}$$

$$+ \sigma h G_\varphi(1, \varphi) = F(\varphi),$$

where the $F(\varphi)$ function, speaking in general, is not unambiguous.

At the boundary G of the considered region in the plane (x, y) the boundary condition can be written in the form $\varphi|_G = f(s)$, where f(s) is some function stipulated at G. The problem of the choice of $F(\varphi)$ and f(s) each time must be investigated separately.

In a barotropic case $G_z \equiv 0$; therefore, without diminishing universality, it is possible to assume that $G \equiv 1$; hence one derives the known equation for the barotropic problem

$$\Delta\varphi + by + \sigma h = F(\varphi). \tag{7}$$

With stipulated kinetic energy and circulation at the boundary of a limited region the form of the $F(\varphi)$ function can be determined from the principle of a minimum of entrophy [5, 23]. It appears that in this case the $F(\varphi)$ function must be linear,

$$F(\varphi) = C\varphi + D. \tag{8}$$

In a baroclinic case, if G allows representation in the form [11]

$$G = A(z)\varphi, \tag{9}$$

the left-hand side of equation (6) also becomes linear. By normalizing $A(z) \geqslant 0$ by the condition

$$\int_0^1 A^2(z)\, dz = 1$$

and assuming

$$S = \int_0^1 A(z)\, dz,$$

from (6) we obtain

$$\Delta\varphi + bSy + \sigma A(1) h = F(\varphi), \tag{10}$$

where, as it is easy to confirm, $0 \leqslant S \leqslant 1$ and $S = 1$ only in a barotropic case $A(z) \equiv 1$.

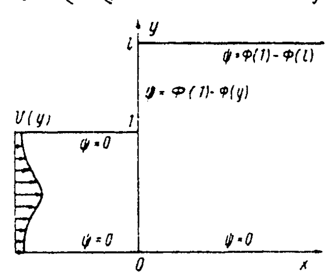

Fig. 1. Geometry of problem.

Equations (7) and (10) are completely similar and therefore the solution of any barotropic problem can be interpreted as a solution of the corresponding baroclinic problem in the case of weak stratification in the form (9). For this reason we will henceforth limit ourselves to study of the solution of equation (7) in the region shown in Fig. 1, when the right-hand side is determined by expression (8). Concentrating attention only on the effect exerted by the form of the boundary, we assume $h(x,y) \equiv 0$. We will also assume that the undisturbed flow is easterly; upstream the disturbances caused by the influence of change in channel width disappear.

Assume that $\Phi(y) = \varphi(-\infty, y)$; then by virtue of geostrophicity the velocity profile of the oncoming flow is determined by the expression $U(y) = -\Phi'(y)$. Assuming $\Phi(0) = 0$, for the total current discharge we obtain

$$Q = \int_0^1 U(y)\, dy = -\Phi(1).$$

From (7) and (8) at the limit $x \to -\infty$ we find

$$\Phi''(y) + by = C\Phi + D. \tag{11}$$

This equation determines the possible transverse profiles of undisturbed flow. As a simplification we will limit ourselves to a case when they are symmetric relative to the axis of the channel upstream: $\Phi'(y) = \Phi'(1-y)$, $0 < y < 1$. The solutions of equation (11) satisfying the mentioned conditions have the form

$$\Phi(y) = \begin{cases} \frac{b}{C}y + \frac{1}{2}\left(Q + \frac{b}{C}\right)\left[\frac{\sin\sqrt{-C}\left(\frac{1}{2} - y\right)}{\sin\frac{\sqrt{-C}}{2}} - 1\right], & C < 0,\ C \neq -4\pi^2 n^2 \\ \frac{b}{12}\, y\,(1 - 2y)(y - 1) - Qy, & C = 0 \\ \frac{b}{C}y + \frac{1}{2}\left(Q - \frac{b}{C}\right)\left[\frac{\operatorname{sh}\sqrt{C}\left(\frac{1}{2} - y\right)}{\operatorname{sh}\frac{\sqrt{C}}{2}} - 1\right], & C > 0. \end{cases} \tag{12}$$

The case $C = -4n^2\pi^2$ (n is a non-negative whole number) is special and corresponds to a shearless flow $\Phi(y) = -b/4\pi^2 n^2\, y$ with the discharge $Q = b/4\pi^2 n^2$. The parameter $D = 1/2(Q + b/C)$ characterizes the velocity shear in the oncoming flow; with $D = 0$ we obtain a shearless current with the velocity $U = Q$. Expressions (12) determine a two-parameter (Q, C) family of on-coming flows and make it possible to investigate the influence of discharge volume and velocity shear.

Using the possibility of an analytical continuation of the $\Phi(y)$ function into the region $1 < y < \ell$, we will seek a solution of the problem in the form

$$\varphi(x, y) = \Phi(y) + \psi(x, y). \tag{13}$$

Substituting this expression into (7) and taking (8) and (11) into account, for disturbance of the stream function $\psi(x, y)$ we obtain the equation

$$\Delta\psi - C\psi = 0. \tag{14}$$

The boundary conditions (see Fig. 1) are:

$$\psi(x, 0)=0, \quad -\infty<x<\infty, \tag{15}$$

$$\psi(x, 1)=0, \quad -\infty<x<0, \tag{16}$$

$$\psi(x, l)=\Phi(1)-\Phi(l), \qquad 0<x<\infty, \tag{17}$$

$$\psi(0, y)=\Phi(1)-\Phi(y), \quad 1<y<l, \tag{18}$$

$$\psi(-\infty, y)=0, \quad 0<y<1, \tag{19}$$

$$\psi(\infty, y) \text{ is limited, } 0<y<l. \tag{20}$$

Mathematically the formulated problem is similar to that in [9], where a study was made of an air flow over a step; we will use the same solution method as in the cited study.

We introduce the functions

$$Y_n(y)=\sqrt{2}\sin n\pi y, \quad n=1, 2, \ldots \qquad \tilde{Y}_m(y)=\sqrt{\frac{2}{l}}\sin\frac{m\pi y}{l}, \qquad m=1, 2, \ldots,$$

forming the full orthonormalized systems for the segments $0<y<1$ and $0<y<l$ respectively. Assume

$$Y(y)=\begin{cases}\dfrac{\sin\sqrt{-C}\,y}{\sin\sqrt{-C}\,l}, & C<0, \quad \sqrt{-C}\,l\neq k\pi\\ y/l, & C=0\\ \dfrac{\mathrm{sh}\,\sqrt{C}\,y}{\mathrm{sh}\,\sqrt{C}\,l}, & C>0.\end{cases}$$

We will exclude from consideration the "resonance" cases $\sqrt{-C}\,l = k\pi$ for whole k, for which there are no solutions disappearing upstream. We will agree on the notations:

$$N=\begin{cases}\left[\dfrac{\sqrt{-C}}{\pi}\right], & C<0\\ 0, & C\geq 0;\end{cases} \qquad M=\begin{cases}\left[\dfrac{l\sqrt{-C}}{\pi}\right], & C<0.\\ 0, & C\geq 0;\end{cases}$$

$$\begin{aligned}\varkappa_n&=\sqrt{C+(n\pi)^2}, \quad n\geq N+1,\\ \omega_m&=\sqrt{-C-\left(\frac{m\pi}{l}\right)^2}, \quad m\leq M,\\ \delta_m&=\sqrt{C+\left(\frac{m\pi}{l}\right)^2}, \quad m\geq M+1,\end{aligned} \tag{21}$$

where [z] is the whole part of the number z. Obviously, in all cases $M\geq N$, since by assumption $l\geq 1$. The solution of equation (14) satisfying the conditions (15)-(17), (19) and (20) has the form

$$\psi(x, y) = \sum_{n=N+1}^{\infty} A_n e^{\sigma_n x} Y_n(y), \quad x < 0, \tag{22}$$

$$\psi = \tilde{\psi}(x, y) \equiv [\Phi(1) - \Phi(l)] \, Y(y) +$$

$$+ \sum_{m=1}^{M} [B_m \cos \omega_m x + C_m \sin \omega_m x] \, \tilde{Y}_m + \sum_{m=M+1}^{\infty} B_m e^{-\delta_m x} \, \tilde{Y}_m(y), \tag{23}$$

where the arbitrary constants A_n, B_m, C_m must be determined from the condition of continuous conjugation of ψ and $\partial\psi/\partial x$ in the section $x = 0$, $0 < y < 1$, and also from the boundary condition (18). These requirements lead to the expressions

$$\tilde{\psi}(0, y) = \begin{cases} \psi(0, y), & 0 < y < 1, \\ \Phi(1) - \Phi(y), & 1 < y < l, \end{cases} \tag{24}$$

$$\frac{\partial \tilde{\psi}}{\partial x}(0, y) = \frac{\partial \psi}{\partial x}(0, y), \quad 0 < y < 1. \tag{25}$$

By multiplying (24) by $\tilde{Y}_k$ and then integrating the result for y in the segment $(0, l)$, with (22) and (23) taken into account, we obtain the equations

$$B_k = \sum_{n=N+1}^{\infty} A_n Q_{kn} + R_k, \quad k = 1, 2, \ldots, \tag{26}$$

where we use the notation

$$Q_{kn} = \int_0^1 \tilde{Y}_k Y_n \, dy, \quad k = 1, 2, \ldots; \quad n = N+1, \ldots,$$

$$R_k = \int_1^l [\Phi(1) - \Phi(y)] \, \tilde{Y}_k \, dy - [\Phi(1) - \Phi(l)] \int_0^l Y \tilde{Y}_k \, dy, \quad k = 1, 2, \ldots,$$

Proceeding in the same way with (25), but now with respect to the Y_s functions in the segment (0, 1), we obtain

$$\sum_{m=1}^{M} \omega_m C_m Q_{ms} - \sum_{n=M+1}^{\infty} \delta_m B_m Q_{ms} = \sum_{n=N+1}^{\infty} \sigma_n A_n \delta_{ns}, \quad s = 1, 2, \ldots, \tag{27}$$

where δ_{ns} is the Kroneker symbol. Using (26) it is possible to exclude B_k, substituting which into (27) we find

$$\sum_{m=1}^{M} \omega_m Q_{ms} C_m - \sum_{n=N+1}^{\infty} \left(\sum_{m=M+1}^{\infty} \delta_m Q_{ms} Q_{mn} + \sigma_n \delta_{ns} \right) A_n = \tag{28}$$

$$= \sum_{m=M+1}^{\infty} \delta_m Q_{ms} R_m, \quad s = 1, 2, \ldots.$$

We will solve the infinite system of equations (28) by the truncation method, assuming $A_n \equiv 0$ for all $n > L \geqslant N + 1$. The total number of unknowns in this case is equal to $K = L + M - N$ and is formed from M coefficients $C_1, C_2, \ldots, C_M$ and $L - N$ coefficients $A_{N+1}, A_{N+2}, \ldots, A_L$. Accordingly, for the closing of system (28) it is necessary to retain the K first equations, s = 1, 2,...,K, which are solved by some standard method. When the C_m and A_n coefficients are found, B_k are determined using formulas (26).

The Q_{kn} values are dependent only on the geometry of the problem and have the form

$$Q_{kn}=\begin{cases}\dfrac{2n(-1)^n\sin\frac{k\pi}{l}}{\pi\sqrt{l}\left[\left(\frac{k}{l}\right)^2-n^2\right]}, & \dfrac{k}{l}\neq n\\ \dfrac{1}{\sqrt{l}}, & \dfrac{k}{l}=n.\end{cases}$$

The R_k coefficients are dependent on the profile of the undisturbed flow, determined by the $\Phi(y)$ function; in the considered case they also are computed in finite form, but we will not cite the corresponding expressions due to their unwieldy form.

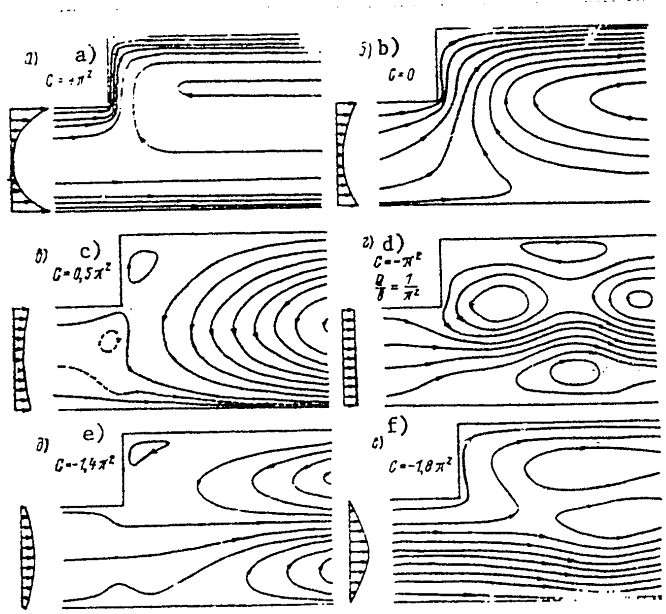

Fig. 2. Computed patterns of streamlines with stipulated discharge and different forms of symmetric profile of undisturbed flow.

Now we will note some general properties of the constructed solution. The behavior of $\varphi(x,y)$ downstream is of the greatest interest. When $x\to\infty$ from (13) and (23) we asymptotically have

$$\varphi(x, y) \approx \Phi_\infty(x, y) \equiv \Phi(y) + [\Phi(1) - \Phi(l)] \; Y(y) + \\ + \sum_{m=1}^{M} (B_m \cos \omega_m x + C_m \sin \omega_m x) \; \tilde{Y}_m(y). \tag{29}$$

Meandering and eddy formation are determined by the wave modes of the last sum in (29). According to (21), the condition for the appearance of these terms has the form

$$C \leqslant -\left(\frac{\pi}{l}\right)^2. \tag{30}$$

It is convenient to separate the considered profiles of on-coming flows (12) into three types: trigonometric ($C < 0$), parabolic ($C = 0$) and hyperbolic ($C > 0$). Thus, for the appearance of wave modes it is necessary that the profile be of the trigonometric type; on the other hand, an adequate condition for the absence of wave modes is that the profile belong to the parabolic or hyperbolic type. If wave modes are absent, in place of (29) we have for large x

$$\varphi(x, y) \approx \overline{\Phi}_\infty(y) \equiv \Phi(y) + [\Phi(1) - \Phi(l)] \; Y(y),$$

that is, purely zonal flow. It may be found that in some part of the channel $\overline{\Phi}_\infty'(y) > 0$; this corresponds to the appearance of return flows. In a more general case (29) with $M \geqslant 1$ the pattern will be complicated by the presence of individual eddies forming a quasiperiodic structure.

On the basis of the described model we carried out several tens of numerical experiments with different values of the determining parameters C, Q and ℓ. The computed fields $\varphi(x,y)$ were printed out in the form of isolines on an automatic digital printout unit in the region $-0.5 < x < 2.0$. Here we will illustrate only one series of experiments carried out with a fixed discharge Q and width ℓ for several values of the C parameter. This makes it possible to trace the influence of curvature (or shear) of the undisturbed flow on the currents in the transition region.

Figure 2 shows the computed streamlines φ/b = const for the case $Q/b = 1/\pi^2$ and $\ell = 1.7$ with $C/\pi^2 = 4$ (a), 0 (b), -0.5 (c), -1 (d), -1.4 (e), -1.8 (f). In all the figures the corresponding velocity profile of the undisturbed flow is shown at the left.

Figure 2a,b shows hyperbolic and parabolic cases with concave profiles of the velocity of the undisturbed flow; here M = N = 0, the wave modes are absent. Return currents can be seen in the expanding part of the channel. In the first case they occupy only the central part, and in the second case -- the lower half of the region.

Figure c-f shows the patterns of currents for trigonometric profiles varying from convex to concave. In these cases N = 0, M = 1 (c), N = M = 1 (d), N = 1, M = 2 (e,f) respectively. The appearance of wave modes makes the field of currents more complex and leads to the appearance of eddy formations.

Theoretical conclusions and numerical experiments show that in the considered model eddy formation and meandering of the flow in a widening channel are essentially dependent on the C parameter. The condition for the appearance of wave modes (30) can assume a different form. Using (11), by differentiation we find the identity

$$-U''(y) + b = -CU(y),$$

by means of which in place of (30) we obtain

$$\frac{U'' - b}{U} \leqslant \left(\frac{\pi}{l}\right)^2. \tag{31}$$

This expression can be interpreted as the instability condition [1], as a result of which a process of meandering and eddy formation arises. It is of interest to apply this criterion to real observation data on the transverse structure of currents.

By multiplying (31) by U(y) and integrating for the width of the channel upstream, we obtain a "softened" condition for the appearance of wave modes

$$U'(1) - U'(0) \leqslant \frac{\pi^2}{l^2} Q + b. \tag{32}$$

Conditions (31) and (32) are equivalent for shearless currents, for which they are always satisfied.

What has been said above leads to the conclusion that any attempts at prediction of the behavior of currents of the Gulf Stream, Kuroshio or Circumpolar Current type must be based on a study not only of changes in their discharge, but (which is evidently more important) also on an analysis of the transverse structure upstream from the considered region.

BIBLIOGRAPHY

1. Arnol'd, V. I., "Conditions of Nonlinear Stability of Plane Stationary Curvilinear Currents of an Ideal Fluid," DOKLADY AN SSSR (Reports of the USSR Academy of Sciences), Vol 162, No 5, 1965.

2. Kozlov, V. F., "Model of Meandering of Inertial Currents in a Baroclinic Ocean," IZV. AN SSSR, FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), Vol 6, No 9, 1970.

3. Kozlov, V. F., Ten, Ye. V., "One Stationary Problem in Mesooceanology," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 1, 1978.

4. Arnason, G., Welsh, J. G., "Numerical Prediction of the Gulf Stream by Means of the Equivalent Barotropic Model," MITT. INST. MEERESKUNDE, Univ. Hamburg, No 10, 1968.

5. Bretherton, F. P., Haidvogel, D. B., "Two-Dimensional Turbulence Above Topography," J. FLUID MECH., Vol 78, No 1, 1976.

6. Gadgil, S., "Structure of Jets in Rotating Systems," J. FLUID. MECH., Vol 47, No 3, 1971.

7. Gadgil, S., "Time-Dependent Topographic Meandering of a Baroclinic Current," DYN. ATMOS. AND OCEANS, Vol 1, No 2, 1976.

8. Godfrey, J. S., Robinson, A. R., "The East Australian Current as a Free Inertial Jet," J. MARINE RES., Vol 29, No 3, 1971.

9. Jones, O. K., "The Flow of a Stratified Fluid Over a Vertical Step," TELLUS, Vol 22, No 5, 1970.

10. Luyten, J. R., Robinson, A. R., "Transient Gulf Stream Meandering. Part II: Analysis Via a Quasigeostrophic Time-Dependent Model," J. PHYS. OCEANOGR., Vol 4, No 2, 1974.

11. McCartney, M. S., "The Interaction of Zonal Currents With Topography With Applications to the Southern Ocean," DEEP SEA RES., Vol 23, No 5, 1976.

12. McCreary, J. P., White, W. B., "On a Theory of the Kuroshio Meander," DEEP SEA RES., Vol 26A, No 3, 1979.

13. McIntyre, M. E., "On Stationary Topography-Induced Rossby Wave Patterns in a Barotropic Zonal Current," DTSCH. HYDR. Z., Vol 21, No 5, 1978.

14. Niiler, P. P., Robinson, A. R., "The Theory of Free Inertial Jets. II. A Numerical Experiment for the Path of the Gulf Stream," TELLUS, Vol 19, No 4, 1967.

15. Porter, G. H., Rattray, M., Jr., "The Influence of Variable Depth on Steady Zonal Barotropic Flow," DTSCH. HYDR. Z., Vol 17, No 4, 1964.

16. Robinson, A. R., Niiler, P. P., "The Theory of Free Inertial Currents. I. Path and Structure," TELLUS, Vol 19, No 2, 1967.

17. Robinson, A. R., Gadgil, S., "Time-Dependent Topographic Meandering," GEOPHYS. FLUID DYN., Vol 1, No 4, 1970.

18. Robinson, A. R., Taft, B. A., "A Numerical Experiment for the Path of the Kuroshio," J. MARINE RES., Vol 30, No 1, 1972.

19. Robinson, A. R., Luyten, J. R., Flierl, G., "On the Theory of Thin Rotating Jets: A Quasigeostrophic Time Dependent Model," GEOPHYS. FLUID DYN., Vol 6, GEOPHYS. FLUID DYN., Vol 6, No 3, 1975.

20. Robinson, A. R., "Dynamics of the Kuroshio Current: Experimental and Theoretical Studies from South of Kyushu to the Izu-Ogasawara Ridge," BRUUN MEM. LECT., Paris, 1976.

21. Saint-Guily, B., "Les Meanders des Veines de Courant dans les Ocean," BULL. INST. OCEANOGR., No 1108, 1957.

22. Solomon, H., "Comment on a Theory of the Kuroshio Meander," DEEP SEA RES., Vol 25, No 10, 1978.

23. Stern, M. E., "Minimal Properties of Planetary Eddies," J. MARINE RES., Vol 33, No 1, 1975.

24. Thacker, W. C., "Spatial Growth of Gulf Stream Meanders," GEOPHYS. FLUID DYN., Vol 7, No 3-4, 1976.

25. Warren, B. A., "Topographic Influences on the Path of the Gulf Stream," TELLUS, Vol 15, No 2, 1963.

26. White, W. B., McCreary, J. P., "On the Formation of the Kuroshio Meander and its Relationship to the Large-Scale Ocean Circulation," DEEP SEA RES., Vol 23, No 1, 1976.

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COMPUTATION OF CURRENT VELOCITY IN THE QUASI-ISOTHERMAL LAYER OF THE EQUATORIAL ZONE IN THE OCEAN

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 78-81

[Article by A. B. Polonskiy, Marine Hydrophysical Institute, submitted for publication 29 Aug 79]

[Text]

Abstract: Within the framework of the model proposed in [5] it was possible to obtain the velocity profile in the quasi-isothermal layer at the equator with different coefficients of turbulent exchange. The balance of turbulent energy is analyzed.

In [5] the author proposed a two-layer stationary model of a baroclinic ocean, including the equator, and carried out a number of numerical experiments for computing the characteristics of the quasi-isothermal layer (QL). Use was made of depth-integrated equations of motion, continuity and heat balance equations. No assumptions were made concerning the mechanism of vertical turbulent exchange of heat and momentum in the QL. Such an approach does not make it possible to determine the current velocity profile in the QL. Using the hypothesis of proportionality of the stress and strain tensors adopted in the theory of currents and stipulating the coefficient of turbulent exchange it is easy to obtain a solution of the equations of motion in a linear approximation and find the current velocity profile in the QL. However, the following question remains open: to what degree is the solution dependent on the behavior of the coefficient of turbulent exchange. Evidently, in the case of a strong dependence it is desirable to confine oneself to computations of the total fluxes within the limits of the QL, as was done in [5], or consider the coefficient of turbulent exchange to be an unknown value and find it from additional considerations.

In [2] a similar problem was solved within the framework of the classical Ekman problem for purely drift currents in the middle latitudes. We will examine a baroclinic ocean, including the equator, with a stipulated temperature model [5]. We will assume that the baroclinic pressure gradients have already been determined from solution of the boundary value problem, and we must find the velocity profile on a stipulated vertical.

The equation of motion and the boundary conditions have the following form:

$$\begin{cases} (ku')' + fv - ru = G_x - g\alpha T^\circ_x z, \\ (kv')' - fu - rv = G_y - g\alpha T^\circ_y z. \end{cases} \tag{1}$$

$$z = 0:\quad ku' = -\frac{\tau^x}{\rho},\quad kv' = -\frac{\tau^y}{\rho},$$
$$z = h:\quad ku' = kv' = 0. \tag{2}$$

Here k is the coefficient of turbulent exchange,

$$G_x = g\alpha\,[T^\circ_x\,(h + c) + T^\circ h_x],$$
$$G_y = g\alpha\,[T^\circ_y\,(h + c) + T^\circ\,(h_y + c_y)],$$
$$g\alpha = 0{,}25\ \frac{\text{см}}{^\circ\text{C}\cdot c^2},\quad r = 10^{-6}\ c^{-1},$$

T° and h are temperature and thickness of the QL, c is the thickness of the thermocline, considered to be a known function of y, the prime denotes differentiation for z, the oz-axis is directed downward.

We recall that in writing (1) use was made of the following model for temperature:

$$T = \begin{cases} T^\circ & \text{when}\quad z \leqslant h \\ T^\circ\ \exp\frac{h - z}{c} & \text{when}\quad z > h. \end{cases} \tag{3}$$

We introduce the characteristic scales: length L (horizontally) and h (vertically), velocity U°, temperature T° and limit ourselves to an examination of a narrow zone near the equator within which $f/r \leqslant 1$. Then for k it is natural to introduce the scale $h^2 r$, and for τ/ρ the scale $U^\circ hr$. We note that it is possible to select the characteristic scales somewhat differently, but for the purposes of this article this choice is not fundamental. Multiplying the second equation of system (1) by i and adding it to the first, we obtain one equation for complex values. Now we will proceed to a dimensionless form. Assuming that $k = k_0$ = const, we obtain

$$m'' - j^2 m = G - Tz. \tag{4}$$

Here
$$m = u + iv,\quad j^2 = \frac{i\mu_0 + 1}{k_0},\quad \mu_0 = \frac{f}{r},$$
$$G = \mu_1\,(G_x + iG_y),\quad T = \mu_2\,(T^\circ_x + iT^\circ_y),$$
$$\mu_1 = \frac{g\alpha T_0\,(c + h)}{k_0 LrU_0},\quad \mu_2 = \frac{g\alpha T^\circ h}{k_0 LrU_0}. \tag{5}$$
$$z = 0:\quad m' = -\tau,\quad z = 1:\ m' = 0,$$

k_0 is the dimensionless value of the coefficient of turbulent exchange.

The solution (4), with the boundary conditions (5) taken into account, has the form

$$m = \frac{(\tau + T/j^2)\ \text{ch}\,[j\,(1 - z)]}{j\ \text{sh}\,j} - \frac{G - Tz}{j^2} - \frac{T}{j^3}\frac{\text{ch}\,(jz)}{\text{sh}\,j}. \tag{6}$$

Now we will assume that the coefficient of turbulent exchange decreases with depth in conformity to the following law:

$$k = k_0 e^{-z}. \tag{7}$$

The dependence (7) models the characteristic behavior of the coefficient of turbulent exchange with depth [3]. In this case system (1) is reduced to the equation

$$m'' - m' - j^2 m e^z = (G - Tz)\, e^z. \tag{8}$$

Solving (8), with allowance for the boundary conditions (5), using power series and limiting ourselves to terms of the order z^3, we obtain the following approximate solution:

$$\begin{aligned} m \cong & -(\tau + 1)\, z \left(1 + \frac{z}{2} + \frac{1 + j^2}{6} z^2\right) + z + \\ & + \frac{G + 1}{2} z^2 + \frac{\tau (j^2 + 5) - 4\, G + T}{4\, j^2} \left(1 + \frac{(jz)^2}{2} + \frac{j^2 z^3}{3}\right) + \\ & + \frac{2\, G + j^2 + 1 - T}{6} z^3. \end{aligned} \tag{9}$$

In expressions (5), (6), (8) τ is normalized to k_0, that is, in actuality is the ratio of the dimensionless wind stress ($\tau = \tau^x + i\tau^y$) to the dimensionless coefficient of turbulent exchange (k_0). Separating the real and fictitious parts in (6) and (9), we obtain an expression for the velocity components.

Now we will examine a current at the equator. As a simplification we will assume that the meridional velocity component is absent (the motion is symmetric relative to the equatorial plane). We note that this limitation is not fundamental. From (6) and (9) we find

$$u = \frac{(\tau + T)\, \mathrm{ch}\, (1 - z)}{\mathrm{sh}\, 1} - (G - Tz) - T \frac{\mathrm{ch}\, z}{\mathrm{sh}\, 1}, \tag{6'}$$

$$u \cong \tau \left(1{,}5 - z + \frac{z^2}{4} + \frac{z^3}{6}\right) + \frac{T}{4}\left(1 + \frac{z^2}{2} - \frac{z^3}{3}\right) - G. \tag{9'}$$

We then assume L = $5 \cdot 10^8$ cm, T = 20°C, c = $2 \cdot 10^4$ cm,

$$U^\circ = 50\ \text{см/с},\ h = 3 \cdot 10^3\ \text{см},\ \tau^x = -0{,}5\ \text{dyne/cm}^2. \tag{10}$$

In accordance with the observational data and the results of computations [5], [6] T° and h decrease from west to east. This means that G and T are negative values. Now it is possible to determine the velocity profiles from equations (6') and (9'), which, with (10) taken into account, are rewritten in the form

$$u \cong \frac{-3{,}9}{\mathrm{sh}\, 1}\, \mathrm{ch}\, (1 - z) - (0{,}6\, z - 4{,}6) + \frac{\mathrm{ch}\, z}{\mathrm{sh}\, 1}, \tag{6''}$$

$$u \cong -3{,}3 \left(1{,}5 - z + \frac{z^2}{4} + \frac{z^3}{6}\right) - 0{,}15 \left(1 + \frac{z^2}{2} - \frac{z^3}{3}\right) + 4{,}6. \tag{9''}$$

In (6'), (6"), (9'), (9") we assumed k_0 = 1. The velocity profiles obtained using (6") and (9") are cited in the figure.

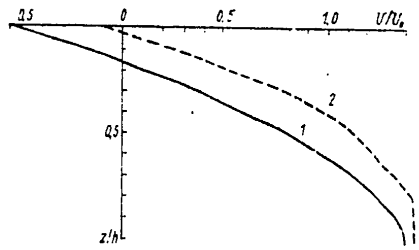

Fig. 1. Current velocity profiles at equator with different coefficients of turbulent exchange.

$$1)\ k = \frac{k_0}{h^2 r} e^{-z/h}, \quad 2)\ k = \frac{k_0}{h^2 r}.$$

The quantitative discrepancies in the distribution of velocity with depth in the two considered cases are extremely great. We note that some difference in velocity at the lower boundary of the QL is associated for the most part with the error in the approximate solution (9), which here is maximum.

Using the derived solutions we will evaluate the principal terms in the equation for the budget of turbulent energy within the limits of the QL -- production due to velocity shear and dissipation.

Determining

$$\int_0^1 k(u')^2\, dz$$

and proceeding to dimensional values, we obtain the integral production of turbulent energy in the QL. It appears that the principal contribution to the production of turbulent energy will be from drift velocity shear. The contribution of the velocity shear arising due to baroclinic effects is less than 10%. The characteristic value for the production of turbulent energy in the QL with the adopted values of the coefficient of turbulent exchange is 10^{-2} cm^2/sec^3, which coincides with the experimental data in [7].

Now we will determi .e integral dissipation within the limits of the QL, using the known expression [4]

$$\varepsilon = c_0 \frac{k^3}{l^4}, \tag{11}$$

where ε is dissipation, l is the turbulence scale, $c_0 = 0.0625$.

Integrating (11) within the limits of the QL, we obtain the integral dissipation. We will assume for the evaluation that the turbulence scale is proportional to the thickness of the QL [1, 5]. Then it appears that integral dissipation with a constant coefficient of turbulent exchange is approximately three times greater than when using dependence (7). The characteristic dissipation value in this case is too low by several orders of magnitude in comparison with the experimental data in [7]. In order to obtain the dissipation of turbulent energy in the QL of the order of magnitude 10^{-2} cm^2/sec^3 it is necessary to increase the coefficient of turbulent exchange to a value of the order of magnitude 10^2 cm^2/sec. (We recall that we assumed $k = h^2 r = 9$ cm^2/sec).

In conclusion we will formulate the principal conclusions:

1. The velocity profile in the QL is determined to a considerable degree by the coefficient of turbulent exchange. Accordingly, it is desirable, using integral models similar to those proposed in [5], to limit ourselves to a determination of the total fluxes into the QL.

2. It is necessary to make precise use of the equation for the budget of turbulent energy in computations of the characteristics of the QL with stipulated coefficients of turbulent exchange, since an insignificant change in these coefficients can lead to an incorrect description of the balance of turbulent energy in the QL.

BIBLIOGRAPHY

1. Kitaygorodskiy, S. A., "Dynamics of the Upper Thermocline in the Ocean: Results of Science and Technology," OKEANOLOGIYA (Oceanology), VINITI, Vol 4, Moscow, 1977.

2. Kozlov, V. F., "Influence of Change in the Coefficient of Vertical Exchange on Drift Currents," IZV. AN SSSR, SERIYA GEOFIZICH. (News of the USSR Academy of Sciences, Geophysical Series), No 7, 1963.

3. Marchuk, G. I., Kochergin, V. P., Klimok, V. I., Sukhorukov, V. A., "Mathematical Modeling of Seasonal Variability of the Surface Turbulent Layer in the Ocean," IZV. AN SSSR, FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), Vol 14, No 9, 1978.

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

5. Polonskiy, A. B., "Computation of the Characteristics of the Quasi-Isothermal Layer in the Equatorial Zone of the Ocean," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 3, 1980.

6. "The Lomonosov Current," CB. STATEY MGI (Collection of Articles of the Marine Hydrophysical Institute Ukrainian Academy of Sciences), Vol 34, 1963.

7. Jones, J. H., "Vertical Mixing in the Equatorial Undercurrent," J. PHYS. OCEANOGR., Vol 3, No 3, 1973.

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COMPUTATION OF THE CONCENTRATION OF GLOBAL ATMOSPHERIC IMPURITIES IN RIVERS AND CLOSED WATER BODIES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 82-89

[Article by Candidate of Physical and Mathematical Sciences K. P. Makhon'ko, Institute of Experimental Meteorology, submitted for publication 14 Mar 80]

[Text]

Abstract: A study was made of the process of contamination of rivers and closed water bodies by an impurity entering from an atmospheric source of a global scale. The contribution of atmospheric fallout of impurity to the contamination of water in the territory of a basin and ocean area, the erosion from the soil surface of the basin and the role of bottom deposits are taken into account. In the case of rivers an allowance is made for the influence of runoff and the dilution of contaminated waters by pure ground water. A comparison of the results of theoretical computations of the change in the concentrations of Sr^{90} in the rivers of the Moscow area and the USSR during 1954-1975 with data from direct observations indicated good agreement.

The contamination of rivers and water bodies occurs as a result of local effluent and the action of sources of a global scale. An example of this type of source is the atmosphere -- the carrier of atmospheric dust, cosmogenic isotopes, radioactive products of powerful nuclear explosions injected earlier into the stratosphere, etc. All these substances have a common criterion -- the global scale of their entry from the atmosphere onto the underlying surface; henceforth for brevity they will be called global atmospheric impurities.

We will examine the entry of global atmospheric impurities into rivers and water bodies.

The contamination of a closed water body is caused by the following factors: a) the entry into it of atmospheric precipitation flowing from the surface of the soil-vegetation cover and carrying a contaminating impurity falling from the atmosphere in the course of the year; b) falling of the impurity directly on the surface of the water body; c) partial washing-out from the soil of an impurity accumulating in it during all the preceding years due to the falling of precipitation [9]; d) sorption interaction between an impurity present in the water and an impurity in the bottom deposits of a water body [10]; e) biogenous migration of an impurity. Biogenous migration can usually be neglected [11]. Then, summing the concentrations created in the water body by the factors enumerated above, the concentration of impurity in the water is represented in the form

$$C = C_p + C_M + C_\sigma + C_E. \tag{1}$$

Now we will examine the individual terms in (1). The concentration created by the runoff of atmospheric fallout of an impurity from the area of a drainage basin into a water body in the current year is represented in the form

$$C_p = \frac{P(t)}{h_p(t)} k_p, \tag{2}$$

where P is the annual atmospheric fallout of the impurity, h_p is the thickness of the annual layer of atmospheric precipitation which in the form of surface runoff carries the impurity falling in its basin to the water body, t is time, k_p is the coefficient of dilution of runoff water contaminated by atmospheric fallout of the impurity in the waters of the water body.

The concentration forming as a result of fallout of an impurity from the atmosphere directly onto the surface of a deep water body, as was demonstrated in [4], can be represented in the form

$$C_M = \int_0^t \frac{P(\theta)}{H(\theta)} e^{-\Lambda_M (t-\theta)} d\theta, \tag{3}$$

where H is the thickness of the quasihomogeneous layer of water in the water body, Λ_M is the constant elimination of impurity from a layer with the thickness H into the deeper layers of the water body. If the impurity decays with time (for example, there is radioactive decay of the isotope), then $\Lambda_M = \Lambda'_M + \lambda$, that is, also includes the constant of this decay λ.

For a shallow water body, mixed to the bottom, the H value acquires the sense of its mean depth and Λ_M is the constant elimination of the impurity from the water into the bed of the water body.

The C_σ value is the concentration of impurity in the water body, forming as a result of removal of that part of the impurity accumulating on the soil surface due to the falling of precipitation:

$$C_\sigma = \frac{\sigma(t)}{\delta(t)\nu(t)} k_\sigma, \tag{4}$$

where σ is the active store of impurity in the surface soil layer with the thickness δ which is subject to washing out, ν is the water volume which is capable of being replaced in a unit volume of soil when the impurity is washed out by atmospheric precipitation, k_σ is the coefficient of dilution of runoff water contaminated by the impurity washed out of the soil in the waters of the water body.

The ν value is dependent on the moisture content and moisture capacity of soils in the basin, the dissection of relief (in particular, due to slope steepness) and similar factors. Some of the impurity in the course of migration gradually leaves the soil layer subject to washing out or passes into fixed form, being held firmly in the soil layer. The temporal change in the active supply of impurity in the surface layer of soil with the thickness δ can be represented in the following way:

$$\frac{d\sigma}{dt} = P(t) - \Lambda_\sigma \sigma, \tag{5}$$

where $\Lambda = \Lambda_\sigma' + \lambda$ is the constant elimination of the mobile part of the impurity from the surface layer of the soil, also taking its decay into account.

Assuming $\sigma = 0$ with $t = 0$, we obtain a solution of equation (5) in the form

$$\sigma = \int_0^t P(\theta)\, e^{-\Lambda_\sigma (t-\theta)}\, d\theta. \tag{6}$$

The concentration of impurity forming as a result of exchange processes with bottom deposits in the water of a shallow water body mixed to the bottom, C_E, is determined by the dynamics of sorption-desorption processes. If the concentration of the impurity in the water is maintained at the same level ($dC/dt = 0$), a mobile equilibrium is established relatively rapidly between the content of impurity in the water and in the bottom deposits; this equilibrium is determined by the distribution coefficient K_d. We will neglect the slow process of diffusion of the impurity into the depth of the layer of bottom deposits [10]. With an increase in the concentration in the water ($dC/dt > 0$) the equilibrium is impaired and a flux of impurity into the bottom deposits arises; with a decrease in the concentration in the water ($dC/dt < 0$) a flux of impurity arises from the bottom deposits into the water. Thus, the C_E value can be assumed to be proportional to the rate of change in the concentration of impurity in the river water and be written in the form

$$C_E = -\frac{1}{\Lambda_E}\frac{dC}{dt}, \tag{7}$$

where $\Lambda_E = \Lambda'_E + \lambda$ is the constant of exchange of the impurity between the active layer of bottom deposits and water. This value, evidently, is determined by the rate and character of the bottom currents, the type of bottom deposits and similar factors. Now substituting (2)-(4), (6) and (7) into (1), we obtain the equation

$$\frac{1}{\Lambda_E}\frac{dC(t)}{dt} + C(t) = k_p \frac{P(t)}{h_p(t)} + \int_0^t \frac{P(\theta)}{H(\theta)}\, e^{-\Lambda_M (t-\theta)}\, d\theta + \tag{8}$$

$$+ k_\sigma \int_0^t \frac{P(\theta)}{\delta(\theta)\, v(\theta)}\, e^{-\Lambda_\sigma (t-\theta)}\, d\theta.$$

Assuming $C = 0$ with $t = 0$, we obtain the general solution (8) in the form of quadratures

$$C = k_p \Lambda_E \int_0^t \frac{P(\theta)}{h_p(\theta)}\, e^{-\Lambda_E (t-\theta)}\, d\theta +$$

$$+ \Lambda_E \int_0^t e^{-\Lambda_E (t-\xi)} \int_0^\xi P(\theta) \left[\frac{e^{-\Lambda_M (\xi-\theta)}}{H(\theta)} + \right. \tag{9}$$

$$\left. + k_\sigma \frac{e^{-\Lambda_\sigma (\xi-\theta)}}{\delta(\theta)\, v(\theta)} \right] d\xi\, d\theta.$$

For water bodies which are great in area but shallow, which are mixed to the bottom, the H value, as already noted above, acquires the sense of the mean depth of the water body and in (9) can be removed from the integral sign. In this case for some climatic zones, where evaporation of moisture from the surface of the water body is great, the right-hand side of formula (9) must be additionally multiplied by the concentration coefficient k_f.

The expressions for the concentration coefficient and the dilution coefficients in (2) and (4) can be written in the form

$$k_f = \left(1 - \frac{\Delta V}{V}\right)^{-1}, \quad k_p = \frac{h_p S_p}{(H + h_p) S} \approx \frac{h_p S_p}{V}, \quad k_з = \frac{\nu}{V + \nu} \approx \frac{\nu}{V}, \tag{10}$$

where ΔV is the quantity of water evaporating from a water body with the volume V, S_p and S are the areas of the drainage basin and water surface respectively.

If the quantity of fallout of the impurity from the atmosphere does not change with time, then assuming the mean values for the parameters h_p, δ and ν varying little from year to year, the solution (9) is easily represented in analytical form

$$C = aP - (be^{-\Lambda_з t} + ce^{-\Lambda_M t} + de^{-\Lambda_E t})\, P, \tag{11}$$

where

$$a = \frac{k_f k_p}{h_p} + \frac{k_f k_з}{\delta\nu\Lambda_з} + \frac{k_f}{H\Lambda_M}, \quad b = \frac{k_f k_з \Lambda_E}{\delta\nu\Lambda_з(\Lambda_E - \Lambda_з)},$$

$$c = \frac{k_f \Lambda_E}{H\Lambda_M(\Lambda_E - \Lambda_M)}, \quad d = \frac{k_f k_p}{h_p} + \frac{k_f k_з}{\delta\nu(\Lambda_з - \Lambda_E)} + \tag{12}$$

$$+ \frac{k_f}{H(\Lambda_M - \Lambda_E)}.$$

In the case of small, shallow water bodies the fallout of impurity onto the water surface can be neglected. Then in (1) $C_M = 0$, the second term on the right-hand side of (8) becomes equal to zero and the concentration of impurity in such a water body will be described by the formula

$$C = k_f k_p \Lambda_E \int_0^t \frac{P(\theta)}{h_p(\theta)} e^{-\Lambda_E(t-\theta)}\, d\theta + k_f k_з \Lambda_E \int_0^t e^{-\Lambda_E(t-\xi)} \times \tag{13}$$

$$\times \int_0^\xi \frac{P(\theta)}{\delta(\theta)\nu(\theta)} e^{-\Lambda_з(\xi-\theta)}\, d\xi\, d\theta.$$

With P = const (13) is transformed into (14):

$$C = aP - (be^{-\Lambda_з t} + de^{-\Lambda_F t})\, P, \tag{14}$$

where

$$a=\frac{k_f k_p}{h_p}+\frac{k_f k_з}{\delta\nu\Lambda_з},\quad d=\frac{k_f k_p}{h_p}+\frac{k_f k_з}{\delta\nu(\Lambda_з-\Lambda_E)}, \tag{15}$$

and the b parameter retains its former value (12).

For large deep water bodies it is possible to neglect the role of bottom deposits in the contamination of water. Then in (1) $C_E = 0$ and the first term on the left-hand side of (8) becomes equal to zero. The concentration of impurity in such a water body is

$$C=\frac{P(t)}{h_p(t)}k_p+\int_0^t P(\theta)\left[\frac{e^{-\Lambda_M(t-\theta)}}{H(\theta)}+k_з\frac{e^{-\Lambda_з(t-\theta)}}{\delta(\theta)\nu(\theta)}\right]d\theta. \tag{16}$$

With P = const (16) can be represented in the form (14) with replacement of Λ_E by Λ_M in (14) and the following values of the parameters a, b, d:

$$a=\frac{k_p}{h_p}+\frac{1}{H\Lambda_M}+\frac{k_з}{\delta\nu\Lambda_з},\quad b=\frac{k_з}{\delta\nu\Lambda_з},\quad d=\frac{1}{H\Lambda_M}. \tag{17}$$

Now we will examine contamination of the river system by an impurity. In this case it is possible to neglect the atmospheric fallout of impurity onto the surface of the river water ($C_M = 0$); however, additionally it is necessary to take into account the runoff of the impurity together with the river water and the dilution of the contaminated river water by pure ground water. If we limit ourselves to an examination of the mean annual water runoff and the water supply of the river and in the first approximation assume these values to be constant, the change in the content of the impurity in the river as a result of runoff, on the one hand, can be represented in the form of the product of water volume V and the change in the concentration of impurity in the river dC. On the other hand, with an intensity of water runoff Q along the river during the time dt there will be a runoff CQdt of the impurity. In a general case it is also necessary to take into account a decay of the impurity during this time of λCdt. Equating these expressions, we obtain an equation describing the change in the contamination of river water as a result of the dilution of contaminated water by pure ground water, compensating the runoff.

$$\frac{dC}{dt}=-\frac{Q}{V}C-\lambda C=-\Lambda_Q C. \tag{18}$$

Here Λ_Q is the dilution constant, characterizing the multiplicity of replacement of the waters in the river during a unit time (with decay taken into account).

Solving (13) and (18) jointly, we obtain an expression for the concentration of impurity in river water:

$$C=\frac{k_f k_p\Lambda_E}{\Lambda_E-\Lambda_Q}\frac{P(t)}{h_p(t)}+\frac{k_f k_з\Lambda_E}{\Lambda_E-\Lambda_Q}\int_0^t\frac{P(\theta)}{\delta(\theta)\nu(\theta)}\times \tag{19}$$

$$\times e^{-\Lambda_з(t-\theta)}\,d\theta+C_0,$$

where C_0 is the "background" concentration in the river from sources of contamination not taken into account, with water evaporation taken into account.

If P = const, with constant values h_p, δ and ν (19) can be represented in analytical form

$$C = aP - be^{-\Lambda_s t} P + C_0, \tag{20}$$

where

$$a = \left(\frac{k_p}{h_p} + \frac{k_s}{\delta\nu\Lambda_s} \right) \frac{k_f \Delta_E}{\Lambda_E - \Delta_Q}, \quad b = \frac{k_f k_s \Delta_E}{\delta\nu\Lambda_s (\Delta_E - \Lambda_Q)}. \tag{21}$$

Now we will compare the computed concentrations of impurity in the water with data from direct measurements. At the present time the greatest amount of material has been accumulated on the Sr^{90} concentration in the rivers of the Moscow area [1, 2] and in the rivers of the USSR as a whole [3]. Source [2] gives an empirical formula for the Sr^{90} concentration in the rivers of the Moscow region

$$C = a \frac{P}{h} + b \frac{A}{h} + c, \tag{22}$$

where h is the annual quantity of precipitation, A is the reserve of Sr^{90} in the soil.

The empirical coefficients a, b and c, refined in [1], are equal to 0.031, 0.012 year^{-1} and $3 \cdot 10^{-13}$ Ki/liter. By comparing formulas (22) and (19) it is possible to note a certain similarity between them. However, the h_p value is dependent on the state of the soil surface and therefore is unrelated linearly to h, and also to $\delta\nu$. In addition, (22) does not take into account the temporal decrease in the quantity of Sr^{90} subject to washing out, although this effect is observed in actual practice [5]. All this explains why in Fig. 1a the concentration curve C(t), computed using (22), after 1965 deviated from the experimental data. This figure shows the results [6] of the computations which we made using formula (19), which is conveniently written in the form

$$C(t) = aP(t) + b\sigma(t) + C_0. \tag{23}$$

The values of the parameters in (23) are as follows:

$$a = \frac{k_f k_p \Delta_E}{(\Delta_E - \Lambda_Q) h_p} = 8{,}1 \cdot 10^{-4} \text{ year/cm}, \qquad b = \frac{k_f k_s \Delta_E}{(\Delta_E - \Lambda_Q)\delta\nu} =$$

$$= 3{,}2 \cdot 10^{-4} \text{ cm}^{-1}, \Lambda_s = 0{,}14 \text{ year}^{-1}, \qquad C_0 = 2 \cdot 10^{-13} \text{ Ki/liter}. \tag{24}$$

Since h_p, δ and ν were assumed to be constant in the first approximation, the errors arising as a result of this are included in the C_0 value, which in a case if the value of the errors is significant, to some degree acquires the sense of a fitting parameter. In computations of C(t) the experimental values P(t) for the Moscow region in 1962-1967 were taken from [2]; data for the remaining years were found for the Moscow region on the basis of the intensity of the fallout onto the territory of the USSR as a whole [3, 7] (the fallout onto the territory of the Moscow region was 10% less).

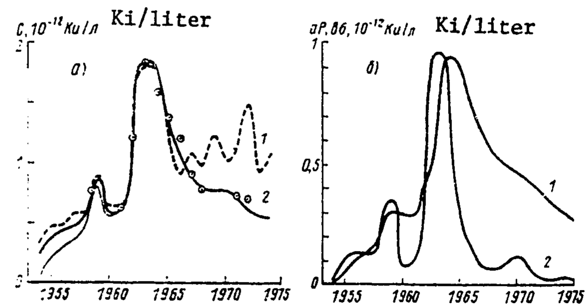

Fig. 1. Concentration of Sr^{90}, averaged for Moscow region. a) in rivers, computed using formula (22) (1) and using formula (23) (2); in surface waters contaminated by runoff of isotope from soil surface, flowing into rivers $b\sigma$ (1) and washing away of atmospheric fallout aP (2).

In Fig. 1a the dots represent data from observations of the concentration of Sr^{90} in the rivers of the Moscow region. It follows from Fig. 1a that the C(t) curve, computed using formula (23), coincides well with the data from direct measurements. This coincidence makes it possible to use formula (23) for computing the Sr^{90} concentration in the rivers of the Moscow region during the preceding years, beginning in 1954, when the radioactive fallout from the atmosphere increased sharply. Figure 1a shows that in 1959 a peak in the concentration of Sr^{90}, approximately corresponding to the 1967 level, should be observed in the rivers of the Moscow region. This peak was caused by the entry of Sr^{90} into the river after falling from the atmosphere in this same year, as can be seen clearly from Fig. 1b, which shows the temporal change in the terms aP and $b\sigma$ in formula (23). It also follows from this figure that the constribution of the washing away of atmospheric fallout aP by precipitation and the contribution of washing away of the isotope $b\sigma$, accumulating in the soil, to the contamination of river water, are quite close in value during the periods of increase in atmospheric fallout, but sharply diverge during periods of a decrease of fallout. In particular, after 1963 washing away from the soil exceeded the entry of Sr^{90} into the rivers with fallout by a factor of 1.5 to 17.

It should be noted that the assumption of a constancy of the C_0 value in (23) which proved to be valid for the C(t) curve segments for 1961-1972 (Fig. 1a) cannot be made for the preceding years, which makes the corresponding segment of the C(t) curve less reliable.

We will make an estimate of the possible changes in C(t), for which we will compute this value under the extreme assumption that during the period 1954-1961 the C_0 parameter increased linearly from zero. In Fig. 1a this part of the curve has been plotted as a thin line. It can be seen that the general character of the curve did not change under these circumstances.

Now we will examine the temporal change in the mean Sr^{90} concentration in the rivers of the USSR.

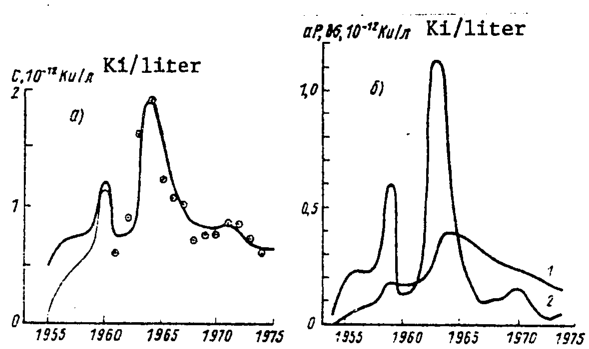

Fig. 2. Sr^{90} concentration averaged for the territory of the USSR. a) in rivers, computed using formula (25); b) in surface waters contaminated by runoff of the isotope from the soil surface $b\sigma$ flowing into rivers (1) and washing away of atmospheric fallout aP (2).

It was demonstrated in [3] that with averaging for the entire territory of the country the maximum of the Sr^{90} concentration in river water is observed in the year following the maximum of fallout of this isotope from the atmosphere. This is attributable to the effect of the lag of Sr^{90} migrating along the system of tributaries into large rivers and along the way in a state of sorption interaction with bottom deposits in the tributaries [8]. Taking the mentioned time shift into account, formula (23) for the rivers of the USSR is represented in the form

$$C(t) = aP(t-1) + b\sigma(t-1) + C_0, \tag{25}$$

where a = $1\cdot10^{-3}$ year/cm, b = $1.4\cdot10^{-4}$ cm^{-1}, Λ_σ = 0.16 $year^{-1}$, C_0 = $4.4\cdot10^{-13}$ Ki/liter. The corresponding C(t) curve is shown in Fig. 2a; the dots in this figure represent observational data taken from [3]; data on the fallout of Sr^{90} were taken from [3, 7]. As earlier, for 1954-1961 the C(t) curve was computed in two extreme variants: with C_0 = const and with C_0 increasing linearly from zero (thin line).

An examination of Fig. 2a makes it possible to conclude that despite the combining of data for extremely diverse rivers flowing through the territory of the USSR, the coincidence of the theoretical curve with the observational data can be deemed entirely satisfactory. The assumption made concerning an increase of the C_0 parameter during the period up to 1961 did not substantially change the nature of the C (t) curve; in particular, a clearly expressed maximum of the Sr^{90} concentration in the rivers of the USSR in 1960 was maintained. It follows from Fig. 2b that this maximum was caused by intensive fallout of the isotope from the atmosphere in 1959. After powerful injections of Sr^{90} into the planetary atmosphere there was also an increase in the intensity of its expulsion from the atmosphere. During these periods the contribution of washing out of fallout of the isotope to the contamination of river water predominated over the contribution from washing out from the soil. After 1965 the picture was the reverse and the contribution of the washing out of Sr^{90} from the soil up to six times exceeded the contribution from washing out of its fallout.

BIBLIOGRAPHY

1. Bobovnikova, Ts. I., Makhon'ko, K. P., "On the Problem of Migration of Sr^{90} in Fresh Waters of the Land," RADIOEKOLOGIYA VODNYKH ORGANIZMOV (Radioecology of Water Organisms), No 2, 1973.

2. Bobovnikova, Ts. I., Sereda, G. A., Shulepko, Z. S., "Relationship Between the Sr^{90} Content in Fallout, Soil and Rivers, According to Measurements During 1961-1967 in the Moscow Region," TRUDY IEM (Transactions of the Institute of Experimental Meteorology), No 5, 1970.

3. Bobovnikova, Ts. I., Avramenko, A. S., Makhon'ko, K. P., Dibtseva, A. V., Volokitin, A. A., Chumichev, V. B., "Sr^{90} Concentration in Surface Waters of the Land in the Territory of the USSR," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 9, 1977.

4. Makhon'ko, K. P., "Computation of Contamination of Surface Waters of the Central Region of the North Atlantic by Atmospheric Fallout of Sr^{90}," METEOROLOGIYA I GIDROLOGIYA, No 3, 1979.

5. Makhon'ko, K. P., Avramenko, A. S., Bobovnikova, Ts. I., Chumichev, V. B., "Runoff Coefficient of Sr^{90} and Cs^{137} from the Soil Surface of a River Basin," METEOROLOGIYA I GIDROLOGIYA, No 10, 1977.

6. Makhon'ko, K. P., Rabotnova, F. A., "Role of Radioactive Contamination of the Soil and Atmospheric Fallout in the Entry of Sr^{90} into River Water," MATERIALY VI VSESOYUZNOGO SIMPOZIUMA PO SOVREMENNYM PROBLEMAM SAMOOCHISHCHENIYA VODOYEMOV I REGULIROVANIYA KACHESTVA VODY. II SEKTSIYA. GIDROKHIMICHESKIYE I SANITARNO-BIOLOGICHESKIYE ASPEKTY SAMOOCHISHCHENIYA, CH. I (Materials of the VI All-Union Symposium on Modern Problems in the Self-Purification of Water Bodies and Regulation of Water Quality. Section II. Hydrochemical and Sanitary-Biological Aspects of Self-Purification, Part I), Tallin, 1979.

7. Makhon'ko, K. P., Avramenko, A. S., Silant'yev, A. N., Polyakova, T. V., Chumichev, V. B., Malakhov, S. G., "Soil Accumulation of Sr^{90}, Cs^{137}, Ce^{144} and Zirconium With $Niobium^{95}$ as an Average for the USSR," TRUDY IEM, No 6(64), 1977.

8. Makhon'ko, K. P., Bobovnikova, Ts. I., Avramenko, A. S., Dibtseva, A. V., Voloktina, A. A., "Vertical Distribution of Sr^{90} and Cs^{137} in the Bottom Deposits of Rivers and Lakes," EKOLOGIYA (Ecology), No 3, 1975.

9. Makhon'ko, K. P., Volokitin, A. A., Avramenko, A. S., Bobovnikova, Ts. I., "Migration of Radioisotopes in the Soils of the Coastal Sectors of Rivers," TRUDY IEM, No 6(64), 1977.

10. Makhon'ko, K. P., "Migration of Radioisotopes of Global Origin in Some Natural Media," RADIOEKOLOGICKA KONFERENCIA, 1 diel, Stary Smokovec, 1972.

11. Rovinskiy, F. Ya., Makhon'ko, K. P., "On the Problem of Migration of a Radioactive Impurity in the Bottom Materials of Water Bodies Without Through Flow," TRUDY IPG (Transactions of the Institute of Applied Geophysics), No 8, 1967.

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WEATHER AND OPTIMUM TIMES FOR THE SOWING OF WINTER CROPS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 90-97

[Article by Professor A. P. Fedoseyev, All-Union Scientific Research Institute of Agricultural Meteorology, submitted for publication 3 Apr 80]

[Text]

Abstract: A study was made of the dynamics of the decrease in the yield of winter wheat and rye, the change in structure of the yield and the effectiveness of mineral fertilizers in connection with deviation of the sowing times from the optimum dates. A tendency to a shifting of sowing times to later dates with an increase in the sophistication of agricultural techniques is demonstrated.

Sowing times exert a great influence on the quality and magnitude of the yield. Sowing dates are related to the conditions of growth and development of plants, resistance to unfavorable meteorological phenomena, harvesting conditions and effectiveness of agroengineering measures. By choosing different sowing times there is a definite possibility for exposure of plants to conditions of a different length of daytime, temperature and moisture content of the air and soil.

The dependence of the yield of grain of winter crops on the sowing times has the character of a single-peaked curve. Figure 1 shows a series of such curves for different natural-economic regions of the European USSR constructed by the author in collaboration with A. I. Snetkova on the basis of a generalization of long-term field data for experimental agricultural stations and scientific institutions (a total of 2,640 experimental years was analyzed). Most of the experiments were carried out during the period 1967-1976 and were carried out using new regionalized varieties of winter wheat and rye. The maximum yield in the case of an optimum sowing time in each case was assigned the value 100%.

The relative curves of the yield of winter crops by sowing times characteristic for the regions of the Nonchernozem zone and the Middle Volga have a sharper peak. Under these conditions the mean long-term range for the optimum period of sowing of winter crops (with a small underharvest of grain in the range 5% of the maximum) is 12-14 days (Fig. 1, curves 1-4). Beyond the limits of this mean interval there is a marked dropoff of the crop yield.

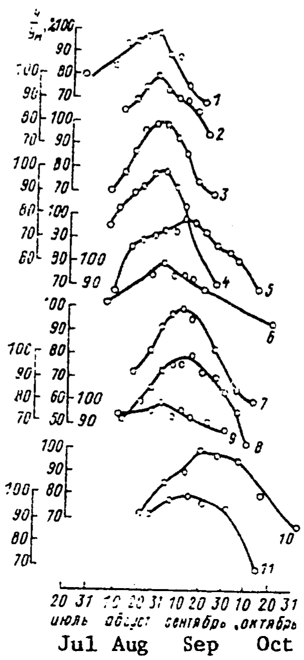

Fig. 1. Yield of winter wheat and winter rye (in the northern regions) in dependence on sowing times for the European USSR. Regions: 1) Northeastern Nonchernozem; 2) Northwestern Nonchernozem; 3) Middle Volga; 4) Central Nonchernozem; 5) Wooded Steppe Zone; 6) Lower Volga; 7) Steppe Zone; 8) Belorussia; 9) Baltic republics; 10) Northern Caucasus; 11) Dry Steppe Zone

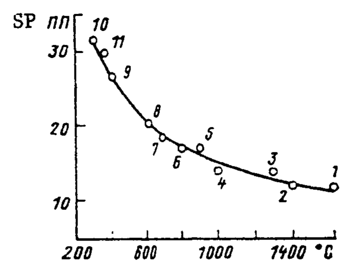

Fig. 2. Correlation between mean duration of period of sowing of winter crops (SP, in days) and sum of negative temperatures during winter (°C). Notations of the regions (figures next to the circles) are the same as in Fig. 1.

The next group of more flattened curves reflects the conditions prevailing in the wooded steppe and steppe zones, Belorussia and the Lower Volga (Fig. 1, curves 5-8). Here the mean range of the sowing period increases to 20 days.

The curves for losses of the grain yield with deviation from the optimum sowing times in the southern regions with warm winters and in the Baltic republics have a still smoother shape (Fig. 1, curves 8-11). The mean range of the period of sowing of winter crops with losses not greater than 5% of the yield attains 27-32 days.

In regions with a prolonged gentle autumn period (Baltic republics, Belorussia, Northwestern Nonchernozem, Northern Caucasus) a deviation in the optimum sowing times by 20 days both in the direction of earlier and later times exerts an effect in an underharvest of grain yields to an approximately equal degree -- in the average range 15-20%. In the remaining regions of the European USSR with a more continental climate the greatest underharvest occurs with late sowing times (on the average by 22-35% with a lag of 20 days) in comparison with early sowing times (15-25%). Accordingly, the greatest loss is from a lag in the sowing of winter crops, which most frequently is actually observed.

It is a legitimate assumption that the duration of the possible sowing period for different regions is related to the severity of winter [11]. In regions with gentle warm winters there can be a periodic winter growing season for plants favoring an intensification of shoot formation in incompletely developed crops. Also of importance for the additional bushing-out of winter wheat is the duration of the early spring period. For example, in the Baltic republics the transition of air temperature from 0 to 10°C occurs on the average in 42 days; in the central regions of the Nonchernozem zone this period is 34 days. Milder conditions of winter and spring to some degree favor a smoothing of the differences in the development of plants sown at different times.

The dependence of the mean duration of the sowing period of winter crops on the sum of negative temperatures in winter is clearly confirmed by Fig. 2. Such a correlation is also traced with other indices of winter severity.

It should be noted that the mean long-term duration of the period of optimum sowing cannot be used for specific years.

Plants sown in the optimum sowing periods are more winter-resistant and productive. With an increase in the deviations of the sowing times from the optimum times the density of the productive stem stand is reduced by the time of the harvest. This is particularly conspicuous at later times, when with a lag of 30 days the density of the stem stand is 40% less than the mean density for optimum sowing times (Table 1).

One of the reasons for the thinning-out of the stem stand is an appreciable decrease in productive bushiness in the case of late sowing times, whereas the bushiness for early-sowed plants remains virtually unchanged. Plantings in late and to a greater degree in early sowing seasons are less resistant to unfavorable wintering conditions. Their survival during the winter on the average is 7-10% less for late and 15-22% less for early sowing times.

The absolute mass of 1,000 grains, according to averaged data, also decreases and this is particularly conspicuous for late times (on the average by 14%). The number of grains in an ear on the average is somewhat greater for plants sown at early times and is somewhat less for plants sown at late times, although under the specific conditions of individual years the infructescence of the grains in the flowers can be more intensive in plants sown at late sowing times.

Sowings at early times are considerably more subject to damage by frit and Hessian flies, are populated by greenbugs, are damaged by blight, brown rust and root rot. In the case of late sowing times there is an increase in the damage inflicted by soil click beetles, surface caterpillars and also inflicted by covered smut. Plants with late sowing times are damaged more intensively by rust during earing.

In addition, with late sowing times the forming of the grain in the southern regions falls in the period of highest temperatures, and in the northern regions, on the other hand, transpires more slowly when there is cold, frequently rainy weather.

The change in the effectiveness of mineral fertilizers in dependence on the times for the sowing of winter crops is of special interest. The displacement of response of winter crops to fertilizers in the direction of later sowing times observed in some experiments is evidently attributable to a slowing of the processes of mineralization cf organic substances in the soil with a decrease in air temperature. However, other factors also become operative which give the reverse effect: a slowing of the entry of nutrients into the plants with a definite minimum of temperatures and a shortening of the period of the active growing season for late-sown crops.

It is known that with a temperature of 8-10°C there is an appreciable decrease in the entry of nitrogen into the roots and movement from the roots into the plant organs above the ground and its use in the formation of organic nitrogen compounds is weakened. At still lower temperatures (5-6°C and below) the root absorption of nitrogen and phosphorus is sharply reduced.

On the average, in the Nonchernozem zone the air temperature drops from 14°C in the 10-day period of optimum sowing times to 12°C by the next 10-day period. With such a sowing time, as a result of weakening of the processes of mineralization of matter, it is possible to expect some increase in the response of winter crops to fertilizer. Sowings in the next (second after the optimum time) 10-day period occur under conditions of a reduced temperature of 10°C, already causing a slowing of the movement of nutrients. In the third and fourth 10-day periods there is a further temperature decrease to 8-6°C. Under these conditions there is a marked impairment in the entry of mineral substances into the roots and the growing season for plants is reduced by 3-4 10-day periods. With these sowing times it is necessary to expect a considerable decrease in the effectiveness of fertilizers.

The ideas expressed here are confirmed well by the actual data (Table 2). On the average, some increase in the response of winter plants to fertilizers is observed only with a small deviation of the sowing times from the optimum. With a lag of sowing by 20 days the effectiveness of the fertilizers is already reduced by 20%. However, some shifting of the maximum effectiveness of fertilizers to sowing times somewhat later than the optimum times does not at all mean a real advantage. This small effect is offset by a far more considerable underharvest associated with deviation from the optimum sowing times (see Table 2). For example, against a fertilized background with a lag in sowing by 10-20 days there is a loss of 10-35% of the crop yield. It is characteristic that on crop areas without fertilizers these losses are somewhat greater -- 15-40%.

Thus, rigorous adherence to the optimum sowing times for winter crops, dependent for the most part on weather conditions, is in essence a considerable reserve for increasing the effectiveness of mineral fertilizers. For this reason it is of great practical importance to develop reliable methods for agrometeorological prediction of sowing times.

The mean long-term times for the sowing of winter crops are rather conservative. One of the founders of Soviet agroclimatology, A. I. Voyeykov, as early as 1884, on the basis of a generalization of the many years of experience of agriculturalists, established the mean sowing times for the territories of Russian provinces [1].

N. N. Yakovlev compared these data with the long-term dates of sowing computed using climatic data on the duration of the period of autumn development of winter crops of 52 days, reckoning from the date of transition of air temperature through 5°C in autumn [17]. The discrepancies between the mean sowing times according to A. I. Voyeykov and the computed dates in most cases were insignificant and with few exceptions fell in the range of four days.

A comparison of the mean sowing times for winter crops by provinces, established in the 1870's-1880's, and the corresponding mean oblast sowing times for a 15-year period taken from agroclimatic reference books (1940-1955), indicates their consistent shift by 2-5 days later and in individual cases a few more days. In turn, a comparison of the mean sowing times for winter crops for the period 1940-1955 (according to data from agroclimatic reference books) with the mean statistical values (50% of the sown area) for 1964-1976 for the most part confirms their shift by 2-9 days later.

The shift of the sowing times for winter crops to a later time must be attributed to the introduction of new varieties of winter wheat of the intensive type and a general increase in soil fertility as a result of the application of fertilizers during recent years.

It is known [9] that intensive varieties of winter wheat form higher yields in the case of sowing at later than the usual times. It is also known that the greater the soil fertility, the more intensive is the growth of plants and the shorter is the time period which is required for the completion of autumn development. For example, in the central regions of the Nonchernozem zone the optimum period for the sowing of intensive varieties of winter wheat without the use of fertilizers is 20-30 August, whereas with the application of fertilizers it is displaced to the third ten-day period in August - first ten-day period in September [5]. With a high background of fertilization and with the formerly recommended earlier sowing times there is a nonproductive large autumn loss of nutrients, an intensified damping of plants and a decrease in their productivity.

With the variable meteorological conditions of specific years the optimum sowing times for winter crops can deviate considerably from those mean times established for each zone. For example, according to experimental data the sowing times for which the maximum grain yield is formed in the northern half of the European USSR in 70-80% of the years fit in the range of four five-day periods. In the southern regions the optimum sowing times are lengthier and in most (80%) cases fit in an interval of five or six five-day periods. In the remaining years (20-30% of the cases) the sowing times can deviate by two or three additional five-day periods.

In agricultural practice recommendations are given for adherence to the mean optimum sowing times established on the basis of experimental data for each soil-climatic zone or its individual parts. The duration of the mean sowing time period is usually determined for winter wheat at 10-15 days and for winter rye at 10-20 days.

Table 1

Change in Yield Structure and Other Winter Wheat Indices in Dependence on Sowing Times (Mean of 200 Experiment-Years)

	Index	Time							
		before optimum days			optimum	after optimum, days			
		30	20	10		10	20	30	40
1	Число продуктивных стеблей, $м^2$	360	365	385	420	360	315	255	250
2	Число зерен	29	26	29	26	24	25	25	—
3	Масса 1000 зерен, г	33,3	36,8	37,5	37,3	36,0	35,7	32,5	32,0
4	Продуктивная кустистость	1,9	2,0	1,9	2,0	1,7	1,6	1,6	1,4
5	Количество сохранившихся растений за зиму, %	62	70	80	84	82	78	75	77
6	Повреждение скрытостебельными мухами, %	42	22	12	5	1	3	0	—
7	Поражение ржавчиной, %	42	34	27	23	24	35	41	—
8	Поражение полосатой мозаикой, %	62	45	17	4	0	0	0	—
9	Содержание белка в зерне, %	—	12.8	12.9	12,9	13,1	13,2	13,8	—

KEY:

1. Number of productive stems, m^2
2. Number of grains
3. Mass of 1,000 grains, g
4. Productive bushiness
5. Number of plants surviving during winter, %
6. Damage by stem flies, %
7. Damage by rust, %
8. Damage by streak mosaic, %
9. Protein content in grain, %

A comparison of the actual annual optimum sowing times established on the basis of large amounts of experimental data with the mean recommended sowing times indicated that their coincidence was observed in 60% of the years; in 20% of the years they differed by 10-20 days. Such deviations lead to considerable underharvests of grain, in individual cases 5-10 centners/hectare or more.

It is entirely obvious that the standard times for the sowing of winter crops recommended by agricultural agencies are in need of annual correction from the point of view of both agroengineering and weather conditions.

Attempts of researchers to give an agrometeorological validation of the sowing times for the most part essentially involved computation (prediction) of the optimum times on the basis of the sum of temperatures necessary for the formation of the optimum

Table 2

Yield of Winter Crops (Centners/Hectare) and Effectiveness of Full Mineral Fertilizers (NPK 120–180 kg/Hectare of Active Substance) for Different Sowing Times (Mean of 116 Experiment-Years)

Variant	Time								
	before optimum				optimum	after optimum, days			
	20	15	10	5		5	10	15	20
Yield without fertilizers	22.4	24.0	24.2	24.9	27.7	24.6	23.8	23.6	16.9
Yield increment due to fertilizers	3.9	4.4	5.8	6.0	6.5	7.2	6.5	5.8	5.2
Yield on fertilized background	26.3	28.4	30.0	30.9	34.2	31.8	30.3	29.4	22.1

number of sprouts of winter crops by the end of the growing season (usually 2-4-6 sprouts, depending on the capacity of the plants for tillering and soil moisture content). The plants of winter wheat and rye attain such a bushiness when a sum of effective (above 5°C) temperatures of 200-300°C is accumulated during the period from sowing to the cessation of the growing season in the autumn [16]. Considerable corrections to determination of the optimum sowing times are introduced by the level of the moisture reserves in the soil, which is taken into account in the computations by different methods [2, 6, 7, 11, 15 and others]. The principle of allowance for definite heat sums necessary for the development of plants also served as a basis for constructing agroclimatic maps of the mean times for the sowing of winter crops [3, 6, 8, 12, 16, 17 and others].

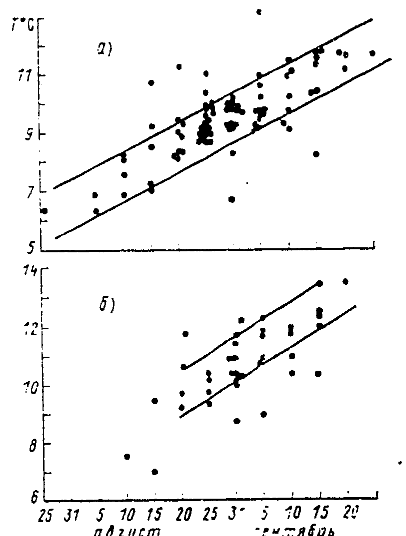

Fig. 3. Optimum times for the sowing of winter crops in dependence on mean air temperature during period from 25 August through 20 October. Nonchernozem zone. a) under conditions of adequate moistening; b) under conditions of inadequate moistening (precipitation during August-September less than 80 mm).

Table 3

Underharvest of Grain Yield (%) for Different Winter Wheat Varieties with Deviation of Sowing Times from Optimum (72 Experiment-Years)

Variety	Time			
	before optimum, days		after optimum, days	
	20	10	20	10
Mironovskaya-808	23	8	7	13
Odesskaya-51	14	6	8	18
Kavkaz	20	10	8	22
Khar'kovskaya-63	18	12	12	--
Bezostaya-1	32	14	12	24
Il'ichevka	17	14	20	--

An analysis of the mass experiments led to the conclusion that the optimality of sowing times is determined not only by the bushiness index at the end of autumn. The advantage of different sowing times in each specific case is dependent on a whole series of factors: agricultural engineering techniques, variety, type of predecessor, multiplication of pests and diseases, wintering conditions, conditions of the spring and summer growing season, including even the resistance of sowings at different times to beating down.

The processing of experimental data shows that the times of sowing of winter crops in the Nonchernozem zone, determined using the criterion of 200–300°C sum of effective temperatures (in the computations this limit was broadened by ±25°C) in 60% of the cases actually corresponded to the maximum yields.

Unfortunately, inadequate methodological possibilities do not make it possible in the determination of the optimum sowing times to introduce corrections for the conditions of the future winter and spring-summer periods. Therefore, for the time being agrometeorologists, as before, have at their disposal only a possibility for taking into account the meteorological conditions developing prior to sowing or for evaluating the subsequent conditions of autumn with the use of synoptic forecasts.

Generally available recommendations on the use of agrometeorological data in the choice of the optimum times for the sowing of winter wheat for the territory of the Ukraine were proposed by I. G. Grushko and V. P. Dmitrenko [3].

For the conditions of the Nonchernozem zone (Belorussia) L. K. Pyatovskaya [10] proposed a method for computing the optimum times for the sowing of winter crops on the basis of mean air temperature for the period from 25 August through 20 October.

A checking of the L. K. Pyatovskaya index on the basis of reliable experimental data for the territory of the Nonchernozem zone indicated a possibility of its use for the choice of optimum sowing times. Under the conditions of an adequate moisture supply the optimality of sowing times was dependent for the most part on the temperature level for the period (Fig. 3a). With a shortage of moisture (precipitation quantity during August-September less than 80 mm) the scatter of points increases and the sowing times as a rule were displaced to earlier times (Fig. 3b).

Different varieties of winter wheat in one and the same climatic region have different optimum sowing times. This is attributable to their photoperiodic reaction. Varieties with a well-expressed photoperiodic reaction must be sown at earlier times. For example, under conditions of a gradually shortening day frost-resistant varieties react more strongly than those with weak resistance to a short day with a lag in their development and should be sown earlier. Wintering Mexican short-stem varieties with a weak photoperiodic reaction can be sown latest [14].

It is known that intensive varieties of winter wheat (Avrora, Kavkaz, Bezostaya-2, Rostovchanka and others) form higher yields when sown 10–12 days later than the usual times [9]. The optimum period of sowing is shortest for them.

The capacity for winter wheat of the Mironovskaya-808 type on a good agricultural background to bush out in autumn and spring broadens the intervals of optimum sowing times. The variety Odesskaya-51 also is characterized by a high plasticity. With a deviation of the sowing times from the optimum its productivity and resistance to winter conditions decrease to a lesser degree than for the varieties Kavkaz, Khar'kovskaya-63, Il'ichevka, Bezostaya-1 (Table 3). Odesskaya-16 and Novomichurinka are plastic varieties. The varieties Odesskaya-26 and Stepovaya occupy an intermediate position.

The optimum times for the sowing of winter crops are dependent on the agroengineering background. For fertilized clean fallow the optimum sowing times are 10-12 days later than for predecessors not on fallow. On the other hand, with an inadequacy of precipitation it is desirable that the sowing begin with clean fallow.

The sowing of winter crops in dry or semidry soil cannot be allowed. According to data from the Zherebkovskaya Experimental Station, with a content of not more than 3 mm of moisture available to plants in the 10-cm soil layer the wheat seeds may not lose their germinating power for a relatively long time. With moisture reserves of 5-6 mm the seeds swell, become moldy and lose their germinating power [13]. When the reserves of productive moisture are less than 15 mm in the cultivated soil layer (0-20 cm) winter crops should not be sown prior to the falling of considerable precipitation because the plantings will be sparse [6]. Accordingly, with a moisture shortage in the soil it is desirable that the sowing be postponed to a later time or that the fields be left in spring crops.

It is possible to await precipitation in dry autumn only within the limits of the admissible times during which the sown winter crops could attain the tillering phase by the end of the autumn growing season.

P. G. Kabanov, for the conditions prevailing in the Volga region, examines the possibility of the sowing of winter rye at early times in a case of falling of rains in late July-early August and dry weather thereafter [4].

As a result, it must be admitted that at the present time the most promising agrometeorological method for computing the optimum times of sowing of winter crops remains the method of computation on the basis of the index of bushiness of winter crops by the end of the growing season. The degree of bushiness characterizes not only the morphological state of the plants but also the size of the apical cones for the autumn sprouts. In its modern modification applicable to the state of the sprouts over great areas this method is being successfully developed at the USSR Hydrometeorological Center [6, 7, 15].

BIBLIOGRAPHY

1. Voyeykov, A. I., "On the Times of Sowing and Harvesting of Field Plants and the Mowing of Hay Fields in European Russia," SEL'SKOYE KHOZYAYSTVO I LESOVODSTVO (Agriculture and Forestry), January, 1884.

2. Grudeva, A. Ya., METODICHESKOYE POSOBIYE PO PROGNOZIROVANIYU SOSTOYANIYU SOSTOYANIYA OZIMOY PSHENITSY I OZIMOY RZHI KO VREMENI PREKRASHCHENIYA OSENNEY VEGETATSII V CHERNOZEMNOY ZONE YeTS (Methodological Aid for Predicting the State of Winter Wheat and Winter Rye by the Time of Cessation of the Autumn Growing Season in the Chernozem Zone of the European USSR), Moscow, Gidrometeoizdat, 1974.

3. Grushka, I. G., Dmitrenko, V. P., "Computation of the Anticipated Times of Sowing of Winter Wheat and Evaluation of its Effectiveness," TRUDY UkrNIGMI, (Transactions of the Ukrainian Scientific Research Hydrometeorological Institute), No 84, 1969.

4. Kabanov, P. G., DIFFERENTSIROVANNOYE PRIMENENIYE AGROTEKHNIKI V POVOLZH'YE (Differential Use of Agricultural Techniques in the Volga Region), Saratov, Privolzhskoye Knizhnoye Izd-vo, 1968.

5. Kozlov, I. V., "Fertilizer, Sowing Times and Yield of Winter Wheat," KHIMIYA V SEL'SKOM KHOZYAYSTVE (Chemistry in Agriculture), No 2, 1978.

6. Maksimenkova, T. A., "A Method for Predicting the State of Winter Grain Crops by the Time of Cessation of Their Growing Season," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 2, 1979.

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

8. Nosatovskiy, A. I., "Theoretical Validation of the Optimum Time of Sowing of Winter Wheat," DOKLADY VASKhNIL (Reports of the All-Union Academy of Agricultural Sciences), No 11-12, 1946.

9. Prutskov, F. M., POVYSHENIYE UROZHAYNOSTI ZERNOVYKH KUL'TUR (Increasing the Yield of Grain Crops), Moscow, Rossel'khozizdat, 1977.

10. Pyatovskaya, L. K., AGROMETEOROLOGICHESKOYE OBOSNOVANIYE SROKOV SEVA (Agrometeorological Validation of Sowing Times), Minsk, Uradzhay, 1977.

11. Svisyuk, I. V., "Sowing of Winter Crops at Optimum Times -- One of the Conditions for Their Favorable Wintering," AGROMETEOROLOGICHESKIYE ASPEKTY PEREZIMOVKI RASTENIY (Agrometeorological Aspects of Wintering of Plants), Leningrad, Gidrometeoizdat, 1977.

12. Stepanov, V. N., "Optimum and Limiting Times of Sowing of Winter Crops in the USSR," DOKLADY NAUCHNOY KONFERENTSII TSKhA (Reports of the Scientific Conference of the Timiryazev Agricultural Academy), No 5, 1947.

13. Tretyuk, A. I., "Times for the Sowing of Winter Wheat," SELEKTSIYA I TEKHNOLOGIYA VOZDELYVANIYA KUKURUZY I NEKOTORYYE VOPROSY AGROTEKHNIKI DRUGIKH KUL'TUR V SEVERO-ZAPADNOY CHASTI STEPI UkrSSR (Selection and Technology for Corn Cultivation and Some Problems in Agricultural Techniques for Other Crops in the Northwestern Part of the Steppe UkrSSR), Dnepropetrovsk, 1976.

14. Turbin, N. V., Fedorov, A. K., "Biological Validation of the Optimum Time for the Sowing of Winter Wheat," DOKLADY VASKhNIL, No 3, 1976.

15. Ulanova, Ye. S., AGROMETEOROLOGICHESKIYE USLOVIYA I UROZHAYNOST' OZIMOY PSHENITSY (Agrometeorological Conditions and Yield of Winter Wheat), Leningrad, Gidrometeoizdat, 1975.

16. Shigolev, A. A., "Method for Preparing Phenological Forecasts," SBORNIK METODICHESKIKH UKAZANIY PO ANALIZU I OTSENKE SLOZHIVSHIKHSYA I OZHIDAYEMYKH AGROMETUSLOVIY (Collection of Methodological Instructions on Analysis and Evaluation of Prevailing and Anticipated Agrometeorological Conditions), Leningrad, Gidrometeoizdat, 1957.

17. Yakovlev, N. N., KLIMAT I ZIMOSTOYKOST' OZIMOY PSHENITSY (Climate and Winter Resistance of Winter Wheat), Leningrad, Gidrometeoizdat, 1966.

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SOME POSSIBILITIES FOR SIMPLIFICATION OF ADAPTIVE ALGORITHMS IN PROGNOSTIC SCHEMES

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[Article by V. V. Plotnikov, Far Eastern Regional Scientific Research Institute, submitted for publication 11 Dec 79]

[Text]

Abstract: In this article, on the basis of an analysis of the variability of elements of a covariation matrix, the conclusion is drawn that so-called homogeneity intervals exist. It is shown that with a change in the volume of the initial samples the static characteristics of the process virtually do not change in this interval. The application of this fact makes possible a substantial decrease in the volume of computations in prognostic problems making use of adaptive algorithms. In addition, the author examines some assumptions relating to the optimum advance period of the prediction and the necessary volume of the sample for statistical experiments.

Adaptive forecasting methods have recently come into broad use in prognostic practice [3-5, 7]. Interest in these methods is attributable primarily to the fact that using them it is possible to solve problems of forecasting under nonstationary conditions. This effect is achieved by constant rescaling of the prognostic operators in each prediction interval. It is necessary to pay for such universality by a considerable complication of the algorithms, and as a result, an increase in the time expended on the forecast.

When using adaptive algorithms in physicostatistical schemes, considering the coefficients of expansion of the initial fields in natural orthogonal components (NOC) as predictors and predictants [1, 7], a considerable part of the time is expended on the scaling of natural components and their coefficients in each individual forecast.

The purpose of this study was an analysis of one of the possible ways to simplify adaptive algorithms for physicostatistical prediction of hydrometeorological objects.

The proposed approach makes it possible to economize the time expended on constant scaling of the natural components as a result of multiple use of the earlier computed values. This is possible in connection with the presence of sectors in the general nonstationary process corresponding to the requirements of statistical homogeneity, with subsequent transfer of the properties of the preceding set to these sectors. As is well known, the determination of the field EOC is a process of finding the eigenvalues and eigenvectors of the covariation (correlation) matrix of this field. In actual practice we almost always deal with sample values of the field elements and accordingly, we can obtain only an evaluation of the covariation matrix; this evaluation will be essentially dependent on the length of the sample.

A sample of the field values $\left\{F\right\}_{i,}^{n}{}_{j=1}^{m}$ (we will call it a sample matrix) can be represented as follows:

$$F = \begin{Vmatrix} F_{11} F_{12} \dots F_{1n} \\ F_{21} F_{22} \dots F_{2n} \\ \dots\dots\dots \\ \dots\dots\dots \\ \dots\dots\dots \\ F_{m1} F_{m2} \dots F_{mn} \end{Vmatrix}. \quad (1)$$

Here m is the dimensionality of the vector F (the number of its components), n is the volume of the sample (the number of observed values of the F vector included in the computations).

The evaluation of the covariation matrix $\left\{M\right\}_{i,}^{m}{}_{j=1}$ is obtained in the form of the averaged product of the matrix (1) and its transposed value.

$$M = \frac{1}{n} F \cdot F^T, \quad (2)$$

where T is the transposition symbol.

It is assumed here that the mean value of each component is already excluded.

Assuming the covariation matrix (2) to be quite well stipulated, we will determine whether the coefficients of expansion of the field in EOC will change with a change in the volume of the sample and if they do, how. In addition, we will attempt to find the maximum possible interval (τ) -- the increment of the sample volume with which these changes are statistically still not significant.

It is evident that the variability of the expansion coefficients is determined, in particular, by the variability of the elements of the covariation matrix. It is necessary to evaluate m(m+1)/2 parameters, of which m are diagonal and m(m-1)/2 are independent nondiagonal values.

Now we will examine one of the elements of the covariation matrix as a function of sample volume.

Without losing universality in our reasoning, in order to simplify the calculations and for greater clarity we will assume that this will be one of the diagonal elements of the matrix

$$M(n)_{jj} = \frac{\sum_{i=1}^{n} (F_{ij} - \overline{F}(n)_j)^2}{n}, \quad (3)$$

where $\overline{F}(n)_j$ is the mean value of the characteristic with a sample volume n.

$$\overline{F}(n)_j = \frac{\sum_{i=1}^{n} F_{ij}}{n}. \quad (4)$$

With a change in the volume of the sample the expressions (3), (4) are written as follows:

$$M(n+\tau)_{jj} = \frac{\sum_{i=1}^{n+\tau} (F_{ij} - \overline{F}(n+\tau)_j)^2}{n+\tau}, \quad (5)$$

$$\overline{F}(n+\tau)_j = \frac{\sum_{i=1}^{n+\tau} F_{ij}}{n+\tau}, \quad (6)$$

where τ is the number of observations added to the sample.

Now we will examine an extremal variant when any subsequent observation of the F vector, added to a sample of n terms, deviates from the mean value by a value which is limiting for the particular process; we will denote it by KA_j. Here A_j is the maximum possible amplitude of the process in the time interval $n + \tau$ and K is a coefficient changing in dependence on the specific field from 0 to 1 and characterizing the degree of nonstationarity of the particular field. Then equations (6), (5) are represented in the following form:

$$\overline{F}(n+\tau)_j = \overline{F}(n)_j + \frac{\tau KA_j}{n+\tau}. \quad (7)$$

$$M(n+\tau)_{jj} = \frac{\sum_{i=1}^{n} \left(F_{ij} - \overline{F}(n)_j - \frac{\tau KA_j}{n+\tau}\right)^2 + \sum_{i=1}^{\tau} (KA_j)^2}{n+\tau} \quad (8)$$

After obvious transformations we obtain

$$M(n+\tau)_{jj} = \frac{M(n)_{jj} \cdot n}{n+\tau} + \frac{n\tau^2 (KA_j)^2}{(n+\tau)^3} + \frac{\tau (KA_j)^2}{n+\tau}. \quad (9)$$

The transformation M_{jj} with a change in the volume of the sample in a general case is related to the influence of nonstationarity of the analyzed processes.

In order to find the shift τ with which the level of variability of the broadened sample in the interval $n + \tau$ with a stipulated confidence coefficient P does not exceed (is commensurable with) the level of variability of the set of the volume n, we will consider the following hypothesis.

We will assume that

$$\frac{M(n+\tau)_{jj}}{M(n)_{jj}} = 1. \tag{10}$$

In this case with an increase (decrease) in the volume of the sample by τ the elements of the covariation matrix do not change statistically. Then the random value

$$\frac{M(n+\tau)_{jj}}{M(n)_{jj}}$$

has a F_{v_1, v_2} Fisher distribution with $v_1 = n + \tau - 1$ and $v_2 = n - 1$ degrees of freedom [4].

Hence with a stipulated confidence coefficient P it is possible to determine that increment in the volume of the sample (so-called homogeneity interval or possible number of prediction intervals without scaling of the EOC) at which the change in the covariation coefficients is statistically not significant.

For this we examine the inequality

$$\frac{M(n+\tau)_{jj}}{M(n)_{jj}} < F_{v_1, v_2, \alpha}, \tag{11}$$

where the significance level is

$$\alpha = 1 - P. \tag{12}$$

With (9) taken into account we obtain

$$\frac{n}{n+\tau} + \frac{n\tau^2 K^2 A_j^2}{(n+\tau)^3 M(n)_{jj}} + \frac{\tau K^2 A_j^2}{(n+\tau) M(n)_{jj}} \leqslant F_{v_1, v_2, \alpha}. \tag{13}$$

In turn it is possible to determine the upper boundary of amplitude of the process (A_j) by having the values of the diagonal elements of the covariation matrix [6]

$$A_j = C\sqrt{M(n+\tau)_{jj}}, \tag{14}$$

where C, using the results presented in [6], can be approximated by the following expression:

$$C = \ln(n+\tau-1) - 0{,}069\sqrt{n+\tau} + 1{,}11. \tag{15}$$

Accordingly, we finally write:

$$\frac{M(n+\tau)_{jj}}{M(n)_{jj}} = \frac{n}{n+\tau} + \frac{M(n+\tau)_{jj}}{M(n)_{jj}}\left[\frac{n\tau^2 K^2 C^2}{(n+\tau)^3} + \frac{\tau K^2 C^2}{n+\tau}\right]. \tag{16}$$

Now the inequality (11) is transformed to the form

$$\frac{n}{n+\tau - \frac{n\tau^2 K^2 C^2}{(n+\tau)^2} - \tau K^2 C^2} < F_{v_1, v_2, \alpha}. \tag{17}$$

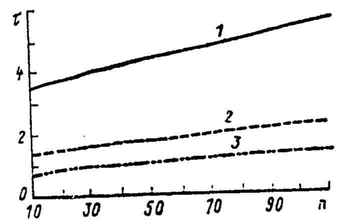

Fig. 1. Dependence of the upper boundary of the region of admissible values τ on length of series n with confidence coefficient 95% for K = 0.5 (1), K = 0.75 (2) and K = 1.0 (3).

It goes without saying that all the proposed computations are correct under the condition $n > \tau$. Otherwise, with $\to n$ there is a redistribution of the contributions of the dispersion of the main (n) and additional (τ) samples, leading to a discontinuity of the function

$$\frac{n}{n+\tau-\frac{n\tau^2K^2C^2}{(n+\tau)^2}-\tau K^2C^2}.$$

Subsequent increases in the interval lead to a dominant influence of the additional sample and the relative fraction of dispersion of the principal sample decreases. But from the point of view of the formulated problems this case is not of interest and is not considered in the study.

With $n \gg \tau$, neglecting the second term in the denominator on the left-hand side of inequality (17), by virtue of its smallness we obtain

$$\frac{n}{n+\tau-\tau K^2(\ln(n+\tau-1)-0{,}069\sqrt{n+\tau}+1{,}11)^2}-F_{v_1,v_2,\alpha}<0. \quad (18)$$

Expression (18) contains τ in implicit form and for its determination as a function of the volume of the initial sample n in the figure we have shown the curves for K = 0.5, 0.75, 1.0 and P = 95%, limiting the region of solution of inequality (18) upward. With different K and P values it is possible to construct a whole family of such curves.

For a stipulated volume of the sample n and the postulated K value, using the constructed curves, we obtain the maximum possible τ value at which there is still satisfaction of the condition of statistical homogeneity of the sample and a conclusion will be drawn concerning the possible reduction of the computations.

The results of the analysis indicate that even in extremal cases the proposed approach makes it possible to reduce the number of computations and save time expended on the prediction by shifting the properties of the initial set to the determined interval, with K = 1 equal to one prediction interval (discreteness of the used data). In the case of weaker nonstationarity (K< 1) the time gain increases significantly. For example, with K = 0.5 (see Fig. 1) the scaling of the eigenvalues and eigenvectors can be accomplished once in 4-5 predictions.

In addition to the enumerated results, the analysis of the discriminated intervals in each specific case makes it possible to evaluate the optimum advance time for which the effectiveness of the regression expressions should be a priori high.

This was associated with the statistical sense of the regression. Applying the regression equations for the prediction, we will assume statistical stationarity, and accordingly, the possibility of extrapolation of the statistical characteristics in time. In this understanding the periods of optimum advance time will be dependent to a considerable degree on the volume n of the sample and the characteristics K.

With respect to the K evaluation, this problem is solved as a result of a preliminary analysis of some a priori information on the statistical structure of the predicted processes. In case of necessity this parameter can be constantly corrected in the calculation procedure.

When determining the volume of work the following must be taken into account: on the one hand, in a prediction using adaptive models under nonstationarity conditions it is necessary that there be a maximum restriction of the volume of initial data, and on the other hand, as follows from the preceding reasoning, in order to obtain effective dependences the length of the sample must be such that with its increase in the advance time interval the statistical evaluations do not substantially change.

The use of the proposed method in a number of cases makes it possible to decrease the uncertainty arising when sampling this parameter.

Thus, the examined principles make it possible not only to simplify the computation scheme and reduce the time expended on computations in problems using adaptive algorithms, but also to evaluate the optimum advance time periods and the necessary volume of the initial sample for obtaining effective prognostic dependences.

The proposed approach is especially preferable in the physical-statistical modeling of hydrometeorological fields of a great dimensionality.

BIBLIOGRAPHY

1. Bagrov, N. A., "Analytical Representation of a Series of Meteorological Fields Using Natural Orthogonal Components," TRUDY TsIP (Transactions of the Central Institute of Forecasts), No 74, 1959.

2. Gmurman, V. Ye., TEORIYA VEROYATNOSTEY I MATEMATICHESKAYA STATISTIKA (Theory of Probabilities and Mathematical Statistics), Moscow, Vysshaya Shkola, 1977.

3. Gruza, G. V., "Adaptive Systems for Statistical Analysis and Prediction of Weather in Meteorology," PRIMENENIYE STATISTICHESKIKH METODOV V METEOROLOGII (Application of Statistical Methods in Meteorology), Leningrad, 1971.

4. Gruza, G. V., Ran'kova, E. Ya., Esterle, G. R., "Model of Adaptive Statistical Forecasting Using a Group of Analogues," TRUDY VNIIGMI—MTsD (Transactions of the All-Union Scientific Research Institute of Hydrometeorological Information-World Data Center), No 13, 1976.

5. Kotov, M. M., "Use of Adaptive Regression Models for Predicting Waters in the Karatal River," TRUDY GMTs (Transactions of the Hydrometeorological Center), No 138, 1974.

6. Tolokontsev, N. A., "Computation of the Standard Deviation for Amplitude, Comparison With Generally Employed Methods," TEZISY DOKLADOV 3-go SOVESHCHANIYA PO PRIMENENIYU MATEMATICHESKIKH METODOV V BIOLOGII (Summaries of Reports of the Third Conference on Use of Mathematical Methods in Biology), Leningrad, 1961.

7. Yudin, M. I., FIZIKO-STATISTICHESKIY METOD DOLGOSROCHNOGO PROGNOZA POGODY (Physical-Statistical Method for Long-Range Weather Forecasting), Leningrad, Gidrometeoizdat, 1968.

UDC 551.542.1

PRESSURE GRADIENTS IN NARROW ZONES OF COLD FRONTS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 102-104

[Article by Candidate of Geographical Sciences M. A. Masterskikh, USSR Hydrometeorological Scientific Research Center, submitted for publication 26 Dec 79]

[Text]

Abstract: The author explains the appearance of considerable pressure gradients in narrow (up to 50 km) zones of cold fronts. The reason for this is the creation of large horizontal temperature contrasts in them. Formulas are given for computing the pressure gradients in narrow zones of cold fronts.

With the passage of atmospheric fronts, especially cold fronts, the wind always is intensified, sometimes to 30 m/sec or more. In the territory of the USSR 91% of all the observed cases of squalls appear on fronts, of which 74% of the cases were along cold fronts [1].

It can be assumed that one of the reasons for such an intensification of the wind is an increase in the pressure gradients in narrow (with a width of several tens of kilometers) frontal zones. In order to estimate the pressure gradient in a frontal zone it is necessary to have detailed ring maps where the pressure field along the front can be analyzed with sufficient accuracy.

In this article an attempt is made to explain the creation of considerable pressure gradients in cold air in narrow (with a width up to 50 km) zones of cold fronts on the basis of the following considerations.

In thermally uniform cyclones the pressure gradient is dependent for the most part on the angle of inclination of the isobaric surface toward the center of the eddy. In the zone of a cold front (in cold air) isobaric surfaces drop down simultaneously in the direction of the warm air. Accordingly, a component of the pressure gradient caused by the difference in densities of contacting air masses with different temperature arises here [3]. Accordingly, the isobars along the rear boundary of the indicated zone experience a corresponding bend, as indicated in Fig. 1.

We will denote the term of the pressure gradient vector directed along the frontal line to the center of the cyclone by $\nabla P_{fr\ line}$ and in the direction of the warm air by $\nabla P_{warm\ fr}$ (Fig. 1). The resultant vector of the pressure gradient in the frontal zone is

$$\nabla P_{fr\ zone} = \nabla P_{fr\ line} + \nabla P_{warm\ fr}. \tag{1}$$

Then

$$\Delta P_{fr\ zone} = \sqrt{\Delta P^2_{fr\ line} + \Delta P^2_{warm\ fr}}, \tag{2}$$

where Δ denotes a scalar value, that is, the length of projection of the vector.

The first component of the pressure gradient in the frontal zone $\Delta P_{fr\ line}$ in (2) is easily determined using data from a synoptic chart; the second component $P_{warm\ fr}$ can be computed using the scheme shown in Fig. 2, which shows the vertical section of the cold front. The following notations are used here: l is the width of the frontal zone; h is the height of columns of air a and b with a unit area of the cross section; ρ_1, t_1 are density and temperature in the warm air, ρ_2, t_2 are density and temperature in the cold column; P_1 and P_2 are the pressures at the base of the air columns a and b.

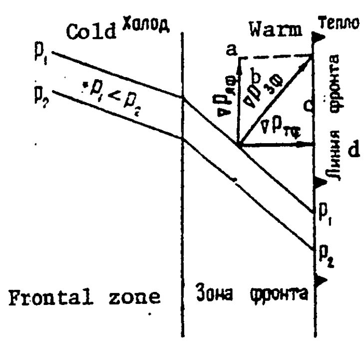

Fig. 1. Resultant vector of pressure gradient in cold front zone $\nabla P_{fr\ zone}$ and its components.

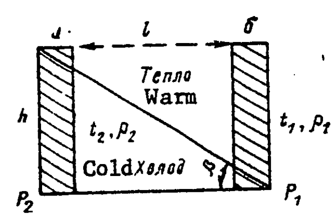

Fig. 2. Diagram for computing component of pressure gradient in cold front zone $\Delta P_{warm\ fr}$.

KEY:
- a) $P_{fr\ line}$
- b) $P_{fr\ zone}$
- c) $P_{warm\ fr}$
- d) Frontal line

The pressure difference $\Delta P = P_2 - P_1$ at the base of the columns a and b is determined from the following expression, corresponding to the equation of statics

$$\Delta P_{\text{тф}} = g\,\Delta\rho\,dz, \tag{3}$$

where $\Delta\rho = \rho_2 - \rho_1$ is the difference in densities between the cold and warm air, g is the acceleration of free falling, dz = h is the height of the air columns.

$$h = l\,\mathrm{ctg}\,\beta. \tag{4}$$

[Note: Strictly speaking, a quasistatic approximation cannot be used near a front. However, evidently, the influence of nonstationarity will be less than the influence of a large thermal contrast. Accordingly, for evaluating $\Delta P_{warm\ fr}$ with a certain degree of caution it is possible to use the equation of statics.]

With a width of the frontal zone ℓ equal to 50 km and a mean slope of the front $\beta = 0.02$ the height h = 1,000 m.

Elementary computations with formula (3) show that with a horizontal temperature difference in the columns a and b ($\Delta t_{fr\ zone} = t_2 - t_1$) equal to 1°C (in this case the $\Delta\rho$ value is equal to approximately $0.005 \cdot 10^{-3}$ g/cm^3 [4]) the value of the component of the pressure gradient in the cold front zone $\Delta P_{warm\ fr}$ is about 0.5 mb/50 km, that is, about 1 mb/111 km. Thus, in cold front zones the value of the horizontal temperature gradient is numerically equal to the pressure gradient component ($\Delta P_{warm\ fr}$), expressed in mb/111 km. Accordingly, for computing the resultant pressure gradient ($\Delta P_{fr\ zone}$ in mb/111 km) in expression (2) $\Delta P_{warm\ fr}$ can be replaced by $\Delta t_{fr\ zone}$ (the horizontal temperature contrast at a distance of 50 km from the frontal line in the direction of the cold air). Then formula (2) assumes the working form

$$\Delta P_{fr\ zone} = \sqrt{\Delta P^2_{fr\ line} + \Delta t^2_{fr\ zone}}\,. \tag{5}$$

This formula was used in computing the $\Delta P_{fr\ zone}$ values with a constant value $\Delta P_{fr\ line}$ = 3 mb/111 km, but with different values of the horizontal temperature gradient in the cold front zone $\Delta t_{fr\ zone}$. The following results of computations were obtained:

$\Delta t_{fr\ zone}$ °C/50 km	1	2	3	4	5	6	7	8
$\Delta P_{fr\ zone}$ mb/111 km	3.1	3.6	4.2	5.0	5.7	6.7	7.6	8.5

Thus, with a relatively small value of the pressure gradient along the frontal line (along which wind velocity is usually determined), but considerable values of the horizontal temperature gradient ($\Delta t_{fr\ zone}$ = 7-8°C per 50 km) the resultant pressure gradient $\Delta P_{fr\ zone}$ attains 7.6-8.5 mb/111 km. However, these pressure gradients in narrow zones of cold fronts, as we have already noted, virtually could not be detected on small-scale synoptic charts. The presence of such pressure gradients can be established reliably only from large-scale circular weather maps for a territory with a dense meteorological network.

We will show this in one of the examples. Figure 3 shows the actual fields of pressure and temperature in a cold front zone on the basis of data from small- and large-scale weather charts for approximately one and the same times. The data on the small-scale chart (a) show that the pressure gradient near the front in the direction of the cold air is about 2 mb/111 km. However, the data from the large-scale map (b) make it possible to see that the pressure trough is expressed more sharply and the pressure gradients in it behind the frontal line attain 5-6 mb/111 km.

The pressure gradients in the cold front zone $\Delta P_{fr\ zone}$, computed using formula (5) with $\Delta P_{fr\ line}$ = 1.3 mb/111 km and $\Delta t_{fr\ zone}$ = 5-6°C per 50 km, constitute 5.3-6.2 mb/111 km, that is, almost coincide with the actual values read from the large-scale chart. Approximately the same results were obtained in an analysis of the thermopressure fields during the passage of 20 cold fronts through Moscow. The values of the horizontal temperature contrasts in the 50-km zone in this case did not exceed 7°C. However, they can also attain 10-11°C [3].

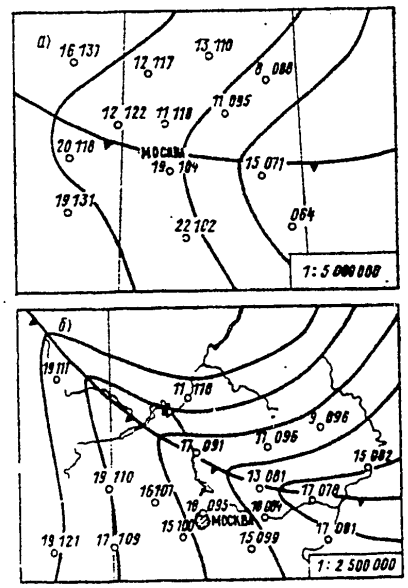

Fig. 3. Actual pressure and temperature fields in cold front zone on weather charts. a) 1800 hours on 5 June 1979; b) 1700 hours on 5 June 1979.

We note that in the cited example the maximum wind gusts were 22 m/sec. The wind speeds, calculated from the pressure field [2] on the small-scale chart (Fig. 3a) and the large-scale chart (Fig. 3b) are 6 and 18 m/sec respectively. Thus, in computing the pressure gradients (and accordingly, also the wind speed) at the time of passage of a cold front it is always necessary to take into account the horizontal temperature contrast in its narrow zone (50 km).

We note in conclusion that an increase in the pressure gradients should also occur in narrow zones of warm atmospheric fronts.

BIBLIOGRAPHY

1. Geokhlanyan, T. Kh., Shakina, N. P., ATMOSFERNYYE FRONTY (Atmospheric Fronts), Moscow, Znaniye, 1978.

2. Zverev, A. S., SINOPTICHESKAYA METEOROLOGIYA (Synoptic Meteorology), Leningrad, Gidrometeoizdat, 1957.

3. Kunits, A. V., SINOPTICHESKAYA METEOROLOGIYA (Synoptic Meteorology), Moscow, Voyenizdat, 1940.

4. Masterskikh, M. A., "On the Prediction of a Frontal Bora in the Novorossiysk Region," TRUDY GIDROMETTSENTRA SSSR (Transactions of the USSR Hydrometeorological Center), No 22, 1968.

5. Khrgian, A. Kh., OCHERKI RAZVITIYA METEOROLOGII (Outlines of the Development of Meteorology), Leningrad, Gidrometeoizdat, 1948.

6. Khromov, S. P., METEOROLOGIYA I KLIMATOLOGIYA DLYA GEOGRAFICHESKIKH FAKUL'-TETOV (Meteorology and Climatology for Geography Faculties), Leningrad, Gidrometeoizdat, 1964.

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INSTRUMENTATION, METHODS AND RESULTS OF INVESTIGATION OF ATMOSPHERIC ICE-FORMING NUCLEI

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 105-110

[Article by N. A. Berezinskiy, V. G. Karpov, G. B. Myakon'kiy, S. D. Smirnov, Candidates of Physical and Mathematical Sciences G. V. Stepanov and V. G. Khorguani, High-Mountain Geophysical Institute and State Special Design Bureau of Thermophysical Instrument Making, submitted for publication 1 Apr 80]

[Text]

Abstract: This article gives a description of instrumentation for investigating the microstructure, condensation and ice-forming properties of aerosols. The authors describe semiconductor thermodiffusion chambers with a broad range of monitorable changes in temperature and humidity which are automatically maintained with a high accuracy. Also examined are methods for investigation of the samples taken in the thermodiffusion chambers. The results of measurements of the concentration of ice-forming nuclei in the surface layer and in the free atmosphere are given.

Introduction. The broad range of artificial modification of clouds and fogs requires a thorough investigation of natural and artificial ice-forming nuclei. In addition to a study of the concentration of natural ice-forming nuclei, their nature and pattern of distribution in the atmosphere, great attention is devoted to the dependence of ice-forming activity on particle size, supersaturation and temperature, and also study of nucleation mechanisms. These investigations meet with considerable difficulties due to the small concentrations and their great variability, diverse chemical composition of the ice-forming nuclei and inadequate study of ice nucleation mechanisms. One of the principal difficulties in study of ice-forming nuclei is related to the lack of a reliable method and instrumentation for this purpose.

At the High-Mountain Geophysical Institute and the State Special Design Bureau of Thermophysical Instrument Making specialists have created a complex of surface and aircraft instrumentation for investigating the microstructural characteristics, condensation and ice-forming properties of natural and ice-forming aerosols. This instrumentation consists of different types of intake devices, making it possible to precipitate aerosols in the size range from 0.01 to 200 μm and thermoelectric diffusion refrigerating chambers with semiconductors operating in a wide temperature range.

Aerosol Sample Intakes

Filtering Apparatus. In the sampling of aerosol we used "Sinpor" millipore filters with a pore size 0.2-1.2 μm. One of the advantages of millipore filters is their relative simplicity and convenience of work with them under field conditions and in an aircraft and also the high efficiency of precipitation of submicron particles.

The design of the aircraft filtering apparatus makes possible the simultaneous taking of two air samples from an undisturbed flow through pipes inserted in the side of the aircraft [1]. Two pairs of filters are installed in the instrument in special holders. After the taking of samples with the first two filters the second pair is introduced into a working position by movement of the holder. Such a design makes it possible to carry out a virtually continuous sampling of the aerosol.

The great handling capacity of the vacuum pump makes it possible to take representative samples of aerosol in a short flight segment, that is, in a local region of space. The exposed filters are stored together with the holders in a special enclosure and are processed in a thermodiffusion cold chamber.

Cascade impactor. Particles in a wide range of sizes are precipitated on the filter, including an enormous number of active condensation nuclei. With the development of filters in a thermodiffusion cold chamber already at a humidity of more than 70% many of the nuclei are transformed into droplets of solutions. As a result of the absorption of water vapor by condensationally active particles it becomes inadequate for the activation of all the ice-forming nuclei, that is, there is a so-called volume effect [10].

In addition, the matte surface of the filter and the penetration of some particles within the pores make it difficult to observe the nucleation processes and investigate the particles themselves.

We designed a three-stage impactor trap (Fig. 1a) which broadens the possibility for investigating ice-forming nuclei using the filters and eliminates some of their shortcomings [3]. In the first stage the escape velocity is 60-70 m/sec and particles with a diameter greater than 0.8 μm are precipitated. In the second stage we made use of the principle of a critical (sound) velocity of escape from a slit nozzle measuring 10 x 0.4 mm and particles of 0.1-0.05 μm are precipitated in it. The remaining particles are precipitated in the third stage — the filter.

An investigation of the effectiveness of precipitation in the impactor trap was described earlier in [6]. Test disks or glasses are covered by a colloid film. Their design makes it possible to investigate the ice-forming properties of the aerosols and their microstructure under optical and electron microscopes, more precisely monitor humidity and temperature and carry out observations of condensation processes and the nucleation mechanism on individual particles. With the precipitation of particles in the impactor on a fixed or slowly rotating plate with a humidity of the investigated air 80-100% there will be a process of a showerlike spraying of droplets from the sample [4]. This is attributable to the condensation enlargement of particles in the nozzle.

In the described instrument the precipitation of particles occurs on a rapidly rotating disk. The probability of fusion of the precipitating droplets in the sample is reduced. During the time of a full rotation the water envelope can be completely evaporated. Such a rate of rotation was experimentally selected at which there was no spraying and flooding of the sample even when working in the clouds. It was 5-10 sec^{-1} and the volumetric air flow attained 30 liters/min.

Thus, a cascade impactor makes it possible to investigate the ice-forming properties and microstructural characteristics of aerosols by fractions in the clouds and in the pure atmosphere. The intakes described above can be used both in an aircraft and on the ground.

Thermoelectric Diffusion Chambers

"Grad" unit. Special chambers with a high accuracy of maintenance of temperature and humidity in a broad range are required for determining the concentration of ice-forming nuclei and investigating the condensation and ice-forming properties of aerosols collected in the form of a sample on millipore filters or some other backings. Our early investigations [1] indicated that chambers operating on freon, liquid nitrogen or solid carbon dioxide do not ensure a high accuracy in maintaining the stipulated temperatures.

Accordingly, the "Grad" apparatus was developed (Fig. 1b). It consists of two thermodiffusion cold chambers -- "Grad-1" and "Grad-2," operating on semiconductor thermopiles. The operating principle of each of these chambers is as follows: a definite temperature gradient is created between two thermostated surfaces, whose temperature is maintained autonomously. One of these surfaces is covered with ice and is a source of water vapor in the chamber. The second surface is a sort of "stage" for the investigated sample (a filter or other backing). Thus, by stipulating a definite temperature of the sample and the gradient between the source of water vapor and the investigated object it is possible to create the regime of temperature and saturation (relative to water and ice) necessary for the investigation.

A change in the temperature of the plates for the vapor source and the stage is accomplished by semiconductor thermopiles. Their working junctions are in thermal contact with the thermostated plates. Depending on the direction of the current passing through the thermopile the thermostated plate is either heated or cooled. The offtake of heat from the thermopile is accomplished by means of a FAK-0.7 cooling apparatus.

Both chambers are controlled by the same electronic automatic control unit. The temperature of the thermostated surfaces can be varied from -35 to +20°C and is automatically maintained with an error to 0.05°C. The measurement is made by thermocouples which have an output to a VK 2-20 digital voltmeter. Up to four samples can be investigated simultaneously in the chambers.

Distinguishing features of the "Grad-1" and "Grad-2" chambers are that in the first the water vapor source is in the same plane with the surface of the investigated object on an annular plate around the sample, and in the second — over the investigated sample. For this purpose an ice layer is frozen on the thermostated surface situated over the sample.

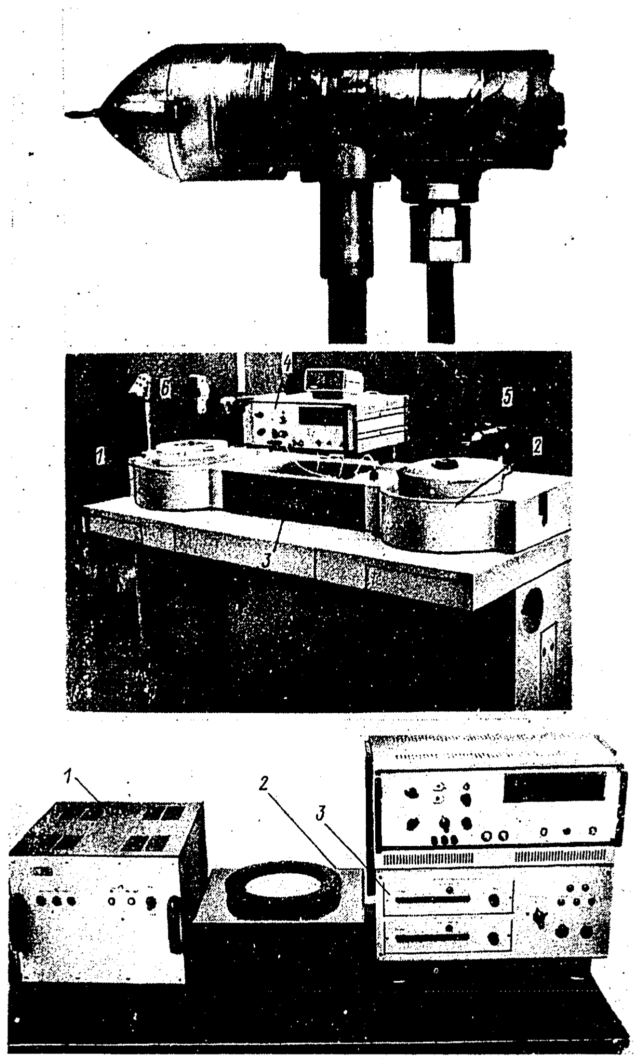


Fig. 1. General view of measuring apparatus complex. a) cascade impactor; b) "Grad" thermoelectric diffusion apparatus — 1) "Grad-1" chamber; 2) "Grad-2" chamber; 3) control panel; 4) digital voltmeter; 5) MBS-2 microscope; 6) camera; c) "Grad-3" chamber: 1) power unit; 2) thermal unit; 3) control unit.

"Grad-3" aircraft thermoelectric diffusion chamber. The "Grad" apparatus is a stationary laboratory instrument. The small thermoelectric diffusion chamber "Grad-3" (Fig. 1c) was developed and created for operation under field conditions and on an aircraft. The principal difference between the "Grad-3" chamber and the "Grad" apparatus is two-stage regulation of the temperatures of the working plates in the chamber by the thermocouples. The use of such a system increases the accuracy of maintaining the temperature (0.05°C) and considerably reduces the size and mass of the chamber. On the heat-releasing surface of the lower stage there is a radiator of the combined type (air or water cooling), making it possible, depending on the operating conditions, to employ one or another heat transfer method.

In the "Grad-3" chamber provision is made for automatic maintenance of the thermostated temperature of the working surfaces in the range from +20 to -25°C. The temperature is controlled by copper-constantan thermocouples and indication is with a digital recording instrument. All the units are small in size and convenient for transport by hand. In the chambers there can be three mechanisms for ice nucleation: sublimation, condensation- freezing and volume freezing. Usually the aerosol samples in the thermodiffusion chamber are processed in the following way: the temperature of the aerosol sample is established, then the temperature of the ice surface (vapor source) slowly increases and accordingly there is an increase in the humidity over the sample. However, as mentioned above, due to the limited rate of vapor diffusion and the high rate of condensation processes there are water vapor depletion effects. In order to avoid this we tested another method for processing the samples in which the chamber is maintained with a definite temperature of the ice surface so that the necessary saturating pressure of the water vapor is established throughout its volume, after which the necessary temperature of the aerosol sample is rapidly established. With development by the proposed method the number of crystals is found to be 30-100% greater than when using the usual method.

The errors in determining the concentration of ice-forming nuclei are related to the accuracy in measuring the volume of the investigated air, the background contamination of the backings, conditions for development of the sample and counting of crystals. According to our estimates, with t = -20°C the error does not exceed approximately 25%. With an increase in temperature to -8°C it increases to 50%.

Research Results

We carried out numerous surface and aircraft measurements of ice-forming nuclei over different physiographic regions of the USSR at altitudes up to 5,500 m above sea level. The samples were developed at temperatures -8, -15 and -20°C and a humidity of 100% using a method developed at the High-Mountain Geophysical Institute, reducing the influence of the volume effect [5].

The investigations which were made [2] indicated that in the surface layer the concentration of ice-forming nuclei is dependent on the measurement region. At a temperature of -20°C it varies from 1 to 3 liter^{-1} in the foothill regions of the Northern Caucasus, over the Black and Caspian Seas and to 7-8 liter^{-1} in the semi-deserts and deserts of Central Asia. The greatest concentration of ice-forming nuclei (up to 15 liter^{-1}) is observed over industrial cities.

The influence of the underlying surface on the concentration of ice-forming nuclei is reflected in the lower layer of the atmosphere to an altitude of 1,000-1,500 m above sea level. In only half the cases is there a positive correlation between the vertical distribution of the ice-forming nuclei and condensation nuclei.

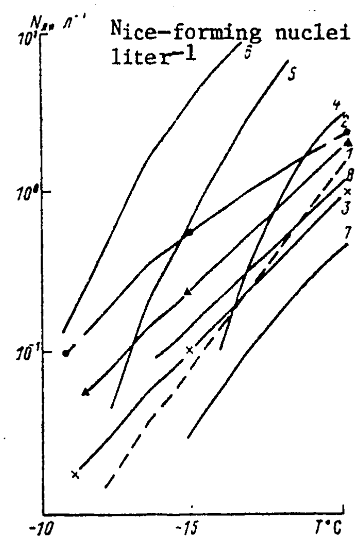


Fig. 2. Dependence of concentration of ice-forming nuclei on temperature. 1) Terskol, 2) Labinsk, Krasnodarskiy Kray, 3) Tul'skaya Oblast; 4) according to data in [13], 5, 6) according to [8] (Japan), 7) data in [9], 8) data in [11] (Tasmania).

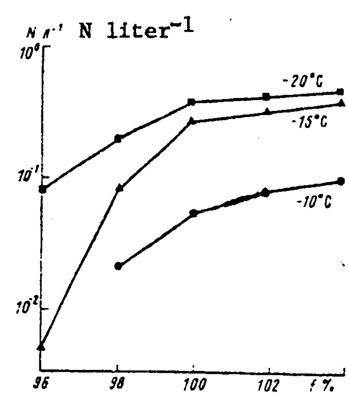


Fig. 3. Dependence of concentration of ice-forming nuclei on supersaturation.

A comparison of the averaged vertical profiles of the concentration of ice-forming nuclei and aerosol particles in different size ranges shows that the decrease in the concentration of ice-forming nuclei with altitude occurs more slowly than the change in Aitken nuclei and agrees well with the variation in the concentration of large nuclei at all altitudes and giant nuclei to an altitude of 2,000 m above sea level.

A series of experiments was carried out for investigating nucleation mechanisms, the dependence of ice-forming activity on temperature, the size of particles and supersaturation.

Figure 2 shows the dependence of the concentration of ice-forming nuclei on temperature on the basis of our surface and aircraft measurements in comparison with the data of other authors for different regions of the earth [8, 9, 11, 13]. From the data of other authors in the literature we intentionally selected and in this figure we have given those which graphically show the limiting values of the measured concentration of ice-forming nuclei. Here it must be noted that the overwhelming majority of the concentrations known in the literature fall between the curves 3 and 4.

Table 1

Number of Natural Ice-Forming Nuclei in Sample Obtained With Development by Different Methods

t = -8°C		t = -15°C	
without moistening	with moistening	without moistening	with moistening
none	14	40	190
3	10	53	120
none	35	5	60
none	4	57	80

Table 2

Concentration ($liter^{-1}$) of Ice-Forming Nuclei by Impactor Stages

	1 Высота от поверхности земли, м											
	100	1000	2000	3000	100	1000	2000	3000	100	1000	2000	3000
	t=—8°C				t=—15° C				t=—20° C			
2 I каскад—гигантские ядра	0,04	0,04	0,03	0,02	0.35	0,36	0,16	0,15	1,5	1,34	0,98	0,41
3 II каскад—крупные ядра	—	—	—	—	0,06	0,06	0,07	0,1	0,49	0.35	0,38	0,25
4 III каскад — ядра Айткена	—	—	—	—	0,98	0,7	0,3	0,53	2,3	2,0	1.02	0,34
5 Сумма	0,04	0,04	0,03	0,02	1,39	1,12	0,53	0.78	4,29	3,69	2,38	1,0

KEY:
1. Altitude from earth's surface, m
2. Stage I -- giant nuclei
3. Stage II -- large nuclei
4. Stage III -- Aitken nuclei
5. Sum

The qualitative shape of the curves for all our measurements coincides and agrees well with the data of other authors. Despite the fact that the concentration of ice-forming nuclei at different points on the earth has a considerable scatter, its change with temperature in the range -10 - -20°C occurs approximately exponentially. This circumstance is clearly illustrated in the figure where the broken line represents the dependence for the concentration of ice-forming nuclei proposed by Fletcher [7]

$$N = N_0 \exp\ (\beta \Delta T),$$

where $N_0 = 10^{-5}$ liter^{-1}, $\beta = 0.6$, ΔT is the supercooling temperature.

In cumulus clouds condensation usually occurs at a positive temperature and accordingly a high water vapor pressure; then the forming droplets and aerosol particles are lifted into the region of negative temperatures. In order to bring the conditions for development of the samples closer to cloud conditions, experimentation was carried out using the following method. A humidity of 100-101% was established over the sample at a temperature of 10°C. Maintaining the mentioned humidity, the sample was cooled to the planned negative temperature. Table 1 gives the results of these experiments [5].

The table shows that with development of the sample with its preliminary moistening the number of crystals increased considerably. This is evidently a result of freezing of droplets containing ice-forming nuclei.

Microscopic observations of condensation and nucleation of the ice phase on giant nuclei precipitated in the first stage of the impactor also indicated that the formation of the crystals passes through the liquid phase. A sector of the aerosol sample with forming crystals at a humidity 100-101% was photographed. Then with undersaturation relative to the ice the crystals evaporated, the sample was heated to +10°C and photographed at a humidity of 100-101%.

A matching of the microphotographs indicated that all the crystals obtained at a temperature of -8 and -15°C coincided with the most condensationally active centers at a positive temperature.

A series of experiments was carried out for investigating the dependence of the ice-forming activity of natural aerosols on particle size. Samples of each stage were developed at temperatures of -8, -15 and -20°C and a humidity of 100-101%. The results of one of the experiments are given in Table 2. The table shows that high-temperature ice-forming nuclei are giant aerosol particles and despite their relatively small concentrations they make a substantial contribution to the total number of ice-forming nuclei.

An analysis of the results of investigations by different authors shows [12] that the great scatter obtained by different counters of ice-forming nuclei for both natural and artificial aerosols can be partially, if not completely, attributed to inexact monitoring of humidity conditions in the investigated chambers. We carried out a series of experiments for determining the dependence of the concentration of ice-forming nuclei on humidity (supersaturation). The samples were taken simultaneously under identical conditions on several filters which were developed at humidities 96, 98, 100, 102 and 104%. Figure 3 shows the averaged curves of the dependence of the concentration of ice-forming nuclei on humidity at temperatures -10, -15 and -20°C. The curves show that with all temperatures the concentration of the ice-forming nuclei increases considerably with an increase in humidity. Accordingly, the ice-forming activity of a natural aerosol is dependent on temperature, water vapor saturation and particle size.

Thus, a unique complex of surface and aircraft apparatus was created for investigating the concentration, spectrum, condensation and ice-forming properties of aerosols. This apparatus was used in carrying out investigations of the ice-forming activity of natural aerosol in dependence on supersaturations, temperature and particle size. Measurements of the concentrations of ice-forming nuclei over different physiographic regions of the country were made.

BIBLIOGRAPHY

1. Myakon'kiy, G. B., Khorguani, V. G., "Methods and Apparatus for Measuring Ice-Forming Nuclei Using Millipore Filters," TRUDY VGI (Transactions of the High-Mountain Geophysical Institute), No 24, 1973.

2. Myakon'kiy, G. B., Stepanov, G. V., Khorguani, V. G., "Results of Measurement of Atmospheric Ice-Forming Nuclei in the Northern Caucasus," METEOROLOGIYA I GIDROLOGIYA (Meteorology and Hydrology), No 3, 1976.

3. Sarkisov, S. L., Stepanov, G. V., Myakon'kiy, G. B., Khorguani, V. G., "Three-Stage Instrument for Investigating the Microstructure and Ice-Forming Properties of Natural and Artificial Aerosols," TRUDY VGI, No 34, 1977.

4. Sarkisov, S. L., Stepanov, G. V., Shvedov, S. V., "Nature of the Precipitation of Solid Aerosols in Impactor Traps," TRUDY VGI, No 8, 1968.

5. Stepanov, G. V., Berezinskiy, N. A., Medaliyev, M. Kh., "Some Results of Investigation of Natural Ice-Forming Nuclei by a Cascade Impactor," TRUDY VGI, No 42, 1979.

6. Khorguani, V. G., Stepanov, G. V., Sarkisov, S. L., "Ground and Aircraft Condensation Impactor Traps for Investigating Aerosols," IZV. AN SSSR, FIZIKA ATMOSFERY I OKEANA (News of the USSR Academy of Sciences, Physics of the Atmosphere and Ocean), Vol 6, No 7, 1970.

7. Fletcher, N. H., THE PHYSICS OF RAINCLOUDS, Cambridge Univ. Press, 1962.

8. Isono, K., Komobayasi, M. S., METEOROL. SOC. JAPAN, SER. II, Vol 37, 1959.

9. Kline, G. W., MON. WEATHER REV., Vol 91, 1963.

10. Langer, J., Rodgers, J., "An Experimental Study of Detection of Ice Nuclei on Membrane Filter and Other Substrate," J. APPL. METEOROL., Vol 14, No 4, 1975.

11. Mossop, S. C., et al., QUART. J. ROY. METEOROL. SOC., Vol 96, 1970.

12. PROC. OF THE SECOND INT. WORKSHOP ON CONDENSATION AND ICE NUCLEI, Fort Collins, Colorado, U.S.A., 1971.

13. Stevenson, C. M., "An Improved Millipore Filter Technique for Measuring the Concentrations of Freezing Nuclei in the Atmosphere," QUART. J. ROY. METEOROL. SOC., Vol 94, No 399, 1968.

UDC 551.(46+51)+556:62-52

REMOTE ACTIVATION AND CONTROL OF INSTRUMENT OPERATION

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 111-113

[Article by V. V. Prokof'yev, Southern Division Institute of Oceanology USSR Academy of Sciences, submitted for publication 26 Nov 79]

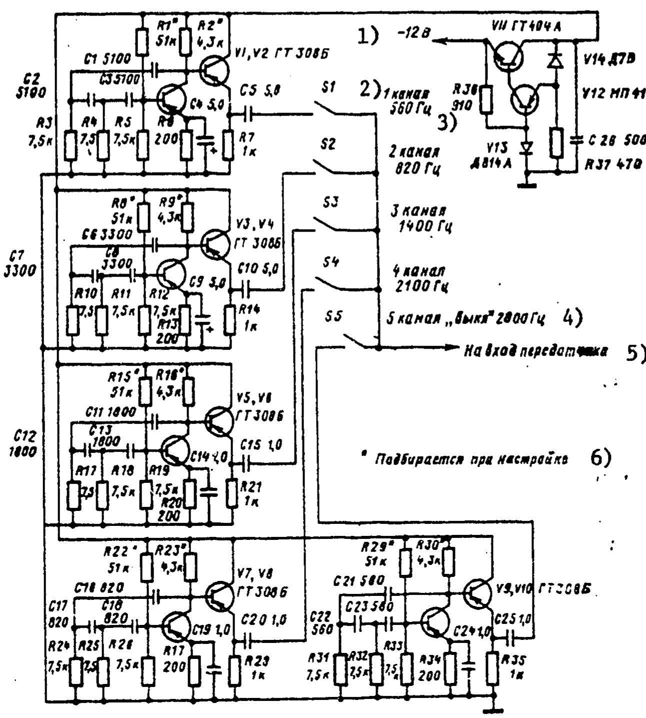
[Text]

Abstract: The article describes the technical realization of the possibility of use of a radio communication channel for the remote control of operation of devices (including telemetric) which are situated at a considerable distance from the data user (collection center). With such a design of measuring systems there are a number of advantages (routineness in collection of information, economy in current consumption, increase in reliability, etc.) in comparison with systems operating under a rigid program.

In carrying out hydrological and meteorological observations from a stabilized buoy the need frequently arises for the brief activation of the instruments at different times of day. With the switching-on of these instruments for constant operation there is an unjustified expenditure of energy from the storage batteries. In order to save electric power and increase the reliability of instrument operation the remote control of instrument operation through a radio channel has been developed.

Radio control was developed on the basis of standard-produced radio stations of the "Niva" or "Karat-M" type with an effective range of 20 km, an output power of 0.5 W and operating in one side band. The transmitting part of the radio control is illustrated in Fig. 1. It consists of five generators of sinusoidal oscillations tuned to frequencies of 560, 820, 1400, 2100 and 2800 Hz. The power supply for the generators is stabilized by a stabilizer (V11-V13), forming a compensation-type stabilizer circuit. The switches S1-S5 are used for connecting the outputs of the generators to the input of the transmitter modulator.

The receiving part of the radio control, situated on the buoy, is illustrated in Fig. 2. The signal is fed from the output of the receiver of the "Karat-M" radio transmitter to an amplifier with a logarithmic scale (A7), from whose output it is fed to the inputs of active filters functioning in K140UT1A microcircuits and tuned to the frequencies 560, 820, 1400, 2100 and 2800 Hz. The signal is fed from

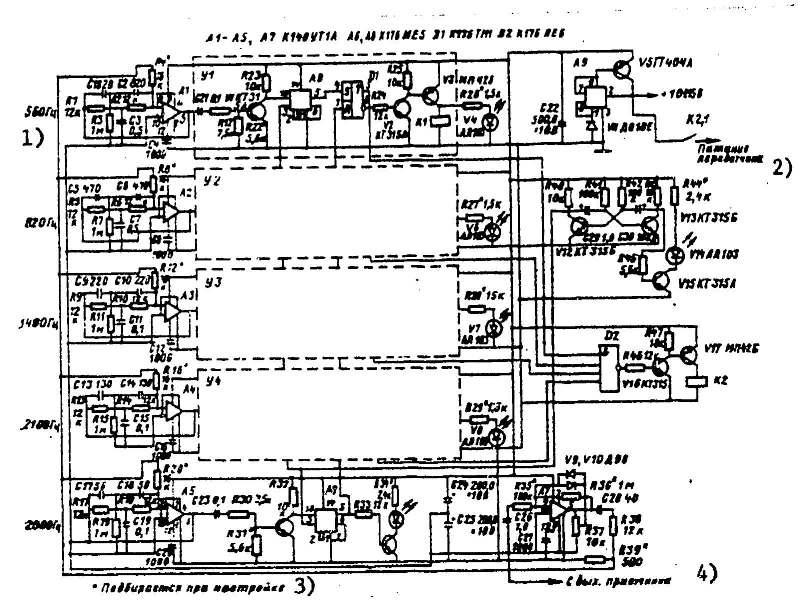


KEY:
1) V
2) Channel
3) Hz
4) cut out
5) to transmitter input
6) is selected during tuning

Fig. 1. Block of modulators for remote control transmitter.

the outputs of the filters to amplifier-limiters based on KTZ15A transistors. The downward level of limitation of the input signal is set by the proper choice of the resistor R22. The signal is fed from the output of the amplifier-limiter to a 15-digit counter, incorporated in the microcircuit A8 (K176IYe5). The overflow pulse from the counter (2^{15} pulses) trips the toggle switch D1 and activates the relay K1 which with its contacts supplies a flow of current to the corresponding instrument. The other channels are activated in the same way. The channel 2800 Hz is used for throwing the toggle switches. An overflow pulse from the counter sets the flip-flops in the initial state, that is, cuts out the relay supplying current to the instruments. Thus, with activation of the channels 560, 820, 1400 or 2100 Hz in the transmitting part there is activation of the corresponding channels in the receiving part and the corresponding instruments are cut in. The channel 2800 Hz is used for general deactivation of the instruments.

Spontaneous activation from radio interference is eliminated by a protective circuit, a multivibrator with the transistors V12, V13 and 15-digit pulse counters assembled in K176IYe5 microcircuits (A8). The multivibrator generates pulses with a repetition rate of 23.0 sec and a pulse duration of about 1 sec. These pulses of positive polarity set the pulse counters in a zero state each 23.0 sec. Accordingly, in order for an overflow pulse to appear at the counter output at its input



KEY:
1) Hz
2) Current supply for transmitter
3) Is selected during tuning
4) From receiver output

Fig. 2. Remote control receiver.

there must be a signal consisting of $2^{15}-2^{9}+1$ pulses (the first nine digits of the counter are not in a zero state and therefore we will assume that after the counter is put into its zero state the first nine digits are already filled). The probability that 31,475 noise pulses will be accumulated in 23 sec is small, whereas with the continuous feeding of a signal with a frequency of 1400, 2100 and 2800 Hz for 23.0 sec this condition is satisfied and the cutting-in of the actuating devices through some channel will occur. Fourteen counter digits are used for more low-frequency control signals (560 and 820 Hz). With the activation of any channel there is simultaneous cutting-in of the current supply to the transmitter, through which information is transmitted from the buoy. This is accomplished using the microcircuit D2, the transistors V16, V17 and the relay K2.

If a zero potential appears across at least one input of the microcircuit (4I-NYe) (the flip-flop is thrown for cutting-in the channel) a positive potential appears across the output of this microcircuit, activating the relay K2, which with its contacts feeds current to the transmitter.

Visual monitoring of operation of the device is accomplished using the photodiodes V4, V6, V7, V8, V14. The current operating regime is governed by resistors connected in series with the diodes.

The receiving part of the radio control, together with the receiver in a duty regime, requires about 50 mA.

Such radio control was installed on a stabilized buoy at a distance of three miles from the shore for the remote activation and deactivation of the instruments installed on the buoy.

BIBLIOGRAPHY

1. Yakubovskiy, S. V., ANALOGOVYYE I TSIFROVYYE INTEGRAL'NYYE SKHEMY (Analog and Digital Integrating Circuits).

HISTORY OF METEOROLOGICAL OBSERVATIONS IN TURKMENISTAN

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 114-115

[Unsigned]

[Text] Abstract: The article discusses the history of organization and development of meteorological observations in a number of regions of Turkmenistan and evaluates the role of meteorological stations in servicing of the Turkmen national economy.

A hundred years have passed since the beginning of meteorological investigations in Turkmenistan. The first meteorological station was established there in 1869 at Krasnovodsk. Russian tsarism was not concerned with the development of scientific research at the margins of the country, but the enthusiasm of Russian scientists, combined in the pre-Revolutionary period by the Russian Geographical Society and the Main Physical Observatory at Peterburg, together with their interest in the unstudied territories of the country, favored much success in the development of the network of meteorological stations in the Transcaspian region, in the territory of present-day Turkmenistan.

The year 1979 marked the 110th anniversary of meteorological observations at Krasnovodsk and the 90th anniversary of observations at Kerki and Bayram-Ali.

A meteorological station was established in 1889 on the left bank of the Amudar'ya, in the southern part of Bukhara emirate, on the outskirts of Kerki village, by the Military-Topographic Section of Turkestanskiy Kray. The methodological direction of operation of the meteorological station was from the Main Physical Observatory in Peterburg. Russian professional meteorologists were highly enthusiastic about the meteorological investigation of Turkestanskiy Kray.

The organization of a meteorological station in the Amudar'ya valley was related to the expansion of use of the resources of the kray and meteorological investigation of the unstudied territory. In July 1912 these uninvestigated places were examined by A. I. Voyeykov. He sailed on a steamer along the Amudar'ya from Termez in Bukharskiy emirate and then to the vassal Khivinskoye khanate.

After the October Revolution the city of Kerki became a part of the Bukharskaya Republic, and in 1924 -- a part of the Turkmen SSR.

During the years of Soviet rule in this part of the republic, as in all of Turkmenia, the development of the national economy proceeded at a rapid pace. And meteorology developed together with development of the national economy: from very narrow elementary observations only of precipitation, temperature, humidity and air pressure the meteorological station proceeded to observations under a broad program, including aerological and agrometeorological.

The Kerki meteorological station today is a station making a great volume of observations, carrying out broad servicing of agriculture, aviation and the multibranch economy of the zone in the first segment of the Karakumskiy Canal imeni V. I. Lenin.

Continuous observations made over the course of almost a century have made it possible to clarify all the peculiarities of climate in the southeastern corner of the republic. The materials of these observations have been included in all the major reference books on the climate of Turkmenistan and in other studies of climatologists.

The Bayram-Ali meteorological station, the first station in Murgabskiy oasis, was organized by the Murgabskoye state estate in 1889. Over a period of years the observations were made in the nursery of this estate.

The city of Bayram-Ali is situated in the northern part of the oasis, fed by the waters of the Murgab. It was constructed near the ruins of old Merv -- one of the most ancient cities of Central Asia. The organization of the meteorological station was associated with the insistent need for studying the geographical and climatic conditions of Murgabskiy oasis. The Transcaspian region at that time was virtually unstudied. The Russians ran railroads into the depths of the region -- from Krasnovodsk to Chardzhou (and then to Samarkand and Tashkent) and from Merv (now Mary) to Kushka.

The Bayram-Ali meteorological station was the reference station for study of the climatic resources in the zone of Russian cotton production. The methodological direction of its operation was from the Main Physical Observatory in Peterburg. The station was visited by outstanding professional meteorologists. For example, on 17 June 1897 the station was inspected by S. I. Savinov -- a Russian professional meteorologist, inventer of a number of meteorological instruments.

During his journeying in Turkestan A. I. Voyeykov -- a famed Russian geographer, the founder of Russian climatology -- visited the Merv oasis. He thoroughly inspected the Murgabskoye state estate and also its administration, under which the meteorological station was organized.

As noted by Voyeykov, "...the Emperor Aleksandr III in 1887 appropriated unto himself all the unworked lands which can be irrigated by the water stored by the Sultan-Bentskaya Dam. The Tsar appropriated a total of 130,000 hectares of lan to himself."

After familiarizing himself in detail with the oasis, Voyeykov thought a great deal about how to extend the boundaries of the oasis and irrigate the desert lands. He came to the conclusion that it was possible to increase the water resources of the Turkmen rivers Murgab and Tedzhen only at the expense of the powerful Amudar'ya.

Voyeykov regarded the future plan of irrigation for an extensive area, including the Merv oasis, by the waters of the Amudar'ya as "one of the most important enterprises of this type on the earth." Voyeykov believed in the future of Turkestan and especially the Transcaspian region. And his dream became a reality.

The waters of the Amudar'ya reached the Murgab valley in 1959. Now the Murgab and Tedzhen oases do not experience a water shortage. The Murgab oasis has become the principal base of valuable fine-fiber varieties of cotton. Already in the late 1920's and the early 1930's work began here on the acclimatization of "guests" from the Nile valley and from California. During the years of the first five-year plans Soviet plant breeders created Russian fine-fiber varieties of cotton, of a higher quality and more rapidly maturing, with a greater yield than the Egyptian and American varieties. Meteorologists also made their contribution to the successes of the plant breeders.

A kidney sanatorium, a remarkable health facility which prior to the Revolution was a royal estate, has operated at Bayram-Ali for 45 years. This sanatorium is well known not only in our country, but also abroad. In the course of all the activity of this sanatorium the data from the Bayram-Ali meteorological station have constantly been used in the investigations. Even in our day it is a reference station which is part of the control network. Its observational data are used extensively for study of the climatic resources of Murgab oasis.

Meteorological observations were begun at Krasnovodsk 110 years ago. Krasnovodsk, a port in the northwestern part of Krasnovodskiy Gulf of the Caspian Sea and the initial point of the Central Asian Railroad, is called the gates of Central Asia. It was founded in 1869 on the shores of the gulf, in small Murav'yevskaya Bay, surrounded by the spurs of the rather large Shagadan Mountains. As early as 1715 a fortress had been constructed on the shores of the Krasnyye Vody embayment prior to the founding of the city. This fortress was constructed by Aleksandr Bekovich-Cherkasskiy, who headed an expedition for studying the eastern shores of the Caspian Sea and seeking the former mouth of the Amudar'ya River. The reddish cliffs surrounding the embayment from the north, northwest and northeast and their reddish reflection in the water gave a name to both the embayment and the city.

The founding of Krasnovodsk was of great economic importance for Russia and Central Asia, which in 1862 became part of Russia. Goods went through this port city from Central Asia to the Caucasus and along the Volga into the central regions of Russia.

With the founding of the city in this largely unknown region the need also arose for climatic investigations. A meteorological station was organized on the very shores of the gulf. And now meteorological observations have already been made for 110 years at Krasnovodsk. Russian scientists -- geographers and meteorologists -- even then devoted great attention to the Transcaspian region.

In 1883 the meteorological station at Krasnovodsk was visited "for inspection," as it was then called, by the director of the Main Physical Observatory in Peterburg, Academician M. A. Rykachev, and in 1897 it was visited by the outstanding Russian meteorologist S. I. Savinov. On 30 September 1910 the station was inspected by the well-known climatologist A. A. Kaminskiy. Early in April of 1912 the Krasnovodsk station was visited by A. N. Voyeykov.

The data from the century-long meteorological observations at Krasnovodsk, at the present time being the sea, railroad and air portal to Central Asia, are extensively used in the national economy and science.

Since 1940 there has also been another suburban meteorological station at Krasnovodsk. It has a synoptic team.

On the basis of the century-long observations by the meteorological stations Krasnovodsk, Bayram-Ali and Kerki it has been noted that since the winter of 1968-1969 a period of strong cooling has begun in Turkmenia. It goes without saying that this applies only to the winter. The last decade has been characterized by particularly cold winters with much snow. For example, during the winters of 1968-1969, 1971-1972, 1973-1974, 1976-1977 the mean annual air temperatures in most of the regions of Turkmenia have been 9-14° below the norm. The absolute minimum temperature values for 16 stations in the republic, including Krasnovodsk, Bayram-Ali and Kerki were equaled. The mean annual air temperatures for 1969 were the lowest for the century-long operating stations.

It can be postulated that since 1969 the climate of Turkmenia has entered a new secular cycle. The springs have become different -- drawn-out, cold, with frequent rains, thunderstorms and hail. Usually hot, summer-like dry weather sets in already with the second half of May. During the last decade, in almost all years, the end of May and the beginning of June can be assigned to the Turkmen spring: there were no hot spells, dryness, and precipitation frequently fell and sharp temperature drops were noted. But in summer, in general, it has been very hot in Turkmenia, as before.

'CLOUD ATLAS' (ATLAS OBLAKOV), EDITED BY A. Kh. KHRGIAN AND N. I. NOVOZHILOV, LENINGRAD, GIDROMETEOIZDAT, 1978, 267 PAGES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 116-117

[Article by L. T. Matveyev and T. V. Ushakova]

[Text] Cloud cover observations are among the most important in the network of meteorological stations. Detailed data on the observed genera, species and varieties of clouds constitute information mandatory for all synoptic stations in the world. The composition of this information is regulated by international requirements [3], in accordance with which the international cloud classification was formulated. The principal manual for determining the species of observed clouds is the cloud atlas, whose principal content is photographs of typical genera, species and varieties of clouds and also a description of the principal structural elements, intended for assisting the observer in establishing a correspondence between the observed cloud and one of the photographs in the atlas.

The CLOUD ATLAS published in 1978 under the editorship of A. Kh. Khrgian and N. I. Novozhilov [1] in this respect differs advantageously from the preceding one [2]. The new atlas includes 101 tables (photographs) of clouds, in particular, a strictly maintained sequence of arrangement of cloud species in tables. The names of a number of clouds are made more precise, in accordance with international practice. These considerably more completely represent combinations of different genera of clouds which are frequently observed in nature. Much work has been done on refining the captions to the tables: in contrast to the preceding edition, the captions have become more specific, with attention being concentrated on the peculiarities of the structural elements of the cloud. In this respect the "arrows" pointing to cloud elements, which should be mentioned, taken from the INTERNATIONAL ATLAS, are useful [see 4].

The possibility of use of photographs from the INTERNATIONAL CLOUD ATLAS made possible a considerable increase in the distinctiveness and degree of expressivity of the tables in the new atlas. Here we should also note the great care exercised in the choice of tables both from the INTERNATIONAL ATLAS (94 tables) and from the atlas of 1957 (6 tables) and the high level of the printing of the color photographs.

The indication of the figures of the KN-01 code in each table for clouds of the lower (C_L), middle (C_M) and upper (C_H) levels is of considerable importance for increasing the reliability and uniformity of determination and coding of the observed cloud genera. These are used in coding observations in the field. Unfortunately, in the descriptive part of the atlas this useful innovation is not mentioned at all.

Among the shortcomings of the new atlas (the same as in all preceding ones) we should mention the absence of methodological instructions in it on the determination of cloud genera when using the atlas. In our opinion, such publications should have a section devoted to the rules for use of the atlas.

It should also be noted that the changes introduced into the new cloud atlas in comparison with past publications are not all pointed out in the foreword or are pointed out without setting forth the significance of the changes. For example, the change in the designation of Ci fil to Ci fib is attributable to the fact that a fibrous structure is more typical for this type of clouds: the variety of these clouds (intortus, vertebratus, uncinus) usually does not have a distinct filamentary structure. Accordingly, such a change in name has been adopted in international practice. In some places in the new atlas a double name of this species has nevertheless been retained in some places — in Russian "voloknistyye, nitevidnyy (see pages 220, 260) and the Latin term "filosus" is already excluded.

No explanation is given for the change in the name fractonimbus clouds to fractocumulus poor-weather clouds (Table 66). It appears that the term fractonimbus clouds in this case is sounder. In the INTERNATIONAL CLOUD ATLAS the new term has not been introduced in a sufficiently full form -- in the Latin name there are no supplementary words separating these clouds ($C_L = 7$) from ordinary fractocumulus clouds ($C_L = 1$), although with respect to the conditions for formation and external appearance fractonimbus and fractocumulus clouds differ substantially. For this reason it seems sounder to call poor weather fractocumulus clouds, as before, fractonimbus clouds.

The expressed comments cannot change the general high opinion of the discussed major CLOUD ATLAS, whose preparation for publication required many efforts on the part of compilers, editors and the publishing house. Although its basis is the INTERNATIONAL CLOUD ATLAS, the CLOUD ATLAS is more compact and convenient for use. However, the inclusion in the CLOUD ATLAS of sections devoted to a description of clouds on the basis of data from radar and satellite observations is useful for the study of cloud systems -- an important direction in modern cloud physics.

In giving high marks to the CLOUD ATLAS we at the same time wish to raise for discussion the problem of the necessity for continuing work on compilation of a new atlas and a reexamination of cloud classification. In existing international and Soviet classifications there is no division of clouds with respect to conditions of formation (genetic principle). In the morphological classification itself a considerable percentage of the species and varieties cannot be related to any physical processes leading to their formation. Since these species and varieties are also not employed in statistical (climatic) generalizations, the question arises: need they be retained in the CLOUD ATLAS?

BIBLIOGRAPHY

1. ATLAS OBLAKOV (Cloud Atlas), edited by A. Kh. Khrgian and N. I. Novozhilov, Leningrad, Gidrometeoizdat, 1978.

2. ATLAS OBLAKOV (Cloud Atlas), edited by A. Kh. Khrgian, Leningrad, Gidrometeoizdat, 1957.

3. NASTAVLENIYE PO KODAM. T. 1 (DOPOLNENIYE II K TEKHNICHESKOMU REGLAMENTU VMO (Instructions on Codes. Vol. 1 (Supplement II to the Technical Regulations of the WMO)), No 306, Geneva, Switzerland, 1974.

4. INTERNATIONAL CLOUD ATLAS, Vol I, Vol 2, 1956.

REVIEW OF MONOGRAPH BY S. S. LEVKOVSKIY:'WATER RESOURCES OF THE UKRAINE. USE AND CONSERVATION'(VODNYYE RESURSY UKRAINY. ISPOL'ZOVANIYE I OKHRANA), KIEV, "VISHCHA SHKOLA," 1979, 200 PAGES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 117-118

[Article by Doctor of Geographical Sciences A. G. Bulavko]

[Text] Data on the water resources of the Ukraine, their use and conservation, to which the monograph of S. S. Levkovskiy is devoted, have been scattered through many sources, each of which is devoted to some one aspect of this problem, which to a considerable degree complicates their scientific evaluation and practical use. The author, on the basis of his own investigations, creative and critical analysis and generalization of data in the literature, created a multisided work, consisting of three chapters in which he has successfully combined the quantitative and qualitative characteristics of surface and ground water resources and problems relating to their use and conservation. This study is particularly timely for the conditions of the Ukraine in connection with the strain on the water balance and the water shortage in individual regions of the republic.

The first chapter contains a description of the natural conditions for the formation of surface waters, data on their resources and the water balance for the principal rivers and natural zones of the territory and for administrative regions of the republic (and with respect to runoff -- also for economic regions), information on the resources of ground water by natural zones and a description of the intra-annual (by seasons) distribution of the runoff of small, medium and large rivers of the northern part of the republic, the rivers of the Crimea and the southern Ukraine, differentiated over the territory. It also gives an analysis of long-term variations in river runoff and gives an evaluation of its changes under the influence of economic activity. The hydrological information given in the chapter for the first time gives a thorough idea concerning the water resources of the republic, their spatial distribution and temporal dynamics, which is of great scientific and practical importance. Unfortunately, the author has devoted unjustifiably little attention (less than 3% of the volume of the book) to the problem of anthropogenic changes in water resources, which in our time are acquiring increasingly greater importance. Many types of economic activity have not been dealt with at all (irrigation, drainage, water consumption, urbanization) although they exert a definite influence on water resources. There are unexplained discrepancies in the number and areas of natural zones (Tables 3 and 5).

The second chapter is devoted to the use of the water resources of the Dnepr and other rivers in such branches of water management in the republic as water supply, irrigation, watering and drainage of lands, fishing, hydroelectric power, water transportation and log rafting. It gives up-to-date data on the use of waters and concise descriptions of the most important water management systems in each of the enumerated branches. An extremely positive aspect of this chapter is that the author does not limit himself to a statement of the present-day status of the use of water resources, but devotes much attention to measures for improving the water economy and the prospects of its development. For a series of aspects of the considered problem the author has defined and formulated unsolved problems requiring study in both scientific and practical respects. This same chapter briefly examines planned and implemented measures for contending with the harmful effect of water (protective measures against high waters, against erosion, etc.), carried out in combination with the use of waters for other purposes.

The third chapter examines the problem of preservation of water resources. Here there is an analysis of the principal reasons for the exhaustion of water resources and ways to eliminate them. Data are given on the present-day system for the state control of waters and monitoring of their conservation. However, this important problem is considered extremely briefly. In particular, the juridical principles for the use and conservation of waters are set forth fleetingly, in abbreviated form, although they merit greater attention.

In general, the reviewed monograph, written at a high scientific level and well printed and bound, gives a thorough idea concerning the water resources of the Ukraine, their use and conservation, and despite individual shortcomings is an important contribution to the literature on the water resources of our motherland. It contains much which is useful to scientists and practical workers concerned with the study, use and conservation of water resources.

I hope that other similar studies devoted to the water resources of different regions of our country will make their appearance.

HIGH AWARD TO YURIY ANTONIYEVICH IZRAEL

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 p 119

[Unsigned]

[Text] Corresponding Member USSR Academy of Sciences Yuriy Antoniyevich Izrael -- Chairman of the USSR State Committee on Hydrometeorology and Monitoring of the Environment -- has been awarded the Order of the October Revolution by the Presidium of the USSR Supreme Soviet for his services to the Soviet state in the field of hydrometeorology and monitoring of the environment and in connection with his fiftieth birthday.

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SIXTIETH BIRTHDAY OF LEONID TIKHONOVICH MATVEYEV

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 119-120

[Article by personnel of the Military Engineering Institute imeni A. F. Mozhayskiy and Leningrad Hydrometeorological Institute]

[Text] Doctor of Physical and Mathematical Sciences Professor Leonid Tikhonovich Matveyev marked his 60th birthday on 18 August.

Beginning his education in the Physics and Mathematics Faculty of Moscow State Pedagogic Institute imeni V. I. Lenin, he completed his training at the Higher Military Hydrometeorological Institute where he attended courses of lectures of such well-known scientists as S. P. Khromov, V. A. Bugayev, V. F. Bonchkovskiy, V. A. Belinskiy and A. Kh. Khrgian. The first scientific article of L. T. Matveyev was written during the period of practical training in the laboratories of I. A. Kibel' and Ye. N. Blinova (Central Institute of Forecasts). In 1947 he defended his Candidate's dissertation and in that same year headed the Department of General Meteorology and Climatology of the Military Hydrometeorological Faculty of the Soviet Army. During the period from 1956 through 1975 Leonid Tikhonovich headed the department and associated laboratory at the Leningrad Military Engineering

Academy imeni A. F. Mozhayskiy, and after 1975 the Department of Climatology and Preservation of the Atmosphere at the Leningrad Hydrometeorological Institute.

The first investigations of L. T. Matveyev were devoted to the peculiarities of structure of the planetary boundary layer of the atmosphere. In these studies he formulated a theory of diurnal variations and altitudinal distribution of wind velocity and turbulence characteristics in the surface and boundary layers of the atmosphere. He developed a method for determining vertical velocity. In comparing the results of computations with observational data use has been made of materials from the first (of his great many later ventures) expedition (Central Asia) in which L. T. Matveyev participated. Then he carried out a series of studies for investigation of the turbulent and thermal regimes in the free atmosphere. In these investigations, on the basis of the concepts of similarity and dimensionality theory there was found to be, in particular, a dependence of both the turbulence coefficient and fluctuating velocity, and also accelerations experienced by aircraft and other flight vehicles on the thermal and dynamic characteristics of the atmosphere. Extensive use was made of materials from a number of air-flight expeditions to different regions of the country (including to the Central Arctic) carried out under the direction of and with the participation of L. T. Matveyev.

The investigations of Leonid Tikhonov in the field of dynamics of formation of large-scale humidity, cloud cover and precipitation fields have received the widest recognition. The method which he developed for analysis and solution of the system of heat and moisture transport in the turbulent atmosphere made it possible to solve a number of problems in the dynamics of clouds and hydrodynamic prediction of the temperature, humidity and cloud cover fields. By means of this method it was possible for the first time to model and theoretically explain the characteristics of formation and structure of stratiform clouds (time of formation, evaluation of the role of different factors, altitude of cloud boundaries and their thickness, vertical profiles of liquid water content, atmospheric moisture content and water reserve in clouds, relationship between the water reserve and the quantity of precipitation, etc.). Much attention (especially during recent years) has been devoted by L. T. Matveyev and his students to the modeling of the conditions for the formation of low clouds and fogs in the atmospheric boundary layer and in frontal zones. The L. T. Matveyev method has been used productively in solving problems of a practical nature.

The results of the many years of study of clouds and fogs were first generalized by Leonid Tikhonovich in his doctoral dissertation (1959), then in a number of chapters in the fundamental textbook entitled FIZIKA ATMOSFERY (Atmospheric Physics) (1965) and in a monograph prepared for publication entitled DINAMIKA OBLAKOV (Cloud Dynamics) which sets forth modern attainments in Soviet and world science in this field. We note in passing that he has published about 120 studies (among them seven textbooks and study aids and three author's certificates for inventions), some of which have been translated into foreign languages.

While heading up different departments, L. T. Matveyev has worked much and productively in the field of training of engineering and scientific-teaching personnel, improvement of instructional methods and the developing of scientific and methodological principles for several lecture courses (in recent years — "Theory of Climate and General Circulation of the Atmosphere" and "Preservation of the Environment") and has also written textbooks and study aids. Under his direction work was completed on more than 20 Candidate's dissertations and not less than 120 diploma studies and projects.

Leonid Tikhonovich combines much scientific and pedagogic work with diversified public activity. In addition to several scientific councils, he is a member of the councils on the problems "Artificial Modification of Hydrometeorological Processes" and "Weather Forecasting" of the Presidium USSR Academy of Sciences and the USSR State Committee on Hydrometeorology, the editorial board of the Hydrometeorological Publishing House (Gidrometeoizdat), etc. During recent years he has headed work on the special council "Preservation and Rational Use of the Atmosphere" in the USSR Ministry of Higher Education system.

L. T. Matveyev was a participant in the defense of Moscow in 1941. His services have been recognized by government awards and also by the award of the honorary title of Meritorious Worker in Science RSFSR.

In Leonid Tikhonovich there is a happy combination of capabilities of a dedicated researcher and highly concerned teacher and instructor of youth. We wish him good health and new creative successes.

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SEVENTIETH BIRTHDAY OF DAVID YAKOVLEVICH SURAZHSKIY

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 120-121

[Article by Yu. F. Ivanov]

[Text] David Yakovlevich Surazhskiy, the oldest scientific specialist of the Scientific Research Institute of Instrument Making of the USSR State Committee on Hydrometeorology, completed his 70th year on 3 July 1980, as well as the 45th year of his scientific and technical activity.

D. Ya. Surazhskiy came to the Hydrometeorological Service from aviation. Graduating in 1934 from the Air Faculty of the Leningrad Institute of Transportation, he worked as a key engineer-designer at the scientific research institute, developing aircraft designs.

Late in 1941 David Yakovlevich was detailed to the Hydrometeorological Service of the Red Army for carrying out a special mission associated with the development of a parachuted automatic radiometeorological station (PARMS -- parashyutnaya avtomaticheskaya radiometeorologicheskaya stantsiya) and its introduction at the fronts of the Great Fatherland War. The PARMS was jettisoned and began to operate in enemy territory and for the first time in meteorological practice regularly transmitted highly important meteorological information so necessary for the parachute units of the air force and partisans. This development work was carried out jointly with A. V. Goreleychenko under the conditions of a besieged Leningrad.

During the years 1942-1952, also in collaboration with A. V. Goreleychenko and other specialists, D. Ya. Surazhskiy developed a series of automatic radiometeorological stations, the ARMS-2, ARMS-3, ARMS-N and a number of others, intended for prolonged operation in uninhabited and inaccessible regions of the Soviet Union and a number of socialist countries. David Yakovlevich directed a number of expeditions to the Far East and to the transpolar regions for the delivery, installation of ARMS and putting them into operation. The ARMS-N station was awarded a large gold medal at the World's Fair in Brussels.

For his work on creation of the ARMS, which came into use in the Arctic and at the fronts of the Great Fatherland War, D. Ya. Surazhskiy in 1946 was awarded the high title of laureate of the USSR State Prize.

In 1954 D. Ya. Surazhskiy defended his dissertation for the scientific degree of Candidate of Technical Sciences. The dissertation was the result of much scientific research work in the field of wind-measuring technology. Among the instruments and apparatus which were created and introduced in the network of the Hydrometeorological Service with the participation of D. Ya. Surazhskiy were the M-63, M-63M and M-27 wind gages. It is interesting to note that the M-27 wind-measuring system installed in the Kazakhstan region registered winds of hurricane force with a wind speed of 96 m/sec.

During 1954-1972 David Yakovlevich headed the ARMS and meteorological instruments laboratory of the Scientific Research Institute of Hydrometeorological Instrument Making. The ARMS M-107 created under his direction and with his direct participation is used extensively in the observation network of the State Committee on Hydrometeorology for obtaining meteorological information on inaccessible regions.

D. Ya. Surazhskiy is the author of more than 25 scientific studies, 15 inventions and two monographs.

One of the recent studies of D. Ya. Surazhskiy is a new programmed self-triggering device for the M-107 station, constructed using an electronic quartz clock and ensuring precise adherence to the daily schedule for the transmission of meteorological information by the station.

A constant search for and the introduction of everything which is new in different fields of science and technology into the field of meteorological instrument making can be regarded as distinguishing characteristics of the creativity of D. Surazhskiy.

The communist D. Ya. Surazhskiy for many years was a member of the Party Bureau and a leading propagandist of the institute. His successes in work have been rewarded by government awards and also by a gold medal from the All-Union Exhibition of Achievements in the National Economy.

Those who know David Yakovlevich note his love of work, modesty, sensitivity to comrades and his great love of life.

Working at the Scientific Research Institute of Instrument Making from the day of its founding in 1942, he devotes many efforts to the education of youth, for whom his experience and knowledge are very valuable. His organizational capabilities and high qualifications contribute greatly to the goals of hydrometeorological instrument making.

We sincerely wish David Yakovlevich new and significant successes in this noble undertaking.

SIXTIETH BIRTHDAY OF VASILIY MIKHAYLOVICH PASETSKIY

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 121-122

[Article by Ye. P. Borisenkov, V. F. Loginov and N. I. Novozhilov]

[Text] Doctor of Historical Sciences Vasiliy Mikhaylovich Pasetskiy, senior scientific specialist at the Main Geophysical Observatory imeni A. I. Voyeykov, marked his 60th birthday on 15 June.

In the field of the history of the geophysical sciences, for the most part on the history of meteorology, he has written 25 books, brochures and collections of articles and more than 40 newspaper articles. In these publications he writes not only as a historiographer, but also as an author on the methodology of the history of science.

Being a participant in the Great Fatherland War from its first to its last days, after demobilization he worked for several years as head of the editorial office "Biblioteka Poeta" in the Leningrad Division of the Publishing House "Sovetskiy Pisatel'." In 1949 he was called to the Arctic Institute (Arctic and Antarctic Scientific Research Institute), where he soon was named scientific secretary of the institute. In addition to a great volume of scientific organizational work, after 1953 he began to carry out systematic investigations of the history of polar investigations. His first book, VLADIMIR RUSANOV, appeared in 1955; it went through several editions and served as a basis for a Candidate's dissertation. During his 24 years of work at the Arctic Institute V. M. Pasetskiy concentrated his main attention on work on the problems involved in the history of study of the Arctic in the 19th and early 20th century. At the same time he wrote a series of scientific biographies of outstanding polar researchers and explorers of the 16th-20th centuries, including Barents, Bering, Sarychev, Kruzenshtern, Vrangel', Reyneke, Litke, Pakhtusov, Toll', Nansen, Nordenshel'd, and others.

In 1973 V. M. Pasetskiy was transferred to the Main Geophysical Observatory imeni A. I. Voyeykov, where he was assigned work on writing the history of geophysical investigations in our country.

During his stay at the Main Geophysical Observatory he developed a methodology for compiling a history of meteorology of past centuries.

Prior to the beginning of the 19th century meteorology developed for the most part as a geographical science. Its establishment was favored to more than a small degree by meteorological observations carried out during the time of expeditions and travels into the remote expanses of the ocean and into the polar regions. A geographical nature of the science also persisted into the later years and the data of expeditionary meteorological observations as before remained the basic data then making it possible to proceed to global generalizations. However, all these valuable materials either remained for a long time in the archives or (if historians knew about them) were treated extremely incompletely in historical descriptions, without fixing their place amidst existing data, without taking into account the tendency for the sciences to be interrelated. Meteorology, for an objective evaluation of its role in the progress of knowledge, had no such thorough treatment of the history of its formation. In marking the birthday of V. M. Pasetskiy it can be said without exaggeration that the noted gap in the history of science was to a considerable extent filled by him.

In many studies he was the first to describe observations remaining unknown to the reader at large. These included the observations made in the 17th century at the court of Tsar Aleksey Mikhaylovich (the first observations in Russia), the observations of Peter I on ships, the observations made by exiled Dekabrists in Siberia and by participants in the Great Northern (Second Kamchatkan) Expedition in the middle of the 18th century, as well as observations on the ships of Kruzenshtern, Golovin, Lazarev, Litke and Vrangel'. In the logs of ships of the Russian fleet for the 18th century and the first half of the 19th century he discovered several million entries concerning the weather.

In a number of books by V. M. Pasetskiy he generalized the history of development of meteorology in Russia from the most ancient times to the beginning of the 20th century, and in one of his last books — the history of the universally known geophysical institution of the country — the Main Physical Observatory, now the Main Geophysical Observatory.

In his description of the arctic travels of Barents, De Long, Toll', Nansen, Nordenshel'd, Rusanov, and others, and also the flight of Andre, in addition to a general history of geographic discoveries he recreated the history of the meteorological study of the Arctic. It became obvious that ideas concerning the climate of this inaccessible teritory had formed long prior to the establishment of regular observations there. During recent years V. M. Pasetskiy has devoted great attention to studies of the climate of the past based on historical data.

Vasiliy Mikhaylovich is meeting his birthday at the height of his creative powers. We wish him further successes!

SIXTIETH BIRTHDAY OF LEV GRIGOR'YEVICH KACHURIN

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 122-123

[Article by P. I. Smirnov, V. I. Vorob'yev and L. T. Matveyev]

[Text] Professor Lev Grigor'yevich Kachurin, Doctor of Physical and Mathematical Sciences, leading Soviet scientist and teacher, outstanding specialist in a number of aspects of physics of the atmosphere, head of the Department of Experimental Physics of the Atmosphere at the Leningrad Hydrometeorological Institute, marked his 60th birthday on 21 June 1980.

All the scientific and teaching activity of L. G. Kachurin has been associated with the Leningrad Hydrometeorological Institute, where he arrived in 1947 after graduating from Leningrad University. Two years later he defended his Candidate's dissertation and in 1962 became a Doctor of Sciences. During the period 1955-1957 he was dean of the faculty. In 1957 a new department was created and he has been its permanent head since then. The department is now a cohesive body of highly qualified specialists carrying out much teaching and scientific work. Most of the specialists in the department are the students and followers of L. G. Kachurin. More than 30 Doctors and Candidates of Science have been trained in the department. They are successfully working in our country and abroad.

L. G. Kachurin has carried out fundamental scientific investigations in the field of physics of the atmosphere. The theory of irreversible phase transitions of water in the atmosphere which he developed for the first time made it possible to explain a number of important aspects of cloud physics and related sciences, in particular, the intense supercooling of disperse systems, the kinetics of condensation processes, electrokinetic phenomena at a moving phase discontinuity, electrification in aerosol fluxes, etc.

His work on the modification of atmospheric processes acquired great reknown. L. G. Kachurin merits great praise for his creation of a new scientific direction associated with active and passive radar observation of thunderstorm clouds and other natural objects. The phenomenon of prethunderstorm radioemission which he predicted is used in the analysis of thunderstorm clouds both in the course of their natural development and in experiments in controlling their intensity.

The development of automatic aerophysical apparatus carried out by L. G. Kachurin and his co-workers is widely known. Such apparatus was used in outfitting the first high-altitude meteorological observatory in the system for the meteorological support of aviation, in designs of anti-fading apparatus, in equipment employed in the modification of atmospheric processes.

L. G. Kachurin is the author of more than 120 scientific studies, 20 inventions and six books, four of which have been republished both in the USSR and abroad. He is a member of a number of national and international special committees. He has repeatedly presented reports at international symposia and conferences and given lectures at foreign universities. Several generations of students in many countries have already studied from his textbooks.

Lev Grigor'yevich is a veteran of the Great Fatherland War. He departed for the front in July 1941 from the schoolroom. He participated in the defense of Stalingrad as the commander of a tank destroyer battery. He received government awards for his military services.

A talented scientist and teacher, a demanding leader, Lev Grigor'yevich Kachurin comes to this birthday youthfully involved with new ideas and plans, full of creative forces and daring.

We wish him successes from our hearts!

CONFERENCES, MEETINGS AND SEMINARS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 123-126

[Article by B. L. Sokolov, M. Sh. Furman, L. F. Yermakova, Yu. G. Slatinskiy and V. M. Pasetskiy]

[Text] The need for more complete hydrometeorological support of construction of the Baykal-Amur Railroad and the intensively developing national economy in Siberia and the Far East is placing before the organizations of the State Committee on Hydrometeorology the task of carrying out a number of new investigations. These include a study of ice encrustations and ice encrustation processes as a dangerous natural phenomenon exerting a harmful effect on engineering structures, transportation lines, and leading to great material losses reckoned in the tens of millions of rubles annually.

In connection with the importance of the problem the State Hydrological Institute and a number of territorial administrations of the State Committee on Hydrometeorology servicing the zone of construction of the Baykal-Amur Railroad have begun planned work on study of ice encrustations and ice encrustation processes in the network of hydrometeorological stations and posts and also by means of expeditionary investigations and aerial photographic surveying under a unified program.

In order to generalize the experience of study of ice encrustations in the station network and the coordination of investigations, on the initiative of the Administration of Hydrometeorological Observatories of the State Committee on Hydrometeorology of the State Committee on Hydrometeorology for the first time there was a conference-seminar on the subject "Study of Ice Encrustations Along the Route of the Baykal-Amur Railroad and in Other Ice Encrustation Regions of the USSR." It was held during the period 9-11 October 1979 at the Irkutsk Administration of the Hydrometeorological Service. The conference was attended by about 60 representatives of institutes of the State Committee on Hydrometeorology and other ministries and departments and also the chiefs and specialists of river stations and zonal hydrometeorological observatories of the Irkutsk Administration of the Hydrometeorological Service. The conference heard 16 reports and communications on the methods and results of study of ice encrustations in the zone of the Baykal-Amur Railroad and other regions of Siberia and the Far East.

B. L. Sokolov (State Hydrological Institute) discussed the principal goals, tasks and methods for investigating ice encrustations and ice encrustation processes. Proceeding on the basis of the needs of practical workers and scientific research, and also work experience, he examined two principal directions in their study: ice encrustations as a dangerous natural phenomenon, exerting a harmful effect on engineering structures, and the role of ice encrustation processes in the formation of the water regime and the resources of surface and ground water.

The methods for carrying out the list of types of observations of ice encrustation phenomena in the station network which he cites are determined by the purposes, technical possibilities and training of the observers.

V. F. Usachev (State Hydrological Institute) examined the results and methodological aspects of study of ice encrustations in the zone of the Baykal-Amur Railroad on the basis of materials from special aerial photographic surveys, the principal requirements in the choice of survey sectors, some procedures in the office processing of materials and obtaining the characteristics of ice-encrusted bodies.

Theoretical investigations of the mechanism of formation of ice encrustations of river waters was the subject of a report by A. N. Chizhov (State Hydrological Institute). The mathematical models of the processes of formation of ice encrustations on rivers developed by the author require the setting up of special observations at stationary posts to ascertain the elements of hydraulics of water flows and the dynamics of growth of river ice proper and ice encrustations. In the future this information will help in making a prediction of ice encrustation phenomena, this being of great practical importance.

A. Ye. Abakumenko and A. M. Chmutov (State Hydrological Institute) reported on experience in preparing a catalogue of ice encrustations in the zone of the Baykal-Amur Railroad for publication. The first number of the catalogue NALEDI BASSEYNA VERKHOV'YEV R. CHARY (Ice Encrustations in the Basin of the Headwaters of the Chary River) was sent for printing and work was completed on compilation of the second number of NALEDI BASSEYNA R. MUI (Ice Encrustations in the Basin of the Muya River). In the near future plans call for the preparation of subsequent numbers of the catalogue of ice encrustations along the route of the Baykal-Amur Railroad from Nizhneangarsk to Tynda.

A report on the results of study of ice encrustations in the zone of the Baykal-Amur Railroad and other regions under the program of the State Hydrological Institute was given by representatives of territorial administrations: Irkutsk (V. V. Kravchenko), Transbaykalian (G. I. Enov), Yakutsk (A. S. Rudnev), Far Eastern (Yu. G. Shikalov), Kolyma (V. N. Dovbyl). They told of the peculiarities of organization of this work in the network of hydrometeorological stations and posts.

A representative of the Irkutsk Geological Administration, V. M. Litvin, told the meeting about the status of work on study of ice encrustations arising under the influence of construction of the railroad in the western segment of the Baykal-Amur line.

The following special reports were also presented: V. V. Kravchenko (Irkutsk Administration of the Hydrometeorological Service) -- "Organization of Observations in Ice Encrustation Polygons" and B. N. Daykin and M. L. Markov (State Hydrological Institute) -- "New Technical Apparatus in Ice Encrustation Investigations." The first report examined problems relating to the organization of detailed observations in ice encrustation polygons for study of the entire complex of physical processes of formation of ice encrustations and their relationship to hydrometeorological conditions, the testing of different observation methods, etc. The first attempt at organization of such an ice encrustation polygon was undertaken by the Irkutsk Administration of the Hydrometeorological Service in the basin of the Uda River (Eastern Sayan). The conferees exhibited lively interest in a communication of representatives of the State Hydrological Institute on the new instruments and technical decisions used by this institute in investigations of ice encrustations in the central sector of the Baykal-Amur Railroad. The authors exhibited diagrams of the design and operating principle of a hydrothermal drill, relatively simple in its construction, developed at the Arctic and Antarctic Scientific Research Institute, which makes it possible to measure the thickness of the ice at a rate of 1-2 m/min, and also a hydrometric flume for registry of the intradiurnal variation of the water yield of ice encrustations, as well as a long-operating automatic recorder of melting of the ice from the surface of ice encrustations.

Participating in the discussions were V. R. Alekseyev (Institute of the Geography of Siberia and the Far East Siberian Department USSR Academy of Sciences), A. N. Leont'yev (State Committee on Hydrometeorology), M. Sh. Furman (Irkutsk Administration of the Hydrometeorological Service), S. M. Bol'shakov (Siberian Central Scientific Research Institute for Construction), V. G. Kondrat'yev (Chita Polytechnic Institute), B. M. Mironov (Giprolestrans Trust, Irkutsk Affiliate), and others. They pointed out the importance of the considered matters for the practical needs of the national economy, as well as the need for standardization of methods for making field observations, preparation and processing of materials on ice encrustation investigations which can be used for the needs of planning, preparation of systematic instructions on making observations of ice encrustations, formulation of long-term investigations of ice encrustations in representative basins, etc. They emphasized the need for organizing large-scale observations of ice encrustations and ice encrustation processes in the network of hydrometeorological stations and posts over the entire territory of Siberia and the Far East.

It was noted that in accordance with a resolution of the State Committee on Hydrometeorology the study of ice encrustation phenomena has been activated in the zone of construction of the Baykal-Amur Railroad. The results of this work have already been used by the State Hydrological Institute, the Siberian Central Scientific Research Institute for Construction, the Khabarovsk Scientific Institute of Railroad Transportation, the Scientific Research Institute of Railroad Transportation, and the All-Union Scientific Research Institute of Road Construction in compilation of the "Recommendations on the Improvement of Construction and More Precise Determination of the Sphere of Application of Anti-Ice Encrustation Measures and Apparatus," intended for use in the planning and construction practice of the Ministry of Transportation Construction and the Ministry of Transportation. However, the status of work on study of ice encrustations and ice encrustation processes by organizations of the State Committee on Hydrometeorology in general does not meet the modern requirements of hydrometeorological support of the national economy in the regions of Siberia and the Far East.

The conference-seminar recommended the adoption of a series of measures for accelerating the development of work by organizations of the State Committee on Hydrometeorology on study of ice-encrustation phenomena, taking into account their importance for the intensively developing national economy in the northern and eastern regions of our country.

The representatives of the territorial administrations of the Hydrometeorological Service participated in a seminar on preparation of a catalogue of ice encrustations, work being done by specialists of the State Hydrological Institute.

B. L. Sokolov and M. Sh. Furman

A scheduled plenary session of the basin section "Indian Ocean and the Southern Seas" of the Scientific Council on the Problem "Study of the Oceans and Seas and Use of Their Resources" of the USSR State Committee on Science and Technology was held on 22 November 1979 in Sevastopol'. The session was attended by representatives of a number of scientific research institutes and other organizations whose activity is related to the sea. Attending the session from the State Committee on Hydrometeorology were specialists of the Sevastopol' Division of the State Oceanographic Institute and the Sea Section of the Crimean Hydrometeorological Observatory. At the session there was discussion of a program for multisided investigations of the Indian Ocean by personnel of the basin section. There was discussion of several scientific reports devoted to the results of studies carried out in the Indian Ocean and in the Mediterranean Sea during recent years.

In discussing the program for research in the Indian Ocean for the coming years G. K. Karatayev (Marine Hydrophysical Institute), A. V. Kovalev (Institute of Biology of Southern Seas), V. A. Bryantsev (Sea of Azov-Black Sea Scientific Research Institute of Fishing and Oceanography) and others noted that the principal objective of the multisided investigations in this region is the formulation of scientific principles for commercial fishing and study of the degree of the effect of natural and anthropogenic factors on the formation of biological productivity of the ocean. In the opinion of the speakers, it is of special interest to study the open regions of the ocean and also underwater banks and other bottom rises. In this connection plans call for the extensive use of the experience in carrying out multisided investigations in the Atlantic Ocean under the POLIMODE program. In addition to a detailed hydrological-hydrochemical survey with occupation of stations approximately each 1°, plans call for development of several special polygons where there will be measurement of eddy formations and study of the influence of synoptic conditions on formation of commercial concentrations of fish. Plans also call for continuing study of topography of the ocean floor for the purpose of detecting new regions of the upwelling of deep waters along the slopes of underwater ridges.

V. A. Khimits (Sea of Azov-Black Sea Scientific Research Institute of Fishing and Oceanography), N. B. Shapiro and G. N. Khristoforov (Marine Hydrophysical Institute Ukrainian Academy of Sciences) reported on some results of investigations carried out in the Indian Ocean during recent years. In particular, it was noted that the most productive zones are associated with those places where there is intensive vertical movement of water and unusual "cells" with a high concentration of silicic acid and an increased content of phosphates in the surface layers.

A great volume of work has been carried out in the region of the Gulf of Aden. A good correlation was noted between the distribution of hydrochemical elements and the biomass of phytoplankton. It was established that biogenous substances in great quantities are transported from the Red Sea by a subsurface current. The collected data have been used in computing the balance of chemical substances in the gulf and calculations of primary production are being made.

A. F. Visotskiy (Law Institute Ukrainian Academy of Sciences) told about the results of work of a UN Conference on Marine Law, on the drawing up of a convention on the use of sea expanses, depths and the sea floor.

I. M. Ovchinnikov (Southern Division Institute of Oceanography) discussed the principal results of hydrological investigations in the Mediterranean Sea. He noted that during recent years much work has been done in systematizing extensive observational data for the Mediterranean basin. Two volumes of materials from different symposia and conferences and several specialized collections of articles have been published. In 1976 the Hydrometeorological Publishing House (Gidrometeoizdat) published the monograph GIDROLOGIYA SREDIZEMNOGO MORYA (Hydrology of the Mediterranean Sea). And nevertheless the available data are inadequate for an exhaustive characterization of the hydrometeorological regime of the sea, especially during winter. The speaker noted that during recent years eddy formations have been discovered in the Mediterranean Sea which in their extent are commensurable with mesoscale eddies in the open ocean.

Ye. V. Grel' (Sevastopol' Division State Oceanographic Institute) reported on some results of hydrochemical observations in the Aegean Sea. Each year since 1974, within the framework of the interdepartmental program for study of the Mediterranean basin, the Sevastopol' Division of the State Oceanographic Institute has been carrying out hydrochemical surveys of the entire surface of the Aegean Sea. During this time a total of 450 deep-water stations have been occupied, including 27 multiday stations, and about 5,000 samples were selected for determining the principal characteristics of the qualitative composition of waters. The collected data make it possible to determine more precisely the relationship of water masses in this dynamically complex region and ascertain the nature of the vertical movements of waters and the peculiarities of the biochemical layers most subjected to anthropogenic modification.

Also examined at the session were the fundamental principles of a program for joint multisided investigations of the Mediterranean Sea in the coming years. During the years 1981-1984 plans call for carrying out several quasisynchronous surveys, primarily during winter, in all the main regions of the sea for the purpose of investigating the dynamics of waters and the spatial-temporal variability of the principal hydrological and hydrochemical characteristics, evaluation of turbulent mixing in straits and more precise determination of concentrations of nutritional salts. It was noted that a problem of great interest is study of the processes of formation of deep waters in all the principal basins of the Aegean Sea, since in each of them waters are formed which are completely different from one another with respect to physicochemical composition.

The resolution which was adopted listed a number of priority problems for further improvement in the coordination of activity of members of the basin section, especially in the formulation of plans for investigations in the next five-year period and the carrying out of major multisided expeditions under interdepartmental programs.

L. F. Yermakova and Yu. G. Slatinskiy

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A joint session of the scientific councils of the Main Geophysical Observatory imeni A. I. Voyeykov and the Central Aerological Observatory was held in Leningrad on 30 January, on the 50th anniversary of launching of the world's first radiosonde. It was devoted to the status and prospects of development of methods for aerological sounding of the atmosphere. The session was opened by the director of the Main Geophysical Observatory, Ye. P. Borisenkov. In his address he noted that the launching of the world's first radiosonde not only was a noteworthy event in study of the atmosphere, but also laid a firm basis for a new branch of science -- radiotelemetry, whose development later made it possible to proceed to the exploration of space. Following the example of the USSR, atmospheric radiosonde observations were organized in all countries of the world and are now being made at more than 1,000 stations.

A report entitled "Fiftieth Anniversary of the World's First Soviet Radiosonde" was presented by Ye. S. Seleznev, who dealt in detail with the history of creation of the world's first radiosonde. As pointed out by the speaker, as early as 1922 appropriations were allocated for developing "a method for transmitting the readings of measuring instruments over a distance for the purpose of obtaining the readings of meteorographs during the time of a flight." This scientific search was assigned to a group of scientists at the Main Geophysical Observatory, headed by P. A. Molchanov, an outstanding specialist in the field of aerology and aerial navigation.

P. A. Molchanov, who beginning in 1919 headed the aerological observatory, and who in 1923 proceeded to the development of a new method for sounding the atmosphere, successfully completed work on this project in late 1929. The launching of the world's first radiosonde took place on 30 January 1930. It was intended for measurements of the physical parameters in the high layers of the atmosphere and their transmission by radio.

The first radiosonde rose to an altitude of 9 km and reached the stratosphere. The invention of the radiosonde opened up a new stage in development of the atmospheric sciences.

A report by Academician S. N. Vernov and G. I. Golyshev, entitled "The P. A. Molchanov Radiosonde and the Development of Space Research," was heard with great attention.

It was noted in the report that the principle for the P. A. Molchanov radiosonde served for later development of other automatic instruments and apparatus, including a special stratospheric balloon for the investigation of cosmic rays. In addition, the P. A. Molchanov radiosonde was the prototype for future automatic instruments transmitting by radio information on the state of the medium over great distances. The development of this method found its reflection in transmissions of geophysical information from space and from the surface of other planets. The report included a detailed discussion of the development of methods for investigating cosmic rays, including geophysical rockets and satellites.

The present-day status and prospects for development of aerological investigations was the subject of a report by A. A. Chernikov. He noted that after the universally known Pavlovskaya Observatory was blown up by the occupying forces the tasks of the aerological center of our country were assigned to the Central Aerological Observatory. During the almost four decades of its activity the Central Aerological Observatory made an outstanding contribution to study of the high layers of the atmosphere. Its scientists worthily continued the work begun by Molchanov and his students. The scientists of the Central Aerological Observatory created more perfect types of radiosondes, modern instruments for rocket sounding of the atmosphere and unique instruments for meteorological satellites, whose use opened up a new era in study of the high layers of the atmosphere.

The session was read telegrams of welcome from the President of the USSR Academy of Sciences, Academician A. P. Aleksandrov, the Chairman of the State Committee on Hydrometeorology, Corresponding Member USSR Academy of Sciences Yu. A. Izrael, and also from scientific organizations in Moscow and Leningrad.

For the purpose of commemorating an outstanding event in the history of Soviet science -- launching of the world's first radiosonde -- a resolution was adopted on the dedication of a memorial obelisk in the form of a granite shaft at Pavlovsk, on the grounds of the universally known Pavlovskaya Observatory. On 30 January 1980 at Pavlovsk there was a symbolic placement of a memorial plaque which was attended by the scientists of Moscow and Leningrad.

V. M. Pasetskiy

NOTES FROM ABROAD

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 pp 126-127

[Article by B. I. Silkin]

[Text] As reported in SCIENCE NEWS, Vol 115, No 20, 1979, specialists at the Livermore (Lawrence) Laboratory in California, J. W. Chan, W. H. Dewer and D. J. Webbles have proposed a new method for modeling the ozone content in the atmosphere using data on the influence of nuclear explosions set off in air space in the 1950's and the 1960's as a "control."

According to the hypotheses made initially, the nitrogen oxides released in such explosions should lead to a decrease in the ozone content in the atmosphere and the maximum of such a decrease should be observed in 1963. However, observational data have by no means confirmed this: the ozone content varied in the usual range ±4% annually. Thus, any mathematical model which shows more than a 4% ozone concentration about 1963 must be regarded as erroneous.

During recent years, as our understanding of the chemical reactions transpiring in the atmosphere has increased and the corresponding information has been fed into an electronic computer, the capacity of the developed models to reproduce the real effects caused by nuclear explosions has increased.

For example, a model which was used in 1973 (it led to a prediction that supersonic aviation should cause a decrease in the ozone content) predicted a 10% decrease of ozone as a result of setting off of a series of nuclear bombs. During 1973-1974 there was found to be several chemical reactions in which NO_x participate and data on them were taken into account in the modeling. By 1976 the same was done with respect to reactions with the participation of atmospheric chlorine (as a result of which the first evidence was obtained on the influence of fluorocarbons on the ozonosphere) and also the intensity of the reactions between OH and HO_2.

As a result of all this there was a reevaluation in the direction of a decrease in the effect of the influence of explosions on ozone by 1963; however, it was nevertheless assumed to be greater than 5%. In 1977 specialists decided that the intensity of the reactions in which NO and HO_2 participate must be regarded as 40 times greater than was assumed earlier. This very decisively changed the evaluation of the influence exerted on ozone by supersonic aircraft and nuclear explosions (both lead to the release of NO).

As a result of such a re-examination the estimated decrease in the quantity of ozone in 1963 was reduced to 2%, which is a value indistinguishable against the background of natural variations. According to the most recent estimates, taking into account the constants computed in 1978, it can even be assumed with some caution that in 1963, as a result of explosions, the ozone content increased somewhat.

As reported in SCIENCE NEWS, Vol 115, No 24, 1979, a group of NASA specialists in the United States carried out the launching of special meteorological pilot balloons in the Palestine polygon in Texas. These carried laser radiometers to an altitude of about 40 km above sea level. These were used in carrying out measurements of solar IR radiation at sunset and in making determinations of the content of oxychloride in the earth's ozonosphere.

R. Menzies, a specialist at the Jet Propulsion Laboratory at Pasadena, California, presented an analysis of the results of these experiments in June 1979 in Washington at a Conference on Laser Engineering. A study of the collected data led him to somewhat unexpected results. It was found that the content of oxychloride, constituting a by-product of fluorocarbons, forming part of many everyday aerosols, exceeds by a factor of two or even three the value postulated up to this time.

According to existing concepts, oxychloride, constituting a component of the fluorocarbons, is filtered through a layer of atmospheric ozone. Under the influence of solar UV radiation the fluorocarbons are photodissociated, in the process setting free the oxychloride which descends into the ozonosphere layer. There it acts on ozone molecules and as a result oxygen molecules are formed.

According to the conclusions drawn by Menzies, the oxychloride concentration, which it is now evident attains 2-3 parts per million, should be extremely uniform at the scale of the entire earth. And nevertheless, despite the extreme activity of oxychloride (one of its molecules, according to the calculations made by Menzies, is capable of destroying 1,000 ozone molecules), the ozone content in the high layers of the atmosphere is not decreasing. It can be postulated that the ozonosphere has a capacity for recovery which is greater than has been assumed.

As reported in SCIENCE NEWS, Vol 115, No 24, attempts to relate solar activity, expressed in spot formation, to weather phenomena on the earth have been made repeatedly, but without special success. Speaking at the annual meeting of the American Geophysical Union, held in Washington in June 1979, D. W. Hoyt, a specialist with NASA in the United States, proposed a new approach to this problem.

He established that there is a statistical correlation between the structure of sunspots and weather on the earth. According to his conclusions, the ratio of the umbra at the center of the spot to the brighter region of the penumbra surrounding it has a definite correlation with the earth's temperature. The analysis was based on a study of data on the ratio of the umbra to the penumbra collected by the Greenwich Observatory during the period between 1881 and 1970 and on their comparison with the mean annual temperatures in the northern hemisphere during this same period.

There was found to be a gradual increase in both the value expressing this ratio and the temperatures during the period from 1881 to 1932. Between 1932 and 1970, when the ratio of the umbra to the penumbra decreased, the mean annual temperatures in the northern hemisphere also experienced a tendency to a decrease.

In developing these investigations, R. A. Sikuig, a scientific specialist at the national center for study of the atmosphere in Boulder, Colorado, compared the data published by D. W. Hoyt with data obtained on mean annual temperatures for the earth as a whole according to a model prepared using an electronic computer. It was found that the mean planetary temperatures react to a change in the ratio of the umbra to the penumbra in sunspots the same as the temperature of the northern hemisphere.

If these conclusions are confirmed by other researchers, variations in spot-forming activity on the sun may prove to be an important key to a precise determination of solar-terrestrial relationships, playing an important role in meteorological and climatological processes.

OBITUARY OF VIKTOR MIRONOVICH SKLYAROV (1916-1980)

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 8, Aug 80 p 128

[Article by a group of comrades]

[Text] Viktor Mironovich Sklyarov, one of the veterans of the Hydrometeorological Service, a member of the CPSU and a Candidate of Geographical Sciences, who was on special pension, has died after a severe illness in his 65th year.

He was born in 1916 in Lugansk (Voroshilovgrad). His work activity began in 1932 as an instructor of geography in the middle school. In 1940, after graduating from the courses for polar workers, Viktor Mironovich was sent to a polar station on Vrangel' Island as a senior hydrometeorologist. Thereafter all his activity was inseparately associated with the Hydrometeorological Service. He occupied responsible posts in the Krasnoyarsk and Novosibirsk Administrations of the Hydrometeorological Service and in 1948 was assigned to responsible work at the central offices of the Main Administration of the Hydrometeorological Service where for a long time he was successfully occupied with solution of personnel problems, working successively as head of the section for the training of personnel, deputy chief and chief of the personnel administration, and deputy head of the Main Administration of the Hydrometeorological Service for personnel. In 1960 he was transferred to the post of head of the section on scientific research institutes of the Main Administration of the Hydrometeorological Service.

In 1963 V. M. Sklyarov, at his own request, was sent to Antarctica as head of the aerometeorological detachment of the Soviet Antarctic Expedition and on return from the expedition was named deputy director of the Scientific Research Institute of Aeroclimatology. In 1969 he successfully defended his dissertation for the academic degree of Candidate of Geographical Sciences.

In the years which followed Viktor Mironovich worked in the post of director of the Central High-Altitude Hydrometeorological Observatory, where with the energy and determination characteristic of him he solved problems relating to the development and improvement of meteorological observations and investigations on high structures. After going on pension in 1976 he continued his activity at the Central High-Altitude Hydrometeorological Observatory as chief of the agrometeorological station.

Viktor Mironovich always was characterized by a high sense of responsibility in the work assigned him and an inexhaustible love of work. Being a demanding work director, at the same he was a responsive and sensitive person.

V. M. Sklyarov has written a number of studies in the field of applied climatology and he has written textbooks on meteorology and meteorological observations.

The bright memory of Viktor Mironovich will forever remain in the hearts of his comrades and friends and numerous workers in the system of the State Committee on Hydrometeorology.

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