Reports of the Academy of Sciences of the USSR

1965

SovietRxiv

Abstract

Full Text

Reports of the Academy of Sciences of the USSR

1965. Volume 163, No. 5

GEOPHYSICS

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SOME ASPECTS OF THE INTERPRETATION AND USE OF CLOUDINESS DATA OBTAINED FROM METEOROLOGICAL SATELLITES

(Presented by Academician E. K. Fedorov, January 18, 1965)

Analysis of numerous photographs of cloudiness obtained with the aid of the American TIROS satellites has shown that these photographs can be used successfully to refine the synoptic analysis of weather maps, especially in those regions where the network of meteorological stations is either sparse or entirely absent ([1,2]). In work ([3]) the author' s proposed method was set forth, which makes it possible to compute the field of vertical currents from cloudiness data, and in work ([4]), where the fundamental possibility of computing the field of absolute topography from cloudiness data was shown for the first time, one variant was presented for using these vertical velocities in determining the field of the stream function from the distribution of cloudiness.

The method presented in work ([3]) for determining vertical currents from the distribution of cloudiness will be used below to compute the vertical component of the vorticity of velocity from cloudiness data.

All further considerations will be referred to the region G at the mean level of the atmosphere, outlined in Fig. 1 by a dashed line. To this same region will be referred the distribution of the total amount of cloudiness in points. In doing so, a regular grid 20×16 with step $h = 300$ km will be used.

Let, at successive observation times $t_1, t_2, \dots, t_n$, in the region G the distribution of the total amount of cloudiness in points and the horizontal components of the wind velocity be given, but at the moment $t_n + \tau < t_{n+1}$ only the distribution of cloudiness is given.

Obviously, from the data on the horizontal components of the wind, say at the level AT—500 mb, for the successive times under consideration one can compute the corresponding values of the vorticity of velocity. Introducing a rectangular

Fig. 1. Cloudiness field according to data from the satellite TIROS-7 for 29 VIII 1963.

Figure 1: Fig. 1. Cloudiness field according to data from the satellite TIROS-7 for 29 VIII 1963.

coordinate system XOY (the OX axis directed eastward, and the OY axis northward), let us denote the functions describing the fields of the total amount of cloudiness and the values of vorticity at time t_v ($v = 1, 2, \ldots, n$), respectively, by $S^{(v)}(x, y)$ and $\Omega^{(v)}(x, y)$. In addition, for convenience, let us denote the values of the functions under consideration for the j-th row of the grid used by $S_j^{(v)}(x)$ and $\Omega_j^{(v)}(x)$, respectively ($j = 1, 2, \ldots, 16$), and consider the following expansions:

$$S_j^{(v)}(x_i) = \frac{s_{j0}^{(v)}}{2} + \sum_m s_{jm}^{(v)} \cos \frac{m\pi x_i}{19h}, \tag{1}$$

$$\Omega_j^{(v)}(x_i) = \frac{\delta_{j0}^{(v)}}{2} + \sum_m \delta_{jm}^{(v)} \cos \frac{m\pi x_i}{16h},$$

where $i = 1, 2, \ldots, 20$.

As in work ([3]), let us assume that between the Fourier coefficients of the cloudiness fields and the vorticity of velocity there exists a linear relation

$$\delta_{jm}^{(v)} = \alpha_{jm} s_{jm}^{(v)} + \beta_{jm}, \tag{2}$$

where α_{jm} and β_{jm} are coefficients of linear regression, which must be

are determined from observational data. In the general case, when $\nu > 2$, system (2) is an overdetermined system of ν algebraic equations with two unknowns α_{jm} and β_{jm}. By the method of least squares, the corresponding system of normal equations is constructed, from which the regression coefficients α_{jm} and β_{jm} are readily found. When the regression coefficients have been determined, the inverse problem can be solved. To do this, from the cloudiness data it is first necessary to find the Fourier coefficients s_{jm}; then, using (2) and the already found regression coefficients, the Fourier coefficients δ_{jm} for the vorticity are determined; and finally, from the second of formulas (1), the function $\Omega_j(x)$ itself is computed for all rows of the grid being used.

Fig. 1. Cloudiness field according to data from the satellite Tiros-7 for 29 VIII 1963.

In an entirely analogous way one can also compute the planar divergence. Incidentally, the planar divergence can be computed in a rough approximation also by using the vertical motions calculated from cloudiness, by the method set forth in [4].

Let us now denote the vorticity, divergence, and horizontal components of the wind velocity computed from cloudiness respectively by Ω_s, D_s, u_s, and v_s, and consider the following problem. Suppose that inside a region G the functions Ω_s and D_s are given, and on its boundary the values u and v of the horizontal components of the actual wind are given. It is required to determine u_s and v_s inside this region. To solve the stated problem, consider the equalities

$$\partial u_s/\partial x + \partial v_s/\partial y = D_s, \qquad \partial v_s/\partial x - \partial u_s/\partial y = \Omega_s. \tag{3}$$

If we introduce into consideration the vector

$$\vec{\mathfrak{B}}_s = \Omega_s i + D_s j, \tag{4}$$

then, as is easy to see, for the function u_s we obtain Poisson' s equation, written in the form

$$\Delta u_s = \mathrm{rot}_z\, \vec{\mathfrak{B}}_s, \tag{5}$$

where $\mathrm{rot}_z\, \vec{\mathfrak{B}}_s$ is the vertical component of the vector $\mathrm{rot}\, \vec{\mathfrak{B}}_s$.

Equation (5) must be solved with the boundary condition

$$u_s = u \quad \text{on the boundary of the region } G. \tag{6}$$

The equation for v_s and the corresponding boundary condition may be written as

$$\Delta v_s = \mathrm{div}\, \vec{\mathfrak{B}}_s; \tag{7}$$

$$v_s = v \quad \text{on the boundary of the region } G. \tag{8}$$

After calculating Ω_s, u_s, and v_s in the region G, the height of the isobaric surface H_s, determined from the cloud-cover field, may be found from the balance equation (provided the sought function is specified on the boundary of the region G), written in the form

$$g\Delta H_s = l\Omega_s - \left[(\partial u_s/\partial x)^2 + (\partial v_s/\partial y)^2 + 2(\partial u_s/\partial y)(\partial v_s/\partial x)\right] - \beta u_s, \tag{9}$$

where $\beta = dl/dy$, l is the Coriolis parameter.

Let us note that, since $\partial\Omega_s/\partial y \gg \partial D_s/\partial x$ and $\partial\Omega_s/\partial x \gg \partial D_s/\partial y$, the terms containing D_s in the right-hand sides of equations (5) and (7) may be excluded from consideration.

In carrying out calculations to determine the function Ω_s, instead of the velocity vorticity Ω the geostrophic vorticity Ω_g, computed from the AT−500 mb chart, was used:

$$\Omega_g = \frac{g}{l}\Delta H, \tag{10}$$

where H is the height of the isobaric surface, g is the acceleration due to gravity. Let the vorticity computed in this way from cloudiness be denoted by Ω_{sg}. Then the function H_s is determined from the equation

$$\Delta H_s = F(x, y), \tag{11}$$

where $F = \frac{g}{l}\Omega_{sg}$.

To determine the function H_s in the region G from equation (11), it is necessary to specify the values of this function on the contour of the region under consideration. We write this boundary condition as follows:

$$H_s = H \quad \text{on the boundary of the region } G. \tag{12}$$

To determine the regression coefficients α_{jm} and β_{jm}, a sample of actual data on total cloud amount, in points, for 15:00 on 25, 26, 27, 28, and 29 VIII 1963 and the corresponding data on geostrophic vorticity was used. The cloudiness data for 28 and 29 VIII were compiled on the basis of observations from the TIROS-7 satellite. For all the remaining times, data from ordinary meteorological observations were used. From the regression coefficients found, the function Ω_{sg} was determined both for all times included in the sample and for 30 VIII 1963, from 15:00 cloudiness data that were not included in the sample. The values of the function Ω_{sg} determined in this way were used in calculating the fields of the components of the geostrophic wind and the heights of the AT−500 mb isobaric surface by means of solutions of equations (5), (7), and (11) with boundary conditions (6), (8), and (12).

For convenience of exposition, let us denote the right-hand side of equation (5) by $f(x, y)$, and in the left-hand side omit the index s on the sought function. Then equation (5) is rewritten in the form:

$$\Delta u = f. \tag{13}$$

This equation corresponds to the system of difference equations

Figure 2a

Figure 2: Figure 2a

Figure 2b

Figure 3: Figure 2b

$$u_{i+1,j} + u_{i-1,j} + u_{i,j+1} + u_{i,j-1} - 4u_{ij} \neq f_{ij}. \tag{14}$$

The system of equations (14) was solved by the Liebmann relaxation method, assuming that the $(\mu + 1)$-th approximation is expressed through the μ-th approximation by the formula

$$u_{ij}^{\mu+1} = u_{ij}^{\mu} + aR_{ij}^{\mu}. \tag{15}$$

Fig. 2. a –actual 500-mb absolute topography (AT) chart and geostrophic wind chart for 15:00, 29 VIII 1963, compiled from the results of objective analysis (according to the scheme of the World Meteorological Centre); **b** –500-mb AT chart and geostrophic wind chart calculated from cloudiness.

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Here α is the relaxation coefficient, R_{ij}^{ν} is the residual, computed by the formula

$$R_{ij}^{\nu} = u_{i+1,j}^{\nu} + u_{i,j+1}^{\nu} + u_{i-1,j}^{\nu+1} + u_{i,j-1}^{\nu+1} - 4u_{ij}^{\nu} - f_{ij}. \tag{16}$$

In solving the problem, $u_{ij}^0 = 0$ was taken as the zero approximation, and the value of the relaxation coefficient was $\alpha = 0.4$. In this case it was required to ensure an accuracy $\varepsilon = 1$ m/sec. In determining the functions v_s and H_s from equations (7) and (11), accuracies of 1 m/sec and 1 dkm, respectively, were required.

We present the results of the calculation and their analysis for 29 VIII 1963. Figure 1 presents the results of a nephanalysis of four orbits (1049, 1050, 1051, 1052) of the Tiros-7 satellite on 29 VIII 1963, obtained from the data of the corresponding telegrams using the instructions and system of notation adopted at the Central Institute of Forecasts of the Main Administration of the Hydrometeorological Service. The cloudiness information obtained for the indicated four orbits did not cover the entire gridded region–the southwestern and northeastern corners of this region were filled with data on total cloud amount from ordinary meteorological observations at 15 h on 29 VIII 1963. Orbit 1051, whose time was 15 h 32 min, is most suitable for this time. The other three orbits, 1049, 1050, and 1052, correspond to 12 h 12 min, 13 h 56 min, and 17 h 14 min, respectively. Thus, the map shown in Fig. 1 was assembled from data from four orbits and, consequently, the cloudiness data under consideration were obtained as a result of nonsynchronous observations. However, we used one of the most important properties of cloud fields—the property of conservativeness (4)—and referred the data from all four orbits to 15 h.

Figure 2 presents the actual (Fig. 2a) and cloudiness-computed (Fig. 2b) maps of AT—500 mb and the geostrophic wind. Comparing these two maps, it is easy to see that the computed and actual AT—500 mb fields are, in general terms, in good agreement. The mean absolute error for the entire map is $\overline{|\delta|} =$ 4 dkm, and the absolute error is $|\Delta| = 13$ dkm. The results of the calculation (including those shown in Fig. 2) showed that, on the computed maps, the geopotential values in the centers of anticyclones are lower than the actual ones. This conclusion agrees with the results obtained earlier by A. I. Burtsev and I. P. Vetlov when computing the geopotential field from the actual wind (5).

As for the fields of geostrophic wind and wind computed from cloudiness, as can be seen in Fig. 2, the largest deviations in wind speed and direction are observed in regions of weak winds or in those regions where the baric formations computed from cloudiness turned out to be displaced relative to their actual positions. The best agreement of the computed wind with the geostrophic wind occurs in the zone of strong winds. For the computed fields u_s and v_s, from which the wind field shown in the indicated figure was constructed, the following estimates were obtained: 1) for the map u_s, $\overline{|\delta|} = 5$ m/sec, $|\Delta| = 17$ m/sec, correlation coefficient $r = 0.69$; 2) for the map v_s, $\overline{|\delta|} = 3$ m/sec, $|\Delta| = 14$ m/sec, $r = 0.91$. For the maps H_s, u_s, and v_s computed from cloudiness data for 30 VIII 1963 (not included in the sample), the following estimates were obtained: 1) for the map H_s, $\overline{|\delta|} = 5$ dkm, $|\Delta| = 16$ dkm; 2) for the map u_s, $\overline{|\delta|} = 9$ m/sec, $|\Delta| = 28$ m/sec, $r = 0.33$; 3) for the map v_s, $\overline{|\delta|} = 7$ m/sec, $|\Delta| = 27$ m/sec, $r = 0.65$.

The problem was implemented on the computer of the World Meteorological Center by N. S. Khazizova.

World Meteorological Center

Received
12 I 1965

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Note: Figure translations are in progress. See original paper for figures.