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TWENTY-FIFTH ANNIVERSARY OF SOVIET RESEARCH IN ANTARCTICA

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 5-12

[Unsigned article]

[Text]

Abstract: The article reviews the principal results of the work of Soviet Antarctic expeditions during the period 1976-1980.

January 1981 marked the 25th anniversary from the beginning of the work of Soviet scientists in Antarctica. The materials of Soviet Antarctic research are published extensively. The results are reflected in the world's first ATLAS ANTARKTIKI (Atlas of Antarctica), published in the USSR, and in numerous books and articles of our polar scientists in different fields of specialization. In this article we summarize the results of some research work carried out in Antarctica during the Tenth Five-Year Plan.

The five-year plan provided for a broad range of scientific research work on the Antarctic continent and in the waters of the Antarctic Ocean. Participating in this work were organizations of seven departments: USSR Academy of Sciences, State Committee on Hydrometeorology, USSR Geology Ministry, USSR Main Administration of Geodesy and Cartography, USSR Ministry of Fisheries, RSFSR Ministry of Higher and Intermediate Special Education and USSR Health Ministry. All the planned work was implemented and considerable investigations were carried out above the plan. Monographs were written and numerous articles were published in periodic scientific journals.

In a brief article it is impossible to set forth the results of all the investigations and therefore we will discuss only some of the most important results.

An analysis of long-term aerometeorological observations made possible a detailed examination of the spatial-temporal structure of the Antarctic atmosphere and a determination of a number of new characteristics. They considerably broadened existing ideas concerning the regime of the southern polar atmosphere, which made it possible to refine earlier concepts, apply the results in a diagnosis and prediction of atmospheric phenomena and use them as initial information for developing a method for predicting weather for three days in advance.

The atlases, reference aids and monographs on aerometeorology of Antarctica, based on the materials of the Soviet Antarctic Expeditions and foreign expeditions, are a reliable basis for developing a theory of climate, methods for predicting weather for different times in advance, and also for the satisfaction, on a high scientific level, of the diversified needs of national economic organizations, and especially support of Antarctic navigation.

Investigations of general circulation of the atmosphere in Antarctica were based on data from artificial earth satellites. For the first time there was a generalization of material on the frequency of recurrence of moving and stagnating cyclones, the frequency of recurrence of anticyclones, the trajectories of cyclones and the distribution of total cloud cover. These data and the results of their analysis can be used for different scientific research studies, and also in the solution of practical problems, including those of a prognostic nature.

Studies related to investigation of the migration of the circumpolar vortex, the slope of its axis and evaluation of the tropospheric sources of vorticity, made it possible to determine the existence of centers of action of the Antarctic circulation.

An empirical model of air exchange in the system of the southern polar circulation cell was developed. A multisided investigation of the mass of atmospheric air in dynamic transformations of the vortex indicated that in the course of the year it varies in a range ±2%, which is equal to $2.1 \cdot 10^{18}$ g. Such significant variations of mass are not observed over any other region of the earth. An investigation of meridional circulation, regulating air exchange, indicated that as an average for the year a balance of masses is attained over Antarctica with an outflow of air from the continent in the layer earth-600gPa in a quantity $4.4 \cdot 10^{13}$ g/sec and an inflow of masses in the above-lying layers (600-50gPa) in the same quantity. It was estimated quantitatively that the compensating flow of masses is directed downward. In this connection, for the first time there was a detailed investigation of the field of distribution of the velocities of vertical movements over Antarctica in different layers of the atmosphere, the principal characteristic of which is a subsidence of air over Eastern Antarctica and rising over Western Antarctica.

The first estimates of the thermal influence of Antarctica on the atmosphere and ocean were made. It was found that for the lower troposphere of the temperate and high latitudes of the Atlantic-Indian Ocean sector there are negative deviations, whereas for the Pacific Ocean sector there are positive deviations.

The use of the developed empirical model of air transfer over Antarctica made it possible to investigate heat and moisture exchange in the south polar region. The characteristics of interseasonal differences in atmospheric heat content were ascertained. For example, over the central part of Eastern Antarctica the rate of cooling of the troposphere between January and April is $21 \cdot 10^6$ J/m^2, whereas over Western Antarctica it is $17 \cdot 10^6$ J/m^2. The greatest rates of heating (between October and January) attain $30 \cdot 10^6$ J/m^2 and are observed over Western Antarctica. An investigation of the meridional heat flows made it possible to note that in the thin surface layer there is predominance of transport beyond the limits of the continent with a maximum intensity on the shores of Amundsen and Davis Seas ($2.4 \cdot 10^6$ W/m^2). Regions of extremal heat exchange in the entire thickness of the atmosphere across the Antarctic coast were also defined. This made it possible to draw definite conclusions concerning the influence of the Antarctic continent on the climate of contiguous regions and indicate the areas through which this influence is exerted. The region through which heat arrives on the continent was ascertained.

The patterns of formation and change in moisture content were determined in dependence on different geographical and circulation conditions. On the whole, Antarctica, constituting about 2.5% of the earth's total area, annually furnishes about 6% of the excess moisture which the world ocean supplies to the atmosphere.

In connection with the development of the system of global climatic monitoring much work was carried out for organizing a program of observations in Antarctica for determining the content of ozone and minor gas admixtures (carbon dioxide, carbon monoxide, methane, nitrous oxide) and transparency of the entire thickness of the atmosphere in the IR spectral region. Since the southern hemisphere, especially its polar region, at the present time experiences virtually no anthropogenic influence, a base was established at Mirnyy Observatory for monitoring the gas and aerosol composition of the atmosphere. Special observations at this station make it possible to evaluate background contamination and use these data for evaluating the stability of climate and investigations of physics of the atmosphere.

Observations of the mentioned components were also carried out on ships of the Arctic and Antarctic Research Institute in collaboration with other organizations of the USSR Academy of Sciences and the State Committee on Hydrometeorology (Institute of Physics of the Atmosphere and Institute of Experimental Meteorology). Even now the results of investigations in combination with data from earlier spectral measurements of atmospheric composition make it possible to determine the principal characteristics of variability of the carbon dioxide content on a global scale and the nature of variations of nitrogen, methane and other gas components.

The first attempts have been made at evaluating the stability of climate in the south polar region. For example, it was possible to detect a tendency to an increase in temperature at the earth's surface and in the lower troposphere during the last 20 years; at the intracontinental stations the warming was 0.5°C; in the region of the Palmer Peninsula -- up to 4.0°C; and on the coast of East Antarctica -- up to 1.5°C.

During the course of the Tenth Five-Year Plan a study was made of the climatic characteristics of formation of the temperature and pressure fields in the troposphere over Antarctica and the southern hemisphere as a whole. Allowance for macrotransformations of the forms of circulation in the method for preparing weather forecasts for three days made possible a better foreseeing of the nature of change of standard processes in Antarctica. The characteristics of the wind and temperature regimes were obtained by stations, making it possible to prepare three-day forecasts in more specific detail.

The characteristics of development of atmospheric processes in the southern hemisphere and Antarctica for long time periods were studied. Group processes of uniform development of circulation over a six-month period were obtained. The characteristics of such group processes are used for preparing forecasts for up to three months in advance.

The experimental introduction of long-range weather forecasts for the hydrometeorological support of navigation and expeditionary work in Antarctica was continued. The weather forecasts for three days in advance prepared sporadically at the weather bureau of the Molodezhnaya aerometeorological center are quite reliable and are employed by a number of users. Beginning in 1977 specialists have prepared long-range forecasts for the spring and summer (September-December) and autumn (March-May) periods.

The study of the Antarctic Ocean during the period 1976-1980 was carried out on the basis of the long-term large-scale program "Polar Experiment-South" (POLEKS-YuG), providing for the carrying out of a number of in situ experiments in different regions of the Antarctic Ocean. This program was carried out at both the national and international levels, primarily within the framework of Soviet-American cooperation, jointly with the American program "International Investigations of the Antarctic Ocean." In the course of 1975-1979 specialists carried out in situ experiments aboard scientific research ships with the implementation of hydrological surveys, placement of buoy stations for prolonged periods with submergible buoys and series of current meters, and investigation of frontal zones with the use of thermohaline probes in the regions of Drake Strait, Scotia Sea, in the ocean areas between Africa and Antarctica, Australia and Antarctica.

As a result of implementation of the "POLEKS-YuG" program, on the basis of instrumental measurements it was possible to establish the vertical structure of the Antarctic Circumpolar Current (ACC), characterized by the penetration of the current to deep horizons, virtually to the bottom of the ocean, with a monotonic decrease of current velocities from 50-100 cm/sec in the surface layers to 5-10 cm/sec in the deep layers.

For the first time on the basis of instrumental measurements it was possible to estimate the transport of waters in the ACC system, being 120-130 sverdrup in Drake Strait, 190-200 sverdrup in the region between Africa and Antarctica, and 160-170 sverdrup in the region between Australia and Antarctica.

The principal scales of temporal variability of currents in the ACC system were established. The mesoscale variability is characterized by the existence of tidal (semidiurnal and diurnal) and inertial fluctuations with periods from 13 to 16 hours.

Fluctuations with periods of 5-6 days were detected in the synoptic region of the spectrum. These represent barotropic waves arising as a result of the passage of macroscale pressure formations over the Antarctic Ocean; these waves are propagated to great depths. Most of the synoptic variability is accounted for by fluctuations with periods of about 14 days and more than 30 days. The appearance of these fluctuations is associated with the influence of long-period tides, but primarily with the origin of frontal eddies forming in the ACC zone as a result of baroclinic instability. Data from instrumental observations were used in establishing the principal parameters of frontal eddies in different regions of the Antarctic Ocean.

On the basis of long series of observations of currents it was possible to evaluate the scales of seasonal (semiannual and annual) fluctuations in the velocity in the ACC system.

Detailed surveys in the region of the polar frontal zone with the use of thermohaline probes made it possible to establish the boundaries of the polar frontal zone from the characterization of the vertical distribution of temperature and salinity, investigate the vertical structure of the polar frontal zone and establish the quasistationary character of its position.

Due to the carrying out of aerological sounding on ships during the period of in situ experiments and collection of aerometeorological information from ground stations surrounding the polygons it was possible to ascertain the principal zones of influence of the Antarctic Ocean on processes in the atmosphere situated to the north of the polar frontal zone in the neighborhood of Drake Strait, to the south of Africa and in the Australia region. Preliminary estimates of heat exchange between the ocean and the atmosphere were obtained.

In the future provision is being made for carrying out a macroscale in situ experiment in the Australian-New Zealand sector of the Antarctic Ocean by personnel on board ships of the Arctic and Antarctic Scientific Research Institute, Far Eastern Scientific Research Institute and Pacific Ocean Institute of Fishing and Oceanography in the southern summer of 1980-1981, implementation of the Soviet-American experiment "POLYN'YA" (Polynia) aboard the scientific expeditionary ship "Mikhail Somov" in the region of upwelling of waters during the winter-autumn period of 1981, investigation of the zone of merging of the waters of the Weddell and Scotia Seas, and the carrying out of experiments in energy-active zones of the Antarctic Ocean.

One of the directions in scientific research is related directly to study of the influence of the ice regime on navigation in the waters of Antarctica. A result of these investigations was the publication of an individual volume of Transactions of the Soviet Antarctic Expedition -- LEDOVYYE USLOVIYA PLAVANIYA V VODAKH ANTARKTIKA (NAUCHNOYE OBOSNOVANIYE I REKOMENDATSII) (Ice Conditions for Navigation in the Waters of Antarctica (Scientific Basis and Recommendations)) and a whole series of scientific studies. A further development of this theme was the development of studies of the hydrometeorological conditions of navigation, optimum times and methods for carrying out loading operations in the roadsteads of Antarctic stations and allowance for hydrometeorological and ice factors in the planning of navigation. The results of this already completed work can be used as a reference aid for navigators, scientific-operational workers, planning and directing agencies of different ministries and departments implementing voyages in Antarctic waters. A noteworthy result of the investigation carried out is the creation of the ATLAS LEDOVYKH KART (Atlas of Ice Maps), which at the present time is being prepared for publication. It contains 60 maps which show the principal features and characteristics of the ice regime of Antarctica during the entire period of the investigations of the Soviet Antarctic Expedition. The Program for Computing the Schedule of Movement of Ships of the Soviet Antarctic Expedition on an Electronic Computer is also to find practical application in the planning of activity of the Soviet Antarctic Expedition. The results of investigations in this scientific direction immediately found practical application in the form of consultations, responses to requests, dispatch of cartographic and other material in accordance with the continuing needs of the Soviet Antarctic Expedition, USSR Ministry of Fisheries and other organizations and departments whose activity is related to voyages to Antarctica.

The study of the glacial cover of Antarctica as the principal component of the Antarctic landscape was carried out primarily within the framework of the International Antarctic Glaciological Project. Practical work was also carried out in the field of engineering glaciology. The latest research methods were used, with employment of modern technical equipment.

On the basis of data from large-scale radar measurements of thickness of the glacial cover it was possible to compile maps of the relief of the bed beneath the ice and the thickness of the glacial cover and on their basis, for the first time with sufficient reliability, it was possible to determine the volume of ice in the glacier: $24.9 \cdot 10^6$ km^3.

The development of methods and equipment for the electrothermal drilling of ice, by means of which about 5 km of boreholes were drilled in the glacial cover with the continuous removal of the ice core, as well as improvement in geophysical and radiochemical methods for studying boreholes and the ice samples extracted from them, created the prerequisites for successful development of the method of paleoclimatic investigations. The first important result of investigations in this direction was the conclusion that there was a synchronicity of the main climatic boundaries in both of the earth's hemispheres.

Glacioclimatic investigations of the energy interaction of the ice cover with the atmosphere had as a result the implementation of a quantitative estimate of the total losses of radiant energy by the glacial cover, annually constituting $3.3 \cdot 10^{22}$ J and characterizing Antarctica as an extensive region of heat loss in the global climatic system.

Many years of practical investigations in the field of engineering glaciology were completed with the formulation of the theoretical principles and implementation of a complex of measures for creating a snow airdrome in Antarctica for heavy wheeled aircraft. It is characterized as an airdrome on deep snow with a "compressible half-space computation scheme" and has no equals in world airdrome construction. The construction of this airdrome at the Molodezhnaya Meteorological Center and the successful accomplishment of a trial run of an Il-18D aircraft along the route Moscow-Molodezhnaya-Moscow should facilitate organization of a modern system for aviation support of a long-term program of scientific investigations and national economic interests in Antarctica.

Investigation of the marginal part of the glacial cover, represented by unusual ice shores along almost the entire extent of the Antarctic continent, made it possible to clarify the principal patterns of interaction between the glacier and the ocean and to develop a genetic classification of Antarctic shores. An analysis of materials obtained by the Soviet Antarctic Expedition made it possible to detect a tendency to degradation of the glacial cover in the coastal zone, in particular, in the region of the Molodezhnaya Aviation Meteorological Center and Mirnyy Observatory.

The region selected for geological exploration work during the Tenth Five-Year Plan was the Weddell Sea region, characterized by an extensive shelf (about 1.2 million square kilometers), information on whose geological structure was completely lacking. In this region the participants of the Soviet Antarctic Expedition carried out an aeromagnetic survey over an area of almost 1 million square kilometers, a gravimetric survey over an area of 235,000 square kilometers which involved aircraft landings on the snow surface, experimental-methodological seismic investigations, radar measurements of thickness of the glacial cover from the air, and also reconnaissance geological investigations in the mountains bordering on the Weddell Sea.

As a result of the investigations which were made the following facts were established:

1. The studied area, in its extent exceeding the shelf of the North Sea, is a component part of a major sedimentation basin. Within its limits the thickness of the sedimentary cover averages 7-10 km, in the deepest depressions of the basement attaining 12-15 km or more;

2. The basement of the studied part of the Weddell Sea basin consists primarily of Archean crystalline rocks locally covered by Lower Proterozoic epicraton complexes;

3. In the thick sedimentary cover there is hypothetically a predominance of Late Mesozoic-Early Cenozoic, primarily molasse formations;

4. In the first approximation it was possible to detect regional structures of the sedimentary cover potentially promising for petroleum and gas;

5. It was possible to establish an intraplatform nature of all folded systems in the southeastern part of the mountainous margins of the Weddell Sea shelf and it was found that this region belongs to the periphery of the Antarctic craton; a study was made of the structure and peculiarities of evolution of the Precambrian zone of folding, with respect to a number of characteristics comparable with structures of the greenstone zones type.

The special studies "Deep Structure of Mac Robertson Land and Princess Elizabeth Land (Eastern Antarctica)" and "Metamorphic and Magmatic Complexes of Western Antarctica," completed at the beginning of the five-year plan, revealed the largest continental rift zone in Eastern Antarctica and characterized its deep structure; a new scheme of evolution of the processes of metamorphism and granitization during the formation of the crystalline basement of Antarctica was created; the principal stages in the endogenous activity and granite formation in different geotectonic provinces of the continent were defined.

Three special maps of Antarctica were published during the Tenth Five-Year Plan: geological (1:5,000,000), metamorphic formations (1:5,000,000) and tectonic (1:10,000,000). Such a series of maps was compiled for the first time in the history of geological study of Antarctica. The series has received a high evaluation from the international scientific community and the Commission on the Geological Map of the World, which adopted a resolution to prepare on the basis of the Soviet publication a series of similar maps for issuance under the aegis of the commission. This work considerably contributed to strengthening of USSR priority in the study of Antarctica.

A dependence of the probability of appearance of the sporadic E layer in the polar regions on the parameters of the interplanetary magnetic field (IMF) was discovered and investigated in detail. On the basis of this dependence it was possible to develop a method for determining the sign of the vertical and azimuthal components of the IMF using data from vertical sounding of the ionosphere at the polar station Vostok. This method makes it possible to ascertain the IMF parameters in all seasons with an identical effectiveness. At the present time this method is undergoing practical checking. These same data made it possible to advance a new hypothesis concerning the nature of the sporadic E layer in the polar caps, and also a hypothesis that the daytime cusp in the southern hemisphere is situated closer to the pole than in the northern hemisphere.

Considerable geomagnetic investigations were carried out in collaboration with the Institute of Terrestrial Magnetism, Ionosphere and Radio Wave Propagation within the framework of the "Antarctic Polygon" project. New information was obtained on the position and intensity of auroral electrojets with different levels of magnetic activity and the degree of geoeffectiveness of IMP components was evaluated. Specialists have developed and are submitting to practical testing a method for the supershort-range (1-1.5 hours) forecasting of geomagnetic activity for 15-minute sums of the vector of magnetic disturbances using observational data for Vostok station.

The scientific-practical principles for creating a broad network of radio paths for slant sounding of the ionosphere in Antarctica have been formulated. The first stage of this plan is being carried out: data are being collected on the radio link Bellinsgauzen Island - Molodezhnaya with an extent of more than 4,000 km, situated alternately first in the main ionospheric gap and then in the zone of auroral absorption.

The collection of materials and the analysis of these data on the transcontinental path Moscow-Molodezhnaya have continued.

A unique experiment was carried out for measuring the angles of arrival of radio waves (in the vertical and horizontal planes) of one of the transmitters situated in the territory of the USSR.

During the Tenth Five-Year Plan scientific research work continued on study of man's adaptation in different regions of Antarctica. The work was carried out for the most part by the scientific specialists of the Siberian Division, USSR Academy of Medical Sciences (Novosibirsk), the Scientific Research Institute of Experimental Medicine, USSR Academy of Medical Sciences, the Scientific Research Institute of Influenza (Leningrad) and other medical institutions. In addition to scientific specialists, the doctors of the expedition participated in these investigations. Definite work in this field was carried out by scientific specialists of the Polar Medicine Laboratory, Arctic and Antarctic Scientific Research Institute.

A number of monographs and more than 200 articles were published as a result of the joint investigations. These studies give materials characterizing the peculiarities of man's adaptation to extremal environmental conditions. Considerable attention was devoted to the health of the winterers, the characteristics of the course of diseases of polar specialists and the organization of medical-sanitary support of antarctic expeditions.

A number of instructions and recommendations were developed and introduced into the work practice of the Soviet Antarctic Expedition on the basis of the results of scientific investigations of the polar medicine laboratory of the Arctic and Antarctic Scientific Research Institute in collaboration with other scientific research institutes.

The complex of scientific research studies and prophylactic measures carried out led to a decrease in the incidence of disease, accompanied by a temporary loss of work capability among the participants of the Soviet Antarctic Expedition. There has been a systematic reduction in the number of persons returning from Antarctica due to medical indications.

The basis for the material-technical support of expeditionary work during this five-year plan was already laid in earlier years, but for successful implementation of the formulated problems it was necessary to take additional measures. The new scientific research ship "Mikhail Somov," replacing the diesel-electric "Ob'," had entered service by the beginning of the five-year plan and was sent on its first antarctic voyage. Major work was undertaken for reconstructing Mirnyy Observatory, Vostok station, Bellinsgauzen and Novolazarevskaya. There was a considerable augmentation of the number of surface transport vehicles, including new "Khar'kovchanka-2" vehicles. Finally, prolonged experimental work on creation of an airdrome on the glacier surface for heavy aircraft at Molodezhnaya was crowned with success and in the summer of 1979-1980 the first flight was made from Moscow to Molodezhnaya. The new, seventh, permanently operating station Russkaya in Western Antarctica was opened on 9 March 1980. The temporary station Druzhnaya was established early in the five-year plan on the shores of the Weddell Sea for supporting seasonal investigations in this region. Scientific equipment and instrumentation were considerably supplemented and renewed.

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USE OF ANALOGUES FOR EVALUATING PREDICTABILITY AND LONG-RANGE FORECASTING OF MEAN MONTHLY AIR TEMPERATURE FIELDS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 13-22

[Article by G. V. Gruza, professor, and E. Ya. Ran'kova, candidate of physical and mathematical sciences, All-Union Scientific Research Institute of Hydrometeorological Information-World Data Center, manuscript submitted 17 Jun 80]

[Text]

Abstract: Different meteorological forecasting algorithms based on use of the analogue principle are examined. The advantages of use of the one "best" analogue and groups of analogues are compared. The limits of predictability of meteorological processes are evaluated. It is shown that the conditions for long-range forecasting in the northern hemisphere are now deteriorating due to the increase in climatic variability and the anomalous degree of the mean monthly temperature field. Algorithms for adaptive forecasting on the basis of a group of analogues with a variable range of predictors and optimization of forecasting using analogues are discussed.

The method for forecasting weather by the use of analogues is widely used in research and in practical work and also in pure (explicit) form as a component (latent) part of the overwhelming majority of synoptic (classification, use of experience, etc.) and statistical procedures.

The fundamental ideas in the field of application of analogues of meteorological fields were unquestionably developed by N. A. Bagrov [1-3]. A number of interesting results and proposals are also given in [4, 5, 13].

The method for use of analogues in monthly weather forecasts [12] assumes choice of the one best analogue for forecasting. The choice of only "computer" analogues is accomplished using quantitative criteria. The combining of "computer" and "subjective" analogues of different meteorological objects is accomplished by the weatherman "visually," although a series of recommendations have also been formulated for this procedure.

In order to predict using analogues, generally speaking, it is possible to use an algorithm proposed in [10], which is a variant of the maximum similarity method or the so-called one best analogue method. Later a considerably more general prediction scheme was developed on the basis of application of the similarity principle, using the group analogues method [6, 8]. The conditions for use of this scheme were as close as possible to the real conditions of prognostic practice. For example, it takes into account, in particular, that usually use is made of not one, but a group of archives of meteorological objects of different types (for example, the fields of temperature, pressure and precipitation anomalies, catalogues of types of processes, etc.). In addition, provision is made for the possibility of using not one analogue, but a group of analogues. On the one hand, this can improve the quality of the forecasts, and on the other hand it can ensure a changeover to forecasts in stochastic form; a categorical forecast is also possible, by which in this case is meant the value of the predictant (with a weighting function dependent on the level of similarity) averaged for the group of analogues. We note at this point that all the empirical results presented in this study were obtained within the framework of use of precisely this numerical model.

Now we will examine some general problems in use of the analogues method in the problems related to long-range meteorological forecasting. One of these problems is an evaluation of the predictability of meteorological processes.

The predictability problem is very timely for meteorologists and it is not without reason that it has been so intensively investigated over the course of the last 20 years (a review of studies up to 1974 can be found in [11]).

It seems extremely desirable to break this problem down into two: the problem of internal predictability of the atmosphere and special predictability, determined by the possibilities of specific prognostic schemes at a given level of completeness and accuracy of information on the initial state [9].

In [7] we refined the basic definitions and introduced some quantitative characteristics, including special predictability by the analogues method. We will recall the principal points.

Assume that Y_t is the predicted process. Without limitations on universality we will consider it as Y_g, that is, with a time interval of 1 year (as if all its states related to a fixed calendar time, season or the year as a whole). Assume that $D(Y_{g0}, Y_g)$ is a measure of the difference (metrics), which can serve as an evaluation of the erroneousness of the categorical forecast Y_g when the state Y_{g0} has been realized. We will examine evaluations of "forecasts by the analogue method" when the analogues to the predictant are selected. Such evaluations correspond to an ideal scheme (method) for selecting the analogues when the best analogues in predictant space Y correspond to the best analogues in predictor space X. These evaluations can be used as predictability indices.

As the initial success level for a comparative evaluation of forecasts we introduce evaluations of the random analogue method when in the role of an analogue use is made of one record of the process randomly selected from the archives of observations. Naturally, for these purposes it is also possible to use evaluations of a climatic forecast when statistical (long-term) means (climatic norms) are used as an analogue and the forecast corresponding to it. Then for the current process Y_{g0} the quantitative characteristics of individual predictability of a specific process will be:

-- for the one best analogue method

$$\mathrm{Mi}^g_{g_0}(D(Y_{g_0}, Y_g)) = \min_g \{D(Y_{g_0}, Y_g)\};$$

-- for a random analogue

$$E^g_{g_0}(D(Y_{g_0}, Y_g));$$

-- for a group of analogues

$$D(Y_{g_0}, EY^a_{g_0}).$$

Here and in the text which follows the letter E denotes the averaging operator and the superscript (if there is one) indicates the region of applicability of the operator; Y_g denotes the analogue to the process Y_{g0} and EY_{g0} is the mean state of the predictant for the sample of analogues.

If these evaluations are now averaged for g_0, that is, from the set of all possible states of the predictant, alternately regarding them as the predictable process, the resulting evaluations

$$E^{g_0}\,\mathrm{Mi}^g\,(D(Y_{g_0}, Y_g)),\ E^{g_0 g}\,(D(Y_{g_0}, Y_g)),\ E^{g_0}\,D(Y_{g_0}, EY^a_{g_0})$$

will be the quantitative characteristics of the special predictability of the mentioned three methods for prediction by the analogues method respectively.

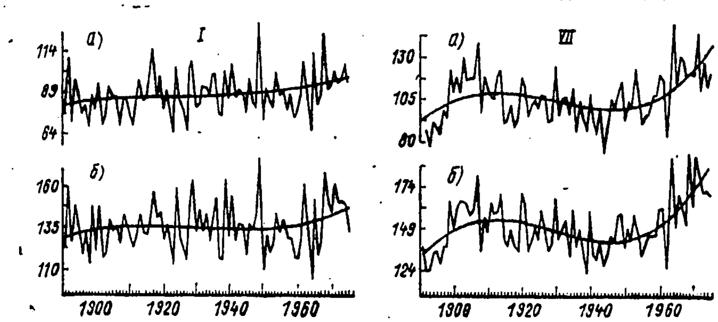

Fig. 1. Evaluations of individual predictability of air temperature fields over the northern hemisphere in January (I) and July (VII):
a) $\mathrm{Mi}^g\,(D^{(1)}_{g_0 g}\,(W^g T))$, б) $E^g\,(D^{(1)}_{g_0 g}\,(W^g T))$.

Figure 1 shows the variation of evaluations of individual predictability of the best and random analogue methods for the air temperature fields over the northern hemisphere (35-70°N). Here the metrics used was Euclidean distance (mean square difference $D^{(1)}$), whereas the temperature fields were represented by normalized W8T anomalies (deviations from the mean long-term value related to the standard deviation at the points).

In this specific case the measure

$$D(Y_{g_0}, Y_g) = D^{(1)}_{g_0 g}(\nabla^g T)$$

can be interpreted as the relative (normalized for σ) quadratic difference ε of the temperature fields, averaged for the field, or as $\sqrt{Q}$, where Q is a parameter widely known in meteorology, used, in particular, as an evaluation of the quality of long-range forecasts of meteorological fields [3]. Evidently, it would be interesting to cite similar results also for the second parameter ρ (similarity of fields with respect to the geographical distribution of the sign of the anomalies), so popular among meteorologists. However, this parameter is very sensitive to evaluations of the "norms," as a result of which it is not used in this study (in this study in computing the normalized anomalies we used the mean long-term anomalies recommended for the WMO for the CLINO period 1931-1960, which as an average for the hemisphere are characterized by a substantial positive displacement of the entire series relative to the means).

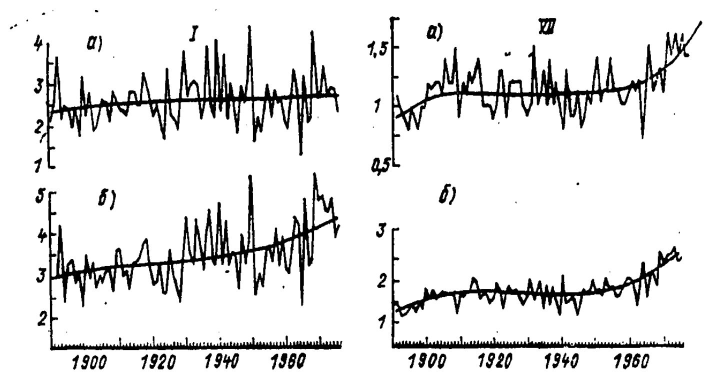

Fig. 2. Characteristics of anomalous character of the temperature field over the northern hemisphere in January and July (I and VII respectively):

$$a)\ \sqrt{\overline{E^{\varphi\lambda}(V^g T)^2}},\quad б)\ \sqrt{\overline{E^{\varphi\lambda}(\Gamma\, V^g T)^2}}.$$

Figure 1 shows that evaluations of individual predictability change substantially from year to year and during recent decades the forecasting conditions for the hemisphere in general deteriorated, especially in summer: both evaluations of individual predictability are characterized by a significant ascending trend (the figure shows approximations of the trend by a cubic polynomial by the least squares method). As a comparison, Figure 2 shows similar curves for the quadratic temperature anomaly, averaged for a hemisphere

$$\sqrt{\overline{E^{\varphi\lambda}(V^g T)^2}}$$

and the averaged horizontal gradient of the field of anomalies

$$\sqrt{E^{\varphi\lambda}(\Gamma V^g T)^2} = \sqrt{E^{\varphi\lambda}(\Gamma_\varphi^2 V^g T + \Gamma_\lambda^2 V^g T)}.$$

Table 1

Correlation Coefficients Between Characteristics of Temperature Regime in the Northern Hemisphere in January (Over Diagonal) and July (Under Diagonal)

Temperature regime characteristics	$E^g D^{(1)} W^g T$	$Ma^g D^{(1)} W^g T$	$Mi^g D^{(1)} W^g T$	$E^{\varphi\lambda} T$	$\sqrt{E^{\varphi\lambda}(\Gamma T)^2}$	$\sqrt{E^{\varphi\lambda}(V^g T)^2}$	$\sqrt{E^{\varphi\lambda}(\Gamma V^g T)^2}$
$E^g L^{(1)} W^g T$	1,00	0,33*	0,90**	—0,16	0,12	0,91**	0,78**
$Ma^g D^{(1)} W^g T$	0,55**	1,00	0,24*	—0,23*	0,17	0,37**	0,13
$Mi^g D^{(1)} W^g T$	0,90**	0,48**	1,00	—0,13	0,15	0,83**	0,78*
$E^{\varphi\lambda} T$	—0,14	—0,33*	—0,18*	1,00	—0,78**	—0,07	—0,09
$\sqrt{E^{\varphi\lambda}(\Gamma T)^2}$	0,03	0,03	—0,03	—0,12	1,00	0,13	0,13
$\sqrt{E^{\varphi\lambda}(V^g T)^2}$	0,91**	0,47**	0,85**	—0,11	0,18*	1,00	0,81**
$\sqrt{E^{\varphi\lambda}(\Gamma V^g T)^2}$	0,83**	0,20*	0,75**	—0,00	0,06	0,86**	1,00

Notes. 1. The asterisks denote the significant correlation coefficients with a significance level $\alpha = 0.1$ (*) and $\alpha = 0.001$ (**). 2. $E^{\varphi\lambda}T$ denotes the mean hemispheric temperature, and $\sqrt{E^{\varphi\lambda}(\Gamma T)^2}$ is the horizontal temperature gradient averaged for the hemisphere. The other notations are explained in the text.

The correlation between these characteristics is very high in both January and July (Table 1). Thus, the evaluations of predictability of individual processes by the random and one best analogue methods are closely related to one another and to the characteristics of anomalousness of the predicted fields.

It is interesting to compare the cited evaluations with the predictability by the groups of analogues method, and also to trace the relationship of these evaluations for different predictants. For this purpose an experiment was carried out for predicting the mean monthly air temperature over the northern hemisphere (35-70°N) and over the territory of the USSR with zero and month advance times. As a result, we obtained evaluations of individual predictability for two extremal seasons (for January-February and July-August) of the ten-year test period 1967-1976. The analogues were selected from a scheme of group analogues [6] (henceforth it will be designated GRAN) directly for the predicted object, which was represented in the archives by the values of the normalized anomalies. As before, use was made of Euclidean metrics so that all the evaluations represented below are adequate for the relative forecasting error (ε).

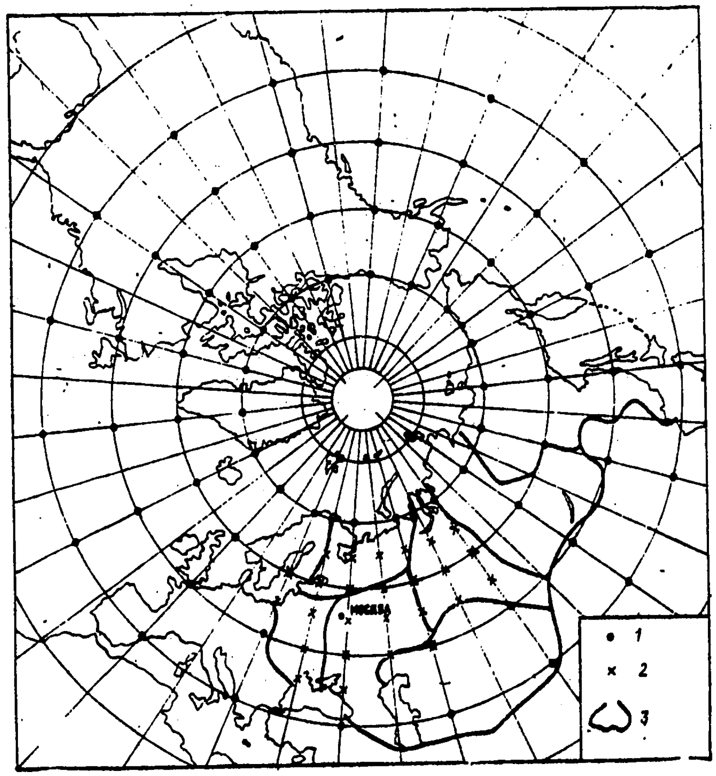

Fig. 3. Network of points of intersection for prediction of temperature field: 1) latitude zone 35-70°N, 2) lowland territory of USSR, 3) seven administrative regions of the USSR.

As the predicted objects the following were examined parallely:

1) the temperature field at 144 points of a regular grid in the latitude zone 35-70°N;

2) the temperature field at 26 points of grid intersection in the lowland territory of the USSR;

3) the values of the normalized temperature anomalies, averaged for the area of seven administrative regions in the USSR.

The network for all these territories is shown in Fig. 3.

Evaluations of predictability for January and July are given in Table 2.

Table 2

Evaluations of Predictability (Relative Error) of Temperature Field Over Different Territories of the Northern Hemisphere by Analogue Methods

	Forecasting method	1967	1968	1969	1970	1971	1972	1973	1974	1975	1976	Mean
	January											
	35—70° N											
1	Климатический	0,78	0,82	1,15	1,26	1,09	1,38	1,25	1,25	1,17	0,91	1,16
2	Случайного аналога	1,17	1,26	1,67	1,52	1,46	1,62	1,51	1,53	1,49	1,33	1,46
3	Лучшего аналога	0,83	0,90	1,25	1,19	1,08	1,13	1,17	1,13	1,11	1,00	1,08
4	Худшего аналога	2,08	1,82	2,25	2,25	2,10	2,29	2,45	2,11	2,22	1,95	2,15
5	Группового аналога	0,74	0,67	1,17	1,05	0,94	1,12	1,08	0,98	0,93	0,81	0,96
	Lowland territory of the USSR											
	Климатический	0,88	1,25	2,15	0,69	1,04	1,86	1,04	0.93	1,19	0,94	1,27
	Случайного аналога	1,15	1,46	2,26	1,04	1,26	1,99	1,31	1,22	1,37	1.22	1,43
	Лучшего аналога	0,42	0,44	0,75	0,47	0,48	0,75	0,63	0,47	0,36	0,56	0,53
	Худшего аналога	2,30	2,61	3,69	2,10	2,98	3,30	2,40	2,32	3,16	2,34	2,72
	Группового аналога	0,35	0,47	0,86	0,44	0,49	0,74	0,57	0,48	0,40	0,51	0,55
	Seven administrative regions of USSR											
	Климатический	0,61	0,77	1,95	0,44	0,66	1,52	0,74	0,74	0,87	0,68	1,00
	Случайного аналога	0,87	0,99	2,03	0,77	0,89	1,63	0,97	0,97	1,02	0,92	1,11
	Лучшего аналога	0,30	0,17	0,83	0,22	0,35	0,54	0,39	0,25	0,25	0,27	0,36
	Худшего аналога	1,71	1,83	3,09	1,61	2,30	2,68	1,84	1,91	2,64	2,21	2,18
	Группового аналога	0,21	0,23	0,84	0,17	0,32	0,48	0,37	0,19	0,21	0,22	0,39
	July											
	35—70° N											
1	Климатический	1,68	1,44	1,15	1,80	1,30	1.75	1,72	1.47			1,56
2	Случайного аналога	1,67	1,60	1,49	1,96	1,61	1,92	1,83	1,70			1,72
3	Лучшего аналога	1.31	1,25	1,17	1,50	1,25	1,48	1,43	1,35			1,34
4	Худшего аналога	2,13	2,11	2,00	2,41	2,08	2.49	2,28	2,22			2,21
5	Группового аналога	1,40	1,22	0,99	1,46	1,13	1,50	1,44	1,24			1.31
	Lowland territory of the USSR											
	Климатический	0,86	1,37	1,06	0,94	0,67	1,75	1,22	1,48			1,21
	Случайного аналога	1,22	1,54	1,36	1,26	1,10	1,90	1,45	1,75			1,67
	Лучшего аналога	0,46	0.59	0,60	0,52	0,51	0,83	0,67	0,89			0,63
	Худшего аналога	2,53	3,15	2,62	2,35	2,08	3,06	2,97	2,83			2,47
	Группового аналога	0,42	0,51	0,59	0,45	0,47	0,81	0,64	0,89			0,62
	Seven administrative regions of USSR											
	Климатический	0,51	1,03	0,71	0,54	0,39	1,04	0,43	1,14			0,78
	Случайного аналога	0,79	1,14	0,92	0,80	0,71	1,18	0,71	1.32			0,95
	Лучшего аналога	0,32	0,39	0,31	0,27	0,17	0,47	0,22	0,56			0,33
	Худшего аналога	2.00	2,14	2,08	1,76	1,57	2.05	1,58	2,39			1,94
	Группового аналога	0,22	0,31	0,32	0,23	0,15	0,43	0,19	0,57			0,30

KEY:

1. Climatic
2. Random analogue
3. Best analogue
4. Worst analogue
5. Group analogue

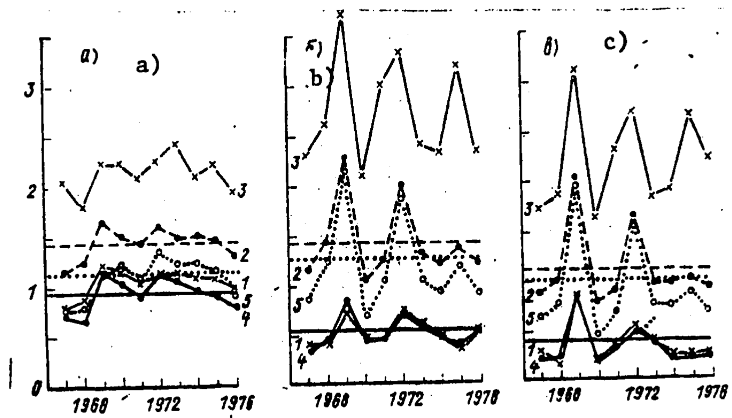

Fig. 4. Evaluations of individual predictability (mean square error of the field of normalized anomalies) of air temperature in January in latitude zone 35-70°N (a), for lowland territory of USSR (b) and for territory of seven administrative regions of USSR (c): 1) $\mathrm{Mi}^{g}(D(Y_{g_0}, Y_g))$, 2) $E^{g}(D(Y_{g_0}, Y_g))$, 3) $\mathrm{Ma}^{g}(D(Y_{g_0}, Y_g))$, 4) $D(Y_{g_0}, EY^{\alpha}_{g_0})$, 5) $D(Y_{g_0}, E^{g}Y)$.

In addition to the three methods indicated above, they include evaluations for the "poorest" analogue

$$\mathrm{Ma}^{g}_{g_0}(D(Y_{g_0}, Y_g)) = \max_{g}\{D(Y_{g_0}, Y_g)\}$$

and for a climatic forecast

$$D(Y_{g_0}, E^{g}Y).$$

The latter was obtained as the mean long-term value $E^{g}Y$ for a 76-year period of observations (1891-1966), from which the choice of analogues was made. In addition, the table gives evaluations of success of the corresponding methods, averaged for the 10-year period of the tests, which in conformity to the terminology introduced above corresponds to evaluations of the special predictability for this method. For greater clarity the results of one of the variants (January) are shown graphically in Fig. 4. The evaluations of special predictability (means for the ten-year test period) are shown here by horizontal straight lines.

The presented materials make it possible to draw the following conclusions:

1. As a rule, the group analogue is not worse (and for large territories -- better) than the one best analogue. Hence the groups of analogues method is preferable even when constructing a categorical forecast.

2. The analogues for global territories (even in an ideal case, that is, selected using an ideally operating system for the selection of analogues) do not ensure any important advantage over a climatic forecast. Evidently, this can be

attributed in part to the inadequate archives of observations (only 76 years!), but in any case it is necessary to reckon with this fact, solving the problem of long-range forecasting by the analogues method.

3. Smoothing by area made it possible to accomplish a greater compression of information on the predicted object and favored an increase in the prognostic possibilities of the analogues. For all methods the success level became higher, although the relationship between them virtually did not change.

4. During summer, the possibilities of analogues for forecasting are greater than in the winter, judging from the relative error. The reason for this is partially concealed in the lesser variability of temperature in summer in comparison with winter.

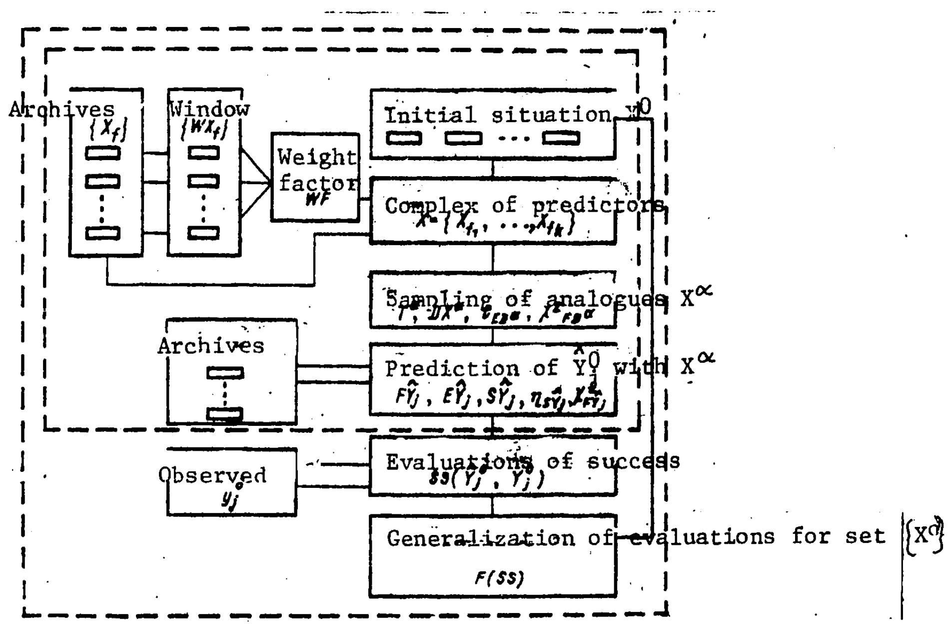

Fig. 5. Principal stages in operation of GRAN group analogues scheme.

In general it follows from the cited materials that the analogues method (and especially the group analogue method) with the availability of an informative system of predictors has a definite advantage in comparison with a climatic forecast and especially a random forecast. This advantage is especially conspicuous with rejection of macroscale analogues in general (for example, at the scale of the extratropical zone of the northern hemisphere, as indicated above).

These two conclusions determine the possible ways to bring about further optimization of automated use of the analogues method for long-range forecasting purposes.

The inadequate effectiveness of macroscale analogues in general agrees with the hypothesis of "local similarity" and indicates the desirability of sectioning of macroscale objects of forecasting into smaller segments and the use of different groups of analogues for individual segments. This, in turn, means the use of different types of information and possibly different smoothing filters in the search for analogues for forecasting in different segments. Moreover, taking into account the substantial variation of individual predictability, it can be assumed that the variation of the used information and filters should be dependent on the characteristics of the current process.

Thus, we arrive at two possible methods for optimizing the parameters of the scheme in accordance with its two possible quality criteria, characterizing the individual (current process) predictability and the special (as an average for the test period) predictability respectively. We will examine them applicable to the GRAN group analogues scheme [6], the principal stages of which are schematically shown in Fig. 5. Here the WF vector of the weighting factors of different factors exerting an effect on the weather (each of which can be regarded as a multidimensional vector), like the weighting functions WX for each of them (playing the role of spatial-temporal filters ensuring the discrimination of components of a definite scale), are the input parameters of the scheme and are designated by the user in accordance with the adopted physical model of the predicted process. It is understandable that these two parameters to a high degree determine the possibility of use of the scheme in each specific case. However, in most cases, unfortunately, we do not have adequate information for adopting such a model and therefore precisely the above-mentioned parameters (WF and WX) are the first to be optimized.

With respect to the optimization criteria, here some refinements are needed. Level optimization for special predictability, that is, as an average for the test period, causes no difficulties. This is the usual optimization procedure, ordinarily employed in the development of all statistical forecasting methods. Here the quality criterion can be any evaluation of the success of the forecast provided for in the scheme, averaged for the period of the tests (it is also the corresponding evaluation of special predictability of the method). With respect to optimization for evaluations of individual predictability, it makes sense only in real time when these evaluations are unavailable, since the predicted process Y_{g0} is still not observed. In this case as the optimization criteria use should be made of indirect a priori evaluations of the quality of a group of analogues, not using actual information on the predicted object. Among the evaluations provided for in the GRAN scheme these include the following.

First, there are evaluations of the quality of the analogues in X (in predictor space). Assume that

$$\left\{T_j^{\alpha},\ j = \overline{1,\ NA}\right\}$$

are the dates NA of the selected analogues and

$$\left\{DX_j^{\alpha},\ j = \overline{1,\ NA}\right\}$$

are the values of the integral measure of the difference in the system of predictors X between the processes corresponding to these dates and the current process. It is understandable that the DX^{α} vector is obtained as the NA vector of the minimum values in the time series $\left\{D(X_0,\ X_g),\ g = \overline{1,\ NG}\right\}$, where X_0 is the current process, X_g is the process from the archives of observations, NG is the volume of observations in the archives. For the two mentioned series

$$\left\{DX_j^{\alpha},\ j = \overline{1,\ NA}\right\} \text{ and } \left\{DX_g,\ g = \overline{1,\ NG}\right\}$$

in the GRAN scheme the statistics of the difference measure are computed, including the mean EDX^{α} and EDX and the distribution functions FDX and FDX (conditional and unconditional respectively), the difference between which, according to criteria known in statistics (for example, Student's t for the means and χ^2 for the distribution functions) can be used as optimization criteria.

The second group of possible optimization criteria for individual predictability is related to evaluations of the anticipated quality of the analogues in Y (in predictant space). These evaluations are evaluations of the statistics of the predictant in a sample of analogues (conditional) and in the entire initial sample (unconditional). Among these the most interesting are the standard deviations SY and SY and the distribution functions FY^{α} and FY. As the optimization criteria in this case it is possible to use $\eta = SY^{\alpha}/SY$, characterizing the relative scattering of the predictants in the sample of analogues, and χ^2 for comparison of the FY^{α} and FY distributions.

Both of the mentioned optimization variants are shown in Fig. 5 by dashed blocks. It is understandable that with optimization for special predictability as the criteria it is also possible to use a priori success evaluations. In addition, it is clear that any of these two optimizations is possible only with powerful electronic computers (at least of the third generation), so that in this stage the development of the corresponding algorithms can be considered a timely and immediate problem. With respect to the use of the now-existing variant of the scheme (without optimization) for solution of problems in practical prediction, its testing has already been carried out for predicting the mean monthly temperature anomalies in the northern hemisphere and their results will be published in an individual article.

In conclusion we feel it desirable to note that a broad range of parameters in the prognostic scheme should remain free, affording the researcher broad possibilities for experimentation in the course of investigations and in the prediction process. In other words, the scheme should make it possible to carry out work on an electronic computer in a dialogue regime, clearly delimiting the results obtained objectively using quantitative methods and their change as a result of variation of the free parameters at the will of the researcher-forecaster. We will regard the analogues method as a variety of an adaptive forecasting scheme when adaptation of the system is accomplished taking into account not only closeness with respect to time characteristics, but also a far broader set of parameters.

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PREDICTION OF SURFACE AIR HUMIDITY

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 23-28

[Article by V. N. Zolotorev and B. D. Uspenskiy, professor, Hydrometeorological Scientific Research Center, manuscript submitted 4 Jun 80]

[Text]

Abstract: The article gives an analysis of the dew point transfer and dew point spread equations. In accordance with its results, a method is proposed for computing the latter in weather diagnosis and prediction. In this method for computing turbulent moisture exchange use is made of empirical parameters -- the amplitude of dew point spread and the extent of lower cloud cover. The results of the computations for four months give some idea concerning the qualities of the proposed method for computing the dew point spread in the surface layer of the atmosphere.

A prediction of air humidity at different levels in the atmosphere is an important part of short-range weather forecasting necessary for precomputing precipitation, the form and extent of cloud cover and other weather phenomena dependent on the phase transformations of water vapor in the atmosphere.

Studies of a number of authors have been devoted to problems involved in the methods for predicting humidity. However, the relatively small number of such studies does not correspond to the role which air humidity plays in atmospheric processes. This can be judged, in particular, from the small number of articles published during the last decade and the complexities in forecasts of weather phenomena associated with changes in the air humidity fields in the atmosphere. The development of methods for computing air humidity has been adversely affected by the lesser accuracy in its measurements in the free atmosphere and the absence of regular observations of the liquid water content of clouds.

As was demonstrated in [3], the use of simplified methods for predicting surface humidity, based on allowance for advection or inertia, has not led to positive results. Accordingly, in the method for predicting surface air humidity set forth in the MANUAL ON SHORT-RANGE FORECASTS [11] and in the METHODOLOGICAL INSTRUCTIONS [8] the results of theoretical [5] and empirical studies, in particular [1, 2, 4, 6, 7, 9], have been included.

The theoretical basis of methods for computing air humidity at different levels in the atmosphere is given by the equations

$$\frac{dq}{dt} = \frac{1}{\rho} \frac{\partial}{\partial z} k \rho \frac{\partial q}{\partial z} - \frac{m}{\rho}, \tag{1}$$

$$\frac{d\Theta}{dt} = \frac{1}{\rho} \frac{\partial}{\partial z} k \rho \frac{\partial \Theta}{\partial z} + \frac{Lm}{C_p \rho}, \tag{2}$$

where q is specific humidity, ρ is air density, k is the turbulence coefficient, m is the mass of water vapor (air) in a unit air volume, condensing (evaporating) in a unit time, θ is potential temperature, c_p is the specific heat capacity of dry air at a constant pressure, L is the latent heat of condensation.

In equation (1) no allowance is made for the influx of water vapor as a result of evaporation of moisture from the underlying surface, and in (2) -- the heat expended in this process, and also the radiant heat influx.

On surface weather charts there are data on temperature and on the dew point, and on high-altitude charts -- on temperature and the dew point spread. Accordingly, instead of q and θ we introduce into equations (1) and (2) the dew point temperature τ and its spread D = $(T - \tau)$ and air temperature T. With the transformation of (1) we use the expression

$$q = 0{,}622 \frac{E(\tau)}{P}, \tag{3}$$

where R/R_n = 0.622, and the Clausius-Clayperon equations, the equations of state and statics

$$\frac{dE}{dT} = \frac{LE}{R_n \tau^2}, \quad P = \rho RT, \ \partial P = -g \rho \partial z. \tag{4}$$

In (3) and (4) E is the elasticity of saturated water vapor, P is pressure, R and R_n are the specific gas constants of dry air and water vapor, g is the acceleration of free falling. After determining the partial derivatives of x, y, z and t from (3), we find that

$$\frac{\partial q}{\partial z} = \frac{0{,}622}{P} \left(\frac{LE}{R_n \tau^2} \frac{\partial \tau}{\partial z} + \frac{Eg}{R\tau} \right), \tag{5}$$

$$\frac{dq}{dt} = \frac{0{,}622}{P} \left(\frac{LE}{R_n \tau^2} - \frac{E}{P} \frac{dP}{dt} \right). \tag{6}$$

The dew point transfer equation, if the right-hand side is transformed by means of (5) and if dq/dt is replaced in accordance with (6), will have the form

$$\frac{\partial \tau}{\partial t} + \left(u \frac{\partial \tau}{\partial x} + v \frac{\partial \tau}{\partial y} \right) + \left(\gamma_{a\tau} - \gamma_\tau \right) w - \underline{\frac{R_n \tau^2}{PL} \left(\frac{\partial P}{\partial t} + u \frac{\partial P}{\partial x} + v \frac{\partial P}{\partial y} \right)} = \tag{7}$$

$$= \frac{\partial}{\partial z} k \frac{\partial \tau}{\partial z} + \underline{k \left(\frac{L}{R_n \tau^2} - \frac{3}{\tau} \right) \left(\frac{\partial \tau}{\partial z} \right)^2} + \underline{\frac{kg}{R\tau} \left(1 - \frac{2R_n \tau}{L} \right) \frac{\partial \tau}{\partial z}} +$$

$$\underline{+ \frac{gR_n \tau}{RL} \frac{\partial k}{\partial z} - \frac{mRR_n \tau^3}{qPL}},$$

where u and v are the horizontal components of wind velocity, w is vertical velocity.

In simplified form equation (7) is given in [13].

Equation (2), taking into account that

$$\Theta = T\left(\frac{1000}{P}\right)^{\frac{\varkappa-1}{\varkappa}}, \quad \gamma_a = \frac{\varkappa-1}{\varkappa}\,\frac{g}{R}, \tag{8}$$

$$\frac{\partial\Theta}{\partial z} = \frac{\Theta}{T}\left(\gamma_a + \frac{\partial T}{\partial z}\right), \quad \frac{\partial P}{\partial z} = -\frac{P}{RT^2}\left(\frac{g}{R} + \frac{\partial T}{\partial z}\right), \tag{9}$$

$$\frac{d\Theta}{dt} = \frac{\Theta}{T}\,\frac{dT}{dt} - \frac{\varkappa-1}{\varkappa}\,\frac{\Theta}{P}\,\frac{dP}{dt}, \tag{10}$$

can be given the form

$$\frac{\partial T}{\partial t} + \left(u\frac{\partial T}{\partial x} + v\frac{\partial T}{\partial y}\right) + (\gamma_a - \gamma)\,w - \underline{\frac{\varkappa-1}{\varkappa}\,\frac{T}{P}\left(\frac{\partial P}{\partial t} + u\frac{\partial P}{\partial x} + v\frac{\partial P}{\partial y}\right)} = $$
$$= \frac{\partial}{\partial z}\,k\,\frac{\partial T}{\partial z} + \underline{\frac{k\gamma_a}{TP}\left(\gamma_a + \frac{\partial T}{\partial z}\right) + \gamma_a\frac{\partial k}{\partial z}} + \underline{\frac{mLRT^2}{Pc_p\Theta}}, \tag{11}$$

where $\gamma = -\partial T/\partial z$, γ_a is the dry adiabatic temperature gradient,

$$\varkappa = \frac{c_p}{c_v},$$

c_v is the specific heat capacity of dry air at a constant volume.

Subtracting (7) from (11), it is possible to find, in general form, the transfer equation for the dew point spread. Terms whose values are 10-10^2 times smaller than the other terms have been underlined in (7) and (11).

In the surface layer of the atmosphere, where the term dependent on w is also of insignificant value, the dew point and dew point spread equations will have the form

$$\frac{\partial\tau}{\partial t} = -\left(u\frac{\partial\tau}{\partial x} + v\frac{\partial\tau}{\partial y}\right) + \frac{\partial}{\partial z}\,k\,\frac{\partial\tau}{\partial z}, \tag{12}$$

$$\frac{\partial D}{\partial t} = -\left(u\frac{\partial D}{\partial x} + v\frac{\partial D}{\partial y}\right) + \frac{\partial}{\partial z}\,k\,\frac{\partial D}{\partial z}. \tag{13}$$

Therefore, the local changes in air humidity in the surface layer are caused for the most part by its horizontal transfer and vertical turbulent moisture exchange.

In synoptic practice use is made of the dew point equation (12) in which the term

$$\frac{\partial}{\partial z}\,k\,\frac{\partial\tau}{\partial z}$$

is assumed to be equal to the amplitude of the dew point. The amplitude values for clear, semiclear and cloudy weather are found from empirical data obtained in [1, 5, 6], and in systematized form, given in [8].

Table 1 lists the correlation coefficients giving some idea about the closeness of the correlation between the amplitudes T, D, τ and the total cloud cover averaged for 0300 and 1500 hours, expressed in tenths of the lower cloud cover.

Computations for the European USSR were made using data on synoptic situations for 1978 at places with weak advection of temperature and air humidity. The number of cases in each month did not exceed 70-80. In July 1979 in the Moscow and Kuybyshev regions there was weather with low wind velocities and therefore 30 cases were used i computing the correlation coefficients at each point.

Table 1

Correlated parameters	European USSR				Moscow	Kuybyshev
	February	April	October	July	July	
A_T and N	-0.33	-0.69	-0.89	-0.67	-0.062	-0.84
A_D and N	-0.13	-0.63	-0.78	-0.51	-0.53	-0.72
A_τ and N	-0.44	-0.06	-0.09	-0.23	0.17	-0.14

Note. A_T, A_D and A_τ are the amplitudes of temperature, dew point spread and dew point, N is the mean cloud cover for 0300 and 1500 hours.

It follows from the data in Table 1 that the amplitudes of temperature and dew point spread are usually in an inverse quite close dependence on the extent of cloud cover. An exception is the data for February, attributable to the fact that in this month the amplitudes of T and D, the same as the quantity of clouds, varied slightly over the territory. The dependence of the amplitude of dew point on cloud cover was weak. However, the increase in r in February can be attributed to the fact that in winter at low temperatures (-10°C or below) even small changes in water vapor elasticity are adequate for causing substantial variations of dew point temperature and its amplitude. In the summer months, however, the dew point is insensitive to small variations in water vapor elasticity. The r values for Moscow and Kuybyshev confirm the results obtained for the European USSR and show that the quite close correlation between (A_T, N) and (A_D, N) is characterized by stability.

It also follows from the data in Table 1 that the dew point spread equation must be given preference in computing air humidity in the surface layer of the atmosphere if as the parameters replacing the value

$$\frac{\partial}{\partial z} k \frac{\partial D}{\partial z}$$

use is made, as in the case of dew point temperature, of statistical data on the amplitudes of the dew point spread and the quantity of clouds.

In the method for computing the dew point temperature in [8], on the basis of equation (12), for determining its transformational changes use is made of the mean monthly amplitudes, computed for nonmoving air masses and a small quantity of clouds. Despite these limitations, they have also come into use in determining the transformational changes in the moving air. This assumption, not validated in methodological respects, naturally led to a decrease in the accuracy of computations of surface air humidity.

Introducing into (13) the notation of an individual derivative

$$\frac{dD}{dt} = \frac{\partial D}{\partial t} + u\frac{\partial D}{\partial x} + v\frac{\partial D}{\partial y} \tag{14}$$

and integrating the transformed equation in time, we obtain

$$D_t - D_{t=0} = \overline{\left(\frac{\partial}{\partial t} k\frac{\partial D}{\partial z}\right)} t, \tag{15}$$

where D_t and $D_{t=0}$ are the dew point spreads at the final and initial points of an air particle trajectory, t is the time interval for which the trajectory was constructed,

$$\overline{\left(\frac{\partial}{\partial z} k\frac{D}{\partial z}\right)}$$

is the turbulent exchange value, averaged for the time t, which is approximately equal to its value at the mid-point of the trajectory.

With u = v = 0 the difference $(D_t - D_{t=0})$ is the local change in the dew point spread, representing a definite transformational change in D during the time t. As noted, this method was used in the method for computing dew point described in [8].

In order to apply the method for determining transformational changes, following from (15), for individual months we prepared maps of the amplitudes of dew point spread (A_D) on the basis of data from the HANDBOOK ON CLIMATE [12] on the diurnal variation of relative humidity and air temperature using the psychrometric tables [10]. The A_D values on these maps, equal to the difference between the D values at 1500 and 0300 hours, are the mean 12-hour transformational changes in the dew point spread. For these same months we constructed maps of total cloud cover, expressed in tenths of lower cloud cover, characterizing the radiation conditions of formation of the amplitudes of the dew point spread. In computations using formula (13) the quantity of clouds is used as a parameter making it possible to convert from the mean A_D values corresponding to the mean tenths of lower cloud cover to the amplitudes for the observed cloud cover.

Equation (13), for computing the D value with advance times of 12, 24 and 36 hours, can be given the following form:

$$\begin{gathered} D_{12} = (D_0)_{12} + k_1(\Delta D)_{12} + a_N(A_D)_1, \\ D_{24} = (D_0)_{24} + k_1(\Delta D)_{24}, \\ D_{36} = (D_0)_{36} + k_1(\Delta D)_{36} + a_N(A_D)_1, \end{gathered} \tag{16}$$

where D_{12}, D_{24}, D_{36} are the D values 12, 24 and 36 hours after the initial time, $(D_0)_{12}$, $(D_0)_{24}$, $(D_0)_{36}$ are advective D values, $\Delta D = D_{end} - D_{beg}$ are the advective changes in D after 12, 24 and 36 hours, taken with the opposite sign, $(A_D)_1$ is the transformational change in D, read from the map of amplitudes at the mid-point of a 12-hour segment of the trajectory closest to the point for which D is computed, $a_N = 1.0$ with $N_{act} = \bar{N}$, $a_N = 0.2(0.3)$ with $N_{act} > \bar{N}$ and $a_N = 1.1(1.5)$ with $N_{act} < \bar{N}$, N_{act} is the mean quantity of actual (prognostic) cloud cover for 0300 and 1500 hours at the point for which the $(A_D)_1$ value is read, $\bar{N}$ is mean monthly lower cloud cover, $k_1 = 0.1$. In computations from the initial weather maps for 1500 hours the $a_N(A_D)_1$ value in (16) is taken with a minus sign. The values of the k_1 and a_N coefficients were found by statistical processing of the actual data. The second term in (16) was introduced in order to exclude the advective contribution from the climatic amplitudes of the dew point spread.

Table 2

Month	a_N	n	$\overline{\lvert \delta D \rvert}$	ρ	E
October	0.3; 1.5	60	1.9	0.60	0.60
February	0.3; 1.5	60	1.2	0.60	0.71
April	0.3; 1.5	36	4.5	0.89	0.50
July	0.2; 1.1	65	1.4	0.89	0.20

Note. a_N is a coefficient in (16), n is the number of cases, $\lvert\overline{\delta D}\rvert$ is the mean absolute error, ρ is an index of coincidence of signs of the computed and actual D changes, E is the relative error.

Table 2 gives some idea concerning the accuracy in computing D_{12} using formula (16) and initial weather maps for 0300 and 1500 hours for July, April 1976, October 1977 and February 1978 in Moscow, Minsk, Gor'kiy, Kazan', Vologda and Khar'kov. It should be noted that in the computations the method for constructing the trajectories of air particles in [11] in all cases was kept regardless of the errors in diagnosis D_{12}. In this connection the data in Table 2 give some idea concerning the possible errors in predicting D_{12}. It follows from Table 2 that the accuracy in computing D_{12} is quite high. Only the absolute error for April was too high. The mean probable success in computing D_{12} for four months, determined in accordance with the instructions on the evaluation of forecasts, with allowance for the errors adopted in them for air temperature, was 88%, exceeding the probable success in computations of dew point temperature by the method given in [8]. Evaluations of individual examples of diagnosis and prediction of the dew point spread using formulas (16) with advance times of 12, 24 and 36 hours were close to those cited, somewhat decreasing with an increase in the advance time.

It follows from the materials in this article that in developing methods for predicting air humidity it is possible to use the dew point transfer or dew point spread equations cited above if we limit ourselves to allowance for the principal factors exerting an influence on its use.

In the case of the surface dew point spread the method for the parameterization of the value

$$\frac{\partial}{\partial z} k \frac{\partial D}{\partial z}$$

by the introduction of empirical data on the amplitude of D and the quantity of lower clouds into (13) is effective. In order to solve the problem of the practical use of the considered method for computing the dew point spread in local weather forecasts it is necessary to have more complete data on its probable success. However, it is clearly desirable to develop a numerical variant of prediction of D in which the use of more complete initial information is one of the possible ways to increase its accuracy.

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USE OF A COMPLEX ANALOGUE IN THE PHYSICOSTATISTICAL METHOD OF WEATHER FORECASTING FOR FIVE-TEN DAYS

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[Article by M. N. Fedulova, candidate of geographical sciences, A. V. Borodina, candidate of geographical sciences, and A. V. Shuvalov, USSR Hydrometeorological Scientific Research Center, manuscript submitted 3 Jun 80]

[Text]

Abstract: The authors propose the use of a complex analogue as the initial sample of processes in the physicostatistical method in the prediction of mean temperature for five days and the H_{500} pressure field for the seventh day. The advantage of the proposed approach in comparison with the use of standard samples is demonstrated on the basis of independent material for winter five-day periods.

In Soviet weather forecasting practice for intermediate times (3-10 days) the analogues of synoptic processes have come into rather broad use. They are employed in predictions for natural synoptic periods [14] and are taken into account in forecasts for calendar time intervals (3, 5, 10 days).

The use of single analogues is always accompanied by risk, despite the high degree of their initial similarity to the current process [1, 14, 17]. Accordingly, recently methods have appeared for the use of so-called "composite" or "group" analogues. There are different ways to use them. For the prediction of temperature in most cases this involves the determination of the mean characteristics from 10-15 records [5, 17]. It was demonstrated in [5] that it is desirable to find weighted means, each of the terms in which is proportional to the measure of similarity with the initial pressure field. However, the most rational method for the use of group analogues for the purpose of predicting circulation and weather for intermediate times is the solution of regression equations within the limits of a sample consisting of an adequately large number of similar situations [4, 11, 15].

In this article we examine the example of the introduction of a composite analogue into the "synoptic-hydrodynamic-statistical method for predicting temperature for 5 and 10 days and precipitation for 5 days" adopted as the fundamental method in the work of the 10-day forecasts section at the USSR Hydrometeorological Center and other long-range weather forecasting subdivisions [6]. In the course of development and testing it was found that the success of application of this method to a definite degree is dependent on the synoptic classification of initial data.

As is known from [6], computations of temperature anomaly forecasts for a five-day period from the dependence

$$Q_R = \overline{Q} + \sum_{h=1}^{h} a_{hR} L_h (T_{h_1}, T_{h_2}, \ldots) \quad (1)$$

are made separately for groups of processes formed on the principle that they belong to definite synoptic types. As the basis use is made of the well-known objective classification using the Kats circulation indices, according to which each process over the first natural synoptic region belongs to one of the five forms of circulation "W," "C," "N," "E" or "Zon" [8]. It was demonstrated in [7, 16] that the use of samples of synoptic processes of a definite form of circulation improves the quality of forecasts of weather elements by 3-10% in comparison with the results obtained using an unclassified sample.

The mentioned gain in quality can be attributed to the fact that within the classes formed on the basis of the similarity principle with standard positions the dispersion of all the elements (both spatial and for time shifts) is less than the climatic dispersion and accordingly the regression equations within these classes must operate better than with a random set of initial data [11]. However, when using the standard samples "W," "C," "N," "E" and "Zon" the success of computations in each individual case is dependent on the measure of similarity between the current H_{500} field and the mean standard field. On a practical basis this similarity is by no means always satisfactory due to the infinite diversity of synoptic situations in nature. In many cases the current process is either at the limit of the zonal and meridional states or the axis of the high-altitude ridge lies between the positions adopted for the circulation forms or the given case simply cannot be assigned to the standards established earlier. In practical work, meeting with difficulties in determining the forms of circulation from prognostic H_{500} charts for 48 hours [18], the team of forecasters has been forced to have recourse to parallel computations in several variants. Then there is an inevitable subjective compilation of the collected (sometimes contradictory) results.

There is no need to explain that such work is carried out by a group of highly qualified specialists and requires high expenditures of every type.

This investigation was undertaken for the purpose of finding a method for improving the quality of the computed forecasts by the method in [6] and objectivization of the prognostic procedure.

The working hypothesis was the assumption of a superiority of dynamic initial samples formed on the principle of similarity to the current process in comparison with earlier prepared reference series similar to the mean standard series (see figure). The proposed approach ensures a position of the current process at the center of the group of synoptic situations related to it, the further development of which must conform to definite patterns. Such a position should lead to the most complete realization of the advantages inherent in samples obtained by the analogue principle [11].

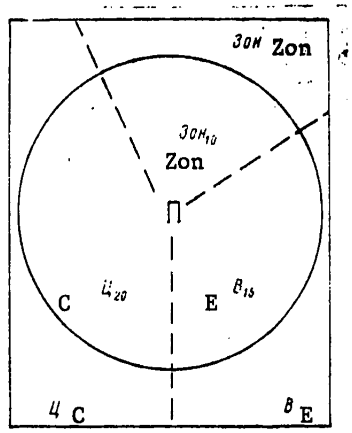

Fig. 1. Example of initial sample of processes by analogy with current process (II), occupying an intermediate position between "C," "E" and "Zon."

In checking this assumption use was made of a program making possible constant revision of the initial information for all the diagnostic and prognostic computations by the method in [6]. The experiment was carried out using material on mean air temperature for a five-day period in winter from 1959 through 1978 (360 five-day periods). The choice of the similarity test was an independent problem. At the present time several objective pressure field similarity criteria are known. In this connection, it is necessary, in particular, to note the studies of N. A. Bagrov [1], K. A. Vasyukov, N. I. Zverev, D. A. Ped' [3, 12], G. D. Kudashkin and M. I. Yudin [10], Kh. Kh. Rafailova [13], G. V. Gruza and his colleagues [4, 5, 15].

In selecting the measure of similarity (difference) in synoptic processes the authors of this article have chosen the G. D. Kudashkin criteria [10] as corresponding most fully to the characteristics of the prediction method [6], based on expansion of the predictor fields in natural orthogonal functions [2].

The indices of the degree of similarity of meteorological fields are selected on the basis of general stochastic considerations taking into account, in particular, the need for discriminating important information on the compared fields and on the possibility of maximum reduction of its volume. The method of natural orthogonal components has definite possibilities in this direction. Using this method the basic information is represented by a relatively small number of the first expansion coefficients [2]. A comparison of these coefficients is more convenient than a comparison of the meteorological field values themselves. As a test of the similarity of meteorological fields it is possible to adopt the M parameter proposed in [10], determined by the expression

$$M = \sum_{j=1}^{h} (T_{j,t_1} - T_{j,t_2})^2, \tag{2}$$

where T_{j,t_1}, T_{j,t_2} are the j-th coefficients of expansion of the field in natural orthogonal functions at the times t_1 and t_2. The lesser the M value, the greater is the degree of similarity of the compared fields.

In a one-dimensional case expression (2) has certain conveniences in the practical realization of the choice of analogues using an electronic computer.

In this article the M parameter is used in determining the similarity of the geopotential fields H_{500}. The initial materials used were the geopotential fields stipulated at 36 points in a territory bounded by 10°W, 100°E, 75 and 35°N; the points were selected at the points of intersection of a geographic grid with an interval of 10° in longitude and 20° in latitude.

The natural orthogonal functions $x_{j,i}$ were computed for a sample of fields including 360 cases for the three winter months from 1959 through 1978 (the middle days of the calendar five-day periods were selected). In expression (2) h was assumed equal to 5.

After the expansion operation the analogues were organized in accordance with the degree of increase in M. A sample of H_{500} fields was organized for computations from (1) from 45 initial terms of the series. The series $\Delta t_{initial}$ [Δt_{in}] and Δt_{5-day} were formed in accordance with the group analogue.

Table 1

Comparative Evaluation of Forecasts of Temperature Anomaly for 5 Days for January-February 1978: a) Inertial, b) Computed Using Standard Samples ("W," "C," "N," "E" and "Zon"), c) Computed Using Complex Analogue

Type of forecast	n	ρ	ε	Q	P	ε_1	ρ_{iner}	ε_{iner}	ρ_{in}	ε_{in}
a а	11	0,01	1,24	4,2	70	—	—	—	—	—
b б	11	0,29	0,98	3,3	81	0,54	0,45	0,87	0,58	0,80
c в	11	0,47	0,76	2,6	91	0,41	0,62	0,66	0,60	0,62

The first testing of forecasts of the mean air temperature anomaly for 5 days, prepared by the proposed method, was carried out from 16 January through 12 February 1978. During this period 11 series of forecasts were prepared, each of which contained an inertial forecast, a forecast computed by the method of 1973 with use of samples of synoptic processes with the circulation forms "W," "C," "N," "E," and "Zon" (without synoptic correction) and a forecast also computed using dependence (1), but with use of a composite analogue including 45 situations as the initial sample. An evaluation of 11 series of forecasts is given in Table 1. The arbitrarily selected testing period was difficult for prediction; the mean evaluation of the inertial forecasts with respect to the coincidence of prognostic and actual temperature anomalies ρ was close to 0 (0.01), the relative error was ε =1.24 (in 8 cases of 11 it was greater than 1) and P = 70%. Computations of forecasts by forms of circulation led to the following mean results: ρ = 0.29; ε = 0.98 and P = 81%.

The composite analogue gave the following: ρ = 0.47, ε = 0.76 and P = 91%. We note all the ρ values in the last group were positive and ε>1 was observed in only two cases.

Table 2

Comparative Evaluation of Forecasts of Mean Temperature Anomaly for Winter Five-Day Periods 1978/1979

	Dec 1978 r.			Jan 1979 r.				Feb 1979 r.			Mean
	15—19	18—22	27—31	8—12	12—16	19—23	24—28	5—9	12—16	23—27	
Inertial forecasts											
Form of circulation	N	N	N	N	E	W	Cen	N	Cen	N	
ρ	0,17	0,22	0,22	—0,17	—0,28	0,00	0,06	0,67	0,67	0,44	0,20
ε	0,85	0,89	0,85	2,50	1,81	1,29	1,08	0,62	0,67	0,99	1,16
Q	5,5	6,0	7,3	9,0	6,3	3,7	6,5	4,3	4,9	3,1	5,7
$P\%$	53	44	28	25	42	81	53	69	67	86	55
Methodological forecasts											
ρ	0,22	0,72	0,50	0,00	0,67	0,28	0,50	0,61	0,61	0,28	0,44
ε	0,92	0,57	0,74	1,40	0,66	1,08	0,60	0,62	0,66	0,90	0,81
Q	5,9	3,9	6,4	5,0	2,3	3,1	3,6	4,2	4,8	2,8	4,2
$P\%$	44	75	56	61	97	92	78	67	53	89	71
ε_1	0,98	0,63	1,09	0,84	0,37	0,49	0,55	0,67	0,84	0,49	0,70
$\rho_{\text{ин}}$	0,17	0,89	0,28	0,67	0,78	0,39	0,72	0,28	0,00	0,28	0,45
$\varepsilon_{\text{ин}}$	1,09	0,64	0,88	0,56	0,37	0,83	0,55	0,99	0,99	0,90	0,78
$\rho_{\text{исх}}$	0,28	0,61	0,22	0,83	0,72	0,33	0,61	0,56	0,44	0,28	0,49
$\varepsilon_{\text{исх}}$	0,86	0,56	0,84	0,54	0,69	0,98	0,57	0,89	0,89	0,98	0,78
Operational forecasts											
ρ	0,76	0,67	0,94	0,45	0,59	0,39	0,53	0,71	0,70	0,28	0,59
ε	0,67	0,65	0,55	0,82	0,73	1,46	0,56	0,47	0,68	0,97	0,76
Q	4,4	4,4	4,8	2,9	2,5	4,2	3,4	3,2	5,0	3,4	3,8
$P\%$	69	56	56	81	89	75	86	86	56	83	74
ε_1	0,75	0,80	0,84	0,46	0,39	0,66	0,54	0,54	0,89	0,57	0,64
$\rho_{\text{ин}}$	0,61	0,89	0,61	0,83	0,78	0,61	0,67	0,56	0,17	0,22	0,60
$\varepsilon_{\text{ин}}$	0,79	0,74	0,66	0,33	0,40	1,13	0,52	0,75	1,02	0,91	0,72
$\rho_{\text{исх}}$	0,56	0,78	0,50	0,89	0,67	0,39	0,43	0,72	0,28	0,28	0,55
$\varepsilon_{\text{исх}}$	0,63	0,65	0,63	0,31	0,76	1,33	0,54	0,68	0,92	0,97	0,74
"GRAN" forecasts											
ρ	0,56	0,83	0,72	0,31	0,61	0,44	0,72	0,56	0,83	0,28	0,59
ε	0,77	0,55	0,65	1,06	0,63	0,74	0,64	0,59	0,72	0,84	0,72
Q	5,0	3,7	5,6	3,8	2,2	2,1	3,9	4,1	5,3	2,6	3,8
$P\%$	64	75	61	75	94	100	78	67	50	83	76
ε_1	0,83	0,61	0,95	0,60	0,35	0,34	0,58	0,66	0,94	0,46	0,63
$\rho_{\text{ин}}$	0,33	0,83	0,39	0,72	0,78	0,56	0,67	0,44	0,00	0,39	0,51
$\varepsilon_{\text{ин}}$	0,91	0,62	0,76	0,42	0,35	0,57	0,59	0,95	1,08	0,85	0,71
$\rho_{\text{исх}}$	0,39	0,71	0,44	0,89	0,83	0,50	0,67	0,61	0,33	0,44	0,58
$\varepsilon_{\text{исх}}$	0,72	0,54	0,73	0,41	0,65	0,68	0,61	0,86	0,98	0,92	0,71

[ИН = iner(tial); ИСХ = in(itial)]

Due to the obvious advantage of the new method for forming the initial sample a second test was carried out in the winter of 1978/1979. In December, January and February we selected 10 5-day periods during which there was an appreciable increase or decrease of temperature over the European USSR. In the majority of cases the temperature anomalies of the initial day and the five-day period following it did not coincide in sign. In selecting cases for comparative testing a necessary

Table 3

Evaluation of H_{500} Forecasts for Seventh Day Computed on Basis of Synoptic-Statistical Forecasting Scheme With Use of Composite Analogue Selected for Forecasting H_{500} for 48 Hours

Initial day of forecast	Date of forecast for 48 hours	Date of forecast for 7th day	Form of circulation for 48-hr forecasts	Evaluation of H_{500} forecasts for 7th day			Error in inertial forecast, ε_{iner}
				ρ	δ	ε	
16 I	18 I	23 I	E	0.17	15.2	1.10	13.8
17 I	19 I	24 I	E	0.58	11.3	0.82	13.8
19 I	21 I	26 I	C	0.36	6.5	0.60	10.8
23 I	25 I	30 I	W	0.47	9.7	0.78	12.4
24 I	26 I	3 I	W	0.64	9.4	0.57	16.5
25 I	27 I	1 II	Zon	0.72	9.4	0.56	16.8
27 I	29 I	3 II	Zon	0.56	13.0	0.70	18.5
30 I	1 II	6 II	C	0.61	9.1	0.64	14.3
3 II	5 II	10 II	C	0.47	8.1	0.80	10.1
7 II	9 II	14 II	C	0.72	10.3	0.61	16.9
8 II	10 II	15 II	W	0.72	9.1	0.60	15.1
Mean				0.55	10.1	0.71	14.5

condition was also the availability of operational forecasts of the USSR Hydrometeorological Center for the corresponding 5-day periods.

Table 2 gives detailed results of an evaluation of 10 series of $\Delta t_{5\text{-day}}$ forecasts carried out with an electronic computer using a program adopted in the long-range forecasting section of the USSR Hydrometeorological Center. The forecasts prepared by the method of 1973 with the use of standard samples in the table are called "methodological," those corrected by the weatherman are called "operational," and forecasts computed with the use of a composite analogue are designated "GRAN" (group of analogues). The evaluation of the inertial forecasts of the selected series was low (ρ = 0.20; ε = 1.16; P = 55%). The methodological forecasts were considerably better (ρ = 0.44; ε = 0.81; P = 71%). The forecasts made using the composite analogue ("GRAN") on the average were better than the methodological forecasts (ρ = 0.59; ε = 0.72; P = 76%). In a comparison of the "GRAN" forecasts with operational forecasts (ρ = 0.59; ε = 0.76; P = 74%) it was found that their probable success is approximately identical. If it is taken into account that in the operational practice of the long-range forecasting section the forecasts are subjected to synoptic corrections prior to issuance, with an equal probable success preference must be given to the objective method.

The traditional method for the use of analogues is a lengthening of the prediction times by their use. The validation for this is as follows: "If a given synoptic process in the course of a given time interval develops the same as some other synoptic process occurring in the past, it can be assumed that in the future as well during some time the development of these processes will transpire in the same way" [1].

If not one, but a group of analogues [4, 5, 11, 15, 17] is used in the forecast, the development of processes in it should have similar characteristics. In this investigation it is postulated that the positive properties of a composite analogue should be manifested not only on the initial day of the forecast, but also during the period when the inertia of the initial state no longer is operative and only the evolutionary characteristics common in the group are operative. These characteristics to the maximum degree must relate to the current process because all the remaining processes are grouped around it.

On the basis of the assumption made an experiment was carried out with a group of analogues for predicting the H_{500} pressure field for the seventh day. The synoptic-statistical scheme proposed in [9] was used for this purpose. The program for computing the predictant field includes expansion of the set of predictor fields in natural orthogonal functions with subsequent finding of the matrix of empirical correlation functions (a_{hj}) between the temporal components of this expansion (T_{hi}) and elements of the predicted field. Computations of the pressure field (H_{500}) for the seventh day ($H_{ij,7}$) were made using the dependence

$$H_{ij,7} = \overline{H}_{ij,7} + \sum_{h=1}^{h} a_{hj} \{H_7,\ T_h(H_3)\}\, T_{hi}(H_3), \tag{3}$$

where H_3 is the H_{500} predictor field for the third day. Prior to each computation of the H_{500} prognostic field for the seventh day the initial series was reformed applying the principle of similarity of predictor fields and the corresponding prognostic chart for 48 hours and on the right side we had the H_{500} fields shifted by another 5 (a total of 7) days forward. During the period from 15 January through 15 February 1978 we prepared 11 forecasts of the H_{500} field by the proposed method. The results of the test are presented in Table 3. The ρ evaluation was computed from the coincidence of the predicted and actual signs of change in H_{500} from the initial to the seventh days. The relative error ε is the ratio of the absolute error in the methodological forecast to the error of the inertial (H_{500} of the initial day) forecast.

During the testing period ρ was quite high -- 0.55 (77.5% of the coincidences), the average relative error was 0.71. The results considerably exceed the probable success of forecasts of changes in the pressure field prepared for approximately the same advance time for the tendency of the future natural synoptic period (ρ = 0.36 and ε = 0.89) by the method of 1962 [14]. We note that modern hydrodynamic schemes for predicting the pressure fields for the fifth, sixth and seventh days on the average have a relative error $\varepsilon \geqslant 0.80$. The advantage of the forecasts obtained using synoptic-statistical forecasting, in which the composite analogue serves as the initial mass of data, agrees with the conclusions presented in a study by G. V. Gruza and R. G. Reytenbakh [5]. They analyzed the dependence of the success of forecasts on their advance time for hydrodynamic schemes and forecasts by the analogue method. In short-range forecasts (1-2 days) the advantage remains on the first day, whereas with an increase in the advance time the errors in forecasts by the analogue method increase far more slowly than in hydrodynamic forecasts. On the third day their values are approximately identical, whereas for times more distant from the initial day forecasts by the analogue method on the average have lesser errors.

In the proposed scheme each of the methods is used specifically in the time range in which it has advantages.

For example, in computations by the scheme

Initial day	Prediction for 48 hours	Prediction for 7th day
(1 January)	(3 January)	(8 January)

the first step is taken using hydrodynamic forecasting for 48 hours [18] and then the principle of forecasting by the analogue method in combination with regression analysis is applied. It is of interest to carry out a qualitative evaluation of forecasts of the development of synoptic processes over Europe and Western Siberia in the course of the investigated month (from 15 January through 15 February 1978) in the example of four periods not overlapping in time. An increase in zonality was actually observed during the period 17 through 24 January. The forecast for 24 January clearly reflected this restructuring. Between 24 and 31 January there was a growth of the ridge over the European USSR and a deepening of the troughs over Western Europe: the "Zon" form was replaced by the "C" form. The pressure field forecast for the seventh day (for 31 January) did not give a similar transformation; it simply indicated a general increase in geopotential heights. True, the regions of the predicted and actual growth maxima coincided, but in general the forecast must be considered unsuccessful. In the next period --

from 31 January -- the ridge over the European USSR intensified and the trough moved from the Urals to Western Siberia. The general sense of the process is reflected in the forecast of 6 February, but there are distortions. And in the course of the last week, from 7 through 14 February, there was a new increase in zonality; the forecast for 14 February was good.

We will allow ourselves to express some conclusions from these individual forecasts which have a formally high evaluation (Nos 2, 4, 8, 10, Table 3). In computing the pressure fields for times beyond the limits of the effect of the inertia of the initial state it is possible to predict zonal processes with success and in all probability, satisfactorily predict the evolution of meridional processes. But the prediction of a changeover of circulation from a zonal state to a meridional state for the time being remains beyond the limits of our capabilities.

Despite the mentioned shortcomings, the forecasting method proposed in this article is of some practical interest. In particular, the prognostic regions of increase and decrease in H_{500} from the first to the seventh days, whose geographic position in most cases corresponds to the actual position, should be used in validating the postulated changes in weather in ten-day periods.

The relative successes attained in this study in prediction of the H_{500} pressure field with a seven-day advance period and also improvement in the results of prediction of the temperature anomaly for five days must be attributed to solution of the regression equations in the limits of a composite analogue ensuring close dependences between meteorological fields. The rational use of the spatial and temporal correlations necessary for the forecast is accomplished by means of centering of the initial sample of processes not in mean standard, but in current synoptic processes.

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PARAMETERIZATION OF THE SURFACE LAYER WITH A VARIABLE HEIGHT

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[Article by A. Ye. Aloyan, candidate of physical and mathematical sciences, D. L. Yordanov, doctor of physical and mathematical sciences, and V. V. Penenko, doctor of physical and mathematical sciences, Computation Center, Siberian Department, USSR Academy of Sciences, manuscript submitted 1 Apr 80]

[Text]

Abstract: The article describes a method for parameterizing the surface layer with a variable height and sets forth a numerical method for its implementation on an electronic computer. Examples of computations of parameters of the surface layer with constant and variable heights are given.

In solving many practical problems in boundary layer dynamics it is necessary to make a detailed allowance for the processes transpiring in the surface layer (SL) and determining the nature of the interaction between the atmosphere and the underlying surface. This includes problems relating to the microclimate of cities, problems relating to study of the transport of impurities in the atmosphere, problems in local weather forecasting, and others. In numerical models the lower level, for which the values of the meteorological elements are computed (wind velocity, temperature, pressure, humidity), is usually situated at the height of the surface layer and the influence of the surface layer itself is parameterized. In most methods for parameterizing the surface layer known at the present time [4, 5, 8, 10] it is assumed that the height of the surface layer is fixed. In actuality, however, it varies in dependence on the meteorological conditions. The influence of these changes on large-scale atmospheric processes is not great, but with the formation of local processes they play a significant role. For example, the distribution of contaminating impurities in the boundary layer is dependent on the relative positioning of the height of the surface layer and the heights of the sources of impurities.

In this article we examine a method for parameterizing the surface layer in which height is dependent on stratification and is determined together with other characteristics. The basis of the proposed method is Monin-Obukhov similarity theory. Its algorithmic realization is accomplished within the framework of a complex of models of local atmospheric processes. An analysis of the results of numerical experiments indicates that allowance for the spatial-temporal changes in height of the surface layer makes possible a more detailed description of the region of constant vertical fluxes of momentum, heat and moisture.

The vertical scalar characteristics of the turbulent regime in a stationary barotropic planetary boundary layer (PBL) over a horizontally uniform underlying surface are unambiguously determined by the internal scales of height of the PBL and the Monin-Obukhov length scale:

$$H=\frac{\varkappa u_*}{l};\ L=-\frac{c_p \rho u_*^3}{\varkappa\beta q_0}, \qquad (1)$$

where $\varkappa$ is the Karman constant, u_* is friction velocity, l is the Coriolis parameter, c_p is the heat capacity of air at a constant pressure, ρ is air density, $\beta = g/\vartheta$ is the buoyancy parameter, g is the acceleration of free falling, ϑ is the potential temperature averaged vertically, q_0 is the vertical heat flow at the ground. [Note: The vector characteristics are also dependent on the angle α between the directions of the geostrophic and surface winds and the α angle is determined from a PBL model.]

For a stratification close to adiabatic ($|\mu|<10$), it follows from (1) that when $L \to \infty$ the height of the surface layer, which we denote by h, is determined only by the height scale H, that is

$$h=a_1 H \quad \text{with} \quad |\mu|<10, \qquad (2)$$

where $\mu = H/L$ is an internal parameter of PBL stratification.

In an investigation of a nonadiabatic case use is made of the following assumption relative to height of the BL. For a highly stable stratification due to the smallness of height of the BL we will assume that the influence of Coriolis force is insignificant and h is proportional only to L [11]:

$$h=a_2 L \quad \text{with} \quad 10\leqslant\mu<100. \qquad (3)$$

In the case of a highly unstable stratification the dynamic factors exert a weak influence on formation of height of the BL and therefore it can be assumed that it is not dependent on u_*. Then it follows from similarity theory that in this case

$$h=a_3 H^{3/2}|L|^{-1/2} \quad \text{with} \quad -400<\mu\leqslant-10. \qquad (4)$$

The inequality $|\mu|<10$ corresponds to conditions close to adiabatic. The limitations $10\leqslant\mu<100$ for equation (3) with stable and $-10\geqslant\mu>-400$ with unstable stratification are attributable to the fact that according to experimental data [7] more extremal values are not observed under nonadiabatic conditions. The constants $a_1 = 0.03$, $a_2 = 0.28$ and $a_3 = 0.01$ were determined from (2)-(4) in [1-2] as a result of integration of the equations of motion and comparison with experimental data on universal functions in the resistance law.

On the basis of the Monin-Obukhov similarity theory it is possible to obtain the following universal profiles of wind velocity and potential temperature:

$$\varkappa \frac{u}{u_*} = f_u(\xi, \xi_0); \quad \frac{\vartheta-\vartheta_0}{\vartheta_*} = \frac{(\vartheta-\vartheta_0)\varkappa^2\beta L}{u_*^2} = f_\vartheta(\xi, \xi_0), \tag{5}$$

where u and ϑ are wind velocity and potential temperature at the altitude z, $\vartheta_* = u^2/\varkappa^2\beta L$ is the temperature scale, f_u, f_ϑ are universal functions of dimensionless altitude $\xi = z/L$ and roughness $\xi_0 = z_0/L$.

For determining the turbulent exchange coefficients we use the expressions

$$k_m = \frac{\varkappa u_* z}{\varphi_u(\xi)}, \quad k_\vartheta = \frac{\varkappa u_* z}{\varphi_\vartheta(\xi)}, \quad \text{where} \quad \varphi_u(\xi) = \xi\frac{\partial f_u(\xi, \xi_0)}{\partial \xi}, \quad \varphi_\vartheta(\xi) = \xi\frac{\partial f_\vartheta(\xi, \xi_0)}{\partial \xi}, \tag{6}$$

where k_m and k_ϑ are turbulent exchange coefficients for the components of the velocity vector and temperature respectively.

If it is assumed that there is an analogy between the transport of moisture and temperature, and if for humidity we write the same expressions as for temperature, and the temperature scale in them is replaced by the corresponding scale for humidity, then under these conditions allowance for humidity does not introduce fundamental changes into the parameterization model and into the scheme for its realization.

Expressions (5), (6) are the basis for different methods for parameterizing the BL in models of the planetary boundary layer. However, in virtually all parameterization methods it is assumed that the height of the BL is constant. In the exposition which follows we will remove this restriction on altitude and will assume it to be variable.

We will direct the x-axis along the vector of stress of turbulent friction. The angle between the geostrophic and surface wind will be determined using a model of the planetary boundary layer, to be used in describing processes in the atmosphere above the surface layer, that is, $z \gg h$. We will examine equation (5) for an adiabatic stratification ($L \to \infty$ or $|\mu| < 10$). It is known that in this case a change in the wind with altitude can be approximated by a logarithmic function. Accordingly, on the basis of (1), (2) for the altitude $z = h = a_1 H$ we obtain the equation

$$a_1 \varkappa^2 r_0 = \frac{h}{z_0} \ln \frac{h}{z_0}, \tag{7}$$

where

$$r = \frac{u_h}{l z_0} \quad \text{is the Rossby number.} \tag{8}$$

For a nonadiabatic stratification ($|\mu| \geqslant 10$) with $z = h$ equation (5) is written in the following way. Dividing u_h^2 by $\vartheta_h - \vartheta_0$ and using the dependences between L and ϑ_*, we obtain

$$\frac{1}{r_t} = \frac{1}{\xi_0} \frac{f_u^2(\xi_h, \xi_0)}{f_\vartheta(\xi_h, \xi_0)}, \tag{9}$$

where

$$r_t = \frac{z_0 \beta (\vartheta_h - \vartheta_0)}{u_h^2}, \tag{10}$$

r_i is the analogue of the Richardson number (z_0 participates in the length scale), $\xi_h = \frac{h}{L}$.

The dimensionless parameters r_0 and r_i in (7) and (8) unambiguously determine the turbulent regime of the surface layer and can be obtained from the model of the PBL for $z \gg h$. For a further examination it is convenient to introduce still another stratification parameter which we will determine using the expression

$$s = r_0 r_i = \frac{\beta(\vartheta_h - \vartheta_0)}{l u_h}. \qquad (11)$$

Thus, the problem is reduced to finding the vertical fluxes of momentum and heat at the ground surface or the dimensionless parameters of the BL

$$\frac{u_*}{u_h} = \frac{\sqrt{\tau_0}}{\sqrt{\rho}\, u_h}; \quad \frac{L}{z_0} = -\frac{c_p \rho u_*^3}{\varkappa g z_0 q_0} \qquad (12)$$

in dependence on the parameters r_0 and r_i. In this case it is also necessary to find the altitude of the BL, determined by expressions (2)-(4) and (12).

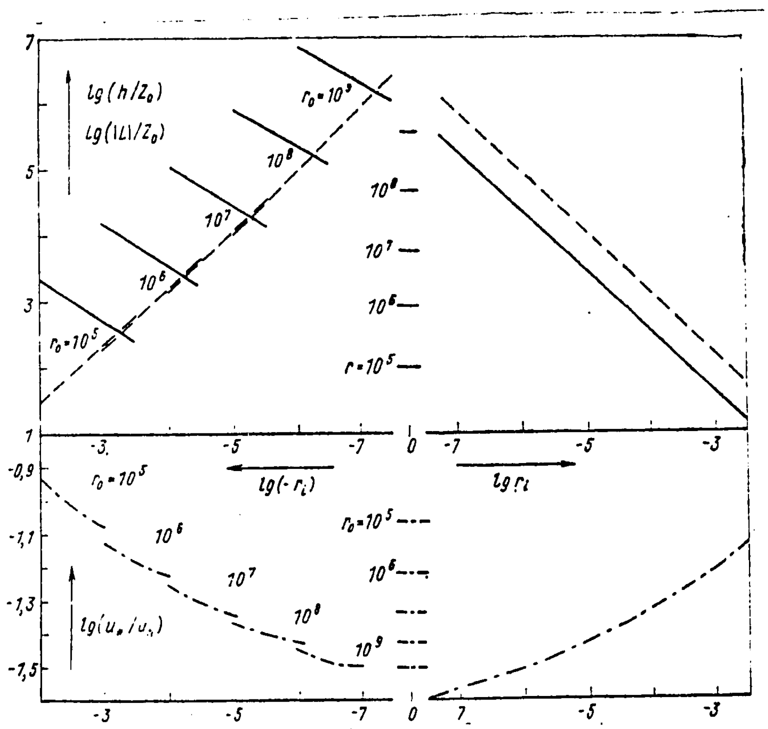

Fig. 1. Functions L/z_0 (dashed line), h/z_0 (solid lines) and u_*/u_h (dot-dash lines) in dependence on r_0 and ri on the assumption that $f_u(\xi, \xi_0) = f_\vartheta(\xi, \xi_0)$.

The nature of the dependence between the parameters of the surface layer (8), (10) and (12) will be demonstrated in a simple model. First we will examine an adiabatic case. Using equation (7) we will determine the dimensionless altitude h/z_0 as a function of the Rossby number r_0 and the drag coefficient $c_u = u_*/u_h$ is found from equations (7) and (2). The curves for these functions with $r_i = 0$ are shown in Fig. 1. In order to find the correlation between the parameters (12) and (8), (10), determining the turbulent regime under nonadiabatic conditions $|\mu| \geqslant 10$, it is necessary that the form of the universal functions (5) be made more specific. As is well known, there are a number of empirical models of the BL [1, 5, 9]. It therefore makes sense to analyze the possibility of using different formulas in parameterization of the surface layer with a variable height. As an example we will examine the universal functions from [5]:

$$f_u(\xi, \xi_0) = f_\theta(\xi, \xi_0) = \begin{cases} \ln \frac{z}{z_0} + \beta_1 \frac{z}{L}, & \frac{z}{L} \geqslant 0 \\ \ln \frac{z}{z_0}, & \beta_2 \leqslant \frac{z}{L} < 0 \\ \ln \frac{\beta_2 L}{z_0} + 3\left[1 - \left(\frac{\beta_2 L}{z}\right)^{1/3}\right], & \frac{z}{L} \leqslant \beta_2. \end{cases} \tag{13}$$

Here $\beta_1 = 10$, $\beta_2 = -0.07$.

The structure of the functions (13) makes it possible to use equation (5) to find the correlation between the parameter (11) and the internal parameter

$$s = r_0 r_i = \frac{u_*}{\varkappa lL}. \tag{14}$$

The internal parameter of PBL stratification μ is related to the external parameter s of the surface layer by the expression

$$\mu = \varkappa^2 s = \varkappa^2 r_0 r_i. \tag{15}$$

Substituting (14) into (3) and (4), we obtain

$$\frac{h}{z_0} = a_2 \frac{L}{z_0} \quad \text{with} \quad 10 \leqslant \varkappa^2 r_0 r_i < 100,$$

$$\frac{h}{z_0} = a_3 \varkappa^{3/2} (r_0 |r_i|)^{3/2} \frac{|L|}{z_0} \quad \text{with} \quad -400 < \varkappa^2 r_0 r_i \leqslant -10. \tag{16}$$

The second expression between the external and internal parameters of the BL is obtained as a result of substitution of (13) into (5) with the use of (14) and (16). Carrying out the necessary transformations, we arrive at the equations

$$\frac{1}{r_i} = \frac{L}{z_0} \left(\ln \frac{L}{z_0} + \ln (a_2 \cdot 10) + a_2 \beta_1 \right) \quad \text{when} \quad 10 \leqslant \varkappa^2 r_0 r_i < 100, \tag{17}$$

$$\frac{1}{r_i} = \frac{L}{z_0} \left(\ln \frac{|L|}{z_0} + \ln |\beta_2| + 3 - 3 \frac{|\beta_2|^{1/3}}{a_3 \sqrt{\varkappa r_0 r_i}} \right) \quad \text{with} \quad -400 < \varkappa^2 r_0 r_i \leqslant -10. \tag{18}$$

The dependence of the dimensionless Monin-Obukhov scale L/z_0 on the analogue of the Richardson number ri with different values of the Rossby number is shown in Fig. 1 by a dashed line. In this same figure we have also shown the dimensionless height of the BL (discontinuous curve) as a function of r_i with different r_0, computed using (16)-(18). In the lower part of Fig. 1 the dot-dash line shows the dependence of the drag coefficient u_*/u_h on r_i and r_0. In the case of an unstable stratification all the curves are given with values of the parameters in the range $-400 < \chi^2 r_0 r_i \leq -10$, and in the case of a stable stratification -- in the interval $10 < \chi^2 r_0 r_i < 100$.

Above we examined the simple case of parameterization of the BL when there was assumed to be an analogy between heat transfer and momentum or the Reynolds analogy $f_u(\xi, \xi_0) = f_\vartheta(\xi, \xi_0)$. In this case the equations (5) for momentum and heat are found to be independent and the correlations (14), (17) and (18) between the characteristics of the stratified BL can be obtained directly.

Now we will examine the more general problem when there is no Reynolds analogy. Using (8) and (10), we reduce equations (5) to the form

$$F_1 = |r_i| - \frac{c_u^2}{\varkappa} |\xi_0| f_\vartheta(\xi_h, \xi_0) = 0, \tag{19}$$

$$F_2 = \frac{1}{c_u} - \frac{1}{\varkappa} f_u(\xi_h, \xi_0) = 0. \tag{20}$$

Replacing ξ_h by a new expression from (10), we rewrite (19), (20) in the form of a system of functional equations

$$F_1(\xi_0, c_u, r_0, r_1) = 0, \tag{21}$$

$$F_2(\xi_0, c_u, r_0, r_1) = 0, \tag{22}$$

where r_0 and r_1 are stipulated, and ξ_0 and c_u are the sought-for parameters, and for solving the derived system we use the method described in [3]. For this purpose we will examine a system of equations linearized in the neighborhood

$$F_1(\vec{X}_k) - \frac{\partial F_1}{\partial \xi_0}(\xi_0 - \xi_{0k}) + \frac{\partial F_1}{\partial c_u}(c_u - c_{uk}) = 0, \tag{23}$$

$$F_2(\vec{X}_k) - \frac{\partial F_2}{\partial \xi_0}(\xi_0 - \xi_{0k}) + \frac{\partial F_2}{\partial c_u}(c_u - c_{uk}) = 0. \tag{24}$$

We introduce the notations

$$A = \left\| \begin{matrix} \frac{\partial F_1}{\partial \xi_0} & \frac{\partial F_1}{\partial c_u} \\ \frac{\partial F_2}{\partial \xi_0} & \frac{\partial F_2}{\partial c_u} \end{matrix} \right\|, \quad \vec{X} = \left\| \begin{matrix} \xi_0 \\ c_u \end{matrix} \right\|, \quad \vec{X}_k = \left\| \begin{matrix} \xi_{0k} \\ c_{uk} \end{matrix} \right\|, \quad \vec{F}_k = \left\| \begin{matrix} F_1(\vec{X}_k) \\ F_2(\vec{X}_k) \end{matrix} \right\| .$$

and we rewrite (23)-(24) with the use of these notations:

$$\vec{F}_k + A(\vec{X} - \vec{X}_k) = 0. \tag{25}$$

If the A matrix is not singular, it is possible to obtain the solution (25) in the form

$$\vec{X}=\vec{X}_k-A^{-1}\vec{F}_k. \tag{26}$$

Finally the solution of equation (26) is found using an iteration scheme in the Newton method

$$\vec{X}_{k+1}=\vec{X}_k-A^{-1}\vec{F}_k \; (k=1,\ 2,\ \ldots). \tag{27}$$

Stipulating the initial approximation, the iterations will be continued to satisfaction of the inequalities

$$\begin{aligned} &|\xi_{0k+1}-\xi_{0k}|<\varepsilon|\xi_{0k}|,\\ &|c_{uk+1}-c_{uk}|<\varepsilon|c_{uk}|, \end{aligned} \tag{28}$$

where $\varepsilon > 0$ is a stipulated number, determining the relative error of the iterations. The approximate solution, satisfying the convergence conditions, will be denoted $\tilde{X}$.

For tabulating the dependences between ξ_0, c_u and r_0, r_1 we will use the iteration procedure (27)-(28) and the perturbations method for evaluating the initial approximations. The cycle of the computation algorithm consists of the following. Assuming r_0, r_1 to be unperturbed values, we will give them some increments Δr_0 and Δr_1 respectively. Then the sought-for vector obtains some increment $\Delta\vec{X}=(\Delta\xi_0, \Delta c_u)$ in the neighborhood of the value $\tilde{X}$. For computing the initial approximation in solving the problem with perturbed values of the parameters we will linearize the functions F_1, F_2 in the neighborhood of the unperturbed values r_0, r_1, $\tilde{X}_k$. As a result we arrive at the system of equations

$$B\psi+A\,\Delta\vec{X}=\vec{\tilde{F}}, \tag{29}$$

where

$$B=\left(\frac{\partial F_j}{\partial r_\alpha}\right),\ j=1,2;\ \alpha=0,1; \tag{30}$$

$$\psi=(\Delta r_0,\ \Delta r_1)^T;\ \vec{\tilde{F}}=(F_1(\vec{\tilde{X}},\ r_0,r_1),\ F_2(\vec{\tilde{X}},r_0,r_1))^T.$$

We solve the derived system relative to $\Delta\xi_0$ and Δc_u:

$$\begin{pmatrix}\Delta\xi_0\\ \Delta c_u\end{pmatrix}=A^{-1}\left|\begin{matrix}\frac{\partial F_1}{\partial r_0}\Delta r_0+\frac{\partial F_1}{\partial r_1}\Delta r_1\\ \frac{\partial F_2}{\partial r_0}\Delta r_0+\frac{\partial F_2}{\partial r_1}\Delta r_1\end{matrix}\right|. \tag{31}$$

The initial approximation is computed using the formula

$$\vec{X}_1^{(0)}=\vec{\tilde{X}}+\Delta\vec{X} \tag{32}$$

and we will carry out the iterations (27) to convergence.

For parameterization of the BL by means of a numerical scheme we used universal Businger functions, etc. [4, 5, 9]

$$f_u(\xi, \xi_0) = 3\eta_u(\xi')\left[1 - \left(\frac{\xi'}{\xi}\right)^{1/3}\right] + \Phi_u(\xi') - \Phi_u(\xi_0), \tag{33}$$

$$\alpha_0 f_\vartheta(\xi, \xi_0) = 3\eta_\vartheta(\xi')\left[1 - \left(\frac{\xi'}{\xi}\right)^{1/3}\right] + \Phi_\vartheta(\xi') - \Phi_\vartheta(\xi_0)$$

in the case of unstable stratification and

$$f_u(\xi, \xi_0) = \ln\frac{\xi}{\xi_0} + \beta_u(\xi - \xi_0), \tag{34}$$

$$\alpha_0 f_\vartheta(\xi, \xi_0) = \ln\frac{\xi}{\xi_0} + \beta_\vartheta(\xi - \xi_0)$$

in the case of stable stratification, where

$$\Phi_u(\xi) = \ln(-\xi) + 2\operatorname{arctg}(\eta_u) - \ln[(1 + \eta_u^2)(1 + \eta_u)^2\eta_u^{-4}], \tag{35}$$

$$\Phi_\vartheta(\xi) = \ln(-\xi) - 2\ln[(1 + \eta_\vartheta)\eta_\vartheta^{-1}]$$

and

$$\eta_u(\xi) = (1 - \gamma_u\xi)^{-1/4}; \quad \eta_\vartheta = (1 - \gamma_\vartheta\xi)^{-1/2}, \tag{36}$$

and the constants have the values $\gamma_u = 15$, $\gamma_\vartheta = 9$, $\beta_u = 4.7$, $\beta_\vartheta = 6.34$, $\chi = 0.35$, $\alpha_0 = 1.35$, $\xi' = -8.21$.

The results of computations using the scheme (21)-(36) are given in Fig. 2. An analysis of the numerical solution shows that when using universal Businger functions the $|L|$ and h parameters decrease, and the drag coefficient increases in comparison with similar results obtained using this method with universal functions in the form cited in [1].

Now we will examine a scheme for applying the proposed parameterization method for the surface layer within the framework of a model of the boundary layer. Using a model of the boundary layer with known values for z = h we determine the Rossby number r_0, the analogue of the Richardson number r_i and the length scale L. Then, using the known values r_0, r_i, L from formulas (16) or from the curves represented in Figures 1 or 2, the height of the surface layer is determined, which in the next time interval is used as the lower boundary in the boundary layer model. At the initial moment in time the height of the surface is assumed to be constant and equal to some $\tilde{h}$ value. In actual practice it is convenient to assume that $\tilde{h}$ = 50 m.

As an illustration of the influence of changes in the height of the surface layer we will demonstrate some results of computations made using the numerical model set forth in [6]. Without citing the equations of this model, we will only describe it briefly. This is a two-dimensional nonstationary model of a mesoscale boundary layer based on a system of full equations in hydrothermodynamics in a quasistatic approximation. As the vertical coordinate we used the function $\xi = (z - h(x, t))/(H-h(x,t)$ h(x,t) is surface layer height; here H is the upper boundary of the boundary layer, x is a horizontal curvilinear coordinate reckoned along the level of the height of the surface layer. Temperature inhomogeneities of the underlying surface are taken into account using the heat balance equation. The roughness parameter over the land is stipulated, whereas over the water it is computed using the Chernok formula. The model also takes into account turbulent exchange. The coefficients of horizontal and vertical turbulence are assumed to be stipulated. We note that in the experiments we used a two-dimensional model only from considerations of convenience in its realization on an electronic computer.

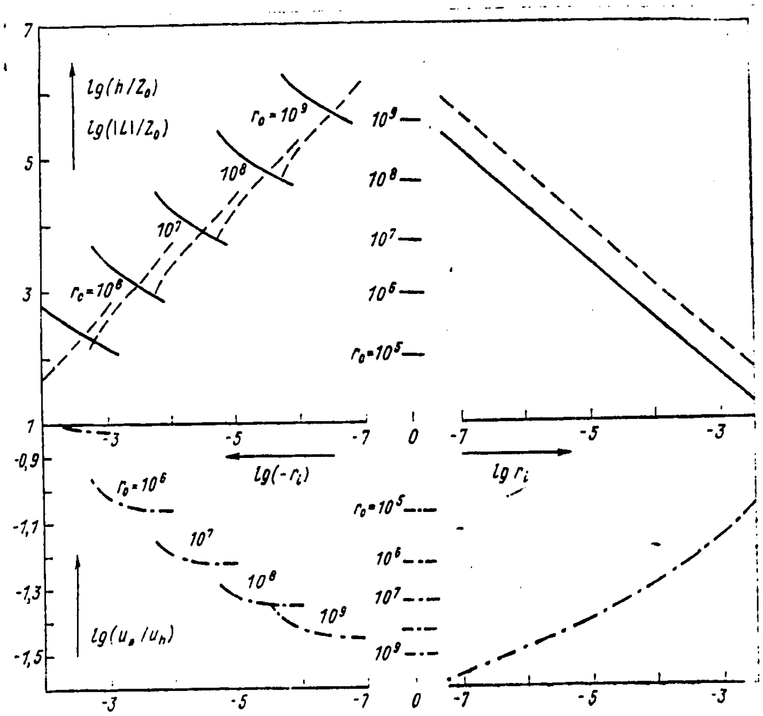

Fig. 2. Same as in Fig. 1, on the assumption that $f_u(\xi, \xi_0) \neq f_\vartheta(\xi, \xi_0)$.

Now we will examine two examples of solution of the problem of a breeze over an island. In the first it is assumed that the height of the BL is constant (h = 50 m), and in the second -- variable. In both cases the computations were made on the assumption that a large-scale background flow is absent with the following values of the input parameters: number of points of intersection of a grid region vertically -- 30, horizontally -- 56, spatial interval -- $\Delta x = 5 \cdot 10^3$ m, time interval -- $\Delta t = 20$ min, boundary layer height H = 2.5 km, stability parameter -- $S = 3 \cdot 10^{-3}$°C/m, convection parameter -- $\lambda = 3.5 \cdot 10^{-2}$ m/(c^2·°C), Coriolis parameter -- $\ell = 10^{-4}$ sec^{-1}, coefficient of horizontal turbulence -- $\mu = 5 \cdot 10^3$ m^2/sec; the coefficients of vertical turbulence for velocity and temperature were assumed to be constant and equal to their values at the height of the surface layer, computed using formulas (6). Figure 3 shows isolines of potential temperature at the time t = 12 hours. The solid curves represent the profiles of potential temperature obtained in the parameterization of a BL with a constant height, whereas the dashed lines are for parameterization of the BL with a variable height. The graph of change in the height of the BL for the time t = 12 hours is shown as Fig. 4.

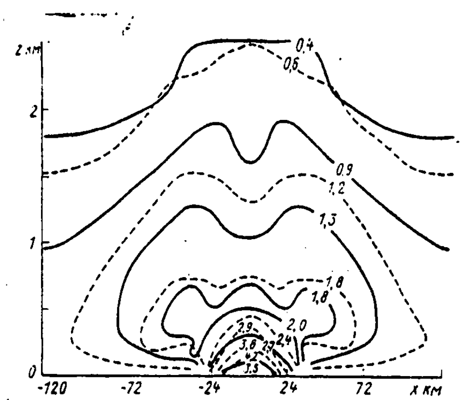

Fig. 3. Isolines of potential temperature with a constant (solid curves) and variable (dashed curves) height of the surface layer at the time t = 12 hours.

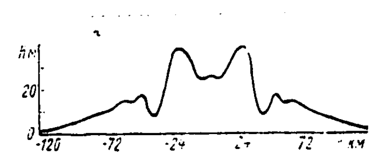

Fig. 4. Height of the surface layer at the time t = 12 hours.

The method for parameterization of the BL proposed in this article for a boundary layer of variable height makes possible a more rational application of the Monin-Obukhov similarity theory. If we examine a BL model with a boundary layer of fixed height, then with stable and neutral stratifications it is possible to obtain an artificial increase in the region of constant turbulent fluxes, which leads to an exceeding of the limits of applicability of the Monin-Obukhov theory, and vice versa, in the case of a highly unstable stratification, when the constancy of the fluxes persists to far greater heights, the possibilities of this theory will not be completely used.

In the ordinary method for parameterization of the BL, the vertical fluxes are usually in dependence on the preselected dimensionless height and on the Richardson number, in which this height is used for determining the length scale. In accordance with the method described above for stipulating the external parameters it is more convenient to select the Rossby number (8) and the modified Richardson number (10), and use the roughness parameter for determining the length scale.

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UDC 551.(511+515.2)

MOTION OF A PARTICLE ON A ROTATING SPHERE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 47-53

[Article by L. Kh. Ingel', candidate of physical and mathematical sciences, and Ye. R. Lepikash, Institute of Experimental Meteorology, manuscript submitted 17 Apr 80]

[Text]

Abstract: The results of an article by A. S. Monin entitled "Inertial Motions on a Rotating Sphere" [2] are generalized for the case of presence of meridional forces arbitrarily dependent on latitude. A general analytical solution of the corresponding nonlinear problem is found. As an example, the author examines a model of the motion of a tropical cyclone under the influence of the Rossby effect.

An analytical solution of the problem of inertial motions of a particle on a rotating sphere was found in [2] (also see the bibliography to that article). In our article this result is generalized for a case when, in addition to inertial forces, the particle is acted upon by other meridionally directed forces arbitrarily dependent on latitude. A geophysical application of this problem is, for example, a model of movement of a tropical cyclone (TC) as a rotating solid body of relatively small size (for example, see [3, 9]).

We will examine a system of equations of particle motion:

$$\frac{du}{dt} - \frac{uv\,\mathrm{tg}\,\varphi}{R} - 2\omega v \sin\varphi = 0, \tag{1}$$

$$\frac{dv}{dt} + \frac{u^2\,\mathrm{tg}\,\varphi}{R} + 2\omega u \sin\varphi = F(\varphi). \tag{2}$$

This system differs from the system of equations in Lagrangian variables [2] examined by A. S. Monin [2] in the addition of the arbitrary force $F(\varphi)$ in (2)*. Here u, v are the velocity components of a particle in the directions to the east and north respectively.

$$u = R\frac{d\lambda}{dt}\cos\varphi, \qquad v = R\frac{d\varphi}{dt}, \tag{3}$$

where φ is geographic latitude, λ is the longitude of the particle, ω is the angular velocity of rotation of the sphere, R is its radius, t is time. [In equation (2), generally speaking, the centrifugal acceleration associated with the earth's rotation is lacking. This term is usually not taken into account on the assumption that it is compensated by the horizontal component of gravity. Accordingly, it is more correct to speak of motion not on a sphere, but on a flattened ellipsoid -- the "normal figure of the earth."]

As the initial conditions we will assume that the coordinates and velocity components of the particle λ_0, φ_0, u_0, v_0 at the time t = 0 are stipulated.

System (1), (2), despite its nonlinearity, can be integrated analytically in general form. We will rewrite (1), (2) with (3) taken into account in the form

$$\frac{du}{dt} - u \operatorname{tg} \varphi \frac{d\varphi}{dt} = 2\omega R \sin\varphi \frac{d\varphi}{dt}, \tag{4}$$

$$R\frac{d^2\varphi}{dt^2} = -\frac{u^2 \operatorname{tg}\varphi}{R} - 2\omega u \sin\varphi + F(\varphi). \tag{5}$$

Equation (4) can be formally integrated, regarding it as a linear equation of the first order relative to u:

$$u = e^{\int_0^t \operatorname{tg}\varphi \frac{d\varphi}{dt'} dt'} \left(C_1 + 2\omega R \int_0^t \sin\varphi \frac{d\varphi}{dt'} e^{-\int_0^{t'} \operatorname{tg}\varphi \frac{d\varphi}{dt''} dt''} dt'\right), \tag{6}$$

where C_1 is the integration constant. Expression (6) is easily simplified taking into account that

$$e^{\int_0^t \operatorname{tg}\varphi \frac{d\varphi}{dt} dt} = e^{\int_0^t \operatorname{tg}\varphi_0 d\varphi} = e^{-\ln\cos\varphi\big|_0^t} = \frac{\cos\varphi_0}{\cos\varphi}. \tag{7}$$

We obtain the result cited in [3]:

$$u = \frac{\cos\varphi_0}{\cos\varphi}\left[u_0 - \frac{\omega R}{\cos\varphi_0}(\cos^2\varphi - \cos^2\varphi_0)\right]. \tag{8}$$

In selecting the integration constant the initial conditions were taken into account.

Substituting (8) into (5), we obtain

$$\frac{d^2\varphi}{dt^2} = -B\frac{\operatorname{tg}\varphi}{\cos^2\varphi} + \frac{1}{2}\omega^2 \sin 2\varphi - \omega^2\cos^4\varphi_0 \operatorname{tg}\varphi + \frac{F(\varphi)}{R}, \tag{9}$$

where

$$B = \frac{u_0^2}{R^2}\cos^2\varphi_0 + \frac{2\omega u_0}{R}\cos^3\varphi_0.$$

We introduce the new variable

$$x = \frac{d\varphi}{dt}. \tag{10}$$

Then

$$\frac{d^2\varphi}{dt^2} = \frac{dx}{dt} = \frac{dx}{d\varphi}\frac{d\varphi}{dt} = x\frac{dx}{d\varphi}. \tag{11}$$

and from (9) we obtain an equation with the separable variables

$$x\frac{dx}{d\varphi} = \frac{F\varphi}{R} - B\frac{\operatorname{tg}\varphi}{\cos^2\varphi} + \frac{1}{2}\omega^2\sin 2\varphi - \omega^2\cos^4\varphi_0\operatorname{tg}\varphi. \qquad (12)$$

Separating the variables and integrating (12), we obtain

$$\frac{x^2 - \frac{v_0^2}{R^2}}{2} = F(\varphi) + B\left(\frac{\operatorname{tg}^2\varphi_0}{2} - \frac{\operatorname{tg}^2\varphi}{2}\right) + \frac{\omega^2}{4}(\cos 2\varphi_0 - \cos 2\varphi) + \omega^2\cos^4\varphi_0(\ln\cos\varphi - \ln\cos\varphi_0), \qquad (13)$$

where

$$F_1(\varphi) = \frac{1}{R}\int_{\varphi_0}^{\varphi} F(\varphi')\,d\varphi'$$

or

$$\frac{d\varphi}{dt} = \pm\sqrt{2F_1(\varphi) + B(\operatorname{tg}^2\varphi_0 - \operatorname{tg}^2\varphi) + D}$$

$$D = \frac{\omega^2}{2}(\cos 2\varphi_0 - \cos 2\varphi) + \omega^2\cos^4\varphi_0(\ln\cos\varphi - \ln\cos\varphi_0) + v_0^2/R^2. \qquad (14)$$

Integrating (14), we obtain

$$t = \int_{\varphi_0}^{\varphi}\frac{d\varphi}{\psi(\varphi)}, \qquad (15)$$

where $\psi(\varphi)$ is the right-hand side of (14).

Knowing $\varphi(t)$, from (8) and (3) we find $\lambda(t)$. Using (8), (3) and (14), it is easy to obtain an equation for the trajectory of the particle

$$\lambda - \lambda_0 = \frac{\cos\varphi_0}{R}\int_{\varphi_0}^{\varphi}\frac{\left[u_0 - \frac{\omega R}{\cos\varphi_0}(\cos^2\varphi - \cos^2\varphi_0)\right]d\varphi}{\cos^2\varphi\cdot\psi(\varphi)}. \qquad (16)$$

The problem in principle is solved.

As an example we will consider the model problem of the influence of the Rossby effect on the motion of a tropical cyclone.

Coriolis force evidently exerts an appreciable influence on the movement of atmospheric vortices of a sufficiently large scale [1, 4, 6, 7]. An important fact here is the dependence of Coriolis force on latitude, which leads to the Rossby effect: Coriolis forces opposite in sign (caused by rotation of the vortex), acting on the northern and southern parts of the vortex, are not equal in magnitude; accordingly, if the vortex is regarded as a relatively solid object, there will be a resultant which for cyclones is directed away from the equator.

It goes without saying that for describing the mentioned effect it is necessary to take into account the finiteness of the horizontal dimensions of the vortex. We will examine a very simple model which takes into account the Rossby effect, the movement of a finite ring rotating with a constant velocity on a rotating sphere. [A description of a tropical cyclone as a rotating axially symmetric body is frequently used in modeling the trajectory of a TC (for example, see [3, 5, 9]), but the limits of applicability of this model are unclear.]

The stipulation of a constant velocity of ring rotation leads to a nonconservation of its total kinetic energy. But for the situations considered in this article this nonconservation is relatively small (in comparison with the kinetic energy of ring rotation). Therefore, the requirement for conservation of energy would result only in a small variability of the velocity of ring rotation and to an insignificant change in the results.

It can be shown (see Appendix) that the variability of the Coriolis parameter, in combination with the finiteness of ring diameter leads to the appearance of an additional force (in comparison with the forces acting on the particle):

$$F_R(\varphi) = \alpha \cos \varphi, \qquad \alpha = \frac{\omega \omega_A A^2}{4R}, \tag{17}$$

where ω_A = const is the angular velocity of ring rotation, A is its diameter.

Now we will examine a particle additionally "loaded" by the force (17). In addition, we will assume that there is a meridional pressure gradient which as a simplification will be considered constant. The corresponding force (to be more precise, acceleration) will be denoted F_0 = const. Thus,

$$F(\varphi) = F_0 + \alpha \cos \varphi.$$

This specific problem allows two important simplifications. First, in (1), (2) it is possible to neglect terms which are quadratic relative to the velocity components, which, as is easy to check for real velocities of movement of TC, are much less than the corresponding components of Coriolis force. Second, below we will be able to confirm that for such velocities the trajectory of the center of the ring falls between two rather close circles of latitude. This makes it possible outside the narrow equatorial zone to replace the functions sin φ and cos φ in equations (1), (2) by the values of these functions at the point $\varphi = \bar{\varphi}$ corresponding to the middle of the region of movement. [In order to avoid misunderstandings, we note that the latter simplification does not mean neglecting of the Rossby effect. But the intensity of this effect is assumed to be constant in the entire region of ring movement.]

The sense of these simplifications can be expressed in the following way. The general solution found above represents nonlinear oscillations of a particle. However, it was found that with real values of the velocity of movement of a TC the oscillations fell between two close φ values for which the radicand in (14) becomes equal to zero, in other words, small oscillations are involved. Accordingly, the nonlinearity of equations (1), (2) in this case is poorly manifested (in particular, leading to a small change in the period of the oscillations).

After the mentioned simplifications we obtain the system

$$R \cos \bar{\varphi} \frac{d^2 \lambda}{dt^2} - Rf \frac{d\varphi}{dt} = 0, \tag{18}$$

$$R \frac{d^2 \varphi}{dt^2} + Rf \cos \bar{\varphi} \frac{d\lambda}{dt} = F_0 + \alpha \cos \bar{\varphi}, \tag{19}$$

where $f = 2\omega \sin \bar{\varphi}$; its solution, as is easy to check, has the form

$$\lambda = \lambda_0 + \frac{1}{fR\cos\bar{\varphi}}\left[\left(u_0 - \frac{F_0}{f} - \frac{\alpha\cos\bar{\varphi}}{f}\right)\sin ft + v_0(1-\cos ft) + (F_0 + \alpha\cos\bar{\varphi})\,t\right]; \tag{20}$$

$$\varphi = \varphi_0 + \frac{1}{fR}\left[\left(u_0 - \frac{F_0}{f} - \frac{\alpha\cos\bar{\varphi}}{f}\right)(\cos ft - 1) + v_0 \sin ft\right]. \tag{21}$$

The determined trajectory represents oscillations with a period $2\pi/f$ in both horizontal directions superposed on zonal movement with the constant velocity

$$\frac{F_0 + \alpha\cos\bar{\varphi}}{f}.$$

The observed tendency to a drift of tropical cyclones away from the equator is frequently associated with the Rossby effect (since the corresponding term in the equation describes a force always directed away from the equator, for example, see [7]). But in the solution found the mean meridional drift is absent.

Assume that at the initial moment the velocity of the center of the ring coincides with the velocity of the geostrophic wind ("steering current")

$$u_0 = -\frac{1}{Rf\rho}\cdot\frac{\partial p}{\partial\varphi} = \frac{F_0}{f}, \quad v_0 = 0, \tag{22}$$

where p is pressure, ρ is air density.

The solution in this case will be

$$\lambda = \lambda_0 + \frac{1}{fR\cos\bar{\varphi}}\left[-\frac{\alpha\cos\bar{\varphi}}{f}\sin ft + (F_0 + \alpha\cos\bar{\varphi})\,t\right]. \tag{23}$$

$$\varphi = \varphi_0 + \frac{\alpha\cos\bar{\varphi}}{f^2R}(1-\cos ft). \tag{24}$$

The Rossby effect (α is different from zero) leads to oscillations of the center of the ring relative to the zonal flow; the amplitude of the oscillations is

$$\Delta y = \left|\frac{\alpha\cos\bar{\varphi}}{f^2}\right|.$$

In addition, there is a drift relative to the background flow with the velocity $\Delta u = \alpha\cos\bar{\varphi}/f$ (in the northern hemisphere $\omega_A > 0, \alpha > 0$ and Δu is directed to the east). Assume that A = 100 km, $\omega_A = 10^{-3}$ sec^{-1}, $\varphi_0 = 15°$N, $F_0 = -2\cdot10^{-4}$ m/sec^2. Then $\alpha = 2.8\cdot10^{-5}$ m/sec^2, $f = 3.6\cdot10^{-5}$ sec^{-1}, $u_0 \approx -5$ m/sec, $\Delta y \approx 30$ km, $\Delta u \approx 1$ m/sec. The trajectory is an ordinary cycloid superposed on the background zonal flow and falls in the interval $\Delta\varphi \approx 2\Delta y/R < 1°$, which justifies the replacement of φ by $\bar{\varphi}$ made above in a number of cases.

If the velocity of the center of the ring at the initial moment "is not matched" with the steering current, then, as can be seen from (20), (21), there can be inertial oscillations of the trajectory with the amplitude

$$\Delta y \approx \frac{1}{|f|}\max\left\{\left|u_0 - \frac{F}{f}\right|, |v_0|, \left|\frac{\alpha\cos\bar{\varphi}}{f}\right|\right\},$$

considerably greater than in the case (22) and the formation of a loop.

The trajectory in the case $u_0 = v_0 = F_0 = 0$ is shown in Fig. 1. It is noteworthy that the ring at rest can acquire a nonzero velocity (to a value of the order $\alpha/f \sim 1$ m/sec) under the influence of the Rossby effect.

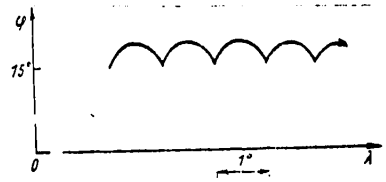

Fig. 1.

It must be remembered that the oscillations of the trajectory and loop discussed above are relatively small and often can be substantially smaller than the extent of the TC themselves. Such peculiarities of the trajectory can remain unnoticed without special observations. Incidentally, some "yawing" of the TC relative to its mean trajectory was already predicted theoretically in connection with other factors [9] and according to data in [8, 9] was actually observed.

The authors express appreciation to N. A. Bagrov, V. N. Ivanov and Yu. T. Sapronov for useful discussions.

Appendix

Derivation of Expression (17)

The Coriolis force acting on an arbitrary element of the ring (Fig. 2) in a meridional direction is

$$dF_\varphi = -2dm \cdot \omega U \sin \Phi, \tag{25}$$

where dm is the mass of a ring element, Φ, U is its latitude and the zonal component of its velocity respectively. The latter consists of the zonal components of the transfer movement of the center of the ring u and rotational movement

$$U = u - \frac{1}{2} \omega_A A \sin \theta \tag{26}$$

(θ is the azimuthal coordinate of the considered ring element read from the east counterclockwise). The latitude of this element, as can be seen easily, is

$$\Phi \simeq \varphi + \frac{A}{2R} \sin \theta . \tag{27}$$

We substitute (26) and (27) into (25):

$$dF_\varphi \simeq -2dm \cdot \omega \left(u - \frac{1}{2} \omega_A A \sin \theta\right) \times \left(\sin \varphi + \frac{A}{2R} \cos \varphi \sin \theta\right) \tag{28}$$

(the smallness of the second term on the right-hand side of (27) in comparison with the first term, that is, the smallness of the size of the ring in comparison with its distance from the equator, is used).

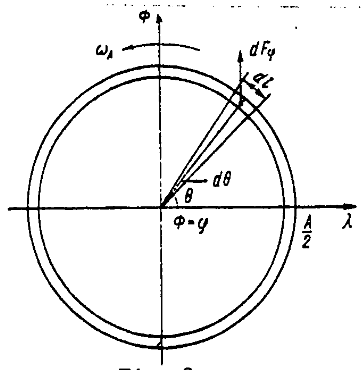

Fig. 2.

The mass of the ring element is

$$dm = \frac{M}{2\pi} d\theta,$$

where M is the total mass of the ring, $d\theta$ is the angular dimension of the ring element.

Integrating (28) for the angle θ from 0 to 2π, we find the meridional component of the Coriolis force acting on the ring. Dividing this value by the mass, we obtain the corresponding acceleration component a_φ. Simple computations give

$$a_\varphi \simeq -2\omega u \sin\varphi + \frac{\omega\omega_A A^2}{4R} \cos\varphi. \tag{29}$$

The first term on the right-hand side of (29) is traditionally taken into account in the equations and is related to movement of the ring as a whole. The second term is associated with the nonpoint character and rotation of the ring and reflects the influence of the Rossby effect.

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METHOD FOR COMPUTING THE HEAT FLUX INTO THE SOIL FROM TEMPERATURE

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 54-60

[Article by V. A. Gordin, candidate of physical and mathematical sciences, USSR Hydrometeorological Scientific Research Center, manuscript submitted 14 May 80]

[Text]

Abstract: The author proposes an economical method for computing soil response to change in the temperature regime of the surface based on use of a Z-transform and the Pade approximation. Numerical experiments which were carried out in which surface temperature, proportional to sin Ωt was stipulated and the heat flux into the soil was determined indicated the advantage of the proposed method in comparison with those known earlier.

1. In the modeling of heat exchange between the atmosphere and soil it is necessary to have a sufficiently economical and precise method for computing the response of the soil to a change in environmental conditions; in the simplest case it is necessary to determine the function $\partial T/\partial z$ $(z = 0, t)$, proportional to the heat flux into the soil, using the function $T(z = 0, t)$ [1, 4, 6].

The simplest model describing the temperature distribution in the soil is a linear thermal conductivity equation

$$\frac{\partial T}{\partial t} = \frac{\partial}{\partial z} D \frac{\partial}{\partial z} T \tag{1}$$

with a constant thermal conductivity coefficient D (on possible generalizations see below).

The boundary conditions for equation (1) on the ray $[-\infty, 0]$ are stipulated in the first kind:

$$T(z \to -\infty, t) = T_{-\infty}(t), \; T(0, t) = T_0(t). \tag{2}$$

In actual practice $T_{-\infty}(t)$ can be determined as

$$\frac{1}{\Delta t}\int_{t-\Delta t}^{t} T_0(\tau)\,d\tau,$$

where Δt greatly exceeds the characteristic time scales of our problem. In particular, assuming Δt to be equal to 1 year, we obtain: $T_{-\infty}(t) = T_{-\infty} =$ const.

In order for the problem (1)-(2) to have a unique solution, it is necessary to supplement it with data on the initial temperature profile.

In this case the solution of the Cauchy problem is determined in the form of the sum of the results of use of an integral operator for the variable z with a kernel dependent on t for the function T(z, 0) and an integral operator for the variable t with a kernel dependent on z for the function $T_0(t) - T_{-\infty}$. Since we are not interested in the behavior of the profile, but only $\partial T/\partial z(0, t)$, we will carry out a simplification of the operators, and in addition, we will neglect the contribution of the initial data to the solution, decreasing with time. Then the response is represented in the form

$$\frac{\partial T}{\partial z}(0,\ t) = (\pi D)^{-1/2}\int_0^t (t-\tau)^{-1/2}\frac{\partial}{\partial \tau}[T_0(\tau) - T_{-\infty}]\,d\tau. \tag{3}$$

2. Since the function $T_0(t)$ is known only at discrete moments in time and at these same moments in time it is necessary to compute $\partial T/\partial z(t)$, the integral in formula (3) must be replaced by a quadrature formula, which is nontrivial due to the singularity of the kernel.

In place of this we will replace the differential equation (1) by a differential-difference analogue (Crank-Nicholson scheme):

$$\frac{T^n - T^{n-1}}{\tau} = D\frac{\partial^2}{\partial z^2}\left(\frac{T^n + T^{n-1}}{2}\right), \tag{1'}$$

where the superscript denotes the number of the time interval. The absolute stability of equation (1') is easily checked.

We apply a $Z_{t\to 1/\omega}$ transform to equation (1') [2-3]. As in section 1, we will neglect the contribution of initial data to the solution. We have

$$\frac{2}{D\tau}\cdot\frac{1-\omega}{1+\omega}\,\tilde{T} = \frac{d^2\tilde{T}}{dz^2} \tag{4}$$

and after solution of this differential equation we find that

$$\frac{d\tilde{T}}{dz}(0,\ \omega) = \sqrt{\frac{2}{D\tau}\,\frac{1-\omega}{1+\omega}}\,(\tilde{T}_0 - \tilde{T}_{-\infty}),$$

and after application of the inverse Z^{-1} transform

$$\frac{\partial T^n}{\partial z}(0) = \sqrt{\frac{2}{D\tau}}\sum_{j=0}^{\infty}\omega_j\,(T_0^{n-j} - T_{-\infty}), \tag{5}$$

where ω_j are the coefficients of expansion of the function $f(\omega) = \sqrt{\frac{1-\omega}{1+\omega}}$ into a Taylor series at the point $\omega = 0$ (in powers of ω). For the ω_j values see Table 1. Other ω_j coefficients were proposed in [1].

It follows from Table 1 that the ω_j coefficients decrease rather slowly with an increase in j (much more slowly than proposed in [1]). This is attributable to the fact that the $f(\omega)$ function has branch points at $\omega = \pm 1$, and therefore the radius of convergence of the Taylor series is equal to 1.

ω_j Coefficients

Table 1

j	ω_j	j	ω_j	j	ω_j	j	ω_j
0	1,0000	6	0,3125	12	0,2256	17	—0,1964
1	—1,0000	7	—0,3125	13	—0,2256	18	0,1855
2	0,5000	8	0,2734	14	0,2095	19	—0,1855
3	—0,5000	9	—0,2734	15	—0,2095	...	...
4	0,5750	10	0,2461	16	0,1964	29	—0,1495
5	—0,5750	11	—0,2461				

Due to the slow decrease of the ω_j coefficients, in the practical application of formula (5) it is necessary to take into account many terms in the series, which in turn leads to large memory requirements and necessitates a large number of electronic computer operations.

3. In order to increase the effectiveness of formula (5) it is proposed, as in [2] (also see [5, 7]) that use be made of the Pade approximation of the function $f(\omega)$ at the point $\omega = 0$ with an inverse Z^{-1} transform:

$$f(\omega) = \frac{P(\omega)}{Q(\omega)} + O(\omega^{M_1+M_2+1}), \tag{6}$$

where

$$P(\omega) = \sum_{j=0}^{M_1} \omega^j p_j, \quad Q(\omega) = \sum_{j=0}^{M_2} \omega^j q_j, \quad p_0 = q_0 = 1.$$

The diagonal Pade approximation ($M_1 = M_2$) is used here because in [2] it gave the best results. [It is possible that in this problem nondiagonal approximations will be more effective, although this is improbable.]

Multiplying $f(\omega)$ by $Q(\omega)$, we obtain the evaluation

$$f(\omega)\,Q(\omega) = P(\omega) + O(\omega^{M_1+M_2+1})$$

and after an inverse Z^{-1} transform -- the computational recurrent formula

$$\frac{\partial T^n}{\partial z}(0) = (D\tau)^{-1/2} \sum_{j=0}^{M_1} p_j \,(T_0^{n-j} - T_{-\infty}) - \sum_{j=1}^{M_2} q_j \frac{\partial T^{n-j}}{\partial z}(0). \tag{7}$$

The p_j and q_j values for different M are given in Table 2.

Since formula (7), in contrast to formula (5), is recurrent, that is, if for the polynomial $Q(z) = \omega^M Q(1/\omega)$ there are roots in absolute value exceeding 1, the formula (7) is unstable, that is, with almost any $T_0^n \lim_{n\to\infty} \left|\frac{\partial T^n}{\partial z}(0)\right| = \infty.$

It follows from Fig. 1, which shows Q(z) curves with real z, that all Q(z) roots are real and fall in the interval (-1, +1), and accordingly, there is no instability. It follows from this same graph that the root which is maximum in absolute value is always negative and that this modulus is the greater the greater the M value. This means that the corresponding transient processes should be accompanied by oscillations and their characteristic time is the greater the greater the M value. A confirmation of this was obtained in test computations.

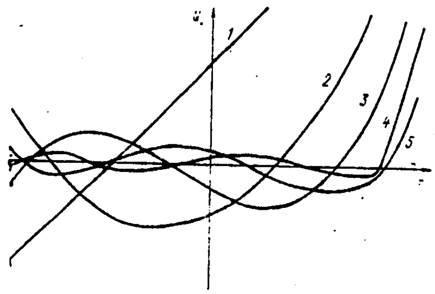

Fig. 1. $Q_i(z)$ polynomials with real z. The subscript is the degree of the polynomial. All the roots fall within a unit interval. These polynomials are evidently orthogonal in the interval [-1, 1] with some weight $d\mu(z)$ (see [5]).

Table 2

p_j and q_j Coefficients

j	p_j	q_j
	$M=1$	
1	−0,5000	0,5000
	$M=2$	
1	−0,5000	0,5000
2	−0,2500	0,2500
	$M=3$	
1	−0,5000	0,5000
2	−0,5000	−0,5000
3	0,1250	−0,1250
	$M=4$	
1	−0,5000	0,5000
2	−0,7500	−0,7500
3	0,2500	−0,2500
4	0,0625	0,0625
	$M=5$	
1	−0,5000	0,5000
2	−1,0000	−1,0000
3	0,3750	−0,3750
4	0,1875	0,1875
5	−0,03125	0,03125

4. Now we will examine the important special case when $T_0(t) - T_{-\infty} = \sin \Omega t$ with $t \geqslant 0$. Here the precise solution of the Cauchy problem with zero initial data is expressed through a generalized hypergeometric series. However, a stationary periodic solution is substantially simpler. In actuality, expanding the solution into a series $T(z,t) = T_{-\infty} + \sum_{-\infty}^{\infty} f_n(z) e^{i\Omega nt}$,

we obtain ordinary differential equations $D f''_n = i\Omega n f_n$ with the boundary conditions $f_n(-\infty) = 0$ with all $n \in Z$ and $f_n(0) = 0$ with $|n| \neq 1$, $f_{\pm 1}(0) = \pm 1/2\, i$. The solutions of these equations are $f_n(z) = 0$ with

$$|n| \neq 1,\ f_1(z) = \frac{1}{2i} \exp \sqrt[+]{\frac{i\Omega}{D}}\, z,$$

$$f_{-1}(z) = \frac{-1}{2i} \exp \sqrt[+]{\frac{-i\Omega}{D}}\, z,$$

where $\sqrt[+]{\ }$ denotes the root with a positive real part.

It therefore follows that

$$T(z, t) = T_{-\infty} + \exp\left(\sqrt{\frac{\Omega}{2D}}\, z\right) \sin\left(\Omega t + \sqrt{\frac{\Omega}{2D}}\, z\right)$$

and

$$\frac{\partial T}{\partial z}(0, t) = \sqrt{\frac{\Omega}{D}} \sin\left(\Omega t + \frac{\pi}{4}\right).$$

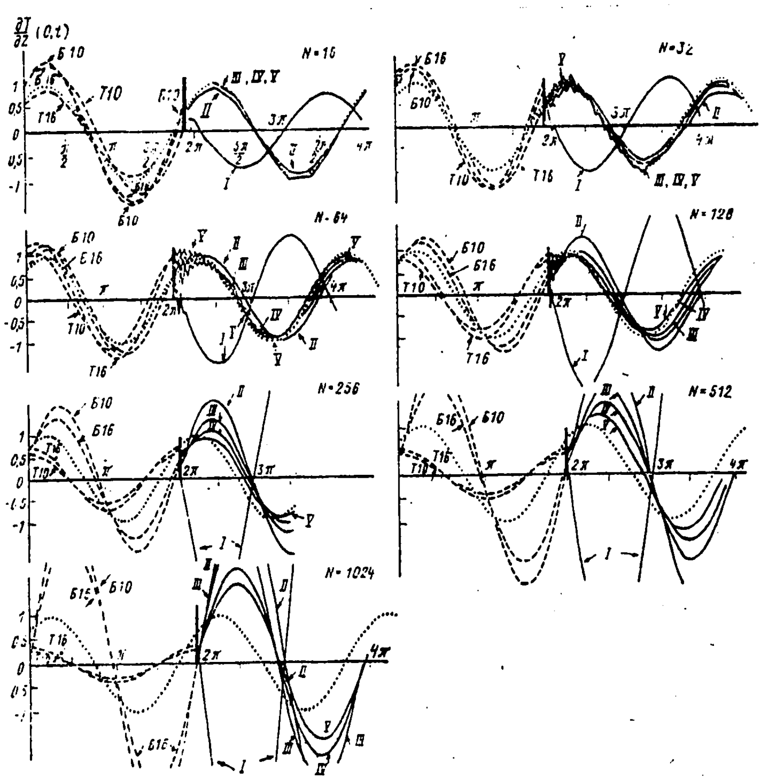

Fig. 2. Heat flow computed by different methods under condition that $T(0,t) - T_{-\infty} = \sin \Omega t$. $N = 2\pi / \Omega \tau$. The precise solution is represented by the dotted curve. Up to the vertical thick line the flow is determined using the explicit formula (5) with K = 9.15: with the coefficients from [1] the curves "B10" and with the coefficients from Table 1 the curves "T10" and "T16." After the thick line use is made of the recurrent formula (7) with M = 1,...5 the curves "1",... "V." The initial data necessary for (7) are taken from "T16."

The solution of equation (1') is found similarly:

$$\frac{\partial T}{\partial z}(0,\ t) = \sqrt{\frac{\operatorname{tg} \Omega \tau}{D \tau}} \sin\left(\Omega t + \frac{\pi}{4}\right).$$

In both cases the flow is shifted relative to temperature by $\pi/4$, regardless of the frequency Ω. The amplitudes of the flows are close with $\Omega\tau \ll \pi/4$. For example, with

$$\Omega \tau = \pi/4 \quad \sqrt{\operatorname{tg} \Omega \tau / \tau} \approx 1{,}13 \ \Omega^{1/2}.$$

With $\Omega\tau = \pi/2$ the equations for $f_{\pm 1}(z)$ degenerate and there is no periodic solution of equation (1').

Formula (5) gives an answer in the form

$$\frac{\partial T}{\partial z}(0, t) = (D\tau)^{-1/2} \sum_{j=0}^{K} \omega_j \sin[\Omega(t - j\tau)] = A \sin \Omega t + B \cos \Omega t,$$

and this answer is the better the closer the expressions

$$A = (D\tau)^{-1/2} \sum_{j=0}^{K} \times \omega_j \cos \Omega j\tau \quad \text{and} \quad B = (D\tau)^{-1/2} \sum_{j=0}^{K} \omega_j \sin \Omega j\tau$$

are to $\sqrt{\Omega/2D}$. With $K\tau \to 0$

$$A = (D\tau)^{-1/2} \sum_{j=0}^{K} \omega_j + O(\tau^{1/2})$$

and $B = O(\tau^{1/2})$ and, accordingly, do not converge to the constant $\sqrt{\Omega/2D}$. This is a result of the obvious fact that with very small τ the operator transforming the function $T_0(t) - T_{-\infty}$ into the function $\partial T/\partial z\,(0, t)$ "cannot give a clear picture of the behavior of the function in a short segment." This same shortcoming is true for the formula (7) which we derived. The conclusion is that it is necessary to use time intervals τ which are not too great ($\leqslant \pi/4\Omega$) and not too small. For this purpose one must first ascertain more precisely specifically what Ω frequencies must be described.

5. Figure 2 shows graphs of the function $\partial T/\partial z(0, t)$ obtained using different formulas. Curves constructed using formula (5) with K = 9 and 15 are shown to the left of the vertical thick line. The ω_j coefficients were computed either in accordance with [1] or as the coefficients of the Taylor series shown in Table 1. The solution is immediately assumed to be stationary since the transient process ends after 9 or 15 intervals. Curves constructed using formula (7) with different M are shown to the right of the thick line. As was predicted in section 3, there is a transient process which is the longer the greater the M value. However, after reaching a stationary solution such curves will be closer to standard. With respect to the necessary electronic computer capabilities the formulas "B10" and "T10" fall between IV and V, being considerably inferior to them in accuracy. Formulas of the type "T" evidently are preferable to formulas of type "B," although the corresponding coefficients also decrease more slowly. For practical use it is evidently necessary to select formula (7) with M = 3, which requires 7 memory units. In a case when the computer memory is not limiting, it is better to use M = 4 or 5. The results obtained with M = 1 must be considered unsatisfactory.

6. The heat balance equation at the soil surface, which is usually used in models of the atmosphere, can be regarded as a condition of the third kind for the equation for thermal conductivity in the soil (1). After linearization this condition has the form

$$\left(aT + b\,\frac{\partial T}{\partial z}\right)\Big|_{z=0} = g(t),$$

the second term is much smaller than the right-hand side and, accordingly, the first term on the left-hand side. It is necessary to find the function T(0, t).

Using the method in section 2, we obtain

$$(\tilde{T}(0)-\tilde{T}_{-\infty})=\frac{\tilde{g}}{a+b\sqrt{\frac{2}{D\tau}}f(\omega)}=\frac{1}{a}[1-\varepsilon f(\omega)+O(\varepsilon^2)]\tilde{g},$$

$$\varepsilon=\frac{b}{a}\sqrt{\frac{2}{D\tau}},$$

from which it follows, as in section 3, that

$$T^n(0)=T_{-\infty}+\frac{1}{a}\sum_{j=0}^{M}p'_j g(n-j\tau)-\sum_{j=1}^{M}q_j(T^{n-j}-T_{-\infty}),$$

where q_j were determined above, and $p'_j = q_j - \varepsilon p_j$.

In this problem, due to the smallness of ε the coefficients ω'_j will be substantially less in absolute value than ω'_0, and accordingly, a good quality of the approximation can be attained with lesser powers of M than in section 5. In particular, M = 1 can already prove to be suitable.

7. Formula (7) can also be used in processing the results of observations for the purpose of solving the inverse problem of finding the thermal conductivity coefficient D. If the series of observations $T^n(0)$ and $_n\partial T^n/\partial z$ (0) are known, it follows from (7) on the assumption that D = const that

$$D=\tau^{-1}\left[\frac{\sum_{j=0}^{M}p_j(T^n(0)-T_{-\infty})}{\sum_{j=0}^{M}q_j\frac{\partial T^n}{\partial z}(0)}\right]^2.$$

It is also possible to solve inverse problems with more complex hypotheses than D = const. It goes without saying that the scaling of $\{p_j\}$ and $\{q_j\}$ is required.

8. Since the thermal conductivity coefficient D enters into formula (7) in simple form, then in a case when D = D(t), but changes little during the time $M\tau$, it is possible, taking into account the continuous dependence of the solution on D, to substitute D(t) into formula (7), derived for constant D.

If D = D(z) and the dependence is such that the fundamental system of solutions of the equation

consists of known analytical functions, for example, of Bessel functions, it is possible to repeat the procedure of constructing the coefficients ω_j, p_j and q_j cited in section 3. It goes without saying that with different dependences D(z) different $f(\omega)$ and ω_j, p_j, q_j will be obtained. It will be possible to compute these coefficients in advance for different characteristic profiles D(z) and use that set of coefficients which corresponds to the most "similar" standard profile D(z).

In constructing the p_j and q_j coefficients there was arbitrariness in choosing the scheme (1'). With its replacement by another (a limitation is the analyticity of the analogue of the $f(\omega)$ function in a unit circle, which should ensure a decrease of the ω_j coefficients) other ω_j, and accordingly other p_j and q_j are obtained. The choice of the best formula is made difficult by the absence of a quality test;

different formulas of type (7) optimum in different frequency ranges. In addition, in these there will be different characteristic times of the transient processes Accordingly, it is either necessary, taking into account the need for further use of formula (7), to formulate a formal quality test, or it is necessary to limit the consideration to curves of the type shown in Fig. 2. In the author's opinion, it follows from Fig. 2 that with the same computer capabilities formula (7) gives an accuracy which is several times greater than a formula of type (5).

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CONCENTRATION OF MINERAL DUST IN THE ATMOSPHERE OVER THE USSR

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[Article by K. P. Makhon'ko, candidate of physical and mathematical sciences, and F. A. Rabotnova, Institute of Experimental Meteorology, manuscript submitted 11 Apr 80]

[Text]

Abstract: The article gives data for ascertaining the concentration of atmospheric dust by weight after ashing of the sample. The samples are taken by the aspiration method from the surface atmosphere. A map of the distribution of mean annual concentrations of mineral dust over the territory of the USSR in 1978 is presented, as are data on the change in the dust concentration for 1969-1979 in the example of the Moscow region. The annual variation of the dust concentration is considered for the southern part of Central Asia, Northern Kazakhstan and the Moscow region. Dependences of the dust concentration on wind velocity are given for dry, wet and snow-covered soils, averaged for 1978-1979 for the Moscow region.

As is well known, atmospheric dust plays an appreciable role in climate-forming processes. It is also an important weather-forming factor [1, 6, 7]. An extensive literature has been devoted to study of the dust content of the atmosphere under extremal conditions (dust storms, wind erosion of cultivated lands, etc.). However, contamination of the atmosphere with dust of natural origin outside of populated places under the ordinary weather conditions usually encountered has not yet been investigated adequately. We studied the behavior of the mineral component of atmospheric dust in the atmosphere, the principal source of which under natural conditions is the soil. This component with respect to its behavior in the atmosphere is more stable than the organic component, whose mass is not constant and is highly dependent on air humidity. The sampling of aerosols from the surface air layer was accomplished by the aspiration method.

Below we give the mean annual concentrations by weight (C) for mineral dust in the surface atmosphere in the Moscow region far from large cities during the period 1969-1979.

Years	1969	1970	1971	1972	1973	1974	1975	1976	1977	1978	1979
C $\mu g/m^3$	22	38	30	38	30	34	28	19	20	19	24

As can be seen from the cited data, the mean annual concentrations vary from year to year in a relatively small range; the extremal values differ from the mean by only 36% (the mean concentration of mineral dust in the Moscow region during the entire considered period is 28 $\mu g/m^3$). A similar variability of the mean annual concentrations is also observed for other regions of the Soviet Union and therefore for clarifying the pattern of spatial distribution of dust in the territory of the USSR in the first approximation it is possible to limit ourselves to data for some one year.

A detailed investigation of the distribution of surface concentrations of mineral dust away from populated places in the territory of our country was carried out in 1978. The observation points were selected in such a way that they were not in the possible aureoles of technogenic contaminations of the atmosphere near industrial centers and large cities. The collected data thus characterize contamination of the atmosphere by local dust of natural and anthropogenic origin arriving at the observation point as a result of distant transport. In winter, when the soil is covered with snow and dust formation does not occur in it, dust is observed in the atmosphere, arriving as a result of distant transport; on the other hand, in summer, with the intensive rising of dust from the soil, in the air there is a predominance of local dust of natural origin. Far from populated places when a snow cover is present the dust concentration in the atmosphere is usually small, whereas in summer, on the contrary, it is great. Therefore, with the averaging of data for the entire year the main contribution to the result of averaging is from dust of natural origin. An exception to this rule is localities with a low intensity of dust formation in the summer (tundra zone, swamp areas, etc.).

Figure 1 shows a map of the distribution of mean annual dust concentrations constructed on the basis of observational data. The high-mountain regions of the Caucasus, Urals, Central Asia and Siberia in this case are excluded from consideration because observations were not made there and the interpolation of data obtained at near-lying points to mountain regions is evidently inadmissible. In the construction of isolines in cases allowing an unambiguous interpretation of measurement data, as additional information use was made of data on dust sources associated with the type of climatic zones.

In an examination of the map we first of all note a clearly expressed tendency to a zonal distribution of the concentrations of mineral dust over the territory of the USSR. The zones of equal concentrations extend primarily from west to east in accordance with the orientation of the climatic zones and the predominating zonal transfer of air masses. The minimum concentrations of mineral dust in the surface atmosphere are observed for the most part in the northern part of the country and the maximum concentrations are located in the south. Thus, the surface concentrations of mineral dust averaged by latitude zones of the country have a clearly expressed latitudinal pattern.

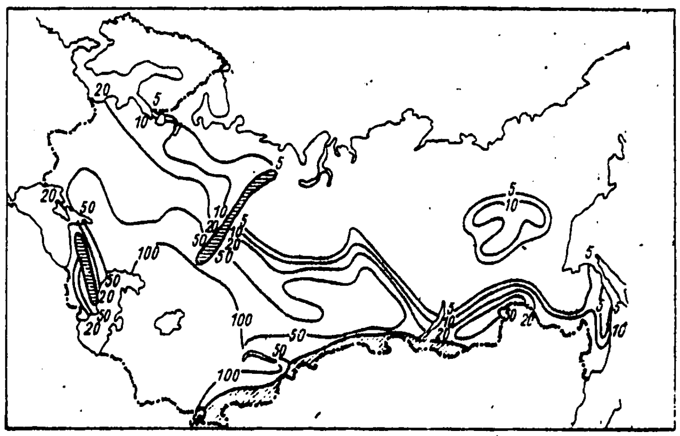

Fig. 1. Map of distribution of mean annual concentrations of mineral dust in the surface atmosphere over the territory of the USSR in 1978, $\mu g/m^3$.

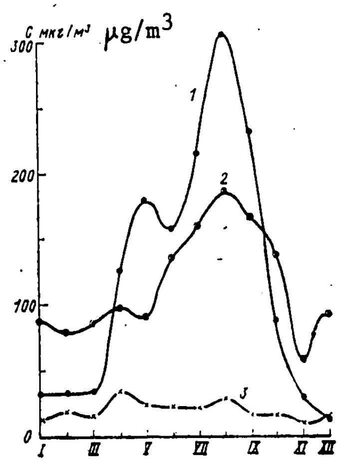

Fig. 2. Annual variation of surface concentration of mineral dust in atmosphere in 1978.

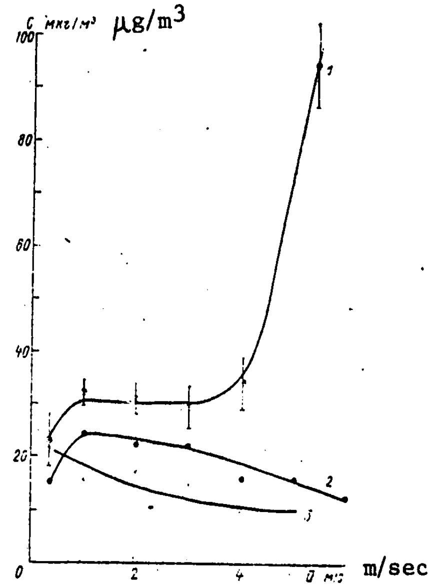

Fig. 3. Change in concentration of mineral dust in surface atmosphere as a function of wind velocity with different states of soil surface. 1) dry; 2) moist; 3) snow covered

This is attributable to the fact that the principal source of mineral dust in the atmosphere, as already noted above, is the soil surface [4, 5], but in the southern part of the country with its steppes and deserts the dust formation process on the soil transpires more intensively than in the north -- in the tundra or taiga. The maximum mean annual concentrations, falling in the interval 100-150 $\mu g/m^3$, are observed in the region of the focus of intensive natural dust formation in the desert zone and in the semideserts of Central Asia and Kazakhstan. On the average the air in Siberia is substantially purer than in the European part of the country, the reason being the weak sources of dust formation: extensive expanses of forests and swamps, and also the presence of the Urals Range, partially screening the northwestern part of Siberia from the westerly zonal transport of dust masses from the European USSR. In general, the pattern of distribution of the dust concentration over the territory of the USSR agrees entirely with the data in [1]. The considered map of the distribution of dust concentration was constructed using data averaged for the year. However, in the course of the year in different regions of the country the dust concentrations will vary differently, and therefore, naturally, the positioning of the isolines on the map will be transformed in dependence on the season of the year.

As an example we will consider a zone of maximum dust concentrations (100-150 $\mu g/m^3$ -- see Fig. 1) taking in a part of Kazakhstan and Central Asia. Figure 2 shows the annual variation of the concentration of mineral dust in the surface atmosphere of the northwestern part of this desert and semidesert zone (curve 1) and in the atmosphere of the oppositely situated southeastern part (curve 2). In the winter months in Northern Kazakhstan, as almost everywhere in the territory of the USSR, the soil is covered with snow and the formation of dust on the underlying surface almost does not occur; therefore, the dust concentration in the atmosphere in winter in this region is low (curve 1). On the other hand, in summer, when dust formation on the soil surface occurs intensively, the dust concentration increases by an order of magnitude. In the southern part of Central Asia there is virtually no snow and therefore the winter dust concentrations in the atmosphere are less than the summer concentrations by only a factor of 2 (curve 2). As a comparison, in this same figure we have shown the annual variation in the concentration of mineral dust in another region differing sharply in its physiographic conditions, in the zone of mixed forests in the southern part of the Moscow region (curve 3). The levels of dust concentration in this region are almost an order of magnitude lower and the annual variation of the concentration is expressed relatively weakly. The small maxima of dust concentration observed in late spring and in late summer are evidently attributable to the intensified formation of dust on the soil in the periods of harvesting and sowing of crops in agricultural fields. The annual curve of dust concentration in the northern regions of our country has a similar shape and is expressed very poorly.

Now, at least qualitatively, we will consider the factors leading to dust formation under natural conditions. The principal factors exerting an influence on dust formation are the state of the underlying surface, wind velocity and the intensity of vertical turbulent exchange in the atmosphere. Figure 3 shows the dependence of the mean daily concentration of mineral dust C on the mean daily wind velocity $\bar{u}$ in the case of a dry soil (curve 1), in the case of moist soil (curve 2) and in the case of a soil covered with snow (curve 3). The wind velocity was measured continuously at a height of 8 m above the soil surface; the samples were taken at a height of 1 m. Data on the dust concentration, presented in Fig. 3, were averaged for the Moscow region for 1978-1979, so that each point on the graph represents the result of a great number of measurements.

Figure 3 shows that with a dry soil surface with an increase in wind velocity $\bar{u}$ the dust concentration at first remains constant at a level of about 30 μg/m^3 up to values $\bar{u}$ = 3-4 m/sec, and then very rapidly begins to increase, attaining 100 μg/m^3 already with $\bar{u}$ = 5.5 m/sec. The existence of such a critical wind velocity at which intensified dust formation begins on the underlying surface has been observed repeatedly for sands and cultivated soils subject to wind erosion [3] and for blizzard transport of snow [2]. The constancy of the dust concentration with a wind velocity varying in the range 1-3 m/sec is attributable to the mutual compensation of two effects directed in opposite directions: with an increase in $\bar{u}$ there is an increase in the intensity of the raising of dust from the underlying surface and a simultaneous increase in dilution of the impurity by the relatively pure air of the above-lying layers of the atmosphere as a result of intensification of vertical turbulent exchange.

In the case of a moist soil surface the correlation between the dust concentration in the air and wind velocity can be different in dependence on the degree of soil moistening, that is, depending on the extent to which it retains a capacity for dust formation. If the soil is so moist that dust formation does not occur at all the dust concentration as a result of the above-mentioned effect of dilution with an increase in wind velocity should drop off. However, if dust formation occurs the concentration will increase. The effect of atmospheric precipitation, clearing the air, is superposed on all this. Figure 3 (curve 2) shows the pattern of change in the mean daily dust concentration with an increase in wind velocity in the case of a moist soil actually observed in 1978-1979. With an increase in wind velocity from 0 to 1 m/sec the dust concentration at first increased somewhat and then began to decrease slowly under the influence of vertical turbulent mixing in the atmosphere. For example, with $\bar{u}$ = 6 m/sec the concentration decreased by half in comparison with the maximum value.

The soil surface, covered with snow, virtually excludes the entry of soil dust into the atmosphere. The contamination of the surface atmosphere in this case is caused by the global background, owing its existence to the distant transport of dust masses at a planetary scale. At the same time, a definite role is played by the capture of dust particles by snowflakes during snowfalls, which thereby clear the atmosphere. It can be seen from Fig. 3 (curve 3) that in the case of a soil covered with snow the dust concentration is minimum. With an intensification of the wind it gradually decreases at approximately the same rate as in the case of a moist soil. This is attributable to an increase in the vertical scattering of dust from surface sources observed with an increase in wind velocity as a result of the accompanying increase in turbulence.

Thus, if dust formation on the underlying surface is difficult, the relationship between the dust concentration in the atmosphere and wind velocity is expressed poorly and is close to linear, whereas if the underlying surface is dry, such a relationship is well expressed and has a nonlinear character. Most of the dust is formed during wind gusts when its velocity attains maximum values.

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UDC 551.465(261)

ONE MECHANISM FOR THE FORMATION OF MACROSCALE WATER TEMPERATURE ANOMALIES IN THE OCEAN

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 66-71

[Article by Ye. S. Nesterov, candidate of physical and mathematical sciences, USSR Hydrometeorological Scientific Research Center, manuscript submitted 2 Jun 80]

[Text]

Abstract: The author investigated the conditions for the formation of macroscale water temperature anomalies in the North Atlantic in the warm season of the year. It is shown that the dynamic characteristics of the near-water layer of the atmosphere during the period of transition from a winter to a summer stratification of the active layer of the ocean play a substantial role in the formation of temperature anomalies during subsequent months. A correlation was found between atmospheric pressure anomalies in May and water surface temperature anomalies in July.

The study of the conditions for formation of macroscale water temperature anomalies in the ocean has long attracted the attention of researchers. Most of them have related the appearance of large anomalies to a change in the intensity of the principal ocean currents. A quite detailed review of these investigations is given in monograph [11].

At the same time, studies have recently appeared in which water temperature anomalies (especially during summer) are regarded as a result of the anomalous redistribution of heat in the upper layers of the ocean as a result of turbulent mixing. Emphasis is on the characteristics of the transitional period from winter stratification of the active layer of the ocean with isothermy to great depths to a summer stratification characterized by the presence of an upper homogeneous layer and a seasonal thermocline. For example, in [14] it was postulated that the earlier this transition occurs, the sooner (all other conditions being equal) will heating of the upper layer of the ocean then begin, which will favor the formation of a positive water temperature anomaly. In this study a good correspondence was obtained between the date of the transition and water temperature anomalies in subsequent months on the basis of the results of 19-year observations at "Rara" station in the Pacific Ocean.

Earlier it was demonstrated in [18] that most commonly water temperature anomalies in the North Atlantic are formed in spring or autumn and retain their sign for a period of several months. For example, the anomalies forming in May in 75% of the cases last to July and in only 10% of the cases change their sign to the opposite.

Thus, there is basis for assuming that the conditions for transition from a winter stratification of the active layer of the ocean to a summer stratification can exert an influence on the formation of water temperature anomalies in the warm part of the year. The purpose of this study is the checking of this hypothesis on the basis of data from field observations.

Now we will examine the thermodynamic processes in the upper layers of the ocean in the warm part of the year in greater detail. The penetration of solar radiation exerts a decisive influence on the thermal regime of the ocean during this period. However, with one and the same quantity of solar radiation incident on the ocean surface the thermal structure of its upper layers can be substantially different in dependence on the heat redistribution regime. Its distinguishing characteristic is the presence of a seasonal thermocline, which, having a considerable density stratification, is a barrier for the penetration of heat into the deeper layers. Accordingly, in the absence of intensive turbulent mixing the heat for the most part is redistributed in the layer above the seasonal thermocline. Accordingly, it can be postulated that the closer the seasonal thermocline is to the surface, the less deep will be the layer in which the heat is redistributed, which leads to a greater heating of this layer.

The formation of the seasonal thermocline occurs during the spring period when there is a transition from winter stratification of the active layer of the ocean to a summer stratification. The depth at which the seasonal thermocline is formed is dependent, evidently, primarily on the wind activity regime during this period. Strong winds cause intensive turbulent mixing which increases the thickness of the upper homogeneous layer and favors a decrease in its temperature, and vice versa, weak winds favor the formation of a thermocline at shallow depths.

If as a result of intensive wind activity the thickness of the homogeneous layer at the beginning of summer exceeds the usual thickness, the heat received from the sun will be redistributed to a depth greater than usual, which reduces the heating of the surface layer and will favor the formation of a negative water temperature anomaly at the surface. Such a situation was observed at the oceanic weather station "India" (59°N, 19°W) in the North Atlantic in 1964. In June strong winds deepened the homogeneous layer to 50 m. Such a depth persisted to the end of the summer [12]. This checked the heating of the upper layer of the ocean and favored the formation of a negative water temperature anomaly, which in July and August was -0.5°C. The opposite situation was observed in this region in 1965 when as a result of weak winds in May-June the thickness of the homogeneous layer in summer was 10-20 m, which favored an intensive heating of the upper layer of the ocean and the formation of a positive water temperature anomaly to +0.6°C.

Thus, an analysis of the experimental data and the investigations made earlier make it possible to postulate that if in the period of transition from winter stratification of the active layer of the ocean to a summer stratification (May-June in the temperate latitudes) there is intensive wind activity, the seasonal thermocline is formed at a greater than usual depth, which will favor the formation of a negative water temperature anomaly at the surface during summer. On the other hand, weak winds result in a shallow depth of deepening of the seasonal thermocline, which will favor the appearance of a positive anomaly with all other conditions being equal (reference is to the quantity of incident solar radiation and advection of heat by currents).

For checking this assumption we checked the formation of water temperature anomalies in the North Atlantic (40-60°N) during the summer months during the period from 1957 through 1979 [8-10]. The regions of the Gulf Stream and the Labrador Current were excluded from consideration because advective factors exert a strong influence on the formation of anomalies in these zones. During the investigated period the largest water temperature anomalies in July (when the isoanomalic line ±1°C occupied an ocean area with a characteristic horizontal scale of several thousands of kilometers) were formed six times: negative anomalies in 1968, 1970, 1972, 1974 and 1976 and a positive anomaly in 1958. For these cases we carried out a comparative analysis of the intensity of wind activity in May and the distribution of water temperature anomalies in July. Wind activity was evaluated indirectly through the atmospheric pressure anomaly [9]. It was assumed that a negative pressure anomaly characterizes active cyclogenesis, the result of which is frequent storms in the ocean, and vice versa, in the case of a positive anomaly cyclogenesis weakened and the wind velocity was low. An analysis of the experimental data indicated that for sufficiently great mean monthly pressure anomalies (more than 4-6 mb) this assumption is satisfactorily realized.

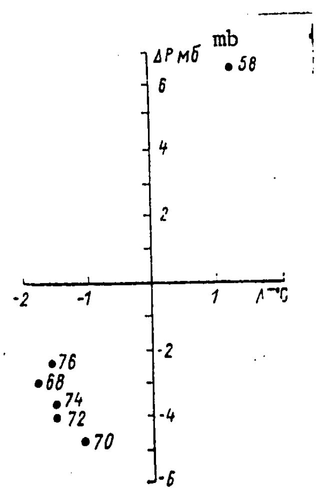

Fig. 1. Correlation between large water temperature anomalies in July and atmospheric pressure anomalies in May.

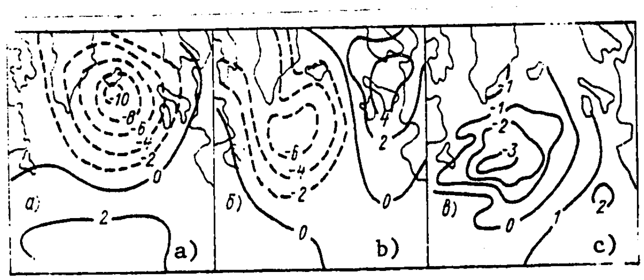

Fig. 2. Mean monthly atmospheric pressure anomalies (mb) in May (a), in June (b) and water temperature anomaly at surface (°C) in July 1976 (c).

We computed the mean water temperature and pressure anomalies for the water expanse occupied by the isoanomalic line ±1°C in July and constructed a correlation graph. (Fig. 1). The graph shows that there is a definite correlation between the pressure anomaly in May and the summer water temperature anomaly. The analysis indicated that as a rule large temperature anomalies are formed when the anomalous wind activity regime which was observed in May (mean monthly pressure anomaly exceeded 4-6 mb at the focus) also persists in June. For example, with the formation of large negative anomalies in the summer of 1968, 1972 and 1976 there were negative pressure anomalies not only in May, but also in June (Fig. 2). This agrees with the results in [15], where it was found from an analysis of observations that an increase in water temperature in the warm part of the year in the temperate latitudes occurs for the most part in those periods when the wind velocity is insignificant and intensive wind activity retards the increase in temperature.

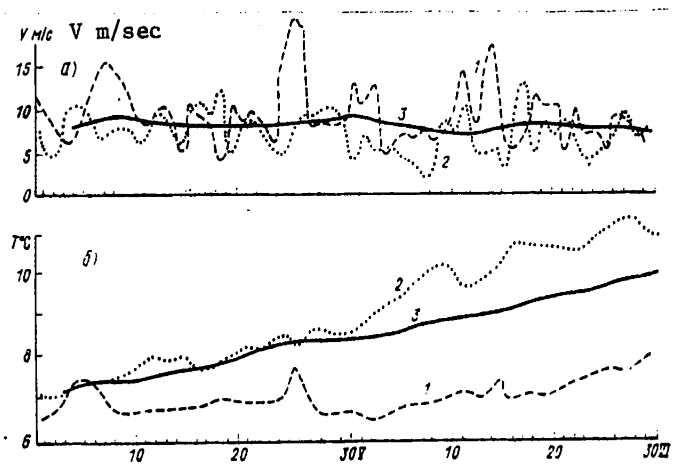

Fig. 3. Wind velocity (a) and surface water temperature (b) at Charlie station in May-June. 1) 1976; 2) 1977; 3) mean long-term values [10].

For a more detailed analysis of the conditions for the formation of temperature anomalies in the active layer of the ocean we will examine the hydrometeorological regime in the region of the oceanic weather station "Charlie" (52°45'N, 35°30'W) in the North Atlantic in May-June 1976-1977, since in these years in this region there was formation of water temperature anomalies of the opposite sign. As already mentioned, in the spring of 1976 in the North Atlantic there was intensive wind activity. This is confirmed by the data cited in Fig. 3, where it can be seen that in May-June 1976 in the considered region there were at least four prolonged periods with a wind velocity considerably greater than the mean long-term value, whereas in 1977 such phenomena were almost not observed.

It is interesting to note that in early May the thermal characteristics of the active layer of the ocean differed insignificantly in these years and the water temperature at the surface was close to the mean long-term value (Fig. 3). Later, however, different dynamic conditions in the near-water layer of the atmosphere favored the formation of a substantially different temperature distribution in the upper

100-m layer of the ocean indicated that whereas in early May the temperature curves in these years were characterized by insignificant differences, by the end of June great differences had arisen in the upper 50-m layer. In the layer from the ocean surface to a depth of 20 m the water temperature in 1977 was 3°C greater than in 1976, at the 30-m horizon -- by 2°C and at the 40-m horizon -- by 1°C, that is, the heat content of the upper layers of the ocean in these years was substantially different.

Thus, our analysis indicated that the hydrometeorological conditions for transition from a winter stratification of the active layer in the ocean to a summer stratification can exert an influence on the formation of the temperature regime in subsequent months. In this case the main role is played by the dynamic characteristics of the near-water atmospheric layer, in particular, the frequency and intensity of the passing cyclones with which ocean storms were associated. It is known that the storms exert a dual effect on the upper layers of the ocean. On the one hand, during a storm there is an increase in the turbulent fluxes of heat and moisture from the ocean into the atmosphere by many times [1, 4], and on the other hand, there is an intensive generation of turbulent kinetic energy, which is expended on the mixing of the surface water layers and the entrainment of colder water from the seasonal thermocline [14, 15]. All these processes result in a decrease in the temperature of the upper layer of the ocean and a deepening of the seasonal thermocline. Since in mid-summer storms in the temperate latitudes are observed considerably less frequently than in spring [3], their number in late spring-early summer, when there is a transition from winter stratification of the active layer of the ocean to a summer stratification, exerts a substantial influence on the further development of the thermal structure of the upper layers of the ocean.

In addition, as was demonstrated in [2, 7], the intensive transfer of heat from the ocean into the atmosphere in winter can exert a considerable influence on the formation of summer water temperature anomalies. However, if atmospheric circulation during the spring-summer period differs greatly from the mean long-term circulation, its influence on the formation of temperature anomalies in the warm part of the year becomes decisive [6]. This is confirmed by the results of statistical analyses [5, 13, 17], in which it is shown that the highest correlation between summer water temperature anomalies and the preceding atmospheric pressure anomalies in the northern parts of the Atlantic and Pacific Oceans is observed with a shift of 1-2 months. It is also necessary to note the results obtained in [16], where by means of a stochastic model of interaction between the atmosphere and ocean it was possible to reproduce the principal characteristics of the formation of large water temperature anomalies in the ocean (the temperature anomalies were regarded as an integral response of the ocean to random short-period atmospheric effects).

In conclusion it must be noted that the considered mechanism of formation of macroscale water temperature anomalies is unquestionably only one of the possible mechanisms of the formation of thermal anomalies in the ocean. There must be a more detailed investigation of other processes responsible for the formation of the thermal structure, and in particular, the anomalous receipt of heat at the ocean surface as a result of cloud cover anomalies and the redistribution of heat by anomalous drift circulation. In zones with an intensive horizontal advection an intensification or weakening of the main ocean currents can exert an influence on the formation of anomalies [11]. In the future it is necessary to evaluate the contribution of these mechanisms to the formation of large thermal anomalies in different regions of the world ocean.

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UPWELLING IN THE SOUTHERN PART OF LAKE ONEGA

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 72-75

[Article by P. M. Boyarinov, Karelian Affiliate, USSR Academy of Sciences, manuscript submitted 17 Jun 80]

[Text]

Abstract: On the basis of the results of a thermal survey in the southern part of Lake Onega the author gives the spatial structure of an upwelling. It is shown that an along-shore wind results in vertical movements in the lake. As a result, the thermocline reaches the surface, forming a thermal front with an average temperature gradient up to 5°C/km at a distance from the shore of 5-7 km. The upwelling phenomenon affects the entire thickness of the water layer, propagating to a depth of 28 m.

By the term "upwelling" is mean the process of vertical movement of water in a sea or lake, as a result of which deep waters, and accordingly, the colder and denser waters, rise to the surface.

An upwelling can be caused by many factors. Among the most important we can mention a wind-driven surge of surface waters into the open part of the sea, divergence of horizontal currents and the rising of waters in the central parts of cyclonic circulations.

What is the essence of an upwelling in the sea or deep lake? In accordance with [3] we will examine the regime of a coastal zone on the assumption that the shoreline is linear and extends infinitely in both directions. The depth of the lake will be considered great and the shore will be considered precipitous. Assume that the wind is directed parallel to the shore in such a way that the shore is situated to the left of it. A drift current arises under the influence of the wind and this transports the water mass in a direction perpendicular to the action of the wind, that is, into the open lake. Along the shore there is a driving away of the water, leading to a lowering of the lake level along the shore, in other words, to the appearance of a pressure gradient. In turn it causes a gradient current which in the main thickness of the water is directed perpendicular to the gradient, that is, along the shore and thereafter to the left (looking from the shore). With the appearance of the gradient current a flow component near the bottom and normal to the shore also appears and this will be directed in the direction opposite the normal component of the drift flow (in this case toward the shore). With some angle of slope of the water body level, dependent on the wind velocity, an equilibrium will set in between these flows. The circulation will be steady. Since the flows are separated by a layer occupied by a gradient current moving along the shore, due to the condition of continuity of movement their combined flow should exist, directed upward. This vertical flow, proportional to wind velocity, is an upwelling.

Due to the diversity of factors causing this, this phenomenon occurs widely in the coastal regions of the seas and oceans and in lakes. Despite the fact that the zone of upwelling of deep waters is a limited region, the rising waters with their properties, different from the properties of the surface waters, lead to a marked restructuring of the thermodynamic, chemical and biological processes in water bodies. Nevertheless, researchers only relatively recently (in the 1950's in the ocean and in the mid-1970's in large lakes) began to devote serious attention to the study of vertical movements. Studies have appeared [7, 8, 11] that have laid a basis of a theory of vertical circulation in the sea.

Recently great successes have been attained in the investigation of upwelling in the coastal regions of the seas and oceans. And nevertheless this phenomenon in lakes has still been studied extremely inadequately. Among the known studies it is necessary to mention [4-6, 9, 10]. They are devoted to the problems involved in study of upwelling and downwelling in the Great Lakes. It is pointed out in these studies that the upwelling of cold deep or the subsidence of warm surface waters lasts an average of 4-5 days in accordance with the cyclonic activity over the Great Lakes. Vertical movements occur when there are winds parallel to the shore line. Observations have shown that with the onset of upwelling the pycnocline begins to move and intersects the surface of the lake, in a number of cases forming a front with exceedingly large temperature gradients, in Lake Ontario attaining 4°C/km [4]. The thermocline, emerging at the surface, can withdraw from the shore a distance as great as 6-10 km. Less frequently the thermocline emerges at the surface with formation of a front during downwelling. Such a case is cited for Lake Superior [9]. The temperature gradient at the front attained 3°C/km and separated coastal waters with a temperature of 15°C from the cold lake waters, where homothermy with a temperature of 5°C was observed.

Among the Soviet studies it is necessary to mention [1], reporting a decrease in water temperature at the surface from 16 to 4°C in Lake Baykal in the coastal region when there is an offshore wind, and [2], examining surge phenomena in Lake Dal'nyy. Due to the small size and depth of Lake Dal'nyy here the vertical movements have had a different mechanism as a result of predominance of frictional forces over Coriolis force and occurred when there were winds directed from the shore.

Information on the emergence of deep waters at the surface during surge phenomena is also contained in a number of other studies, but as a result of lack of specific observational data none of them make it possible to examine the mechanisms responsible for the upwelling of water, to evaluate the intensity and limits of propagation of the phenomenon.

Taking into account the poor degree of study of mesoscale vertical movements in lakes, as well as the almost complete lack of studies describing the spatial structure of an upwelling, it seems important to describe the horizontal and vertical distribution of temperature, serving as an indirect indicator of vertical movements in the water body, for the case of an upwelling in the southern part of Lake Onega.

The investigations were made from the scientific research ship "Poseydon" (type SChS-150) in late July 1979 during stable density stratification with a clearly expressed thermocline. The surface waters had a temperature of 16°C; in the bottom layers it was below 6°C.

The distribution of temperature in the upper layer was obtained when towing a temperature sensor. As the latter use was made of a copper resistance sensor with a time constant of about 30 sec. The signals from this sensor were fed to an EMP-09 bridge which registered temperature with a discreteness of 5 sec. With a speed of movement of 8 knots the spatial resolution in registering temperature was 120 m. The response of the measuring circuit was regulated in such a way that the full scale of the bridge was 25°C (from 0 to 25°C). This ensured an accuracy in reading temperature not poorer than 0.15°C. The sensor of the resistance thermometer was towed from the ship on a rigidly attached rod at a depth of 1.0 m.

Despite the small response of the temperature sensor and the errors inevitably introduced by towing behind the ship, such temperature registry in the surface layer is entirely adequate for investigating mesoscale processes, the upwelling phenomenon entering into this class.

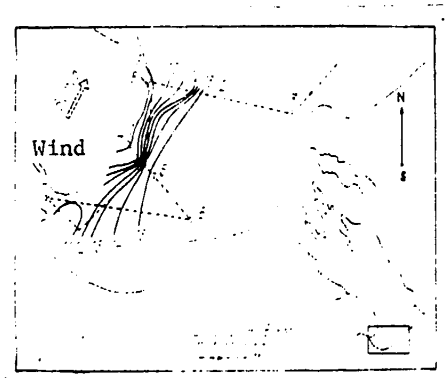

Fig. 1. Distribution of water temperature at 1-m horizon during the upwelling of 22 July 1979.

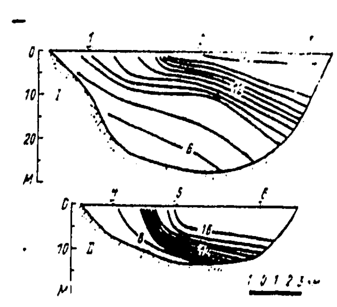

Fig. 2. Vertical distribution of water temperature in sections I and II intersecting upwelling front.

Deep-water reversing thermometers were used for evaluating the distribution of temperature with depth.

In the course of the three days preceding the survey and at the time it was carried out there was a predominance of a SSW wind with a velocity 7-8 m/sec. The wind direction coincided with the general direction of the shoreline in this region.

Such a wind caused a transport of surface waters from the western shore into the open part of the lake and the upwelling of cold deep waters with a temperature of less than 8°C at the surface along the shore.

It follows from an analysis of the horizontal distribution of temperature (Fig. 1) that the isotherm 8°C, showing the lower boundary of the thermocline, withdrew from the shore for a distance of 4 km, whereas the isotherm 15°C, characterizing the temperature of the surface layer prior to the action of the wind, withdrew from it 7-9 km. In the southwest, shallower part of the region with depths not exceeding 10 m there were waters with a temperature 11-12°C, rising from the bottom. In the central and eastern parts of the region there was concentration of waters with a temperature 15-16°C with a mean horizontal gradient 0.07-0.08°C/km. In the frontal zone it was approximately 20 times greater (1.5-2.0°C/km).

Figure 2 shows the temperature distribution in the vertical plane for two sections intersecting the upwelling front.

Data on the vertical distribution of temperature, obtained for a small number of stations, were supplemented by materials on the direct registry of its spatial change accomplished during towing. This made it possible to refine the pattern of temperature distribution in the surface layer and more precisely define the frontal zone.

It was established in the course of the investigations that simultaneously with the upwelling observed in the western part of the water area, in its eastern region during presence of a wind directed parallel to the shore, there was a subsidence of the warm surface waters in depth, that is, the downwelling phenomenon occurs. The thermocline, characterized by the isotherms 12-13°C, subsides to a depth of 12-14 m (instead of 4-5 m in an undisturbed state), whereas the waters with a temperature of 15°C are at a depth up to 10 m.

The upwelling phenomenon mentioned above was propagated in the considered region to the entire thickness of the waters to a maximum depth in this part of the water area of 28 m (Fig. 2). For example, the isotherm 6°C, characterizing bottom waters in section I, assumed a slope as a result of upwelling of waters in the western part and subsidence in the eastern part, equal to 12 m at a distance of 9 km, that is, on the average 1.3 m/km.

The emergence of the thermocline at the lake surface is accompanied by the formation of an exceedingly sharp thermal, and therefore density front. The zone with maximum temperature contrasts, attaining 5°C at a distance of 5-6 km from the shore in section II, has a width of about 1 km. The thermocline creating it has a slope equal to 0.007 or 7 m/km. It should be noted that the temperature gradient is not identical in the entire section of the front and in one of the sectors is 1°C per 80 m. Such temperature drops undoubtedly cause a strong gradient current directed in the plane of the figure. Unfortunately, the absence of data on currents in this region does not make it possible to evaluate its intensity.

It can be postulated that a stable NNE wind, parallel to the shoreline in this region, will cause inverse phenomena, that is, will be accompanied by an upwelling on the eastern and a downwelling on the western side of the water body.

Thus, it follows from the above that a stable SSW wind, parallel to the shoreline in the southern part of Lake Onega, causes an upwelling phenomenon in the western part of the region and a downwelling in the eastern part of the region. These phenomena occupy the entire water layer, propagating to a depth of 28 m. The emergence of the thermocline at the surface causes the formation of a thermal front with a temperature gradient up to 5°C/km and a distance from the shore of 5-7 km.

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LABORATORY INVESTIGATIONS OF THE INFLUENCE OF STRUCTURES FOR PROTECTION OF LENINGRAD AGAINST INUNDATIONS ON WATER LEVEL RISES IN THE GULF OF FINLAND

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 76-81

[Article by V. G. Noskov, candidate of geographical sciences, State Hydrological Institute, manuscript submitted 16 May 80]

[Text]

Abstract: The article gives the results of investigations (with a hydraulic model) of the influence of structures which are being constructed for the protection of Leningrad against sea inundations on storm-induced level rises in the Gulf of Finland to the west of the structures. The experiments revealed that with the most unfavorable meteorological and hydrological conditions an increase in level rises directly before the structures along the line Gorskaya-Kronshtadt-Lomonosov will be not more than 13% (32 cm) in the Northern Gates and 8% (20 cm) in the Southern Gates. With increasing distance from the structures their influence lessens and 100 km to the west of these structures no influence is detected.

The project for the protection of Leningrad against sea inundations provides for the construction of protective structures along the line Gorskaya-Kronshtadt-Lomonosov [1]. In this connection the need arose for determining their influence on the magnitude of storm-induced water level rises in the Gulf of Finland, clarifying how far to the west this influence will be propagated. In the event of a substantial increase in level rises it must be taken into account in calculating how high the protective structures must be and an evaluation must be made of the degree of danger of storm-induced surges when protection has been provided for the populated places situated on the shores of the Gulf of Finland to the west of the structures (Fig. 1).

This problem, together with the use of computation methods, was solved using a large-scale hydraulic model of the Gulf of Finland (Fig. 2), the most important information on which was given in [4]. Long-wave and seiche oscillations, observed both in nature and in the model under the influence primarily of gravitational and inertial forces, were modeled with adherence to Froude similarity.

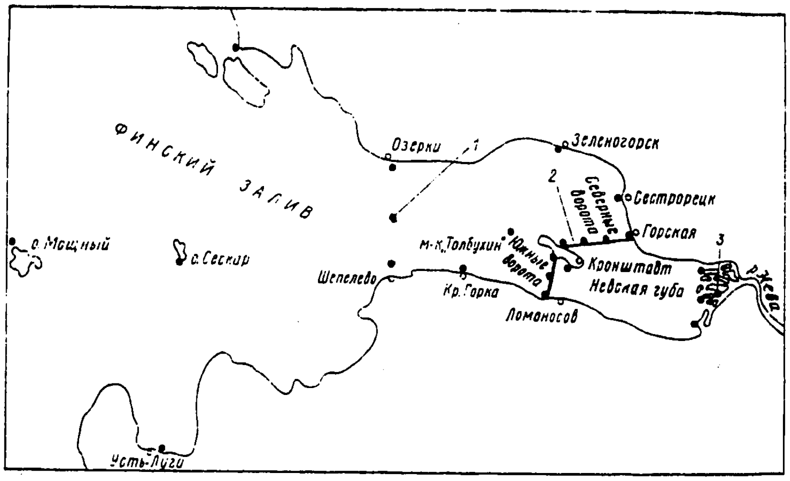

Fig. 1. Plan of fragment of model of Gulf of Finland. 1) sites of water level registry, 2) lines of protective structures, 3) gaging station at Mining Institute.

A subsequent comparison of some of the most important model and field parameters characterizing the processes of propagation, transformation and reflection of long waves and subsequent inertial level oscillations indicated their entirely satisfactory coincidence. For example, one of the most important prognostic characteristics is the ratio of the maximum water level rise at the Mining Institute to the rises at other points in the gulf -- and both in nature and in the model it was identically equal in Kronshtadt (1.3) and at Ust'-Luga (1.9) and only at Tallin was there some discrepancy: 2.5 in nature, according to N. I. Bel'skiy [2], and 2.8 according to the model. The time required for the propagation of long waves from the entrance to the Gulf of Finland to the Neva delta in nature and in the model was also identical — 8.0-8.1 hours. The period of uninodal seiche oscillations in nature and in the model is 26-28 hours. In the model, as under field conditions, there were repeated level rises at the head of the gulf, following approximately 8 hours after the peak of the first rise. Finally, at all points in the model the variation in water level during the flood of 15 October 1955 was reproduced quite well, which all researchers relate primarily to inundations of a long-wave origin.

In the modeling of wind surges, since the similarity criteria for reproduction of the wind are unreliable, recourse was to an artificial procedure -- the comparison method: for one and the same conditions for reproduction of the wind we compared the level rises in the eastern part of the Gulf of Finland in the absence and presence of protective structures.

The experiments with the reproduction of long waves in the model indicated that under the planned conditions the height of the water level rise to the west of the structures can be both greater and smaller than the rises under natural conditions, depending on the magnitude of the wave itself. An increase in the level rises in the Gulf of Finland to the west of the structures is noted in those cases when the wave period is less than 16 hours. [The characteristics relating to water level fluctuations in the model are given scaled to natural conditions.]

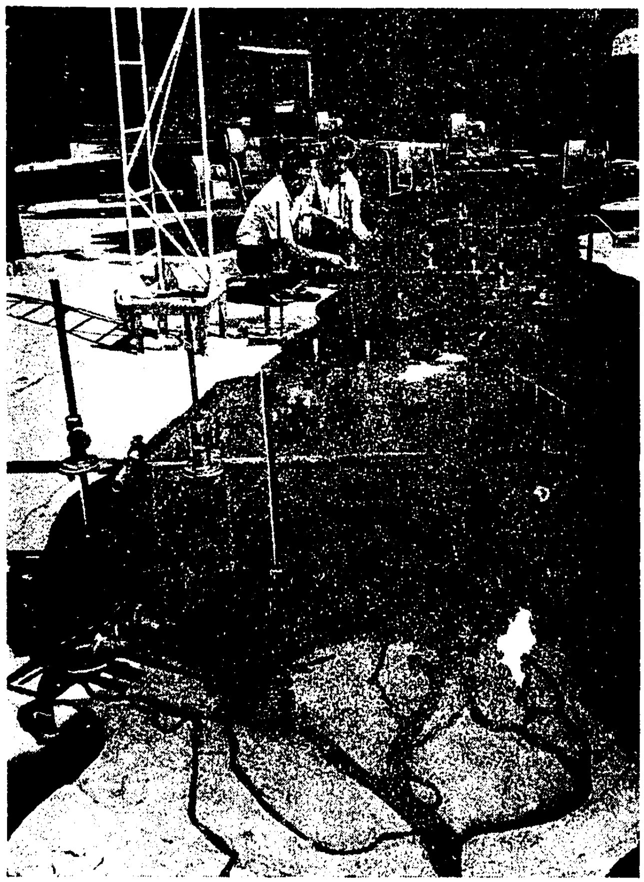

Fig. 2. Model of Gulf of Finland (view from the direction of the Neva delta). Reproduction of wind surge under planned conditions.

The level increment (Δh) before the structures evidently occurs at the expense of that volume of water which under natural conditions is expended in filling the Neva Inlet and the Neva delta above the initial level. Under the planned conditions the water volume not entering into the Neva Inlet is expended on additional filling of the part of the Gulf of Finland adjacent to the structures to the west.

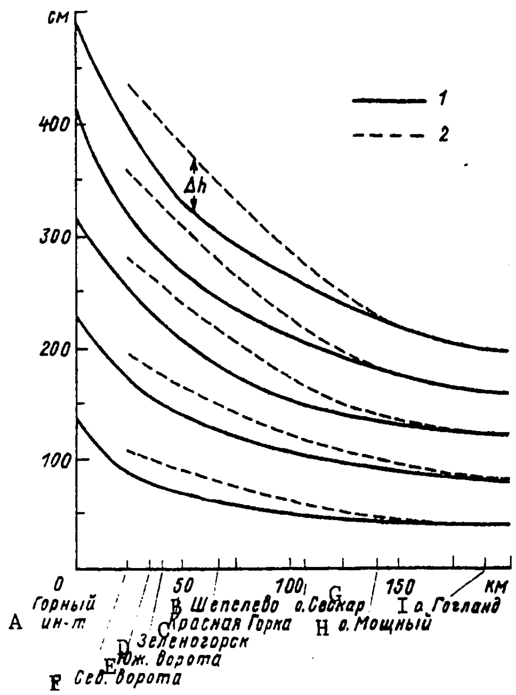

Fig. 3. Curves representing increase in water level rises in eastern part of Gulf of Finland during propagation of long waves of different height. 1) natural conditions; 2) planned conditions.

KEY:

A) Mining Institute
B) Shepelevo
C) Krasnaya Gorka
D) Zelenogorsk
E) Southern Gates
F) Northern Gates
G) Seskar Island
H) Moshchnyy Island
I) Gogland Island

Figure 3 shows an example of change in level rises along the length of the eastern part of the Gulf of Finland caused by the propagation of long waves of different height with their identical period (t) under natural and planned conditions.

An examination of Fig. 3 makes it possible to express the following judgments.

1. The effect of protective structures on the level rise is discovered approximately to Moshchnyy Island (about 108 km from the structures).

2. The greatest rise is observed directly before the structures, but the greatest level increment does not occur at the structures themselves, but at a distance of 15-20 km to the west of them. This is attributable to the change in the conditions for the reflection of long waves: under the planned conditions the reflection front is moved 30 km to the west, into a region of relatively greater depths and

the gradual decrease in the depths of Neva Inlet and the low shores are replaced by a virtually vertical wall.

3. The magnitude of the increase in water level under the influence of structures with an identical period of the waves is the greater the greater the height of the wave.

Table 1

Increments of Water Level Rises (Δ h cm) in Gulf of Finland to West of Protective Structures for Incidence of Long Waves With Different Periods With Same Heights

Region	Wave period t, hours				
	6	8	12	16	20
Northern Gates	12	6	14	14	-20
Southern Gates	12	6	10	14	-10
Zelenogorsk	40	28	26	28	6
"Tolbukhin" beacon	34	18	34	34	10
Shepelevo	32	18	22	26	14
Seskar Island	8	16	14	14	0

Table 2

Increment of Water Level Rises (Δh cm) in Gulf of Finland to West of the Protective Structures With Incidence of Long Waves of Different Heights at t = 16 hours

Region	Water level rise in Neva delta H, cm				
	100	200	300	400	500
Northern Gates	17	27	36	42	46
Southern Gates	7	15	24	33	41
Zelenogorsk	16	25	34	42	51
"Tolbukhin" beacon	22	31	40	50	59
Shepelevo	14	20	26	32	38
Seskar Island	10	12	14	16	21
Moshchnyy Island	0	0	0	0	0

Other experiments, confirming the judgments made, indicated, in addition, the special nature of measurement of the Δh value with a change in wave period. The Δh value with a change in t from 6 to 16 hours remained essentially unchanged (Table 1). With a transition to waves with a period of 20 hours the Δh value at all points in the gulf from the structures to Seskar Island decreased sharply, assuming negative values directly at the structures. In these experiments the height of all the waves was identical. With their incidence in the head of the gulf the level rise in the Neva delta for natural conditions was 230 cm.

An explanation of this phenomenon must probably be sought in the circumstance that with a period of more than 16 hours the length of the wave begins to exceed substantially the length of the model and possibly therefore qualitatively changes the picture and the results of superposing of the preceding, reflected part of the wave on the unreflected part of the wave [3].

For use in planning computations we recommend the means of the maximum registered Δh values (Table 2). Their values were determined for different wave heights, conditionally characterized by the level rise in the Neva delta under natural conditions and for wave periods not exceeding 16 hours.

It follows from an analysis of the reasons for the storm level rises in Neva Inlet made by Bel'skiy [2] that level rises occurring from the incidence of free long waves, more than 250 cm on the line of structures and more than 230 cm in the neighborhood of "Tolbukhin" beacon, have not been observed during the entire 250-year period of observations. They also were not noted after publication of the study by Bel'skiy (1954). This gives basis for assuming that the probability of their appearance in the future is insignificant. Accordingly, any Δh values in the Northern Gates of more than 32 cm (13%), in the Southern Gates of more than 20 cm (8%) and at the position of "Tolbukhin" beacon of more than 36 cm (16%) are improbable. The somewhat greater (by 5-12 cm) increment in level rises in the Northern Gates in comparison with the Southern Gates is possibly attributable to the differences in topography of the Northern and Southern Gates when protective structures are present.

In order to detect the influence of protective structures on the level rise caused by seiche oscillations a single-node seiche with a period of 26-28 hours, most frequently observed under natural conditions, was reproduced. This study revealed that protective structures exert virtually no influence either on the magnitude of the level rises or on the period of the seiche, since they cut off only 30 km (about 2% of the total extent of the most narrow and shallow-water part of the basin).

In order to determine the magnitude of the level rise before the structures when there are wind surges a "westerly" wind of the same force was created, as a result of whose action a level rise up to 145 cm developed in the Neva delta and up to 225 cm developed when six and nine fans respectively were activated. Even with such exceedingly high wind surges no substantial increase in level rises was discovered in the model before the structures in comparison with natural conditions.

Since the wind surge is dependent primarily on wind force and on the depth of the water body and the protective structures do not change either, at the time of wind surges there is no basis for expecting changes in the level rises before the structures. True, in the Gorskaya and Sestroretsk region in the experiments there was an increase in level rises under the planned conditions in comparison with the natural conditions by 10-20 cm, but this, it can be surmised, is attributable to the same factors which cause an increase in the rises here with the incidence of long waves.

The results of experiments for determining the Δh value during the level rises formed by the mutual superposing of long-wave, wind and seiche rises are set forth in [4]. They are essentially as follows: in the Neva delta and in Neva Inlet, including the Northern and Southern Gates, under natural conditions and directly before the structures under the planned conditions the resultant rise with superposed phenomena in most cases is less than the sum of the rise components, that is, the superposition principle at the head of the gulf is not observed. Under the planned conditions the approximate "level loss" with mutual superposing of different kinds of rises is 10-20%.

Thus, the experiments with a model of the Gulf of Finland revealed the following:

1) The protective structures increase the storm level rises before the structures only with the incidence of free long waves on them with a period of 16 and less hours and decrease them in comparison with rises under natural conditions with the incidence of waves with a period greater than 16 hours.

2) In the presence of wind surges and seiche oscillations the influence of protective structures on water level rises is not reliably detected.

3) When there are rises of mixed origin, when there is a mutual superposing of the three principal types of water level fluctuations -- long-wave, seiche and those caused by the westerly wind, it is found that the resultant of the rises coinciding in phase is always less than the sum of the individually taken rise components.

Dangerous rises occur with the superposing of especially significant long-wave, seiche and wind-induced changes in the level surface of the sea and therefore it is recommended that it be acknowledged that under the most unfavorable, but physically possible conditions the magnitude of the additional rise under the influence of structures cannot be greater than that which is produced by some long waves, that is, 8% in the Southern Gates and 13% in the Northern Gates or 20 and 32 cm in absolute expression. Under real conditions, however, if the "level loss" phenomenon is taken into account, this being discovered with the mutual superposing of rises of different origin, the indicated values will be essentially less.

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UDC 556.536

DEPENDENCE OF FREE SURFACE SLOPES ON THE MORPHOMETRIC CHARACTERISTICS OF A CHANNEL AND FLOODPLAIN

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 82-88

[Article by N. B. Baryshnikov, doctor of geographical sciences, and Ye. S. Subbotina, Leningrad Hydrometeorological Institute, manuscript submitted 9 Jun 80]

[Text]

Abstract: The problems involved in computing the free surface slopes for maximum water levels are examined taking into account the morphometric characteristics of the floodplain and channel. The authors propose a new interpretation of the reasons for the formation of loops on the discharge and mean flow velocity curves during the passage of high waters along the floodplains of lowland rivers, based on allowance for the volume of filling and emptying of a floodplain and channel and the effect of interaction of the flows in them. A computational graphic dependence is obtained showing the dependence of the slopes of the free surface of a channel flow with different levels of floodplain inundation on the angle of intersection of the dynamic axes of channel and floodplain flows.

In calculations of maximum water discharges one of the most important but least studied characteristics is the slope of the free water surface. There is no method for computing slopes and in practical work its value is assumed equal to the averaged bottom slope of the river or the slope of the free surface of the watercourse during the low-water period. The imperfection of such a method has been noted by many researchers (M. A. Velikanov [4], G. V. Zheleznyakov [7], D. Ye. Skorodumov [9] and others). In particular, M. A. Velikanov, as early as 1948, indicated the dependence of slope of the free surface of a river during the high-water period on the morphometric characteristics of structure of the channel and floodplain [4]. This point has been developed more fully by one of the authors of this article, who made an analysis of the nature of change of the free surface slopes of a channel flow with its interaction with a floodplain flow in dependence on the morphological structure of a channel and floodplain below the reach used in the computations and establishing that narrowing of the floodplain (valley) causes a decrease in the free surface slopes with an increase in its levels, whereas its broadening, on the other hand, causes their increase [1, 3].

In actuality, these studies have demonstrated the invalidity of using the averaged value of the slope of the free surface of the flow, equal to the bottom slope or the slope of the free surface of the flow at the low-water period in computations of maximum water discharges.

Accordingly, we made an analysis of the influence of the morphological characteristics of a channel and floodplain on the slopes of the free surface of a channel flow during its interaction with the floodplain flow. It should be noted that the lack of measurement data, and, indeed, a method for measuring the free surface slopes of a floodplain flow, especially in the early stages of floodplain inundation, makes it impossible to carry out a joint analysis of information on the change in the slopes of channel and floodplain flows. However, an analysis of special field investigations carried out by specialists of the State Hydrological Institute [10] indicates substantial differences in their change with an increase in water levels, the presence of considerable transverse level drops on the floodplain and on the boundary of the channel and floodplain flows. The transverse slopes in young channels are particularly great.

In this article particular attention will be given to two problems: the influence of the morphometric characteristics of the floodplain and channel on the nature of change in the free surface slopes of the channel flow and the reasons for the formation of loops on the curves for discharges, velocities and free surface slopes during the interaction of channel and floodplain flows.

On the basis of observational data collected by the Hydrometeorological Service on 66 rivers in the Soviet Union we carried out an analysis of information on high-water measurements in channels with floodplains. As a result of the analysis, depending on the reliability of the initial information and the amplitude of level variations revealed by measurements of water discharges, we selected information on 33 rivers. On 13 of these there was a second type of interaction between the channel and floodplain flows, on 19 -- the third type, and on one -- the first type. At all posts the axes of the channel and floodplain flows intersected at angles less than or equal to 50°. Accordingly, all the observational data for rivers with type-IV interaction of flows, in accordance with the nature of change in the floodplain below the computation point, were assigned either to the second or to the third types.

The basis of the classification proposed by N. B. Baryshnikov in [2] was the change in the widths of the floodplains, characterized by the relative orientation of the dynamic axes of the channel and floodplain flows. The first type corresponds to parallelism of the flow axes; the second is characterized by their divergence in the lower-lying reach (with broadening of the floodplain); the third corresponds to a convergence of the axes (with narrowing of the floodplain). The fourth and fifth types correspond to flows whose dynamic axes intersect at different angles (α). With small α angles and broadening of the floodplain below the computation reach the process of interaction of the flows in the fourth type is closer to the second, but with narrowing of the floodplain is closer to the third. The fifth type differs from the fourth in having in having a stepped structure of the floodplains.

It should be noted that observations of free surface slopes of the channel flow at most posts either are not made or have a low accuracy. As an example we can cite the Northwestern Administration of the Hydrometeorological Service where measurements of slopes at floodplain posts are made only at Tolmachevo station, situated on the Luga River. The situation is still worse at the Krasnoyarsk, East Siberian and a number of other administrations where high-water measurements in floodplain sectors are not made at all.

As the principal computation characteristic N. B. Baryshnikov proposed the values of the angles of intersection (divergence or convergence) of the dynamic axes of the flows (α) [1, 2]. The α angle is computed most precisely from maps of currents constructed on the basis of field data as the angle between the averaged vectors of the mean velocities of the channel and floodplain flows. Approximately, with an accuracy to 3-5°, this angle is determined as the angle between the geometrical axes of the channel and floodplain from the (contoured) map of the area of a post situated below the computation sector.

The value of the α angle sometimes varies very greatly with a change in water levels and its value as a rule is determined by the nature of the change in the widths of the floodplain in the lower-lying reach. For the second type of interaction of flows the value of the α angle was taken with a negative sign, whereas for the third it was taken with a positive sign. This is attributable to the fact that the effect of interaction of flows in the second type is opposite of that in the third. In actuality, for the third type with an increase in the α angle and constant depth values the velocities and the free surface slopes of the channel flow decrease and the resistances increase. In the second type the reverse picture is observed, that is, with an increase in the α angle the velocities and slopes increase, whereas the resistances decrease [3].

In developing the method the principal computation parameter used was the free surface slope of the channel flow at the level of flooding of the channel bank brow ($I_{chan\ br}$). The level of flooding of the channel bank brow is determined rather simply from the cross-sectional profiles at a hydrological post for rivers with a channel process -- free and uncompleted meandering and a multibranching floodplain process. The determination of this level for other types of channel process, described in a number of special studies [1, 7, 9], is somewhat more complex.

For each watercourse we computed the values $I/I_{chan\ br}$ and constructed curves of the dependence $I/I_{chan\ br} = f(H')$ and $I/I_{chan\ br} = f(h_{chan}/h_{chan\ br})$, which supplemented the data published by N. B. Baryshnikov in [1], where H' is the water level over the channel bank brow. As indicated by an analysis of these curves, their position is determined by the morphological structure of the reach situated below the point for which the computations were made.

Using the collected data for constant values of relative depths ($h_{chan}/h_{chan\ br}$), we constructed graphic dependences $I/I_{chan\ br} = f(\alpha)$ (Fig. 1). The figure shows that it is possible to trace a clear pattern of change in the relative values of the free surface slopes with an increase in the α angle. Thus, with an increase in α from -45 to 0° (second type of interaction of flows) there is an intensive decrease in slopes ($I/I_{chan\ br}$). The $I/I_{chan\ br} = f(\alpha)$ curves for different values of relative depths are almost parallel to one another and only with values of the α angle close to 0° does the intensity of changes in the slopes decrease sharply. In actuality, with an increase in the α angles in the cited range the values of the slopes ($I/I_{chan\ br}$) change from 3.5 to 1.0 respectively.

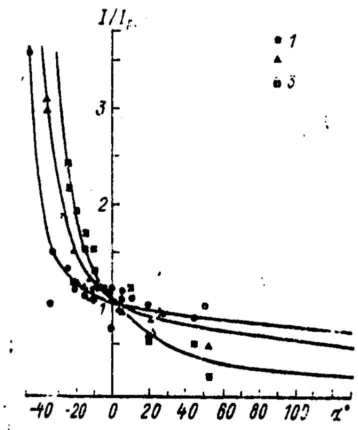

Fig. 1. Curves of the dependence $I/I_{chan\ br} = f(h_{chan}/h_{chan\ br}, \alpha)$.
1) with $h_{chan}/h_{chan\ br} = 1.1$; 2) with $h_{chan}/h_{chan\ br} = 1.25$; 3) with $h_{chan}/chan\ br = 1.5$.

The nature of the change in the free surface slopes is essentially different with an increase in the α angles from 0 to 50° (third type of interaction of flows). The curves of the dependence $I/I_{chan\ br} = f(\alpha)$ for different relative depths were also close to parallel and have an insignificant slope to the x-axis (α). The free surface slopes in this range of change in angles decrease from 1.0 to 0.9 with $h_{chan}/h_{chan\ br} = 1.10$ and to 0.4 with $h_{chan}/h_{chan\ br} = 1.50$.

The resulting dependences $I/I_{chan\ br} = f(\alpha, h_{chan}/h_{chan\ br})$ are quite close, their correlation ratios (for different depths) vary from 0.82 to 0.94, and the standard deviations vary from 0.59 to 0.79.

The α angle evidently inadequately characterizes the morphological structure of the computed reach. A significant influence is also exerted by the relative widths of the floodplain (B_{fl}/B_{chan}) and other morphometric characteristics. However, the correlation coefficients between the values of the free surface slopes ($I/I_{chan\ br}$) and these morphometric characteristics are small (< 0.5), which makes it impossible to recommend their use in computations.

The second objective of this study was an investigation of the problem of the formation of loops on the curves of the dependences $Q = f(H)$, $v = f(H)$ and $I = f(H)$ with the passage of high waters along the floodplain. As is well known, this problem has concerned a number of Soviet [4, 5 and others] and foreign researchers. The general conclusions from these studies can be summarized as follows: "... with an identical high-water wave in each section of the flow there is first a velocity maximum, then a discharge maximum and then a height maximum" ([4], p 330).

Thus, the authors assert that with the passage of a high-water wave the slopes, current velocities and water discharges in its frontal part exceed the corresponding values in the rear part. Accordingly, the dependences Q = f(H); v = f(H) and I = f(H) for the high-water period should be looped and the curve for the rise must be situated to the right of the dropoff curve. This finding is only partially confirmed by data from field observations (Ob' River at Barnaul, etc.), because at a number of posts (Luga River at Tolmachevo village and elsewhere) on the mentioned curves there is an opposite arrangement of curves, that is, the branch for the rise is situated to the left of the branch for the dropoff, but for the majority of rivers unambiguous curves of the dependence of the mentioned parameters on water levels are usually drawn. Such a character of the dependences Q = f(H), v = f(H) and I = f(H) is difficult to explain on the basis of the classical concepts set forth by M. A. Velikanov and other authors. It is particularly difficult to explain this situation with the passage of high waters along an inundated floodplain.

In 1978 N. B. Baryshnikov proposed a hypothesis explaining the formation of loops for rivers with floodplains, especially loops with an opposite positioning of the branches for the rise and dropoff [1]. However, this hypothesis also requires further investigations and improvement.

The process of formation of the velocity field of channel flow and the carrying capacity of a channel is influenced to a considerable degree by two oppositely directed factors. The first is the accumulation of water by the floodplain and channel during the period of rising of levels and its entry into the channel from the floodplain during its evacuation. The question arises as to the choice of length of the reach in which the computed volume is determined. Taking into account the "beaded" nature of the change in widths of floodplains [6], the length of the reach used for computational purposes is assumed to be the distance between successive narrowings of the floodplain in the region where hydrological measurements are made.

The second factor is the slowing of channel flow by the floodplain flow during the dropoff of high water and an increase in its velocities during the rising of levels due to the outflow of masses of water along the floodplain.

Thus, the balance equation will be written in the following form:

$$W_{drop_i} - W_{rise_i} \mp W^*_{i\,drop} \pm W^*_{i\,rise} = \Delta W_i, \quad (1)$$

where W_{rise_i} is the volume of outflow of water from the channel onto the floodplain during the period of rise of levels; W_{drop_i} is the volume of water inflow into the channel from the floodplain during the period of the dropping-off (equal to the volume W_{rise_i}, excluding losses in the filling of undrainable relief depressions, infiltration and evaporation); $W^*_{i\ drop-rise}$ is the change in the volume of runoff in the channel during the rising ($W^*_{i\ rise}$) or dropoff ($W^*_{i\ drop}$) of levels due to interaction between the channel and floodplain flows; i is a subscript denoting the phase of filling or emptying of the floodplain and assuming the values 1, 2, 3; ΔW_i is the algebraic sum of these volumes, characterizing the divergence of the branches of the curves of the dependence Q = f(H) with the rising and dropoff of levels.

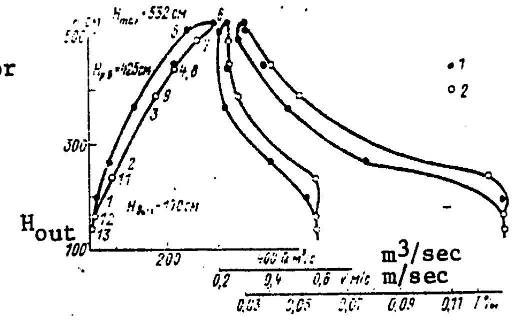

Fig. 2. Curves of the dependence Q = f(H), v = f(H) and I = f(H). Luga River, Tolmachevo village, 1977.

With the formation of a unified flow the nature of the interaction of its channel and floodplain components will be determined by the morphological structure of the reach used in the computations [1, 2].

In order to confirm these concepts we carried out computations in the example of a number of rivers, especially the Luga River at Tolmachevo village.

Figure 2 shows that with the passage of a high water of low guaranteed probability in 1977 the water discharges, velocities and free surface slopes during the period of rising of levels were considerably less than during the dropoff. Accordingly, in constructing curves of the dependences of these parameters on water level we obtained looped curves.

Such a nature of the change in discharges, velocities and free surface slopes can be explained on the basis of the advanced concept.

Floodplains are exceptionally diversified and their structure, dependent on their type, determines the nature of the interaction between the channel and floodplain flows. According to data published by the State Hydrological Institute in [8], the most common are the floodplains of freely meandering rivers, which includes the floodplain of the Luga River at Tolmachevo village. Precisely for this type of floodplains it was possible to discriminate three phases of their filling and emptying in [1].

In the first phase of inundation the water enters onto the floodplain from the channel through breaks situated in the lower parts of the floodplain sectors. On the latter there will be currents opposite in direction to the channel flow, intensive infiltration, filling of different isolated water bodies and depressions in the relief, and on this almost the entire mass of water coming from the channel will be expended. At the same time, as a result of the outflow of masses of water from the channel onto the floodplain there will be some increase in velocities in the channel part of the flow (interaction of channel and floodplain flows close to the second type).

The reverse phenomenon is observed in the similar phase of emptying of the floodplains. The water masses enter the channel from the floodplain sectors and the channel flow is slowed.

Thus, for this phase of filling and emptying of floodplains equation (1) can be written in the form

$$W_1^* \text{ rise} - W_{\text{rise}_1} - W_1^* \text{ drop} + W_{\text{drop}_1} = \Delta W_1. \tag{2}$$

Depending on the morphological characteristics of the channel and floodplain these characteristics can vary in a considerable range and the ΔW_1 value can have a positive, negative or zero value.

The second phase of inundation is characterized by the entry of masses of channel flow onto the floodplain through upstream breaks and low places in the banks and the formation of a transitory flow on it. In this case the channel and the floodplain flows are separated by a longitudinal river bank.

Equation (1) will also be valid for the second phase of floodplain inundation. However, the values of the terms in the equation change considerably. For example, $W_{\text{drop}2}$ in absolute value will be close to W_{rise_2} (the magnitude of the losses is determined only by evaporation and infiltration), but has an opposite sign. The determination of W_2^* rise, 2 drop is more complex because in the upper parts of the floodplain sectors the water on them arrives from the channel, whereas in the lower parts the reverse picture is observed, that is, the flow is from the floodplain sectors into the channel. Thus, the value and the sign of W_2^* rise and W_2^* drop are dependent on the morphological structure of the reach and the location of the computation point on the floodplain sector.

Equation (1) for the second phase of filling and emptying of floodplains can be represented in the form

$$W_{\text{drop}2} - W_{\text{rise}2} \pm W_2^* \text{ rise} \pm W_2^* \text{ drop} = \Delta W_2. \tag{3}$$

It should be noted that in all phases of filling and emptying of floodplains the ΔW_i value is considerably influenced by the measurement errors and the computation methods.

The third phase is characterized by inundation of the channel bank brows. The channel and floodplain flows, merging, form a unified channel-floodplain flow. The values of the terms W_{rise_i} and W_{drop_i} in equations (2), (3) are determined by the depth of flooding and the dimensions of the floodplain in the computation sector. The same as in the second phase, $W_{\text{rise}3}$ in absolute value is close to $W_{\text{drop}3}$, differing only with respect to the magnitudes of the losses in evaporation and filtration during periods of the rise and fall in levels:

$$W_{\text{drop}3} - W_{\text{rise}3} \pm W_3^* \text{ rise} \pm W_3^* \text{ drop} = \Delta W_3. \tag{4}$$

Taking into account that the brows of the channel banks are inundated, the nature of the interaction between the channel and floodplain flows will be similar for this phase of filling and emptying of the floodplain and therefore in the first approximation we will use $W^*_{3\,rise} = W^*_{3\,drop}$. However, an unsteady regime of movement of the flows will exert a substantial influence on the process of their interaction, but it is difficult to take it into account. Special laboratory and field investigations of this process are required. Accordingly, at the present time only a rough estimate of these parameters has been made.

Thus, equation (4) for the third phase of filling and emptying of floodplains can be represented in the form

$$W_{drop_3} - W_{rise_3} \approx \Delta W_3. \tag{5}$$

We used field data for confirming the proposed hypothesis. In particular, in 1977 a high water with a close to 1% guaranteed probability passed along the Luga River at Tolmachevo village. The individial phases of this high water are quite completely and reliably covered by measurement data (Fig. 2).

The W_{rise_i} and W_{drop_i} values were determined for a computation reach of the Luga River with a length of 9 km situated between two successive narrowings of a valley in which there are virtually no floodplains. The computed $W^*_{i\,rise}$ and $W^*_{i\,drop}$ values were determined from the refined graphic dependences

$$v_{chan}/v_{chan\ br.} = f(h_{chan}/h_{chan\ br},\ \alpha,\ B_{flood}/B_{chan}),$$

proposed by N. B. Baryshnikov in [3] for the second and third types of interaction of channel and floodplain flows. The results of the computations are given in the table.

Values of Parameters ($10^6 m^3$) in Equations (2)-(5) for Luga River at Tolmachevo

Phase of floodplain inundation	$W_{drop_i} - W_{rise_i}$	$W^*_{i\,rise} - W^*_{i\,drop}$	ΔW_i	ΔW_{cr}	$\frac{\Delta W_i - \Delta W_{cr}}{\Delta W_{cr}}$ %
First and second	-3.8	-2.1	-5.9	-5.5	7.3
Third	37.2	--	37.2	18.4	102

The computed ΔW_i value was compared with the ΔW_{cr} value, determined as the difference in the runoff volumes during the period of rising and dropoff of water levels in different phases of filling and emptying of the floodplain (Fig. 2).

Due to the absence or low accuracy of some initial data the computations were made for the first-second and separately for the third phases of filling and emptying of the floodplain. The table shows that for the first and second phases there is a good agreement between the computed and field data. The discrepancy of 7.3% is in the limits of accuracy in measuring water discharges in the high-water period. At the same time, the deviation of the computed data from the actual data for the third phase is 102%, evidence of shortcomings in the proposed method and the need for its further improvement for this phase.

It is evidently necessary to make additional allowance for the influence of unsteady movement, introducing substantial corrections into the processes of interaction between channel and floodplain flows, filling and emptying of floodplains.

However, the proposed method makes it possible to evaluate the position of the rise and dropoff branches on the water discharge and velocity curves during the passage of high waters at floodplain measurement points.

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ERRORS IN MEASURING WATER DISCHARGES BY THE 'VELOCITY-AREA' METHOD

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[Article by V. A. Rumyantsev, candidate of technical sciences, submitted for publication 17 Jun 80]

[Text]

Abstract: The article gives expressions making it possible to ascertain the systematic errors and dispersions of random errors in determining elementary water discharges and in measuring the total water discharge by the "velocity-area" method.

An improvement in methods for determining runoff requires the ability for making a proper evaluation of the accuracy in measuring water discharges. Recently this subject has been devoted much attention in the hydrological literature both in our country [1-11] and abroad [12-15].

Specialists at the State Hydrological Institute, beginning in 1966, have carried out a major complex of studies for evaluating the accuracy and optimizing measurements of water discharges. However, the discussion recently arising on the pages of this journal [4, 10] has shown that there are still many disputable aspects of this problem attributable primarily to the fact that the solutions obtained up to the present time have been based on different kinds of assumptions and it is not possible to evaluate the degree of their approximation. Accordingly, the need has arisen for a further continuation of these studies. In this article an attempt has been made in the most general formulation to derive expressions for determining the systematic error and dispersion of random errors in measuring water discharge by the "velocity-area" method, the method used most commonly in practical work.

When using the "velocity-area" method the total water discharge through the cross section of the flow $\hat{Q}$ is obtained by summation of the water discharges $\hat{q}$ passing through elementary surfaces

$$\hat{Q} = \sum_{i=1}^{n+1} \hat{q}_i, \tag{1}$$

where n is the number of verticals. We will designate the error in measuring the elementary water discharge by $e_{\hat{q}_i}$. Then the error in the total discharge will be

$$e_{\hat{Q}} = \sum_{i=1}^{n+1} e_{\hat{q}_i}. \tag{2}$$

The error in measuring the water discharge $e_{\hat{Q}}$ is a random parameter conforming to some distribution law. We will then characterize this distribution law by two parameters -- the mean value $\bar{e}_{\hat{Q}}$ and the dispersion $De_{\hat{Q}}$. We note that the mean value in this case is the systematic error in measuring the water discharge.

Taking the mathematical expectation of both sides of expression (2), we obtain

$$\bar{e}_{\hat{Q}} = \sum_{i=1}^{n+1} \bar{e}_{\hat{q}_i}, \tag{3}$$

where $\bar{e}_{\hat{q}_i}$ is the systematic error in measuring the discharge of water flowing through an elementary surface bounded by the (i-1)-st and i-th verticals. For the dispersion $De_{\hat{Q}}$ we have the expression

$$De_{\hat{Q}} = \sum_{i=1}^{n+1} De_{\hat{q}_i} + 2 \sum_{i<j}^{n+1} R(e_{\hat{q}_i}, e_{\hat{q}_j}), \tag{4}$$

where $De_{\hat{q}_i}$ is the dispersion of the random errors in measuring the elementary discharge,

$R(e_{\hat{q}_i}, e_{q_j})$ is the covariation between the errors in measuring the elementary discharges.

The error in measuring an elementary discharge $e_{\hat{q}_i}$ is the difference between the measured $\hat{q}_i$ and actual q_i values

$$e_{\hat{q}_i} = \hat{q}_i - q_i, \tag{5}$$

where the true q_i value is determined as

$$q_i = \iint_{(\omega_i)} v_{(\alpha,\varphi)}\, d\alpha\, d\varphi, \tag{6}$$

and the value

$$\hat{q}_i = \hat{v}_i \hat{\omega}_i. \tag{7}$$

In expression (6) integration is carried out for the area of the section ω_i bounded by the (i-1)-st and i-th velocity verticals, $v(\alpha, \varphi)$ is the water velocity at the point of the cross section of the flow with the coordinates $(\alpha, \varphi) \in (\omega_i)$, and in expression (7) $\hat{v}_i$ is the mean measured water velocity in the section, $\hat{\omega}_i$ is the adopted area of the section.

In reckoning the water discharge Q for obtaining the mean velocity in the section $\hat{v}_i$ and its area $\hat{\omega}_i$ it is customary to use the following expressions:

$$\hat{v}_i = \begin{cases} \mu \tilde{v}_1 & \text{when } i = 1, \\ 0{,}5\,(\tilde{v}_{i-1} + \tilde{v}_i) & \text{when } i = 2 - n, \\ \gamma \tilde{v}_n & \text{when } i = n+1, \end{cases} \tag{8}$$

$$\hat{\omega}_i = \begin{cases} 0{,}5\ \tilde{h}_i \tilde{L}_1 & \text{when } i = 1, \\ 0{,}5\ (h_{i-1} + \tilde{h}_i)\ \tilde{L}_i & \text{when } i = 2 - n, \\ 0{,}5\ h_n L_{n+1} & \text{when } i = n + 1, \end{cases} \tag{9}$$

where μ and γ are the boundary coefficients for the shore sections, $\tilde{v}_i$ is the mean water velocity at the vertical,

$$\tilde{v}_i = \frac{1}{m} \sum_{k=1}^{m} \tilde{v}_{ik} \quad (i = 1, 2, \ldots, n), \tag{10}$$

v_{ik} is the measured water velocity at the k-th point of the i-th vertical, m is the number of points of velocity measurement at the verticals, equal to 1, 2, 3 or 5, depending on the depth, velocity distribution at the verticals and measurement conditions (channel overgrown with aqueous vegetation, presence of an ice cover, etc.), $\tilde{h}_i$ is the measured depth, $\tilde{L}_i$ is the measured distance between the (i-1)-st and i-th verticals.

Thus, for computing the elementary discharge we have the evaluation

$$\hat{q}_i = \begin{cases} \dfrac{0{,}5\,\mu}{m}\ \tilde{h}_1 \tilde{L}_1 \sum\limits_{k=1}^{m} \tilde{v}_{1,k} & \text{when } i = 1, \\ \dfrac{0{,}25}{m}\ (\tilde{h}_{i-1} + \tilde{h}_i)\ \tilde{L}_i \sum\limits_{k=1}^{m} (\tilde{v}_{i-1,k} + \tilde{v}_{i,k}) & \text{when } i = 2 - n, \\ \dfrac{0{,}5\,\gamma}{m}\ \tilde{h}_n \tilde{L}_{n+1} \sum\limits_{k=1}^{m} \tilde{v}_{n,k} & \text{when } i = n + 1. \end{cases} \tag{11}$$

However, in this study it seems desirable to examine an evaluation of an elementary discharge $\hat{q}_i$ of a more general form, specifically

$$\hat{q}_i = \hat{\omega}_i \left(\sum_{k=1}^{m_{i-1}} a_{i-1,k}\ \tilde{v}_{i-1,k} + \sum_{k=1}^{m_i} a_{i,k}\ \tilde{v}_{i,k} \right), \tag{12}$$

which allows a different number of points for measurement of velocity m_i at each vertical and also stipulation of the weighting coefficients $a_{i,k}$, taking into account current concepts concerning the hydraulic conditions of the flow. In this case the expression for computing the error in an elementary discharge has the form

$$e_{\hat{q}_i} = \hat{\omega}_i \left(\sum_{k=1}^{m_{i-1}} a_{i-1,k}\ \tilde{v}_{i-1,k} + \sum_{k=1}^{m_i} a_{i,k} \hat{v}_{i,k} \right) - \iint_{(\omega_i)} v_{(\alpha,\varphi)}\, d\alpha\, d\varphi. \tag{13}$$

The measured values enter into expressions (11)-(13). These can be represented in the form of the sum of the actual value (represented henceforth without a prime) and the error in its measurement (with a prime)

$$\hat{\omega}_i = \omega_i + \omega'_i, \quad \tilde{v}_{i,k} = v_{i,k} + v'_{i,k}, \left(\begin{matrix} i = 1, 2, \ldots, n \\ k = 1, 2, \ldots, m \end{matrix} \right). \tag{14}$$

Proceeding on the basis of an analysis of the measurement conditions, in the ratio of errors we will assume the following:

$$\overline{\omega_i'} = \Delta\omega_i; \quad \overline{v_{i,k}'} = 0; \tag{15}$$

$$\overline{v_{jk}' \cdot v_{j,l}'} = \begin{cases} \overline{D}v' & \text{with } i = j \wedge k = l, \\ 0 & \text{in the remaining cases;} \end{cases} \tag{16}$$

$$\overline{v_{i,k} \cdot v_{j,l}'} = \begin{cases} R(v, v') & \text{with } i = j \wedge k = l, \\ 0 & \text{in the remaining cases;} \end{cases} \tag{17}$$

$$\overline{\omega_i' \cdot \omega_j'} = \begin{cases} D\omega_i' & \text{with } i = j \\ 0 & \text{in the remaining cases;} \end{cases} \tag{18}$$

$$\overline{\omega_i' \, v_{j,k}'} = \overline{\omega_i' \, v_{i,k}} = \overline{\omega_i' \, \omega_j} = \overline{\omega_i' v_{j,k}} = 0 \quad \text{with } \forall\ i, j, k. \tag{19}$$

Here the line at top denotes averaging for an infinite set of records, $\Delta\omega_i$ is the systematic error of the section area, Dv' is the dispersion of the random errors in measuring velocity at a point, R(v, v') is the covariation between the velocity of water at a point and the error in its measurement, $D\omega_i'$ is the dispersion of the random errors in measuring the area of the section.

With the assumptions made (14)-(19) taken into account, taking the mathematical expectation of both sides of expression (13), we obtain the systematic error in measuring the elementary discharge.

$$\overline{e}_{\hat{q}_i} = (\omega_i + \Delta\omega_i)\left(\sum_{k=1}^{m_{i-1}} a_{i-1,k}\,\overline{v}_{i-1,k} + \sum_{k=1}^{m_i} a_{i,k}\,\overline{v}_{i,k}\right) - \iint_{(\omega_i)} \overline{v}_{(\alpha,\varphi)}\, d\alpha\, d\varphi, \tag{20}$$

where $\overline{v}_{(\alpha,\varphi)}$ and $\overline{v}_{i,k}$ are the current velocity values averaged for an infinite set of records with a given value of the water discharge Q at the point with the coordinates $(\alpha, \varphi) \in (\omega_i)$ and at the k-th point of the i-th vertical respectively. Substituting the latter expression into (3), we will have an expression for finding the systematic error $e_{\hat{Q}}$ in the total water discharge

$$\overline{e}_{\hat{Q}} = \sum_{i=1}^{n+1} (\omega_i + \Delta\omega_i)\left(\sum_{k=1}^{m_{i-1}} a_{i-1,k}\,\overline{v}_{i-1,k} + \sum_{k=1}^{m_i} a_{i,k}\,\overline{v}_{i,k}\right) - \iint_{(\omega)} \overline{v}_{(\alpha,\varphi)}\, d\alpha\, d\varphi, \tag{21}$$

and here integration is already carried out for the entire area of the flow cross section.

In order to derive expressions for the dispersion of the random errors in measuring the elementary discharge $De_{\hat{q}_i}$ and the covariations between the errors in measuring discharges in different sections $R(e_{\hat{q}_i}, e_{\hat{q}_j})$ it is more convenient to use the following representations:

$$De_{\hat{q}_i} = D\hat{q}_i - 2R(\hat{q}_i, q_i) + Dq_i \tag{22}$$

$$R(e_{\hat{q}_i}, e_{\hat{q}_j}) = R(\hat{q}_i, \hat{q}_j) - R(\hat{q}_i, q_j) - R(q_i, \hat{q}_j) + R(q_i, q_j), \tag{23}$$

where, as before, D represents the dispersion operator and R represents the operator for covariation between the random values in the parentheses. Omitting the intermediate computations, we immediately give the final expressions for De_{q_i} and $R(e_{\hat{q}_i}, e_{\hat{q}_j})$:

$$\begin{aligned} De_{\hat{q}_i} &= (\omega_i^2 + D\omega_i')\Bigg[\sum_{k=1}\sum_{l=1}^{m_{i-1}} a_{i-1,k}\, a_{i-1,k}\, R(v_{i-1,k}, v_{i-1,l}) + \\ &+ 2\sum_{k=1}^{m_{i-1}}\sum_{l=1}^{m_i} a_{i-1,k}\, a_{i,l}\, R(v_{i-1,k}, v_{i,l}) + \sum_{k=1}^{m_i}\sum_{l=1} a_{i,k}\, a_{i,l}\, R(v_{,k}, v_{i,l}) + \\ &+ (2R(v, v') + Dv')\left(\sum_{k=1}^{m_{i-1}} a_{i-1,k}^2 + \sum_{k=1}^{m_i} a_{i,k}^2\right)\Bigg] - 2\omega_i \iint_{(\omega_i)} \Bigg[\sum_{k=1}^{m_{i-1}} a_{i-1,k} \times \\ &\times R(v_{i-1,k}, v_{(\alpha,\varphi)}) + \sum_{k=1}^{m_i} a_{i,k}\, R(v_{i,k}, v_{(\alpha,\varphi)})\Bigg]\, d\alpha\, d\varphi + \\ &+ \iint_{(\omega_i)}\iint_{(\omega_i)} R(v_{(\alpha_1,\varphi_1)}, v_{(\alpha_2,\varphi_2)})\, d\alpha_1\, d\varphi_1\, d\alpha_2\, d\varphi_2, \quad j - i = 1, \end{aligned} \tag{24}$$

$$\begin{aligned} R(e_{\hat{q}_i}, e_{\hat{q}_j}) &= \omega_i\omega_j \Bigg[\sum_{k=1}^{m_{i-1}}\sum_{l=1}^{m_{j-1}} a_{i-1,k}\, a_{j-1,l}\, R(v_{i-1,k}, v_{j-1,l}) + \\ &+ \sum_{k=1}^{m_{i-1}}\sum_{l=1}^{m_j} a_{i-1,k}\, a_{j,l}\, R(v_{i-1,k}, v_{j,l}) + \sum_{k=1}^{m_i}\sum_{l=1}^{m_{j-1}} a_{i,k}\, a_{j-1,l} \times \\ &\times R(v_{i,k}, v_{j-1,l}) + \sum_{k=1}^{m_i}\sum_{l=1}^{m_j} a_{i,k}\, a_{j,l}\, R(v_{i,k}, v_{j,l}) + (2R(v, v') + Dv') \times \\ &\times \sum_{k=1}^{m_i} a_{i,k}^2\Bigg] - \omega_i \iint_{(\omega_j)} \Bigg[\sum_{k=1}^{m_{i-1}} a_{i-1,k}\, R(v_{i-1,k}, v_{(\alpha,\varphi)}) + \\ &+ \sum_{k=1}^{m_i} a_{i,k}\, R(v_{i,k}, v_{(\alpha,\varphi)})\Bigg]\, d\alpha\, d\varphi - \omega_j \iint_{(\omega_i)} \Bigg[\sum_{k=1}^{m_{j-1}} a_{j-1,k}\, R(v_{j-1,k}, v_{(\alpha,\varphi)}) + \\ &+ \sum_{k=1}^{m_j} a_{j,k}\, R(v_{j,k}, v_{(\alpha,\varphi)})\Bigg]\, d\alpha\, d\varphi + \\ &+ \iint_{(\omega_i)}\iint_{(\omega_j)} R(v_{(\alpha_1,\varphi_1)}, v_{(\alpha_2,\varphi_2)})\, d\alpha_1\, d\varphi_1\, d\alpha_2\, d\varphi_2, \quad j - i > 1, \end{aligned} \tag{25}$$

$$R(e_{\hat{q}_i}, e_{\hat{q}_j}) = \omega_i \omega_j \left[\sum_{k=1}^{m_{i-1}} \sum_{l=1}^{m_{j-1}} a_{i-1,k}\, a_{j-1,l}\, R(v_{i-1,k}, v_{j-1,l}) + \right.$$
$$+ \sum_{k=1}^{m_{i-1}} \sum_{l=1}^{m_j} a_{i-1,k}\, a_{j,l}\, R(v_{i-1,k}, v_{j,l}) + \sum_{k=1}^{m_i} \sum_{l=1}^{m_{j-1}} a_{i,k}\, a_{j-1,l}\, R(v_{i,k}, v_{j-1,l}) +$$
$$\left. + \sum_{k=1}^{m_i} \sum_{l=1}^{m_j} a_{i,k}\, a_{j,l}\, R(v_{i,k}, v_{j,l}) \right] - \omega_i \iint_{(\omega_j)} \left[\sum_{k=1}^{m_{i-1}} a_{i-1,k}\, R(v_{i-1,k}, v_{(\alpha,\varphi)}) + \right. \tag{26}$$
$$\left. + \sum_{k=1}^{m_i} a_{i,k}\, R(v_{i,k}, v_{(\alpha,\varphi)}) \right] d_\alpha\, d_\varphi - \omega_j \iint_{(\omega_i)} \left[\sum_{k=1}^{m_{j-1}} a_{j-1,k}\, R(v_{j-1,k}, v_{(\alpha,\varphi)}) + \right.$$
$$\left. + \sum_{k=1}^{m_j} a_{j,k}\, R(v_{j,k}, v_{(\alpha,\varphi)}) \right] d_\alpha\, d_\varphi +$$
$$+ \iint_{(\omega_i)} \iint_{(\omega_j)} R(v_{(\alpha_1,\varphi_1)}, v_{(\alpha_2,\varphi_2)})\, d\alpha_1\, d\varphi_1\, d\alpha_2\, d\varphi_2 .$$

In these expressions $R(v_{j,k}, v_{j,\ell})$ is the covariation between the water velocities at the i-th vertical at the k-th point at the time t_{ik} and the j-th vertical at the ℓ-th point at the time $t_{j\ell}$; $R(v_{i,k}, v_{(\alpha,\varphi)})$ is the covariation between the water velocities at the k-th point of the i-th vertical at the time t_{ik} and at the flow cross-sectional point with the coordinates (α, φ) at the time $t_{ik} + t_{nm}/2$ (the remaining notations are as before).

Substituting expression (24)-(26) into (4), we obtain a final expression for the dispersion of the random errors in measuring water discharge Q by the "velocity-area" method:

$$De_{\hat{Q}} = \sum_{i=1}^{n+1} (\omega_i^2 + D\omega_i') \left[\sum_{k=1}^{m_i} \sum_{l=1} a_{i-1,k}\, a_{i-1,l}\, R(v_{i-1,k}, v_{i-1,l}) + \right.$$
$$+ 2 \sum_{k=1}^{m_{i-1}} \sum_{l=1}^{m_i} a_{i-1,k}\, a_{i,l}\, R(v_{i-1,k}, v_{i,l}) + \sum_{k=1}^{m_i} \sum_{l=1} a_{i,k}\, a_{i,l}\, R(v_{i,k}, v_{i,l}) +$$
$$\left. + \left(\sum_{k=1}^{m_{i-1}} a_{i-1,k}^2 + \sum^{m_i} a_{i,k}^2 \right) (Dv' + 2R(v, v')) \right] + 2 \left[\sum_{i<j}^{n+1} \left\{ \omega_i \omega_j \times \right. \right. \tag{27}$$
$$\times \left[\sum_{k=1}^{m_{i-1}} \sum_{l=1}^{m_{j-1}} a_{i-1,k}\, a_{j-1,l}\, R(v_{i-1,k}, v_{j-1,l}) + \sum_{k=1}^{m_i-1} \sum_{l=1}^{m_j} a_{i-1,k}\, a_{j,l} \times \right.$$
$$\left. \times R\, v_{i-1,k}, v_{j,l}) + \sum_{k=1}^{m_i} \sum_{l=1}^{m_j-1} a_{i,k}\, a_{j-1,l}\, R(v_{i,k}, v_{j,l}) + \sum_{k=1}^{m_i} \sum_{l=1}^{m_j} a_{i,k} a_{j,l} R(v_{i,k}, v_{j,l}) \right] -$$

$$-\omega_l \iint_{(\omega_j)} \left[\sum_{k=1}^{m_{l-1}} a_{l-1,k} R(v_{l-1,k}, v_{(\alpha,\varphi)}) + \sum_{k=1}^{m_l} a_{l,k} R(v_{l,k}, v_{(\alpha,\varphi)}) \right] d\alpha\, d\varphi - \tag{27}$$

$$- \omega_j \iint_{(\omega_l)} \left[\sum_{k=1}^{m_{j-1}} a_{j-1,k} R(v_{j-1,k}, v_{(\alpha,\varphi)}) + \sum_{k=1}^{m_j} a_{j,k} R(v_{j,k}, v_{(\alpha,\varphi)}) \right] \times$$

$$\left. \times d\alpha\, d\varphi \right\} + (2R(v, v') + Dv') \sum_{l=1}^{n-1} \omega_l \omega_{l+1} \sum_{k=1}^{m_l} a_{l,k}^2 \Bigg] +$$

$$+ \iint_{(\omega)} \iint_{(\omega)} R(v_{(\alpha_1,\varphi_1)}, v_{(\alpha_2,\varphi_2)})\, d\alpha_1\, d\varphi_1\, d\alpha_2\, d\varphi_2$$

(all the notations are as before).

We note that the error in measuring water discharge is dependent on its quantity. The covariation coefficient is

$$R(e_{\hat{Q}}, Q) = \sum_{l=1}^{n+1} \omega_l \iint_{(\omega)} \left[\sum_{k=1}^{m_{l-1}} a_{l-1,k} R(v_{l-1,k}, v_{(\alpha,\varphi)}) + \sum_{k=1}^{m_l} a_{l,k} \times \right. \tag{28}$$

$$\left. \times R(v_{l,k}, v_{(\alpha,\varphi)}) \right] d\alpha\, d\varphi - \iint_{(\omega)} \iint_{(\omega)} R(v_{(\alpha_1,\varphi_1)}, v_{(\alpha_2,\varphi_2)})\, d\alpha_1\, d\varphi_1\, d\alpha_2\, d\varphi_2.$$

For the usual evaluation of the elementary water discharge (11), the systematic error $\bar{e}_{\hat{Q}}$ and the dispersion $De_{\hat{Q}}$ we have

$$\bar{e}_{\hat{Q}} = \frac{1}{2m} \left[\gamma(\omega_1 + \Delta\omega_1) \sum_{k=1}^{m} \bar{v}_{1,k} + \sum_{l=2}^{n} (\omega_l + \Delta\omega_l) \sum_{k=1}^{m} (\bar{v}_{l-1,k} + \bar{v}_{l,k}) + \right.$$

$$\left. + \mu(\omega_{n+1} + \Delta\omega_{n+1}) \sum_{k=1}^{m} \bar{v}_{n,k} \right] - \iint_{(\omega)} \bar{v}_{(\alpha,\varphi)}\, d\alpha\, d\varphi, \tag{29}$$

$$De_{\hat{Q}} = \frac{1}{m^2} \left\{ m(Dv' + 2R(v, v')) \left[\gamma^2(\omega_1^2 + D\omega_1') + \mu^2(\omega_{n+1}^2 + \right. \right.$$

$$\left. + D\omega_{n+1}') + \gamma\omega_1\omega_2 + \mu\omega_n\omega_{n+1} + 0{,}5 \left(\sum_{l=2}^{n} \omega_{l-1}\omega_l + \sum_{l=2}^{n} (\omega_l^2 + D\omega_l') \right) \right] +$$

$$+ \sum_{k=1}^{m} \sum_{l=1} \left[\gamma^2(\omega_1^2 + D\omega_1^1) R(v_{1,k} + v_{1,l}) + \mu^2(\omega_{n+1}^2 + D\omega_{n+1}') R(v_{n,k}, v_{n,l}) + \right.$$

$$+ 2\gamma\mu\omega_1\omega_{n+1} R(v_{1,k}, v_{n,l}) + \sum_{l=2}^{n} \omega_l [\gamma\omega_1 (R(v_{1,k}, v_{l-1,l}) +$$

$$+ R(v_{1,k}, v_{i,l})\Big) + \mu\omega_{n+1}\Big(R(v_{l-1,k}, v_{n,l}) + R(v_{l,k}, v_{n,l})\Big] +$$

$$+ 0{,}25\Big[\sum_{i=2}^{n}\sum_{j=2}\omega_i\omega_j\Big(R(v_{i-1,k}, v_{j-1,l}) + R(v_{i,k}, v_{j,l}) + \qquad (29)$$

$$+ 2R(v_{i-1,k}, v_{j,l})\Big) + \sum_{i=2}^{n} D\omega_i'\Big(R(v_{i-1,k}, v_{i-1,l}) + R(v_{i,k}, v_{i,l}) +$$

$$+ 2R(v_{i-1,k}, v_{i,l})\Big)\Big]\Big\} - \frac{1}{m}\sum_{k=1}^{m}\iint_{(\omega)}\Big[2(\gamma\omega_1 R(v_{1,k}, v_{(\alpha,\varphi)}) +$$

$$+ \mu\omega_{n+1}R(v_{n,k}, v_{(\alpha,\varphi)})\Big) + \sum_{i=2}^{n}\omega_i\Big(R(v_{i-1,k}, v_{(\alpha,\varphi)}) +$$

$$+ R(v_{i,k}, v_{(\alpha,\varphi)})\Big)\Big]d\alpha\, d\varphi + \iint_{(\omega)}\iint_{(\omega)} R(v_{(\alpha_1,\varphi_1)}, v_{(\alpha_2,\varphi_2)})d\alpha_1 d\varphi_1 d\alpha_2 d\varphi_2. \qquad (30)$$

If it is assumed that the error in measuring depth h and the distance L between verticals conforms to the normal law

$$h_i' \sim N(\Delta h,\ \overline{Dh'}),\quad L_i' \sim N(\Delta L,\ \overline{DL'}) \qquad (31)$$

and there is satisfaction of the condition

$$\overline{h_i'\ h_j'} = \overline{L_i'\ L_j'} = \overline{h_i'\ L_j'} = \overline{h_i'\ L_j} = \overline{h_i'\ h_j} = \overline{L_i'\ L_j} = \overline{L_i'\ h_j} = 0, \qquad (32)$$

then with these assumptions the expressions for computing the systematic error in determining the area of the section $\Delta\omega_i$ and the dispersion of the random errors $D\omega_i'$ have the form

$$\Delta\omega_i \approx \Big(\frac{h_{i-1} + h_i}{2} + \Delta h\Big)\Big(L_i + \Delta L\Big) - \int_{l_{i-1}}^{l_{i-1}+L_i} h(l)\, dl, \qquad (33)$$

$$D\omega_i' \approx \frac{1}{4}\{2\, Dh'\,[(L_i + \Delta L)^2 + DL'] + DL'\,(h_{i-1} + h_i + 2\,\Delta h)^2\}. \qquad (34)$$

In these expressions h and L are the real values for depth and the distance between verticals, l_{i-1} is the distance from the channel edge to the (i-1)-st velocity vertical, Δh and ΔL are the systematic errors in measuring depths and distances between verticals respectively, Dh' is the dispersion of errors in measuring the depth of the flow, DL' is the dispersion of errors in measuring distance between velocity verticals.

The $\Delta\omega_i$ value is dependent to a considerable degree on channel bottom relief: the greater the dissection of the bottom in the distance between the verticals, the greater is the $\Delta\omega_i$ value, whereas the $D\omega_i$ value is determined by the depths and the distances between the velocity verticals, with an increase in which it increases.

It can be seen from expressions (29) and (33) that the systematic error in measuring water discharge by the "velocity-area" method is dependent on the number of velocity verticals n and the number of readings at them m and also on the variability of the flow in the section and in time. The greater the distance between the velocity verticals L or the lesser the number of points for the measurement of velocity m, the greater is the change in the time-averaged velocity in the cross-sectional area and the greater is the time shift between measurements and the greater will be the value of the systematic error $\bar{e}_{\hat{Q}}$.

The $De_{\hat{Q}}$ value is determined to a considerable degree by the statistical structure of the water velocity field in the flow cross section. The lesser the coherence between the water velocities at different points in the cross section, the greater will be the dispersion of the random errors in measuring water discharge. The $De_{\hat{Q}}$ value is also influenced by the number of velocity verticals and measurement points, the ratio between the areas of the sections, the random errors in measuring velocity and the area of the sections.

It follows from expressions (27)-(29) that for finding the dispersion $De_{\hat{Q}}$ it is necessary to have the covariation function of the velocity field. Such a function can be obtained using either data from special frequent measurements of velocities within the limits of the flow cross section, or, which is less reliable, by the joint processing of archival data from standard velocity measurements in some stipulated range of water discharges or water levels. We note that the covariation functions with one and the same water discharge volumes can be dependent on the phase of the water regime due to the differences in the course of channel processes, hydraulic resistance of the channel, etc. Accordingly, they must be formed for selected typical periods (high water, low water, periods of overgrowth, etc.).

The expressions cited above for the systematic error $e_{\hat{Q}}$ and the dispersion of the random errors $De_{\hat{Q}}$ can find use both in evaluating the accuracy of the measured water discharges and in ascertaining the influence exerted on it by different factors and in developing simplified engineering methods for evaluating accuracy and also in improving computation formulas for determining water discharges.

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INFLUENCE OF METEOROLOGICAL CONDITIONS ON THE QUALITY OF POMEGRANATE FRUITS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 98-101

[Article by A. D. Eyyubov, doctor of geographical sciences, and Kh. Sh. Ragimov, Geography Institute Azerbaijan Academy of Sciences, submitted for publication 25 Mar 80]

[Text]

Abstract: A study was made of the influence of meteorological conditions during the growing season on the quality of pomegranate fruits. It was found that when the pomegranate crop is irrigated the principal factor responsible for the accumulation of a particular quantity of sugar and acid in the pomegranate juice is the thermal conditions during the maturing period (September). Equations are derived giving the relationship between the content of sugar and acid in the juice and also their ratio (sugar-acid coefficient) and the temperature sums during the maturing period. The best sour-sweet juices are obtained with temperature sums during the maturing period exceeding 690°. These correlation equations can be used in an agroclimatic evaluation of the territory for the purposes of pomegranate cultivation and also for predicting the anticipated quality of the yield.

The pomegranate is among the most valuable and promising fruit crops for the dry and arid subtropics of the USSR. Pomegranate juice is used in the treatment of a whole series of diseases, including cardiovascular, angina, fever, etc. Recently different varieties of pomegranate have been put into more and more new types of use.

With respect to their acid content, pomegranate juices can be classified as sweet (acid content up to 0.9%), sour-sweet (0.9-1.8%) and sour (more than 1.8%). However, a more perfect index of juice quality is the sugar-acid coefficient, expressed by the ratio of the sugar present in the juice to the acid. In the presence of other positive indices A. M. Murzayeva [5] proposed that the best sour-sweet juices be regarded as those with a sugar-acid coefficient in the range

7.5-11. B. S. Rozanov [7], generalizing the results of numerous taste tests of pomegranate juice, regards as acidic those juices with a sugar-acid coefficient from 1 to 6, sour-sweet juices -- those with a coefficient from 6.1 to 15 and sweet juices -- those with a coefficient greater than 15.

In this article an attempt is made to ascertain the quantitative relationships between the principal indices of the quality of pomegranate juice (content of sugars and acids and their ratio) and meteorological conditions.

An analysis of the influence of meteorological conditions on the chemical composition of pomegranate juice was made using the variety Gyuleysha Azerbaydzhanskaya (planting of 1949), which occurs widely in the Azerbaijan SSR and which has been adopted as a standard by the State Commission on Strain Testing of Agricultural Crops. Use was made of the materials [6] from the Geokchayskiy, Mir-Bashirskiy, Apsheronskiy and Lenkoranskiy State Strain Testing Stations, situated in different agroclimatic zones of the republic and the results of a chemical analysis of the fruits which we selected in 1975-1976.

A precise determination of the chemical composition of pomegranate juice characterizing the quality of the entire harvest involves great difficulties since under the influence of different factors the chemical composition of individual fruits on one and the same bush can differ very greatly. Among the factors responsible for these differences are primarily the time of infructescence, the degree of maturity and exposure of the fruits on the bush.

A chemical analysis of juice of pomegranate fruits of the harvest of 1976, selected in the orchard of the Apsheronskiy Strain Testing Station, indicated that the difference in the sugar content in the juice of fruits taken from bushes with a southerly and northerly exposure was 3.13%, whereas the acid content was 1.02%. These data are close to the results obtained by A. B. Kerimov [3], who noted that the difference in the content of sugar and acid in the juice between the shaded and unshaded fruits can attain 2-3%. Approximately the same differences are discovered in the content of sugar and acid in the juice of fruits of different generation. In the juice of pomegranate fruits infructescent during the period of the second flowering (15 days after the fruits of the first generation) the sugar was less by 2.54% and the acid was greater by 3.11%. For later fruits this difference attained still greater values and was 3.86 and 3.83%.

The time of harvesting of the fruits exerts a substantial influence on the accumulation of sugar and acid in the pomegranate juice. A difference in the times of harvesting of the fruits by 1.5 months (in the case of harvesting in September-October) leads to an increase in the content of sugars by 45-64% in comparison with the initial quantity and to a decrease in acids by a factor of almost 2 [7, 8].

Thus, an inadequate allowance for the above-mentioned factors in the selection of fruits can lead to considerable errors in determining the chemical composition of the juice.

An analysis of the material shows that during the warm and dry years more sugar is accumulated in the pomegranate juice and less acid is accumulated than during the cool and moist years. A comparison of data from a chemical analysis of pomegranate juice from the harvests of 1975 and 1976, which we collected in the orchard of the

Geokchayskiy control point of the Azerbaijan Scientific Research Institute of Agriculture and Seed Selection indicated that in the juice of pomegranates of the harvest of 1975 there was 2.4% more sugar and 0.62% more acid, or almost 1.5 times less than in the fruits of the harvest of 1976, which was attributable to the difference in meteorological conditions during the growing season of the plants during these years. The growing season of 1975 was warmer and drier. The sum of active temperatures during the growing season during this year was 186°C higher, and during the period of maturing (September) was 12°C higher than in 1976. During 1975 almost 2.5 times less precipitation fell than in 1976. Similar results were also obtained in [4], where it is noted that a decrease in the temperature sums during the maturing period by approximately 50°C leads to a decrease in sugar in the pomegranate juice by 4.46% and to an increase in the acid content by a factor of 3.3.

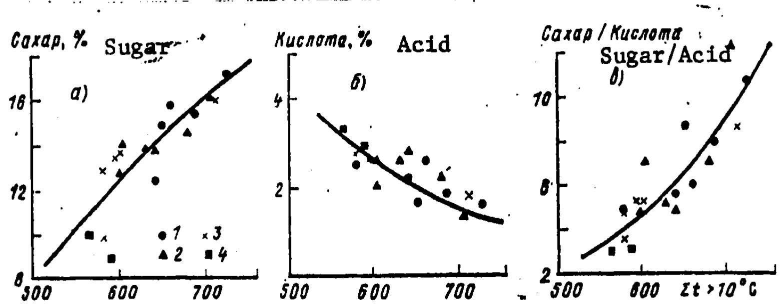

Fig. 1. Graphs of the dependence of sugar content (a), acidity (b) and sugar-acid coefficient (c) of pomegranate juice on sums of active temperatures during September. 1) Geokchayskiy SSSS (State Seed Selection Station), 2) Mir-Bashirskiy SSSS, 3) Apsheronskiy SSSS, 4) Lenkoranskiy SSSS.

A comparison of the indices of the quality of juice with such meteorological elements as precipitation, absolute and relative humidity and dew-point spread, and also the moistening indices Md and HTC (humidity-temperature coefficient), both for the entire growing season and for individual development phases, indicated that when the pomegranate crop is irrigated, when the soil air and moisture content is regulated by watering, the influence of variability of atmospheric moistening on the content of sugar and acid in the juice is insignificant. An exception is the years when the moistening attains anomalously high values and leads to a sharp deterioration of the quality of the harvest.

Since under the conditions prevailing in Azerbaijan the harvesting of pomegranate begins late in September, this month was adopted as the period of maturing of the fruits. The correlation between the sugar content of the pomegranate juice and the sums of active temperatures for September (see Fig. 1a) is approximated well by a logarithmic function in the following form:

$$y = 55{,}273 \lg x - 141{,}11; \quad \eta_{yx} = 0{,}737; \quad E_y = 6{,}89\%,$$

where y is the sugar content in the juice in percent in the wet mass, x is the sum of active temperatures during September, η_{yx} is the correlation ratio, E_y is the mean relative error of the equation.

In contrast to sugar, the acid in the pomegranate juice does not increase continuously. An increase in the content of acids occurs right up to the onset of the maturing phase when it attains its maximum [1, 2]. Then, beginning approximately from late August, the acidity of the juice decreases and the more rapidly the greater are the temperature sums which are accumulated during the maturing period. Accordingly, the correlation between the quantity of acid present in the pomegranate juice and the temperature sums for September is inverse (Fig. 1b) and is expressed by an exponential function in the form

$$y_1 = 32{,}96\, e^{-0{,}004225\, x}; \quad \eta_{y.x} = 0{,}947; \quad E_y = 9{,}87\%,$$

where y_1 is the acid content in the juice in percent of wet mass.

Since individual components of the quality of pomegranate juice (sugar, acid) are, as we ascertained, in close dependence on the thermal background of the maturing period, naturally the sugar-acid coefficient (y_2) of the pomegranate juice is dependent on the variability of the temperature conditions (see Fig. 1c). This correlation is expressed by the following equation:

$$y_2 = 0{,}21 \cdot 10^{-10}\, x^{4{,}32}; \quad \eta_{y_2 x} = 0{,}887; \quad E_y = 15{,}46\%.$$

In analyzing the cited dependences it can be noted that with one and the same thermal conditions the sugar content of the juice of fruits from the Apsheronskiy and Lenkoranskiy State Strain Testing Stations is somewhat less and the acid content is somewhat higher than in the fruits of the Geokchayskiy and Mir-Bashirskiy types, which is attributable primarily to the difference in soil conditions and to some degree to the increased air humidity in the coastal zone. However, the derived equations with sufficient accuracy can be used for the territory of Azerbaijan as a whole.

Proceeding on the basis of these equations, it can be assumed that the best sugar content (more than 12%) in the juice of pomegranates of the variety Gyuleysha Azerbaydzhanskaya will be in years when $\sum t > 10°C$ during September will exceed 590°C and the best acid content (less than 1.8%) with $\sum t > 10°C$, exceeding 690°C. The best sour-sweet [7] juices will be obtained with $\sum t > 10°C$, not less than 670°C. Thus, the minimum sums of active temperatures during September ensuring the optimum content of sugar and acid and their best ratio will be 690°C. With this temperature sum the quantity of sugar in the juice will exceed 15.8%, the acid will be less than 1.8%, whereas the sugar-acid coefficient will be 8.8, that is, will be close to the values when the pomegranate juice is characterized by the best taste [5].

Summary

1. The influence of meteorological conditions on the formation of the quality of pomegranate fruits is reflected primarily during the period of maturing of the fruits.

2. When the pomegranate crop is irrigated the influence of the variability of precipitation and humidity on the chemical composition of the pomegranate juice is insignificant.

3. The principal factor responsible for the accumulation of different quantities of sugar and acid in pomegranate fruits is the thermal conditions of the maturing period. The best sour-sweet juices from the pomegranate fruits are obtained with sums of active temperatures during the period of their maturing exceeding 690°C.

4. The derived correlations between the elements of the chemical composition of pomegranate juice and the sums of active temperatures during the maturing period (September) can be used in an agroclimatic evaluation of a territory for the purpose of cultivating pomegranate and also in predicting the anticipated quality of the yield of pomegranate fruits.

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PARAMETERS OF THERMAL STRATIFICATION OF THE PLANETARY BOUNDARY LAYER

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[Article by N. Ye. Dosycheva and A. M. Krigel', candidate of physical and mathematical sciences, Leningrad Hydrometeorological Institute, submitted for publication 25 Jul 80]

[Text]

Abstract: The least squares method was used in processing data from the "Wangara" experiment. The accuracy of different analytical expressions proposed for approximation of the temperature profile is compared. The diurnal variations and phase shifts of variations of the stratification parameters characterizing thermal stability separately in the upper and lower parts of the boundary layer are determined.

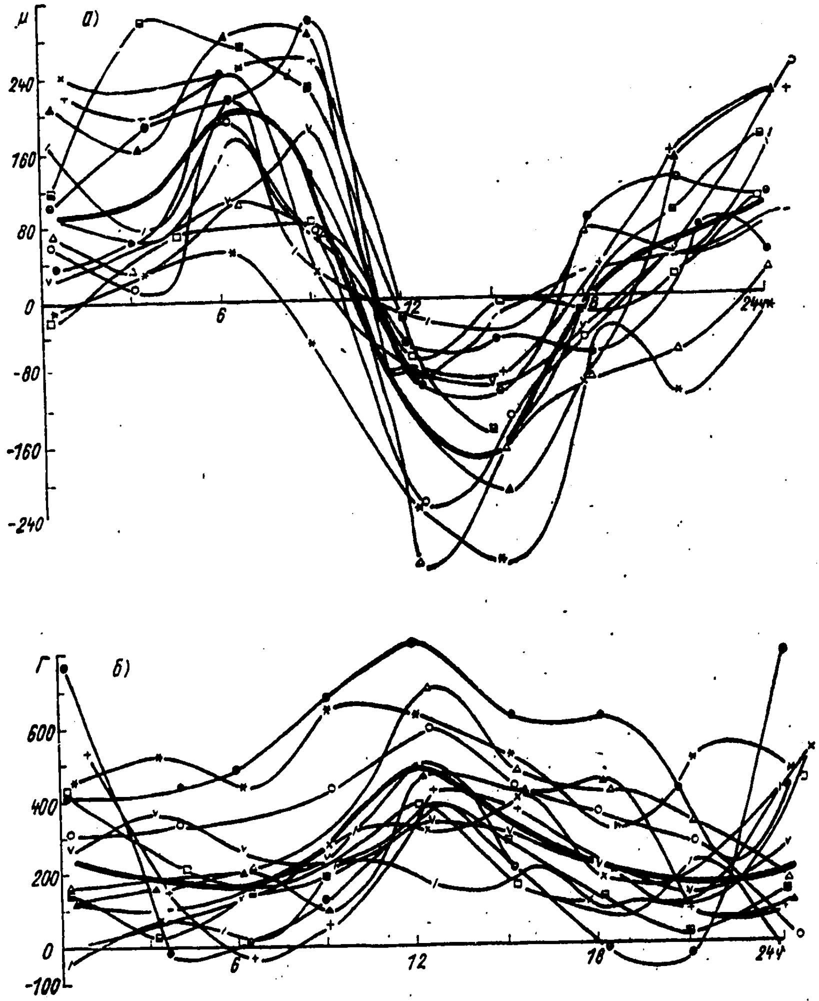
A considerable percentage of the investigations of the planetary boundary layer of the atmosphere is devoted to determination of the influence of the thermal structure of the boundary layer on its dynamics. This influence is attributable to the presence of a term in the equation for the balance of turbulent energy describing thermal generation, proportional to the vertical turbulent heat flux. Different analytical expressions have been proposed for stipulation of thermal stratification in semiempirical boundary layer models. In early studies of the theory of the atmospheric boundary layer use was made of the assumption of a constancy of the vertical turbulent heat flux P with altitude:

$$P_0 = -\rho c_p \alpha_T K d\theta/dz = \text{const}, \tag{1}$$

where ρ is density, c_p is specific heat capacity at a constant pressure, α_T is the Prandtl turbulent number, K is the coefficient of turbulent viscosity, $d\theta/dz$ is the vertical component of the potential temperature gradient. Within the framework of this model the thermal stratification is stipulated by the single parameter P_0 or by the dimensionless Kazanskiy-Monin parameter [2]

$$\mu = \varkappa u_* / \lambda L = -\varkappa^2 \beta P_0 / u_*^2 \lambda \rho c_p.$$

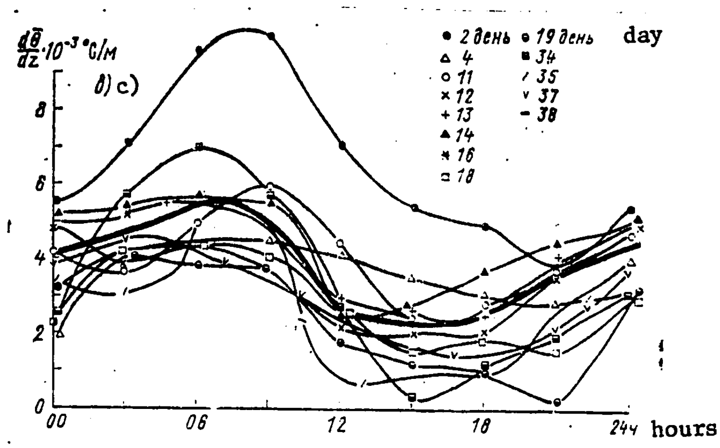
Here $\varkappa$ is the Karman constant, u_* is friction velocity, λ is the Coriolis parameter, L is the Monin-Obukhov scale, β is the buoyancy constant.



Diurnal variation of parameters of thermal stratification of the planetary boundary layer according to data from "Wangara" experiment. Thin lines -- individual days, thick line -- average for several days. (Figure continued on next page.)

In the lower part of the atmospheric boundary layer there is actually a constancy of the vertical turbulent heat flux with altitude. However, its upper part is usually characterized by a constancy of the potential temperature gradient

$$d\Theta/dz\,|_{z\to\infty} = \text{const.} \tag{2}$$



Taking into account both characteristics of the boundary layer, Laykhtman [3, 4] proposed an approximation synthesizing (1) and (2). In somewhat modified form [5, 6] this approximation can be written

$$\frac{d\Theta}{dz} = \frac{u_* \lambda}{\alpha_T \beta \chi^3 (z + z_0)} \left(\mu + \Gamma \frac{z \lambda}{\chi u_*} \right), \tag{3}$$

where $\Gamma = \alpha_T \beta \chi^4 (\gamma_a - \gamma_{ub}) / \lambda^2$ is the Laykhtman parameter, characterizing thermal conditions when $z \to \infty$, γ_a is the absolute value of the adiabatic temperature gradient ($\approx 9.8\cdot 10^{-3}$K/m), γ_{ub} is the temperature gradient at the upper boundary of the atmospheric boundary layer, taken with the opposite sign, z_0 is the roughness parameter. Height is reckoned from the ground surface.

The authors of [5, 7, 8] noted the important role of stratification of the upper part of the atmospheric boundary layer, especially in the low latitudes. In this connection it is necessary to investigate the variability of the Γ parameter and its typical values on the basis of experimental data.

The purpose of this study is an investigation of diurnal variations of the parameters μ and Γ, and also checking the comparative accuracy of approximations of temperature profiles by different formulas. For this purpose we processed temperature profiles measured in the "Wangara" experiment [9]. From the 44 days of the experiment we selected 13 for which frontal surfaces were absent at the observation point and the isobars had a minimum curvature. The friction velocity values were obtained by a graphic method from the profile of mean wind velocity, measured at heights up to 8 m. The temperature profiles were processed by the least squares method using formula (3). The entire region of measurements (0-2000 m) was used for obtaining the μ and Γ values. This method is characterized by a greater accuracy in comparison with the usually practiced evaluation of the μ and Γ parameters separately on the basis of several points in the lower and upper parts of the boundary layer. In addition, there is no need for solving difficult problems in determining the values of the α_T parameter and determining the height of the boundary layer, since the Γ parameter was determined by us directly from (3), and not on the basis of a determination of this parameter. Since in the numerical modeling of dynamics of the boundary layer of the atmosphere use was made not of the stratification parameters themselves, but the related turbulent heat flow, the indicated method for determining them corresponds to the formulated problem.

Figure 1 a,b shows the diurnal variation of the μ and Γ parameters, determined by approximation of the temperature profiles by formula (3). At 0600-0900 hours at the earth's surface there is a stability which at 1200-1500 hours is replaced by instability. The mean variation of μ averaged for 13 days for the "Wangara" experiment falls in the range -120-200. In contrast to the parameter μ, which has now been studied quite well, little is known about the Γ parameter. Figure 1b shows the diurnal value of this parameter on the basis of data from the "Wangara" experiment. As in the case of μ, the Γ value varies in the course of 24 hours in a quite broad range. The Γ parameter, averaged for 13 days, varies in the course of 24 hours from 150 to 500 with a mean daily value 265. A comparison of Figures 1a and 1b shows that the diurnal variations of μ and Γ have a phase shift. The development of instability at the earth's surface during 0600-1200 hours is accompanied by an increase in stability in the upper part of the boundary layer. The Γ maximum is observed at 1200 hours. The change in Γ lags by approximately 6 hours in comparison with the change in μ.

Figure 1c represents the diurnal variation of the mean gradient of potential temperature, determined by averaging of the gradient in the entire interval of change in the height of observation in the "Wangara" experiment (from the ground surface to a height of 2000 m). The mean daily potential temperature gradient is $-6.1° \cdot 10^{-3}$ K/m. It is interesting that the diurnal variations $d\theta/dz$ are in phase with the μ variations. Variations of the mean temperature gradient have a lesser relative amplitude than the μ and Γ variations. This is natural because, as can be seen from a comparison of Fig. 1a and Fig. 1b, stratification in the lower and upper parts of the boundary layer varies with a considerable phase shift, in part compensating one another.

The phase shift of the diurnal variations of μ, Γ and $d\theta/dz$ has a simple explanation. The diurnal temperature variations at the upper boundary of the boundary layer have a considerably lesser amplitude than at the ground surface. Accordingly, the diurnal temperature variations of the earth's surface are in phase with variations of the potential temperature gradient $d\theta/dz$, averaged for the entire boundary layer. The phase shift of the variations in Γ relative to the phase of the μ variations is related to the propagation of the diurnal temperature wave. In actuality, according to measurements by Devyatova [1], the velocity of propagation of diurnal temperature variations in the atmospheric boundary layer is about 100 m/hour. According to data from the "Wangara" experiment [9, 10, 11], the velocity of movement of the phase of temperature variations at 0600-1000 hours is about 30-50 m/hour, whereas at 1200-1800 hours it is about 100 m/hour. Accordingly, variations of the temperature difference in the layer 0-600 m during the daytime hours will outrun the change in stratification in the layer 600-1200 m by approximately 6 hours, which also explains the origin of the phase shift in the diurnal variations of μ and Γ.

Formula (3) fairly well reflects the characteristic features of thermal stratification, especially during instability. The mean square error in approximation of the temperature profile by formula (3) was equal (averaged for all observation periods) to 0.55 K.

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MODEL OF FORMATION OF A STATIONARY ZONE OF CONTAMINATION IN WATER BODIES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 105-107

[Article by A. V. Karaushev, professor, and L. N. Meyerovich, State Hydrological Institute, submitted for publication 12 Jun 80]

[Text]

Abstract: The article gives a model of formation of a stationary zone of contamination in water bodies due to the discharge of a nonconservative dissolved contaminating substance. The model is based on a numerical solution of the turbulent diffusion equation in cylindrical coordinates. The results of the computations are presented in the form of a nomogram which can be used in solving practical problems. Formulas are also proposed for estimating the total mass of contaminating substance in a water body and the time required for stabilizing the region of propagation of the contaminating substances.

Methods making it possible to estimate or predict the indices of water quality in water bodies which are the receivers of waste waters are of great importance for evaluating the sanitary state of water bodies, in planning and projecting water conservation measures. The proposed model of the propagation of matter over the surface of a water body makes it possible to compute the parameters of the contamination zone forming in a water body due to the discharge of a nonconservative dissolved contaminating substance in it. The model is based on the turbulent diffusion equation in cylindrical coordinates proposed in [2]. The equation was derived on the basis of the balance of mass of contaminating substance under the following assumptions: the currents in the water body in the region of discharge of the waste waters are small in magnitude and variable in direction; the depth of the water body in the discharge region is small; there are no tributaries in the discharge region; the waste water source is situated at the center of coordinates; the decay (or formation) of a nonconservative substance is described by the first-degree equation

[H = non(conservancy)]

$$\frac{ds}{dt} = k_H s, \tag{1}$$

where s is the concentration, t is time, k_{non} is the nonconservancy coefficient (the value is negative for decaying matter).

The turbulent diffusion equation with the indicated assumptions has the form

[H = non]

$$\frac{\partial s}{\partial t} = D\frac{\partial^2 s}{\partial r^2} + \frac{\beta}{r}\frac{\partial s}{\partial r} + k_H s. \tag{2}$$

Here D is the turbulent diffusion coefficient, r is a coordinate of the cylindrical system (distance from the discharge site), β is a parameter determined by the equation

$$\beta = D - \frac{Q_{CT}}{\varphi H}, \tag{3}$$

[CT = waste]

where Q_{waste} is the discharge of waste water, H is the mean depth of the water body in the discharge region, φ is the angle of the sector of propagation of waste waters (in the case of discharge along a linear shore $\varphi = \pi$; in the case of discharge distant from the shore $\varphi = 2\pi$).

The concentration field at any moment in time will be symmetric relative to the origin of coordinates; the concentration isolines are concentric circles with the center at the origin of coordinates. If it is assumed that the background concentration of a particular substance in a water body is equal to zero and the dimensions of the forming region of propagation of contaminating substance are less than the dimensions of the water body, the boundary and initial conditions can be written in the form

$$s|_{r=0} = s_{CT}, \quad s|_{r\to\infty} = 0, \tag{4}$$

[CT = waste]

$$s|_{t=0} = 0, \tag{5}$$

where s_{waste} is the concentration of this substance in the waste waters. In the case of a nonzero background concentration the solution of the formulated problem will give the excess of the concentration above the background.

Equation (2) with the boundary conditions (4), (5) is solved numerically. The corresponding difference scheme has the form

$$s_{i-1}^{j+1}\mu_i - s_i^{j+1}\nu_i + s_{i+1}^{j+1}\gamma_{il} = -F_i^j \tag{6}$$

$$(i = 1, 2, \ldots, N-1;\ j = 0, 1, 2, \ldots),$$

$$s_0^j = s_{CT}, \quad s_N^j = 0\ (j = 0, 1, 2, \ldots), \tag{7}$$

$$s_i^0 = 0\ (i = 1, 2, \ldots, N). \tag{8}$$

The superscript denotes the number of the time interval, the subscript denotes the interval along the radius:

$$\left.\begin{aligned} \gamma_{il} &= \sigma D \tau \\ \mu_i &= \sigma D \tau\left(1 - \frac{a}{i}\right) \\ \nu_i &= \sigma D \tau\left(2 - \frac{a}{i}\right) + 1 - \sigma b D \Delta t \end{aligned}\right\}, \tag{9}$$

$$F_l^j = s_{l-1}^j (1-\sigma) D \tau \left(1 - \frac{a}{l}\right) - s_l^j \times$$
$$\times \left\{ (1-\sigma) \left[D \tau \left(2 - \frac{a}{l}\right) - bD \Delta t \right] - 1 \right\} + s_{l+1}^j (1-\sigma) D \tau, \tag{10}$$

$$a = \frac{\beta}{D}, \quad b = \frac{k_H}{D}, \tag{11}$$

$$\tau = \frac{\Delta t}{\Delta r^2}, \tag{12}$$

Δt is the time interval, Δr is the interval along the radius, σ is a parameter of the scheme (in the computations it was assumed that $\sigma = 0.5$).

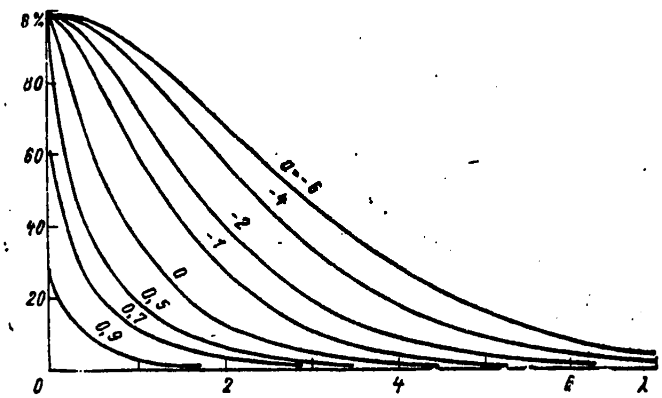

Fig. 1. $s(a, \lambda)$ nomogram.

The problem (6)-(8) is solved by the step method

$$s_l^{j+1} = s_{l+1}^{j+1} \chi_{l+1}^{j+1} + \psi_{l+1}^{j+1} \tag{13}$$
$$(l = 0, 1, \ldots, N-1;$$
$$j = 0, 1, \ldots).$$

These "step" coefficients are computed using the recurrent formulas

$$\chi_{l+1}^j = \frac{\gamma_l}{\nu_l - \mu_l \chi_l^j},$$
$$\psi_{l+1}^j = \frac{\mu_l \psi_l^j + F_l^j}{\nu_l - \mu_l \chi_l^j} \tag{14}$$
$$(l = 1, 2, \ldots, N-1),$$
$$\chi_1^j = 0, \quad \psi_1^j = s_{\text{ст}}. \tag{15}$$

The proposed computation scheme makes it possible to obtain the concentration field in the water body at stipulated time intervals after the onset of waste water discharge.

Numerical experiments show that in the case of discharge into a water body from a source with a constant intensity of the decaying contaminating substance ($k_{non} < 0$) its distribution zone is stabilized after a certain number of computation time intervals. This corresponds to theoretical concepts according to which stabilization sets in at the time when the mass of matter decaying in a unit time in the water is in equilibrium with the receipt of the particular substance with the waste water in the water body. In such a situation the diffusion process is described by the equation

$$D\frac{\partial^2 s}{\partial r^2} + \frac{\beta}{r}\frac{\partial s}{\partial r} + k_n s = 0 \tag{16}$$

with the boundary conditions (4). The solution of the corresponding difference scheme was also found by the "step" method and the results indicated a good agreement with the results of solution of the already described scheme with stabilization of the concentration field. As indicated by the computations, the concentration s with fixed values of the a parameter is a function of the dimensionless parameter $\lambda = r\sqrt{-b}$. This made it possible to construct a nomogram (see Fig. 1) by means of which it is possible to obtain the values of the different characteristics of the region of propagation of the nonconservative contaminating substance in the water body directly. The nomogram makes it possible to determine the concentration of the particular substance at an arbitrary distance from the site of discharge of the waste water or to find the distance to an isoline of an arbitrary concentration. Thus, it is possible to estimate the dimensions of the zone of contamination (at its boundary s = MAC), compute the mean concentration in this zone, etc. The total mass of contaminating substance in the entire forming stationary region of its propagation M_s is determined by the equation

$$M_s = -Q_{waste}s_{waste}/k_{non}. \tag{17}$$

The stabilization of the region of propagation of contaminating substance forming in the water body occurs over the course of a prolonged time (theoretically equal to infinity). On a practical basis it is possible to estimate the time after which the relative deviation of the total mass of contaminating substance in the water body from M_s will not exceed the stipulated ε value. This time t_s is determined by the equation

$$t_s = \ln \varepsilon / k_{non}. \tag{18}$$

The computation method, nomogram and formulas (17)-(18) proposed in the article can be used for solving practical problems.

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STANDARD INSTRUMENTS FOR MEASURING AIR HUMIDITY AT NEGATIVE TEMPERATURES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 108-113

[Article by V. A. Usol'tsev, Scientific Research Institute of Instrument Making, manuscript received 17 Jun 80]

[Text]

Abstract: A study was made of the desirability of developing standard instruments for measuring air humidity with reproduction of the composition and properties of water vapor over plane water and ice surfaces, as well as the composition and properties of moist air. In particular, it is shown that in the reproduction of the triple point of purified natural water the error in reproducing the pressure of saturated water vapor is more than 1,000 times less than the error in modern working instruments for measuring air humidity. The principle for the reproduction of standard samples of moist air set forth in this article served as a basis for developing standard instruments at the Scientific Research Institute of Instrument Making. This article gives brief descriptions of static and two-temperature moist air generators. These instruments, constituting nonstandardized measurement instruments, have undergone state certification and are used as sampling devices in carrying out scientific research and experimental design work and also in state acceptance tests for new working instruments for measuring humidity.

Meteorological hygrometers are intended for measuring humidity in a broad temperature range, but their metrological support in standard production does not meet modern requirements. Hygrometers are checked only at a temperature of about 20°C in a PO-34M hygrostat, which is obviously inadequate.

The problems involved in the metrological support of scientific research and experimental design work in the development and modernization of hygrometers are still more acute. At the present time the availability of sampling techniques and instruments for making the required measurements is a decisive factor determining the development of hygrometry.

Now we will examine the problems relating to the creation of standard instruments intended for the testing and checking of hygrometers in the range of negative temperatures where metrological support encounters substantial difficulties.

The system for testing hygrometers is still in the development stage at the State Committee on Hydrometeorology. There is a proposal for such a system [8], but its approval is being held up due to the lack of high-accuracy standard instruments.

The development of standard instruments can be accomplished both in the direction of creation of standard hygrometers or with orientation on the use of measurement instruments with the reproduction of a vapor-air mixture with a normalized humidity.

Repeated attempts to develop standard instruments for the purpose of increasing the accuracy of hygrometers have not yielded positive results. With a decrease in temperature the water vapor content in the air is sharply reduced. For this same reason it is impossible to make effective use of standard instruments for measurement of the graviametric apparatus type employed at the United States Bureau of Standards [14] because with its use the measurement errors increase substantially at temperatures below 0°C.

It is more realistic to develop measurement instruments with reproduction of a vapor-air mixture with normalized humidity. The realization of this direction requires having standard moist-air samples reproducible with a sufficiently high accuracy in the entire measurement range.

There is a draft of recommendations on the working out of a practical relative humidity scale in the temperature range from 0 to 100°C at atmospheric pressure on the basis of use of aqueous saturated solutions of inorganic salts, that is, the use of these solutions for the reproduction of standard humidity samples.

On an international scale considerable work has been done on measuring equilibrium humidity over saturated solutions of salts, but the data obtained by different authors have discrepancies and there is no international agreement on this problem. But the most important consideration here is that aqueous saturated solutions of inorganic salts are difficult to use in the reproduction of standard air relative humidity samples at low negative temperatures. With a temperature decrease the equilibrium relative humidity increases and at a eutectic temperature and below it it is possible to obtain standard samples close to 100% relative to ice. In this case the use of "contaminated" ice instead of natural ice loses sense.

The properties of water have been investigated more fully. There is international WMO agreement determining the formulas and constants for computing water vapor pressure over the plane surfaces of water and ice as a function of temperature, which in our opinion should be adopted as standard reference data. This is all the more necessary because at the present time virtually all measurements of the humidity of atmospheric air in one way or another involve the use of data on the pressure of water vapor over water (ice) as a function of temperature.

On the basis of these standard reference data it is possible to develop a testing system and select standard samples not having the shortcomings enumerated above.

In our opinion, standard humidity samples should reproduce the composition and properties of water vapor, especially its pressure, over the plane surface of water or ice with known temperature values of the single-component equilibrium two-phase system "vapor + water" or "vapor + ice." As the means for reproducing these samples it is desirable to use apparatus in which there is sufficiently rigorous satisfaction of the conditions for reproduction of the mentioned system; the vapor pressure in such apparatus during the reproduction of standard samples is computed using the formulas recommended by the WMO [10], in other words, is determined using the tables in [6] for system temperature. In confirmation of the desirability of the proposed method for the reproduction of humidity we can mention the following.

The most important processes responsible for the presence of water vapor in the air in one way or another are related to water vapor evaporation and condensation phenomena. In turn, the transpiring of evaporation and condensation processes is dependent on the saturation deficit E - e, where the pressure E of saturated vapor is determined by the characteristic equation for the equilibrium of the two-phase system "vapor + water" or "vapor + ice" $E = f(T)$ [1], whereas vapor pressure e is determined by the dew point T_τ, that is, $e = f(T_\tau)$. Both parameters E and e are reproduced with a sufficiently high accuracy under laboratory conditions. It therefore follows that the calibration of the hygrometers should be accomplished on the basis of the characteristic equation $E = f(T)$.

An equally important conclusion is that the accuracy in reproducing a vapor-air mixture under laboratory conditions with an equilibrium of the phases can be considerably higher than the accuracy in measuring humidity by any measurement instrument, especially in the range of negative temperatures, where water vapor pressure is insignificant.

Tables of water vapor saturation pressures over water and ice are usually used in determining the partial pressure of water vapor in moist air on the basis of the dew (frost) point. However, in developing a testing system it must be taken into account that the partial pressure of water vapor in a moist gas over water (ice), the same as over saturated solutions of salts, is dependent on the composition of the gas, temperature and its pressure. The formulas recommended by the WMO [10] for computing the pressure of saturated vapor over water and ice are applicable only to a single-component equilibrium two-phase system, whereas real systems are multicomponent.

At pressures and temperatures which are observed in the earth's atmosphere the partial pressure of saturated water vapor in moist air differs from the pressure E of saturated vapor of a single-component system with the same values by not more than 0.5% [10].

From the Clausius-Clayperon equation

$$\Delta\tau = \frac{(1 + a\tau)^2}{a} \frac{\Delta E}{E},$$

where $a = 0.072°C^{-1}$ in the system "vapor + water" and $a = 0.082°C^{-1}$ in the "vapor-ice" system, $\alpha = 1/273°C^{-1}$, τ is the dew point in °C. Assuming that $\Delta E/E = 0.005$ we find that the maximum error when using the WMO formulas for a system with moist atmospheric air will not exceed 0.08°C of the dew point when $\tau = 30°C$ and 0.06°C of the dew point when $\tau = 0°C$.

It therefore follows that even at the present time, when apparatus has not yet been developed for reproducing the composition and properties of the vapor-air mixture with an adequately high accuracy, this error, systematic in character, in calibration work cannot be taken into account or evaluated on the basis of theoretical computations. In the future it will be possible to develop apparatus for reproducing the composition and properties of water vapor in a single-component equilibrium two-phase system and then by means of a sensitive and adequately stable instrument there will be a possibility for experimental determination and allowance for the mentioned systematic error.

In order to reproduce standard humidity samples it is necessary to use water with a definite isotopic composition, for example, purified natural water. One of the indices of the suitability of water can be the reproducibility of the temperature of its triple point of the melting point of ice. The discrepancy between the temperatures of the triple points of ocean and ordinary continental surface water is about $4 \cdot 10^{-5}$ °C. The maximum discrepancy in the temperatures of the triple points of purified natural water is $2.5 \cdot 10^{-4}$ °C [7]. It is usually assumed that the error in reproducing the triple point temperature of purified natural water is $1 \cdot 10^{-4}$ °C [9]. And this means that at a temperature of 0.01°C the dew point can be reproduced with the same error.

At a temperature of 0.01°C the pressure of saturated water vapor is 611.14 Pa [6] but with a change in the dew point by $1 \cdot 10^{-4}$ °C the water vapor pressure changes by 0.0044 Pa. Accordingly, with the use of purified natural water at the triple point it is possible to reproduce water vapor pressure with an error of 0.0044 Pa, which is $7.2 \cdot 10^{-4}$% of the pressure of saturated water vapor at a temperature of 0.01°C. This error is more than 1000 times less than the error of modern working instruments for measuring air humidity, evidence of the possibility of using purified natural water for the reproduction of standard humidity samples in standard high-accuracy instruments.

With respect to the melting point of ice, it is a very close approximation to the temperature determined as a temperature 0.01°C below the triple point of water [7]. Accordingly, in standard measurement instruments having an error greater than 0.05°C of the dew point and operating on the principle of dynamic equilibrium of moist air with a water vapor condensate the considered error can be neglected.

The described principle for the reproduction of moist air samples was used in developing apparatus necessary in the metrological support of scientific research and experimental design work for the development of hygrometers (sensors) at the Scientific Research Institute of Instrument Making. Nonstandard measurement instruments have been developed, fabricated and certified: a static generator of moist air and a two-temperature moist-air generator [12, 13]. With respect to metrological characteristics the mentioned generators correspond to "second-order" measurement instruments in conformity to the Socialist Economic Bloc recommendations RS Z118-71.

Static Moist Air Generator

The static generator makes use of the dynamic equilibrium of water vapor over the surface of water (ice), in a closed thermostated space containing water (ice) making it possible to create constant air humidities near 100%.

Apparatus operating on this principle is known. As the dew point use is made of the air temperature [4]. However, in such apparatus the total saturation condition cannot be rigorously satisfied since the mixing of air is necessary and this leads to an increase in its temperature. In most cases the tested instrument also releases heat. When there is a heat flow from the air to the chamber walls the internal surfaces of the chamber walls, the condensate on them and the chamber air will have different temperatures. The air temperature will be higher than the condensate temperature, the relative humidity will be less than 100%, but the dew (frost) point will not be equal to the air temperature. A substantial error can arise as a result of this.

In order to eliminate this error in the static generator [11] on the coldest part of the wall within the working chamber there was installation of a thin polished metal plate, one part of which (a small mirror) is curved at a small angle and is present in the air, whereas the other part has a thermal contact with the wall. A metal disk in the form of a radiator is attached on the outer surface of this part of the wall in order to reduce its temperature. This unit in essence is a dew-point hygrometer.

The working chamber, that is, the chamber in which the instruments to be tested are placed, in the course of thermostating is moistened and a condensate with a fixed boundary is created on the surface of the small mirror. There is a clear temperature gradient on the small mirror along the plate (the base of the small mirror is colder than its end by tenths of a degree) and therefore it is possible to create such conditions that the temperature of the base of the small mirror will be below the dew (frost) point and the temperature of the end will be above the dew (frost) point.

With the presence of such a fixed boundary the temperature of the condensate on it is measured. Such a device and measurement method make possible the complete exclusion of the error arising due to noncorrespondence of air temperature to the temperature of the condensate. The presence of a fixed condensate boundary is evidence of its dynamic equilibrium on the boundary with the water vapor in the working chamber. With this condition the temperature of the condensate at its boundary will be equal to the dew (frost) point.

With temperatures of the small mirror below 0°C an error can arise due to an incorrect determination of the phase state of the condensate. In order to exclude the possibility of appearance of this error measurements at negative temperatures are made only after a condensate is formed on the small mirror in the form of small ice crystals of an acicular structure which with a change in the angle of illumination is readily distinguished from its external appearance from a condensate consisting of droplets of supercooled water. A special method has been developed for obtaining a condensate in the form of small ice crystals with an acicular structure [12].

In the development of the generator use was made of an evaluation of the errors arising as a result of the influence of curvature of the elements of the condensate on the small mirror and contamination of the surface of the small mirror. These particular errors are insignificant. It was demonstrated in [12] that with adherence to the generator operating regime they can be neglected and that the error in measuring humidity in the working chamber is determined by the error in determining temperature t_τ of the condensate at its boundary on the small mirror.

Temperature t_τ is determined on the basis of the results of measuring air temperature t in the working chamber, the temperature difference Δ_1 between the thermometer bulb and the center of the small mirror plate and the temperature difference Δ_2 on the small mirror plate between two points situated symmetrically relative to the center of the small mirror plate in the limits of the zone of possible positioning of the boundary of the condensate.

In measuring air temperature use was made of an M-34 aspiration psychrometer in which the electric motor fabricated by the Safonovskiy Plant has been replaced by a DPM-25-N1-02 electric motor which creates a heat flow not greater than 2 W, that is, 5.5 times less than the heat flow of the electric motor in the M-34 psychrometer. The temperature difference was measured using differential thermocouples and an instrument with a measurement range from 0 to 50μV.

Temperature is computed using the formula

$$t_\tau = t - \Delta_1 + \frac{\Delta_2}{L} n,$$

where t is the temperature according to the thermometer of the aspiration psychrometer, read using a telescope mounted outside the climatic chamber: Δ_1 is the temperature difference according to the main thermocouple; Δ_2 is the temperature difference according to the auxiliary thermocouple; L is the distance on the small mirror between the junctions of the auxiliary thermocouple; n is the distance from the junction of the main thermocouple on the small mirror to the boundary of the condensate.

In order to shorten the time of moistening of the working chamber it contains a moistener, a fabric wetted with water, heated during moistening of the chamber by an electric current. The heating of the moistener makes it possible in a short time interval to introduce into the chamber the necessary quantity of water vapor, part of which is sorbed on the walls of the apparatus and another part of which is expended on the formation of a condensate on the chamber walls and on the small mirror.

The thermostating of the working chamber is accomplished in the 3001 GDR climatic chamber. The climatic chamber is supplied with a special heat regulator with a contact mercury thermometer [5] and makes possible the automatic maintenance of constant stipulated air temperature values in the working chamber of the generator with deviations not greater than ±0.03°C.

The static moist-air generator ensures the reproduction of the dew point at an atmospheric pressure from -30 to -6° and from 0 to +5°C. The absolute error in reproducing the dew point is ±0.2°C; the absolute error in such measurements of air temperature in the working chamber is ±0.1°C; the volume of the working chamber is 150 liters.

On the basis of the described generator it is possible to develop a standard measurement instrument with an absolute error in reproducing the dew point of about 0.05°C. A substantial decrease in the error can be attained by use of a standard resistance thermometer for measuring air temperature and more careful calibration of the thermocouples.

Two-Temperature Moist Air Generator

The two-temperature generator consists of three principal units: a "Polyus-1" standard generator with a dessicator and moistener, working chamber, thermostated in a 3001 climatic chamber, and a measuring support.

Air humidity in the working chamber is determined from dew point values set by the "Polyus-1" generator and air temperature in the working chamber.

The "Polyus-1" standard generator [2, 3] is intended for the checking of hygrometers and also for other purposes requiring the creation and maintenance fixed dew (frost) point values in the flow of air or nitrogen. The working range of fixed dew point values at the generator output is from -60 to (t-5)°C, where t is air temperature in the enclosed space. The absolute error of this generator does not exceed ±0.2°C of the dew point.

The air from the standard "Polyus-1" generator is fed through a metal pipeline into the working chamber through a ventilator-mixer. The air emerges from an annular collector of the ventilator-mixer in thin jets which are broken up by an air flow from a bladed disk. This makes it possible, without using a heat exchanger with an intricate surface, to obtain an adequately uniform temperature field with an air temperature in the pipeline differing from the air temperature in the working chamber.

The bladed disk of the ventilator-mixer also accomplishes mixing of the air in the working chamber; the rotating disk with the blades sucks the air into its central part and then expels it. Then the air flows out primarily along the walls of the working chamber and thus there is adequately effective mixing, precluding the appearance in the working chamber of zones of stagnant air. The electric motor of the ventilator-mixer is situated outside the working chamber.

The air is fed from the working chamber into a valve intended for converting the open system communicating with the external air into a closed system. Then there is a discharge stimulator (a pump with an airtight housing) and a preparation valve. The preparation valve makes it possible, using a moistener and dessicator, on the basis of readings of the hygrometer-indicator placed at the input of the "Polyus-1" generator, to bring the air humidity to the dew point, 5-20°C higher than the normalized value set by the generator. With a dew point below -15°C the work is accomplished with a closed system without the monitoring of humidity at the input to the "Polyus-1" generator and without drying or moistening of the air entering into the "Polyus-1" generator.

The air discharge through the "Polyus-1" generator can be set from 10 to 40 liters/m. At the same time air pressure is measured in the "Polyus-1" generator. The measurement results are used with the introduction of corrections if the pressure difference in the "Polyus-1" generator and in the working volume exceeds 15 gPa. The normalized air dew point value at the output of the "Polyus-1" generator is determined using the standard resistance thermometer of the "Polyus-1" generator and an R363-3 d-c potentiometer.

With the operation of the generator an excess pressure of not less than 50 Pa is created in its working chamber, which impedes the penetration of external air into the working chamber in the case of formation of a microflow in it. The excess pressure is determined using the manometer on the working chamber.

In the working chamber there is a second hygrometer-indicator which is an auxiliary means for monitoring the stability of humidity and also three differential thermocouples for increasing the accuracy in determining temperature in the zones of placement of the sensing elements during the testing of relative humidity hygrometers. The measurement of air temperature in the working chamber is accomplished using an aspiration psychrometer in which a DPM-25-N1-02 electric motor has been installed, and a telescope. An N391-1 outfit is used in measuring the temf of the differential thermocouples and the resistance of the thermoelement of the hygrometer-indicator.

The two-temperature generator ensures the reproduction of the following in its working chamber at temperatures from -25 to (t - 4)°C:
-- relative humidity from 10 to 90% at positive temperatures and from 20 to 90% relative humidity relative to ice at negative temperatures;
-- dew point from -45 to (t - 5)°C with the above-mentioned relative humidity values, where t is the temperature in the enclosed space.

The absolute error in determining relative humidity is not more than ±2% and for the dew point -- not more than ±0.2°C. The useful volume of the working chamber is 12.5 liters.

In the measurements there must be assurance that there is no condensate in the working chamber. Accordingly, it is inadmissible that there be even a brief increase in relative humidity in the chamber above 90%. For this same reason in the preparation of the generator for operation the working chamber is dried out. For testing hygrometers at a humidity exceeding 90% provision is made for use of a static generator.

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Summary

The results provide a basis for constructing standard instruments and for developing a testing system at the level of modern requirements for instruments for measuring air humidity (hygrometers).

This is a further step on the path to improving the metrological support of measurements of the humidity of atmospheric air, the development of hygrometers and the improvement of measurement methods.

The standard instruments described in the article, existing at the Scientific Research Institute of Instrument Making, made it possible to complete the development of new air humidity measurement instruments and carry out their state acceptance tests.

The experience accumulated in work with these standard instruments confirmed their high metrological characteristics, but such instruments are relatively complex, suitable for use only at base organizations and at factories making hygrometers, that is, in places where it is necessary to check the metrological characteristics of instruments in the entire working range of negative temperatures and carry out research work.

In order to check hygrometers during their operation, especially for outfitting the testing bureaus of the administrations of the Hydrometeorological Service of the State Committee on Hydrometeorology and Environmental Monitoring, there is a need for less complex standard instruments, for whose development the results provide practical possibilities.

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AUTOMATIC GENERATION OF PROGRAMS FOR THE DECODING OF METEOROLOGICAL SUMMARIES

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[Article by S. L. Rudenko and A. Yu. Solomakhov, USSR Hydrometeorological Scientific Research Center, manuscript received 8 Jul 80]

[Text]

Abstract: A new method is proposed for preparing programs for the decoding of meteorological telegrams. A specialized language is described which is applied using a macroprocessor designed for this purpose.

Meteorological data are received at the data processing centers of the World Weather Service in the form of coded telegrams. The processing of this information on digital computers requires programs for the decoding of telegrams. At the present time there are a great number of different coded forms (about 100) [3] and their number can increase. In addition, the coded forms can change. At the same time it is known that the writing of decoding programs is a time-consuming process. Accordingly, it is extremely desirable to have means for facilitating this work.

The process of decoding of meteorological telegrams can be broken down into two stages. The first is in the telegram, which gives the value (it can be coded) of the necessary meteorological element, and the second stage is its decoding, which frequently involves very simple operations.

However, knowing the structure of the telegrams it is possible to create a specialized language for decoding. It is convenient to regard it as a "macrolanguage" [1, 2] because this also makes it possible to use all the means of the basic language.

We created such a language and a corresponding specialized macroprocessor for FORTRAN, which is also an application of FORTRAN programming, which facilitates the transfer from computer to computer not only of the macroprocessor unit, but also the programs generated by it.

The macroprocessor scans the text which is fed to it, which consists of FORTRAN operators and decoding operators. In the processing of the latter there is generation of some sequence of FORTRAN operators incorporated in the initial text. The FORTRAN operators of the initial text are not processed by the macroprocessor. The resulting program can be dispatched for compilation, editing and execution.

(We will use the term "decoding language operator" instead of the more precise "macrocommand" because this better conveys the purpose of the proposed language.)

The lines of the blank (punch card) in which the decoding language proposals are placed are set off by the symbols C+ in the first two positions. A marker can be placed in the next three positions. The language operators, placed on one line, are separated by a semicolon.

The language user must adhere to some special conventions when writing programs. For example, the text of the telegram to be processed must be in the data complex TEXT. The marker M must indicate that byte of this data complex from which it is proposed that the processing begin. The use of the identifiers NE, I00000, M00000 is disallowed, other than the marker 77777. The use of this marker, as well as the TOBE and LENGTH blocks will be described below.

In the decoding language there are 6 operators, 2 of which are means for the control of printout:

I. Description operator.
II. Obligatory group operator.
III. Nonobligatory group operator.
IV. Checking operator.
V. Print-out suppression operator.
VI. Begin print-out operator.
Now we will describe these operators.

I. Description operator. This operator has the form C+DCP (in the first five positions). It must be present in each program containing the decoding language operators and be placed before the first applied Fortran operator and the decoding language operators, other than the print-out control operators. In processing the description operator the macroprocessor does not generate any applicable Fortran operators.

II. Obligatory group operator. The simplest form of this operator can be written in the following form:

<OGRO> :: = <FIGURE>
<DESCRIPTOR> <FORTBL>
<FIGURE> :: = 1|2|3|4|5|6|7|8|9
<DESCRIPTOR> :: = |N|D|T (1)

The syntax of <FORTBL> and the action of the operator are described separately for each of the three types of <DESCRIPTOR>.

a) Descriptor N

<FORTBL> :: = Fortran identifier.

As a result of application of the operator in the block with the designation <FORTBL> there will be a value of a number present in the text of a telegram, beginning from the place indicated by the marker M and with a length determined by <FIGURE>, in the form INTEGER or REAL (depending on how the block is described). If in the indicated positions there are no figures, the block contains the value -9999. In both cases the marker is increased by the value < FIGURE> .

[Note: The term "application" is as inexact as the term "operator." The program is "applied" in a computer language, obtained after compilation of the generated FORTRAN operators. However, as a simplification in the exposition we will henceforth say that the decoding language operators are "applied."]

Example 1.

Assume that the telegram marker and text are as follows:
....6753A8...
    ↑
    M = 13

Then after applying the command 1NJ in the block J there will be the number 7 and the marker becomes equal to 14.

If the operator 2NRAB is applied in this same situation, in RAB there will be 75, marker = 15.

After the operator 5NABC the value ABC = -9999., M = 18.

In all the examples we will assume that the types of blocks are implicitly determined.
b) Descriptor D
<FORTBL> : : = < LETTER> <FORTSP>
<LETTER> : : = one of the 26 letters of the Latin alphabet
<FORTSP> : : = Fortran identifier.

The action of an operator with this descriptor differs from the preceding case in that the transformed value serves as the actual parameter when consulting the subprogram with the designation <FORTSP> . The output parameter of this subprogram is the block <FORTBL> . The user must write the corresponding subprogram with two parameters: first -- input, second -- output.

Example 2.

Assume that the marker and the text of the telegram are the same as in example 1. Then, with application of the operator 3DISUBR there will be consultation of the subprogram SBUR with the value 753 as the input parameter. The result will be placed in the block ISUBR.
c) Descriptor T.
< FORTBL> : : = Fortran identifier.

In this case in a<FORTBL> block there will be an untransformed text with positions determined by the momentary position of the marker and <FIGURE> .

Frequently it will be convenient to place a value in an element of the data mass. In this case <FORTBL> : : = a FORTRAN variable with an index. For example, 3DISUB (1, MET). Then the necessary value is placed in an element of the data mass ISUB (1, MET). (In this case the name of the subprogram is SUB).

In those cases when the decoding of the element is quite simple (it is necessary to divide by 10, multiply by 3, 5, etc.) it is infeasible to consult the subprogram. Such decoding can be accomplished using the decoding language. For this purpose it is possible to use an obligatory group operator, described by the rules (1) with the supplementation:
<OGRO> : : = <FIGURE>
<DESCRIPTER><FORTBL> <SIGN>
<OPERAND>
<SIGN> : : = +|-|* |/
<OPERAND> ::Fortran identifier | number in the form INTEGER or REAL.

Example 3

The marker and text of the telegram are the same as in example 1.
a) 2NI10*33.

In block 110 there is the number 75*33.

b) 2DRDEC (10, J, L)+AGBC.

The element of the data mass RDEC (10, J, L) will be equal to the sum of the result of application of the subprogram DEC with the input parameter 75 and the number AGBC.

III. Nonobligatory group operator
<NOGRO> : : = (< KEY>) (<KEY> , <LIST>),
<LIST> : : = list of operators of the obligatory group, separated by commas,
<KEY> : : = any set of symbols not containing periods, parentheses, comma and semicolon
<MASS> .<BEGIN> . <LENGTH>
<MASS> : : = FORTRAN identifier
<LENGTH> : : = FORTRAN identifier | number in the form INTEGER
<BEGIN> : : =< LENGTH> .
With application of this operator the following occurs.

The symbols in the telegram, beginning with the position indicated by the marker are checked for coincidence with the key. If < KEY> is a textual literal not containing periods, it is assumed that the key is explicitly incorportated in < KEY> . However, if the key has the form <MASS>. <BEGIN>. <LENGTH>, this means that the key is present in the data mass designated <MASS> (a LOGICAL*1 description is assumed), beginning with < BEGIN> and with the length of <LENGTH> bytes.

If a key is not found in the telegram TOBE (INTEGER) is assigned a zero value and the marker M does not change. Otherwise TOBE = 1, M advances by the length of the key and application of operators from < LIST> begins, if such exist.

Example 4

Assume that the data mass TEXT and the marker M have the following form:
....99741....
M = ↑6

and the data mass S is stipulated by the FORTRAN operator LOGICAL*1/1H1, 1H9, 1H9, 1H8... /. We will examine the action of the following operators in the described situation:

a) given (99, 3NJ), we obtain
TOBE = 1, M = 11, J = 741;
b) given (A78, 2TEX (20)), we obtain
TOBE = 0, M = 6;
c) given (S.2.2, 1NN100, 2NLM(K)*10), we obtain
TOBE = 1, M = 11, N100 = 7, LM(K) = 410;
d) given I = 3, LEN = 2,
(S.I.LEN, 1NN100, 2NLM(K)*10),
we obtain
TOBE = 0, M = 6.

The nonobligatory group operator should be used for decoding that part of the telegram where the data are key coded. In decoding the information falling between the limiters it is convenient to use a nonobligatory group operator in seeking the limiters and the information itself is processed by obligatory group operators, incorporating the decoded values into the data mass.

IV. Checking operator.
<CH OP> :: = O

This operator performs a comparison of the marker M with the content of the block LENGTH. In a case when M> LENGTH, there is shifting to the marker 77777. This operator is conveniently used for checking at the end of the telegram.

V. Print-out suppression operator. The operator has the form C+1. These symbols must be in the first three positions in the line.

With the appearance of this operator the printout of the FORTRAN elements generated by the macroprocessor ceases.

VI. Begin print-out operator. The operator has the form C+O (in the first three positions).

This operator cancels the action of the preceding operator.

All the generated elements are printed out prior to appearance of the first print-out control operator.

The proposed language substantially facilitates the process of writing of decoding programs. There is a considerable decrease in the volume of the initial program and the time required for its writing. This considerably increases the programmer's work efficiency.

As an example we will give a program for decoding the code FM-14V SYNOP [3].

```
          SUBROUTINE FM14
         *(TEXT, M, LENGTH, IU1L)
C+DCP
C+!
C+       ( ); INIST1L; 1NIR1L;
C+       1NRN1H; 1NV1H; 1DRH1L;
         RN1=RN1H * 1.25,
         CALL FM2408
        * (TEXT, M, LENGTH, IU1L)
         RETURN
         END
         SUBROUTINE FM2408
        *(TEXT, M, LENGTH, IU1L)
C+DCP
C+!
         TZ=0.514
         IF (MOD(IU1L, 2). EQ. 0)
        *TZ=1.0
C+       ( ); (0, 2NRD2L*10,
        2NRF2L*TZ);
C+       O;
C+       ( ); (1, 2NIWPLL, 2NIW2H);
C+       O; ( );
C+       (2, 1NISN1L, 3NRT3H*10); O;
C+       ( );
C+       (3, 1NJSN1L, 3NRTD3H*10); O;
C+       ( );
C+       (4, 4NRP04H); O; ( ); (5,
C+       4NRP4H); O; ( );
C+       (6, 1NIA1PLL, 3NRPV3L); C;
C+       ( );
C+       (7, 3DRR3H. 1NITR1L); C; ( )
C+       (8, 1NINH1N*1.25, 1NICL1H,
C+       *1NICM1H, 1NICH1N);
         IF (ISN1L. EQ. 1) RT3H

        *=-RT3H
        IF (JSN1L. EQ. 1)   RTD3H
       *=-RTD3H
        IF (RP04H. NE. -9999. AND.
       * RP04H. LT. 1000)
      * RP04H=RP04H+10 000.
        IF  (RP4H. NE.-9999.   AND.

          * RP4H. LT. 1000)
         * RP4H=RP4H+10 000.
77777      RETURN
           END
```

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EXPERIENCE IN APPLICATION OF A COMPLEX WORK QUALITY CONTROL SYSTEM AT THE AVIATION METEOROLOGICAL STATIONS OF THE UKRAINIAN ADMINISTRATION OF HYDROMETEOROLOGY AND ENVIRONMENTAL MONITORING

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[Article by N. P. Skripnik, head of the Ukrainian Republic Administration of Hydrometeorology and Environmental Monitoring, manuscript received 27 May 80]

[Text]

Abstract: The article describes 18-month experience with the introduction of a complex work quality control system at the aviation meteorological stations of the Ukrainian Administration of Hydrometeorology and Environmental Monitoring.

In solving the problems involved in the further raising of the material and cultural life of the Soviet people formulated by the 25th CPSU Congress, a factor of great importance is the systematic and undeviating increase in the effectiveness of production and the quality of work in all branches of the national economy.

Among the means and work methods directed to solution of the key problems of the Tenth Five-Year Plan we should include the complex work quality control system created through the initiative and experience of workers in leading units of the Ukrainian Administration of Hydrometeorology and Environmental Monitoring.

The necessity for a substantial improvement in work quality in the service is felt acutely particularly at aviation meteorological stations engaged in meteorological support of flights of civil aviation, whose requirements for meteorological support are constantly increasing. It is natural that the experience of the leading industrial and transportation enterprises of the country and the recommendations of the Central Committee CPSU on quality control problems have found response and have been caught up by the workers of the Ukrainian Administration of Hydrometeorology and Environmental Monitoring responsible for and dedicated to the meteorological support of civil aviation.

The fundamental principles and criteria for the complex system of work quality control, the formulas for evaluations and the quality standards were developed in such a way that they stimulated:
-- the scientific organization of labor;
-- the introduction into work and the use, on a large scale, of the latest technical advances in observations, information and communication;

-- more complete computational validation of the forecasts and warnings which are issued;
-- more active work of rationalizers and inventors;
-- exchange of leading experience;
-- development of instructional work, creative initiative, activation and improvement of forms of socialist competition;
-- creation of more favorable working conditions with respect to work hygiene and moral aspects, etc.;
-- more rational use of working time and successful implementation of production plans.

The basis for the development of evaluation principles was the use of formulas and coefficients of fault-free work which have come into wide use.

The principal formula for the evaluation of work quality is

$$K_{qual} = 1 - \sum K_{dec} + \sum K_{inc},$$

where K_{qual} is the work quality coefficient; 1 is the evaluation of fault-free work; $\sum K_{dec}$ is the sum of the coefficients of decrease in the evaluation of work quality during a definite period; $\sum K_{inc}$ is the sum of coefficients of increase in the evaluation of work during this same period.

The decrease and increase in the evaluation of work quality are determined employing special standards (coefficients), on the one hand taking into account the shortcomings, omissions and errors in work, and on the other hand, the positive results and attainments exceeding the established norms and requirements on the work done. All the workers at the aviation meteorological stations (forecasters, meteorological observers, communications specialists, aerologists, specialists on new equipment, workers with meteorological radars, etc.) have their own particular standards (coefficients) of decrease and increase in the evaluation of work quality for a shift and a month, making possible a differentiated evaluation for all aspects and important components in the work of each person. Standards were also formulated for evaluating the quality of work separately for all groups at aviation meteorological stations, for group leaders and the entire work force of aviation meteorological stations during a month.

In accordance with the "Instructions on the Complex Work Quality Control System at Aviation Meteorological Stations," the evaluation of the work of each worker is made in three steps:
1) daily, by one another, and by the shift supervisor (duty officer);
2) each 3-5 days by the group supervisor together with the trade union group organizer;
3) each 10 days by the chief of the aviation meteorological station and the chairman of the local committee or shop committee.

In order to publicize the results and for moral stimulation the evaluations for all workers and duty periods and for the month are posted on a special bulletin board.

In accordance with the "Instructions on the Complex Work Quality Control System," when a worker at an aviation meteorological station or in a group in a month meets the conditions for the award of a prize, as established in the "Instructions on the

Awarding of Prizes to Workers in Network Subdivisions of the Ukrainian Administration of the Hydrometeorological Service," the size of the prize is determined in dependence on the value of the quality coefficient for his work during the month ($K_{qual\ month}$) in the following way:

if $K_{qual\ month} \geqslant 1$ the prize is reckoned at 15% of the base pay;
if $0.99 \geqslant K_{qual\ month} \geqslant 0.90$ the prize is in the sum of 10% of the base pay;
if $0.89 \geqslant K_{qual\ month} \geqslant 0.80$ the prize is in the sum of 5% of the base pay;
if $K_{qual\ month} \leqslant 0.79$ no prize is awarded.

At the Ukrainian Administration of the Hydrometeorological Service work on the creation of a complex work quality control system was initiated in 1977. After appropriate elaboration during routine work in the network of aviation meteorological stations it was introduced beginning in October 1978.

The experience in routine introduction of the complex work quality control system has now made it possible to clarify the degree and mechanism of its effect on work quality. Our analysis of materials from the complex system has revealed the following:

First, the introduction of a complex work quality control system exerted a great positive effect on the quality of all important aspects of operation of aviation meteorological stations, in particular:
-- a higher quality was attained in the finalization of meteorological and operational documentation; there was a decrease in the number of errors in making aviation meteorological observations by more than half (by 60-70%);
-- there was a considerable improvement in the quality of processing of aerosynoptic material and almost a twofold decrease in the number of errors in its analysis;
-- there was a decrease in the number of interrupted flights and landings of aircraft at alternative airports due to incorrect forecasts or other shortcomings in the meteorological support of flights due to shortcomings of workers at aviation meteorological stations;
-- as a result of the more regular use of computational methods under complex meteorological conditions there was some improvement in the quality of aviation weather forecasts;
-- there has been almost complete exclusion of cases of deviation from terminology and NMO GA-73 requirements, as well as the malfunctioning of meteorological instruments and equipment due to shortcomings of workers at aviation meteorological stations;
-- there has been an increase in the quality of the various kinds of meteorological information received by flight personnel in meteorological briefings; the number of delays in the relaying of meteorological information has been reduced to a minimum;
-- there has been an increase in the competence of workers and their dedication to duty, as well as an improvement in the interrelationships with airport services.

Second, the basis for the positive effect of the complex work quality control system is attributable primarily to the following:

-- a constant, daily, complete and objective monitoring and evaluation of the work of each worker in all aspects of his productive activity, taking into account not only errors and omissions in work, but also his positive results and achievements

which earlier had not always been noted or were taken into account incompletely;
-- obtaining objective quantitative evaluations of the work of each worker, the group and the entire work force, which are easily comparable and make it possible to determine the contribution of each worker and the group as a whole in the work of the organization;
-- proper material stimulation, developing creative initiative and increasing the interest of each worker in fault-free work, the implementation of socialist obligations and production plans.

Third, the complex work quality control system exerts a controlling effect, that is: the constant analysis of the quality of work of individual workers and groups as a whole enables the administration and social organizations to concentrate attention on the "bottlenecks," as well as to carry out necessary measures directed to improvement of the work of both individual workers and the entire work force (instruction and shifting of personnel, assistance to those working below the standards, creation of the necessary working conditions, introduction of improved work organization, etc.).

Fourth, the most positive effect from the routine introduction of the complex work quality control system with relatively small expenditures of working time has been registered at those aviation meteorological stations where the chiefs and group supervisors have dealt seriously and responsibly with the introduction of the complex system and have constantly monitored it, devoting proper attention to it, where they have clearly thought out and perfected the order for regular daily and two-level monitoring, evaluation and consideration of all the important stages in work in all groups (shifts, work periods), taking into account their specific circumstances and personnel resources, where the summarization of the results of work of individual workers and groups during the month, with determination of the winners of the socialist competition, is accomplished with extensive publicity on "quality days" and there is appropriate determination of measures for moral and material approval.

Fifth, the considerable number of proposals received at the administrations of the Hydrometeorological Service from many aviation meteorological stations for the improvement of the complex work quality control system, enlargement of the table of quality standards, supplementation of these with a section taking into account production discipline, social behavior of workers, etc. indicate, on the one hand, that the problems involved in work quality control are important and urgent, that serious attention is being devoted to them at many aviation meteorological stations and that constant thought is given to them, and on the other hand, that the possibilities of improving the complex system have not been completely exhausted and that the administrations of the Hydrometeorological Service have definite work to do along these lines.

In conclusion we consider it necessary to emphasize once again that the complex work quality control system involves a wide range of organizational, methodological and economic measures which make possible a systematic increase in the quality of work and its maintenance at a high level and the successful implementation of the major and responsible tasks of meteorological support of Civil Aviation assigned to the aviation meteorological stations.

REVIEW OF BOOK 'COMPUTATIONS OF RUNOFF OF RIVERS AND INTERMITTENT WATERCOURSES' (RASCHETY STOKA REK I VREMENNYKH VODOTOKOV), IZD-VO VORONEZHSKOGO UNIVERSITETA, 1979, 200 PAGES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 120-121

[Review by G. A. Alekseyev, professor]

[Text] This book was prepared in the Department of Hydrology of the Land at Voronezh University by a group of authors under the direction of Professor A. G. Kurdov.

In the foreword it is stated that the first part of the book (Chapters 1-3), being of scientific interest, is devoted to the theoretical principles for computation of the norms of runoff of intermittent watercourses, permanent and seasonal rivers, and the development of new methods. The second part (six chapters) is of practical importance. It makes use of recommendations on the application of the proposed methods for computing all categories of runoff [the reviewer stresses this] for individual regions of our country and gives examples of computations.

Chapter 1 gives 55 scientific equations for the water balance of rivers and intermittent watercourses (1.1)-(1.55) for the periods of formation of low (minimum), high (maximum) and mean annual runoff, and also for suspended sediments and matter dissolved in water. These equations include components which for all practical purposes cannot be determined. For example, for spring runoff the mean channel water discharge during high water is

$$\overline{Q}_{spr} = (X' - P'' - Z)\Omega_{spr} \cdot 10^3 \qquad (1.20)$$

and the runoff volume is

$$\overline{W}_{spr} = \overline{Q}_{spr} T_{spr}, \qquad (1.21)$$

where Ω_{spr} is the "mean surface area (km^2) at high water in the entire intermittent network situated in the area of the drainage basin F_0," T_{spr} is the "mean duration of spring high water of an intermittent watercourse," X' is the mean intensity of water supply (lateral inflow and precipitation) per unit area of the water surface, P" is the mean intensity of filtering of water along the moistened perimeter of the channel flow, Z is the mean intensity of evaporation from the water surface.

At the end of Chapter 1 (#4) there are 55 purely symbolic correlation equations (1.56)-(1.110) for making corrections ($\Delta Q = \overline{Q} - \tilde{Q}$, $\Delta W = \overline{W} - \tilde{W}$) to the mean regional runoff values ($\tilde{Q}$, $\tilde{W}$), averaged for the observation period, as a function of the "anomalous (to be more precise, azonal — editor's note) physiographic factors in the drainage basin and human activity."

However, subsequently (Chapters 2-9), instead of the 110 "theoretical" equations (1.1)-(1.110) use is made of the empirical relationships between the mean runoff values ($y \equiv \overline{Q}$ or $\overline{W}$) at the observation points and the runoff factors ($x_j \equiv F$, $\overline{H}_{mean}$, f_{forest}, and others) which within the limits of the discriminated hydrological regions are approximated by linear or fractional-power law empirical equations in the form

$$\tilde{y} = ax + b, \quad \tilde{y} = \sum_j a_j x_j + b,$$
$$\tilde{y} = K(F - F_0)^n. \qquad (*)$$

It should be emphasized that the structure and parameters of these empirical equations do not follow from the "theoretical" equations (1.1)-(1.110).

An empirical equation of the type (*) was evidently used for the first time by A. G. Kurdov in [2, 3] for computing the minimum runoff of small rivers (in the example of rivers in the Central Chernozem oblasts). At a later date, an appropriate equation of the type (*)

$$Q_{80\%} = 10^{-3}\, a\, (F + f_0)^n \qquad (**)$$

with allowance for the conversion factors $\lambda_p = Q_p Q_{80\%}$ to the minimum 30-day discharges with the guaranteed probability $p = 75$-97% was used for the entire territory of the USSR by A. M. Vladimirov in [1, 4] without citations to the work of A. G. Kurdov [2, 3]. This fact is noted in the reviewed monograph, but at the same time it is asserted without basis that A. G. Kurdov gave a theoretical validation of the purely empirical formula (*).

It is also necessary to note a number of other substantial shortcomings of the reviewed monograph:

1. The study examines only the mean long-term runoff values ($\overline{Q}$, $\overline{W}$) and does not give any recommendations for determining the computed runoff values (Q_p, W_p) with a stipulated excess probability p.

2. No consideration is given to the most progressive method for mapping or regionalization of parameters (quasiconstants -- to use the terminology of M. A. Velikanov), "purified" of the influence of azonal factors, widely used in practical work.

3. No consideration is given to and no recommendations are made concerning the method for determining the computed parameters on the basis of observations on adjacent analogue rivers, with allowance for the difference in zonal and azonal factors entering into the computation formula, widely used in hydrological computations.

4. In the water balance equations the author always places ± in front of its terms, whereas it is necessary to write only one sign, "plus" or "minus," otherwise it is impossible to write a program for computations on an electronic computer. The values of some terms can be positive or negative.

The materials presented above make it possible to conclude that the reviewed monograph, with respect to both its content and its form of exposition, was written at a low theoretical level, without regard for modern methods for computing river runoff.

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REVIEW OF THE MONOGRAPH 'MULTISIDED INVESTIGATIONS OF RESERVOIRS. NO III. THE MOZHAYSKOYE RESERVOIR' ('KOMPLEKSNYYE ISSLEDOVANIYA VODOKHRANILISHCH. VYP III. MOZHAYSKOYE VODOKHRANILISHCHE'), EDITED BY V. D. BYKOV AND K. K. EDELSHTEYN, MOSCOW, IZDATEL'STVO MGU, 1979, 399 PAGES

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 121-122

[Review by S. L. Vendrov, professor]

[Text] The construction of the Mozhayskoye Reservoir, situated in the upper course of the Moskva (Moscow) River, took place in 1955-1960. Together with other reservoirs of the Moskvoretskaya Hydrographic System it plays an important role in the water supply of Moscow. And this role has been increasing during recent years because the southwestern and western parts of the capital are rapidly developing. Now already more than 1/4 of the total volume of water used in Moscow water lines comes from the Moscow River -- due to its own runoff and also from a right-bank tributary of the Volga River, the Vazuza River through the water-divide Vazuzkaya Hydrotechnical System.

In 1965 the Geography Faculty of Moscow State University organized the Krasnovidovskaya Scientific Research Laboratory for the study of reservoirs. It has grown into a fully developed scientific institute in which there are hydrological, hydrochemical, hydrobiological and ichthyological sections. The Krasnovidovskaya Laboratory organized an interfaculty complex expedition for the study of reservoirs which included specialists of the geography, biology and soil science faculties, etc. As a rule 50-70 specialists work annually on the expedition. The Hydrometeorological Service opened a meteorological station at the Mozhayskoye Reservoir. One of the important objects of expeditionary and laboratory work is the Mozhayskoye Reservoir, where the laboratory has developed its research methods.

Among the great many publications on different aspects of the hydrological, hydrochemical and biological regimes of reservoirs there are not many investigations having a complex character which would shed light on the above-mentioned processes in their interrelationship and interaction for the purpose of revealing the factors governing water quality.

The monograph is characterized by a highly varied, multisided analysis of field investigations and results closely related to the economic role played by the Mozhayskoye and other reservoirs of the Moskvoretskaya System in the water supply of

Moscow. It is therefore not without reason that five chapters (VI-X) of the twelve are devoted to different processes governing definite properties of water and characterizing its quality, including: radiation and heat balance, dynamics and formation of winter and summer water masses; mineralization and salt composition of the water; balance of principal ions; regime of biogenous substances, nitrogen in silts, trace elements and organic matter; suspended sediments and accumulation of alluvium; primary production of photosynthesis, phyto- and zooplankton, micro- and macrobenthos. All these sections are tied in well to the earlier presented characteristics of the watershed, runoff from it and the water regime of the reservoir.

In connection with the recreational role of the Mozhayskoye Reservoir a large chapter is devoted to the ichthyofauna, this being done in the interests of recreational fishermen.

The book is well documented with factual material, is illustrated with more than 70 figures and is supplied with a long bibliography. The group of authors, consisting of professors, instructors and scientific specialists in many departments at the university, made use of publications and research materials on the hydrology and hydrobiology of other reservoirs in this same geographic zone, which has greatly enriched the book.

By way of critical comments it can be said that the range of problems discussed in the final chapter, devoted to the influence of the Mozhayskoye Reservoir on the environment, should be expanded, with an examination of the mutual influence of the water body and the adjacent terrain. The publishing house of Moscow State University is to be reproached for the fact that the book, as indicated by the publication data, was in production for almost four years.

However, we assure the reader that it is by no means out of date, for which it is necessary to thank its editors, who ensured the necessary revision of some data.

The reviewed monograph is a useful book in both scientific and practical respects. It will undoubtedly also be used in other colleges in the country.

NINETIETH BIRTHDAY OF YEVGENIYA SAMOYLOVNA RUBINSHTEYN

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 p 123

[Article by T. V. Pokrovskaya and L. G. Polozova]

[Text] Professor Yevgeniya Samoylovna Rubinshteyn, doctor of geographical sciences, the oldest worker of the Hydrometeorological Service, an outstanding climatologist of the Soviet Union and a Meritorious Scientific Worker of the RSFSR, marked her 90th birthday in January 1981. Yevgeniya Samoylovna began to work at the Main Geophysical Observatory as a scientific specialist 65 years ago.

In pre-revolutionary Russia the path to science was not easy for a woman, but Yevgeniya Samoylovna advanced along this path with success and without restraints due to her diversified capabilities, which she manifested early, her quest for knowledge, exceptional capacity and love for work. In 1908 she graduated from secondary school with a gold medal. In 1913 she completed the Higher Female (Bestuzhevskiye) Courses and in 1914 took a university examination in the physical and mathematical sciences, receiving a first-degree diploma. That same year she began her work at the Main Physical Observatory as a nonstaff computer. In this modest post Yevgeniya Samoylovna immediately undertook a major task: the collection and processing of data on air temperature in Russia for the purpose of subsequent major generalization. This task was included in the plan for a climatography of Russia prepared with the participation of A. I. Voyeykov, with whom Yevgeniya Samoylovna came into personal contact during these years.

The first monograph by Yevgeniya Samoylovna was published later, in 1926-1927. It was devoted to a characterization and analysis of air temperature in the European part of the USSR.

During the course of her activity Yevgeniya Samoylovna has been concerned with major scientific problems of climatology and carried out fundamental investigations and generalizations as an author, scientific director and work organizer. To a considerable degree as a result of her energy there was assurance of such important aspects of the activity of the Hydrometeorological Service as the rationalization of the work of the meteorological network, creation of a unified group of climatologists working by themselves and at observatories, and carrying out major generalizations of the climate of the USSR, as well as supporting the national economy with climatological information.

Over the course of a number of decades Yevgeniya Samoylovna headed climatological subdivisions of the Main Geophysical Observatory, and during the prewar years was director of the Institute of Climatology, at that time, together with other branch institutes, forming part of the observatory.

The lists of personal scientific works of Yevgeniya Samoylovna are extremely long. They contain more than 100 titles, including major monographs, articles and textbooks. She has performed great services in the training of professional climatologists.

In 1976, after working at the Main Geophysical Observatory for 62 years, Yevgeniya Samoylovna left its walls for reasons of health. But she still remains a scientist, retaining interest in that field of science to which she devoted her life, and is actively working in this field.

Her book entitled ODNORODNOST' METEOROLOGICHESKIKH RYADOV VO VREMENI I PROSTRANSTVE V SVYAZI S ISSLEDOVANIYEM IZMENENIYA KLIMATA (Uniformity of Meteorological Series in Time and Space in Relation to an Investigation of Climatic Change) is dated 1979. Yevgeniya Samoylovna shares her great experience and knowledge in that field of climatology which is destined to ensure a proper approach to study of the nature of climate in its most complex, interesting and timely manifestations -- temporal variations and changes.

The creative activity of Yevgeniya Samoylovna is truly surprising. She continues to live for the interests of climatology, concerned as to completion of studies conceived long ago but still incomplete and realizing many of her scientific ideas. Yevgeniya Samoylovna is interested in all events of our time, both scientific and social.

We wish Yevgeniya Samoylovna health, strength, and all the conditions for realizing her scientific plans.

SEVENTY-FIFTH BIRTHDAY OF VASILIY ALEKSEYEVICH BELINSKIY

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 p 124

[Article by members of the Geography Faculty at Moscow State University]

[Text] Professor Vasiliy Alekseyevich Belinskiy, doctor of physical and mathematical sciences, a member of the CPSU since 1925, marked his 75th birthday and 50 years of scientific and teaching activity on 30 December 1980.

V. A. Belinskiy was born in a poor peasant family. In 1923 he graduated from intermediate school and in 1930 from Moscow State University in the Geophysics Department in the Physics Faculty. The Moscow Hydrometeorological Institute was created on the basis of this department with the active participation of V. A. Belinskiy. He became its first director. At the same time he headed the Moscow Division of the Aerological Institute.

After 1930 V. A. Belinskiy, in addition to an enormous amount of scientific and scientific-organizational activity, did much teaching work as an instructor and as a department head at the Moscow Hydrometeorological Institute and later at the Higher Military Hydrometeorological Institute, Moscow Finance Institute and Moscow State University. Vasiliy Alekseyevich was the author of the first Soviet textbook on dynamic meteorology, in which he simply and meticulously, at a high scientific level, dealt with the most complex subject matter. He authored textbooks on higher mathematics and aerology. A great number of course and diploma projects, as well as candidate's dissertations, were prepared under his direction.

V. A. Belinskiy is also known as a tireless researcher and experimenter, filled with energy. While working in the post-war years at the Central Aerological Observatory, he repeatedly made research flights in balloons. While a professor in the Geography Faculty at Moscow State University he organized and headed the scientific research work of the Department of Meteorology and Climatology in the Pamirs, in the Caucasus and in many other regions of the USSR.

He published more than 150 scientific studies devoted to the most timely problems in hydrometeorological science. Beginning in the 1950's, V. A. Belinskiy undertook studies of radiation climatology, laying the principles and developing a new direction -- investigation of the regime of ultraviolet radiation and the bioclimatic aspects associated with it. The monograph UL'TRAFIOLETOVAYA RADIATSIYA SOLNTSA I NEBA (Ultraviolet Radiation of the Sun and Sky) (1968), written by him in collaboration with his students, is widely known. He was awarded the D. N. Anuchin Prize for the atlas UL'TRAFIOLETOVAYA RADIATSIYA SOLNTSA I NEBA (Ultraviolet Radiation of the Sun and Sky).

Numerous students, friends and colleagues warmly and sincerely congratulate beloved Vasiliy Alekseyevich Belinskiy on his noteworthy anniversary and wish him good health and great successes in implementation of his creative plans.

AWARDS OF THE USSR ALL-UNION EXHIBITION OF ACHIEVEMENTS IN THE NATIONAL ECONOMY

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 124-126

[Article by M. M. Kuznetsova]

[Text] The Main Committee of the USSR Exhibition of Achievements in the National Economy has presented awards to participants in the USSR All-Union Exhibition of Achievements in the National Economy in the specialized exhibitions "USSR Hydrometeorological Center -- 50 Years," "The Environment -- Reliable Monitoring" and the specialized exposition "Investigation of the Free Atmosphere," held in the pavilion "Hydrometeorological Service."

First-Degree Diploma:

-- USSR Order of Lenin Hydrometeorological Scientific Research Center: for developing a system for processing routine meteorological information used in preparing numerical forecasts of the fields of meteorological elements, making it possible to accelerate the time required for the preparation of weather forecasts and increasing their quality;

-- Central Aerological Observatory: for work on systems engineering for the measurement system and development of the scientific-methodological and technical principles for the nosecone of the MMR-06M rocket. The results of the investigations made it possible, when using a standard engine, to increase the altitude for ascent of the payload to 20 km and reduce the zone of danger from the falling of the nosecone and engine;

-- Order of the Red Banner of Labor Institute of Applied Geophysics: for theoretical and experimental development of a system for monitoring the transport of substances contaminating the air across the national boundary of the USSR, ensuring the functioning of this system and carrying out the international obligations of the USSR following from the "Convention on the Trans-Boundary Contamination of Air Over Great Distances With Respect to the Monitoring and Exchange of Data on the Trans-Boundary Flows of Substances Contaminating the Air";

-- Central Design Bureau of Hydrometeorological Instrument Making: for developing and introducing an automatic station for monitoring atmospheric contamination (ASKZA -- avtomaticheskaya stantsiya kontrolya zagryazneniya atmosfery).

The annual savings from use of the station was 28,700 rubles.

Second-Degree Diploma:

-- Order of the Red Banner of Labor Main Geophysical Observatory imeni A. I. Voyeykov: for developing a method for the meteorological forecasting of air contamination and formulation of recommendations on reduction of effluent at enterprises for the purpose of preventing an increase in the concentrations of impurities during unfavorable periods;

-- State Oceanographic Institute: for formulating the principles for the organization of a hydrobiological network in the seas and introduction of a method for monitoring the quality of sea waters on the basis of hydrobiological indices, preparation of reviews of the quality of sea waters on the basis of hydrobiological indices and the carrying out of joint Soviet-Swedish expeditions in the Baltic;

-- Institute of Experimental Meteorology: for developing and introducing a system for observations of soil contamination in the USSR by metals and chlororganic pesticides, development and introduction of methods for determining DDT and its metabolites in soils and methods for ascertaining the gross quantities of metals in soils;

-- Laboratory for the Monitoring of the Environment and Climate: for participation in investigations and coordination of work on the problem "Global System for Monitoring the Environment" in the scientific-technical cooperation of the member countries of the Socialist Economic Bloc and implementing joint studies in the field of many-sided background monitoring, for the preparation and development of a program for a joint expeditionary experiment carried out in the territory of the Hungarian People's Republic in 1979, the introduction of the methods tested in the course of the experiment in the network of complex stations in the member countries of the Socialist Economic Bloc.

Third-Degree Diplomas:

-- Uzbek Republic Administration of Hydrometeorology and Environmental Monitoring: for developing and introducing a method for predicting air contamination in Tashkentskaya Oblast, a complex of measures for reducing effluent into the atmosphere during the effective period of the forecast, the effectiveness of which is evaluated on the basis of the decrease in the concentration of impurities in the atmosphere;

-- Lithuanian Republic Administration of Hydrometeorology and Environmental Monitoring: for operational prediction of high levels of atmospheric contamination, and also routine introduction of forecasts of background contamination for the city of Vilnyus;

-- Ural Territorial Administration of Hydrometeorology and Environmental Monitoring: for carrying out investigations and refinement of a method for predicting high levels of air contamination and introduction of these forecasts into routine practice;

-- Irkutsk Territorial Administration of Hydrometeorology and Environmental Monitoring: for prediction and prevention of high levels of atmospheric contamination and introduction into routine practice of a method for predicting high levels of contamination of atmospheric air under the conditions prevailing in Eastern Siberia;

-- Upper Volga Territorial Administration of Hydrometeorology and Environmental Monitoring: for the introduction, on a routine basis, of a method for predicting high levels of atmospheric contamination and work on the organization of the monitoring of atmospheric contamination in the region of the "Yasnaya Polyana" Botanical Reserve;

-- Kazakh Republic Administration of Hydrometeorology and Environmental Monitoring: for the introduction, on a practical basis, of a method for predicting high levels of atmospheric contamination and organization of a system of observations and warnings of anticipated dangerous conditions of air contamination over the territory of Kazakhstan;

-- Sakhalin Territorial Administration of Hydrometeorology and Environmental Monitoring: for carrying out varied hydrobiological observations over the territory of Sakhalinskaya Oblast, the introduction of the results of investigations in the national economy, the development of a scale for evaluating the contamination of rivers in the southern part of Sakhalin with respect to functional characteristics for an expert evaluation of the quality of waters and the state of water ecosystems.

A number of workers of the USSR State Committee on Hydrometeorology and Environmental Monitoring have been awarded the Diploma of Honor and medals of the USSR All-Union Exhibition of Achievements in the National Economy.

The Diploma of Honor was awarded to:

Sh. A. Musayelyan, M. A. Petrosyants, V. G. Shishkov, D. A. Drogaytsev, P. S. Lineykin, V. V. Rakhmanov, N. A. Aristov (USSR Hydrometeorological Center), I. S. Moshnikov, G. A. Kokin (Central Aerological Observatory), A. A. Shidlovskiy (Central Design Bureau of Hydrometeorological Instrument Making)

A Gold Medal was awarded to:

Ye. G. Lomonosov, V. N. Parshin (USSR Hydrometeorological Center), V. I. Yermakov (Central Aerological Observatory), I. M. Nazarov (Institute of Applied Geophysics), A. I. Mekhovich (Central Design Bureau of Hydrometeorological Instrument Making).

A Silver Medal was awarded to:

S. L. Belousov, B. S. Chuchkalov, P. P. Vasil'yev, A. A. Akulinicheva, A. I. Fotiyev, V. P. Sadokov, A. A. Vasil'yev, Z. K. Abuzyarov, T. A. Pobetova, M. G. Lubnin, V. A. Fedorov (USSR Hydrometeorological Center), Ye. A. Besyadovskiy, S. P. Perov (Central Aerological Observatory), V. G. Khvostov, I. V. Gryts'kiv,

A. G. Roshchin, I. P. Kuz'minykh (Central Design Bureau of Hydrometeorological Instrument Making), A. Ya. Pressman, A. F. Yakovlev, V. A. Abakumov, I. B. Pudovkina, M. M. Novikov, Sh. D. Fridman (Institute of Applied Geophysics), L. R. Son'kin (Main Geophysical Observatory), A. V. Tsyban' (State Oceanographic Institute), E. I. Babkina, E. P. Makhon'ko, M. I. Allenov (Institute of Experimental Meteorology), F. Ya. Rovinskiy, L. M. Filippova (Aerial Methods Laboratory), I. V. Tsvetkov (State Committee on Hydrometeorology), V. I. Kuznetsov (Northwestern Administration of the Hydrometeorological Service).

A Bronze Medal was awarded to:

L. V. Berkovich, T. G. Ivanidze, T. S. Kruzhkova, R. G. Petukhova, I. A. Petrichenko, Ye. S. Kuryleva, N. N. Bel'skaya, A. D. Chistyakov, N. Ye. Minakova, K. A. Vasyukov, M. A. Sorochinskiy, A. Snitkovskiy, N. M. Zakharova, G. K. Veselov, G. T. Ivanova, A. I. Bostrykina, T. M. Fedunova, A. I. Ugryumov, N. A. Chuzavkova, M. V. Rubinshteyn, K. M. Sirotov, M. A. Suchkov, N. S. Nechayeva, Ye. S. Zmiyeva, N. D. Yefremova, N. F. Dement'yev, V. N. Pupkov, A. N. Derevyanko, Ye. S. Zverintseva, V. V. Brezhnev, O. M. Kastin, Yu. L. Shmel'kin, Yu. K. Fedorov, V. I. Mamontov, N. I. Rumyantsev, A. I. Vanyushin, N. S. Yefimov, L. M. Neronova (USSR Hydrometeorological Center), Yu. V. Vasil'yev, A. P. Popov, V. M. Ignatov, V. N. Alin, V. I. Tatarenko, A. V. Komotskov, S. A. Vyazankin, V. I. Kon'kov, A. I. Zolkin, A. F. Chizhov (Central Aerological Observatory), V. A. Mal'skiy, A. M. Rudometkin, N. T. Romanenko, V. G. Pilipyuk, V. P. Smyk, G. F. Markov, V. I. Timofeyev, V. G. Avdeyev, Ye. M. Pakhomchik (Central Design Bureau of Hydrometeorological Instrument Making), V. F. Shkilev (Far Eastern Scientific Research Institute), A. V. Lysak, V. F. Brendakov, N. I. Kholikova, V. P. Krivchikova, A. I. Koloskov, S. B. Iokhel'son, V. M. Artemov, V. I. Rozhdestvenskaya, O. S. Renne, V. N. Vasilenko, V. P. Chirkov (Institute of Applied Geophysics), E. Yu. Bezuglaya, I. A. Yankovskiy (Main Geophysical Observatory), O. A. Klimenko, V. V. Fadeyev, V. M. Mukhin, N. I. Katalevskiy (Geochemical Institute), G. V. Panov, Yu. L. Volodkovich, N. A. Afanas'yeva (State Oceanographic Institute), Ts. I. Bobovnikova, N. D. Tret'yakov, S. G. Malakhov, Ya. I. Gaziyev, G. G. Belov, E. G. Tertyshnik, G. N. Klochkov (Institute of Experimental Meteorology), L. I. Buyanova, A. Kh. Ostromogil'skiy, L. V. Burtseva (Aerial Methods Laboratory), A. I. Pankov, M. N. Popov (State Committee on Hydrometeorology), N. V. Koroleva, L. A. Pavlenko (Uzbek Administration of the Hydrometeorological Service), R. P. Chivilite (Lithuanian Administration of the Hydrometeorological Service), A. N. Andrianov, S. L. Basova, I. M. Markovets (Northwestern Administration of the Hydrometeorological Service), I. A. Shevchuk (Western Siberian Scientific Research Institute), N. A. Shapareva (Ural Administration of the Hydrometeorological Service), Yu. I. Onanko (Ukrainian Administration of the Hydrometeorological Service), I. N. Ponomarenko (Ukrainian Scientific Research Institute), B. B. Chebanenko, V. P. Zinov'yev (Irkutsk Administration of the Hydrometeorological Service), D. V. Vinokurova, P. A. Borisova (Upper Volga Administration of the Hydrometeorological Service), K. Zh. Omarova (Kazakh Administration of the Hydrometeorological Service), L. P. Gorelova, V. S. Gladkov (Central Volga Hydrometeorological Observatory), V. S. Kuznetsov (Northern Administration of the Hydrometeorological Service), L. V. Daksh (Latvian Administration of the Hydrometeorological Service), L. I. Sokhina (Murmanskoye Administration of the Hydrometeorological Service), N. P. Usova (Sakhalinskoye Administration of the Hydrometeorological Service), V. G. Prokacheva (State Hydrological Institute), O. N. Frantsuzov (Leningrad Division of the State Oceanographic Institute).

The total number of participants from the USSR State Committee on Hydrometeorology and Environmental Monitoring was 478 persons. In addition to workers of the State Committee on Hydrometeorology and the main exhibit of the USSR Exhibition of Achievements in the National Economy in the "Hydrometeorological Service" pavilion, awards were also given to outside organizations directly participating in the development of a number of themes.

CONFERENCES, MEETINGS AND SEMINARS

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 pp 126-127

[Article by N. S. Shapovalova and I. V. Trosnikov]

[Text] An international conference on the scientific results of the Atlantic Tropical Experiment (GATE), carried out in 1974 within the framework of the GARP tropical subprogram, was held in Kiev during the period 17-23 September 1980.

The conference was organized by a joint scientific committee working under the aegis of the World Meteorological Organization and the USSR National GARP Committee with the collaboration of the Administration of the Hydrometeorological Service Ukrainian SSR and was held in accordance with resolutions of the Joint Organizing Committee (JOC) of the WMO and collaborating agencies (JOC-XIII, 14-20 April 1974, Stockholm and JOC-XIV, 13-19 April 1978, Mexico City). According to these resolutions the most important GATE results were to be set forth in a monograph and its contents should be discussed at the conference.

In accordance with the recommendations of the 17th session of the working group on numerical experimentation the preparation of the monograph, which was given the name "Synthesis of GATE Scientific Results," was to be carried out in three stages. In the first stage the invited experts prepared the sections of the monograph assigned to them. The second stage was the calling of a conference at which there was to be presentation and discussion of the reports of invited experts. The third stage provided for the experts to present the finalized texts of sections of the monograph, taking into account the proposals and comments expressed at the conference.

Participating in the work of the conference were scientists from Great Britain, Brazil, East Germany, Mexico, Rumania, United States, France, West Germany and the USSR. The chairman of the international organizing committee for preparing for and holding the conference was M. A. Petrosyants, at the same time representing the joint scientific committee under WMO auspices. Among the members of the international organizing committee at the conference were P. Roundtree (Great Britain), N. P. Skripnik (Ukrainian SSR) and I. G. Sitnikov (WMO). Yu. V. Tarbeyev represented the USSR National GARP Committee.

The conferees heard 13 reports of invited experts who discussed all chapters of the monograph.

The monograph begins with the section "Principal Tasks of the Atlantic Tropical Experiment GARP," prepared by M. A. Petrosyants (USSR). This section sets forth the history of planning of GATE, the objectives and components of the basic program and its five subprograms (synoptic scale subprogram, convection subprogram, boundary layer investigation subprogram, radiation subprogram, oceanographic subprogram).

The tasks under the basic GATE program were defined in the following way:
1) ensure means for evaluating the influence of microscale weather systems in the tropics on macroscale circulation (synoptic scale);
2) facilitation of the development of numerical modeling and forecasting methods.

A report by the former director of the GATE scientific-administration group, I. Kutner (United States), entitled "Observational Strategy of GATE (Retrospective Look)," in addition to scientific results, examined organizational, technical and economic problems of GATE. In discussing this report the conferees expressed the unanimous opinion that the success of GATE was favored by a properly selected strategy for implementation of the experiment and the spirit of international cooperation.

I. Kutner also presented a report by T. N. Krishnamurti and R. G. Pash (United States) entitled "Macroscale Mean State of the Atmosphere During GATE." The report gives the results of an analysis of the wind and ocean surface temperature fields and some other thermodynamic characteristics of the atmosphere and ocean surface in the course of the three phases of GATE. A report by R. W. Burpee and R. G. Reid (United States), entitled "Synoptic Scale Movements," was presented by P. Roundtree (Great Britain). It was devoted to a review of investigations of tropical waves and disturbances made on the basis of GATE data. The review included the problems involved in the development and energy characteristics of easterly waves in the West Africa and Eastern Atlantic regions and their relationships to convection and precipitation. The problems involved in macroscale disturbances of the ICZ and the development of hurricanes were also discussed.

Ye. M. Dobryshman (USSR), in a report entitled "Theoretical Investigations of Tropical Waves," gave a review of existing movements in the equatorial atmosphere and gave the results of his own investigations of wave disturbances arising in the equatorial region. Much interest was shown in the results obtained by the author in his study of nonlinear effects.

A report by A. Gilchrist, P. Roundtree and D. Shaw (Great Britain), entitled "Macroscale Numerical Modeling," gave a review of studies (prepared using GATE data) on the collection of sets of data and on macroscale modeling.

The theoretical and observational aspects of study of the upper layers of the tropical ocean on the basis of GATE data were covered in reports by H. Zilder (West Germany) and G. Philander (United States).

A report by A. I. Fal'kovich (USSR), entitled "Movements of Scales A/B and B of the Balance in the ICZ Region," gave data on the structure and development of the ICZ during the GATE period, and also on components of the energy balance for the atmospheric column of air during different periods of ICZ development. The problems involved in interaction of movements of different scales are discussed.

A report by A. Betts and R. House (United States), entitled "Clouds, Convection and Convective Models," devoted to a review of the results of study of tropical convective systems, obtained on the basis of GATE data, was of considerable interest.

Investigations of the atmospheric boundary layer in the GATE region were discussed in reports by Yu. A. Volkov (USSR), entitled "Surface Layer (Interaction Between the Ocean and the Atmosphere) and Its Parameterization," and H. Hintzpeter and E. Augstein (West Germany), entitled "Structure of the Atmospheric Boundary Layer Under Different Convective Conditions (Observations and Models)."

In the section "Radiation Processes and Their Parameterization" reports were presented by Ye. M. Feygel'son and K. Ya. Kondrat'yev (USSR). The latter report was read by M. A. Prokof'yev.

The results of GATE were summarized in the "Final Comments," a report prepared by M. A. Petrosyants (USSR).

In addition to the reports of the invited experts, the conferees heard 23 additional reports. Twenty of these were related to the subject matter of individual chapters of the monograph and three were devoted to methodological problems related to the reliability and representativeness of observational data obtained during the GATE period. Taking into account the importance of studies of a methodological character, the conference adopted a resolution to the effect that methodological problems must be reflected in the monograph.

After the work of the conference was completed a conference of invited experts and organizers was called on problems involved in further preparation of the monograph (Kiev, 24 September 1980). This conference was attended by J. Smagorinsky, head of the WMO joint scientific committee, and Professor B. R. Deez, director of the WMO joint planning group. This conference summarized the results of the earlier conference, heard the opinions of experts on individual chapters of the monograph and discussed the content of the "Final Comments" presented by M. A. Petrosyants. Three reviewers for each chapter of the monograph were named and a plan for completing work on its final preparation and publication was outlined.

NOTES FROM ABROAD

Moscow METEOROLOGIYA I GIDROLOGIYA in Russian No 1, Jan 81 p 128

[Article by B. I. Silkin]

[Text] As reported in SCIENCE NEWS, Vol 116, No 19, p 324, 1979, about a half-century ago the Serbian scientist Milutin Milankovic formulated the hypothesis that the onset of glaciation on the earth was associated with cyclic changes in the orbit of our planet around the sun, affecting the quantity and distribution of the energy received from the sun. Milankovic mentioned four such cycles: with a duration of 23,000 years, leading to changes in the time of closest approach of these two celestial bodies (now the earth and the sun are closest to each other in January, but in 10,000 years this will occur in July); with a duration of 41,000 years, during which the inclination of the earth's orbit becomes almost perpendicular to the axis of its characteristic rotation, which decreases the contrast between the seasons of the year and hinders the thawing of polar ice; cycles with a duration of 93,000 and 413,000 years, during which the orbit is transformed from an almost perfect circle into a more elliptical orbit, which leads to changes in the distances between the sun and the earth, exerting an influence on the season of the year.

Different researchers have been able, using cores taken during drilling, to show the reality of existence of the three Milankovic cycles, except the last, the longest, which remained hypothetical. And only now have scientific specialists at the University of Cincinatti (Ohio, United States) M. Briskin and J. Harrell found evidence of climatic variations corresponding to a cycle with a duration of 413,000 years.

Studying cores of sediments taken in the Atlantic and Pacific Oceans and taking in the last 2 million years, they determined the quantity of ice present in the earth's polar caps. They also measured the ratio of oxygen isotopes 0-18 and 0-16 in remnants of fossil plankton. The periods of glaciation were established from the relative increase in the quantity of 0-18, attributable to the fact that 0-16 is held by the ice. The content of the coarse fraction (heavier particles) in the sediments indicated periods with more active erosion, caused by melting of the ice. And in any of these parameters it was possible to trace three complete cycles with a duration of approximately 413,000 years each.

Similar evidence was also obtained in a study of the periodicity of changes in magnetic declination, detected in magnetic rocks. It is possible that this is not associated directly with climatic phenomena, but a correlation between the changes in the earth's orbit and variations of the earth's magnetic field is entirely probable.

Earlier a cycle with a duration of 413,000 years could not be detected due to the absence of a corresponding statistical method. Existing methods required the availability of a long series of data, from which the researcher should select samples equally distant in time from one another. The method employed by M. Briskin and J. Harrell makes it possible to overcome the lack of part of the data. The fact that their conclusions are based on materials of regions of the Atlantic and Pacific Oceans which are remote from one another and on different measurement parameters makes the conclusions, confirming the M. Milankovic hypothesis, quite reliable.

As reported in NATURE, Vol 286, p 114, 1980, and in NEW SCIENTIST, Vol 87, No 1211. p 287, 1980, a group of British participants in FGGE, headed by R. Hyde, has carried out a study of the influence of circulation of air masses on the rate of the earth's rotation. It is known that the seasonal movements of such masses at a global scale are capable of leading to changes in the length of day, amounting to milliseconds.

The observations now being made are some of the first cases of direct measurement of interrelated changes in atmospheric circulation and the length of day at the scale of several days, not a season. These measurements confirmed well the theory that the atmosphere, despite its considerably lesser mass than of the earth's solid body, is sometimes capable of causing changes in its rotation. This, evidently, is attributable to its positioning over the very surface of the planet and as a result of this there is an adequate angular momentum for the effect.

The measured effects attain only milliseconds. For example, atmospheric processes evidently caused a change in the duration of a day by 1 msec, occurring between 21 January and 7 February 1979, and by 1.5 msec, noted between 18 May and 2 June 1979.

It is noted that the contribution of the atmosphere to the total angular momentum of the earth's rotation changes together with the change in zonal circulation of the air envelope, when large air masses are driven in the direction of the equator or away from it.

It is also possible to trace a definite relationship between these processes and the solar wind. It was demonstrated earlier that activization of processes on the sun has consequences reflected in the weather in the high latitudes of the earth's northern hemisphere. It is evident that the arrival in the earth's neighborhood of an additional large stream of particles forming the solar wind somehow stimulates the development of systems of low atmospheric pressure which thereafter move rapidly from west to east at the latitudes of Alaska and Northern Europe.

From the point of view of atmospheric circulation this is equivalent to the movement of air masses precisely of the type which, as is now confirmed by FGGE data, can cause changes in the length of day which can be measured.

Thus, an observational confirmation has been obtained of the existence of relationships between changes in the level of spot formation on the sun and variations in the length of day, amounting to milliseconds. Until now many astronomers and geophysicists have regarded these relationships to be improbable.

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