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USSR Report

EARTH SCIENCES

(FOUO 2/81)

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JPRS L/9621

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USSR REPORT

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(FOUO 2/81)

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METEOROLOGY

UDC 551.5

RADIATION FACTORS IN CONTEMPORARY CHANGES IN GLOBAL CLIMATE

Moscow RADIATSIONNYYE FAKTORY SOVREMENNYKH IZMENENIY GLOBAL'NOGO KLIMATA in Russian 1980 signed to press 3 Jun 80 pp 2, 279

[Annotation and table of contents from book "Radiation Factors in Contemporary Changes in Global Climate", by K. Ya. Kondrat'yev, Gidrometeoizdat, 1,350 copies, 280 pages]

[Text] [Annotation] The most important external factors in contemporary changes in climate are possible variations of the solar constant and also the gas and aerosol composition of the atmosphere. This range of problems is the main content of the book. Emphasis is on the properties of atmospheric aerosol and its possible influence on climate, which is manifested in changes in the earth's albedo and the radiant heat influx in the atmosphere. Anthropogenic effects on the ozone layer are discussed (reference is primarily to halocarbons and their products), as well as the consequences of these modifications from the point of view of variations of the radiant heat influx in the stratosphere.

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OCEANOGRAPHY

UDC 551.466.81

MODEL OF THE LIMITING SPECTRUM OF INTERNAL WAVES

Moscow IZVESTIYA AKADEMII NAUK SSSR: FIZIKA ATMOSFERY I OKEANA in Russian Vol 16, No 11, 1980 pp 1173-1178

[Article by S. P. Levikov, State Oceanographic Institute, manuscript submitted 12 Mar 79, resubmitted after revision 19 Dec 79]

[Text]

Abstract: The author proposes a model of the high-frequency part of the spectrum of internal waves propagating in thin stratified layers of the thermocline. The field of internal waves is modeled by an ensemble of algebraic solitones with random amplitudes and phases. The article gives a comparison of the proposed theoretical curve with known evaluations of the spatial spectrum of internal waves in the ocean obtained on the basis of observational data.

It is known that the heights of surface gravitational waves developing under the influence of the wind are limited due to collapse. Some universal interval is formed in the spectrum of wave heights. It is assumed that the form of the spectral function in the spectrum can be obtained from an analysis of the dimensionalities [1]. Assuming that the one-dimensional frequency spectrum $\Phi(\omega)$ is determined only by the acceleration of gravity g and the frequency ω, we obtain

$$\Phi(\omega) \sim g^2\omega^{-5}.$$

The spectrum of wave numbers $\Psi(k)$ (we recall that the waves are two-dimensional) in this case has the form

$$\Psi(k) \sim k^{-4},$$

for sections of the wave field (plane waves) $\psi(k) \sim k^{-3}$.

Internal waves can also attain states of destruction so that it is necessary to expect the formation of a similar equilibrium interval in the spectrum of internal waves. In this article an attempt is made to derive an analytical expression for such an equilibrium (limiting) spectrum. In examining this problem we will use an analogy with a better studied process -- surface waves.

The correctness of the Phillips "ω^{-5} law" for the equilibrium interval in the spectrum of surface waves has been confirmed by numerous in situ measurements [2]. However, up to the present time this law has not been derived from any rigorous theoretical model. The results in [3] give the dependence $\Phi(\omega)\sim\omega^{-4}$ for the equilibrium interval in the spectrum of the slightly turbulent field of surface gravitational waves. The difference from the Phillips "ω^{-5} law" is associated with the fact that the authors of [3] examined slightly nonlinear processes of interaction of random wave fields; highly nonlinear interactions of the collapse type were not taken into account.

For an explanation of the equilibrium interval V. A. Krasil'nikov and V. I. Pavlov used a model of interaction between gravitational and capillary waves. Using as a point of departure some stipulated form for turbulence in the near-water layer of the atmosphere, they obtained the following dependence for the frequency spectrum of surface gravitational waves in the equilibrium interval:

$$\Phi(\omega)\sim\omega^{-11/4},$$

which is close to the Phillips law, but nevertheless differs from it.

For the time being no theoretical expression has been derived for the equilibrium interval in the spectrum of surface waves and therefore it is necessary to limit ourselves to less rigorous solutions, for example, those based on an analysis of dimensionalities.

However, it is impossible to use an analysis of dimensionalities for obtaining the spectrum of internal waves of interest to us. Phillips [5] proposed another approach based on the assumption that the increase in amplitudes of the spectral components in the field of the lower mode of internal waves is limited due to shear instability. The limiting amplitude of some spectral component a_s is determined by the critical Richardson number, which, as for steady plane-parallel flows, is assumed equal to 1/4

$$a_s^2 = 4\frac{\omega^2}{k^2 N_m^2}\sim\begin{cases} k^{-1} & \text{for short waves } (\omega^2\sim k), \\ k^0 & \text{for long waves } (\omega\sim k), \end{cases}$$

where ω is frequency; k is the wave number; N_m is the maximum Väisälä frequency; it is assumed that $\omega/N_m \ll 1$. The spectral density is determined as the square of wave amplitude per unit area in wave number space. We obtain

$$S(k)\sim\frac{a^2}{\Delta k_1 \Delta k_2}\sim\begin{cases} k^{-3} & \text{for short waves,} \\ k^{-2} & \text{for long waves.} \end{cases}$$

The model of a limiting spectrum of internal waves proposed by Phillips [5] is linear and has limitations following from its linearity — the entire solution is stipulated in the form of the superposing of independent spectral components. This assumption evidently ceases to be correct for waves close to collapse. Accordingly, in this case rigorously coupled sets of spectral components arise which in physical space correspond to a real nonlinear wave.

The model of the limiting spectrum of internal waves proposed in [6] has similar limitations associated with linearity.

We will not discuss the results obtained by Garrett and Munk [8,9], who proposed an expression for the "climatic" spectrum of internal waves, correct in the low-frequency range, where the assumptions made concerning the linearity of the field of internal waves are soundly validated.

We will use another approach: for computing the limiting spectrum of internal waves we will select a solution in the form of some superposing of nonlinear formations, each of which is a solution of the nonlinear problem corresponding to a wave with a limiting amplitude.

Spectrum of internal waves propagating through density layers. We will derive an expression for the high-frequency part of the spectrum of internal waves propagating in sharply defined layers of the thermocline. As is well known, it is precisely in thin stratified layers where there is an increase in the nonlinear effects favoring wave collapse [9, 10].

Now we will turn to an equation describing the dynamics of unsteady slightly nonlinear internal waves propagating through a stratified thin layer bounded above and below by a homogeneous deep fluid [11, 12]:

$$A_t=-C_0A_x+\alpha AA_x+\beta\frac{\partial^2}{\partial x^2}H[A],\ H[A]=-\frac{1}{\pi}\int_{-\infty}^{\infty}\frac{A(\xi)}{x-\xi}d\xi. \quad (1)$$

Here $A(x,t)$ is the horizontal velocity component in the wave field or the deviation of the isopycnic line from the position of equilibrium; C_0 is the velocity of wave propagation (in a linear approximation). The coefficients α and β are determined from the solutions of the linear boundary-value problem relative to the vertical coordinate and are thus dependent on the characteristics of stratification and the vertical velocity shear in an undisturbed state. Equation (1) was derived on the assumption of long waves relative to D-layer thickness, that is, the ratio D/ℓ is a small value (ℓ is the characteristic wave length). According to the generally accepted classification these internal waves are "high-quality" short waves (their lengths can be from several meters to several hundreds of meters). The effects of the earth's rotation and viscosity are not taken into account.

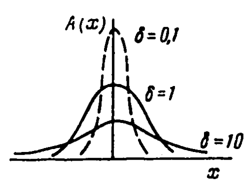

Fig. 1. Form of algebraic solitones for different values of amplitude parameter δ.

It is known that equation (1) has solutions in the form of stationary solitary waves, so-called algebraic solitones. In a moving coordinate system $x = \tilde{x} - C_0t$

$$A(x)=\frac{4\beta}{\alpha}\frac{\delta}{x^2+\delta^2}. \quad (2)$$

Figure 1 shows the shape of the wave for different values of the δ parameter. It can be seen that the lesser the δ value, the higher and narrower is the solitone. The slightly nonlinear theory allows the existence of solitones with an arbitrary amplitude (although it must be remembered that equation (1) was derived on the assumption of a small amplitude of the waves, it follows from the restrictions of slightly nonlinear theory). It was impossible to derive an expression for the limiting amplitude of the wave in this approximation; it must follow from the theory of strongly nonlinear waves. Such an evaluation, for example, was obtained for waves in rarefied plasma propagating across a magnetic field [14]. It was established in this case that a stationary wave exists with $H_{max} < 3H_0$, where H_0 is the magnetic field in undisturbed plasma, H_{max} is the maximum magnetic field in the solitone (here H_{max} is similar to the height of a wave of the limiting amplitude and H_0 is similar to the depth of the undisturbed fluid).

We will estimate the limiting height of a solitone, using an analogy with solitones, on the surface of the fluid. For example, for the latter the following expression is correct for the relationship between the depth h of the fluid and the maximum height ζ_{max} of the solitone [15]:

$$\zeta_{\text{макс}} = 0{,}78h \tag{3}$$

(similar estimates are also given in other sources). We will also assume that the height of the internal solitone is of the order of the thickness of the stratified layer D. In dimensional variables equation (1) assumes the form

$$\eta_t + \frac{rC_0}{D}\eta\eta_x + C_0 sD\frac{\partial^2}{\partial x^2}H[\eta] = 0, \tag{4}$$

where η is the height of the internal wave. (Conversion from (4) to (1) is accomplished by replacement of the variables $\eta = DA$, $x = Dx'$, $t = Dt'/C_0$). The solution (2) is transformed to

$$\eta(x) = \frac{4s}{r}D^2\frac{\delta}{x^2+\delta^2}. \tag{5}$$

From the condition $\eta(x = 0) \sim D$ we have $\frac{4s}{r}\,\frac{D^2}{\delta} \sim D$, that is

$$\delta_{min} \sim D. \tag{6}$$

We will estimate the spectral density of a one-dimensional record (series) formed from the superposing of algebraic solitones with random amplitudes and phases. The total one-dimensional wave field is stipulated in the form

$$\zeta(x) = \frac{4\beta}{\alpha}\sum_j \frac{\delta_j}{(x+r_j)^2+\delta_j^2}, \tag{7}$$

where r_j is a random phase. The Fourier transform of expression (7) will have the form

$$F[\zeta(x)] = \frac{4\beta}{\alpha}\int_{-\infty}^{\infty}\sum_j \frac{\delta_j}{(x+r_j^2)+\delta_j^2}e^{ikx}dx = \frac{4\beta}{\alpha}\sum_j\int_{-\infty}^{\infty}\frac{\delta_j}{z^2+\delta_j^2}e^{ikr}e^{ikr_j}dz =$$

$$= \frac{4\beta}{\alpha}\sum_j e^{ikr_j}\int_{-\infty}^{\infty}\frac{\delta_j e^{-ikz}}{z^2+\delta_j^2}dz = \frac{4\beta}{\alpha}\sum_j e^{ikr_j}(-\pi e^{-k\delta_j}).$$

We multiply by the complex-conjugate expression

$$FF^{*} = \frac{16\beta^{2}}{\alpha^{2}} \sum_{j} \sum_{j'} e^{ik(r_j - r_{j'})} (\pi^{2} e^{-k\delta_j} e^{-k\delta_{j'}}).$$

Now we will average it for the random phase, which we assume to be uniformly distributed in the interval (0.2π). The spectral density can be evaluated using the formula

$$\hat{S}(k) = \frac{\langle FF^{*} \rangle}{L} = \frac{16\beta^{2}\pi^{2}}{L\alpha^{2}} \sum_{j} e^{-2k\delta_j} \qquad (8)$$

(here the expression $\langle e^{ik(rj-rj')} \rangle \simeq \delta_{jj}$, where δ_{jj} is the Kronecker symbol, L is the length of the series to be analyzed).

The form of the spectrum $\hat{S}(k)$ with large k will be determined by the contribution of the one term corresponding to $\delta_j = \delta_{min}$, which attenuates relative to k more slowly than the others. Thus, the spectrum of fluctuations of the horizontal component of velocity with large k is described by an exponential dependence. The spectrum of fluctuations of the isopyc or the spectrum of density (temperature) fluctuations at a fixed depth has a similar form. The latter, however, can be attenuated due to the influence of the nonlinear vertical distribution of density (temperature) in the neighborhood of the measurement point.

Now we will compare the proposed theoretical curve with evaluations of the spectrum of internal waves cited in [17]. They were obtained on the basis of temperature measurements made using thermistors towed in the upper layer of the open region of the Atlantic Ocean. The vertical distribution of density in the region of the experiment approximately corresponded to a three-layer model within whose framework the theoretical curve for the spectra was obtained. The thickness of the mean stratified layer can be assumed equal to 80 m. As an evaluation of the δ_{min} value we can use a value equal to the thickness of the stratified layer. Figure 2 shows a theoretical curve which is compared with the evaluation of the spectrum given in [17]. The straight lines represent the power law dependences (k^{-2}, k^{-3}, k^{-4}).

The numerical constant in formula (8) before the summation symbol is dependent both on the stratification parameters r, S and the thickness of the layer D and on the length of the analyzed series L (that is, on the number of solitones in the series). Since its evaluation is not presented here, the position of the theoretical curve on the y-axis remains arbitrary, so that in the figure the curve is drawn from the condition of best agreement with the experimental evaluations (this is admissible since the graph is given in logarithmic coordinates).

The observational data make it possible to assume only that the field of internal waves in this case was saturated, that is, it can be modeled by a superposing of the algebraic solitones with a limiting amplitude. However, here we observe a far better agreement with the model curve than with the power-law dependences.

The form of the spectrum of fluctuations of the vertical velocity component is interesting. The following evaluation can be obtained for it. The vertical velocity component W in the field of an algebraic solitone is stipulated by the expression

$$W = A_*(x, t)\,\Phi(z) \tag{9}$$

At a fixed depth z_0 $\qquad W(z_0;\, x,\, t) = A(x,\, t)\,\Phi(z_0).$

In an algebraic solitone of the form $A(x,t) = 4\beta/\alpha)\,\delta/\lambda^2 4\delta^2$ we have (in a coordinate system moving together with the wave)

$$\begin{aligned}
&W(r_0, x) = \\
&= \Phi(z_0)\frac{\partial}{\partial x}\left(\frac{4\beta}{\alpha}\frac{\delta}{x^2+\delta^2}\right), \\
&F[W(x)]_{z=z_0} = \\
&= \frac{4\beta}{\alpha}\int_{-\infty}^{\infty} e^{-ikx}\frac{\partial}{\partial x}\left(\frac{\delta}{x^2+\delta^2}\right)dx = \\
&= \frac{4\beta}{\alpha}e^{-ikx}\frac{\delta}{x^2+\delta^2}\Big|_{-\infty}^{\infty} + \\
&+ \frac{4\beta}{\alpha}\int_{-\infty}^{\infty} ike^{-ikx}\frac{\delta}{x^2+\delta^2}dx = \\
&= \frac{4\beta ik}{\alpha}(-\pi e^{-k\delta}), \\
&\hat{S}(k) \sim \langle FF^* \rangle = \frac{16\beta^2\pi^2}{\alpha^2}k^2 e^{-2k\delta}.
\end{aligned} \tag{10}$$

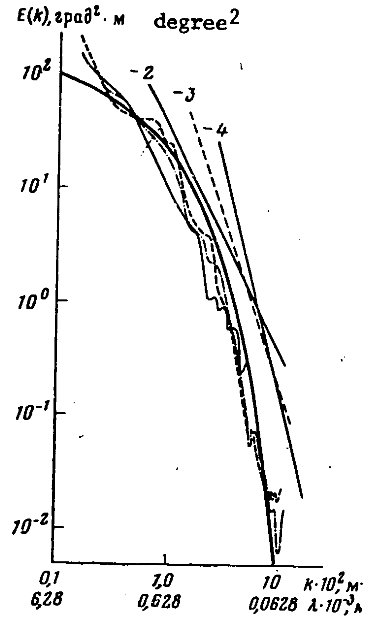

Fig. 2. Comparison of experimental evaluations of the spatial spectrum of internal waves [17] with model curve $\hat{S}(k)$. The straight lines represent the corresponding power-law dependences.

The schematic function $\hat{S}(k)$ is represented in Fig. 3. The spectral density $\hat{S}(k)$ has a maximum at the wave number $k_0 \sim \delta^{-1}$. For solitones with a large amplitude (lesser δ parameter) k_0 is shifted into the region of higher k. For solitones of a maximum amplitude, corresponding to $\delta = \delta_{min}$, we obtain

$$k_0 \sim \delta_{min}^{-1} \sim D^{-1}.$$

Thus, in the spectrum of fluctuations of vertical velocity measured in the field of algebraic solitones with a maximum amplitude it is necessary to expect a maximum with $k_0 \sim D^{-1}$, where D is the characteristic thickness of the layer with a sharp stratification.

We note in conclusion that the equation used in the study, which is now usually called the Benjamin-Ono equation, has recently attracted the attention of researchers. For example, quite recently a precise N-solitone solution [17, 18] was found for it, from which it follows that in the case of interaction of two solitones their phases do not change (in contrast to solitones of the Cortevega-de Vries equation). In addition, with the interaction of two solitones their heights in the interaction region decrease substantially. Thus, the assumption of a random distribution of phases made in this study is correct if there was such a distribution at some initial time. With respect to the contribution due to waves situated in the interaction region, it is evidently small. However, this problem requires investigation.

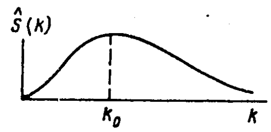

Fig. 3. Schematic form of spectrum of fluctuations of vertical velocity in the field of internal waves of a limiting amplitude.

The simple model of the spectrum constructed above can be considered the first step on the path of creation of models of internal waves as a solitone gas. There are natural means for its further complication and improvement: use of a precise N-solitone solution, allowance for interaction in the set of solitones and examination of a set of two-dimensional solitones.

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[38-5303]

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SCATTERING OF RADIATION INCIDENT ON A SECTOR OF THE SEA SURFACE. STATISTICAL APPROACH TO THE EVALUATION OF ACCURACY

Moscow IZVESTIYA AKADEMII NAUK SSSR: FIZIKA ATMOSFERY I OKEANA in Russian Vol 16, No 11, 1980 pp 1189-1197

[Article by M. Kh. Rafailov, manuscript submitted 3 Jan 80]

[Text]

Abstract: When using the Kolmogorov statistical test, taking into account the finite divergence of incident radiation, it is possible to ascertain the dependence between the area of the sector of the wave-covered sea surface scattering radiation and the statistical parameters of the spatial structures making up the surface. It is demonstrated that if the area of the sector of interaction of incident radiation with the surface for a given divergence of the incident radiation is greater than or equal to the definite area of the sector of surface stationarity given in the article there will be no distortions of the parameters of the scattered field in comparison with the scattering on a similar (from the statistical point of view) surface of infinitely great area. An evaluation of the accuracy of the parameters of radiation scattered by the wave-covered sea surface in a case when the area of the sector of interaction is less than the area of the sector of surface stationarity was obtained. Within the framework of correlation theory it was possible to derive expressions for the area of the sector of stationarity of the wave-covered sea surface for the individual spatial structures making up the surface. Numerical estimates of the area of the sector of stationarity for a number of cases are given.

At the present time there has been extensive development of theoretical methods for determining the parameters of the field of radiation scattered by the wave-covered sea surface. The basis of these methods is the use of statistical models of the wave-covered surface, by means of which, with allowance for the spatial structure of the incident radiation, it has been possible to determine the parameters of the scattered field [1-3]. There is also a great volume of experimental data on the scattering of the radiation incident on the wave-covered sea surface, taking into account interaction with different components of the spectrum of sea waves. These data are used extensively when sounding the sea surface for the purpose of determining wave parameters. It is assumed that the scattering surface is quite representative in a statistical sense, that is, in a sector of interaction between radiation and the surface all the scattering elements of the surface whose probability of existence is greater than zero participate in the scattering process. In other words, the scattering surface in the sense of its statistical representativeness is equated to the surface of an infinitely great area [4].

Under real conditions the interaction of radiation with the surface occurs in a sector of limited area. This sector is oriented in a definite way relative to the general direction of propagation of sea waves. In this case the representativeness of the statistics of scattering elements is limited and frequently is too small for the results to be correct for the entire surface, and not only for the considered sector.

All this in a number of cases results in substantial discrepancies between the theoretical and experimental results in an investigation of the radiation scattered by the sea surface and errors in determining the parameters of sea waves. Factors of this sort are particularly important when during sounding of the sea surface on the basis of the results of one or more experiments it is necessary to obtain reliable information on the state of the sea surface.

The problem of the relationship between the spatial structure of the radiation field and the change in wave parameters in a finite sector of the sea surface was examined in [5] from the point of view of hydrodynamic conditions for the development of sea waves, taking into account the bottom profile in shallow waters. However, the problem of changes in the structure of the scattered field in dependence on the area of the sector of interaction of radiation with the sea surface in the case of a real three-dimensional random structure of the sea surface when waves are present has not been solved.

The purpose of this study was a determination of the conditions under which the parameters of the radiation scattered on the wave-covered sea surface, with allowance for the properties of the radiation and the statistical characteristics of the surface, are not dependent on the position of the scattering sector at the surface. [Or the conditions of nondependence of the parameters of the scattered radiation on time if the surface moves relative to the interaction sector].

A result of the study was the derivation of formulas giving an evaluation of the accuracy of the parameters of the scattered field in the case of a definite surface area interacting with radiation. The inverse problem was also solved: the requirements on the area of the sea surface ensuring determination of the statistically stable characteristics of the scattered field were determined under conditions of a different state of the wave-covered sea surface.

1. Statistical Analysis of Scattering Conditions

We will examine the wave-covered sea surface, representing it in the form of several spatial structures of different scale superposed on one another. In a general case the scattering of radiation on such a surface is determined by the ratio of the radiation wavelength λ to the characteristic dimension l of a scattering nonuniformity of the surface. If $\lambda \ll l$, the principal contribution to scattering is from reflection in accordance with the laws of geometrical optics; however, if $\lambda \sim l$, the diffraction mechanism of radiation scattering predominates.

We will assume that both mechanisms of radiation scattering exist. Our task is to evaluate the influence of the finite dimensions of the sector of interaction of the incident radiation with the surface on the characteristics of the scattered field. For its solution it is necessary to examine the statistically least representative (in the sense of presence of scattering elements) structure of the surface components. A macroscale structure on which ripples, in turn, are already superposed is such a critical structure from the point of view of statistical representativeness. Accordingly, in our problem the mechanism of reflection from the surface in conformity to the laws of geometrical optics will be decisive.

In reflection the most important surface characteristic is the distribution of slopes. We will assume that a wave with the divergence $\Delta\alpha$ is incident on the surface. Then $\Delta\alpha$ will determine (Fig. 1) the ambiguity measure Q_0 in the value of the integral distribution function $F(\alpha)$

$$Q_0 = \int_{\alpha-\frac{\Delta\alpha}{2}}^{\alpha+\frac{\Delta\alpha}{2}} f(\alpha)\,d\alpha = F\left(\alpha+\frac{\Delta\alpha}{2}\right) - F\left(\alpha-\frac{\Delta\alpha}{2}\right), \quad (1)$$

where $f(\alpha)$ is the differential distribution function for surface slopes, regardless of azimuth.

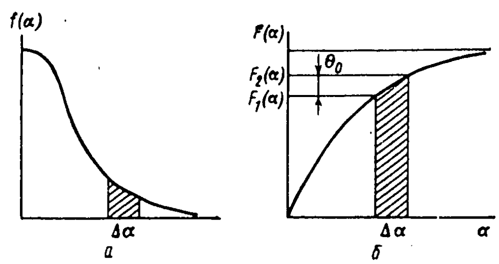

Fig. 1. Appearance of ambiguity in determining the integral distribution function with a finite angular divergence $\Delta\alpha$ of incident radiation: a) differential distribution law, b) integral distribution law.

For practical purposes the two determined distribution functions of slopes will not differ if with the angular divergence $\Delta\alpha$ their integral distribution functions differ from one another by a value less than Q_0. As the ambiguity measure it is possible to use the maximum possible value of the difference of two integral distribution functions, one of which -- $F_1(\alpha)$ -- corresponds to the integral distribution function for the entire general set of slopes, that is, for an infinitely great area of the scattering surface, whereas the second -- $F_2(\alpha)$ -- corresponds to the integral distribution function for the considered sample set of slopes, that is, the really irradiated sector of a finite area: $Q_0 = \sup \{|F_1(\alpha) - F_2(\alpha)|\}$.

Thus, it is possible to determine the volume of the sample set ensuring determination of distribution of slopes whose difference from the distribution for the general set for a stipulated slope α does not exceed (with a prestipulated probability) the ambiguity measure Q_0 for the integral distribution function of slopes.

The Q_0 value can be regarded as a parameter of the Kolmogorov statistical test of agreement between two integral distributions $F_1(\alpha)$ and $F_2(\alpha)$. Using this test, for the volume of the sample N_0 -- the number of the independent observations of surface slopes -- we obtain

$$N_0 = \nu^2(u) Q_0^{-2}, \tag{2}$$

where $\nu(u)$ is the argument of the Kolmogorov distribution, having the form [6]:

$$A(\nu) = \sum_{r=-\infty}^{\infty} (-1)^r \exp(-2r^2\nu^2)$$

with a definite significance level u corresponding to the condition $A(\nu) = 1 - u$.

We will consider the sea surface in the presence of waves as a random surface whose y-coordinates have a normal distribution with zero means. Since the surface is statistically symmetric relative to the mean level $f(\xi) = f(-\xi)$, the density of the maxima n_{max} for the surface is equal to the density of the minima and the total number of the stationary surface points is $n_{st} = 4n_{max}$.

A continuous gently sloping random surface can be represented as a set of plane microareas tangent to the surface; the distribution of slopes of these microareas coincides with the distribution of surface slopes. With such an approximation of the random surface the density of the microareas, having the slopes ξ_x, ξ_y, will be equal to

$$n(\xi_x, \xi_y) = 4n_{max} \exp\left[-\frac{\xi_x^2 m_{02} + \xi_y^2 m_{20}}{2m_{02}m_{20}}\right],$$

or, dividing and multiplying the right-hand side of this expression by the normalizing factor $2\pi\sqrt{m_{20}m_{02}}$, we obtain

$$n(\xi_x, \xi_y) = 4n_{max} 2\pi\sqrt{m_{20}m_{02}} f(\xi_x, \xi_y), \tag{3}$$

where m_{dq} is the initial moment of the spectrum of surface rises, of the orders d and q, and is equal to

$$m_{dq} = \int_0^\infty \int_{-\pi}^{\pi} W(k, \varphi) k^{d+q} \sin^d \varphi \cos^q \varphi \, dk \, d\varphi, \tag{4}$$

k is the wave number; φ is the azimuthal angle relative to the general direction of wave propagation; $W(k,\varphi)$ is the spectrum of surface rises; $f(\xi_x, \xi_y)$ is the differential distribution function of slopes.

The number of microareas whose slopes fall in the intervals from ξ_x to $\xi_x + \Delta\xi_x$ and from ξ_y to $\xi_y + \Delta\xi_y$, in a unit area of the random surface is

$$n'(\xi,\Delta\xi)=4n_{max}2\pi\sqrt{m_{20}m_{02}}\int\limits_{\xi}\int\limits_{\Delta\xi} f(\xi_x,\xi_y)\,d\xi_x\,d\xi_y. \tag{5}$$

In the surface area S_0 the total number of points of definite slopes is equal to

$$N(\xi,\ \Delta\xi)=n'(\xi,\ \Delta\xi)S_0. \tag{6}$$

The N value can also be determined from the total number of microareas on the random surface taken from the condition that their number is statistically representative for ensuring the stipulated degree of accuracy

$$N(\xi,\ \Delta\xi)=N_0F(\xi,\ \Delta\xi), \tag{7}$$

where N_0 is determined in accordance with formula (2); $F(\xi,\Delta\xi)$ is the probability of appearance of slopes in the intervals from ξ_x to $\xi_x+\Delta\xi_x$ and from ξ_y to $\xi_y+\Delta\xi_y$.

Taking expressions (2), (6), (7) into account, for the area of a sector of stationarity -- the sector of minimum area ensuring the obtaining of a distribution within the framework of a stipulated accuracy not dependent on the movement of the sector over the investigated surface -- we obtain

$$S_0=\frac{\nu^2}{Q_0^2 4n_{max}\,2\pi\sqrt{m_{02}m_{20}}}. \tag{8}$$

If the area of the surface sector with which the radiation interacts is less then S_0, due to a decrease in the volume of the sample set there will be a distortion of the determined experimental distribution in comparison with the distribution for a general set with these same ν, α, $\Delta\alpha$ values. When the quantity of reflectors in a sector of interaction of radiation with the surface is $\gg 1$ it can be assumed that the change in the experimental distribution is determined by the change in the sample dispersion m_2' with retention of the theoretical form of the distribution law.

The relationship between the change of the experimental integral distribution function $F(m_2')$ in comparison with the theoretical distribution $F(m_2)$ can be expressed through the ambiguity measure Q_e, that is $|F(\alpha,\ m_2)-F(\alpha,\ m_2')|=Q_e$. In this case Q_e is determined from formula (8) with the surface area S_i

$$Q_e=\nu(8\pi S_i n_{max})^{-1/2}(m_{02}m_{20})^{-1/4}.$$

Hence, knowing the form of the distribution law, we determine the value of the sample dispersion parameter from the equation $F(m_2')=F(m_2)+Q_e(S_i)$.

The value of the possible deviation of the experimental value from the theoretical value will be

$$\delta=\int\limits_{\alpha,\Delta\alpha}[f(\alpha,m_2')-f(\alpha,m_2)]\,d\alpha \tag{9}$$

with the confidence coefficient $A(\nu)$.

2. Correlation With Statistical Parameters of Wave-Covered Surface

A Gaussian random error has a law of distribution of slopes not dependent on azimuth in the form [7]

$$f(\alpha)=\frac{\alpha}{\sqrt{m_{02}m_{20}}}\exp\left[-\frac{\alpha^2(m_{20}+m_{02})}{4m_{20}m_{02}}\right]I\left[\frac{\alpha^2(m_{20}-m_{02})}{4m_{02}m_{20}}\right], \qquad (10)$$

where I(z) is a modified Bessel function of a zero order.

In examining the wave-covered sea surface it must be noted that the individual components of its structure have different statistical properties. Thus, there are isotropic or nearly isotropic structures [8]. There are also highly anisotropic structures consisting of waves having an approximately identical length and direction of propagation.

The energy in the spectrum of such structures is grouped around one wave number on the spectral plane, that is, the spectrum is narrow-band.

For isotropic structures $m_{02} = m_{20}$ and the distribution of slopes conforms to the Rayleigh law

$$f(\alpha)=\frac{\alpha}{m_2}\exp\left[-\frac{\alpha^2}{2m_2}\right]. \qquad (11)$$

If the structure has a narrow-band spectrum, that is $m_{20} \gg m_{02}$, the distribution of slopes has the form

$$f(\alpha)=\sqrt{\frac{2}{\pi m_{20}}}\exp\left[-\frac{\alpha^2}{4m_{02}}\right]. \qquad (12)$$

The density of the maxima of the statistically isotropic surface is equal to

$$n_{\max}=\frac{1}{6\pi\sqrt{3}}\frac{m_4}{m_2}, \qquad (13)$$

and for a surface having a narrow-band spectrum

$$n_{\max}=\gamma\bar{k}^2C(a), \qquad (14)$$

where $\gamma = \sqrt{m_{02}/m_{20}}$ is the index of three-dimensionality of the surface; $\bar{k} = m_1/m_0$ is the mean wave number around which the narrow-band spectrum is grouped; C(a) is the function of the parameter of spectral steepness [7].

With small $\Delta\alpha$ values and with a considerable slope α the uncertainty value is $Q \approx f(\alpha)\Delta\alpha$. In this case formula (8) for the area of the stationarity sector for a statistically isotropic structure is equal to

$$S_0=\frac{3\sqrt{3}\,v^2m_2\exp[\alpha^2/m_2]}{4\alpha^2m_4\Delta\alpha^2}, \qquad (15)$$

and for a structure with a narrow-band spectrum

$$S_0=\frac{v^2\exp[\alpha^2/2m_{02}]}{16k^2C(a)\Delta\alpha^2\gamma^2}, \qquad (16)$$

in this case for a structure with a narrow-band spectrum the stationarity sector has the form of an ellipse whose area is $S_0 = \pi r_1 r_2$, where r_1 and r_2 are the semiaxes of the ellipse, determined with the levels $\alpha/\sqrt{m_{02}}$ and $\alpha/\sqrt{m_{20}}$ respectively for the general direction of wave propagation and the direction perpendicular to it.

In order to determine the values of the initial moments of the spectrum and the characteristics of the sea surface related to them we will use the spectrum of rises of the wave-covered surface in the form $W(k) = W_t(k)$, $k_{t-1} < k < k_t$ corresponding to structures of different scales making up the wave surface [9]:

$$W_1(k)=4{,}05\cdot10^{-3}\exp\left[-\frac{0{,}74g^2}{V^4k^2}\right],\quad 0<k<k_1, \tag{17a}$$

$$W_2(k)=4{,}05\cdot10^{-3}k_1^{-1/2}k^{-5/2},\quad k_1<k<0{,}359, \tag{17b}$$

$$W_3(k)=4{,}05\cdot10^{-6}V_*k_3^{-p}k^{p-3},\quad 0{,}359<k<0{,}942, \tag{17c}$$

$$W_4(k)=4{,}05\cdot10^{-6}V^{5/4}k^{-3},\ 0{,}942<k<k_4, \tag{17d}$$

$$W_5(k)=3{,}37\cdot10^{-1}V_*^3k^{-8},\ k_4<k<\infty, \tag{17e}$$

where V_* is dynamic wind velocity, cm/sec; $p = 2.4 \lg (12V_*^{5/4})$; $k_1 = 51.7/V_*^2$, $k_4 = 2.1V_*^{1/8}$. The function of angular distribution of energy relative to the general direction for low-frequency structures is used in the form $\cos^n\varphi$, where n is a function of the wave number k [10]. The spectral width ε of the structures is evaluated in the form [7]

$$\varepsilon=\left(1-\frac{m_2^2}{m_0m_4}\right)^{1/2}. \tag{18}$$

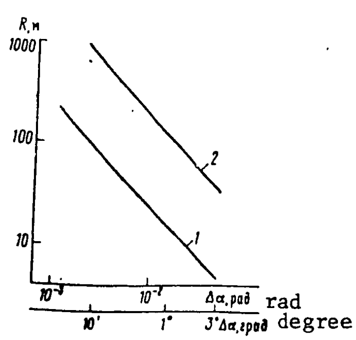

Fig. 2

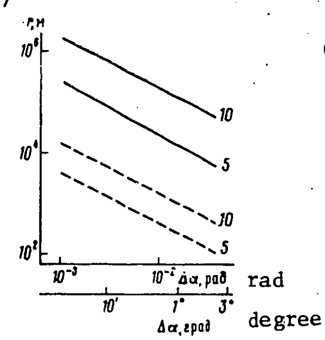

Fig. 3

Fig. 2. Radius of circle of area of sector of stationarity for gravitational range of waves: 1) for relative slope $\alpha/\sqrt{m_2} = 2$, 2) for relative slope $\alpha/\sqrt{m_2} = 3$.

Fig. 3. Semiaxes of ellipse of area of sector of stationarity for swell waves: parallel to general direction (dashed curve) and perpendicular to general direction (solid curve). The figures indicate wind velocity, m/sec; relative slope $\alpha/\sqrt{m_2} = 2$.

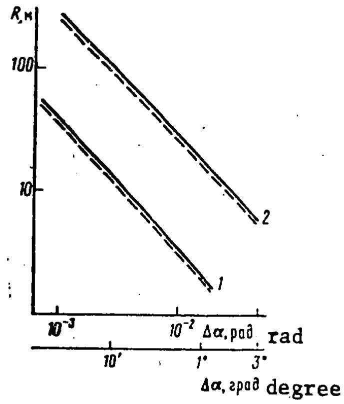

Fig. 4

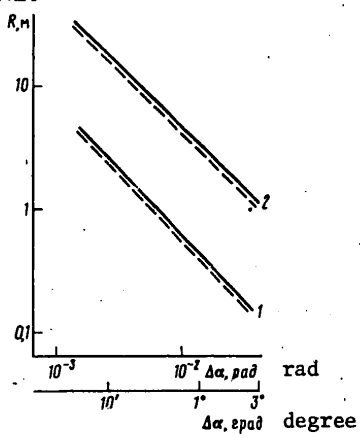

Fig. 5

Fig. 4. Radius of circle of area of sector of stationarity for gravitational-capillary range of waves (ripples): 1) for relative slope $\alpha/\sqrt{m_2} = 2$, 2) for relative slope $\alpha/\sqrt{m_2} = 3$, with wind velocity 5 m/sec (continuous curve) and 10 m/sec (broken curve).

Fig. 5. Radius of circle of area of sector of stationarity for capillary range of waves: 1) for relative slope $\alpha/\sqrt{m_2}$, 2) for relative slope $\alpha/\sqrt{m_2} = 3$, with wind velocity 5 m/sec (continuous curve) and 10 m/sec (continuous curve).

3. Analysis of Structures and Numerical Evaluations

The availability of experimental data on the fine structure of individual spectral components of sea waves makes it possible to carry out numerical evaluations for the most important spectral structures.

In the case of extinction of the high-frequency structures due to an increase in the viscosity of the surface water layer from oil and petroleum films [11] the statistical structure of the scattering surface becomes virtually isotropic ($\gamma >$ 0.9) with $\varepsilon = 0.96$, having a wide-band spectrum. In this case it is possible to neglect the spectral components described by expressions (17c)-(17e), limiting ourselves to an examination of purely gravitational waves and employing formula (15). Figure 2 shows the computed values for the radius of a circle in area corresponding to a sector of surface stationarity in the form of the angular resolution function $\Delta\alpha$ with a significance level u = 0.1 for two values of relative slopes $\alpha/\sqrt{m_2}$.

For the considered interval of waves the area of the sector of stationarity is invariant relative to wind velocity, the same as the width of the spectrum ε. This is attributable to the fact that with an increase in the steepness of the slopes with an increase in wind velocity there is also an increase in the curvature of the surface, that is, with a broadening of the distribution law for slopes there is a corresponding increase in the density of the reflecting elements per unit surface area.

With the presence of swell waves on the sea surface the surface spectrum is limited by the region of the main maximum of the gravitational interval of waves described by (17a). Determining the index of three-dimensionality of waves in the form [10] $\gamma^2 = (n+1)^{-1}$, for $\bar{k} = 0.85-0.22\ m^{-1}$ we obtain the value $\gamma^2 = 0.2-0.16$. The spectrum is narrow-band and we use formula (16) for the computations. The computed values for the semiaxes of the ellipse corresponding in area to the sector of stationarity are given in Fig. 3. It can be seen that the dimensions of the sector of surface stationarity are very highly dependent on the steepness of the swell waves.

The spectral interval (17c) is of great importance for a number of problems in radiophysics. Waves corresponding to this part of the spectrum are also widely used in scale physical modeling. As indicated in [8], the wave structure corresponding to this interval is statistically virtually isotropic. The values of circle radius R for the sector of stationarity for this structure, computed using formula (15) for two wind velocities, are given in Fig. 4. It can be seen that the dimensions of the sector of stationarity decrease insignificantly with an increase in wind velocity.

In the study of roughness of the water surface, and also in problems relating to the scattering of radiation in the radio range, an important role is played by the interval of capillary waves, described by expression (17d) for the spectrum. The structure of the capillary waves is anisotropic and the spectrum is wide-band ($\varepsilon > 0.9$). However, the angular dependences of energy distribution in the wave-covered surface have been poorly studied for this interval. But in evaluating the sector of stationarity this problem can be avoided, being guided by the values of the spectral moments most critical from the point of view of the area of the sector of stationarity. In this case the sector of stationarity can be stipulated in the form of a circle in which an ellipse corresponding to the sector of stationarity of a particular anisotropic surface surely will be inscribed, that is, the radius of the circle will correspond to the semimajor axis of the ellipse. The values of this radius, computed using (15) for two wind velocities, are given in Fig. 5. For this interval an increase in wind velocity leads to a more substantial decrease in the dimensions of the sector of surface stationarity.

An increase in wind velocity leads to an increase in the probability of appearance of large slopes, but at the same there is a more considerable increase in curvature of the surface. This dependence becomes more significant with an increase in the spatial frequency of the waves.

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TERRESTRIAL GEOPHYSICS

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RELATIONSHIP BETWEEN INHOMOGENEITIES IN THE UPPER MANTLE AND TECTONICS

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[Article by I. A. Rezanov, All-Union Scientific Research Institute of Nuclear Geology and Geophysics]

[Text] The study of structure of the earth's crust and the earth's upper mantle for the purpose of clarifying the patterns of distribution of mineral deposits is one of the most important problems in the earth sciences.

During the last decade considerable successes in study of the earth's mantle have been attained in our country. Mantle inhomogeneities are investigated by different geophysical methods characterized by different resolution and measurement accuracy. For example, attempts have been made to ascertain the extent and depth of density inhomogeneities in the upper mantle by means of an analysis of maps of gravity anomalies. However, using gravimetric data alone it is impossible to make a reliable determination of the depth at which these inhomogeneities lie, their vertical extent and the degree of decrease in density. More reliable information on structure of the upper mantle is obtained on the basis of seismic measurements.

The materials from study of the upper mantle by the deep seismic sounding (DSS) method, shot seismology and using waves from earthquakes have been generalized most completely by V. Z. Ryaboy [16, 17, 18]. L. P. Vinnik [4, 5] proposed the use of new methods for mapping horizontal inhomogeneities in the mantle using waves from earthquakes. All these materials, and also earlier studies, make it necessary to introduce substantial corrections into the existing concepts concerning structure of the upper mantle. For example, A. S. Alekseyev and V. Z. Ryaboy [1] note that the widely disseminated concept of the universal existence of a thick layer with a reduced velocity (asthenospheric channel) in the earth's upper mantle does not correspond to reality. Within the limits of tectonically quiet regions (platforms, ocean basins) layers with reduced velocities in the mantle are absent or have a very small thickness (10-40 km), as is confirmed by data on both body and surface waves.

A number of seismic profiles have now been published which were obtained by the DSS method and which clarify the structure of the upper mantle to a depth of 100 km. Several tens of seismic sections of the earth's upper mantle to a depth as great as 150-250 km below the Mohorovicic discontinuity are also known. Most of these sections indicate a change in the velocity of longitudinal waves (in our article in those cases when the section was prepared on the basis of transverse waves it is so stated). The value of all these materials for geology is that they reveal the structure of the mantle under specific folded zones and platforms.

We note that the constructed vertical sections of the mantle cannot always be tied in to an individual tectonic zone (geosynclinal downwarp, graben, anticlinorium, range), since in this case an averaged description of the mantle is given for a space with a width of 100-200 km. In such an area there are usually several tectonic zones with different structure. In clarifying the relationship between mantle inhomogeneities and tectonics we can compare available seismic sections only with the largest tectonic elements -- platforms, ocean plates, folded zones of different age, mountain zones of neotectonic activation, mid-oceanic ridges.

In analyzing the relationship between mantle inhomogeneities and large tectonic elements of different age and nature, we will attempt to answer the following questions: does the mantle differ under structures with different geological history; when did inhomogeneities develop in the mantle; how long are they retained in it; to what degree do these inhomogeneities correlate with crustal structure; how do they exert an influence on the direction and intensity of modern processes?

We will examine the structure of the mantle under lowland (platform) territories. Inhomogeneities in the uppermost layer of the mantle, extending for 2-4 km below the Mohorovicic discontinuity, can be judged from the boundary velocities along this surface. Maps of boundary velocities were compiled for the territories of the United States and the USSR. The velocity at the Mohorovicic discontinuity with mean values 8.1-8.2 km/sec varies from 7.5 to 8.5 km/sec. We note the circumstance that within the limits of the lowland-platform region the velocities at the Mohorovicic discontinuity vary in a broad range -- from 7.8 to 8.5 km/sec.

On the Baltic shield, to the north of the Gulf of Bothnia and in the region between Lake Onega and the White Sea, the velocities are high: 8.3-8.4 (in individual places 8.6) km/sec. But within the limits of this same shield, to the south of the Gulf of Bothnia, there is a region where the velocities are lower than average: 7.9-8 km/sec. The velocity on the Russian platform within the limits of the Pechorskaya syneclise is reduced (8 km/sec), and within the limits of the Moskovskaya syneclise -- increased (8.3-8.4 km/sec). Some regions of lowland Kazakhstan are characterized by very high velocities (8.4-8.6 km/sec), whereas others are characterized by reduced velocities (7.8-8 km/sec). A zone of reduced velocities at the Mohorovicic discontinuity is traced along the Urals in a southward direction, beyond its limits. Accordingly, it is not associated with the morphologically well-expressed Ural Range but is correlated with its ancient (Paleozoic) structure.

The fact that within the limits of structures with a close history of development in the Phanerozoic and with an identical nature of neotectonic movements the mantle velocities differ so greatly is evidence that these inhomogeneities are not directly associated with the last stages in the development of the mentioned structures, with neotectonic structures in particular. Inhomogeneities in the upper layer of the mantle are not manifested clearly in the gravity field. They are isostatically in equilibrium and must be regarded as relict structures, that is, those already existing for hundreds of millions of years.

The most detailed investigations of the upper mantle were made by the DSS method in the territory of the USSR. On the European platform (Moskovskaya syneclise) at a depth of 10–20 km below the Mohorovicic discontinuity there was found to be a thin (15-20 km) waveguide where the velocity was 0.2 km/sec lower than at its top [6]. In the northern part of the seismic profile (Pechorskaya syneclise) there is no waveguide. The same thin (10-12 km) waveguide was discovered by V. Z. Ryaboy under the Turanskaya platform to the north of the Kopetdag. It lies at a depth of 20 km below the Mohorovicic discontinuity; the velocity in it decreases by 0.2 km/sec. Beneath the Ural Mountains there is no waveguide in the mantle; at the Mohorovicic discontinuity the velocities are somewhat reduced, but then at a depth of 55-70 km they increase rapidly to 8.5-8.7 km/sec.

According to less detailed investigations (in a study of waves from earthquakes) a similar picture is also established under other lowland-platform territories. Layers with a small thickness (10-15 km) in the uppermost part of the mantle with a decrease in velocity by 0.3-0.5 km/sec were discovered on the Baltic shield, within the limits of the Barents Sea platform, in the western regions of the West Siberian platform. In many other regions of the European and North American platforms, and also the epi-Hercynian Ural-Siberian platform, there are no thick waveguides in the upper mantle. The most significant changes in mantle structure in the lowland territories are associated with the boundaries of major geological structures, such as the European (where the velocities are greater) and the Ural-Siberian (where they are somewhat reduced). Inhomogeneities in the upper 10-30 km in the mantle of lowland-platform regions correlate with major geological structures formed long before the neotectonic stage. Accordingly, they, like the already considered inhomogeneities at the Mohorovicic discontinuity, must be regarded as relict structures.

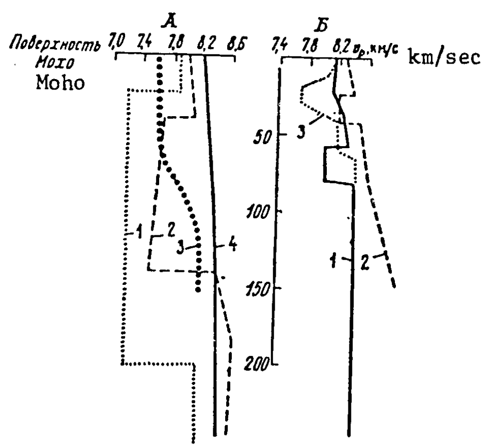

Fig. 1. Sections of structures with different tectonic development in the upper mantle. A) Pacific Ocean folded zone; 1) Cordillera Coastal Range [18]; 2) Rocky Mountains [18]; 3) Eastern Kamchatka [10, 21]; 4) example of mantle without waveguide -- one of sections of Canadian shield [18]; B) platform territories: 1) Canadian shield [18]; 2) Moscow syneclise of East European platform [6]; 3) Central French massif [32].

Now we will turn to the structure of the upper mantle of mountain zones. Within the limits of the Pacific Ocean zone a study was made of two of its sectors: in the North American Cordilleras and in the Kurile-Kamchatka zone. Now we will trace the changes in mantle structure from east to west under the North American continent (Fig. 1). Under the Canadian shield there is no thick waveguide in the mantle. The velocity at the Mohorovicic discontinuity on the average has a value 8.2 km/sec and in the depths of the mantle increases very smoothly, at a depth of 200 km attaining 8.4-8.5 km/sec.

Under the Colorado Plateau the velocity at the Mohorovicic discontinuity is reduced to 7.8 km/sec; this velocity persists to depths 100-120 km. Then the velocity begins to increase and at a depth of 200 km is approximately the same as under the Canadian shield. In the more westerly regions of the Cordilleras (Basin and Range Province) and under many other mountain regions the velocity changes with depth from 8 to 7.5-7.6 km/sec; then it increases and at a depth of 120 km becomes the same as under the Canadian shield.

Within the limits of the Coastal Range the mantle is different. On the two seismic sections of the mantle known for this zone near the Pacific Ocean coast there is a waveguide of great thickness (up to 180 km) with a marked decrease in velocity in it (up to 7.2 km/sec).

Thus, under the North American continent it is possible to discriminate three types of upper mantle:

a) a mantle without a waveguide or layer with reduced velocity of a small thickness (20-40 km) -- the lowland territories of the Canadian shield; b) regions with a waveguide of great thickness (more than 80 km) -- the Rocky Mountains, eastern and central Cordilleras; c) Coastal Ranges zone, where the waveguide in the mantle is expressed sharply and its thickness is enormous (180 km). The Cordilleras differ from the lowland territories situated to the east in having reduced velocities at the Mohorovicic discontinuity (less than 8 km/sec) and in that the mantle has layers with sharply reduced velocities. These peculiarities of mantle structure were registered under geological zones of different ages in the Cordilleras -- Paleozoic, Mesozoic and Cenozoic, which makes it possible to consider the detected mantle inhomogeneities to be newly formed, arising approximately simultaneously with the formation of mountain relief.

The second sector of the Pacific Ocean zone, where a detailed study was made of mantle structure, is the Kurile-Kamchatka zone. A crust of a continental type with a thickness of 30 km has developed under the Kamchatkan Peninsula. The velocities below the Mohorovicic discontinuity are reduced here. They are especially low in a zone with a width 50-90 km where modern Kamchatkan volcanoes are situated (7.3-7.6 km/sec).

In the deeper layers of the mantle the velocity of the longitudinal waves changes in the following way. Under Eastern Kamchatka the velocity in the mantle is first constant (7.6 km/sec), and then, at depths 60-70 km (30-40 km below the Mohorovicic discontinuity), increases gradually [10, 20]. As a result of low velocities at the Mohorovicic discontinuity the waveguide in the upper mantle is not clearly registered here. However, its existence in this part of the Pacific Ocean zone is evidenced by data on the structure of the mantle under Southern Sakhalin and the regions adjacent to it. The velocity in the mantle here decreases from 8 km/sec at the Mohorovicic discontinuity to 7.4-7.7 km/sec at a depth of 70 km, and then increases to 8-8.3 km/sec. At depths 150-200 km there is a second, very poorly expressed decrease in the velocity of seismic waves.

If the structure of the upper mantle under the Cordilleras of North America and Eastern Kamchatka is compared, the picture is similar: with a velocity at the Mohorovicic discontinuity of less than 7.8 km/sec the waveguide in the mantle is not manifested; if the velocity is greater it is manifested clearly. Under Eastern Kamchatka the velocities at the depth of the Mohorovicic are still lower than under the Cordilleras and the waveguide in the mantle is situated at lesser depths. In both cases this is attributable to the unusually high volcanic activity in the eastern part of the Kamchatkan Peninsula. It is characteristic that in the Asiatic part of the zone there is not such a thick waveguide as under the Coastal Range of the Cordilleras.

Within the limits of the western part of the Mediterranean-Himalayan zone the structure of the mantle has been studied under most mountain ranges. A waveguide was discovered under the Pyrenees, at depths of 120-200 km. Under the Alps it was possible to discriminate two layers with reduced velocities -- in the ranges 700-100 and 130-180 km. In the mountains of Southeastern Europe, and also in the Eastern Carpathians (Vrancha Mountains) a zone of reduced velocities is associated with depths of 100-150 km. In the Dinaric Alps the top of the waveguide rises to a depth of as much as 85 km. In the Pamir-Hindukush region there is a waveguide at depths of 80-200 km. In the eastern part of the zone, under the Himalayas, there is a waveguide which is registered in the range of depths 100-150 km. Thus, it can be concluded that the structure of the mantle in the Mediterranean-Himalaya zone in general is similar to the structure of the mantle in the Pacific Ocean zone.

A detailed study was made of a segment of the Afro-Asiatic mountain zone from the Pamirs to the Baykal region. Along the Pamir-Baykal profile there are two zones with reduced velocities at depths 100-200 km [14]. One of them is situated under the Pamirs and is gradually wedged out in the direction of the Tien Shan. The second begins under the Sayan, but it is expressed most sharply in the region of Lake Baykal, where the velocity is reduced to 7.9 km/sec. In the northeastern part of the considered zone V. A. Rogozhina [15] defined a body which is complex in configuration and which is characterized by velocities of longitudinal waves reduced by 0.3 km/sec. An anomalous zone in the mantle is situated under the Altay, Eastern and Western Sayans, Baykal rift zone and in the high-mountain zones of Northern Mongolia, that is, in the limits of the entire Afro-Asiatic mountain zone. In the vertical section the zone of decreased velocity is a lens with a thickness up to 200 km or more.

If one compares data from investigations of the mantle along the profile Pamirs-Baykal with materials published by V. A. Rogozhina it is easy to see that to depths 200 km the results of these investigations in general coincide, and specifically, from the Baykal mountain region to the Sayans at depths of 100-200 km below the Mohorovicic discontinuity there is a zone of reduced velocities. According to V. A. Rogozhina, the sharp plunging of the anomalous layer to depths of 300-600 km below the Sayans is not reflected on the Pamirs-Baykal profile.

However, it must be remembered that V. A. Rogozhina deals with a relative decrease in velocity, whereas the true values are shown on the Pamirs-Baykal profile.

All three described mountain zones are characterized by one and the same mantle characteristics, namely, that in the upper 100-200 km below the Mohorovicic discontinuity there are layers with reduced velocities in the form of lenses with a thickness of 50-100 km. Their horizontal dimensions, judging from the Afro-Asiatic zone, attain 500-1,000 km. The two considered mountain zones -- Pacific Ocean and Mediterranean-Himalayan -- are ancient; they were formed in the Riphean and developed as geosynclinal systems in the Phanerozoic. The Afro-Asiatic mountain zone was formed in the Cenozoic. The similarity of structure of the upper mantle under mountain zones with a different prehistory indicates that these mantle inhomogeneities were formed recently and are associated with the formation of mountain relief, not with their ancient history.

In a comparison with tectonics it is of special interest to use seismic methods, making it possible to obtain the areal distribution of mantle inhomogeneities. L. P. Vinnik [4, 5] developed a method for the mapping of horizontal inhomogeneities of the top of the mantle by an analysis of the travel time of a seismic ray between seismic stations. A merit of this method is that in its use it is possible to register inhomogeneities in an area; the minimum width of the discriminated zones is about 50 km. This makes possible a horizontal comparison of zones of increased and reduced velocities at the top of the mantle with a small-scale tectonic scheme.

Particularly interesting results were obtained with respect to the Caucasus isthmus. It was found that the strikes of the velocity inhomogeneities in the upper layer of the mantle do not coincide with the direction of the geological structures and the location of the mountain ranges. Zones with increased velocities in the mantle have a northwesterly orientation under the Caucasus (Fig. 2). It is easiest to trace clearly the zone with reduced velocities passing from the Stavropol'skoye Plateau to the Dzirul'skiy complex and then to the Lesser Caucasus through the cities of Bakuriani and Yerevan. Parallel to it (to the east of Tbilisi and Kirovabad) there is a zone of increased velocities. Farther to the east, parallel to the shores of the Caspian Sea, there is still another zone of low velocities.

The general Caucasian strikes of geological structures were formed partially in the Baykalian geotectonic stage, but they were finally formed in the Hercynian stage, that is, 250 million years ago, and persisted stably during subsequent history [21]. The noncoincidence of zones of equal velocities in the mantle with the Hercynian-Alpine plan is evidence that inhomogeneities in the upper mantle reflect a very ancient structural plan existing in the Caucasus prior to formation of downwarps and uplifts with a general Caucasian strike. The structures with a northwesterly strike developed on the Caucasus isthmus in the Proterozoic [11]. To some degree they even now "show through" the medium of younger structures. For example, the Trans-Caucasian uplift has such a strike; this strike coincides with the strike of the low-velocity zone at the top of the mantle.

Now we will proceed to the Pamirs-Hindukush region (Fig. 3). In the western part of the Pamirs and Hindukush the mantle is high-velocity, whereas under the Eastern Pamirs, Karakorum and Himalayas there are reduced velocities at the top of the mantle. The eastern boundary of the high-velocity mantle block passes in a meridional direction, not taking orography and neotectonics into account. It cuts the unified high-mountain country -- the Pamirs-Hindukush mountain node -- into two parts. Accordingly, both in this region and in the Caucasus velocity inhomogeneities in the mantle are unrelated to mountain formation.

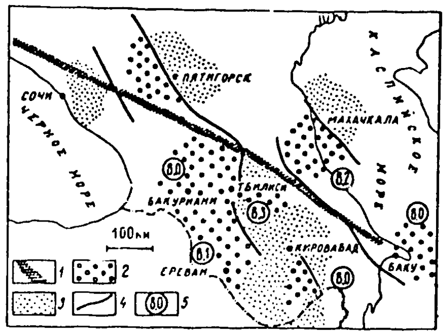

Fig. 2. Horizontal inhomogeneities in the upper mantle under the Caucasus (according to L. P. Vinnik). 1) watershed of the Greater Caucasus Range; 2) zone of reduced velocities in the upper mantle; 3) zone of increased velocities in the upper mantle; 4) boundaries of zones; 5) velocities at Mohorovicic discontinuity according to N. A. Belyayevskiy and V. Z. Ryaboy.

Comparison of the pattern of mantle inhomogeneities obtained by L. P. Vinnik with geological and tectonic maps indicates that the high-velocity block in the upper mantle virtually completely coincides with the system of Precambrian median masses of the Hindukush and Western Pamirs. The Nuristanskiy (Afghanistan) and Badakhshan-Vakhanskiy (for the most part in the southwestern Pamirs) median masses fall within the limits of the high-velocity mantle [13]. The first consists for the most part of Lower- Middle-Proterozoic gneisses, schists and quartzites. The second was formed of Archean, Lower- and Middle-Proterozoic and Riphean gneisses, migmatites, granite-gneisses, amphibolites, schists and marbles. Low-velocity zones were discovered where thick strata of Paleozoic-Mesozoic deposits have developed.

Thus, tectonic zones, being somewhat uplifted over a prolonged time (over the course of 500 million years), are characterized by greater velocities in the mantle than the territories which during this period experienced considerable subsidence. Mantle inhomogeneities, detected by the L. P. Vinnik method in the Pamir-Hindukush region, are ancient, already arising in the Paleozoic.

Now we will turn to the third region -- the Northern Tien Shan. It was discovered that beneath the depressions (Iliyskaya, Issykkul'skaya) the upper mantle is characterized by increased velocities, and under the high-mountain ranges separating them (Terskey Alatau and Kungey Alatau) the mantle is low-velocity. In this case there will seemingly be observed a relationship between the structure of the mantle and the mountain relief -- under the rising ranges the velocities in the mantle are lower. However, such a conclusion [4] was drawn without taking into account the ancient history of the region.

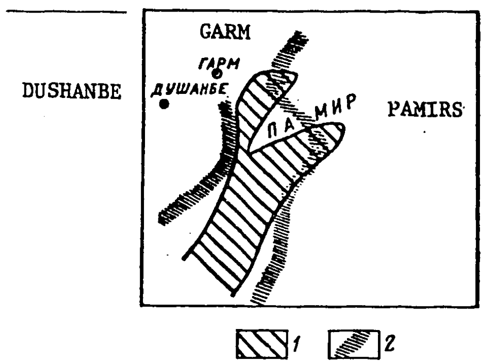

Fig. 3. Horizontal inhomogeneities in the upper mantle under the Pamirs according to L. P. Vinnik. 1) areas occupied by Precambrian median masses; 2) contours of zone of increased velocities.

A paleotectonic analysis indicates that there were somewhat uplifted blocks (of the median masses type) at the site of the Iliyskaya and Issykkul'skaya depressions in the Upper Proterozoic and Lower Paleozoic, and at the site of the Terskey Alatau and Kungey Alatau there were deep geosynclinal downwarps [19]. Taking this into account, we will draw the conclusion that for the Tien Shan, as for the Pamirs, there is a relationship between zones of ancient uplifts and sectors of increased velocities in the mantle and zones of downwarps with sectors of reduced velocities. Mountain formation in the Tien Shan, as in many other regions, was controlled by the ancient structural plan: in the orogenic stage the median masses subsided and became depressions (Iliyskiy, Issykkul'skiy), and ranges were formed at the site of Paleozoic downwarps. The modern location of the Tien Shan ranges and depressions (like the zones of low and high velocities in the mantle) was governed by the ancient structural plan.

Thus, we draw the conclusion that in all three considered regions horizontal inhomogeneities at the top of the mantle, detected by L. P. Vinnik, are not associated with recent (Neogene-Quaternary) processes, but reflect the ancient structural plan, that is, are relict processes. These inhomogeneities formed during the period of genesis of Proterozoic and Paleozoic uplifts and downwarps. The great velocities under the uplifts are possibly attributable to the fact that denser (high-velocity) rocks were uplifted there. On the other hand, the lesser velocities registered under the Paleozoic downwarps can be attributed to the fact that the high-velocity rocks are relatively subsided. If such a conclusion is correct, it therefore follows that the L. P. Vinnik method makes it possible to register the horizontal distribution of inhomogeneities in the first tens of kilometers below the Mohorovicic discontinuity. Only with such an assumption can there be an explanation of such a clear spatial correlation between the ancient geological structures visible on the earth's surface and the inhomogeneities at the top of the mantle preserved from those times.

The next important means for investigating lateral inhomogeneities in the mantle is measurement of the absorption of seismic waves using seismograms of remote earthquakes [4]. This method is valuable in that it makes it possible to form some idea concerning the properties and processes in the upper mantle under those regions where there are no seismic stations.

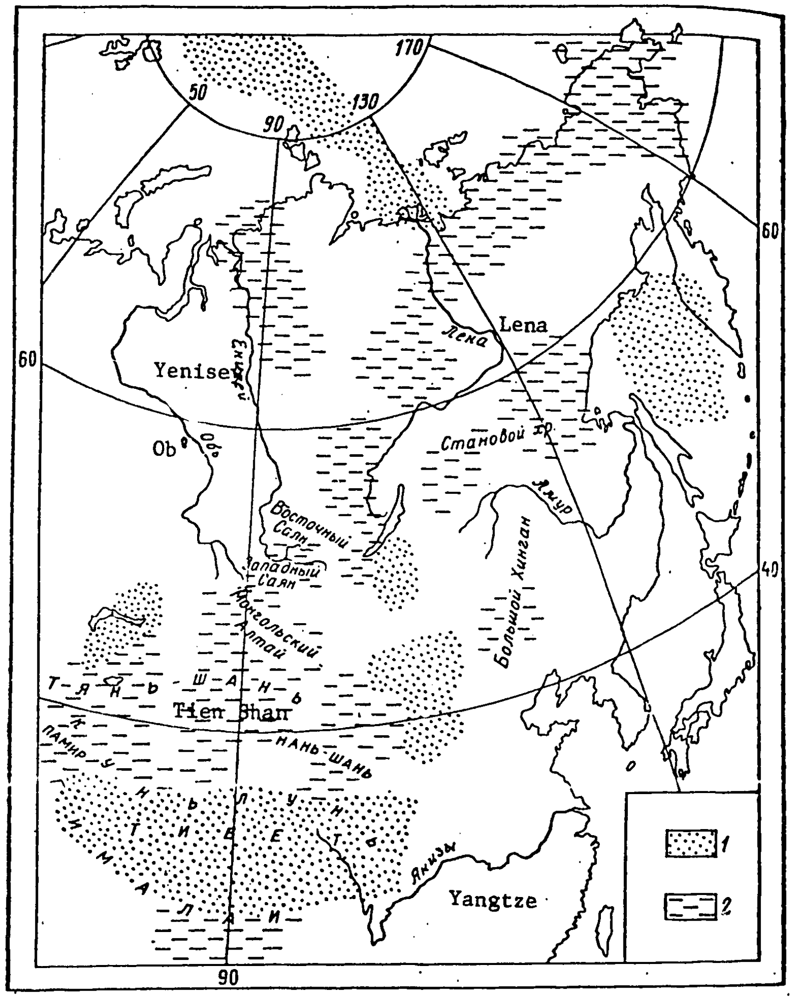

Fig. 4. Variations in absorption of seismic waves in the earth's mantle in the territory of Asia according to L. P. Vinnik and A. A. Godzikovskaya. 1) regions of increased absorption; 2) regions of reduced absorption.

Figure 4 shows the nature of absorption of elastic waves in the mantle in Asia. Over the greater part of Asia the upper mantle has a low absorption. Against this background it is possible to discriminate anomalous zones with high absorption -- Tibet, Tien Shan, Inner Mongolia, Southern Trans-Baykalia, Gakkel' Range, Laptev Sea, Sea of Okhotsk. The largest and most clearly expressed anomaly with high absorption in the mantle is situated under Tibet. Spatially it takes in the

Tibetan Plateau, Himalayas and central sector of the Kunlun. There are no anomalies to the north and south of Tibet under depressions and lowland territories. Thus, an anomaly of increased absorption coincides with a region of intensive mountain formation. The conclusion can be drawn in the example of Tibet that regions of intensive mountain formation are situated where the mantle is characterized by an increase in the absorption of elastic waves. This conclusion is confirmed by the discovery of increased absorption in the mantle under the Northern Tien Shan. However, the mantle has weak absorption beneath the Pamirs and under the Sayan mountain region.

In the example of the southern part of Siberia we will compare a zone of anomalies of velocity in the mantle (according to V. A. Rogozhina) with an absorption anomaly. The coincidence of a zone of high absorptions and a zone of low velocities occurs where the latter is situated at depths of 0-100 km below the Mohorovicic discontinuity. However, in places where a low-velocity mantle is discovered at depths of 100-200 km or more there is no increased absorption. If it is assumed that the L. P. Vinnik method makes it possible to register increased absorption only in the upper 100 km of the mantle, the absence of such a zone under the Sayans, Mongolian Altay, Stanovoye Plateau is regular, since the layer of the anomalous mantle, according to V. A. Rogozhina, is situated deeper.

Evidently, zones of increased absorption mark places where the anomalous mantle approaches the bottom of the crust. This also explains the presence of zones of high absorption of seismic waves under the Sea of Okhotsk where, judging from the heat flow, the anomalously heated layer is situated directly beneath the crust. The discovery of a zone of increased absorption beneath the Northern Tien Shan is also becoming understandable: there the upper surface of the mantle of reduced density, according to data obtained along the profile Pamirs-Baykal, has also approached the surface. The absence of a zone with high absorption beneath the Pamirs for the time being remains unexplained.

It is proposed that the concept "waveguide capacity" be introduced for a quantitative evaluation making possible a simultaneous determination of the thickness of the layer and the decrease in velocity. By this term is understood the product of the decrease in velocity and the volume within which the velocity is reduced. The dimensionality of "waveguide capacity" is volume multiplied by velocity (km^4/sec). If this value is determined in a plane (along the seismic profile), then it is used in km^3/sec, if only along a vertical line -- in km^2/sec.

Now we will compare "waveguide capacity" along verticals (see Table 1). The waveguide capacity under mountain zones is 10 times or more greater than the capacity of waveguides under platforms. However, if we take the Coastal Range of California, here the waveguide capacity is almost 50 times greater. This once again confirms that in this case we are dealing with phenomena of a different scale, and probably of a different nature.

A comparison of inhomogeneities in the upper mantle detected by seismic methods with the tectonic structure of the earth's surface leads to the conclusion that there are two groups of inhomogeneities -- relict and newly formed. In the upper layer of the mantle, whose thickness is 10-30 km, virtually everywhere there are relict inhomogeneities. Their horizontal dimensions are a few hundred and their vertical dimensions are a few tens of kilometers. These inhomogeneities are registered by three independent methods: 1) change in velocity along the Mohorovicic discontinuity; 2) layers and lenses of reduced velocities along extended profiles registered by deep seismic sounding; 3) variations in the travel time of waves between stations as registered by the L. P. Vinnik method.

Relict inhomogeneities are detected more reliably within the limits of platform regions where they are not suppressed by thicker newly formed inhomogeneities which are superposed on them. However, there is no doubt but that relict inhomogeneities exist in the mantle and under mountain zones. In addition to the maps prepared by L. P. Vinnik, a proof of this is the absence of any clear patterns between mountain relief and boundary velocity. For example, under the high-mountain part of the Caucasus there were both increased velocities (8.2-8.4 km/sec) and low velocities (8 km/sec). In the eastern part of the Tien Shan there is a predominance of velocities below the norm (7.6-7.8 km/sec), and in the high-mountain regions of the Western Tien Shan -- average velocities (8.1-8.2 km/sec).

Anomalously high (8.3-8.4 km/sec) velocities at the Mohorovicic discontinuity were registered under the Fergana depression. However, it would be hasty to relate this causally to its formation. It is more probable that in this case the high values are attributable to the ancient structural plan -- the presence of an ancient median mass under the depression. We note that not all the intermontane depressions are characterized by increased velocity at the Mohorovicic discontinuity. In the Kura depression, separating the Greater and Lesser Caucasus, the velocities are low (8 km/sec).

In the mantle there are newly formed inhomogeneities directly associated with modern tectonic processes. These are thick (80-200 km) layers with reduced velocities under all mountain-folded zones. The thickness of such waveguides, their number and extent, vary within the limits of the mountain zone. In the example of the Sayan-Baykal region it can be seen that such inhomogeneities coincide in area with territories having mountain relief, which makes it possible to relate them causally. Additional information on newly formed mantle inhomogeneities is given by the outlining of zones of increased absorption which in many cases are registered in places where thick layers of reduced velocities approach the Mohorovicic discontinuity. Newly formed inhomogeneities are evidently absent under lowland-platform territories.

Thus, there are two types of inhomogeneities of different origin in the upper mantle. The first (relict) were formed long ago, in the process of geotectonic development of the lithosphere, and coincide with the most ancient structural plan. Their existence must be attributed to differences in the mineral composition of the rocks making up the mantle. The second (newly formed) are associated with relatively recent processes transpiring uner the earth's mountainous zones. They are correlated with the three most active processes in the life of the planet: 1) volcanism; 2) mountain formation; 3) formation of deep depressions with a thinned crust. Their formation is more logically attributed to the state of mantle matter, most probably its partial fusion.

Now we will turn to structure of the mantle beneath the oceans. It has been studied to a considerably lesser extent than under the continents. Within the limits of the island arcs and abyssal trenches the structure of the upper mantle is the same as on the margin of the Pacific Ocean mountain zone (the zone of the Coastal

Ranges in the Cordilleras, Eastern Kamchatka). As indicated by body waves in the marginal parts of the Pacific Ocean the zone of reduced velocities in the upper mantle frequently begins directly from the bottom of the earth's crust and is traced to a depth of not less than 150-200 km.

Table 1

Dimensions and Capacity of Waveguides in Upper Mantle Under Different Tectonic Zones

	Investigated region	Thickness, km	Decrease in velocity, km/sec	Capacity, km^2/sec
Platform territories	Baltic shield, Voronezhskiy complex, eastern part of Turanskaya platform, Chu-Sarysuyskaya depression, Scythian platform, Balkhashskaya depression	0	--	0
	Baltic shield	10	0.4	4
	Moscow syneclise	20	0.2	4
	Cis-Ural downwarp	10	0.4	4
	Northwestern Kazakhstan	10	0.4	4
	Turanskaya platform (central part)	10-15	0.2-0.3	3
	Northern and Central Europe	20-30	0.2-0.3	5-8
Mountain-folded zone	Pamirs	120	0.3	36
	Cis-Baykalia	180	0.3	54
	Eastern Kamchatka	80	0.4	32
	Cordilleras (inner regions)	80	0.5	40
	Coastal Range of Cordilleras	180	0.8	144

The boundary velocities at the Mohorovicic discontinuity under the oceans vary in a still greater range than under the continents (from 7.2-7.6 to 8.6-8.8 km/sec). This is indicative of the presence of considerable horizontal inhomogeneities, but their spatial position is by no means always detected. It can only be noted that mid-oceanic ridges are characterized most frequently by reduced velocities, whereas under ocean basins the velocity is increased.

S. M. Zverev, for the northwestern part of the Pacific Ocean, on the basis of DSS data, determined the nature of change in velocity in the upper mantle. It was found that the mantle of this ocean basin is characterized by higher velocities than on the continents. High velocities in the mantle in the western part of the Pacific Ocean, at least to depths of 250-300 km, are indicated by seismological data [24]. In the depth range up to 300 km the mantle velocities gradually increase from 8.2-8.3 to 8.8 km/sec. At a depth greater than 100 km there is a poorly expressed waveguide with a thickness not greater than 40-45 km. However, the velocities in this waveguide are anomalously high (8.4-8.6 km/sec), which in no way compares with waveguides in the mantle of folded mountain zones. We should state that under this part of the ocean the waveguide in the mantle is, in the usual understanding of the word, absent (Fig. 5,A).

Velocity models of the upper mantle under the oceans have been constructed in limited numbers on the basis of surface waves. Since earthquakes in the oceans are known only within the limits of the mid-oceanic ridges, the resulting sections illustrate the structure of the mantle beneath them (Fig. 5,B). An analysis of transverse waves revealed a layer of reduced velocity, the depth at which it occurs, velocity parameters; its thickness varies greatly. Within the limits of the East Pacific Ocean Rise (near California) the velocity of the transverse waves in this channel is reduced from 4.4-4.8 to 3.5-4.1 km/sec [29]. It was discovered that the thickness of the waveguide gradually decreases (due to plunging of its top) from the axis of the East Pacific Ocean Rise to its periphery and a low-velocity layer is no longer registered beyond the limits of the rise [30]. The mentioned study is still another evidence that the structure of the upper mantle of mid-oceanic ridges and oceanic basins is fundamentally different.

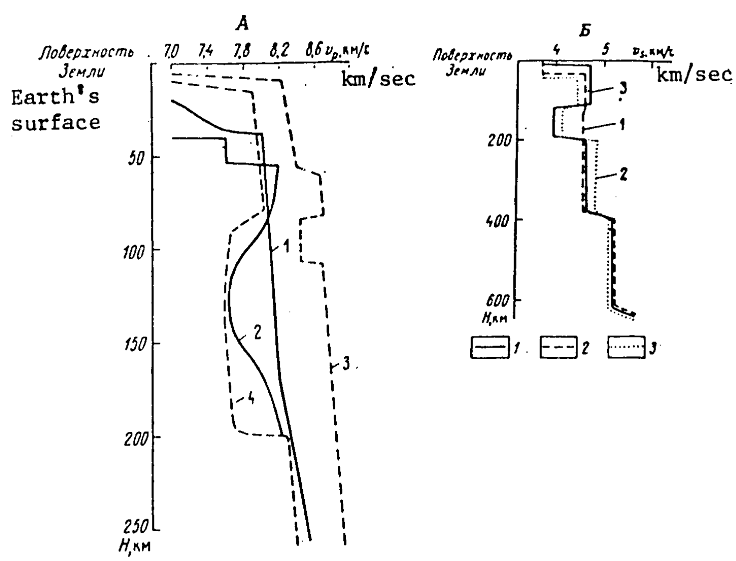

Fig. 5. Seismic sections of crust and upper mantle of continents and oceans. A) sections constructed on the basis of longitudinal waves (solid curve -- continents, broken curve -- oceans): 1) East European platform, according to V. Z. Ryaboy, 2) Baykal mountain region, according to S. V. Krylov, B. P. Mishen'kin, P. N. Puzyrev, 3) Western part of Pacific Ocean [25]; 4) Sea of Japan (compiled by V. Z. Ryaboy using data in the literature); B) sections constructed on the basis of transverse waves (constructed using data on surface waves [27]); 1) continental shields, 2) tectonically active regions of continents, 3) oceans (mid-oceanic ridges).

Under the mid-oceanic ridges of the oceans the mantle is constructed approximately the same as under folded zones affected by mountain-forming movements. It is possible to speak of a definite similarity between the upper mantle under platform territories and oceanic plates. In the mantle there are no thick layers with reduced velocities under either of these. Under oceanic plates in the upper mantle, judging from limited data, the velocities are even higher than under platforms.

The widespread idea that there are two types of upper mantle -- the mantle of the continents and the mantle of the oceans -- is contradicted by new data. In all probability the picture is the reverse -- continental platforms and oceanic basins have a similar mantle structure (there are no thick waveguides).

If ancient (relict) mantle inhomogeneities are left to one side, we can list the following three principal models of the upper mantle associated with definite types of geological structures.

1. The mantle of ancient platforms on the continents and in oceanic basins -- layers of reduced velocity are absent or thin.

2. The mantle of orogenic zones of the continents and mid-oceanic ridges in the oceans -- thick layers of reduced velocity. On the continents they developed during a period of mountain formation; accordingly, their lifetime is a few tens of millions of years.

3. Mantle of geosynclines. Judging from modern geosynclines (island arcs), a mantle of this type constitutes extended steeply dipping layers of a higher-velocity matter extending to a depth of 200-700 km and accompanied by earthquakes. The lifetime of these inhomogeneities, taking into account the duration of the geosynclinal stages, is measured in the hundreds of millions of years.

In nature there can be a spatial matching of different types of mantle. For example, within the limits of the transition zone from the Asiatic continent to the Pacific Ocean the mantle inhomogeneities characteristic of orogenic zones and geosynclines were superposed on one another. This, in our opinion, explains the complexity of deep structure in the mentioned region.

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