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Abstract

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GEOPHYSICS

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ON THE POSSIBLE DECOMPACTION OF THE MATERIAL OF THE UPPER MANTLE

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In the classical Bullen model "A" and in a number of other works, the density distribution in the upper mantle is represented by a general section, averaged for oceanic and continental territories ($d\rho/dz = 8.65 \cdot 10^{-4}$ g/cm$^3 \cdot$ km). A characteristic feature of this model and similar ones is the absence of negative values of $d\rho/dz$. In the latest Bullen-Haddon model ([11]), with a density gradient $+2.64 \cdot 10^{-4}$ g/cm$^3 \cdot$km, which is consistent with the free oscillations of the Earth, preference is given to Gutenberg' s velocity distribution ([13]), but nevertheless a negative density gradient in the upper mantle is not admitted. Nor was any reflection found in it of the difference in the distribution of velocities beneath oceans and continental structures. Until very recently it was believed that density sections are not sensitive to changes in velocities ([10]). E. V. Artyushkov ([1]) and M. S. KrasS ([5,6]), studying the rheological properties of the Earth' s crust and upper mantle in connection with the postglacial uplifts of Fennoscandia and the Canadian Shield, proposed a three-layer model for the Earth' s tectonosphere. The upper layer (lithosphere), extending to a depth of 80-100 km, has substantially increased values of effective viscosity $\eta = 10^{22}$—10^{23} poise ([6]) and $\eta = 10^{25}$ poise ([16]). The middle layer of reduced viscosity (asthenosphere), with $\eta = 1$—$5 \cdot 10^{20}$ poise, confined to the seismic waveguide, rests on a rigid foundation (layer C of the upper mantle). The effective viscosity of the latter is estimated in different ways: from 10^{23} poise ([16]) to 10^{26}—10^{28} poise ([5]). While holding different points of view on the nature and physical state of the asthenospheric layer, both researchers do not allow decompression and the existence in the upper mantle of lateral density inhomogeneities. In work ([16], p. 106) it is even emphasized that the density in the asthenosphere increases with depth more rapidly than according to the adiabatic law. One cannot agree with E. V. Artyushkov and M. S. KrasS regarding the passive role of the asthenosphere in magmatic, seismological, tectonic, and other processes that transform the lithosphere and the Earth' s surface of our planet.

In recent foreign publications ([8a—b,9,12,14,16]) the question is discussed, from various points of view, of the possible existence in the upper mantle of a decompression zone that would correspond to the seismic waveguide established from seismological data.

Clark and Ringwood ([12]) showed that density sections calculated by them to a depth of 400 km, taking into account the influence of temperature and pressure, irrespective of the petrological composition of the upper mantle (pyrolite, eclogite) and the structural position of the region, contain decompression zones. The depth, thickness, and degree of expression of these zones depend on the temperature conditions, composition, and structure of the Earth' s interior.

Density ρ and its vertical gradient are usually considered functions of independent variables–pressure P and temperature T, or pressure and entropy S, which in turn are continuous functions of depth z. For variation with depth in a chemically homogeneous region of the Earth one may write

$$\frac{d\rho}{dz} = \left(\frac{\partial \rho}{\partial P}\right)_S \frac{dP}{dz} + \left(\frac{\partial \rho}{\partial S}\right)_P \frac{dS}{dz}. \tag{1}$$

After simple transformations, expression (1) takes the form

$$d\rho/dz = g\rho/\Phi - d_P \rho\tau; \tag{2}$$

τ is the superadiabatic temperature gradient; α_P is the coefficient of thermal expansion at constant pressure; g is the acceleration of gravity; Φ is the elastic parameter; $\Phi = \rho/K_S = V_p^2 - {}^4/_3 V_s^2$; K_S, K_T are the adiabatic and isothermal incompressibilities; V_p, V_s are the velocities of longitudinal and transverse waves.

However, with this method of expressing $\rho(P, S)$, it is practically impossible to take into account the influence of the temperature term with τ on the character of the distribution of densities in the tectonosphere region under study (to a depth of 40 km), where the temperature gradient differs from the adiabatic one.

Birch ([9]), considering the possibility of the occurrence of a zone of decompaction of material in the upper mantle, proposed proceeding from the existing dependence between density and velocity. The simple relationship between V_p, V_s, and ρ indicates that discontinuities in the quantities V_p, V_s, dV_p/dz, and dV_s/dz should apparently be accompanied by discontinuities in ρ and $d\rho/dz$. Proceeding from velocity models, it is recognized that the curve $\rho(z)$ is continuous for the mantle, while the gradient $d\rho/dz$ undergoes a discontinuity only at the boundary between zones B and C ([2]).

Anderson ([8B]) and Thomson ([16]), to explain the density minimum in the upper mantle, used Debye theory and the Grüneisen law, assuming that for a homogeneous* mantle with a positive value of the coefficient of thermal expansion the density gradient has the same sign as the gradient of the velocity of transverse waves. This follows from the general expression for the Grüneisen coefficient

$$\gamma = \left(\frac{\partial \ln \theta}{\partial \ln \rho}\right)_T = {}^1/_3 + \left(\frac{\partial \ln V_s}{\partial \ln \rho}\right)_T - {}^1/_3 \left\{\frac{\partial \ln[2 + (V_s/V_p)^3]}{\partial \ln \rho}\right\}_T, \qquad (3)$$

and since the 3rd term in the given equation is, as a rule, small,

$$\gamma \approx {}^1/_3 + (\partial \ln V_s / \partial \ln \rho)_T, \qquad (3')$$

where θ is the Debye temperature; γ is the Grüneisen coefficient. For the Earth' s interior $\gamma > {}^1/_3$ ([9,16]). Consequently, $(\partial \ln V_s / \partial \ln \rho)_T > 0$. Thus, the negative gradient dV_s/dz, established from seismological data at depths of 70–200 km, must be accompanied by a negative density gradient.

Anderson ([8B]), believing that density in a solid is a function of the quantity V_s, at the same time erroneously neglects the thermal effect, which is significant in the upper mantle because of the large geothermal gradient. To estimate the magnitude of $d\rho/dz$ in the low-velocity zone of the upper mantle beneath shields and oceanic areas, Thomson ([16]) proposed an expression of another form, which also takes into account Debye theory and Gutenberg' s ([13]) velocity distribution:

$$\frac{d\rho}{dz} = \left[\rho \frac{dV_s}{dz} - \frac{\rho g}{\Phi}(1 + T\alpha_P\gamma)V_s\left(\frac{\alpha_P \Phi}{c_p} - {}^1/_3\right)\right]_{\alpha_P}\left(-\frac{\partial T}{\partial V_s}\right)_P + \frac{\rho g}{\Phi}(1 + T\alpha_P\gamma), \qquad (4)$$

where c_p is the heat capacity at constant pressure.

In our opinion, equation (4) is to a significant extent symbolic in character, since it contains a large number of elastic and thermodynamic parameters which, even if they have been measured, were measured only on rock samples under laboratory conditions at low pressures and tempera–

* The Grüneisen law is also applicable to the model of an inhomogeneous (pyrolitic) upper mantle, for which the mean atomic weight remains practically constant ([8B]).

temperatures not characteristic of the Earth' s interior. Of course, there is no reason to think that the upper mantle includes precisely those rocks that have been studied in laboratories. Thus, in applying expression (4) to estimate the magnitude of $d\rho/dz$, one has to assume that the mantle material must have properties varying within the same limits as those established experimentally from rock samples. Thomson, solving equation (4), obtained broad limits for variation of $d\rho/dz$ in the upper mantle, both for oceanic regions (from -9.3 to $+2.3 \cdot 10^{-4}$ g/cm$^3 \cdot$ km) and for continental regions (from -2.6 to $+5.7 \cdot 10^{-4}$ g/cm$^3 \cdot$ km). These results can hardly be used to explain differences in the deep structure of the Earth' s planetary structures.

Using the functional dependence of ρ on V_s and T and the velocity models of the upper mantle proposed in works ([8a], [13]), we find the critical values of the geothermal gradient in the tectonosphere of the oceanic and continental sectors of the Earth at which negative density gradients are possible. The general superposition equation for the vertical density gradient may be written as follows:

$$\frac{d\rho}{dz} = \left(\frac{\partial\rho}{\partial V_s}\right)_T \frac{dV_s}{dz} + \left(\frac{\partial\rho}{\partial T}\right)_P \frac{dT}{dz}. \tag{5}$$

Using the thermodynamic identities

$$(\partial\rho/\partial V_s)_T = (\partial\rho/\partial P)_T(\partial P/\partial V_s)_T, \tag{6}$$

$$(\partial\rho/\partial T)_{V_s} = (-\partial V_s/\partial T)_\rho(\partial\rho/\partial V_s)_T \tag{7}$$

and replacing the isothermal coefficient $(\partial\rho/\partial P)_T$ by the parameter Φ and the adiabatic exponent $\left[\left(\frac{\partial\rho}{\partial P}\right)_T = \frac{1}{\Phi}\frac{c_p}{c_v}\right]$, we obtain the final expression for the vertical density gradient in the form

$$\frac{d\rho}{dz} = \frac{1}{\Phi}\frac{c_p}{c_v}\left(\frac{\partial P}{\partial V_s}\right)_T \left[\frac{dV_s}{dz} + \left(-\frac{\partial V_s}{\partial T}\right)_\rho \frac{dT}{dz}\right]. \tag{8}$$

The critical value of the geothermal gradient is determined from the general expression (8). If we assume that $d\rho/dz = 0$ at $\rho = \max$, then, after simple transformations, we find

$$\frac{dT}{dz} = \frac{dV_s}{dz} \Big/ \left(\frac{\partial V_s}{\partial T}\right)_\rho. \tag{9}$$

The value of the temperature derivative of velocity can be obtained from laboratory measurements performed at high pressures and temperatures. Thus, for example, Zoss ([15]), who investigated the properties of oxides at temperatures from 0 to 1200°, obtained for MgO $(\partial V_s/\partial T)_\rho = -5.1 \div -6.9 \cdot 10^{-4}$ km/sec · deg. For 4 dunite samples $(\partial V_s/\partial T)_\rho = -5.7 \cdot 10^{-4}$ km/sec · deg ([8r], p. 113). We have chosen $(\partial V_s/\partial T)_\rho = -5.7 \cdot 10^{-4}$ km/sec · deg, corresponding to dunites and close to the mean value of this parameter for MgO. In calculating the critical value of the geothermal gradient for oceanic regions and the continental sector, different values of the vertical velocity gradient dV_s/dz were adopted in accordance with existing velocity models for the upper mantle: for oceanic regions $dV_s/dz = -4.9 \cdot 10^{-3}$ sec^{-1} (model CIT–11A ([8a])), and for the continental sector $dV_s/dz = -2.2 \cdot 10^{-3}$ sec^{-1} ([13]). The critical geothermal gradient for continents is equal to 3.85 deg/km, and for oceanic regions 8.60 deg/km. The lower boundary of the region with $d\rho/dz < 0$ is determined from the obvious relation

$$|dV_s/dz| \leq |(\partial V_s/\partial T)_\rho dT/dz| \tag{10}$$

and lies in the transition zone of the velocity-gradient from negative values to zero.

Within the oceanic sector of the Earth, at a higher value of the critical gradient (8.65 deg/km), loosening of the material of the upper mantle apparently begins at substantially shallower depths (80–110 km) than beneath the continents.

The results of seismological and electromagnetic investigations indicate that beneath shields and other stable structures of the continents the region of thermodynamic instability and the associated loosening of the material are located at depths of 200–300 km ([7]).

Table 1

Physicomechanical parameters; type of geostructure	Φ, (km/sec)^2	$\frac{dV_s}{dz} \cdot 10^{+3}$, sec^{-1}	$\frac{c_p}{c_v}$	$\frac{dT}{dz}$, deg/km	$\left(\frac{dP}{dV_s}\right)_T \cdot 10^{-5}$, bar/km·sec (olivine)	$\left(\frac{dV_s}{dT}\right)_\rho \cdot 10^{+4}$, km/sec·deg	$\frac{d\rho}{dz} \cdot 10^{+8}$, g/cm^3·km
Oceanic sector of the Earth	36.6 (8a)	-4.9 (8a)	1.2 ([4])	0.75 ([4])	2.00 ([3])	-5.7 (8 , 15)	-3.0
Continental sector	38.4 ([13])	-2.2 ([13])	1.2 ([4])	0.75 ([4])	2.00 ([3])	-5.7 (8 , 15)	-1.10

Let us quantitatively estimate the magnitude of the negative density gradient in the tectonosphere for oceanic regions and the continental sector, taking as the limit for the geothermal gradient the value for an adiabatic temperature distribution, 0.75 deg/km ([4]). For this purpose we shall use equation (8), the velocity models of the upper mantle proposed in (8a, [13]), and some results of experimental measurements.

The calculated values of $d\rho/dz$ and all the physicomechanical characteristics used in the calculations, with indication of the bibliographic source, are summarized in Table 1.

The numerical values obtained for the negative density gradient characterize the degree of loosening of the material of the upper mantle beneath oceanic regions and stable continental structures. The processes of loosening beneath the oceans proceed considerably more intensely than beneath shields. This should lead to more contrasting changes of the Bouguer anomalies within the former and to a smooth change of the gravitational field over the territory of ancient platforms and shields.

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CITED LITERATURE

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