ISSN 1069-3513, Izvestiya, Physics of the Solid Earth, 2010, Vol. 46, No. 11, pp. 966–973. © Pleiades Publishing, Ltd., 2010.
Original Russian Text © V.N. Senachin, A.A. Baranov, 2010, published in Fizika Zemli, 2010, No. 11, pp. 61–68.

Estimation of the Deep Density Distribution in the Lithosphere of Central and Southern Asia using Data about Free Mantle Surface Depth

V. N. Senachin[a] and A. A. Baranov[b]

[a] *Institute of Marine Geology and Geophysics, Far East Division, Russian Academy of Sciences, ul. Nauki 2, Yuzhno-Sakhalinsk, 693002 Russia*
e-mail: geodyn@imgg.ru

[b] *Schmidt Institute of Physics of the Earth, Russian Academy of Sciences, Bol'shaya Gruzinskaya ul. 10, Moscow, 123995 Russia*
e-mail: baranov@ifz.ru

Received October 12, 2009

Abstract—The results of the selection of a model of the deep density distribution in the lithosphere of Central and Southern Asia, which explains the previously revealed dependence of the free mantle surface depth on the thickness of the crust [Artemjev, 1975], are described. It is shown that this dependence can be caused by variations in the mantle's density with depth. Models of the continental and oceanic mantles with an increase in the linear density over depth are selected for the region of Asia. The level of the free surface depth in the oceanic mantle is higher than in the continental mantle. The observed dependence on the crustal thickness can also be used for determining nonlinear density variations with depth under the assumption that lateral density variations in this dependence are of a random character.

DOI: 10.1134/S1069351310110066

INTRODUCTION

As is well-known, density is one of the most important parameters of the Earth: all processes proceeding inside and at the surface of our planet are in one way or another related to the nonequilibrium distribution of masses. However, this parameter cannot be determined directly. Density can be measured only in the upper part of the Earth's crust with the use of samples from drilled boreholes. In deeper parts of the Earth, density is determined from velocities of seismic waves, and on the whole, on our planet, the density distribution is controlled by well-known values of the mass and moment of inertia of the Earth.

At the same time, the intense development of geological–geophysical and geochemical methods over recent decades has made it possible to form a rather clear insight into the inner structure of the Earth, its physical state, and the material compositions of its shells. Therefore, our interrelated notions about the petrological composition and the main physical parameters inside the Earth (density, pressure, temperature) should be refined.

This paper presents the results of the work on estimating the possibilities for determining the deep density distribution in the lithospheric mantle on the basis of the observed dependence of the free mantle surface depth on the crust thickness [Artemjev, 1975]. In this case, the well-known processes of the redistribution of substance inside the Earth, one of which is the isostasy, should not be neglected.

DETERMINATION OF THE FREE MANTLE SURFACE: RELATION TO THE CRUSTAL THICKNESS AND PROBABLE CAUSES

The isostatic state of the upper shell of the earth has been studied rather well. Thus, it is known that on the major part of the surface of our planet, where there is no tectonic activity, the lithosphere is isostatically compensated and, accordingly, is not subject to any vertical motions.

The free mantle surface (FMS) depth is one of the characteristics of the isostatic state of the Earth's lithosphere. This parameter shows the crustal uplift or lowering with respect to its normal state necessary for the isostatic leveling of the lithosphere with the density-homogeneous mantle [Artemjev, 1975; Artemjev and Belousov, 1983]. However, an anomalous uplift or lowering of the FMS level does not mean that that the isostatic position of the lithospheric block is upset; most frequently, such deviations point to an anomalous consolidation or deconsolidation inside the lithosphere [Senachin, 2006]. This fact allows us to use FMS data for studying the density distribution in the mantle.

The FMS depth is calculated by the formula:

$$H_{\mathrm{FMS}} = H_m - \frac{1}{\rho_m}\sum_{1}^{n} m_i \rho_i, \qquad (1)$$

where H_{FMS} is the FMS depth; H_m is the Moho boundary depth; ρ_m is the mantle density; ρ_i and m_i are, respectively, the density and thickness of the ith layer of the crust at the calculated point; and n is the number of layers of the crust.

As is seen from this formula, the given characteristic is one-dimensional, and in the case of using area-averaged data, it does not show any substantial local disturbances of the isostasy. In addition, it is impossible to determine from calculated FMS anomalies whether an anomaly is caused by a disturbance of the isostatic state or by the presence of density heterogeneities in the mantle. It is necessary to use gravity field data or other additional information for such determinations. Errors in density distributions for crustal layers also introduce a considerable error in the determination of the FMS depth. Density is calculated from seismic wave velocities data using the well-known velocity–density dependences [Barton, 1986]. These dependences have an averaged character, and therefore, they are not capable of taking into account differentiated contributions of each factor affecting density (temperature, material composition, etc.).

Nevertheless, the study of FMS depth anomalies and regular features of their distributions over continents and oceans made it possible to obtain a number of interesting results [Artemjev, 1975]. Thus, it was revealed that the FMS depth increases with the increasing of crustal thickness. This could be related to the difference between the real mantle density and that used in the FMS calculations; however, according to the estimation presented in [Artemjev, 1975], for eliminating this dependence, it is necessary to reduce the mantle density to 3.0 g/cm^3, i.e., practically to the density of the lower crust. Such a state of the crust means a more complicated system of isostatic compensation than is incorporated into the Airy model, namely, the isostatic compensation is partially caused by density heterogeneities in the mantle and increasing of

these heterogeneities directly connected with crustal thickening. This conclusion follows from the FMS definition in accordance with formula (1). However, this formula does not take into account the influence of deep (radial) density variations, which can take place in the Earth's lithosphere. As it will be shown below, mantle density variations with depth change the FMS depth in crustal blocks with different thicknesses. Such dependence makes it possible to determine the deep density distribution from the observed FMS dependence on the crustal thickness. However, for this determination, it is necessary to separate the influences of lateral and radial density heterogeneities, which is very difficult. That is why, as the first step on the way to the solution of this problem, we estimated the possible influence of deep density variations in the mantle on the FMS depth with different crustal thicknesses.

The effect of dynamic topography produced by mantle flows serves as a correction to the FMS depth in the real Earth. The FMS is uplifted over ascending mantle flows and lowered over descending ones. The dynamic topography is studied in [Anderson, 2006; Ricard et al., 2006]. According to the estimation obtained by these authors, topographic variations attain hundreds of meters with maximums in the zones of mid-ocean ridges and continental rifts (+500 m) and minimums over descending mantle flows (–500 m). However, during the averaging, the addition of this effect to the FMS calculated by formula (1) virtually did not change either the FMS or the density distributions in the mantle obtained below.

Density models of isostatic compensation were calculated on the basis of the data provided by the AsCrust-08 crustal model [Baranov, 2010]. This regional model for Central and Southern Asia contains more exact data about the crustal structure than the well-known CRUST 2.0 crustal model [Bassin et al., 2000]. In addition, this model is quite representative in respect of the crustal thickness and age, i.e., the region covered by this model contains the crust of the entire age and thickness ranges met on the Earth.

THE ASCRUST-08 MODEL

The AsCrust-0.8 crustal model of Central and Southern Asia [Baranov, 2010] covers tectonically complex regions of the Earth's surface, where the collision of the Indian and Eurasian plates, which resulted in the formation of the Plateau of Tibet (the largest continental uplift on the Earth), plays the key role.

New seismic data obtained in recent years has provided the basis for constructing a much more detailed crustal model that yields seismic velocity distribution in crustal layers and can be applied in gravity modeling and other applications. In the AsCrust-08 model, particular attention was given to the regions of Arabia, China, India, and Indochina. Multiple diverse data were checked for their mutual correlation, and the most reliable of these data were used for constructing a unified model of the entire region.

This refined numerical model of the consolidated crust includes the Moho boundary depth, thicknesses of the crustal layers, and the Vp velocity distributions in these layers. A great amount of new seismic data (reflected, refracted, and surface waves from earthquakes and explosions) were analyzed during the construction of this model. All these data were integrated into a unified model with the resolution $1° \times 1°$. The results were presented in the form of ten numerical maps determining depths to the Moho boundary, the thicknesses of the upper, middle, and lower crust, as well as the densities and Vp velocities in these layers.

The new Moho map is shown on Fig. 1. The black line indicates the boundaries of the region, within which new seismic data were analyzed. In oceans, the minimum depth to the Moho is 7 km, whereas its maximum (75 km) is attained under Tibet. The thickness of the standard continental crust is usually on the order of 38–44 km. The Arabian Peninsula, Small Asia, northern and central India, eastern China, and Indochina are regions with a normal continental crust.

The South China Sea, Sunda Sea, Sea of Japan, and other marginal seas of Southeastern Asia have three mixed types of the crust: oceanic (7–9 km), stretched continental (10–38 km), and mixed.

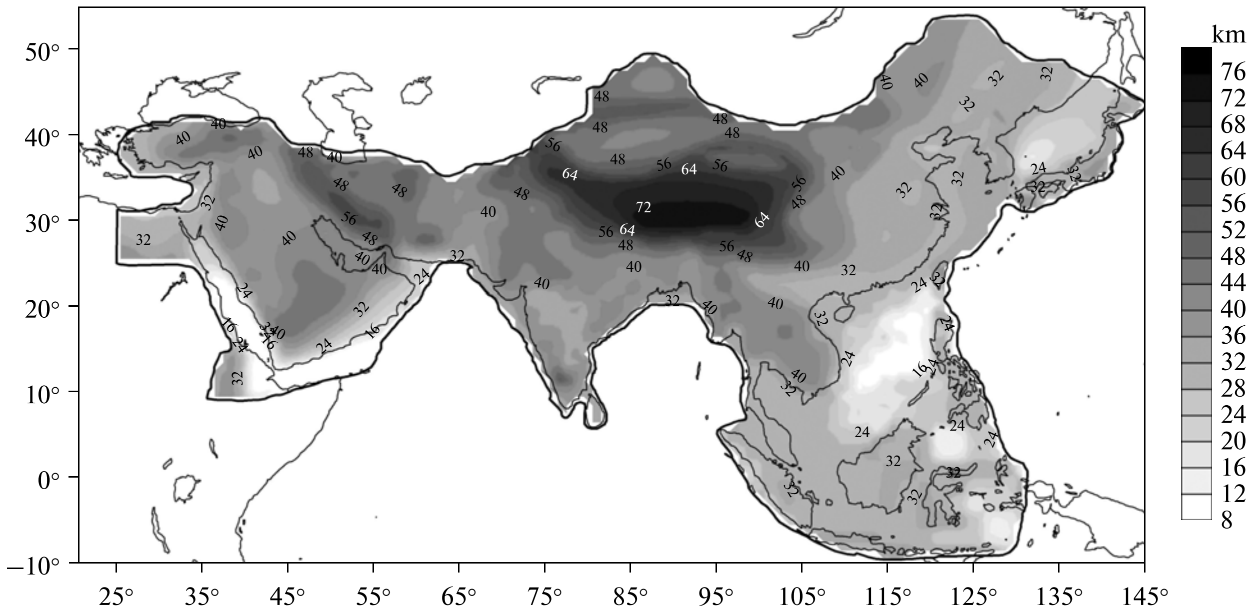

Fig. 1. Map showing the depth to the Moho boundary for Central and Southern Asia according to the AsCrust-08 model. The black line marks the boundaries of the region. Gray color shades indicate the Moho depth in accordance with the right-hand scale.

MANTLE FREE SURFCE OF CENTRAL AND SOUTHERN ASIA

The distributions of FMS depth in the Asian region calculated from data of the AsCrust-08 model and from data of the CRUST 2.0 model [Bassin et al., 2000] outside the region covered by the new model are presented in Fig. 2.

The FMS depth was calculated by formula (1). There are seven layers in our model: a water layer (wherever it is present), three layers of sediments from the model described in [Laske and Masters, 1997], and three layers of the crust from the AsCrust-08 model. All data for these layers were taken from numerical models with the resolutions $1° \times 1°$ for sediments and the crust, and $0.1° \times 0.1°$ for the water layer (bathymetry).

According to the calculations performed, the FMS depth in Central and Southern Asia (see Fig. 2) varies within a considerable range (from 2 to 7 km), which is explained by the presence of recent tectonic activity in the Alpine–Himalayan fold belt and the development of rifts in the northeastern edge of Africa. The largest FMS uplift is observed in the Red Sea, the Gulf of Aden, and the adjoining them northern part of the East-African rift valley. The largest FMS depth is encountered in the east of the Tien Shan Mountains. The Himalayas are characterized by a narrow zone with an increased FMS level within 4–5.5 km. Parallel to this zone, the zone with an increased FMS depth (up to 6 km) is traceable in the south. The latter corresponds to the boundary of the thrust zone of Asia (Main Boundary Thrust [Mishra and Rajesh, 2003]) on the Indian plate. The Plateau of Tibet in the major part of its area is characterized by FMS depths from 4.5 to 5.5 km, and only at the boundary with the Tarim basin, a narrow zone with FMS uplift up to 3 km is observed. In the east, the FMS deepens to 6.5 km. The normal FMS depth in the Tarim basin is within 4.5–5 km.

On the whole, it can be noted that the FMS depth decreases in rifting zones with the thinned crust and increases in mountainous structures of the Asian region, where the crustal thickness attains 75 km. Accordingly, it can be suggested that the relation of the FMS depth to the crustal thickness, which was revealed by M.E. Artemjev and his colleagues, is caused by the conditions of the present-day tectonic development. At the same time, in Tibet, which is a mountainous structure with the thickest crust on the planet, and which is located, in the opinion of some researchers, in an isostatically compensated state [Fang et al., 1997], the FMS depth shows no tendency to increase with an increase of the crustal thickness (see Fig. 2).

THE MANTLE MODEL WITH DENSITY VARIATIONS

Suppose that the crust thickness is the same everywhere, and the mantle density varies in accordance with the following law:

$$\rho_m(h) = \rho_0 + \alpha h, \qquad (2)$$

where ρ_0 is the mantle density at the FMS level, h is the depth from the FMS level, and α is the coefficient of mantle density variations with depth. In this case, the equilibrium of load and compensation masses, which are separated by the FMS level, in the mantle with density variations over depth (see Fig. 2) can be presented in the following way:

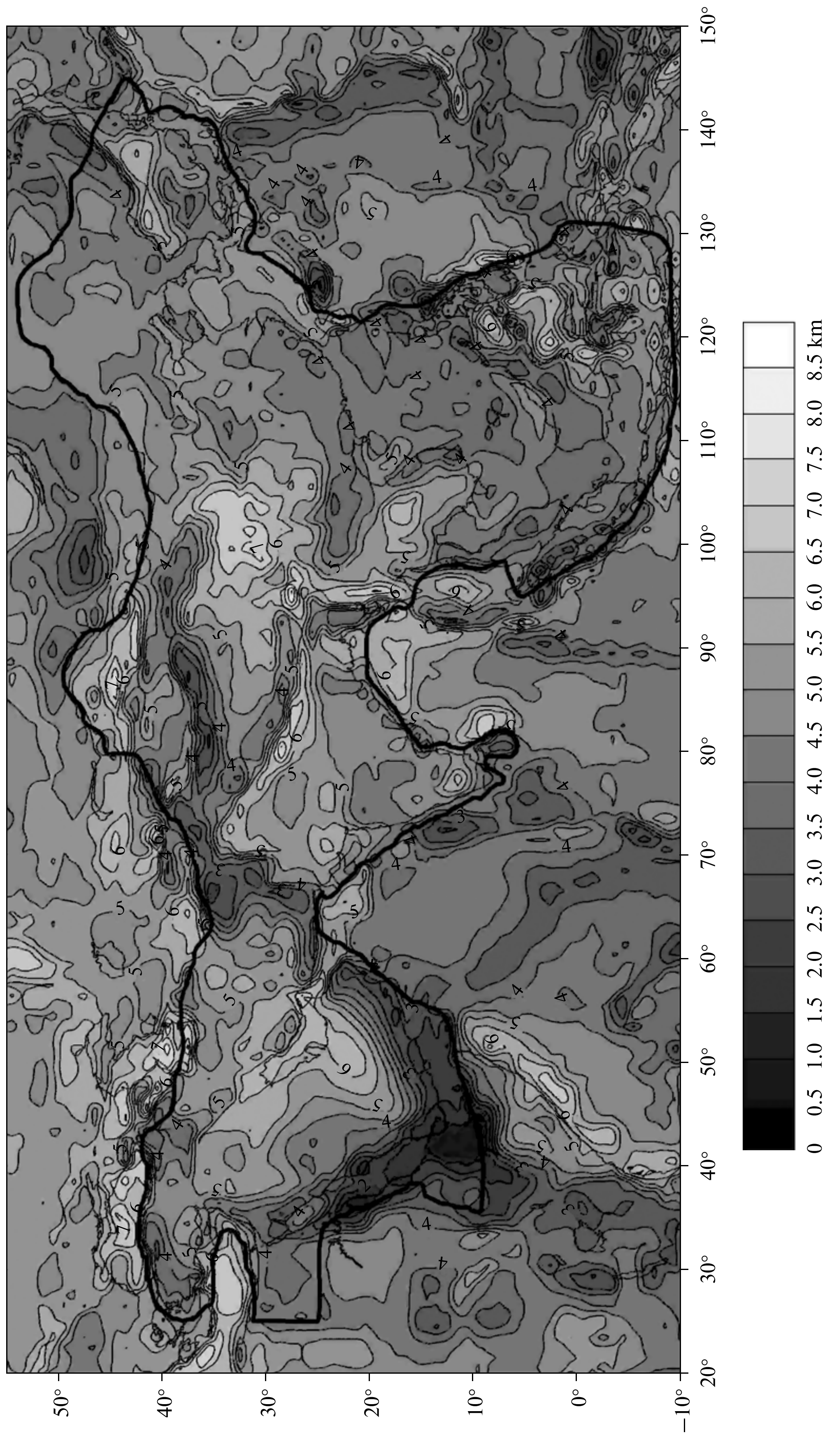

Fig. 2. Depths of the free mantle surface in Central and Southern Asia. The thick black line marks the boundaries of the region covered by the AsCrust-08 model; outside this region, FMS depth were calculated with the use of the CRUST 2.0 model [Bassin et al., 2000].

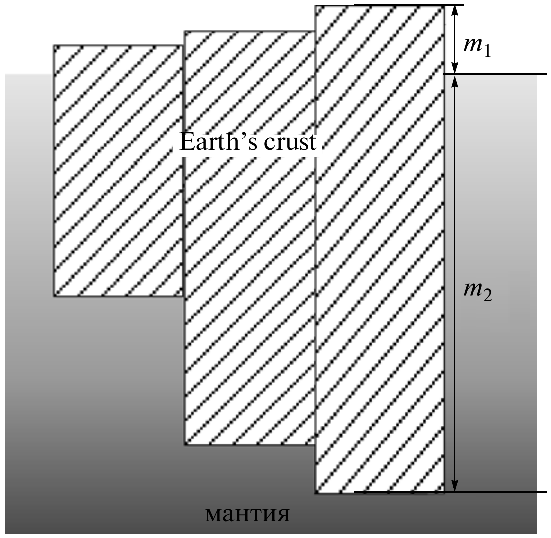

Fig. 3. Illustration to the derivation of the formula for calculating the FMS depth in the mantle with the density gradient.

$$m_1\rho_k = m_2\left(\rho_0 + a\frac{m_2}{2} - \rho_k\right), \tag{3}$$

where m_1 is the thickness of the load layer (the upper part of the crust to the FMS level); m_2 is the thickness of the compensation layer (the lower part of the crust from the FMS level) (see Fig. 3); ρ_0 is the mantle density at the FMS level; and ρ_k is the mean density of the crust. Taking into account that the thickness of the entire crust is $M_k = m_1 + m_2$, Eq. (2) can be rewritten in the form:

$$M_k\rho_k = m_2\left(\rho_0 + a\frac{m_2}{2}\right). \tag{4}$$

The FMS depth, which must be constant in the model with radial density variations (let us designate it as $H'_{\rm FMS}$), can be calculated by subtracting the compensation layer thickness (m_2) from the Moho depth:

$$H'_{\rm FMS} = H_m(M_k) - m_2 = H_m(M_k) - \frac{1}{\alpha}\left(\rho_0 - \sqrt{\rho_0^2 + 2\alpha\rho_k M_k}\right). \tag{5}$$

The last equation allows us to determine changes in the FMS depth calculated traditionally by formula (1) in the gradient medium, i.e., under the assumption that the mantle is homogeneous. For this purpose, it is necessary to specify all parameters of the gradient medium, including $H'_{\rm FMS}$, and to determine the depth to the Moho boundary by using expression (5).

In the oceanic lithosphere, the water layer produces an additional load. The thickness of this layer depends on the thickness of the solid crust and is controlled by the condition of isostatic leveling. Therefore, the expression will be slightly more complicated:

$$H'_{\rm FMS} = H_m(M_k) - \frac{1}{\alpha}(\rho_0 - \rho_{\rm V}) - \sqrt{(\rho_0 - \rho_{\rm V})^2 + 2\alpha(M_k\rho_k + \rho_{\rm V}(H'_{\rm FMS} - M_k))}. \tag{6}$$

Figure 4 shows variations in the FMS depth depending on the increase of the crustal thickness, which are obtained from experimental data (Fig. 4a), and the curves of the theoretical dependence of the FMS depth on the crustal thickness, based on the calculations with the use of formulas (1), (5), and (6) at $H'_{\rm FMS} = 4$ km as well as at different values of α and the initial density at the FMS level.

In models with the initial mantle density $\rho_0 = 3.3$ g/cm^3 (Fig. 4b), positive and negative values of α correspond to the increase and decrease of density with depth, respectively. It is evident that in this case, the degree of FMS depth variations changes nonlinearly. Comparison of the trend obtained from experimental data with the model plots of FMS depth variations in Fig. 4b shows that the curve with the coefficient $\alpha = -0.0017$, which yields a density decrease to 3.2 g/cm^3 at a depth of 70 km, best corresponds to experimental data. However, the shape of the experimental dependence of the FMS depth, as distinct from this theoretical curve, suggests that the degree of FMS depth variations tends to decrease with crustal thickness increasing.

If we assume that the mantle density in the lower part of the lithosphere is 3.3 g/cm^3 and changes linearly in such a way that the density at the FMS level can be higher or lower than this value, the shapes of the theoretical curves will change, and they will become similar to the experimental dependence. A number of such curves at different values of α and the initial density ρ_0 are shown in Fig. 4c (for the curves with $\alpha > 0$ and $\alpha < 0$, the initial densities ρ_0 are assumed to be 3.2 and 3.4 g/m^3, respectively). In this case, the curves coincide rather well, but only within the thickness of the continental crust. This coincidence takes place with the curve corresponding to an increase of the linear density from 3.23 g/cm^3 at a depth of 30 km to 3.28 g/cm^3 at a depth of 80 km.

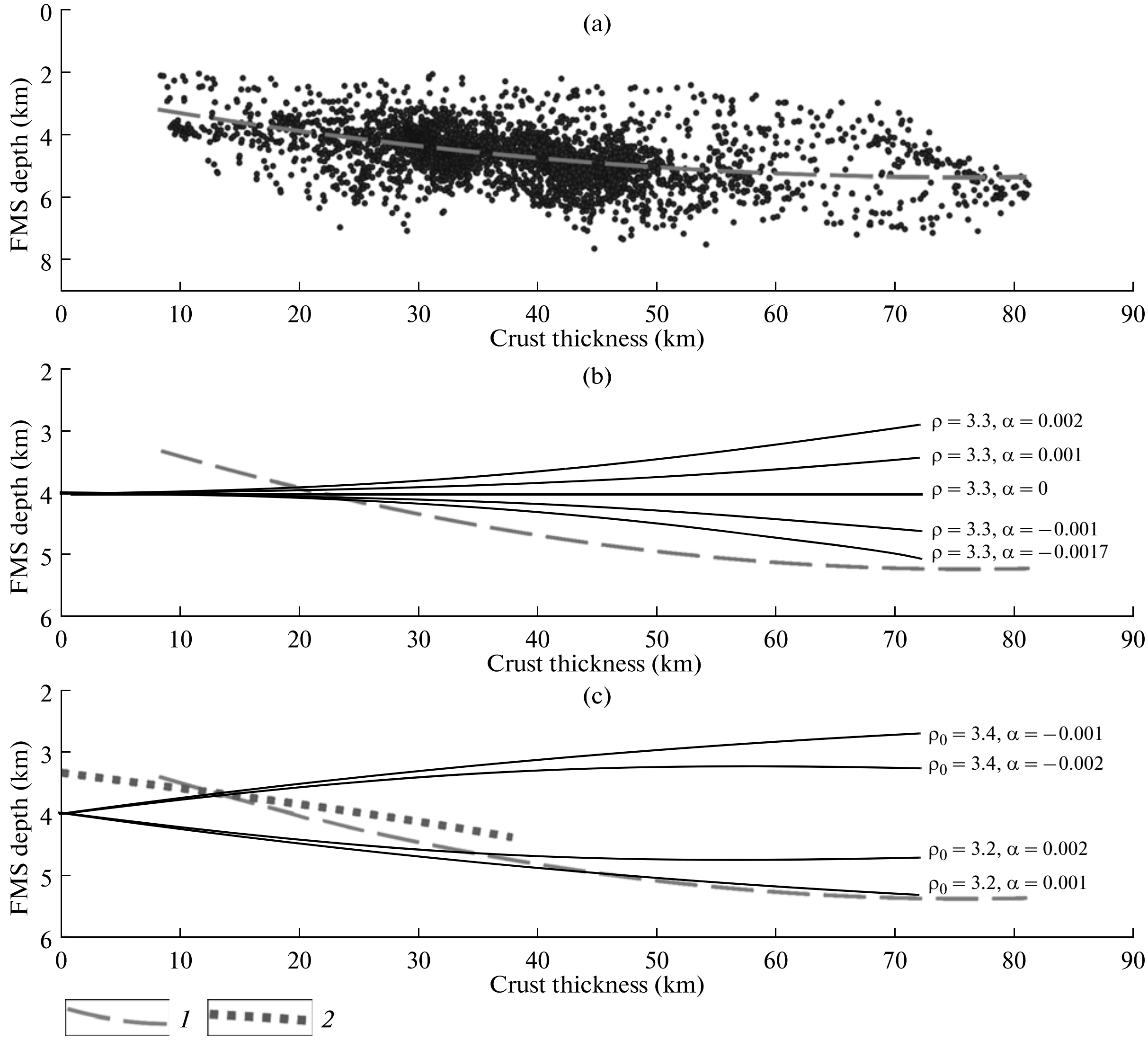

Fig. 4. Changes in the FMS depth calculated with the use of formula (3) for the model with radial density variations at different increasing ($\alpha > 0$) and decreasing ($\alpha < 0$) coefficients: (a) experimental dependence of the FMS based on the AsCrust-08 model data; (b) curves calculated under the assumption that the density at the FMS level is constant and equal to 3.3 g/cm^3; and (c) curves calculated under the assumption that the density in the lower part of the lithosphere is constant and equal to 3.3 g/cm^3 (see the text for details). Notation: (*1*) polynomial trend controlling the dependence of the FMS on the crustal thickness from experimental data (taken from Fig. 7a); (*2*) curve of the FMS depth selected for an oceanic mantle with an increase in the linear density (see the text for details).

The lower part of the upper mantle, which affects the FMS depth but is disregarded in our calculations, is colder under continents than under oceans. Therefore, the FMS level must be higher in oceanic regions. The theoretical curve of the FMS for the oceanic crust calculated at $H'_{\text{FMS}} = 3.2$ km and the mantle density, which increases from 3.2 3.2 g/cm^3 at the FMS level to 3.3 g/cm^3 at a depth of 33 km, best corresponding to our experimental curve are shown in Fig. 3c.

Geophysical data on the character of density variations in the lithosphere with depth are contradictory. Temperature and pressure increasing with depth affect the density in opposite directions. A. Ringwood [1972, p. 15], with reference to F. Burch and D. Anderson, writes that the Vs velocities in the lithosphere predominantly decrease with depth, and Vp velocities decrease to a lesser degree with depth.

In parametric models of the Earth, such as the PREM [Dziewonski and Anderson, 1981] and Ak135 [Montagner and Kennett, 1996] models, the density and velocity in the lithosphere linearly decrease with depth. However, in the earlier PEM model [Dziewonski et al., 1975], the density in the lithosphere was specified to increase with depth.

As noted, models with a linear density increasing best correspond to our data about the FMS depth. However, the density is found to be smaller than its normal values. At the same time, the density distribution over depth in

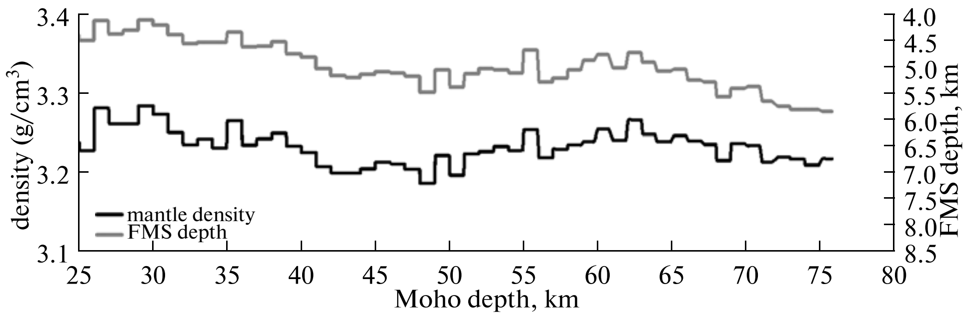

Fig. 5. Variations in the mean density of the subcrustal mantle and the FMS level in the continental lithosphere depending on the depth of the Moho boundary, according to AsCrust-08 model data.

the mantle is not necessarily linear. Further, we will show changes in the mantle density corresponding to FMS depth variations with crustal thickening.

ESTIMATION OF THE DEEP DENSITY DISTRIBUTION FROM FMS DATA

The estimates of the density distribution over depth presented above are based on the revealed general tendency toward a change in the linear density with depth. Our data allow us to estimate a nonlinear density distribution over depth on the basis of two assumptions: (1) the entire study region is isostatically compensated and (2) the mantle density varies only over depth and does not vary along the lateral. In these conditions, the real FMS depth, as is noted above, must be the same everywhere (as distinct from the calculated FMS depth, i.e., the depth, which is obtained under the assumption that the mantle is homogeneous in density), and the observed variations in the FMS values calculated by formula (1) with crustal thickness increasing will show variations in the mean density within the range from the real FMS depth to the current Moho depth. In order to eliminate the scatter of FMS depth at the points with the same values (or at values close to each other) of the crustal thickness, the calculated FMS values will be averaged within a 1-km range.

For estimating the FMS level in the model with radial density variations, we will calculate this level for a crust 33 km thick. As different models show, the crust with such thickness occupies an intermediate position between the continental and oceanic crust, and its upper level is located at about 0 km. Assuming that the crustal density is 2.85 g/cm^3 and the mantle's density is 3.3 g/cm^3, we will obtain that the FMS depth is 4.5 km.

Let us designate in formula (1) the weight of the density column at a certain specified point as $V_i = \sum_{i=1}^{N} M_i \rho_i$. Then, the formula for calculating the FMS can be rewritten in the following form:

$$H'_{\text{FMS}} = H_m - \frac{V_i}{\rho_m}, \qquad (7)$$

hence it follows that

$$\rho_m(H_m) = \frac{V_i}{H_m - H'_{\text{FMS}}}. \qquad (8)$$

The last expression allows us to calculate *the mean value of the mantle density* in the depth range from the FMS (i.e., from 4.5 km) to the Moho boundary at each point of the model. We calculated the V_i and H_m values on the basis of the data provided by the AsCrust-08 model. The calculated distributions of the mean density of the continental mantle and the FMS depths in the range from 25 to 75 km are shown in Fig. 5. As is seen from the lower plot in Fig. 5, the mean mantle density under continents varies from 3.18 to 3.28 g/cm^3, averaging 3.24 g/cm^3. Such low densities are most likely explained by the presence of lateral density heterogeneities under the crust in some ranges of its thickness (that are disregarded by us).

DISCUSSION AND CONCLUSIONS

Our calculations of the deep density distribution in the mantle most simply explain the observed dependence of the FMS depth on the crust thickness. It was also shown that the existence of this dependence makes it possible to determine density variations in the mantle with depth.

The density distribution in the mantle of Central and Southern Asia estimated in Fig. 5 should be regarded as preliminary, because (1) density heterogeneities in the crust are studied scantily, which leads to large errors in FMS determinations; and (2) the method used does not take into account the possible influence of lateral density heterogeneities in the mantle, which can be significant.

In addition, it should be remembered that the obtained density distribution shows the mean density value in the depth range from the FMS level to the Moho boundary, rather than the density at a specified depth level. Therefore, some local anomalies of the density distribution curve in Fig. 5 seem to be improbable, because very large and sharp density differences in depth are required for their formation.

The calculated models of the deep density distribution in the mantle show that the observed dependence of the FMS depth on the crust thickness in regions of Central and Southern Asia can be, on the whole, explained by the use of the model of the underlying mantle with an increase in the linear density; however, in this case, the basic FMS levels in the continental and oceanic lithosphere will be different.

The selection of the deep density distribution on the basis of averaging of the dependence of the FMS depth on the crust thickness with a step of 1 km, which is presented in Fig. 5, yields obviously underestimated density values. This underestimation is due to the presence of lateral density heterogeneities depending on the crust thickness, which were disregarded by us. Such lateral density heterogeneities can be caused by variations in the temperature, chemical composition, phase transition boundaries, and by other factors.

The presence of density heterogeneities in the mantle is beyond doubt, because it is confirmed by multiple data of seismic tomography. However, it is rather difficult to explain the existence of such density heterogeneities in the mantle, whose values depend on the thickness of the overlying crust. It can be suggested that the origin of such heterogeneities is associated with the formation process of the crust. In this case, we can expect distinctions in the relation of the FMS depth to the crust thickness in the continental and oceanic lithosphere, where different mechanisms of the crustal formation can take place.

Digital data about FMS anomalies and the crust model are accessible on the sites: http://www.ifz.ru/personel/baranov.htm and www.geodynamics.ru.

ACKNOWLEDGMENTS

We are grateful to S.A. Tikhotsky for his constructive criticism, which made it possible to clarify the sense of our work and improve its presentation; as well as to T.V. Romanyuk, who pointed out errors and drawbacks in the work and ways of correcting them.

The work was supported by the Russian Foundation for Basic Research, grants nos. 08-05-99098 and 10-05-00579, and by the Russian Science Support Foundation, grant "Best Candidates of Sciences of RAS".

REFERENCES

Anderson, D., Speculations on the Nature and Cause of Mantle Heterogeneity, *Tectonophysics,* 2006, vol.416, pp. 7–22.

Artemjev, M.E., *Izostaziya territorii SSSR* (Isostasy in the USSR Territory), Moscow: Nauka, 1975.

Artemjev, M.E. and Belousov, T.R., Recent Tectonics and Isostasy in the Pamirs, Southern Tien Shan and the Zone of Their Junction, *Bull. Geoph.Theor and Appl.*, 1983, vol. 25, nos. 99–100, pp. 401–408.

A. A. Baranov., A New Model of the Earth's Crust in Central and South Asia, *Fiz. Zemli*, 2010, no.1, pp. 37–50 [*Izv. Phys. Earth* (Engl. Transl.), 2010, vol. 46, no. 1, pp. 34–46.]

Barton, P.J., The Relationship between Seismic Velocity and Density in the Continental Crust—a Useful Constrain? *Geophys. J. R. Astron. Soc.,* 1986, vol. 87, pp. 195–2008.

Bassin, C., Laske, G., and Masters, G., The Current Limits of Resolution for Surface Wave Tomography in North America, *EOS Trans. AGU*, 2000, vol. 81, no. 48, Fall Meet. Suppl., Abstract F897 (http://mahi.ucsd.edu/Gabi/ rem.html).

Dziewonski, A., Hales, A., and Lapwood, R., Parametrically Simple Earth Models Consistent with Geophysical Data, *Phys. Earth Planet. Int.*, 1975, vol. 10, pp. 12–48.

Dziewonski, A. and Anderson, D., Preliminary Reference Earth Model, *Phys. Earth Planet. Inv.*, 1981, vol. 25, pp. 297–356.

Fang, S.-X., Feng, R., Tian Ch.-Zh., et al., Features of Isostatic Gravity Anomaly and Seismic Activity in the Central Asian Region, *Acta Seismologia Sinica*, 1997, vol. 10, no. 6, pp. 801–805.

Laske, G. and Masters, G., A Global Digital Map of Sediment Thickness, *EOS Trans. AGU*, 1977, vol. 78, F483.

Mishra, D. and Rajesh, R., Admittance Analysis and Modeling of Satellite Gravity over Himalayas—Tibet and Its Seismic Correlation, *Current Science*, 2003, vol. 84, no. 2, pp. 224–230.

Montagner, J.-P. and Kennett, B.L.N., How to Reconcile Body-Wave and Normal-Mode Reference Earth Models? *Geophys. J. Int.,* 1996, vol. 123, pp. 229–248.

Ricard, Y., Chambat, F., and Lithgow-Bertelloni, C., Gravity Observations and 3D Structure of the Earth, *Comtes Rendus Geosciences*, 2006, vol.338, nos. 14–15, pp. 992–1001, 228: 992–1001, doi: 10.1016/j.crte.2006.05.013.

Ringwood, A. E., Composition and evolution of the upper mantle, in: *The Earth's Crust and Upper Mantle: Structure, Dynamic Process, and Their Relations to Deep-Seated Geological Phenomena,* Hart, P.J. (Ed.), Geophys. Mongr. Ser. 13, Washington, D. C., pp. 1–17; Moscow: Mir, 1972, pp. 7–26.

Senachin, V.N., *Svobodnaya poverkhnost' mantii kak indicator geodinamicheskikh protsessov* (The Free Mantle Surface as an Indicator of Geodynamic Processes), *Vestnik DVO RAN*, 2006, no. 1, pp. 18–25.