STRUCTURE OF THE LITHOSPHERE OF THE MONGOLIAN-SIBERIAN MOUNTAINOUS PROVINCE

YU. A. ZORIN, M. R. NOVOSELOVA, E. KH. TURUTANOV and V. M. KOZHEVNIKOV

Institute of the Earth's Crust, U.S.S.R. Acad. Sci., Siberian Branch, Irkutsk, U.S.S.R.

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ABSTRACT

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The Mongolian-Siberian mountainous province distinguished by Florensov (1978) includes the Sayan-Baikal, the Altai-Sayan regions and East Mongolian high ranges, and those of Mongolian Altai, Goby Altai and Khangai. In the Late Cenozoic all this province was involved in an intensive orogeny, the latter occurring both in the area under extension and in that under compression (the Sayan-Baikal domal uplift and the Altai uplift system, respectively). The study of the deep structure of the mountainous province and of relatively stable adjacent regions must contribute to a more complete understanding of the causes of intracontinental orogeny. The main object of this study was to map the Moho depth and the thickness of the lithosphere as a whole, based on the interpretation of geophysical data covering the Mongolian-Siberian mountainous province, the Southern Siberian platform and the East Mongolian high plains.

CRUSTAL THICKNESS

A relatively large number of deep seismic soundings has been performed in the U.S.S.R. part of the province (Pusyrev, 1981). But the data available are not sufficient to map the crustal thickness throughout the whole area. Based on deep seismic soundings, with the profiles following the strike of recent topographic features, a regression equation was obtained for the Baikal rift zone, the Southern Siberian platform and for the Trans-Baikal region of moderate orogeny (Zorin *et al.*, 1986):

$$H_M = 39.3 + 4.4 \cdot h_{\mathrm{eff}}, \tag{1}$$

where H_M = the crustal thickness, measured from sea level (Moho depth); h_{eff} = effective height of relief (km). In order to determine h_{eff}, one should add the heights, averaged by $30 \times 30\ \mathrm{km}^2$ areas, their increments calculated

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from the formula $\Delta h = \Delta g_d / 2\pi f \sigma_T$, where Δg_d = the decompensative gravity anomaly (Zorin *et al.*, 1985, 1986); f = a gravity constant; $\sigma_T = 2670\ \mathrm{kg/m^3}$, the topographic mass density. Thus, topographic relief and upper crustal density inhomogeneities, converted to equivalent height increments, are reflected in effective heights.

The analysis of isostatic gravity anomalies shows that isostasy is maintained both in the U.S.S.R. (Florensov, 1977) and in the Mongolian (Zorin *et al.*, 1982) parts of the area under study, with topographic masses balanced, as well as large density inhomogeneities of the upper crust. At the same time, comparison of the regression coefficient from equation (1) with its theoretical value, calculated with regard to the density jump at the base of the crust (this jump is determined based on the known relationship between density and *P*-wave velocity (Subbotin *et al.*, 1979), indicates that the geometry of the Moho provides only 40% of the isostatic compensation (Zorin *et al.*, 1986). The remaining 60% is produced by density inhomogeneities in the mantle and the lower crust. Both the latter relation and equation (1) are valid only for the U.S.S.R. territory, where the effective heights in an axial part of the Sayan-Baikal domal uplift do not exceed 1.8 km*, with the top of the asthenosphere ascending to the Moho discontinuity (Zorin *et al.*, 1986).

In West Mongolia, in Khangai and in Mongolian Altai, effective heights of relief amount to 2.9 km. As will be shown below, there are reasons for believing that beneath these regions the mantle part of the lithosphere is fully replaced by the asthenosphere. If one takes as constant the asthenosphere density deficit related to that of the lithosphere for the whole area of the Mongolian-Siberian mountainous province, it should be considered that the amount of mantle-compensating masses per unit of area (down to 200 km) is almost the same as that beneath the Sayan-Baikal domal uplift, Khangai and Mongolian Altai. Hence, the increment of the effective height for the latter two regions related to its extreme value for the former one must be compensated at the expense of an additional crustal thickening.

Proceeding from these ideas, the Moho depts for the Mongolian regions with effective heights of relief less than or equal to 1.8 km were determined using equation (1). For those regions with the effective heights exceeding the latter value, an additional term in the equation was used, namely $\sigma_T/\Delta\sigma_M \cdot (h_{\mathrm{eff}} - 1.8)$. Here σ_T = topograhic mass density given above; $\Delta\sigma_M$ = the density change at the base of the crust amounted to $240\ \mathrm{kg/m^3}$ (Zorin *et al.*, 1986). Adding this term to equation (1), and assuming that in

* Since these heights are calculated using the altitudes, averaged by the sliding window of $30 \times 30\ \mathrm{km^2}$, they turned out to be less than the maximum height values for the dissected relief.

this equation the value of h_{eff} amounts to 1.8, after some transformation we have the following:

$$H_M = 27.1 + 11.1 \cdot h_{\mathrm{eff}}. \tag{2}$$

The map of the lithospheric thickness beneath the Mongolian-Siberian mountainous province and adjacent regions was constructed based on calculations using equations (1) and (2) (Fig. 1).

Beneath the Siberian platform, crustal thickness ranges from 39 to 43 km, and beneath the high plains of East Mongolia it amounts to from 42 to 45 km. Beneath the major part of the Baikal rift zone (of the Sayan-Baikal domal uplift), the Moho is at depths ranging from 34 to 47 km, with some local uplifting beneath rift depressions. In the south-western rift zone,

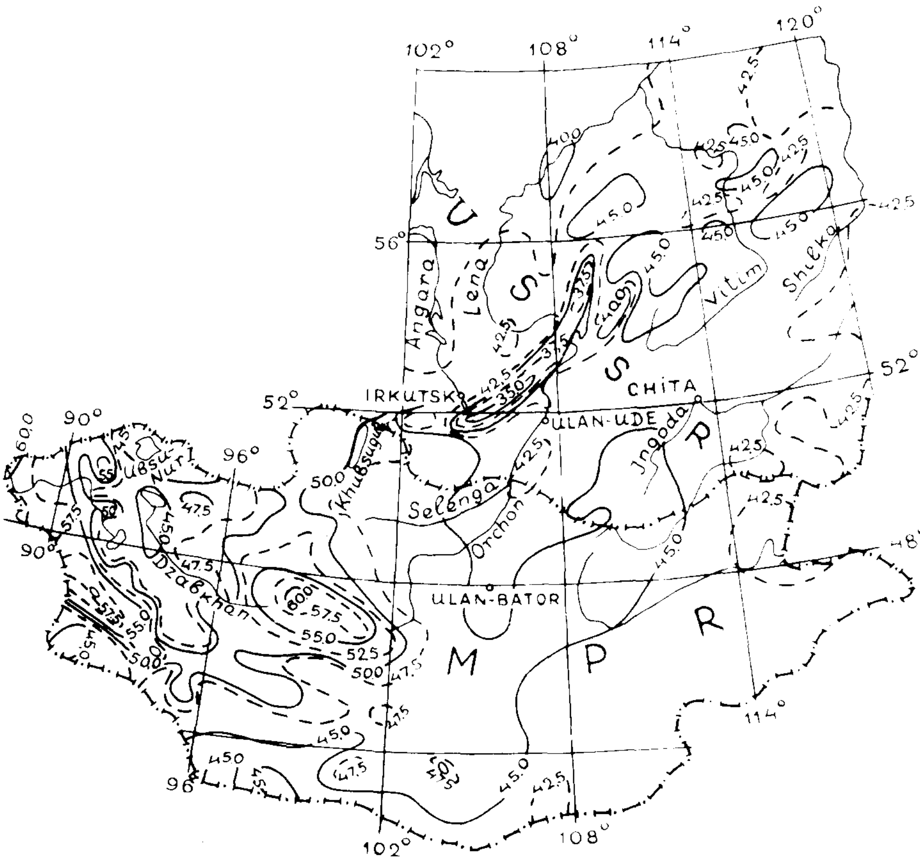

Fig. 1. Map of the crustal thickness. The interval of solid contours is 5 km.

including the East Sayan ridge and the pre-Khubsugul area, the crustal thickness ranges from 46 to 52 km. Beneath Khangai and Mongolian Altai, the Moho discontinuity deepens to more than 60 km. The crustal thickness beneath the Great Lake Valley amounts to 44–45 km.

Thus, we have crustal thicknesses beneath all ridges of the mountainous province somewhat increased as compared to depressions and relatively stable regions. At the same time, for West Mongolia, where compression is present, a maximum crustal thickness is obtained that correlates well with the maximum heights of ridges.

LITHOSPHERIC THICKNESS

There are schematic maps available of lithospheric thicknesses (of depth of mantle solidus temperature) for the U.S.S.R. part of the province, based on heat-flow measurements taking account of the assumption that the thermal field is steady-state (Čermák, 1982; Duchkov and Sokolova, 1986). The estimation of the lithospheric thickness based on heat-flow data is, however, rather conventional, because the thermal properties of the lithosphere are not well studied. On the one hand, the vagueness is greater for regions with low heat-flow values, i.e., for platforms. On the other hand, in the regions of present-day tectonic activity, where relatively recent deep mass displacements are possible, the thermal field can be transient, and this may lead to overestimations of the lithosperic thickness.

The definition of the asthenosphere as a layer in which partial melting of the mantle occurs, however, allows of making the inference that it must differ from the lithosphere in its seismic velocities, electrical conductivity and density. Thus, seismological, magnetotelluric sounding and gravity data can be used to determine lithospheric thickness. It can be determined with higher resolution from deep seismic soundings with long profiles and from surface-wave data. But deep seismic soundings are rather laborious and surface-wave investigations are restrained by the limited number of stations equipped with long-period seismographs. This is the reason why the lithospheric thickness is estimated using seismological methods in some regions only and the data available are not sufficient to map this parameter throughout the area. Magnetotelluric soundings of a wides range of recorded electromagnetic variation periods have also been carried out in a few regions. Observation results of such soundings are strongly distorted by surface electrical inhomogeneities, and this leads to the semi-quantitative character of estimations of lithospheric thickness.

In the present study, we propose to map lithospheric thickness based on regional gravity anomalies, using seismic constraints (Zorin *et al.*, 1986). We assume that the anomalies related to lithospheric thickness variations can be defined as follows:

$$\Delta g_L = \Delta g_B - \Delta g_{uc} - \Delta g_{lc}, \tag{3}$$

where Δg_B = Bouguer anomalies, Δg_{uc} = gravity effect of upper crust inhomogeneities, Δg_M = effect of the Moho configuration, Δg_{lc} = effect of lower crustal density inhomogeneities. We used so-called decompensative gravity anomalies as Δg_{uc}. These are isostatic anomalies specifically transformed to diminish the effect of local compensation of density inhomogeneities of the upper crust (Zorin *et al.*, 1985, 1986).

The effect of Moho configuration is calculated using the map of crustal thickness (Fig. 1). For this, the crust is presented as a combination of vertical rectangular prisms with horizontal section areas of $30 \times 30\ \mathrm{km}^2$. The density jump at the Moho discontinuity is taken as $240\ \mathrm{kg/m}^3$, based on seismological data of *P*-wave velocities in the crust and the mantle of the Siberian platform (Zverev *et al.*, 1980) and on the known relationship between density and *P*-wave velocity (Subbotin *et al.*, 1979).

When the upper-crust inhomogeneities and the Moho discontinuity effects were subtracted from Bouguer anomalies, it turned out that the long-wave residual anomaly in the Baikal rift zone is complicated by local disturbances of the gravity field: the central part of a broad low associated with the Sayan-Baikal domal uplift is superimposed by relatively narrow highs associated with rift depressions. It is apparent that these highs are produced by an additional isostatic compensation of rift depressions not accounted for in formula (1). These compensating masses occur in the lower crust and appear to be related to its density increase caused by the intrusion of mantle dyke swarms (Florensov, 1978). Such basic zones of the lower crust were modelled by a layer with variable density excess. The thickness of the layer is assumed to be 18 km, its base coinciding with the Moho discontinuity. The distribution of the density excess within the layer is obtained based on following ideas. The fourth-order trend surface describing the Sayan-Baikal domal uplift heights, ignoring rift depressions, was constructed based on the map of effective relief, using the least-square method. The difference between effective heights and their trend values characterizes the mass deficit in rift depressions. This deficit, owing to isostasy, is equal (with opposite sign) to the local compensating mass excess. Therefore, the variable density responsible for additional compensation of rift depressions can be presented as follows: $\rho_{lc} = 0.6 \cdot \Delta h_{eff} \cdot \sigma_T / m$, where Δh_{eff} = the above difference between effective height values and their trend; σ_T = topographic mass density; m = layer thickness. The coefficient 0.6 reflects the portion of compensating masses not considered by formula (1).

The effect of an additional compensating layer of the lower crust (Δg_{1c}) was calculated by separating the crust into vertical prisms with horizontal sectional areas of $30 \times 30\ \mathrm{km}^2$, based on the density distribution obtained. This effect was then subtracted from the residual anomalies. This is done for the rift zone only.

When all the above calculations are performed, the anomaly field is obtained which is supposed to be associated with the "lithosphere-asthenosphere" boundary. In order to make unique inversions of these anomalies, one should have independent lithosphere thickness determinations for at least several sites. As we have already mentioned above, the most authentic information can be obtained based on seismological data.

Very long travel-time curves for refracted waves were obtained from deep

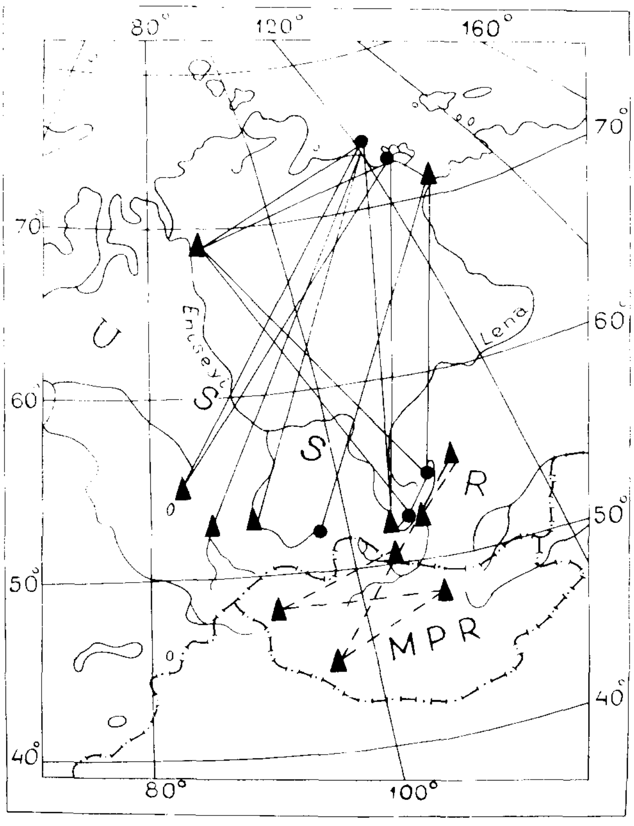

Fig. 2. position of the Rayleigh wave paths. Solid lines indicate the paths used for the determination of group velocities; dashed lines indicate the paths used for the determination of group velocities; dashed lines indicate the paths used for the determination of phase velocities. Solid circles show earthquake epicenters; triangles show seismic stations.

seismic soundings on the Siberian platform. Based on their analysis jointly with data on the arrival times of reflected and alternated waves, two (Zverev and Kosminskaya, 1980) or four (Yegorkin *et al.*, 1984) low-velocity layers can be revealed in the upper mantle. All versions of interpretations show that the layer with its top at a depth of 200 km is the thickest and possesses the maximum relative decrease of *P*-wave velocity and *Q*-factor.

An analysis of dispersion curves of *Rg*-wave group velocities for 12 paths with their main parts situated within the Siberian platform was performed by one of the authors of this study (Kozhevnikov, 1987) (Fig. 2). These curves (fundamental and two first highest modes) turned out to be practically similar for all paths, indicating the homogeneity of deep structures of

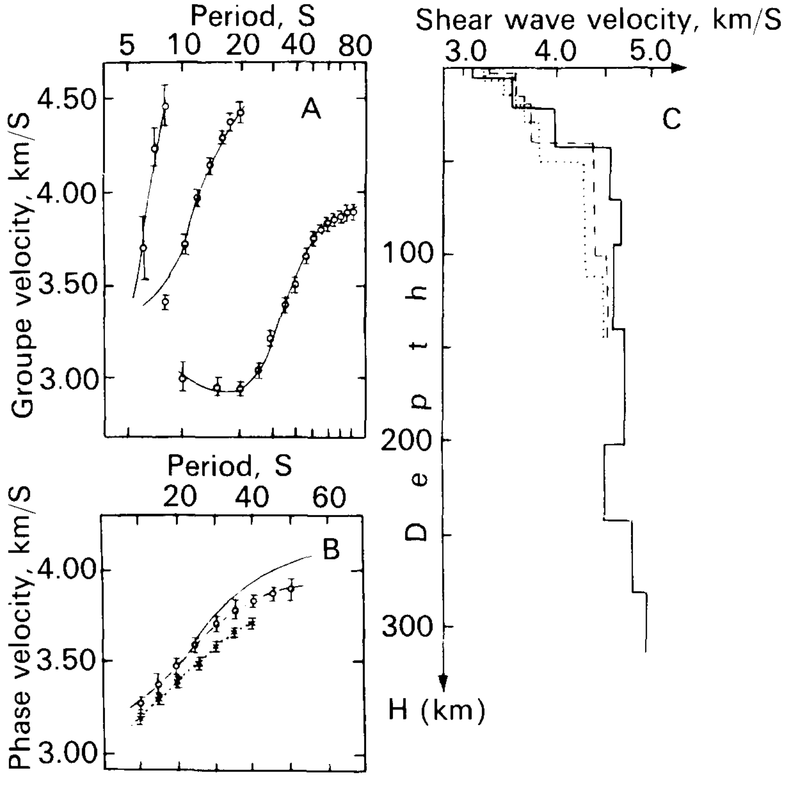

Fig. 3. Rayleigh wave dispersion curves and the results of their inversion. A) Group velocities for the Siberian platform; B) Comparison of phase velocity curves for the Siberian platform, the Baikal rift zone and the Khangai region; C) Patterns of *S*-wave velocities versus depth. Solid lines are referred to the Siberian platform, dashed lines to the rift zone, and dotted lines to the Khangai region; vertical bars denote the confidence intervals of mean velocity values of the *Rg*-waves.

the platform. The results of averaging data for all paths are given in Fig. 3. The confidence intervals of mean values of the group velocity are also indicated. The *S*-wave velocity distribution versus depth, obtained from the inversion of averaged curves, is similar, in the position of discontinuities, to the *P*-wave velocity distribution obtained from seismic soundings (Yegorkin, 1980). The *S*-wave velocity within the layer with its top at a depth of 200 km amounts to 4.45 km/s, and in the overlying medium it reaches 4.70 km/s (Fig. 3).

Electrical conductivity values obtained from magnetotelluric soundings with a wide frequency range indicate that partial melting can occur at depths exceeding 180–200 km (Pusyrev, 1984). Finally, a mathematical simulation of cooling of the Earth's upper layer, where heat transfer is conductive, shows that the known relationship between heat flow and the age of geological units (Polyak and Smirnov, 1968) is reproduced if the layer thickness is taken as about 200 km (Zorin, 1987). The latter value is evidently the estimation of maximum continental lithospheric thickness, implied to be a typical value for old platforms. Thus, combined data of different geophysical methods (deep seismic soundings, surface wave analysis, magnetotelluric soundings, geothermal modelling) allow us to assume that the lithospheric thickness beneath the Siberian platform amounts to 200 km.

Deep seismic soundings indicate that the *P*-wave velocity in the mantle immediately beneath the crust of the Baikal rift zone is estimated as 7.7 km/s, and that of the Siberian platform amounts to 8.2 km/s (Pusyrev, 1981). In the rift zone, the mantle velocity increases with depth. A similar tendency is also relevant for the Siberian platform (Yegorkin, 1984); thus, at depths of about 60–80 km, the velocity is still lower beneath the former region than that below the latter one. Our analysis of dispersion curves of *Rg*-wave phase velocities indicates that an analogous *S*-wave velocity difference beneath these regions still remains down to depths exceeding 120 km. The *S*-wave velocity distribution for the rift zone is shown in Fig. 3c. The lower left segment of the figure displays confidence intervals of mean phase-velocity values obtained from the data of the Uoyan-Kabansk and Uoyan-Zakamensk paths (Fig. 2). The phase-velocity curve theoretically calculated for the Siberian platform is also given for comparison. This curve is obtained based on the inversion of group velocity data.

Some ten years ago, it was ascertained that all seismic stations within the Sayan-Baikal domal uplift trace the delays of *P*-waves from distant earthquakes and explosions. If one uses only the most exactly determined delays from distant great explosions (Rogozhina *et al.*, 1983), with almost vertical angles of emergence of seismic rays and precise data on mean velocities, then the anomalous low-velocity mantle beneath the Sayan-Baikal domal uplift involves the depth range from 40–50 km (bottom of the crust) to

200 km. Similar results were obtained using the method of spectral ratios of seismic waves (Zorin *et al.*, 1986). Thus, the lower boundary of the anomalous (low-velocity) mantle beneath the rift zone and the base of the lithosphere underneath the Siberian platform are at the same depth. For the area of the anomalous mantle, a significant attenuation of seismic waves is characteristic (Vinnyk, 1976). This indicates a partial melting of mantle material. Therefore, this area actually can be treated as an asthenospheric upwelling, above which the lithosphere is thinned to a crustal thickness.

This inference is supported by the results of determinations of hypocenter depths. These depths range here from 0 to 20–25 km (Golenetsky, 1977). The latter values estimate the thickness of the layer accumulating elastic stresses and being subdued due to deformation with faulting. Deeper, a subsolidus creep is evident. It is ascertained for some regions that the mantle solidus temperature corresponding to the base of the lithosphere occurs at depths exceeding the thickness of the elastic layer by a factor of 2 (Peltier, 1984). Thus, the lithospheric thickness beneath the Sayan-Baikal domal uplift can amount to 40–50 km. This corresponds to the above depths of the top of the anomalous mantle.

An averaged dispersion curve of *Rg*-wave phase velocity is obtained for the Khangai region. It is constructed for the paths Ulan-Bator—Khovd, Ulan-Bator—Altai, Zakamensk-Altai and Zakamensk-Khovd (Fig. 2b). The phase velocities for the Khangai region are less than that for the rift zone (Fig. 3a). Its inversion shows that the crust is thickes beneath Khangai than beneath the rift zone, and that the mantle *S*-wave velocity is much lower here than that beneath the Siberian platform. Such a difference remains down to depths exceeding 100–120 km (see Fig. 3c).

When seismic rays from distant large explosions emerge subvertically, *P*-wave delays in mountainous regions of West Mongolia amount to 1 s, this being almost the same as for similar values for the Baikal rift zone (Rogozhina *et al.*, 1983). Earthquakes in these regions are also shallow (Solonenko and Florensov, 1985). The above data permit the inference that beneath Khangai, as well as beneath the Sayan-Baikal domal uplift, the lithosphere is thinned to a crustal thickness.

Lithospheric thickness estimations based on seismic data are emphasized here because they are used as constraints for gravity-anomaly inversion and they determine considerably the reliability of the latter. The above estimations, compared to the regional mantle anomalies, have shown a good accordance: the gravity-field values are increased in the platform part and decreased in axial parts of the Sayan-Baikal and the Khangai uplifts. There is a good reason then to suppose that Δg_L anomalies reflect lithospheric thickness variations.

An inversion of these anomalies was performed taking account of the assumption that they are produced by deflection of the lithospheric thickness from its "normal" value, taken as 200 km. The lithosphere was divided into vertical rectangular prisms with horizontal sectional areas of $120 \times 120\ km^2$ in order to determine the depth of the contact discontinuity corresponding to the asthenospheric roof. Such large dimensions of prisms are imposed by reasons of saving computer time.

With regard to the above data, the following constraints are made beneath the inner area of the Siberian platform (240–480 km from the roots of the mountainous range) the lithospheric thickness is 200 km, and underneath the axial parts of the Sayan-Baikal and the Khangai uplifts, the lithosphere is thinned to a crustal thickness (see Fig. 1). In these places density discontinuity depths were fixed. Such constraints were imposed for almost 15–20% of the prisms considered, implying that both minimum and maximum values of lithospheric thickness were fixed. These conditions are rather strict and make it possible to determine the configuration of the discontinuity in the remaining area, as well as the density jump at this interface. The former problem was solved using the least-square method in a 3-D version (Bulakh *et al.*, 1984). The interface depths and the constant regional field were determined with fixed values of asthenosphere density deficit compared to lithosphere. Then the density difference was changed and the process of automatic inversion was repeated. Minimization of mean-square deviations of the calculated field from the residual regional anomalies was the criterion for determining the final version, implying that the depths obtained should not transcend fixed extreme values.

The theoretical field was obtained in the best conformity with residual anomalies when the asthenosphere was less dense than the lithosphere by $20\ kg/m^3$. This value is apparently averaged both by the area and by the depth range change of the asthenospheric roof. Beneath the regions of active tectonics, small-scale convection is possible in the asthenosphere. It transfers to the base of the lithosphere deep material with density lowered merely owing to its heating and partial melting.

The interface depths determined are referred to the centers of horizontal sections of prisms, and the lithospheric thickness is mapped based on the digital field obtained (Fig. 4). Isolines are drawn using linear interpolation, except for the 200-km isoline countouring the area within which the lithospheric thickness corresponds to, but does not exceed this value.

A lithosphere thinned to crustal thickness is typical for almost the whole Mongolian-Siberian province. The only exception is the Large Lake Valley separating Mongolian Altai from Khangai. Beneath this valley the lithospheric thickness increases to 100–110 km.

The zone of translation from the thin lithosphere beneath the Sayan-Baikal domal uplift to a thick one beneath the internal part of the Siberian

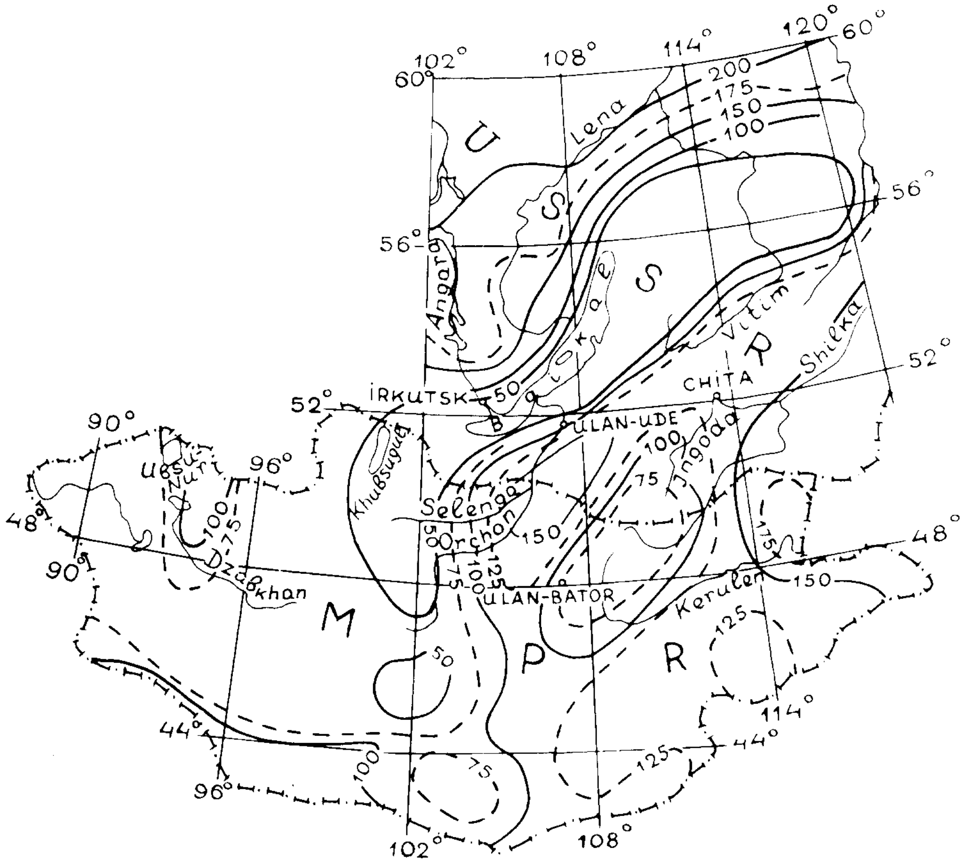

Fig. 4. Map of lithospheric thickness. The interval of solid contours is 50 km.

platform embraces the relatively low Presayan, Upper Lena and Baikal-Patom plateaus. In the Trans-Baikal area, the mean lithospheric thickness amounts to 125 km, ranging from 70 km beneath the Khantai-Dauria uplift to 150 km beneath the Preargun region. Beneath the East Mongolian plains the top of the asthenosphere occurs at depths of 140–175 km, uplifting to 110 km beneath the Dariganga volcanic plateau. In the Trans-Altai Goby region, the lithospheric thickness amounts to 120–130 km (Fig. 4).

DISCUSSION

It seems reasonable to refer the estimations of lithospheric thickness obtained to the regional heat-flow values given by Lysak (1983) and

Khutorskoy *et al.* (1986) and to correlate these data with the theoretical relationship between heat flow and depths to the asthenospheric roof. The above relationship is deduced using the formula obtained by Crough and Thompson (1976) for a steady-state case:

$$q_s = H_0 D + K_l(T_m - H_0 D^2/K_c/L - C(K_l/K_c - 1), \tag{5}$$

where q_s = surface heat flow, H_0 = surface radiogenic heat production, D = characteristic depth (a parameter of exponential decrease with depth of radiogenic heat producton in the crust), T_m = mantle solidus temperature, K_c = thermal conductivity of the crust, K_l = thermal conductivity of the lithosphere, C = crustal thickness, L = lithospheric thickness. The relationship between q_s and L is determined for the following parameter values: $H_0 = 2\ \mathrm{Wm}^{-3}$, $D = 10$ km, $T_m = 1350°$C, $K_c = 2.5\ \mathrm{W(mK)}^{-1}$, $K_l = 3.0\ \mathrm{W(mK)}^{-1}$, $C = 40$ km.

Fig. 5 shows that for the regions adjacent to the Mongolian-Siberian mountainous province (the Siberian platform, the Trans-Baikal area, southeast Mongolia) the empirical data are compatible with the results of theoretical calculations. As for mean heat-flow values for this province (the regions of the Baikal rift zone and the Khangai uplift), they are considerably lower than those calculated by formula (5). This allows us to suppose that lithospheric thermal field of the latter regions is transient. The asthenospheric top here must have come close to the Moho rather recently

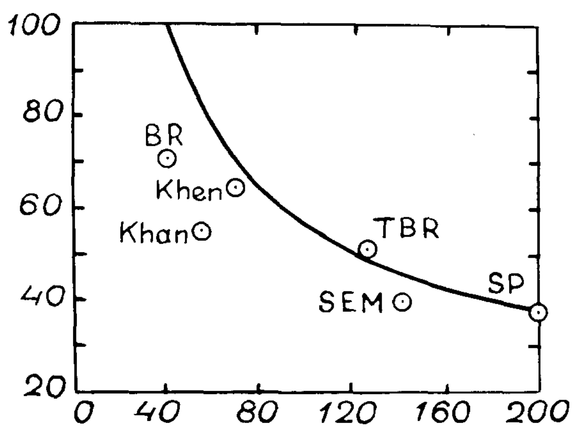

Fig. 5. Heat-flow data versus the lithospheric thickness. Solid line shows the theoretically calculated dependence of heat flow on lithospheric thickness. Experimental data are for: SP: the Siberian platform, TBR: the Trans-Baikal region, SEM: Southeast Mongolia, KHEN: the Khentai uplift, KHAN: the Khangai uplift, BRZ: the Baikal rift zone.

(on the geological time scale) and the crust has not yet heated to temperatures corresponding to thermal equilibrium. A thermal model of the development of asthenospheric upwelling has been made earlier for the Baikal rift zone (Zorin and Osokina, 1981; Zorin and Lepina, 1986). The regional heat-flow value turned out to be in line with the theoretical one, in the case that it is assumed that the lithosphere was mechanically replaced by the asthenosphere starting from 30–35 Ma, and that the asthenospheric roof reached the base of the crust about 3–4 Ma ago. Such time estimations can be attributed to major Late-Cenozoic geological events: the former coincides with the beginning of the development of the Sayan-Baikal domal uplift (the Baikal rift zone), and the latter corresponds to the time when the highs of the uplift reached the recent values. Geothermal modelling results may be valid for the whole Mongolian-Siberian mountainous province, since all uplifts within it were developing almost simultaneously (Zorin and Florensov, 1984).

The idea of the quite recent approach of the asthenosphere to the Moho can explain why the products of lower-crust melting are practically lacking among Cenozoic volcanics within the Mongolian-Siberian mountainous province. A considerable volume of the lower crust could not then have heated to melting temperatures. Moreover, a broad development of Paleozoic granitoids is typical of mountainous uplifts. Below we shall consider the reasons. Extraction of silicic material from the lower crust in the Paleozoic may be responsible for the mainly refractory residual composition of the lower crust in later geological periods.

The significant role of the asthenospheric upwellings in isostatic compensation of uplifts, as well as the similar character of the development of these uplifts and upwellings, allows us to suppose that the orogeny in the region under study was strongly conditioned by asthenospheric diapirism. The above mechanism is sufficient to account for the mean height level of the Sayan-Baikal domal uplift. The increased altitude within this uplift, oriented northeastwards, was accompanied by extension of the crust occurring at its narrow, weakend segments. Above these segments, the rift depressions have appeared.

In East Mongolian uplifts, oriented northwestwards and sublatitudinally, the extension was blocked by compression from the Indian peninsula (Zorin and Florensov, 1984). Together with the lithospheric thinning, a considerable crustal thickening is revealed beneath these uplifts (see Fig. 1). Such thickening may (at least partly) have arisen in Pre-Cenozoic development of the region. However, if one assumes that some additional crustal thickening (shortening) occurred in the Cenozoic, caused by the Indian plate pressing against Eurasia (Molnar and Tapponnier, 1975), one should suppose that such deformation required a condition of lithospheric weakening resulting from its thinning and heating. A more considerable relative contrast between the altitudes of different Central Asian regions and the oceanic parts of the Indian plate with lithospheric thicknesses of 100–120 km appears not to be fortuitous.

Lithospheric thinning due to isostasy must itself lead to the rise of uplifts. Hence, we believe that asthenospheric diapirism is the main cause of the uplifting of the West Mongolian territory. Compression from the Indian plate could induce some additional elevation of the ranges owing to crustal thickening. Such compression was also responsible for strike-slip faulting (Molnar and Tapponnier, 1975). The magnitudes of strike-slip faults were hardly significant since they attenuate along the strike.

An interesting peculiarity of the uplifts of the Mongolian-Siberian province appears noteworthy in making a conclusion. These uplifts are inherited, to this or that extent, from Paleozoic uplifts within which either Caledonian or Hercynian granitic magmatism is manifested. Granite formation is evidently related to an increased heat flow, i.e., to a comparatively small lithospheric thickness. Thus, in the Middle and the Upper Paleozoic, the lithosphere of the uplifts must have been thinned. Since lithospheric thickening at the expense of the crystallization of liquid phases is rather slow (Zorin and Lepina, 1987), a relative thinning of the lithosphere beneath these regions could take place during 250–300 Ma. Low-density material coming into the asthenosphere from hot spots inevitably became caught up in such "traps" (Artushkov, 1979) which served as initial disturbances for the new development of gravity instability, i.e., for asthenospheric diapirism. This is why the uplifts were rejuvenated. The hot spots could exist not only directly beneath the uplifts but also in their vicinity. Therefore, the inheriting of uplifts is determined by a "structural memory" of the slowly cooling asthenosphere rather than by the geometry of deep material sources.

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