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# Thickness of the crust along the Irkutsk–Ulan-Bator–Undurshill profile from spectral ratios of body seismic waves

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**Abstract.** The thickness of the crust is estimated from the frequency intervals between the maxima of spectral ratios of body waves. This method was applied to processing data obtained at 15 observation sites by using digital seismic stations located along the Irkutsk–Ulan-Bator–Undurshill profile. The results obtained are consistent with the previous estimates from correlation of deep seismic sounding (DSS) data for southeastern Siberia with effective heights of topographic relief.

## Introduction

Until recently, the deep structure of Mongolia could be determined only from gravity anomalies [*Stepanov and Volkhonin,* 1969] and correlation of average heights of relief with DSS data for adjacent Russian regions [*Zorin et al.,* 1988, 1990].

In the framework of the Russian-American project Teleseismic Tomography of Mantle in the Baikal Rift, seismic observations were carried out at 15 temporary digital stations along the profile Irkutsk–Ulan-Bator–Undurshill, during 3.5 months in 1992. Although the prime objective of the observations was to study the mantle, the field material obtained can be used for solving many other seismic problems. In particular, the thickness of the crust at observation sites can be estimated from the spectral ratio of body seismic waves. The thickness can be determined from complete inversion of the spectral ratio with fitting of both velocity and thickness of the crustal layers [*Kurita,* 1973; *Phinney,* 1964]. However, we restrict ourselves only to evaluating the crustal thickness at a given average velocity of transverse waves; the thickness can be found directly from the position of the resonance peaks in the graph of the spectral ratio versus wave frequency. Our paper tests the validity of this method and reports results of its application to data of seismic observations along the profile indicated above. The profile crosses the border of the Siberian Platform, Baikal Rift zone, and Mongolia mountain ranges and high plains.

## Seismic Data From the 1992 Survey Network

Seismic events were recorded by three-component short-period seismometers with a natural frequency of about 1 Hz. Two sites also were instrumented with STS 2 receivers to record oscillations at greater amplification in a broad frequency band.

American digital stations Passcal-Reftek were used in recording. The internal quartz crystal clocks were synchronized to the Omega navigation system, resulting in a time error no greater than 0.02 s. The ground velocity was recorded by high-frequency instruments, which worked in two regimes: continuous and trigger. The sampling frequency in the continuous regime was 10 scans per second, which is quite sufficient for using the spectral ratio method. The broadband seismometers continuously recorded signals at the same frequency.

The seismic instruments were calibrated at regular intervals during the field season. Standard pulses were fed to the input, and the calibrating characteristics were detected at the output. This knowledge enabled us to consider instrumental noise in data processing.

The stations recorded on average 20 teleseismic events at a high signal-to-noise ratio during 1 month. However, only a third of the records met the requirements of the method used, which places severe restrictions on the seismic recording. Following the spectral method of



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**Table 1.** Data on earthquakes

| No. | Date | $T_0$, UT | $\varphi$ | $\lambda$ | $h$, km | Magnitude | Station Number |
|---|---|---|---|---|---|---|---|
| 1 | July 12, 1992 | 2340:59.7 | 3.11°N | 122.00°E | 616 | 5.6 | 22, 23, 24, 36, 83, 85, 86, 87, 88, 89, 90 |
| 2 | July 13, 1992 | 0010:08.1 | 3.26°N | 121.86°E | 612 | 5.1 | 22, 23, 36, 83, 85, 86, 87, 88, 89 |
| 3 | July 14, 1992 | 0703:09.7 | 4.74°S | 125.45°E | 464 | 5.5 | 11, 12, 22, 23, 36, 85, 86, 87, 88, 89, 90 |
| 4 | July 15, 1992 | 1804:18.8 | 8.82°S | 121.55°E | 125 | 5.7 | 11, 12, 86, 87, 88, 89, 90 |
| 5 | July 28, 1992 | 1730:18.0 | 4.41°S | 127.67°E | 305 | 5.3 | 11, 12, 23, 36, 80, 83, 84, 85, 87, 88, 89 |

$T_0$ is origin time, and $\varphi, \lambda$, and $h$ are coordinates and focal depth, respectively.

determination of the crustal thickness [*Kosarev,* 1971; *Mordvinova,* 1983], we analyzed the 40-s interval of the seismogram beginning from the $P$ wave arrival, as well as (to account for noise effect) approximately the same interval from the previous record. The analysis dealt only with the events for which the indicated interval may be divided in two parts: the initial part with large amplitudes, and the tail of irregular oscillations with essentially smaller amplitudes. The former, identified with a direct $P$ wave, lasted no longer than half the total duration of the desired signal.

Data on the chosen earthquakes are listed in Table 1. Their epicenters are in a narrow range of azimuth ($150-155°$ and $155-160°$) and epicentral distance ($45-60°$). This approach allows the use of averaged spectral ratios, which weakens unknown random effects on the results obtained.

The observation stations were arranged along the south-southeastern direction, almost coinciding with the direction of the observed seismic rays (Figure 1). Thus the formation zone of the spectral ratios was displaced south-southeast of the corresponding observation sites because of seismic migration. The value of this migration may readily be estimated.

## Estimation of Crustal Thickness by the Spectral Ratio Method

This method uses the ratio of the spectrum of the vertical component of body seismic waves, $SW(f)$, to the spectrum of their horizontal components $SU(f)$ [*Haskell,* 1962; *Phinney,* 1964]. This provides for elimination of the undesired factors affecting the spectrum shape, and for extraction of the part of the spectrum that is determined only by properties of the medium in the observation region.

Many authors believe that it is not possible to directly relate the observed ratio $R(f)$ to the structural parameters of the crust [*Bath,* 1980]. However, our theoretical calculations and experimental data show that such a relation can be derived.

The maxima of $R(f)$ (Figure 2) are repeated with the same frequency as the minima of the amplitude spectrum $SU(f)$ of the horizontal component $U(t)$ formed mainly by converted transverse waves [*Phinney,* 1964]. If two waves with their close amplitudes and phase spectra interfere with each other, the position of the minimal (in amplitude) components $f_n$ in the spectrum of the interference wave is governed by the difference between the arrival times of these waves and their mutual polarity. *Khudzinsky* [1961] derived a relation between the position of the spectrum minima and parameters of a particular layer forming multiple longitudinal waves. For the given model of a one-layer crust lying on a homogeneous half-space, the frequency difference $\Delta f = f_{n-1} - f_n$ is mainly determined by the delay time $\tau_{P2S}$ of a multiple converted wave $P_2P_12S_1$ relative to the wave $P_2P_1$ (Figure 3a) and by their polarity (Figure 3b):

$$\tau_{P2S} = 2H \cos i_1^S / V_S \tag{1}$$

where $i_1^S$ is the incident angle of $S$ wave having velocity $V_S$ in a layer of thickness $H$. The mutual polarity ($\Delta\varphi$) is equal to the product of the angular frequency times $\Delta\varphi_1 = \omega\tau = 2\pi f_n \tau$ plus an additional phase shift $\Delta\varphi_2 = -\pi$ produced by reflection of the multiple converted wave from the Earth's surface as a medium with smaller acoustic rigidity. The antiphase oscillations,

$$\Delta\varphi = \Delta\varphi_1 + \Delta\varphi_2 = (2n-1)\pi \tag{2}$$

where $n$ is any integer number, are completely destroyed. Hence we have the frequency of the maxima of $R(f)$ (or of the minima in the amplitude spectrum of the horizontal component):

$$f_n = n(V_S / 2H \cos i_1^S) \tag{3}$$

$$\Delta f = f_1 = V_S / 2H \cos i_1^S \tag{4}$$

Thus the crustal thickness can be estimated from the frequency of peaks ($\overline{\Delta} f$) of the amplitude ratio $R(f)$ at a given average $S$ wave velocity.

Past experience of applying the spectral ratio in studying crustal structure mostly concerned using long-period instruments [*Fernandez and Careage,* 1968;

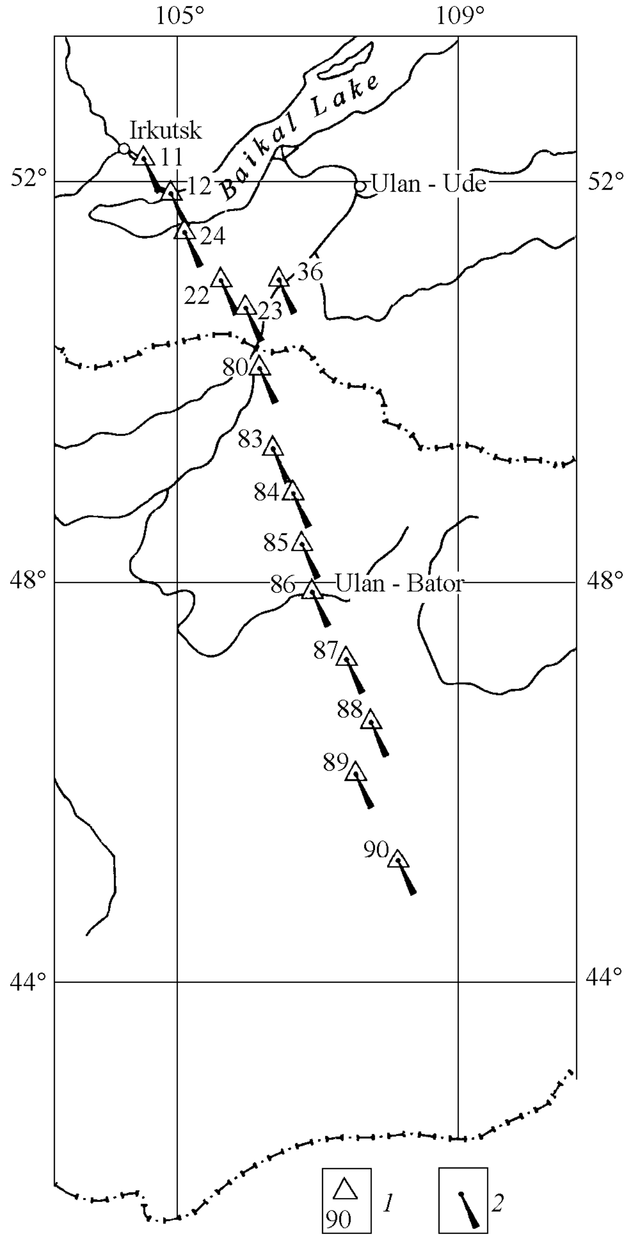


**Figure 1.** Observation region: 1, seismic station; 2, projection of the proposed zone of formation of spectral ratios for body waves in the crust (according to the model in Figure 3a).

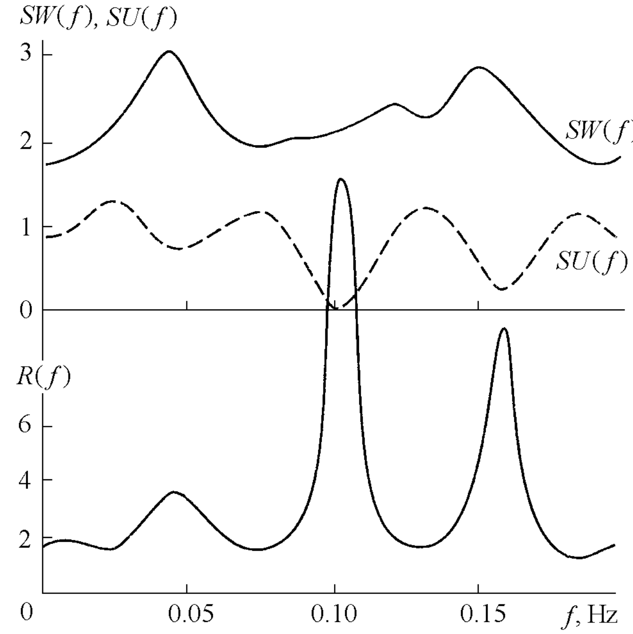


**Figure 2.** Spectral characteristics of the one-layer model [*Phinney,* 1964]. Layer parameters are $Vp_1 = 6.1$ km s$^{-1}$, $Vs_1 = 3.55$ km s$^{-1}$, $\rho_1 = 2.80$ g cm$^{-1}$, and $H = 35$ km (notations are given in text). Half-space: $Vp_2 = 8.1$ km s$^{-1}$, $Vs_2 = 4.68$ km s$^{-1}$, $\rho_2 = 3.30$ g cm$^{-1}$, and apparent velocity $C = 20$ km s$^{-1}$.

*Kosarev,* 1971; *Kurita,* 1973; *Lap Sau,* 1975; *Mordvinova,* 1983, 1988]. However, *Lap Sau* [1977] showed that short-period wave data confirm the structure of the crust inferred from the long-period records.

Note that broadband instrumental records revealed that the frequency ($\overline{\Delta f}$) of the observed ratios $R(f)$ is invariable in different frequency ranges, a finding consistent with theoretical conclusions. Furthermore, sufficiently close dependences $R(f)$ in the frequency range of 0.20–0.96 Hz (Figure 4) were found by comparing the spectral ratios on short- and long-period records at the sites where instruments of both types were installed. The crustal thicknesses estimated from measurements by these instruments are also close to each other.

## Method of Interpretation

The results from the previous section and many theoretically obtained spectral ratios [*Mordvinova,* 1983, 1988] allow us to determine the thickness of the crust and its deep structure from the form of the experimental spectral ratio. The equally spaced peaks in the graph of $R(f)$ may imply the absence of sharp seismic interfaces beneath the station and of the $P$ and $S$ wave inversion, as well as the possibility for the crust to be represented by a one-layer model, or by a medium with constant velocity. It follows from (4) that the power in our case depends on the average distance $\overline{\Delta f}$ between the first two to three peaks of the spectral ratio

$$H = V_S / 2\overline{\Delta f} \cos i_1^S \qquad (5)$$

Both theory and experiment show that the overall crustal thickness can be found from the measured graph

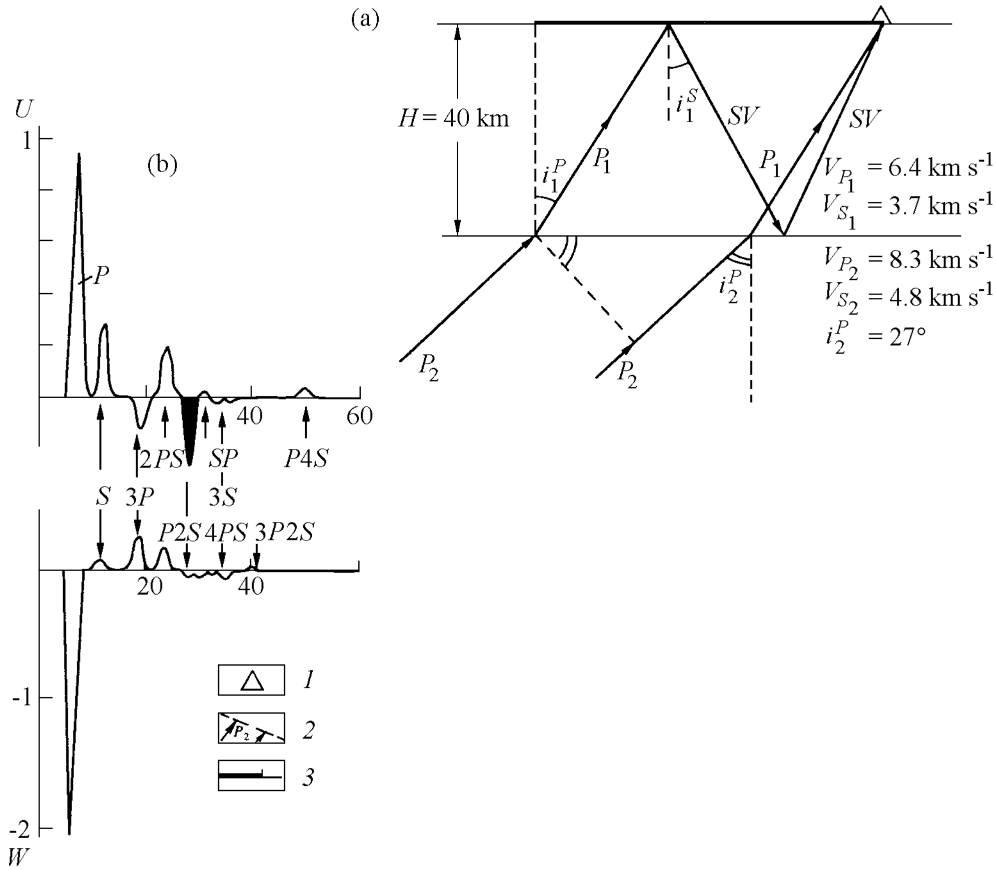


**Figure 3.** (a) Reflection of converted waves in the crust and (b) surface oscillations produced by a longitudinal wave (bell-shaped pulse): 1, seismic observation site; 2, longitudinal wave front; and 3, projection of formation zone model of spectral ratios. Shaded area is a pulse $P_2P_12S_1$ corresponding to Figure 3a; $U$ and $W$ are the horizontal and vertical oscillation components, respectively.

$R(f)$ in more complicated cases, when the distance between the adjacent peaks is not constant. Then $\overline{\Delta f}$ is determined from 10–15 peaks, i.e., in a greater frequency range.

The average velocity $V_P$ in the Southern Siberia crust, obtained by the DSS technique [*Krylov et al.,* 1981; *Yegorkin et al.,* 1984], varies in fairly narrow limits: 6.3–6.4 km s$^{-1}$. The velocity ratio is $V_P/V_S = 1.73 - 1.75$. Therefore the greatest error in the velocity $V_S$ assumed for different regions is $\pm 2.5\%$. Then the estimated average thickness of the crust should not be greater than $\pm 1$ km, provided $R(f)$ was reliably determined from (5). We calculated $H$ for each of the observation sites as an average of a few events over the spectral ratio, or as an average over different events. In the latter case, the maximal scatter in the experimental estimates of $H$ did not exceed $\pm 2$ km.

Results of similar determinations are in good agreement with DSS data. For example, data on $R(f)$ give the crustal thickness 38 km in the Novosibirsk region [*Mordvinova,* 1988], 39 km in Kabansk, and 48 km in Zakamensk; for comparison, we have 38-, 37-, and 46-km thicknesses, respectively, for these regions from DSS data [*Krylov et al.,* 1981; *Puzyrev and Krylov,* 1977].

We accepted the value 3.7 km s$^{-1}$ for the transverse wave velocity in the crust of the profile segment involved ($V_P$ = 6.4 km s$^{-1}$); this value is compatible with DSS data for the southern Baikal and Zabaikal'ye zones. On the basis of the resemblance to geological structures of the observation region and Zabaykal'ye zone, the same velocity value was assumed for Mongolia. The incidence angle $i_1^S$ was determined from its relation to the epicentral distance for each particular case.

The estimates obtained from (5) are confirmed by comparing the experimental ratios $P(f)$ with the theoretical values found from the given velocities $V_P$ and $V_S$ and calculated $H$. The average density of the crust (the fourth parameter necessary to solve the forward problem) was calculated from the formula suggested by *Semenova* [1978] for rocks without melts:

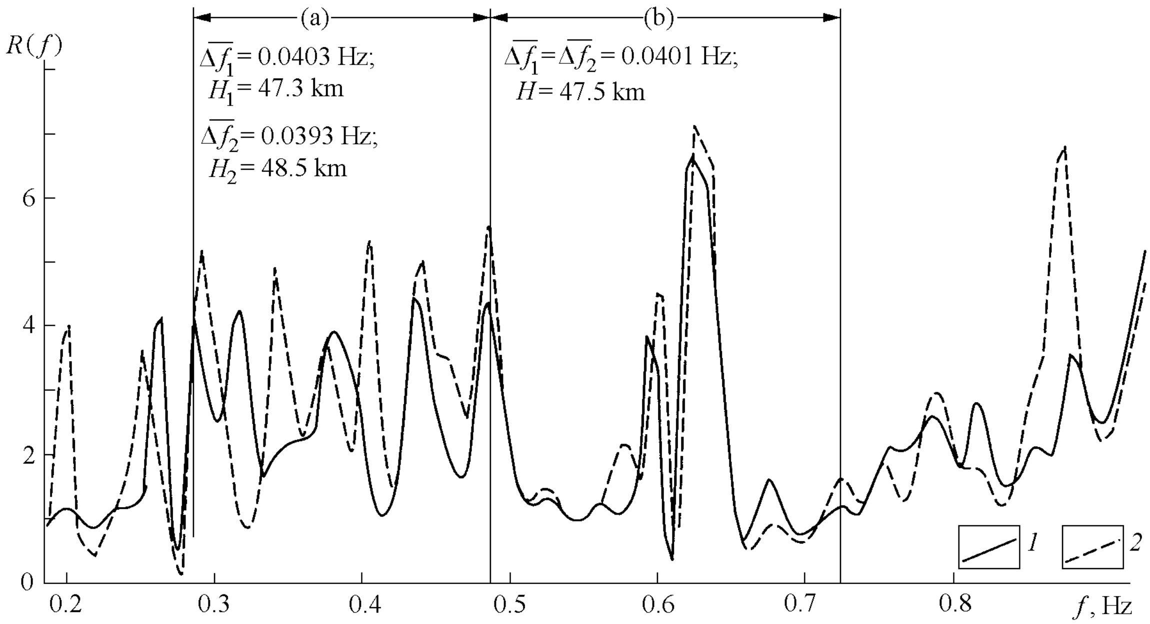


**Figure 4.** Comparison between spectral ratios from broadband (1) and short-period (2) instruments. Station 36, earthquake on July 14, 1992.

$$\rho = 0.292 V_P + 0.929 \quad (6)$$

Calculation of the theoretical ratio in question also showed that the sediment effect is negligible in the region, since the sediments in the mountain range are of small thickness, and seismic velocities in the ancient cover of the Siberian platform are slightly distinct from those in the basement.

Figure 5 demonstrates the coincidence between the positions of the maximum ratios determined by the travel times. Achieving coincidence for the peal level is also possible by fitting detailed models of the velocity structure. Development of such models for Mongolia is our future problem. However, preliminary estimates of the crustal thickness are also useful for fitting a layered model, since they give additional restrictions on possible interpretation variants.

## Discussion of Results

Before turning to discussion of our results, we note that the estimated thickness of the crust is referred to an area somewhat displaced from the station in the direction of the seismic events we used, not directly to the observation point. Such displacements in our case may be 25–35 km (depending on the epicentral distance) toward the south-southeast (Figure 1).

We see from Table 2 and Figure 6 that the crustal thickness in the 900 km profile segment essentially varies only near the rift zone. The crust is 42 km thick under the border of the platform, becomes thinner to 38 km under Baikal Lake, and is thickened to 49 km under the Khamar-Daban Range. Further away the thickness is diminished and reaches 42 km in northern Mongolia. The latter value is retained over almost all the rest of the profile, and only in a 130-km interval, near Ulan-Bator, does the thickness increase to 43–44 km. This latter interval corresponds approximately to the south-western closing of the Khentey arched uplift. The results obtained agree with DSS data [*Krylov et al.,* 1981] for the Siberian platform and with surface wave data [*Kozhevnikov et al.,* 1992] for southern Mongolia.

Figure 6 also shows the crustal thickness inferred previously from the correlation of DSS data for South-eastern Siberia with the so-called effective relief heights [*Zorin et al.,* 1988]. The relief heights were obtained by adding the topographic heights averaged over the $30 \times 30$ km areas to their increments such that loads are equivalent to the upper crust density heterogeneities estimated from local (decompensation) gravity anomalies [*Zorin et al.,* 1988, 1990]. From Figure 6 one can see that these estimates of the crust thickness differ not more than 2–3 km from our seismic results described above. Thus it can be said now with assurance that the relation between the topographic heights and crustal thickness, which was previously established for Eastern Siberia, remains valid for the Mongolia territory. The basic relief forms in Central Asia were developed during the Late Cenozoic [*Florensov,* 1978; *Logatchev and Zorin,* 1992]. Therefore we feel that the basic structure

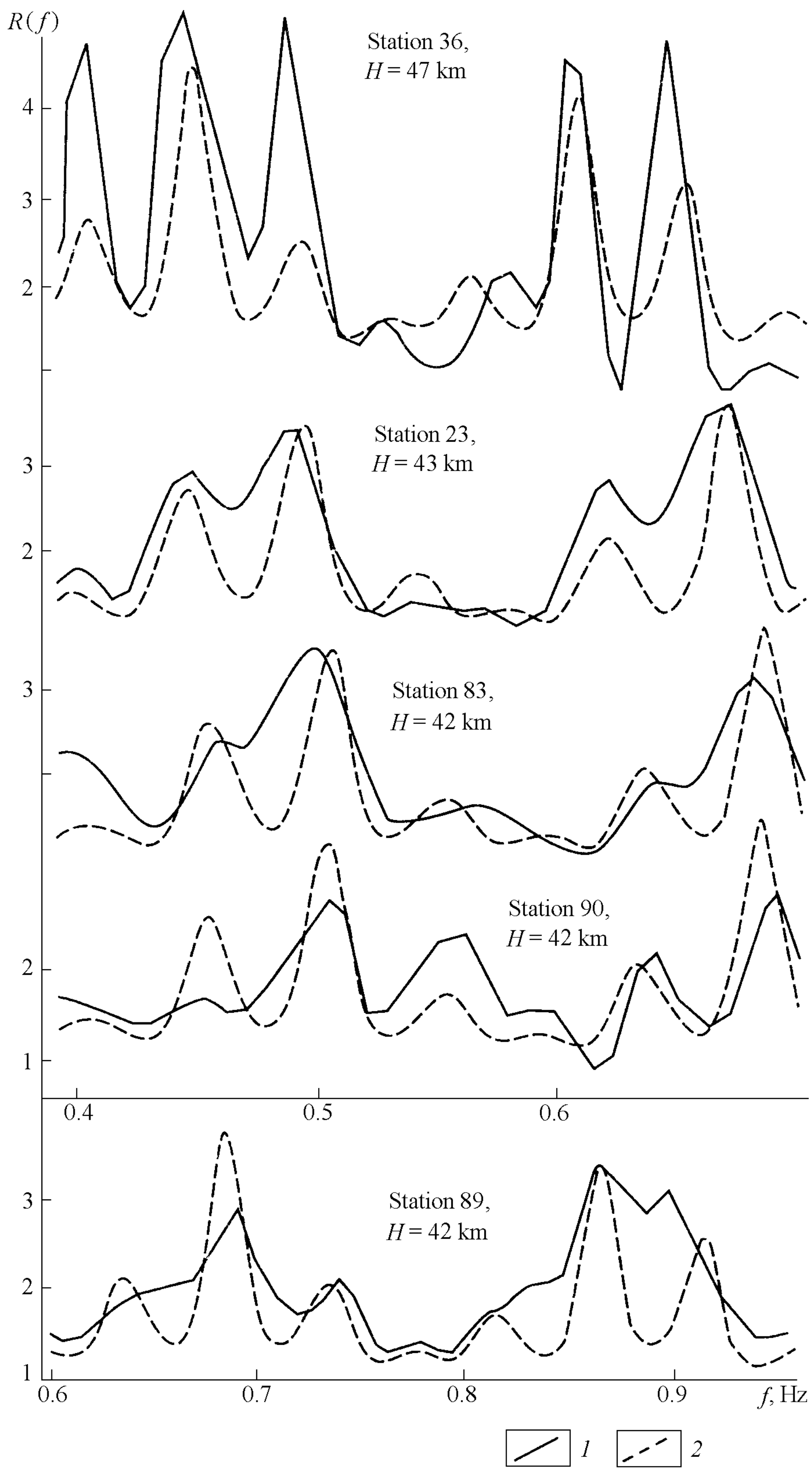


**Figure 5.** Comparison between average observed (1) and theoretical (2) ratios for one-layer models described in text.

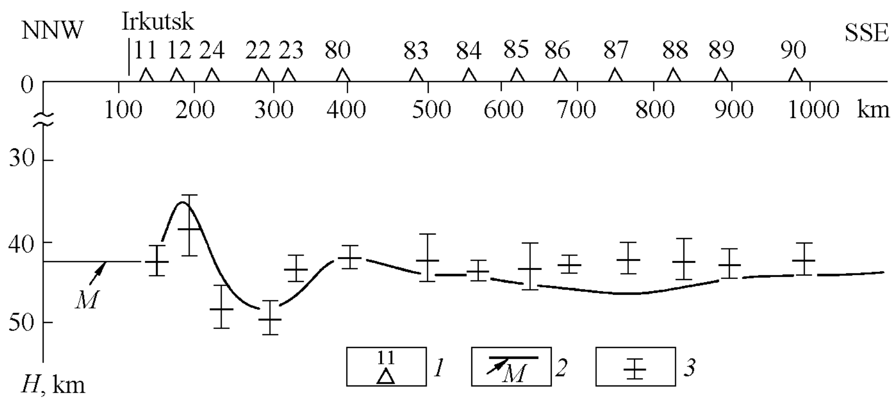


**Figure 6.** Estimated thickness of the crust: 1, digital seismic stations (1992); 2, Moho [*Zorin et al.,* 1990]; 3, crustal thickness determined from spectral ratio for body waves; bars show the 95% confidence intervals (see Table 2).

of the Moho boundary was finally formed at the same time.

However, the amplitudes of the crustal thickness variation were found to be insufficient for the explanation of the observed isostasy [*Zorin,* 1971; *Zorin et al.,* 1982, 1988]. Thus change in the deep structure of the lithosphere including the crust and upper mantle played an important role in formation of the recent tectonic basis for the present relief.

**Table 2.** Thickness of the crust ($H$) from spectral ratio

| Station Number | Station Coordinates | | $H$, km | $\pm\sigma$, km |
|---|---|---|---|---|
| | $\varphi_s$, deg | $\lambda_s$, deg | | |
| 11 | 52.168 | 104.385 | 42.0 | 1.0 |
| 12 | 51.847 | 104.893 | 37.5 | 2.0 |
| 22 | 51.022 | 105.682 | 49.5 | 1.0 |
| 23 | 50.790 | 105.970 | 43.0 | 0.8 |
| 24 | 51.527 | 105.122 | 49.0 | 1.4 |
| 36 | 51.148 | 106.025 | 47.3 | 1.0 |
| 80 | 50.193 | 106.253 | 42.1 | 0.8 |
| 83 | 49.288 | 106.412 | 42.3 | 1.5 |
| 84 | 48.932 | 106.682 | 43.7 | 0.5 |
| 85 | 48.383 | 106.783 | 43.4 | 1.6 |
| 86 | 47.955 | 106.955 | 42.5 | 0.5 |
| 87 | 47.208 | 107.422 | 41.9 | 1.0 |
| 88 | 46.635 | 107.758 | 41.8 | 1.2 |
| 89 | 46.115 | 107.620 | 42.3 | 0.9 |
| 90 | 45.262 | 108.260 | 42.0 | 1.1 |

$\sigma$ is error in $H$.

## Conclusions

Both theory and experiment show that the thickness of the crust can be determined from the positions of peaks of the spectral ratio of body waves without fitting seismic section parameters. The crust thickness estimated by this method agrees with estimates previously obtained from correlation of DSS data for Southeastern Siberia with the effective height of the present relief, whose basic forms were developed during the Late Cenozoic. Therefore it is believed that Late Cenozoic processes played an important role in the formation of the present crustal structure (at least of their areal thickness variations). These processes are, first, thinning of the Baikal crust due to stretching of the crust and some thickening of the crust under the Khentey archbend due to crustal contraction, or of magmatic melts forming at depth.

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