REFLECTIONS FROM THE CORE DURING ATOMIC EXPLOSIONS AND PROBLEMS OF THE EARTH’ S INTERNAL STRUCTURE

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Abstract

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REFLECTIONS FROM THE CORE DURING ATOMIC EXPLOSIONS AND PROBLEMS OF THE EARTH' S INTERNAL STRUCTURE

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The problem of the distribution of density with depth is one of the most difficult in the physics of the Earth. In constructing specific sections, a number of hypotheses are used; as a result there are models that differ substantially from one another at depths greater than 2000 km; nevertheless, practically all competing models assume a jump-like increase in density at the core-mantle boundary by approximately a factor of 1.7. Such, in particular, is Bullen' s model A′, which best accords with present-day ideas and has the following parameters in the vicinity of the core boundary: mantle $v_p = 13.6$ km/sec, $v_s = 8.1$ km/sec, $\rho = 5.66$ g/cm^3; core $v_p = 8.1$ km/sec, $\rho = 9.7$ g/cm^3. In what follows we shall call this model the standard one.

Reflections from the core attract attention because they make it possible to obtain direct data on the parameters of the media on both sides of the boundary. An extensive literature is devoted to the study of these waves ([1–6] and others). An insurmountable difficulty in interpreting the observations is presented by the inexplicably large values of the ratio PcP/P (the amplitudes of the longitudinal reflected and refracted waves) and, consequently, of the reflection coefficient from the core boundary.

Analyzing records of atomic explosions obtained by seismic stations of the USSR, we came to the conclusion that the inexplicably large values of PcP/P, which gave rise to the discussion, are the product of incorrect processing of the observational results. The scatter of individual values of the quantity PcP/P is so great that on half of the seismograms the PcP wave is not distinguished at all, whereas on others it is comparable in amplitude with P. A substantial improvement of the method used up to now consists in the fact that records of the first group do not in any way participate in the formation of the statistics. In the past, when earthquakes were the main source of seismic signals, cases of absence of the PcP wave were attributed to asymmetry of the source and were not taken into account ([5]). However, the experience now available from observations of symmetric sources (atomic explosions) shows that such a view

Fig. 1

Figure 1: Fig. 1

is in principle incorrect, and the cause of the scatter of seismic-wave amplitudes should be sought elsewhere. The obvious cause of such scatter is random inhomogeneities of the medium. The theory of wave fields in media with random inhomogeneities has been considerably advanced in acoustics and optics ([7,8]); a model of this type has been used in interpreting TSS data ([9]).

The initial material for our analysis consists of about 60 independent records of explosions. We consider records from one station to be independent if they refer to explosions at different test sites, and records of explosions from one test site at different stations. If several explosions from one test site were recorded at one station, one of them is included in the statistics. Let $\lambda = A^0_{PcP}/A^0_P$ be the ratio of the amplitude of PcP to P at the Earth' s surface that would be observed if the medium were homogeneous, and $\beta = A_{PcP}/A_P$ the result of an individual measurement, as a rule differing

from λ. As a result of measuring the seismograms, data of two types were obtained: in N cases, when PcP is identified, the quantity β_i is measured; in those M cases when PcP is not identified, α_j is measured–the ratio of the background amplitude to the amplitude P. It is assumed that $\alpha_j > \beta_j$. The measurements were made for $29^\circ \leq \Delta \leq 71^\circ$; the results are shown in Fig. 1. In a medium with random inhomogeneities the amplitude of the signal is distributed lognormally ([8]). One can estimate, for $\ln\beta$, the mean μ and variance σ^2 by constructing the logarithmic likelihood function L

$$L = -\sum_{i=1}^{N} \frac{(\ln\beta_i - \mu)^2}{2\sigma^2} - N\ln\sigma + \sum_{j=1}^{M} \ln\left[1 + \Phi\left(\frac{\ln\alpha_j - \mu}{\sqrt{2}\sigma}\right)\right] + c, \qquad (1)$$

where

$$\Phi(x) = \frac{2}{\sqrt{\pi}} \int_0^x e^{-t^2}\, dt;$$

c is a constant chosen in such a way that $L_{\max} = 0$. The best estimates, in a certain sense, of μ and σ correspond to $L_{\max}$ ([10]).

Fig. 1

Figure 2a shows function (1) for the sample $\Delta \sim 29$–40°. The estimates of μ and σ are -2.1 and 1.2, respectively. For distances 40–50° and 60–71° very similar estimates and approximately the same confidence regions were obtained. For distances 50–60°, $\mu \approx -1.75$, $\sigma \approx 1.75$ were obtained; however, in this case the confidence regions are much larger, indicating a significantly lower reliability of

Fig. 2

Figure 2: Fig. 2

the result. Naturally this result should be excluded from further analysis. The remaining results allow one to consider that μ and σ practically do not depend on Δ. Therefore we combined all the data for $\Delta \sim 29$–50° and 60–71° into one sample, for which function (1), shown in Fig. 2b, was constructed; for μ and σ the estimates -2.2 and 1.3 were obtained.

Fig. 2

For any wave (for example P) the relation ([8]) holds

$$\sigma_P^2 = -\bar{\chi}_P = -\overline{\ln A_P} + \ln A_P^0,$$

where the overbar denotes averaging. Hence

$$\lambda = \exp(\mu + \sigma_{PcP}^2 - \sigma_P^2).$$

It is obvious that $\sigma^2 = \sigma_{PcP}^2 + \sigma_P^2$; to determine λ it is necessary to obtain σ_{PcP}^2 and σ_P^2 separately. To estimate σ_P^2 we used data on the amplitudes of P in the Bilby and Longshot explosions, published in ([11]). The results of our processing are presented in Table 1 in the form of point estimates and 60% confidence intervals.

Table 1

Δ, degrees	25–45[1]	67–87[2]	50–70[3]	50–70[4]	40–60[5]	40–50[6]
	0.3	0.5	0.31	0.42	0.22	0.65
σ_P^2	0.37	0.64	0.4	0.5	0.28	0.76
	0.48	0.78	0.58	0.61	0.39	0.95

[1] Bilby. [2] Bilby. [3] Longshot, WWSSN network. [4] Longshot, stations in Europe and North America. [5] Longshot, LRSM network. [6] Longshot, stations in Europe and North America.

The data of Table 1 show that σ_P^2 does not reveal any noticeable dependence on Δ; for further calculations we take $\sigma_P^2 = 0.45$. Then $\sigma_{PcP}^2 = 1.25$, $\lambda = \exp(-1.4) \approx 0.25$. For comparison with theory, it is necessary to convert the quantity λ into the reflection coefficient r by the formula

$$r = \lambda b \exp[-\pi f H],$$

where

$$H = \int_{\text{ray } P} \frac{ds}{vQ} - \int_{\text{ray } PcP} \frac{ds}{vQ}.$$

Here b is a geometrical factor; H takes account of the effect of absorption; f is the frequency, taken to be equal to 1 Hz; v is the velocity of longitudinal waves; Q is the quality factor of the medium for longitudinal waves. The quantity b for $\Delta \sim 30$–70° is close to 2. The distribution of Q adopted by us (see Table 2) is close to the values obtained by Anderson et al. ([12]) for long-period transverse waves.

Table 2

Depth of the top of the layer, km	0	100	400	500	600	700	800	900	1000
Depth of the bottom of the layer, km	100	400	500	600	700	800	900	1000	2898
Q	228	150	160	180	250	450	500	600	2300

The grounds for this choice are as follows. According to (6), a distribution of Q of this kind satisfactorily explains the difference between the spectra of P and PcP at frequencies of 1–0.5 Hz. A similar distribution of Q was obtained by Mikumo and Kurita ([13]) from the spectra of long-period longitudinal waves and by I. P. Pasechnik ([14]) from the spectra of longitudinal waves from underground explosions.

In Fig. 3, line I shows the value of r obtained with the aid of (2). This curve differs substantially from curve II—the reflection coefficient for the standard model. We shall assume that this discrepancy is not entirely the result of errors in the estimate of r. The theoretical reflection coefficient can be reduced by assuming that the core material has a shear modulus of considerable magnitude;

Fig. 3

Figure 3: Fig. 3

however, this assumption is difficult to reconcile with a number of other seismic results. One may suppose that the surface of the core is not a sharp boundary between two media, but a thin-layered transition zone several kilometers thick; however, the results of (6) make this assumption unlikely. One may allow that at the base of the mantle there is a low-velocity zone about 30 km thick ([15]), but, in combination with the density jump at the core boundary, this model does not yield a reduction of the reflection coefficient in comparison with the standard one. Let us suppose, finally, that at the core-mantle boundary there is a velocity jump, while the density jump is absent. Correspond-

···the curve III corresponding to this model is in excellent agreement with curve I. This assumption does not contradict any seismological data; indeed, quite the contrary. A model without a density discontinuity implies a phase reversal of the PcP wave at $\Delta < 30^{\circ}$; precisely such a phenomenon was observed by Buchbinder in analyzing records of the Longshot explosion [16]. True, the fact of a phase reversal, taken by itself, as determined from the signs of the first arrivals, is not very reliable: the PcP wave arrives against a disturbed background, when it is difficult to say exactly which phase of the record is the first arrival. Nevertheless, it is unlikely that such an agreement of the results was accidental. The model under consideration allows certain modifications; in particular, one may admit the existence of a layer of reduced velocities at the base of the mantle; it is impossible to distinguish this modification (Fig. 3, curve IV) from our data.

Fig. 3

A second important conclusion may be obtained from comparing σ_P^2 and σ_{PcP}^2; according to our data, σ_{PcP}^2 is approximately three times greater than σ_P^2. The increase in the path traversed by PcP in comparison with P cannot be the cause, since σ_P^2 does not increase with distance. Evidently, the point is that the main inhomogeneities responsible for the scatter of the amplitudes of P are located in the crust and upper mantle, so that although the path traversed by this wave increases together with Δ, the length of the active segment is practically constant. Then, in order to explain the greater scatter of the PcP amplitudes, it is necessary to assume that at the base of the mantle there is a second, strongly inhomogeneous zone. A qualitatively similar effect may be caused by irregularity of the core boundary.

Analysis of reflections from the core during explosions makes it possible to obtain an idea of the characteristic scales of mantle inhomogeneities. According to (7), for a Gaussian correlation function of the inhomogeneities, the correlation interval of the amplitude fluctuations coincides with the correlation interval of the inhomogeneities. Analysis of records from one and the same station, when the epicenters are scattered over distances on the order of the wavelength (10 km), shows that in this case the quantity β changes insignificantly. Apparently, the characteristic scale of the inhomogeneities is considerably greater than 10 km.

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