

MAPPING THE LOWERMOST MANTLE BY MEANS OF DIFFRACTED WAVES

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A B S T R A C T

Diffracted seismic waves bear information about the base of the mantle and the core-mantle boundary. Digitally recorded P_{diff} phases show distinct dispersion in dependence on the source-receiver configuration. For the source region of Fiji Islands azimuthal variations of the frequency dependent velocities of P_{diff} are observed. The comparison with modeling results enables to indicate regions with different seismic velocities and velocity gradients. For diffracted S-waves a splitting of vertically and horizontally polarized waves is observed depending on the path at the core-mantle boundary. Results indicating a laterally heterogeneous structure of the core-mantle transition region are presented.

ΧΑΡΤΟΓΡΑΦΗΣΗ ΤΟΥ ΚΑΤΩΤΕΡΟΥ ΜΑΝΔΥΑ ΜΕ ΚΥΜΑΤΑ ΠΕΡΙΘΛΑΣΗΣ

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Π Ε Ρ Ι Λ Η Ψ Η

Τα κύματα περίθλασης περιέχουν πληροφορίες για τη περιοχή του κατωτάτου τμήματος του μανδύα και το όριο μανδύα-πυρήνα. Ψηφιακές καταγραφές κυμάτων P_{diff} παρουσιάζουν σκέδαση που εξαρτάται από τη θέση πηγής-σεισμομέτρου. Παρατηρούμε αζιμουθιακή μεταβολή των ταχυτήτων που εξαρτώνται από την συχνότητα των κυμάτων P_{diff} για σεισμούς της περιοχής των νησιών Fiji. Η σύγκριση με θεωρητικά μοντέλα μας δείχνει περιοχές που έχουν διαφορετικές ταχύτητες και διαφορετικές μεταβολές των ταχυτήτων με το βάθος. Παρατηρούμε διαχωρισμό των επιπέδων πόλωσης των οριζόντια και κατακόρυφα πολωμένων και περιθλωμένων S κυμάτων που εξαρτάται από το δρόμο διάδοσης στο όριο μανδύα-πυρήνα. Τέλος παρουσιάζονται αποτελέσματα που δείχνουν μιά πλευρικά ανομοιογενή δομή στην περιοχή του ορίου μανδύα-πυρήνα.

INTRODUCTION

One of the most important results of seismological and global geophysical research is the creation of spherical symmetric Earth models. The core-mantle boundary (CMB) and the inner core boundary (ICB) are principal features of these models. They are influenced by the evolution of the Earth and by processes at great depths. Recent investigations have shown that dynamic

processes near the CMB are closely related to possible inhomogeneous structure of the core-mantle transition (Stacey and Loper, 1983; Knittle and Jeanloz, 1989, 1991). Therefore one of the actual tasks of seismology is the resolution of possible inhomogeneous structures of the Earth's interior, especially of core-mantle transition. Seismological studies of the CMB and the ICB are based mostly on the analysis of reflected and refracted waves (Lay and Helmberger, 1983; Dziewonski, 1984; Morelli and Dziewonski, 1987; Weber and Kornig, 1992) and on free oscillations (Kumagai et al., 1992). Alexander and Phinney (1966), Chapman and Phinney (1972), Mula and Muller (1980), Mula (1980) and Wysession et al. (1992) have used diffracted P- and S-waves for studying the CMB. Although diffracted waves are complicated for observation and analysis they have the advantage of long paths near the CMB. Recent studies indicate a complicated structure of the lowermost mantle (D"-layer) (see e.g. Weber and Kornig, 1992; Wysession et al., 1992). This study is directed to the problem of using asymptotic methods of modeling diffracted waves for analysis of corresponding data.

METHOD

In this study theoretical modeling of diffracted waves is performed by means of the so-called asymptotic boundary layer method. The principal approach of the method includes not a direct solution of the wave equation for interference waves traveling along the curved interface of inhomogeneous media. A high frequency approximation of these wave phenomena is chosen, describing the propagation of inhomogeneous waves in a layer of thickness of about $O((2\pi f)^{-2/3})$ near the curved interface. The solution of this approximation is extended by means of ray theory into the whole Earth model. A detailed description of the scalar approach is given by Babic and Buldyrev (1991). The propagation of interference waves on a curved interface between a solid and liquid was firstly described with the help of the asymptotic boundary layer method by Molotkov and Krauklis (1971). Krauklis and Kowalle (1989), Krauklis et al. (1991) and Kowalle and Krauklis (1992) have applied this method for modeling propagation of diffracted waves at the core of an inhomogeneous Earth model. Doornbos (1992) has proposed to use diffraction for seismic tomography. By means of the asymptotic boundary layer method it will be possible to include into consideration inhomogeneous waves with different velocities (P-, S-, and surface wave velocity) propagating in the vicinity of the inhomogeneous core-mantle transition. The general solution is obtained in the form

$$U_i = \sum_n U_i^{(n)}(\tau, \alpha, \sigma) \exp[i2\pi f\tau + (2\pi f)^{1/3}\delta(\alpha, \sigma) + 2\pi f\mu\phi(\alpha, \sigma)], \quad (1)$$

where τ, α, σ describe the curved coordinate system connected with the interface, f is the frequency, $U_i^{(n)}$, δ, ϕ are functions to be determined. After substituting (1) into the equation of motion and the boundary conditions, one obtains a system of equations for the unknown functions. The effective radius of curvature of the interface is determined by

$$R_{eff}^{-1} = \frac{1}{R} - grad(v) / v \quad (2)$$

where R is the radius of the core, v is the P-or S-velocity and grad(v) the respective vertical velocity gradient. The amplitude of a wave traveling along the interface is given by

$$U_m = \frac{1}{G} \exp \left[i 2 \pi f \int \frac{ds}{v} - i (\pi f)^{1/3} j_m \int \frac{ds}{(v^{1/3} R_{eff}^{2/3})} - \int \frac{\Theta}{R_{eff}} ds \right] \quad (3)$$

with G - geometrical spreading function, j_m - the m-th root of the Airy-equation,

$$\Theta = \frac{\beta^3 (1 - \beta^2)^{1/2}}{(1 - 2\beta^2)^2}, \quad \beta = \frac{v_s}{v_p} \quad (4)$$

The integration of (3) is performed along the wave path at the CMB. Due to the fact that velocities and gradients are functions of coordinates, inhomogeneous models can be treated. It is evident that the amplitude is influenced not only by geometrical spreading but also by the last term of equation (3). This term describes the radiation of waves into the above medium. The wave propagation velocity is determined by the first two terms of (3). It is obvious that the velocity will be frequency dependent. Due to the asymptotic approach the expansion of the solution is provided by terms of $(vT/R_{eff})^{n/3}$ and the above formulae will be applicable only for values $(vT/R_{eff}) \ll 1$.

MODELING

The principal situation at the CMB is described by solid lower mantle and a liquid outer core. Using the above formulae one is able to determine the seismic velocities of the waves traveling along the CMB. P_{diff} is the longitudinal diffracted wave at the Earth's core with

$$v = v_p \left[1 - \frac{j_1}{q} \left(\frac{v_p T}{2\pi R_{eff}} \right)^{2/3} \right], \quad (5)$$

where j_1 is the first root of the Airy-equation, and q is a constant determined by the expansion. Fig.1 shows the behaviour of P_{diff} -velocity for different models of D". There is found a decrease of the velocity of diffracted P-waves with increasing period. This dependence is strongly influenced by the vertical velocity gradient in the layer just above the core-mantle boundary. If there is a positive gradient, the influence will be weaker than for a negative gradient.

The analysis of (5) shows that the P_{diff} velocity is influenced mainly by changes of the P-velocity and the velocity gradient in the D"-layer. Variations of the core radius show smaller influence (Tab. 1). These results are in agreement with those of

Doornbos (1992). P_{diff} -velocity decreases with period. Positive P-velocity gradients in the D"-layer will lower the period dependence of the P_{diff} -velocity, whereas negative gradients will produce larger influence of period.

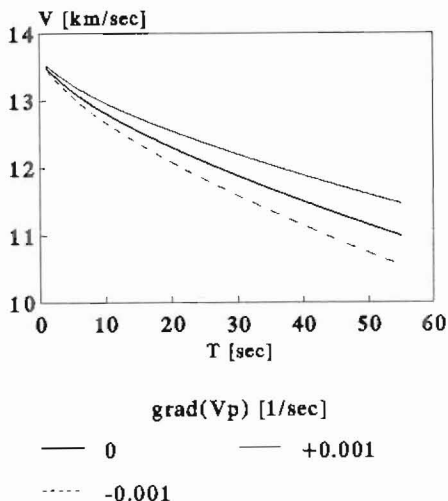


Fig.1. Dependence of P_{diff} -velocity on period for models of core-mantle transition with different vertical P-velocity gradients in the D"-layer.

Table 1. Factors influencing the velocity of diffracted core phases.

factor	change in %	influence on diffracted waves (%)
velocity in D"	1	1
velocity gradient in D"	1	10
core radius	1	10 ⁻⁵

The same procedure is applicable to diffraction of S-waves. S_{diff} -velocities are determined by the following formulae:

$$v = v_s \left[1 - \frac{j_1'}{Q_2} \left(\frac{v_s T}{2\pi R_{eff}} \right)^{2/3} \right], \quad \text{for } SH_{diff}, \quad (6)$$

$$v = v_s \left[1 - \frac{j_1'}{Q_3} \left(\frac{v_s T}{2\pi R_{eff}} \right)^{2/3} \right] - W \left(\frac{v_s T}{2\pi R_{eff}} \right), \quad \text{for } SV_{diff}, \quad (7)$$

$$W = \frac{4(1-b^2)^{1/2}(1-c^2)^{1/2}}{(1-c^2)^{1/2} + d(1-b^2)^{1/2}} \quad (8)$$

Constants are indicated by q_2 and q_3 . The factor j_1' represents the first root of the second Airy-equation. The ratio of S- to P-velocity in the D"-layer is denoted by b , that of S-velocity in D" to P in the outermost core by c , and that of core density to density of D" is d . From formulae (6) and (7) it is obvious that horizontally and vertically polarized diffracted S-wave propagate with different velocities. For every period T velocity of SH_{diff} is larger than that of SV_{diff} . The principal behaviour of velocities of S_{diff} is quite similar to that of P_{diff} (Fig. 2). Between SH_{diff} and SV_{diff} there is one significant difference. SH_{diff} is influenced only by properties of D", whereas SV_{diff} is also affected by physical properties of the outermost part of the liquid core. If there exists a possibility for recording and analysing of SH_{diff} and SV_{diff} separately, one will be able to investigate by means of diffracted waves not only the structure of the D"-layer but also of the outer core.

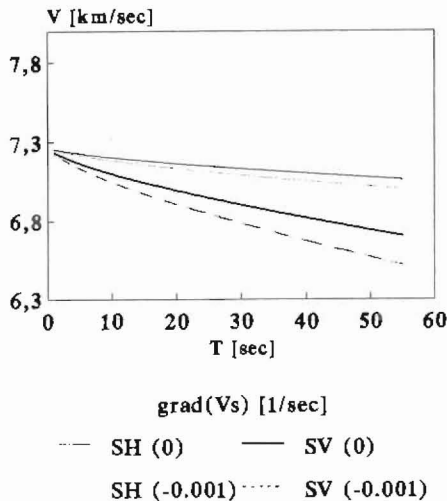


Fig.2. S_{diff} velocities for models with different S-velocity gradients in D". Due to different velocities of SV_{diff} and SH_{diff} a splitting of S_{diff} occurs.

Similar to the case of P_{diff} a negative vertical gradient of v_s in the region above the CMB will cause strong decrease of the velocity of S_{diff} -velocity with period. This dependence is much more pronounced for the vertically polarized S_{diff} -phases (SV_{diff}).

DATA

In this study about 500 seismograms of deep earthquakes recorded by the Global Digital Seismological Network (GDSN) were

analyzed. Earthquakes with focal depth greater than 450 km have been chosen to avoid interference of phases of interest, especially S_{diff} with SKS, SKKS, and the corresponding surface reflections. For determining the dependence of P_{diff} -velocity on period the seismograms have been band pass filtered. The central frequencies of the filters were equal to 0.25 Hz (4 sec), 0.125 Hz (8 sec), 0.0625 Hz (16 sec), 0.0417 Hz (24 sec), 0.03125 Hz (32 sec), and 0.02083 Hz (48 sec), respectively. The corresponding onset times of the diffracted P-waves were picked from the filtered seismograms. The data have been grouped in accordance with the focal region and the azimuth. Azimuth ranges of about 20 degrees have been found optimal for this data set. Smaller ranges would contain less data and larger ones would smear out the chosen effect of azimuthal dependence. Fig.3 shows the dispersion curves for different azimuth ranges for Fiji Islands source region. The experimentally determined dispersion curves have been compared with theoretically computed ones. For the computations the asymptotic boundary layer method and the IASP91 model have been used. The velocity within the D"-layer was determined by the intercept value of the dispersion curve with the v-axis ($T=0$). The velocity gradient within D" is forced by the slope of $v(T)$. For the source region of Fiji-Islands for the azimuth of 45 deg (station pair RSON-GAC) a P_{diff} -velocity of 13.67 km/sec in the D"-layer and a vertical velocity gradient of zero could be determined (model 1). This represents the CMB-structure under the Pacific ocean near the west coast of North America.

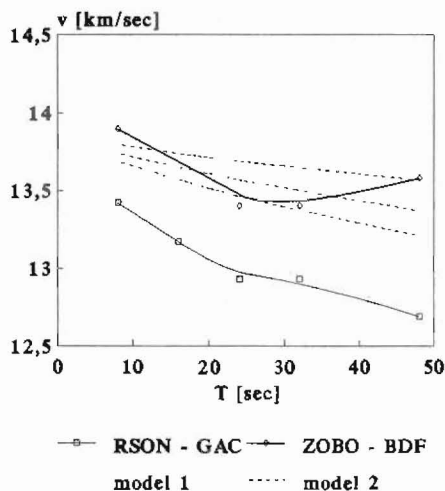


Fig.3. Azimuthal variations of P_{diff} dispersion for deep earthquakes at Fiji Islands (azimuth 45° - RSON-GAC, azimuth 350° - ZOBO-BDF).

For 350 deg azimuth (station pair ZOBO-GAC) a slightly positive gradient of +0.0002 sec⁻¹ to +0.0004 sec⁻¹ (model 2) fits the

observational data. The P_{diff} -velocity is increased by 1.5% in this direction. This indicates a slightly higher vertical P-velocity gradient within the D"-layer under Asia and southern Siberia. By means of P_{diff} , Wyss et al. (1992) have determined an increased P-velocity in the D"-layer under the Arctic region, under Asia, and southern Siberia. Under the Pacific ocean near the west coast of Northern America they have found a normal v_p within the D"-layer. For other source-receiver configurations a similar azimuthal dependence of the structure of D" has been found. These results are in a good agreement with observations of S_{diff} -splitting.

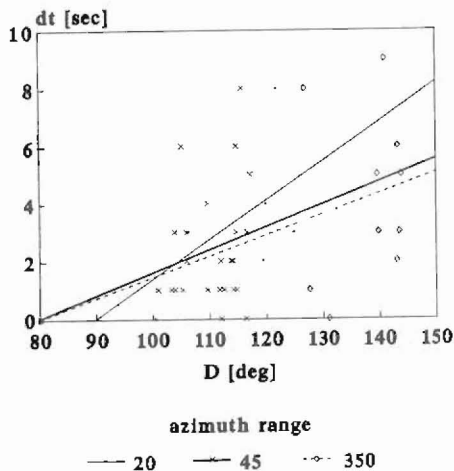


Fig.4. Splitting of S_{diff} for the source region of Fiji (dt is the time difference between SH_{diff} and SV_{diff}).

To find the splitting the seismograms have been transformed into ray-centered coordinates. Then the different polarized S_{diff} phases have been analyzed visually and also by matching. A dependence on azimuthal source - receiver configuration for the source regions of Fiji and South America was observed. For 45 deg azimuth of the source region of Fiji the vertical S-velocity gradient in the D"-layer (Pacific ocean near the west coast of North America) is zero, for the azimuth of 20 deg of about -0.002 sec^{-1} , and for 350 deg azimuth the vertical v_s -gradient is of about -0.0004 sec^{-1} . The stronger negative vertical v_s -gradient indicates an anomalous structure of the core-mantle transition (D"-layer) under the Arctic Region. For the azimuth direction of 45 deg the vertical P- and S-velocity gradients in the D"-layer are similar, whereas in the 350 deg azimuth an opposite behaviour of P- and S-velocity gradients is observed.

The observed S_{diff} splitting times for earthquakes in South America show also azimuthal dependence. The D"-layer under the southern Atlantic ocean shows a strong negative vertical v_s -gradient (azimuth 40 deg.), whereas under the Pacific ocean

(azimuth 210 deg) the gradient is slightly positive. The analysis of S_{diff} provided by Wyssession et al. (1992) has shown an increased v_s under the Pacific ocean and southern Asia. For the Arctic region a normal v_s was determined. In general the coverage of collected data for selected source regions as well as in the global scale is poor. Due to the lack of good data the mapping of the behaviour of velocities and velocity gradients within the D"-layer seems to be complicated. For the Pacific and Atlantic region different behaviour of P_{diff} as well as of S_{diff} has been found, indicating different structure of the core-mantle transition below these regions.

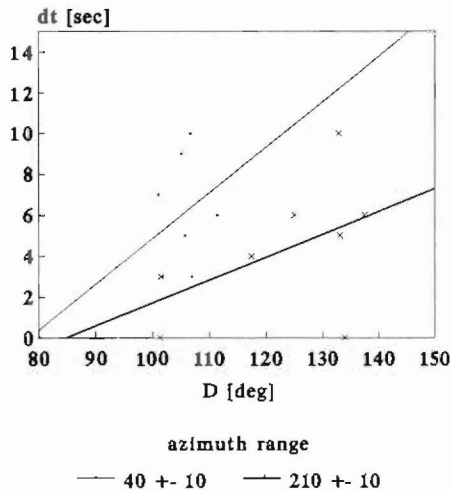


Fig.5. S_{diff} splitting for earthquakes in South America.

CONCLUSIONS

The asymptotic boundary layer method allows to model the propagation of diffracted seismic waves in an inhomogeneous model of the core-mantle transition zone. Inhomogeneities above the core-mantle boundary (D"-layer) are influencing the diffracted P- and S-phases. The vertically polarized S_{diff} phases (SV_{diff}) are also influenced slightly by the structure of the outermost core. Observations of diffracted waves indicate inhomogeneities in the D"-layer. From the data one concludes that within distinct regions of the D"-layer there are different P- and S-velocities as well as corresponding gradients influencing the propagation of diffracted phases. The inhomogeneous structure of the core-mantle transition seems to be a global feature influencing the dynamics of the Earth (e.g. core-mantle coupling). These results are in a good agreement with similar studies of Wyssession et al.(1992). A further improvement of the knowledge of inhomogeneous CMB structures depends on the progress in seismological observations. The number of observations of

diffracted P- and S-phases should be increased, and the global coverage of those observations should be improved.

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