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Seismic Modeling of the Velocity Components at the Sea Floor in Frequency Domain

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Abstract: The seismic modeling in Forward problem consists of generate seismic data from a given model. It is a useful tool to investigate the nature and the distribution of the subsurface. It includes two main parts: the physical modeling and the numerical modeling. These two techniques are also the basic tools, which have brought tremendous progress in solving the inverse problem. The inverse problem consists to build a model from a set of seismic data. The reflectivity method combining the theoretical seismic records is one kind of the solving methods in frequency domain; consequently in this study first of all we need to solve the wave equation in frequency domain, taking into account the appropriate boundary conditions. So we will get in frequency domain the horizontal and the vertical coordinates of the soil particles. Finally we will use the inverse Fourier transform to get the solution of the wave equation in time domain.

Key words: Elastic waves, frequency domain, reflectivity method, numerical modeling, wave equation

INTRODUCTION

Reflectivity method: Doing geophysical exploration, the basic objective is the investigation of subsurface layers, for that, processing the observed data for a given subsurface model is another way to do the seismic inversion. The inverse problem has many solutions; so these solutions can be used to adjust the model's parameters of the subsurface layers built from the error differences of observed values during data processing. So the inversion process almost depends on the solution of forward modeling.

That is to say prior to inversion process we need the forward modeling result. In regards of previous causes, during the forward modeling the variations of seismic wave's field are 10 times more important; in addition to what precedes the singularities of numerical modeling keep the attention of people and its fast development is surprising. In regard to the computing techniques progress, this kind of processes reduces most of the restrictions and really boosts the production.

The reflectivity method introduced by FUCHS is deduced from one kind of the frequency spectrum methods, where in the frequency domain we look for the interval in which there is an appropriate interpretation of boundary conditions and after the use of inverse Fourier transform we get the seismic waves field in time domain. Special methods are needed to calculate the displacement of the seismic wave field. The reason why people are trying to improve the degree of precision in various kinds of special methods of processing data is to get the more realistic result imaging the subsurface layer. The modified reflection coefficients allow overcoming the loss of some significant digits during the computing process. The goal is related to the previous practical methods to make the forward modeling of the seismic wave field components in the sea floor for multi-waves exploration.

Consequently, first of all we need to solve the wave equation in frequency domain taking to account the appropriate boundary conditions to get in frequency domain the Earth surface displacement, then after the inverse Fourier transform gives us the particle displacement or the potential displacement in time domain.

The main task for us is to calculate the reflection coefficients, find the displacement formula and finally the modified reflection coefficients and caring to satisfy the compilation process used[1].

Seismic modeling: After the seismic wave field has been modeled, we will use the algorithm of the reflectivity to build a model of the sea floor media It is necessary to calculate the excitation of the sea floor, the reception conditions and their influence on the displacement and on the potential displacement of the seismic wave field, including the particle velocity field and the component of water pressure. We must pay attention to the seismic wave field characteristics for analysis, then after we deduce same significant conclusions. The analysis of various components of the seismic wave field shows that the record becomes complicated, the main reason is

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essentially due to the fact that the source produces in the sea floor interface same down going propagating waves, these waves are transmitted as down going P-waves and down going SV waves and for each reflection there is wave conversion phenomenon[2].

There is also reflections for the up going P-waves in the sea floor interface which produce down going P-waves and down going P-SV waves, the up going P-SV waves produce the down going SV-P waves and down going P-SV waves. The amplitudes of these waves are directly proportional to the reflection coefficients in the reflector interface. Simultaneously, the sea floor interface reacts as high impedance and the sea water level reacts as a good reflector. So in seawater there are many multiple waves, that is one of the various causes making the seismic records complicated.

BASIC RELATIONS

P and S waves potential: In a homogeneous, isotropic and horizontally layered medium only two kinds of wave propagation exist, namely the P-waves propagation and the S-waves propagation and two elastic constants are needed. In the ith layer the potential of P-waves is given by the first Eq. 1 and the potential of S-waves is given by the second Eq. 1.

The constants of integration A_i, B_i, C_i, D_i depend on the variable k and the layer parameters α_i, β_i ρ_i, d_i and $J_1(r)$ is the first order Bessel function. The first term in each sum of equation (1) describes the wave propagation along the negative part of the Z-axis and the second term describes the wave propagation along the positive part of the Z-axis.

$$\left.\begin{aligned} \bar{\varphi}(r,z,\omega) &= \bar{F}(\omega)\int_0^\infty A_i e^{j\nu_i(z_i - z_{i-1})} j_0(kr)\frac{kdk}{j\nu_1} + \\ &\quad \bar{F}(\omega)\int_0^\infty B_i e^{-j\nu_i(z_i - z_{i-1})} j_0(kr)\frac{kdk}{j\nu_1} \\ \bar{\psi}(r,z,\omega) &= \bar{F}(\omega)\int_0^\infty C_i e^{\frac{j\nu_i'}{(z_i - z_{i-1})}} j_1(kr)\frac{kdk}{\nu_1} + \\ &\quad \bar{F}(\omega)\int_0^\infty D_i e^{\frac{-j\nu_i'}{(z_i - z_{i-1})}} j_1(kr)\frac{kdk}{\nu_1} \end{aligned}\right\} \quad (1)$$

BOUNDARY CONDITIONS

The displacement continuity: The displacement continuity expresses the fact that the displacement above the layer must be equal to the displacement under the layer, as well as for the horizontal component q than for the vertical component w.

$$\left.\begin{aligned} \bar{q}_i &= \bar{q}_{i+1} \\ \bar{w}_i &= \bar{w}_{i+1} \end{aligned}\right\} \quad (2)$$

The stress continuity: The stress continuity expresses the fact that the stress above the layer must be equal to the stress under the layer, as well as for the diagonal component than for the non-diagonal component.

$$\left.\begin{aligned} (\bar{\sigma}_{zz})_i &= \lambda_i \nabla^2 \bar{\varphi}_i + 2\mu\left(\frac{\partial^2 \bar{\varphi}_i}{\partial z^2} + \frac{\partial^2 \bar{\psi}_i}{\partial r \partial z} + \frac{1}{r}\frac{\partial \bar{\psi}_i}{\partial z}\right) \\ &= (\bar{\sigma}_{zz})_{i+1} \\ (\bar{\sigma}_{zr})_i &= \mu_i\left(2\frac{\partial^2 \bar{\varphi}_i}{\partial r \partial z} - \frac{\partial^2 \bar{\psi}_i}{\partial z^2} + \frac{\partial^2 \bar{\psi}_i}{\partial r^2} + \frac{\partial \bar{\psi}_i}{r \partial r} - \frac{\bar{\psi}_i}{r^2}\right) \\ &= (\bar{\sigma}_{zr})_{i+1} \end{aligned}\right\} \quad (3)$$

THE REFLECTION-TRANSMISSION COEFFICIENTS

Reflection coefficients: The basic task of interpreting reflection sections is that of selecting those events on the record that represents primary reflections, translating the travel times of these reflections into depths and dips and mapping the reflecting horizons. In addition, the interpreter must be alert to other types of events that may yield valuable information, such as multiple reflections and diffractions. Reflections exhibit normal move outs that must fall within certain limits set by the velocity distribution.

The apparent velocity (distance between two geophones divided by the difference in travel times) is very large for reflections, usually greater than 50 km sec^{-1}. Reflection events rarely involve more than two or three cycles and are often rich in frequency components in the range 15 to 50 Hz. Deep reflections may have considerable energy below this range.

The basic problem in reflection seismic surveying is to determine the position of a bed that gives rise to a reflection on a seismic record.

Transmission coefficients: When a wave arrives at a surface separating two media having different elastic properties, it gives rise to reflected and refracted waves. At the boundary, the stresses and displacements must be continuous .The reflection coefficients given by the two Eq. 4 and the transmission coefficients given by the two Eq. 5 are not so accurate because we loose some significant digits. Furthermore some corrections are needed.

$$\left.\begin{aligned}\tilde{R}_{PP} &= \frac{(m_{12}m_{22} - m_{13}m_{22})}{\det M_{11}} \\ \tilde{R}_{PS} &= \frac{(m_{11}m_{23} - m_{13}m_{21})}{\det M_{11}}\end{aligned}\right\} \quad (4)$$

$$\left.\begin{aligned}\tilde{T}_{PP} &= m_{31}R_{PP} + m_{32}\tilde{R}_{PS} + m_{33} \\ \tilde{T}_{PS} &= m_{41}\tilde{R}_{PP} + m_{22}\tilde{R}_{PS} + m_{34}\end{aligned}\right\} \quad (5)$$

MODIFIED REFLECTION COEFFICIENTS

To overcome the loosing of some significant digits, in the case of reflection for example, we will use the matrix representation of the reflection coefficients. In formula (6) ij means the ith and the jth row of the matrix M and kl means the kth and the lth row of the matrix M. So we introduce the notation (7) and (8) than from formula using the matrix properties we can for example write the reflection coefficient of P-P wave given by formula (9) and $t_i(i=1,\ldots,5)$ are the coefficients deduced from the matrix multiplication. Finally the Eq. 9 allows us to obtain the reflection coefficient of a no converted P-wave without loosing some significant digits. In the same manner all kinds of P-waves and S-waves reflection components can be deduced[2].

$$m\Big|_{kl}^{ij} = m_{ik}m_{jl} - m_{il}m_{jk} \quad (6)$$

$$\left.\begin{aligned}\tilde{R}_{PP} &= \frac{m_{12}m_{23} - m_{13}m_{22}}{m_{11}m_{12} - m_{12}m_{21}} = \frac{m\Big|_{23}^{12}}{m\Big|_{12}^{12}} = \frac{\hat{m}_{14}}{\hat{m}_{11}} \\ \tilde{R}_{PS} &= \frac{m_{11}m_{23} - m_{13}m_{23}}{m_{11}m_{12} - m_{12}m_{21}} = -\frac{m\Big|_{13}^{12}}{m\Big|_{12}^{12}} = \frac{\hat{m}_{12}}{\hat{m}_{11}}\end{aligned}\right\} \quad (7)$$

$$\left.\begin{aligned}\hat{m}_{11} &= \hat{T}_{11}^{(1)} \cdot t_1 + \hat{T}_{21}^{(1)} \cdot t_2 + \hat{T}_{31}^{(1)} \cdot t_3 + \hat{T}_{51}^{(1)} \cdot t_4 + \hat{T}_{63}^{(1)} \cdot t_5 \\ \hat{m}_{14} &= -\hat{T}_{13}^{(1)} \cdot t_1 - \hat{T}_{21}^{(1)} \cdot t_2 - \hat{T}_{33}^{(1)} \cdot t_3 - \hat{T}_{51}^{(1)} \cdot t_4 - \hat{T}_{63}^{(1)} \cdot t_5\end{aligned}\right\} \quad (8)$$

$$\tilde{R}_{PP} = \frac{-\hat{T}_{13}^{(1)} \cdot t_1 - \hat{T}_{21}^{(1)} \cdot t_2 - \hat{T}_{33}^{(1)} \cdot t_3 - \hat{T}_{51}^{(1)} \cdot t_4 - \hat{T}_{63}^{(1)} \cdot t_5}{\hat{T}_{11}^{(1)} \cdot t_1 + \hat{T}_{21}^{(1)} \cdot t_2 + \hat{T}_{31}^{(1)} \cdot t_3 + \hat{T}_{51}^{(1)} \cdot t_4 + \hat{T}_{63}^{(1)} \cdot t_5} \quad (9)$$

THE INVESTIGATED MODEL

The basic technique of seismic exploration consists of generating seismic waves and measuring the time required for the waves to travel from the sources to a series of geophones, usually disposed along a straight line directed toward the source. From the above displayed data after plotting we can easily differentiate the low velocity layer and the high velocity layer for the P-waves and S-waves. In the same way the low and high density layer are displayed. The model used in this numerical investigation for a horizontally layered medium.We have 10 layers just below the sea floor; the P-wave velocity and S-wave velocity are respectively 2000 and 550 ms^{-1} (Table 1 and Fig. 1). The density of the sea floor is equal to 2.0 g cm^{-3}[3].

RECORDING PARAMETERS

Seismic source function: The source is located 8 m below the free surface and generates Ricker wavelets with a dominant frequency approximately equal to 30 Hz. The wavelet length is 60 m, the offset is 10 m, the time sampling is 2 s, the maximum record length is 4 s. The water depth is 300 m. The minimum offset and maximum offset are, respectively 0 and 350 m (Fig. 2). The hydrophones and multi-components geophones are located at the sea floor. As the medium has a symmetrical presentation, only the radial and vertical components of particle velocity are calculated.

Magnetic tape recording spread rapidly in the next few years, especially after digital recording and processing were introduced in 1960s. Magnetic tape recording made it possible to combine the data from several recordings made at different times and this made the use of weaker energy sources feasible. As the capability of recording the data from more geophones grew, recordings became spaced so closely those reflection reflections could be followed continuously along lines of profile and continuous coverage became the standard seismic reflection method.

The Ricker wavelet has been chosen as the seismic source function; mathematically it can be expressed, respectively in time domain and frequency domain as

$f(t) = (1 - 2\pi^2 v_M t^2)\ \exp(-\pi^2 v_M t^2)$, the time domain function and

$$F(v) = \left(\frac{2}{\sqrt{\pi}}\right)\left(\frac{v^2}{v^2 M}\right)\exp\left(-\frac{v^2}{v^2 M}\right),$$

the frequency domain function, With two particular values given by

$$T_D = \frac{\sqrt{6}}{\pi} v_M;\quad T_R = \frac{T_D}{\sqrt{3}}\ \text{ and }\ F_{max} = F(v_M).$$

The velocity analysis: Figure 3 clearly shows the local high velocity layer HVL (1200-1400 m) and the local low velocity layer LVL (950-1200 m) of P-wave and we can also differentiate the high velocity layer of S-waves located in

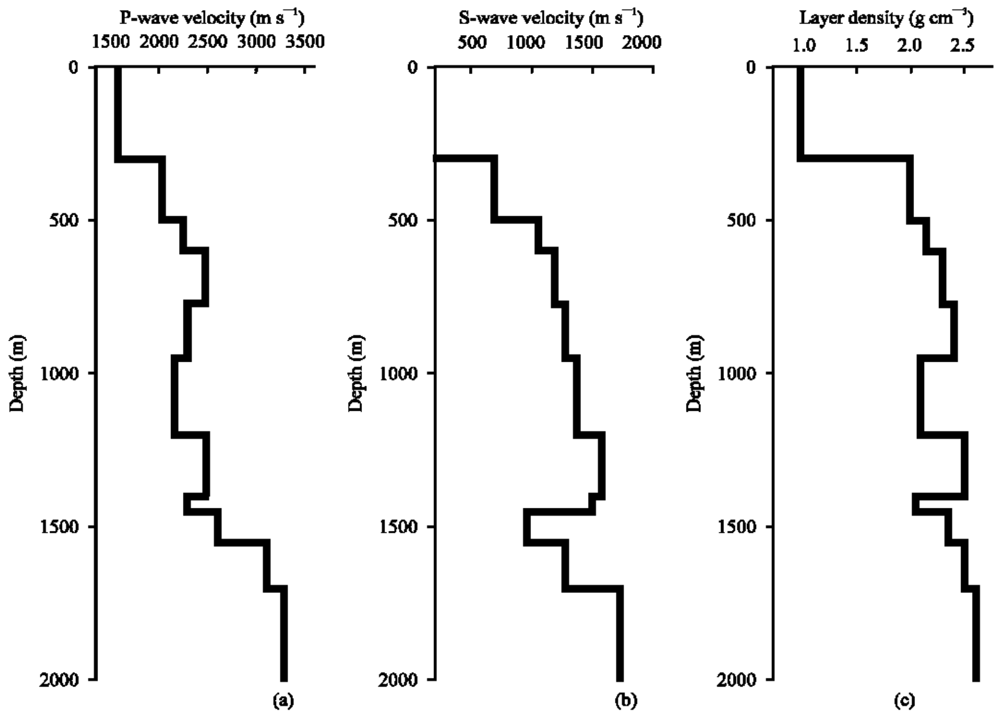

Fig. 1: The model used to generate data. (a) P-wave velocity versus depth. (b) S-wave velocity versus depth. (c) Layer density versus depth

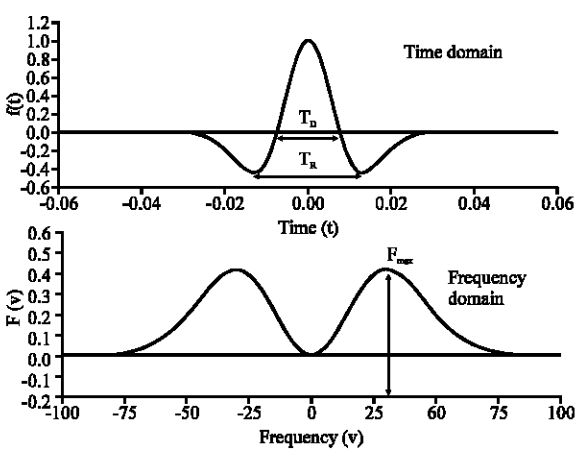

Fig. 2: Ricker Wavelet in time and frequency domain

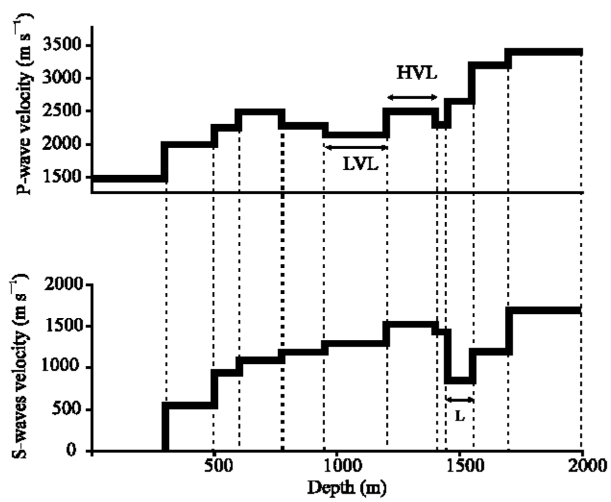

Fig. 3: Velocity analysis of P and S-waves

Table 1: The investigated model parameters

Layer No.	Depth (M)	P-wave velocity (m sec^{-1})	S-wave velocity (m sec^{-1})	Layer density (msec^{-1})
1	300	1500	0	1.00
2	500	2000	550	2.00
3	600	2250	950	2.15
4	775	2500	1100	2.30
5	950	2300	1200	2.40
6	1200	2150	1300	2.10
7	1400	2500	1530	2.50
8	1450	2300	1440	2.05
9	1550	2650	850	2.35
10	1700	3200	1200	2.50
11	Half space	3400	1700	2.60

the same depth interval with the S-wave (1200-1400 m) and the local low velocity layer of S-wave is located in the depth interval (1450-1550 m). The high velocity layer of P and S wave corresponds with the local high density layer and the two low velocity layers each one correspond to a local low density layer. About the factor affecting velocity we know that the velocity of P-waves in homogeneous solid medium is a function only of the elastic constants and the density.

One might expect that the elastic constants, which are properties of the intermolecular forces, would be relatively intensive to pressure, whereas the density should increase with pressure because rocks are moderately compressible. In actual rocks, the pore spaces are filled with a fluid whose elastic constants and density also affect the seismic velocity. Oil is slightly more compressible than water, so oil-filled pores result in slightly lower velocity than water-filled pores. Gas is considerably more compressible than water and so gas-filled pores often result in much lower velocity. Even a small amount of gas may lower velocity appreciably. These effects are used as hydrocarbon indicators.

Subjecting a porous rock to high pressure results in both reversible and irreversible changes in porosity, that is, when the pressure is removed, a small part of the original porosity is regained whereas most is permanently lost, perhaps because of crushing of the grains, alteration of the packing, or other permanent structural changes. Empirical data suggest that the maximum depth to which a rock has been buried is a measure of the irreversible effect on porosity. In summary, porosity appears to be the dominant variable in determining the velocity in sedimentary rocks and porosity in turn is determined principally by the existing differential pressure and the maximum depth of burial[5].

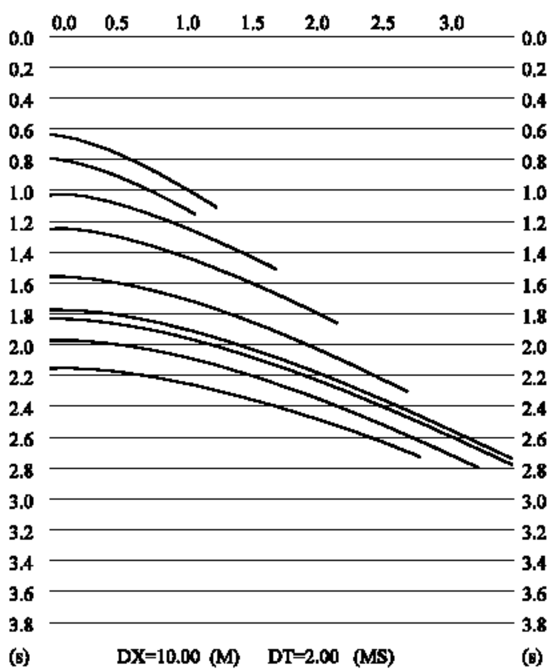

Fig. 4: Radial component of P-S waves

RESULTS OF RECORDS MODELED

The pressure field above the sea floor is displayed in (Fig. 6) and the radial component of up-going P-SV waves in (Fig. 7). Below the acoustic layer also called elastic

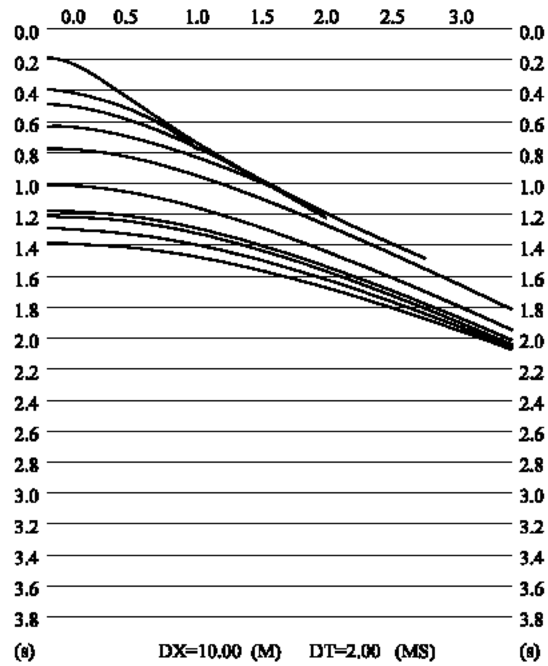

Fig. 5: Vertical component of P-P waves

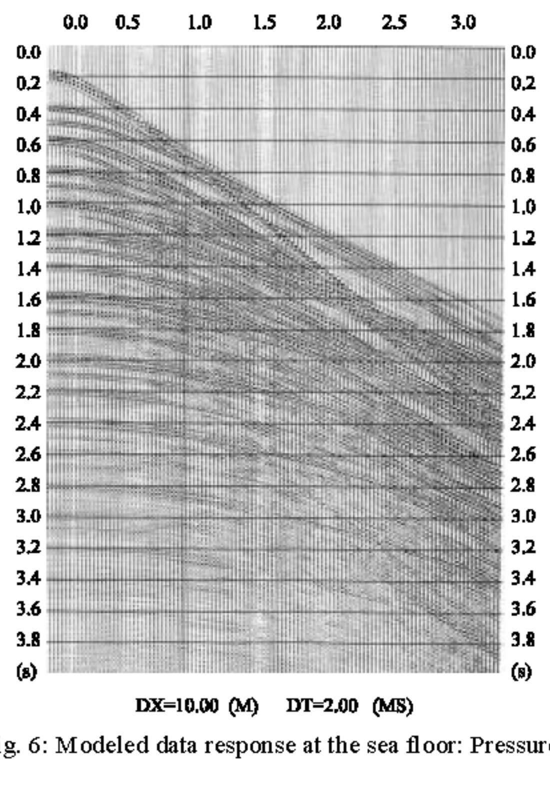

Fig. 6: Modeled data response at the sea floor: Pressure

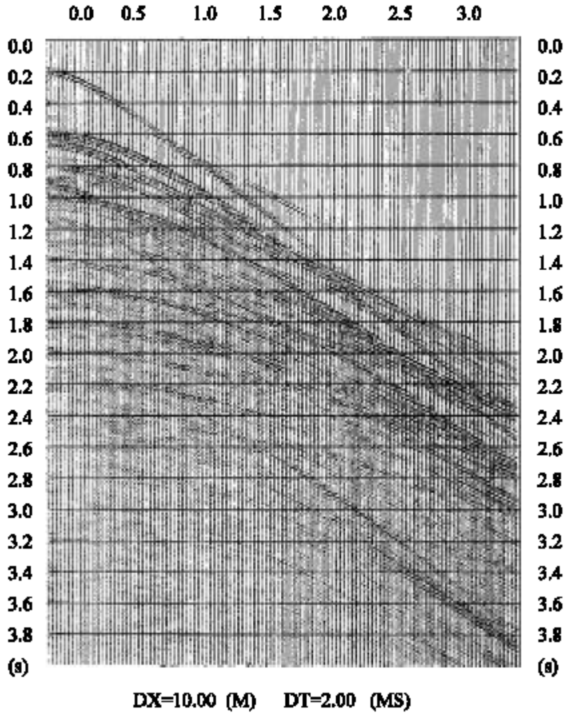

Fig. 7: Radial component of up-going P-SV waves

interface, two particle velocity components are displayed in (Fig. 4) and (Fig. 5). At the sea floor, the down-going waves and multiple waves are refracted in P-waves and S-waves. These waves reverberate in the elastic layers embedded between the acoustic top layer and the elastic half space. For each reflection the energy conversion takes place and the recorded wave's pattern at the sea floor becomes quite complicated.

Just below the sea-floor, up-going P-waves are reflected into both: down-going P-waves and down-going S-waves, with the relative amplitudes determined by reflection coefficients for an up-going P-wave.

In the same way, up-going S-waves reflect into down-going P-waves and S-waves. Hence, it is important to enable the separation of the down-going events from the up-going events when we are interpreting the multiple-component seismograms.

CONCLUSIONS

A short overview of the reflectivity method has been demonstrated, the algorithm for the program is omitted. The pressures, the radial and vertical components of the particle velocities recorded by the geophones planted on the sea floor have been modeled. The energy conversion takes place and the recorded wave patterns at the sea floor as seen become complicated. The modeled data shows that the reflectivity method is quiet correct.

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