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SEISMIC SOURCE STUDY BY ANALYSIS OF BODY WAVE SPECTRA

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A new technique for seismic source study by analysis of teleseismic body wave spectra is presented. The reconstruction of seismic rupture is usually carried out by fitting a parametric source model to the observations. The method we are going to discuss is characterized by rejecting, even though in part, the parametric approach. Most techniques rely on minimization of the differences between synthetic and observed seismograms. The technique we are discussing fits the model to the observations in the frequency domain. We use for inversion the initial part of every record, formed by direct P, pP and sP body wave phases. We consider the approximation of time dependent point source. In that case the body waves of this same type recorded at different stations ought to have identical (apart from the sign and the time delays) shapes coinciding with the source time function. It is this model property of the observations, "coherence of wave phases at different stations", which underlies the method to be discussed.
Results of application of this technique for Koryakiya, 1991 and Kirgizia, 1992 earthquakes study are presented.

We will start from the illustration of main ideas of the method considering an unreal but simplest case of isolated *direct P-wave inversion,* recorded at different stations (see Fig.1).

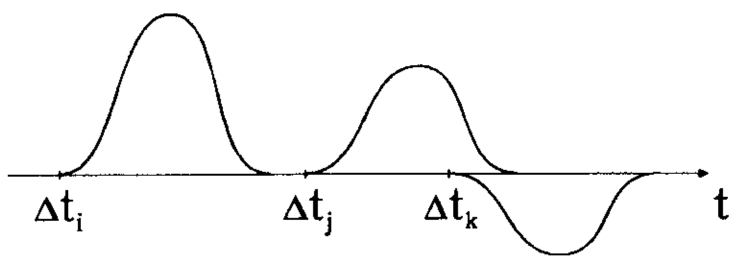

Fig.1 Direct P-wave recorded at different stations.
Δt_i, Δt_j, and Δt_k, - arrival times at i-th, j-th and k-th stations.

Let $K_i(\omega)$ - complex spectrum of z-component of direct P-wave recorded at i-th station, corrected for attenuation and divided by value $\dfrac{\sin e^{(i)} \chi^{(i)} J^{(i)}}{4\pi \sqrt{\rho_0 \alpha_0^5 \rho\alpha}}$.

Here ρ_0, α_0 - density and P-wave velocity near the source; ρ, α - values of this same parameters near the point of registration; $J^{(i)}$ - geometrical spreading; $e^{(i)}$ - the angle between the direction of P-wave propagation and free surface; $\chi^{(i)}$ - the ratio of the amplitude of free surface oscillations to the amplitude of the

Constant A will be defined later.
Let us consider the way of estimation of "polarities" $\Delta\psi_i$. We will use equation (4). Averaging it with respect to stations we have:

$$\langle \psi(\omega) \rangle = \psi_0(\omega) - \omega \langle \Delta t \rangle + \langle \Delta \psi \rangle + 2\pi \langle k \rangle, \quad (7)$$

where $\langle \psi(\omega) \rangle = \left(\frac{1}{n}\right)\left[\psi_1(\omega) + \cdots + \psi_n(\omega)\right]$.

Subtracting (7) from (4) we obtain

$$\psi_i(\omega) - \langle \psi(\omega) \rangle = -\omega \left(\Delta t_i - \langle \Delta t \rangle\right) + \qquad (8)$$
$$+ \Delta \psi_i + 2\pi k - \langle \Delta \psi \rangle - 2\pi \langle k \rangle$$

Curves (8) are presented on the Fig. 2a. We approximate these curves by straight lines in the spectral band where the signal to noise ratio is high. After it we shift every line by $l_i\pi$ (l_i - integer) in vertical direction in such a way that they are intersecting in the same point at the axes $\omega=0$ in the interval $(0,\pi)$ as shown at the Fig. 2b.

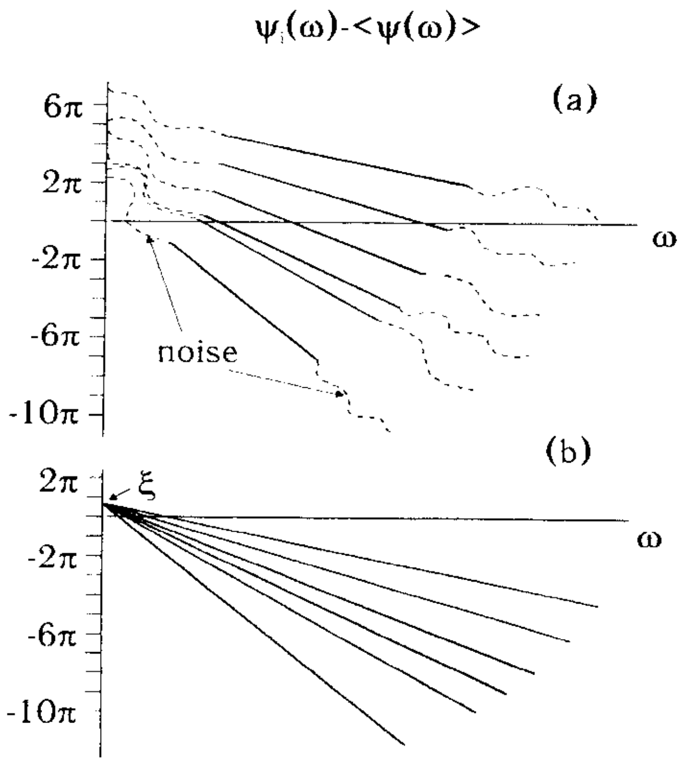

Figure 2. Determination of relative polarity.

The existence of such a point ξ follows from (8): the difference of values $\psi_i(0) - \langle \psi(0) \rangle$ and $-\langle \Delta\psi \rangle - 2\pi \langle k \rangle$ for every station is fold to π.
So, for every i following equation is valid:

incident P-wave. In applications we used IASPEI-91 model of the Earth to calculate all these parameters.

If $K_0(\omega)$ - complex spectrum of the source time function and Δt_i - arrival time, then

$$K_i(\omega) = C_i K_0(\omega) e^{-i\omega\Delta t_i} e^{i\Delta\psi_i}. \qquad (1)$$

C_i , $\Delta\psi_i$ are defined by the source radiation:

$$C_i e^{i\Delta\psi_i} = \mathbf{r}_i^T \mathbf{m}\, \mathbf{r}_i , \qquad (2)$$

$C_i \geq 0$, $\Delta\psi_i = 0, \pi$, $\mathbf{m}$ - normalized moment tensor, $\mathbf{r}_i$ - radiation direction, $\mathbf{r}_i^T$ - transposed vector.

As one can see from equations (1)-(2) the amplitude spectra in direct P wave recorded by different stations differ from each other by scalar factor, and phase spectra - by linear function of ω. This property of signals is called "coherence". If C_i and $\Delta\psi_i$ are estimated from observations then equation (2) gives a linear system for moment tensor elements calculation.

Let $F_i(\omega)$ and $\psi_i(\omega)$ - amplitude and phase spectra of observed signal and $F_0(\omega)$, $\psi_0(\omega)$ - amplitude and phase spectra of the source time function, *i.e.*

$$K_i(\omega) = F_i(\omega) e^{i\psi_i(\omega)},$$

$$K_0(\omega) = F_0(\omega) e^{i\psi_0(\omega)}.$$

We assume that source time function is positive, then $\psi_0(0)=0$.
Applying Ln to equation (1) we have:

$$\mathrm{Ln}\,F_i(\omega) = \mathrm{Ln}\,F_0(\omega) + \mathrm{Ln}\,C_i , \qquad (3)$$

$$\psi_i(\omega) = \psi_0(\omega) - \omega\Delta t_i + \Delta\psi_i + 2\pi k_i , \qquad (4)$$

where $i = 1, \ldots , n$.
System (3) gives us following estimates for C_i and $F_0(\omega)$:

$$C_i = A \frac{\{F_i\}_\omega}{\{F\}_{i\omega}} \qquad (5)$$

$$F_0(\omega) = \frac{\{F(\omega)\}_i}{A} , \qquad (6)$$

where notations are following. Let $\omega = \omega_1, \ldots ,\omega_N$ and $i = 1, \ldots , n$, then

$$\{F_i\}_\omega = \left(F_i(\omega_1)\cdot F_i(\omega_2)\cdots F_i(\omega_N)\right)^{1/N};$$

$$\{F(\omega)\}_i = \left(F_1(\omega)\cdot F_2(\omega)\cdots F_n(\omega)\right)^{1/n};$$

$$\{F\}_{i\omega} = \left(\{F(\omega_1)\}_i \cdot \{F(\omega_2)\}_i \cdots \{F(\omega_N)\}_i\right)^{1/N}$$

$$K_i(\omega) = K_0(\omega)\,Q_i(\omega)\exp(-i\omega\, t_P^{(i)}), \qquad (12)$$

where $Q_i(\omega)$ is defined by the source radiation pattern of phases P, pP, sP and by time delays $\tau_1^{(i)} = t_{pP}^{(i)} - t_P^{(i)}$ and $\tau_2^{(i)} = t_{sP}^{(i)} - t_P^{(i)}$ of phases pP and sP with respect to the direct P wave; $t_P^{(i)}, t_{pP}^{(i)}$ and $t_{sP}^{(i)}$ - arrival times of P, pP and sP waves. $Q_i(\omega)$ can be described by formulae

$$Q_i(\omega) = a_i + b_i \exp(-i\omega\tau_1^{(i)}) + c_i \mathrm{xp}(-i\omega\tau_2^{(i)}), \qquad (13)$$

$$a_i = \mathbf{r}_P^{(i)^T}\,\mathbf{m}\,\mathbf{r}_P^{(i)} \;,$$

$$b_i = k_{pP}^{(i)}\,\mathbf{r}_{pP}^{(i)^T}\,\mathbf{m}\,\mathbf{r}_{pP}^{(i)} \;,$$

$$c_i = k_{sP}^{(i)}\sqrt{\frac{\alpha_0^5}{\beta_0^5}}\,\mathbf{r}_{VsP}^{(i)^T}\,\mathbf{m}\,\mathbf{r}_{sP}^{(i)} \,.$$

Here $\mathbf{r}_P^{(i)}, \mathbf{r}_{pP}^{(i)}, \mathbf{r}_{sP}^{(i)}$ - direction of radiation of P, pP and sP waves; $\mathbf{r}_{VsP}^{(i)}$ - direction of displacement in sP phase in the initial moment; α_0 , β_0 - P-wave and S-wave velocities near the source; $k_{pP}^{(i)}$, $k_{sP}^{(i)}$ - reflection coefficients of pP and sP waves from the free surface above the source.

Long period inversion

Comparing (12)-(13) with (1)-(2) one can see that the sums of three phases recorded by different stations are coherent in low frequency domain of the spectra where following inequalities are valid

$$\omega\tau_1^{(i)} << 1 \quad \text{and} \quad \omega\tau_2^{(i)} << 1$$

For these frequencies function $Q_i(\omega)$ can be substituted by the sum $a_i+b_i+c_i$ and spectrum $K_i(\omega)$ can be expressed in the form of equation (1):

$$K_i(\omega) = g_i K_0(\omega)\, e^{-i\omega t_P^{(i)}} e^{i\Delta\psi_i} , \qquad (14)$$

where g_i and $\Delta\psi_i$ are the amplitude and phase of the sum $a_i+b_i+c_i$, and can be estimated from observed spectra in the same way as C_i and $\Delta\psi_i$ in the case of direct P-wave inversion. After g_i and $\Delta\psi_i$ are determined we have from formulae (13) following linear system for moment tensor elements calculation:

$$\mathbf{r}_P^{(i)^T}\mathbf{m}\,\mathbf{r}_P^{(i)} + k_{pP}^{(i)}\mathbf{r}_{pP}^{(i)^T}\mathbf{m}\,\mathbf{r}_{pP}^{(i)} + k_{sP}^{(i)}\sqrt{\frac{\alpha_0^5}{\beta_0^5}}\,\mathbf{r}_{VsP}^{(i)^T}\mathbf{m}\,\mathbf{r}_{sP}^{(i)} = g_i e^{i\Delta\psi_i} . \qquad (15)$$

$$\psi_i(0) - \langle \psi(0) \rangle - l_i \pi = \xi \ ,$$

and averaging it by "i" we have

$$\xi = -\langle l \rangle \pi \ . \quad (9)$$

In fact we are searching for such a point where the intersections of all these lines with axes ω=0 are mostly concentrated. The values of l_i give us relative polarities of P-waves: all signals with odd l_i have this same polarity as well as all signals with even l_i . Polarities we found are not absolute but relative because we could search for point ξ in any interval $(k\pi, (k+1)\pi)$ and we used $k = 0$. To obtain the absolute polarities (see formula (8)) we estimate the integer number m, where m is such that

$$\xi + \langle \Delta \psi \rangle + 2\pi \langle k \rangle = m\pi \ ,$$

or using (9)

$$\langle \Delta \psi \rangle + 2\pi \langle k \rangle - \langle l \rangle \pi = m\pi \ . \quad (10)$$

If $m + l_i$ is even number then $\Delta\psi_i$=0, if $m + l_i$ is odd number then $\Delta\psi_i=\pi$. To estimate m we use function $\varphi_i(\omega)$, where $\varphi_i(\omega) = \psi_i(\omega) - l_i\pi$, $\psi_i(\omega)$ are given by equation (4), and l_i - numbers obtained before. Averaging with respect to station number i we have

$$\langle \varphi(\omega) \rangle = \psi_0(\omega) - \omega \langle \Delta t \rangle + \langle \Delta \psi \rangle + 2\pi \langle k \rangle - \langle l \rangle \pi \ ,$$

or using (10):

$$\langle \varphi(\omega) \rangle = \psi_0(\omega) - \omega \langle \Delta t \rangle + m\pi \ . \quad (11)$$

Taking into account the equality $\psi_0(0)=0$ we approximate the curve (11) by a straight line. The inclination of this line defines the value of $\langle \Delta t \rangle$ and the intersection with the axes ω=0 (the closest point $m\pi$) defines the number m.

After C_i and $\Delta\psi_i$ are estimated from observations equation (2) gives a linear system for moment tensor elements calculation. Constant A in (5) and (6) is defining by the condition of tensor **m** normalisation $\mathrm{Tr}\ \mathbf{m}^T\mathbf{m} = 2$. $F_0(\omega)$ and $\psi_0(\omega)$ define the source time function spectrum, $M_0=F_0(0)$ - seismic moment.

Superposition of P, pP and sP phases

Let us consider the typical case when three body wave phases P, pP and sP forming the initial part of every teleseismic record can not be separated. In this case $K_i(\omega)$ - the complex spectrum of z-component of the sum of these three phases can be expressed in the form similar to equation (1):

$$\Phi(h) = \left\{ \frac{1}{2k} \sum_{i \neq j} \sum_{m=1}^{M} \left[v_m^{(i,j)}(h) - v_m^{(j,i)}(h) \right]^2 \right\}^{\frac{1}{2}}, \quad (18)$$

where k is the number of pares $i \neq j$.
We take as the estimate of the source depth the value of h, correspondent to the minimum of the functional $\Phi(h)$.

Determination of the source time function.

We determined the source time function spectrum in low frequency domain. But it is important to know this spectrum for higher frequencies as well. Using estimated moment tensor, source depth and following formulae (13) we calculate function

$$\tilde{Q}_i(\omega) = Q_i(\omega)\left[a_i + b_i + c_i\right]^{-1}.$$

Let us consider function

$$\tilde{K}_i(\omega) = K_i(\omega)\left[\tilde{Q}_i(\omega)\right]^{-1}, \quad (19)$$

which can be presented by formula similar to the formula (14):

$$\tilde{K}_i(\omega) = g_i K_0(\omega) e^{-i\omega t_p^{(i)}} e^{i\Delta\psi_i}.$$

As one can see from formula (19) while $\omega \to 0$ function $\tilde{K}_i(\omega)$ converge to the function $K_i(\omega)$, so the estimates of parameters characterising observed spectra $K_i(\omega)$ (obtained from long period spectra analysis) can be related to $\tilde{K}_i(\omega)$.
Substituting observed phase spectra in equation (7) by the phase spectra of "corrected observations" $\tilde{K}_i(\omega)$ we estimate the phase spectrum of the source time function $\psi_0(\omega)$.
To estimate the amplitude spectrum of source time function we use formula

$$F_0(\omega) = \frac{\sum_{i=1}^{N} q_i(\omega) F_i(\omega)}{\sum_{i=1}^{N} \left[q_i(\omega)\right]^2}. \quad (20)$$

After the spectrum of source time function is calculated, we obtain this function applying Fourier transformer.

Determination of the source depth

Let us consider the spectrum at periods comparable with delays of pP and sP phases with respect to direct P-wave. At these periods the dependence of function $Q_i(\omega)$ on the frequency ω is significant and defined by moment tensor and by the source depth. As follows from formula (12) the amplitude spectrum $F_i(\omega)$, observed at the station i, can be represented in following form

$$F_i(\omega) = q_i(\omega) F_0(\omega) , \qquad (16)$$

where $q_i(\omega) = |Q_i(\omega)|$, and $F_0(\omega)$ - amplitude spectrum of the source time function.

Using moment tensor obtained from long period spectra we can calculate $q_i(\omega)$ for any value of the source depth and find which of them will satisfy equation (16) in the best way. We could compare the ratios $F_i(\omega)/q_i(\omega)$ which have to be equal to each other and to the function $F_0(\omega)$, and use the average (with respect to the subscript i) difference of them as the minimising functional. But functions $q_i(\omega)$ can become too small at any frequencies. In such a case even small noise will cause a big error. To avoid it we compare products $F_i(\omega) q_j(\omega)$ for different pares i and j. As one can see from formula (16) when the depth takes its true value following equality should be valid for every pare i and j:

$$F_i(\omega) q_j(\omega) = F_j(\omega) q_i(\omega) . \qquad (17)$$

To make the technique more stable with respect to the noise we compare normalized left-hand and right-hand parts of the equality (17). Let us consider functions $V^{(i,j)}(\omega_m) = F_i(\omega_m) q_j(\omega_m)$ given at a discrete set of ω_m (m=1,...,M) as M-dimensional vectors $V^{(i,j)}$ with components

$$V_m^{(i,j)} = V^{(i,j)}(\omega_m) \cdot$$

Let us define the unit vectors $\mathbf{v}^{(i,j)}$ by formula

$$v_m^{(i,j)} = \frac{V^{(i,j)}(\omega_m)}{\left\{ \sum_{m=1}^{M} \left[V^{(i,j)}(\omega_m) \right]^2 \right\}^{\frac{1}{2}}} .$$

Unit vectors $\mathbf{v}^{(i,j)}$ depend on the source depth h , that is $\mathbf{v}^{(i,j)}=\mathbf{v}^{(i,j)}(h)$. When the source depth takes its true value vectors $\mathbf{v}^{(i,j)}(h)$ and $\mathbf{v}^{(j,i)}(h)$ should coincide for any (i,j). We chose the mean-root-square (among the pares (i,j)) value of the sine of the half of the angle between the vectors $\mathbf{v}^{(i,j)}(h)$ and $\mathbf{v}^{(j,i)}(h)$ as a minimising functional $\Phi(h)$. This sine is equal to the half of the length of vector $\mathbf{v}^{(i,j)}(h)$-$\mathbf{v}^{(j,i)}(h)$:

3 km/s. Rupture is a segment parallel to the bigger side of rectangle. Its length is 40 km. The arising time at every point is equal to 1 s. The moment tensor is uniform in all the source: seismic moment $1.3 \cdot 10^{19}$ n·m, strike angle 225°, deep angle 60°, and slip angle 90° - all these values were obtained from the surface wave inversion. Synthetic seismograms were calculated for all the stations presented in Fig. 3, and they were processed in the same way as real records accordingly to described technique.
Analysing the synthetic spectra in the low frequency domain from 0.007 to 0.02 Hz we obtained following results: seismic moment $1.4 \cdot 10^{19}$ n·m, strike angle 225°, deep angle 63°, and slip angle 87°. We consider the accuracy of these results of inversion as high enough to make a positive conclusion about the applicability of this point source approximation technique for estimation of moment tensor of real events.

From the analysis of real P-wave records spectra in the same spectral domain we obtained following focal mechanism: strike angle 219°, deep angle 64°, and slip angle 97°. The seismic moment was estimated as $1.9 \cdot 10^{19}$ n·m. As one can see the difference between this focal mechanism and synthetic one is small enough. But the P-wave estimate of seismic moment is about 1.5 times higher then the surface wave estimate. It may be explained by the fact that the instrument amplitude responses as well as the sum $\sum_{i=1}^{N}\left[q^{(i)}(\omega)\right]^2$ in formula (20) become too small on low frequencies and because of existence of low frequency noise the amplitude spectra are overestimated. The stereographic projections of nodal planes on the lower hemisphere, correspondent to focal mechanisms obtained from different data, are shown in the Fig. 5. (a) corresponds to the solution obtained by surface wave inversion and used for synthetic source model; (b) - CMT solution; (c) - solution obtained by describing technique from synthetic seismograms; and (d) - solution obtained by this same technique from real P-wave records.

To determine the source depth we analysed P-wave spectra in frequency domain from 0.01 to 0.12 Hz. The function $\Phi(h)$ (see formula (18)) obtained from synthetic spectra is shown in the Fig. 6a. It attains minimal value at the depth correspondent to the depth of spatial centroid of the source model used for calculation of synthetic seismograms. The function $\Phi(h)$ obtained from real records spectra is shown in the Fig. 6b. The depth 18 km, where function attains minimum we consider as estimate of equivalent point source depth. The CMT solution depth is equal to 15 km. The depth 21 km, where function $\Phi(h)$ is broken, corresponds to the boundary between granite and basalt in IASPEI-91 model.

The spectrum of source time function was calculated in this same frequency domain, which was used for source depth estimate. Correspondent source time function calculated from synthetic records is given in Fig. 7a, and that one calculated from real P-wave records is given in Fig. 7b. The oscillations around the main peaks on both curves may by explained by the approximate character of formulae used for calculations - they were obtained under assumption that the source is a point. Nevertheless even the function obtained from the real records has a strongly pronounced main peak which can be considered as approximation of source time function f(t). As one can see from

Misfit between observed and synthetic spectra as function of the source depth

Now we will describe another then minimisation of the functional (18) algorithm for source depth determination. For a set of values of the source depth h we calculate by formula (13) functions $q_i^{(h)}(\omega) = |Q_i(\omega)|$, using moment tensor obtained from long period spectra. Estimating amplitude spectra of source time function by formula (20), we obtain from (16) an equation for estimation of amplitude spectra for every value of h, which we will call synthetic amplitude spectra $F_i^s(\omega,h)$:

$$F_i^s(\omega, h) = q_i^{(h)}(\omega) \frac{\sum_{i=1}^{N} q_i^{(h)}(\omega) F_i(\omega)}{\sum_{i=1}^{N} \left[q_i^{(h)}(\omega)\right]^2} . \qquad (21)$$

Considering functions $F_i^s(\omega_m,h)$ and $F_i(\omega_m)$ given at a discrete set of ω_m (m=1,...,M) we define the normalised residual between observed and synthetic amplitude spectra by formula

$$\varepsilon(h) = \frac{\sqrt{\sum_{i=1}^{N}\sum_{m=1}^{M}\left(F_i^s(\omega_m,h) - F_i(\omega_m)\right)^2}}{\sqrt{\sum_{i=1}^{N}\sum_{m=1}^{M} F_i^2(\omega_m)}} \qquad (22)$$

We take as the estimate of the source depth the value of h, correspondent to the minimum of the function $\varepsilon(h)$.

Results of application

1. This technique was used for analysis of the records of Hailino earthquake, 1991 in Koryakiya, which we studied before using surface waves records. On the base of results of this previous study we developed a model of the source of this event. We calculated synthetic seismograms for the sum of P, pP and sP phases correspondent to this model and applied the technique under discussion for analysis of these synthetic seismograms and records of real earthquake.

As it was mentioned we solve the problem in approximation of time dependent *point* source. Applying the technique for analysis of synthetic seismograms generated by a source of a finite size, we tested its applicability for the study of real events.

The distribution of stations which records were used in analysis is given in the Fig. 3. VBB channels are marked by squares, and LP channels - by triangles. The source model developed on the base of surface waves study is presented in the Fig. 4. Arrows with letters N and E are indicating the directions to the North and East. Vector ΔU shows the direction of slip. Its value is equal to 0.4 m, and slip angle is 90°. The source shape is a rectangle 40x20 km, the depth of the spatial centroid is equal to 20 km. Vector **V** indicate the direction of rupture propagation. The rupture velocity is equal to

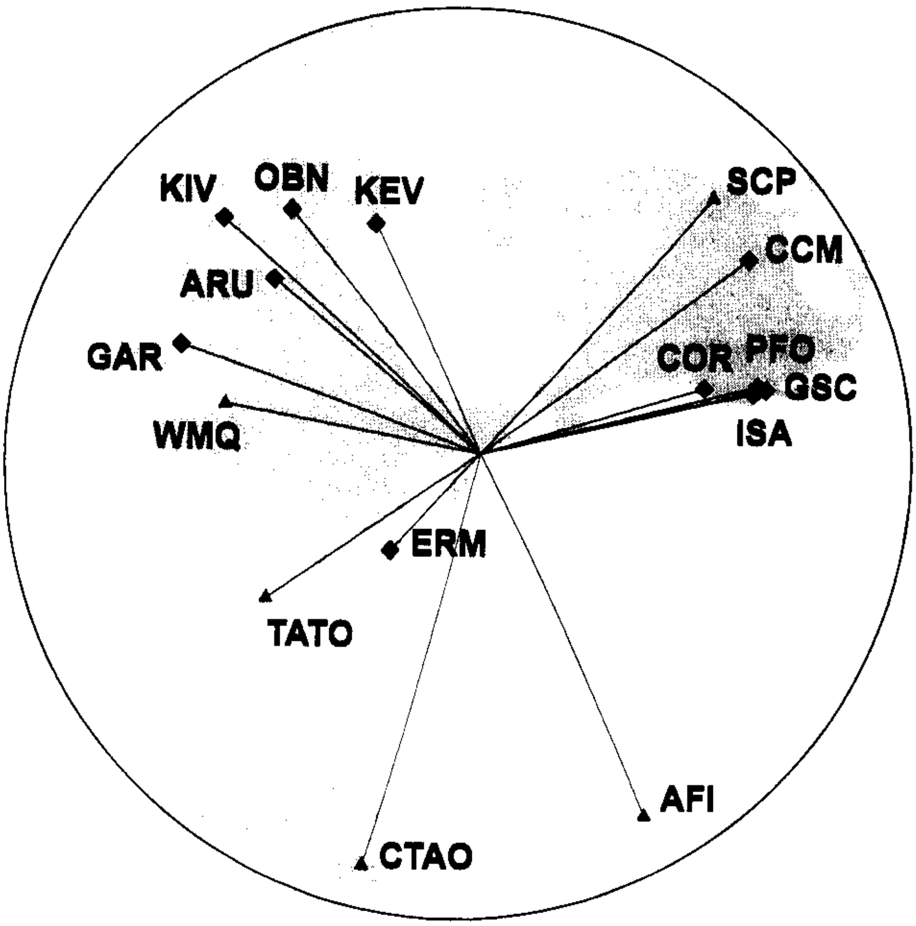

Fig.3 Distribution of stations used for Hailino earthquake study

Fig. 7b, the equivalent point source duration is about 10 s. The CMT duration for this event was reported equal to 10.2 s.

1. As another example of application of this technique we present here the results of study of Susamyr earthquake of 19 August, 1992 in Kirghizstan (M_s=7.4, M_b=6.8). The distribution of stations which VBB records were used in analysis is given in the Fig. 8. Analysing the synthetic spectra in the low frequency domain from 0.008 to 0.02 Hz we obtained moment tensor solution. The stereographic projections of nodal planes on the lower hemisphere are shown in the Fig. 9a. For comparison we present in Fig. 9b CMT solution. This solutions are in a good agreement. Our estimate of seismic moment is $8.9 \cdot 10^{19}$ n·m. CMT seismic moment is equal to $7.1 \cdot 10^{19}$ n·m. To determine the source depth we analysed P-wave spectra in frequency domain from 0.008 to 0.05 Hz. The residual function $\varepsilon(h)$, calculated by formula (22) is shown in the Fig. 10. It attains minimal value at the depth 19 km. The CMT solution depth is equal to 17 km. The spectrum of source time function was calculated from 0.001 to 0.1 Hz. Correspondent source time function is given in Fig. 11. The source duration can be estimated by 17 - 18 s. This value is in a good agreement with results of surface waves analysis and body waveform inversion, which will be presented in next lecture.

The comparison of results of application of technique under discussion with results of application of other techniques, and testing of this technique on synthetic data show its efficiency.

Acknowledgements

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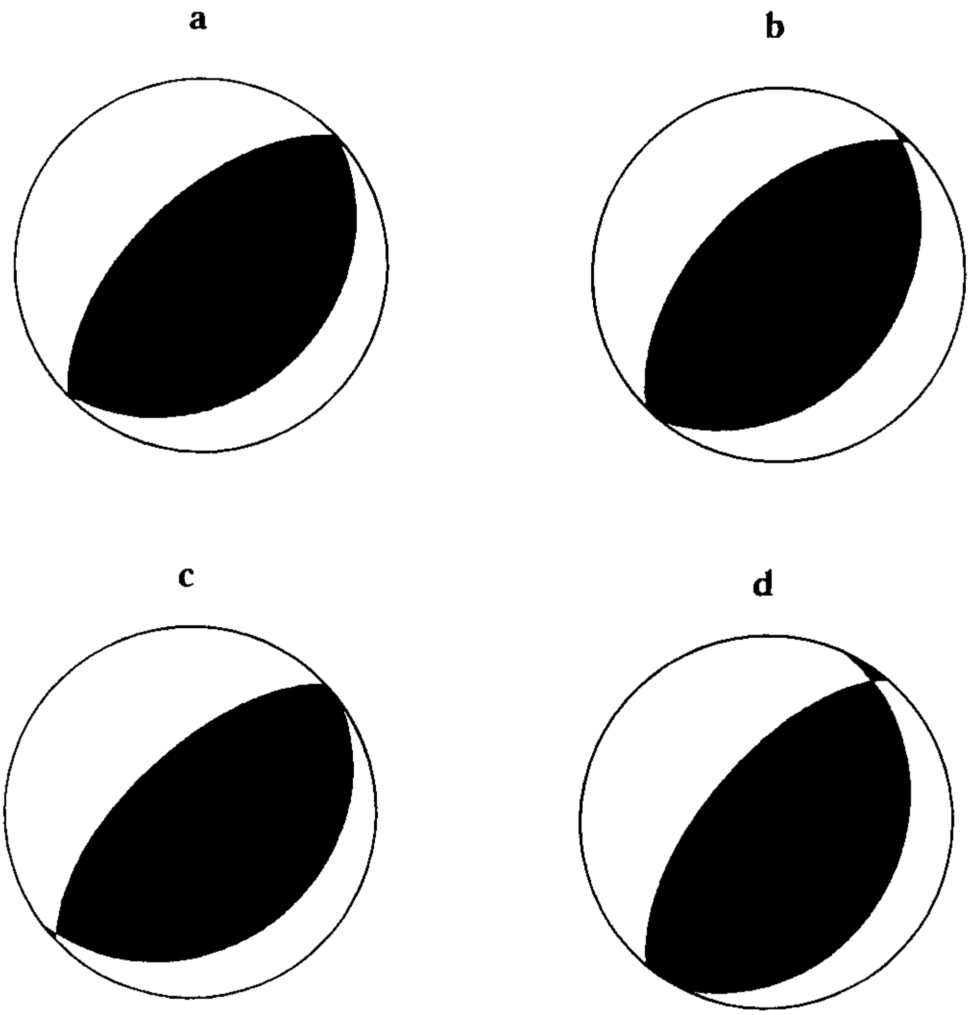


Fig.5 Focal mechanisms from analysis of different data.

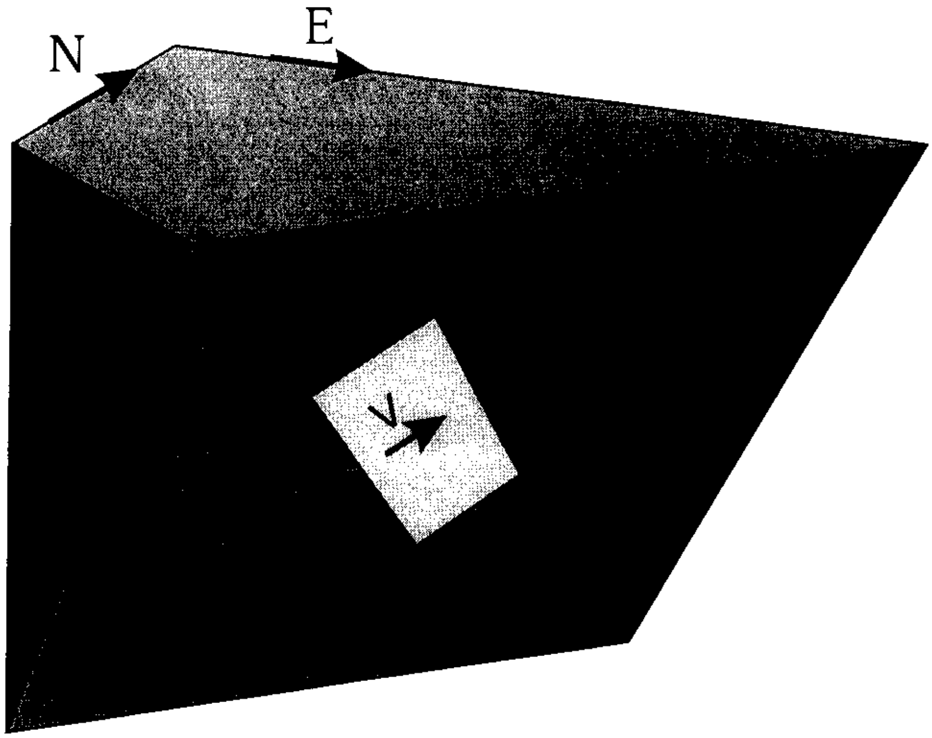


Fig.4 Synthetic source model.

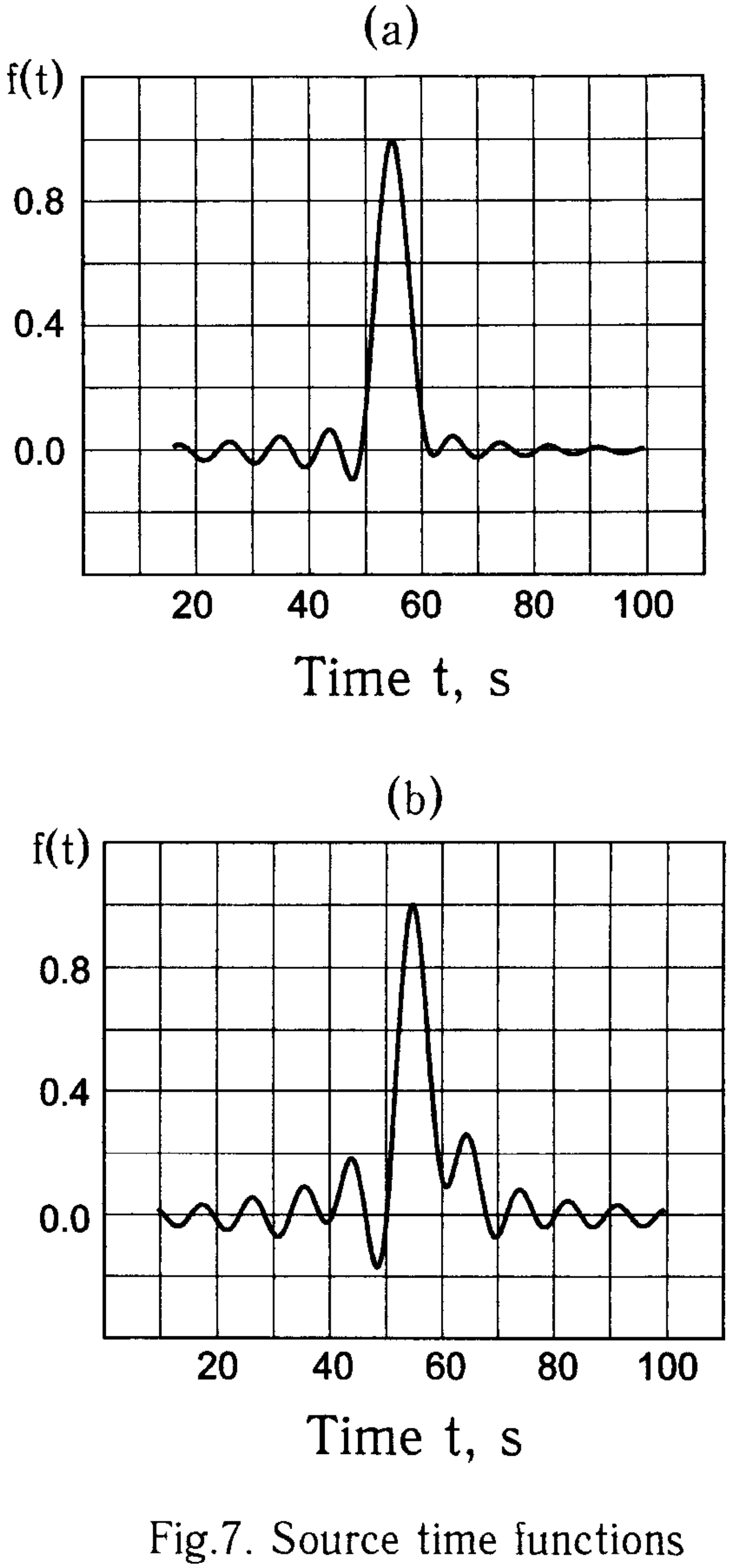


Fig.7. Source time functions
obtained from synthetic and observed data.
(a) - result of analysis of synthetic spectra;
(b) - result of analysis of real Pwave spectra.

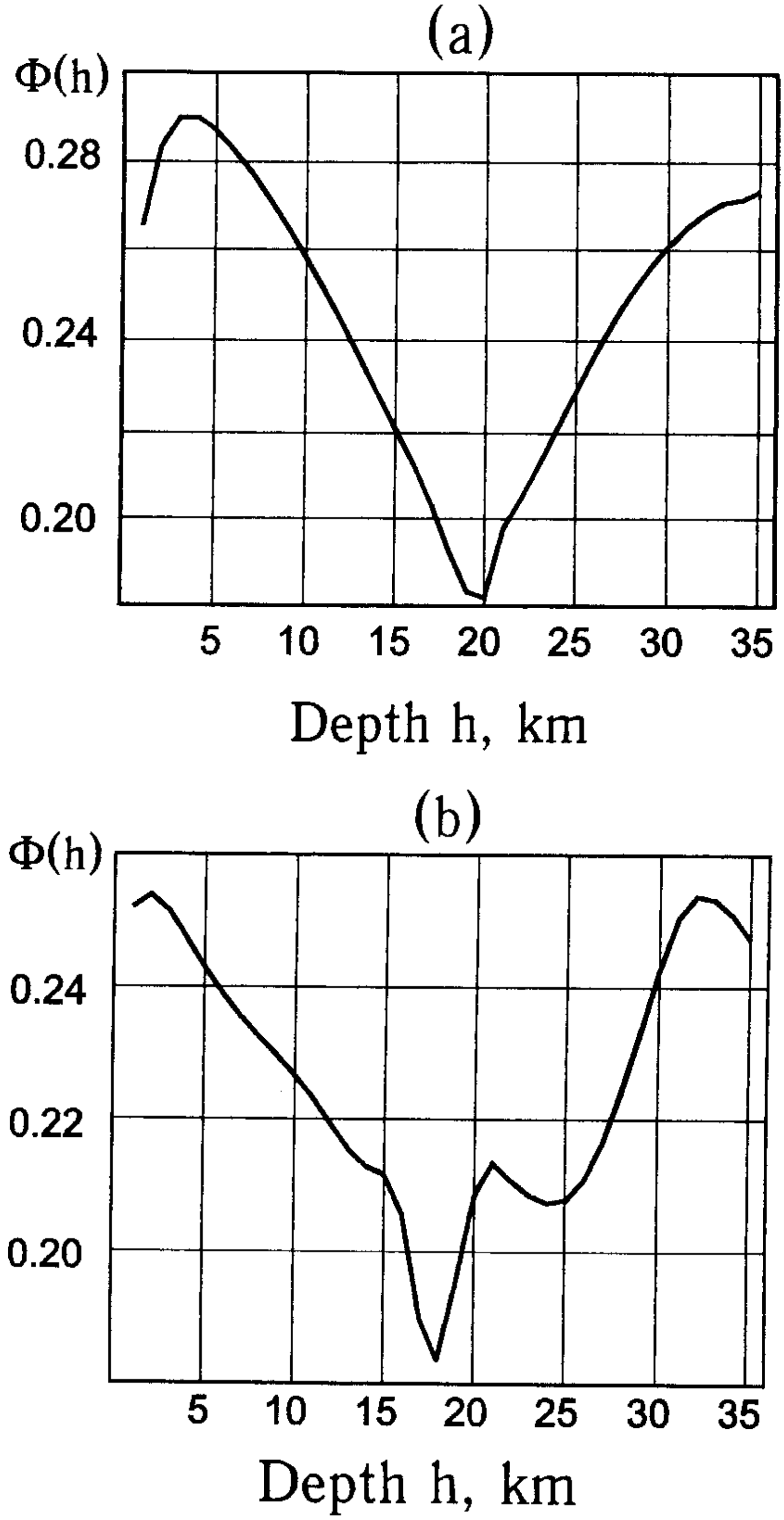
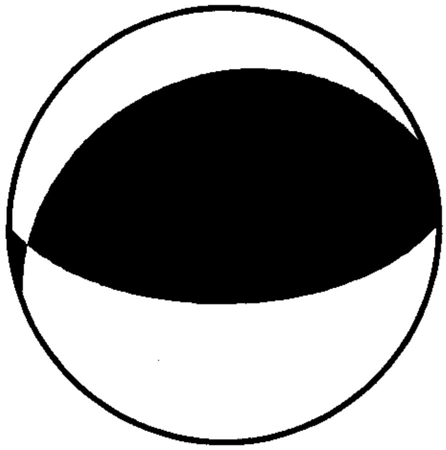


Fig.6. Functions $\Phi(h)$ obtained from synthetic and observed data.
(a) - result of analysis of synthetic spectra;
(b) - result of analysis of real Pwave spectra.

(a)



(b)

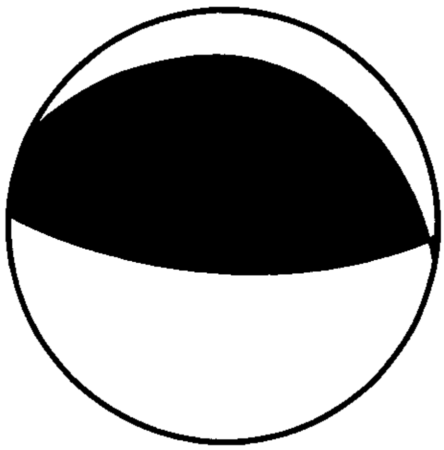


Fig. 9. Fault plane solutions for Susamir earthquake.

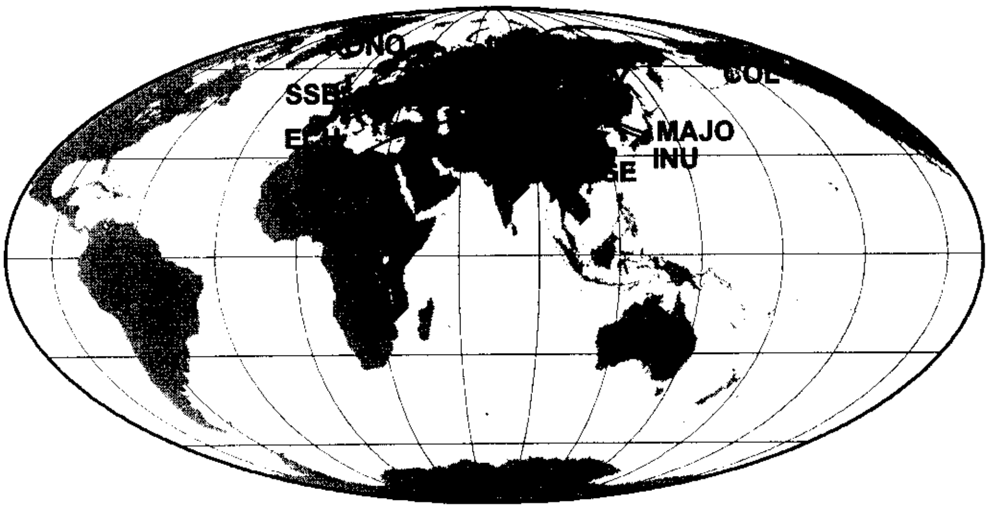


Fig. 8. Distribution of stations used for study of Susamyr earthquake.

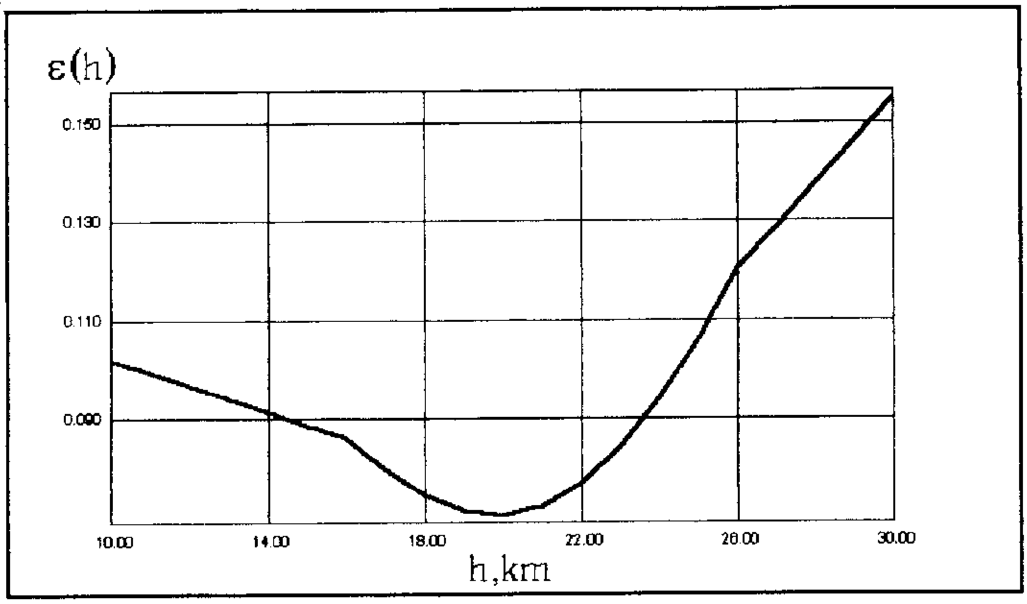


Fig. 10. Residual-depth function

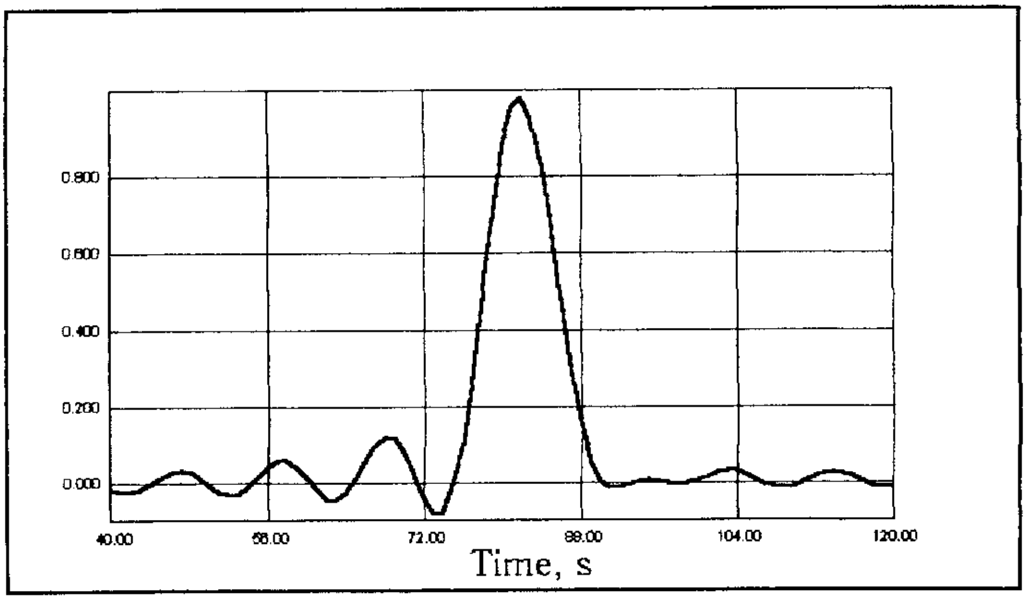


Fig. 11. Source time function