

INVERSION OF INCIDENT ANGLE AND Q VALUE OF SEDIMENTS FROM DEEP BOREHOLE SEISMOGRAMS USING ADATPTIVE SIMULATED ANNEALING METHOD

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SUMMARY

I propose a method to invert frequency-dependent Qs as well as an incident angle at a depth of a downhole sensor and S-wave velocities assuming obliquely incident SH-waves and SV-waves. Transverse, radial, and vertical components of surface-to-downhole spectral ratios are used for the analyses. As an inversion technique, adaptive simulated annealing (ASA) method (Ingber [1,2]) is introduced. This method is applied to surface and deep (GL-1206m) downhole seismograms in the Sendai basin, Japan. The inverted Qs is modeled as $Qs=Vs/32f^{0.57}$ where S-wave velocity Vs in m/s and frequency f in Hz. The inverted incident angle of S-waves agrees with the dominant direction of a particle orbit in a vertical plane calculated from radial and vertical components at GL-1206 m.

INTRODUCTION

To theoretically evaluate site effects of strong motions, it is necessary to obtain information on subsurface structures such as S-wave velocity (Vs) and quality factor for S-waves (Qs), especially for sedimentary layers overlying bedrock. However, it is difficult to estimate Qs for sediments compared with S-wave velocities. Waveform modeling in the time domain is very effective to estimate frequency-independent Qs in the long period range and applied to some deep borehole seismograms (e.g., Bonilla et al. [3]). In the short period range (< 0.5 sec), however, waveform modeling has difficulty in phase fitting with increasing frequency due to scattering effects. In the short period range, frequency-independent Qs is often estimated from the gradient of spectral ratios between two downhole seismograms on the log-linear plots in the frequency domain (e.g., Abercrombie [4]). Fukushima et al. [5] estimate frequency-dependent Qs of sedimentary rocks from spectral ratios between a direct wave and a reflected wave observed at a deep downhole. These methods in the frequency domain are applicable to estimate Qs for homogeneous rock where site amplifications are negligible. Since site amplification effects can not be negligible at sites on soft sediments, these methods would be not suitable to estimate Qs in general.

Since the pioneering work done by Ohta [6], Qs and S-wave velocities of shallow sediments have been inverted from horizontal components of vertical array (downhole array) seismograms by considering onedimensional site amplification effects (e.g., Satoh et al. [7]; Kobayashi et al. [8]). In the methods, the Qs

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Figure 2. P- and S-wave velocity structures surveyed by logging (NIED [11]) and inverted in this study.

and S-wave velocities are inverted to minimize the misfit between observed surface-to-downhole spectral ratios and those computed by one-dimensional wave propagation theory assuming vertically incident S-waves. By introducing nonlinear least square method or generic algorithm to the methods as inversion techniques, Qs of shallow (< 100 m) and soft (Vs < 500 m/s) sediments have been inverted in several regions effects (e.g., Ohta [6]; Satoh et al. [7]; Kobayashi et al. [8]). When the downhole sensors are installed at still soft layers, the assumption of vertical incident of S-waves is a good approximation in most cases. However, when the downhole sensors are installed at hard rock, the assumption would not be acceptable in some cases. Note that the Qs values are much more influential on strong motion prediction with increasing thickness of sediments.

In this study, to estimate *Qs* of deep sediments I propose a method to invert an incident angle at a depth of the downhole sensor as well as S-wave velocities and frequency-dependent *Qs* assuming obliquely incident SH-waves and SV-waves. Transverse, radial, and vertical components of surface-to-downhole spectral ratios are used for the analyses. Adaptive simulated annealing (ASA) method (Ingber [1,2]) is introduced as an inversion technique. ASA is a global optimization algorithm to statistically find the best global fit of a nonlinear cost-function and has been found to be useful through applications for many optimization problems (e.g., Ingber [9]; Yamanaka [10]). The proposed method is applied to seismograms observed at a deep dwonhole array, MYGH01, in the Sendai basin, Japan. This downhole array is one of so called KiK-net stations deployed at about 660 places across the all of Japan by National Research Institute for Earth Science and Disaster Prevention (NIED [11]) mainly from 1998 to 1999. At MYGH01 accelerometers are set at a surface and at a depth of 1206 m which is one of the deepest downholes among the KiK-net stations. Seismograms with high S/N ratios in the wide frequency range were observed during the Miyagiken-oki earthquake (Mw 7.0) occurred in May, 2003 there.



Figure 3. Observed accelerograms and the velocity seismograms integrated from the accelerograms. The S-wave windows with the duration of 15-sec (arrows) are used for the inversion.

DATA

Downhole array site

Figure 1 shows the location of a downhole array station MYGH01 and a fault model of the Miyagiken-oki earthquake (Mw 7.0) by Aoi et al. [12]. This event is a down-dip compression type intraslab earthquake whose focal depth estimated by the Japan Meteorological Agency is 72 km. The epicentral distance to MYGH01 is 86 km. Figure 2 shows the P- and S-wave velocity structures surveyed by velocity logging by NIED [11] and inverted in this paper. The logging results are open in public as temporal results. Digital accelerometers are deployed at GL0m and GL-1206 m. The S-wave velocities by the logging are 210 m/s at GL0m and 3,260 m/s at GL-1206m. The upper 65-m soil consists of sand in the Cenozoic Quaternary era. In the depth range of 65 m to 250-m silt, sandstone, and tuff in the Cenozoic Neogene era exit. The lower 250-m layer consists of shale and sandstone in the Meosozonic Triassic era.

Observed records

Firstly, the horizontal accelerograms observed at MYGH01 during the Miyagiken-oki earthquake are transformed to radial and transverse components. Figure 3 (a) shows the accelerograms and Figure 3 (b) shows the velocity seismograms integrated from the accelerograms. The S-wave windows with the duration of 15 sec are taken from observed records. Then the downhole-to-surface spectral ratios for radial, transverse, and vertical components are calculated for the inversion analyses.

METHOD

Adaptive simulated annealing (ASA) method

Simulated annealing (SA) algorithms are general Monte Carlo approximation methods that are suitable for optimization problems when a desired global minimum is hidden among many poorer local minimums in objective function. Adaptive simulated annealing (ASA) is a global optimization algorithm based on an associated proof that the parameter space can be sampled much more efficiently than by using other previous simulated annealing algorithms (e.g., Ingber [1,2,9]). Here ASA method is explained based on papers by Ingber [1,2].

Individual states vectors $\mathbf{x}(t)$ are formed by *D* scalar parameters $x_i(t)$. The $x_i(t)$ is generated from the following generating function using random variables Δx_i .

$$x_{i}(t) = x_{i}(t-1) + (B_{i} - A_{i})\Delta x_{i}, \quad x_{i} \in [A_{i}, B_{i}]$$
(1)

If a generated $x_i(t)$ is less than A_i or greater than B_i , it is thrown away and a new $x_i(t)$ is regenerated. Δx_i are generated from generating temperature $T_{i,gen}$ and uniform random values u_i ,

$$\Delta x_{i} = sgn(u_{i} - \frac{1}{2})T_{i,gen} \left[\left(1 + 1/T_{i,gen} \right)^{2u_{i} - 1} - 1 \right], \quad \Delta x_{i} \in [-1, 1], \quad (2)$$

where $T_{i,gen}$ is calculated from the temperature reduction function from

$$T_{i,gen}(k_i) = T_{i,gen}(0) \exp(-c_i k_i^{q_i/D}).$$
(3)

Here c_i are constants defining the relative widths of the generating distributions, k_i are annealing indices used for each generating temperature, and q_i are quenching factor. In this study I set $T_{i,gen}(0)=1.0$, $c_i=1$, and $q_i=3$. Quenching factor is used to speed up the calculations without affecting the annealing proof of ASA.

Then the difference ΔE of the objective functions with newly generated values $\mathbf{x}(t)$ and the previous values $\mathbf{x}(t-1)$ is calculated from

$$\Delta E = E(\mathbf{x}(t)) - E(\mathbf{x}(t-1)). \tag{4}$$

The newly generated values $\mathbf{x}(t)$ is accepted if $\Delta E < 0$. If $\Delta E > 0$, $\mathbf{x}(t)$ is accepted according to the probability

$$P = \exp\left(-\frac{\Delta E}{T(k)}\right) \tag{5}$$

where the acceptance temperature T(k) is calculated from

$$T(k) = T(0) \exp(-ck^{q/D}).$$
(6)

Here k are annealing indices used for cost temperature. I set final k of 10,000, T(0)=1.0, c=1.0, and quenching factor q=3. If the uniform random value is less than P, newly generated values x(t) is accepted. Otherwise previous value x(t-1) is accepted.

Under the same cost temperature, five times of iteration is performed. The total number of generating vectors is 50,000. The values of user setting parameters are determined through several numerical experiments and an application to shallow vertical array seismograms (Satoh et al. [13]).

Objective function

Based on analyses of vertical array seismograms it has been pointed out that S-wave windows of vertical components are mainly composed of SV-waves and P-waves converted from obliquely incident SV-waves at the bedrock (Takahashi et al. [14]). Therefore observed surface-to-downhole spectral ratios for vertical components $O_3(f)$ can regard as ratios of vertical components of SV-waves and P-waves $M_3(f)$ converted from obliquely incident SV-waves at the bedrock. In the same way, observed surface-to-downhole spectral ratios for radial components $O_1(f)$ can regard as ratios of horizontal components of SV-waves and P-waves $M_1(f)$ converted from obliquely incident SV-waves at the bedrock. Observed surface-to-downhole spectral ratios for transverse components $O_2(f)$ can regard as ratios of SH-waves M_2 (f) generated from obliquely incident SV-waves at the bedrock. In the computed spectral ratios, incident and reflected waves are taken into account.

Using $O_1(f)$, $O_2(f)$, $O_3(f)$, $M_1(f)$, $M_2(f)$, and $M_3(f)$, the objective function $E(\mathbf{x})$ is defined as

$$E(\mathbf{x}) = \frac{1}{3} \sum_{i=1}^{3} \left(\frac{\sum_{i=i_{\min}}^{i_{\max}} \left| \frac{M_{j}(f_{i}) - O_{j}(f_{i})}{w(f_{i})} \right|}{\sqrt{\sum_{i=i_{\min}}^{i_{\max}} \frac{M_{j}(f_{i})}{w(f_{i})} \sum_{i=i_{\min}}^{i_{\max}} \frac{O_{j}(f_{i})}{w(f_{i})}} + \frac{\sum_{i=i_{\min}}^{i_{\max}} \left| \frac{\log_{10} M_{j}(f_{i}) - \log_{10} O_{j}(f_{i})}{w(f_{i})} \right|}{\sqrt{\sum_{i=i_{\min}}^{i_{\max}} \frac{\log_{10} M_{j}(f_{i})}{w(f_{i})} \sum_{i=i_{\min}}^{i_{\max}} \frac{\log_{10} O_{j}(f_{i})}{w(f_{i})}}}{w(f_{i})}} \right)}$$
(7)

where f_i is *i*th frequency, $f_{i \min}$ and $f_{i \max}$ correspond to minimum and maximum frequencies. In this study f_i min =0.5 Hz and $f_{i \max}$ =10 Hz both of which are determined from S/N ratios. The Partzen window with a width of 0.3 Hz smoothes all of the Fourier spectra used for calculations of $O_1(f)$, $O_2(f)$, $O_3(f)$, $M_1(f)$, $M_2(f)$, and $M_3(f)$. The $w_i(f)$ is a weight function

$$w(f_i) = \frac{1}{(\log_{10} f_i - \log_{10} f_{i-1})}$$
(8)

Parameter Settings

The velocity model based on the logging is composed of five layers with different *Vs* as shown in Figure 2. In the inversion, thickness for each layer is fixed by the values by the logging and *Vs* for each layer is inverted. Density is assumed based on the density and P-wave velocity relationship of $\rho=0.31Vp^{1/4}$ proposed by Gardner et al. [15], where density ρ is in g/cm³ and P-wave velocity *Vp* is in m/s. Since only one downhole seismometer is installed, it is difficult to estimate *Qs* for each layer. For the long period (> 0.5 sec) strong motion simulation, Qs=Vs/10 (m/s) (e.g., Olsen et al. [16]) is commonly used. For site amplification calculation based on one-dimensional wave propagation theory, Qs=Vs/20 or Qs=Vs/30 has been used in the engineering field. In addition, recent previous researches on *Qs* for shallow sediments show the dependence of *Qs* on *Vs* or depth as well as frequency (e.g., Kobayashi et al. [8]; Satoh et al. [17]). Therefore in this study *Qs* is modeled as the function of frequency *f* and *Vs*. (9)

where a and b are model parameters. It is difficult to independently invert
$$Qs$$
 and quality factor for P-waves (Qp) , because sensitivity of Qp on the objective functions is relatively small. Therefore Qp is modeled by using the same model parameters as Qs .

$$Qp = cVp/bf^a.$$
(10)



Figure 4. Comparison of observed and computed surface-to-downhole spectral ratios. Inverted spectral ratios are computed from five different initial values.

Firstly, two cases of inversions with c=1 (Case-1) and c=2 (Case-2) are performed by fixing P-wave velocity of each layer. As Case-3, P-wave velocity of each layer is inverted by fixing c=1. In the Case-1 and Case-2, the total number of model parameters is 8, that is, an incident angle of S-waves at a depth of 1206 m, five S-wave velocities, a, and b. In the Case-3, the total number of model parameters is 13.

The bounds used for the generating function $[A_i, B_i]$ are [0.0, 40.0] for an incident angle in degree, [0.0, 2.0] for a, [0.0, 4.0] for $\log_{10}(b/2)$, [0.8Vs, 1.2Vs] for Vs for the layer with Vp > 1500 m/s, [0.7Vs, 1.3Vs] for Vs for the layer with Vp < 1500 m/s, and [0.8Vp, 1.2Vp] for Vp. The Poisson ratio for each layer is bounded between 0.2 and 0.5. The initial value for each model parameter is generated using the uniform random number within these bounds. The inversions are performed using five different initial values.

		Vs1	Vs2	Vs3	Vs4	Vs5	а	b	ϕ	$E(\mathbf{x})$
		m/s	m/s	m/s	m/s	m/s			degree	
Case-1	AV.	187.8	466.7	2099.8	2939.7	3410.3	0.817	34.2	16.34	0.760
Qp / Qs = Vp / Vs	S.D.	± 0.5	± 11.0	± 0.0	± 0.0	± 11.6	± 0.04	± 2.0	± 1.48	± 0.000
Case-2	AV.	193.1	472.0	2099.8	2889.5	3339.0	0.756	31.0	16.47	0.763
Qp /Qs = Vp /(2Vs)	S.D.	± 11.8	± 7.9	± 0.0	± 0.0	± 148.3	± 0.15	± 2.4	± 0.72	± 0.000

Table 1 Average (AV.) and standard deviations (S.D.) of model parameters in the Case-1 and Case-2.

Table 2 Average (AV.) and standard deviations (S.D.) of model parameters in the Case-3.

Case-3.										
		Vs1	Vs2	Vs3	Vs4	Vs5	а	b	φ	$E(\mathbf{x})$
		m/s	m/s	m/s	m/s	m/s			degree	
Case-3	AV.	214.6	595.3	1582.6	2938.9	3619.3	0.570	31.8	17.23	0.744
Qp /Qs = Vp /Vs	S.D.	± 7.9	± 92.8	± 217.3	± 0.0	± 342.5	± 0.146	± 3.0	± 2.95	± 0.006
and		Vp1	Vp2	Vp3	Vp4	Vp5				
<i>Vp</i> are inverted		m/s	m/s	m/s	m/s	m/s				
	AV.	1401.2	1741.4	2905.5	5388.7	6207.1				
	S.D.	± 11.8	± 7.9	± 0.0	± 0.0	± 148.3				

INVERSION RESULTS

Frequency domain inversions

Figure 1 compares the observed and computed surface-to-downhole spectral ratios for three cases. The first peak frequencies of observed spectral ratios are different between radial and transverse components. The difference is reproduced by the computed spectral ratios by considering obliquely incident SV- and SH-waves. The average and the standard deviation of model parameters inverted using five initial values are shown in Tables 1 and 2 where *Vsi* is *i*th layer S-wave velocity. In the Case-1 and Case-2, Qp/Qs=Vp/Vs and Qp/Qs=Vp/(2Vs) are assumed, respectively, as derived from equations 9 and 10.

The difference of spectral ratios computed for Case-1 and Case-2 is small because the model parameters inverted for Case-1 and Case-2 are almost the same within the standard deviations as shown in Table 1. The objective function value $E(\mathbf{x}(t))$ in the Case-1 is slightly smaller that in the Case-2. The relation of Qp/Qs=Vp/Vs in the Case-1 is comparable with Qp/Qs=9/4 for intrinsic attenuation in a dry, elastic, Poisson solid (e.g., Aki and Richards [18]). However, the difference of $E(\mathbf{x}(t))$ is not well constrained due to small sensitivity of Qp on the $E(\mathbf{x}(t))$. To estimate well constrained Qp or Qp/Qs, the use of P-wave windows will be helpful.

In the Case-3, P-wave velocities are inverted by assuming Qp/Qs=Vp/Vs. The agreement between observed and computed spectral ratios in the Case-3 is better than that in the Case-1, especially for vertical components. The objective function value $E(\mathbf{x}(t))$ in the Case-3 is also smaller that in the Case-1. The spectral ratios computed from five initial values have greater variation in the Case-3 than the Case-1. The standard deviations of inverted model parameters are also greater in the Case-3 than the Case-1. These differences may be caused by the trade-off between model parameters in the Case-3 because the number of model parameters in the Case-3 is greater than that in the Case-1.



The inverted Vs and Vp in Case-1 and Case-3 are shown in Figure 2 as mentioned before. Figure 5 shows a particle orbit in a vertical plane calculated from radial and transverse components of downhole records bandpass-filtered in the frequency range from 0.5 to 10 Hz. Although the shape of particle orbit is complex maybe due to scattering effects, the incident angle ϕ can be estimated to about 15 to 20 degrees from the most dominant direction. The angles agree with the inverted values of 17-degree (See Tables 1 and 2).

The inverted Qs in Case-1, Case-2, and Case-3 are shown in Figure 6 as the values of Qs/Vs. The Qs/Vs in the Case-1 and Case-2 is almost the same. The differences of Qs in the Case-1 and Case-3 are not so large whether P-wave velocities are inverted or not. In Figure 6, an empirical Qs/Vs relation for sedimentary rocks proposed by Kobayashi et al. [8] using seismograms observed by shallow (< 100 m) vertical arrays at about 10 sites in the Sendai basin. It is found the Qs for deep sediments inverted in this study is larger than the Qs for shallow sedimentary rocks in the same basin. The depth dependence of Qs can say as pressure dependence. The pressure dependence of Qs and Qp has been interpreted as resulting from the closure of cracks in the rocks with increasing confining pressure (e.g., Abercrombie [19]).

In Figure 7 the Qs models ($Qs=Vs/32f^{0.57}$) inverted in the Case-3 are shown for S-wave velocities of 300 m/s, 1,000 m/s, and 3,000 m/s. Frequency-dependent Qs models derived from deep downhole seismograms in the other regions (Kinoshita [20]; Fukushima et al. [5]; Takemura et al. [21]) are also shown in Figure 7. All Qs models are proportional to $f^{0.5 \text{ to } 1.0}$. Absolute Qs values have some scatter among the models in the different regions even if Vs difference is considered. This result suggests that Qs values are strongly site dependent.



Figure 7. *Qs* models inverted in the Case-3 ($Qs=Vs/32f^{0.57}$) and derived from deep downhole seismograms in the other regions (Kinoshita [20]; Fukushima et al. [3]; Takemura et al. [21]).

Time domain synthetics

Using the relation of $Qs=Vs/32f^{0.57}$ and inverted P- and S-wave velocity structures, three components of velocity seismograms at GLOm are computed from the S-wave widow of velocity seismograms observed at GL-1206m assuming obliquely incident SH- and SV-waves with the inverted incident angle of 17 degree. The top traces in Figure 8 are the synthetic velocity time histories at GLO m. The middle and bottom traces are the seismograms observed at GLO m and GL-1206 m, respectively. All seismograms are bandpass-filtered in the frequency range from 0.5 to 10 Hz. The phases and amplitudes match between the data and the synthetics reasonably well, though the matching of vertical components is worse than that of horizontal components. This may be due to the contamination of P codas to the S-wave window of vertical components (See Figure 2).

CONCLUSIONS

In this study I propose a method to invert frequency-dependent Qs we well as an incident angle at a depth of the downhole sensor and S-wave velocities assuming obliquely incident SH-waves and SV-waves. Transverse, radial, and vertical components of surface-to-downhole spectral ratios are used for the analyses. As an inversion technique, adaptive simulated annealing (ASA) method (Ingber [1,2]) is introduced.

This method is applied to surface and deep (GL-1206m) downhole records in the Sendai basin, Japan. The inverted incident angle of S-waves, 17 degree, agrees with the dominant direction of a particle orbit in a vertical plane calculated from radial and vertical components at GL-1206 m. The inverted Qs is modeled as $Qs=Vs/32f^{0.57}$. It is found that the Qs model for deep sediments is much larger than the Qs model for shallow sediments derived by Kobayashi et al. [8] in the same Sendai basin. This result can be interpreted as pressure dependence of Qs, which has been interpreted as resulting from the closure of cracks in the rocks with increasing confining pressure (e.g., Abercrombie [19]). The inverted Qs model reproduces three components of observed seismograms in the time domain reasonably well. However, Qp is not well



Figure 8. The top traces are the synthetic velocity time histories at GL0 m computed from the S-wave widow of velocity seismograms observed at GL-1206m using the inverted $Qs=Vs/32f^{0.57}$, velocity structures, and incident angle of 17 degree. The middle and bottom traces are the velocity seismograms of S-wave windows observed at GL0 m and GL-1206 m, respectively.

constrained in this study. In the future research, the use of P-wave windows will be helpful to invert Qp quantitatively.

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REFERENCES

- 1. Ingber L. Very fast simulated reannealing. Math. Comput. Modeling, 1989, Vol.2: 967-973.
- 2. Ingber L. Adaptive simulated annealing. http://www.ingber.com/ ASA-README.pdf, 2002.
- 3. Bonilla LB, Steidl JH, Gariel J-H, Archuleta RJ. Borehole response studies at the Garner Valley Downhole Array, Southern California. Bull. Seism. Soc. Am., 2002, Vol.92: 3165-3179.
- 4. Abercrombie RE. Near-surface attenuation and site effects from comparison of surface and deep borehole recordings. Bull. Seism. Soc. Am., 1997, Vol.87: 731-744.
- 5. Fukushima, Y, Kinoshita S, Sato H. Measurements of Q⁻¹ for S waves in mudstone at Chikura, Japan : Comparison of incident and reflected phases in borehole seismograms. Bull. Seism. Soc. Am., 1992, Vol.82: 148-163.
- 6. Ohta Y. Application of optimization technique for earthquake engineering. Journal of Struct. Eng., Transactions of Architectural Institute of Japan, 1975, Vol.299: 35-41 (in Japanese).
- 7. Satoh T, Kawase H, Sato T. Evaluation of local site effects and their removal from borehole records observed in the Sendai region, Japan. Bull. Seism. Soc. Am., 1995, Vol.85: 1770-1789.

- Kobayashi K. Estimation of velocity structures and attenuation characteristics of sedimentary layers. Proc. 27th Symposium of Earthquake Ground Motion, Architectural Institute of Japan, 1999: 29-39 (In Japanese with English abstract).
- 9. Ingber L. Rosen BE. Genetic algorithms and very fast simulated reannealing : A comparison. Math. Comput. Modeling, 1992, Vol.16: 87-100.
- Yamanaka H. Application of simulated annealing to inversion of surface wave phase velocity -Comparison of performances between SA and GA inversions -. Butsuri-Tansa, 2001, Vol.54: 197-206 (in Japanese with English abstract).
- 11. NIED. Kiban-Kyoshin Network. http://www.kik.bosai.go.jp/kik/index_en.shtml.
- 12. Aoi S, Sekiguchi H, Kunugi T, Honda R, Fujiwara H. Ground Motion and Rupture Process of the 2003 Off-Miyagi Earthquake Obtained from Strong Motion Data of K-NET and KiK-net. http://www.k-net.bosai.go.jp/k-net/news/miyagi_e/.
- 13. Satoh T. Identification of damping factors of soil from vertical array data considering obliquely incident S waves. J. Struct. Constr. Eng., Transactions of Architectural Institute of Japan, 2003, Vol.569: 37-45 (in Japanese with English abstract).
- 14. Takahashi K. Ohno S. Takemura M. Ohta T. Sugawara Y. Hatori T. Omote S. Observation of earthquake strong-motion with deep borehole generation of vertical motion propagation in surface layers after S wave arrival. Proc. 10th World Conference of Earthquake Engineering, 1992, 1245-1250.
- 15. Gardner GH, Gardner LW, Gregory AR Formation velocity and density the diagnostic basics for stratigraphic traps, Geophysics, 1974, Vol.29: 770-780.
- Olsen KB, Nigbor R, Konno T. 3D viscoelastic wave propagation in the upper Borrego Valley, California, constrained by borehole and surface data. Bull. Seism. Soc. Am., 2000, Vol.90: 134-150.
- 17. Satoh T, Hayakawa T, Sato T. Underground structure inferred from observed records." Proc. first Symposium of study on the master model for strong ground motion prediction toward earthquake disaster mitigation, 2002: 61-68 (in Japanese).
- 18. Aki K, Richards PG. Quantitative Seismology, second edition, University Science Books, 2002, pp.700.
- 19. Abercrombie RE. A summary of attenuation measurements from borehole recordings of earthquakes: The 10 Hz transition problem. Pure and Applied Geophysics, 1998, Vol.153: 475-487.
- 20. Kinoshita S. Frequency dependent attenuation of shear waves in a sedimentary layers-basement system in the Kanto area, Japan. Proc. International Symposium on Earthquake Disaster Prevention, Mexico, 1992, Vol. I: 212-226.
- 21. Takemura M, Ikeura T, Takahashi K, Ishida H, Ohshima Y. Attenuation characteristics of seismic waves in sedimentary layers and strong motion estimation. J. Struct. Constr. Eng., Transactions of Architectural Institute of Japan, 1993, Vol.446: 1-11 (in Japanese with English abstract).