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S-wave velocity structure estimated from long-period coda waves

Hitoshi Oda*and Aki Ito**

*Department of Earth Sciences, Okayama University,
Okayama 700-8530, Japan
**Observation Center for Prediction of Earthquakes and
Volcanic Eruptions, Tohoku University,
Sendai 980-8578, Japan

Long-period coda waves were recorded on the vertical-component seismograms of aftershocks of the Hyogo-ken Nanbu earthquake, 1995. We identify the long-period coda waves as Rayleigh wave, because they appear after the S-arrival times and exhibit the normal dispersion that propagation velocity of the coda waves increases with an increase in period. By applying the moving window analysis to the coda waves from nine aftershocks, the group velocities are determined as a function of period within the range of 2 to 8 s. The group velocity dispersion data are inverted to investigate the S-wave velocity structure of the upper crust. The S-wave velocity structure is consistent with those obtained in previous studies using travel time analysis of body waves.

Keywords: surface wave dispersion, coda wave, S-wave velocity structure, upper crust, Hyogo-ken Nanbu earthquake

1. Introduction

The coda wave that appears after the arrival time of S-wave from near local earthquake provides us with information of the earthquake source mechanism, seismic velocity structure and attenuation of the crust (e.g. Aki and Chouet, 1975, Sato and Fehler, 1998). The long-period coda wave has been recognized to be a kind of surface waves, while the short-period coda wave is considered to arise from scattering of the S-wave due to structural heterogeneity. If the group velocity of the long-period coda waves is measured as a function of period, the dispersion curve is useful for estimating the seismic velocity structure beneath a region including station and earthquake hypocenter. However the excitation of the long-period coda waves by small local earthquakes is usually so weak that it has been difficult to measure the dispersion curve of surface wave. An advantage in using surface waves for studying the seismic velocity structure is that the structure can be determined only from seismic records at a single station, whereas a number of seismic stations are required for the study of velocity structure by travel time analysis of the P- and S-waves. In addition, the short period surface waves are suitable for estimating the S-wave structure in shallow depth, because the surface wave group velocity is sensitive to the S-wave velocity structure.

We found long-period coda waves on the seismograms of aftershocks of the Hyogo-ken Nanbu earthquake in 1995. In this study, the coda waves are identified as Rayleigh wave and their group velocities are determined by the moving window method (Landisman et al., 1969) as a function of period within the range of 2 to 8 s. The group velocity data are inverted to estimate the S-wave structure of the upper crust, and the structure is compared with those reported in previous studies.

2. Waveform data and Rayleigh wave group velocities

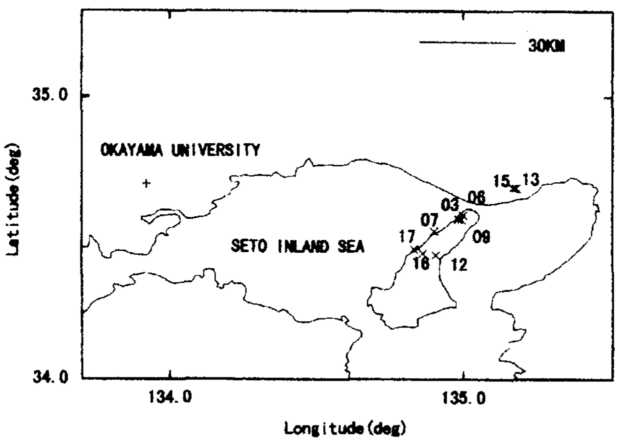

Fig. 1. Epicentral distribution of aftershocks (cross) and station location (plus). The numbers correspond to the event numbers listed in Table 1.

Table 1. Source parameters and magnitudes of aftershocks

No.	Date	Latitude	Longitude	M	Dep. (km)
3	Jan. 17	34.57° N	134.96° E	4.2	11
6	Jan. 17	34.59° N	135.00° E	4.5	19
7	Jan. 17	34.53° N	134.90° E	4.2	10
9	Jan. 17	34.57° N	134.99° E	4.3	12
12	Jan. 18	34.44° N	134.90° E	4.0	5
13	Jan. 18	34.68° N	135.18° E	4.3	16
15	Jan. 18	34.68° N	135.17° E	4.3	13
16	Jan. 18	34.45° N	134.87° E	4.2	8
17	Jan. 18	34.46° N	134.83° E	4.1	10

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Seismic waves from aftershocks of the Hyogo-ken Nanbu earthquake were recorded by three component seismometers with natural period of 5 s at Okayama University. Only the vertical-component seismograms were analyzed because of trouble in the recording system of horizontal components. We selected records of nine events of which amplitudes were not scaled off. The source parameters of the events are listed in Table 1. The epicentral distances are less than 100 km, and the focal depths fall between 5 and 20 km. The earthquake magnitudes are between 4.0 and 4.5. The epicenters of the nine events and seismic station are shown in Fig. 1. Figure 2 shows the vertical-component seismograms of #3 and #9 events. Long-period coda waves appear after the S-wave arrival in the vertical component and they exhibit the characteristic of normal dispersion that propagation velocity becomes higher for longer period, though the coda waves are contaminated with short-period seismic waves. Therefore we identify the long-period coda waves as Rayleigh wave.

The moving window method (Landisman et al., 1969) was applied to the long period coda waves in order to determine the group velocities as a function of period. In

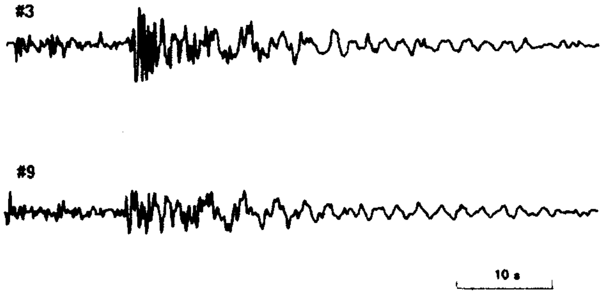

Fig. 2. Vertical-component seismograms of #3 and #9 events.

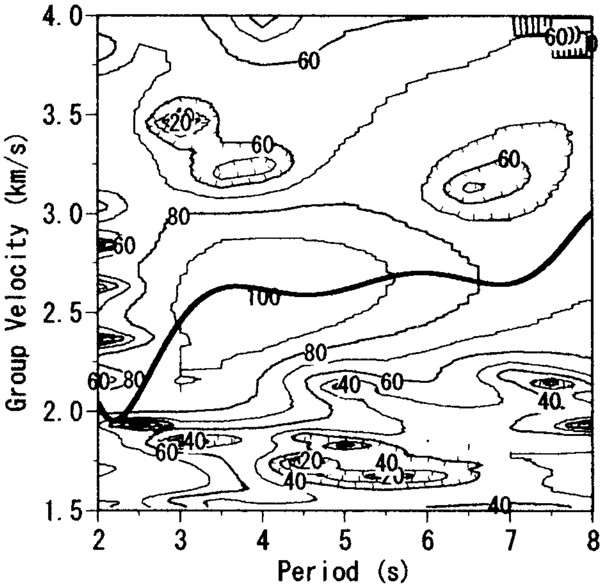

Fig. 3. Wave energy arrival time chart of #3 event by the moving window analysis. The contours with interval of 20 dB correspond to the isolines of the amplitude level. Thick line represents the group velocity dispersion curve.

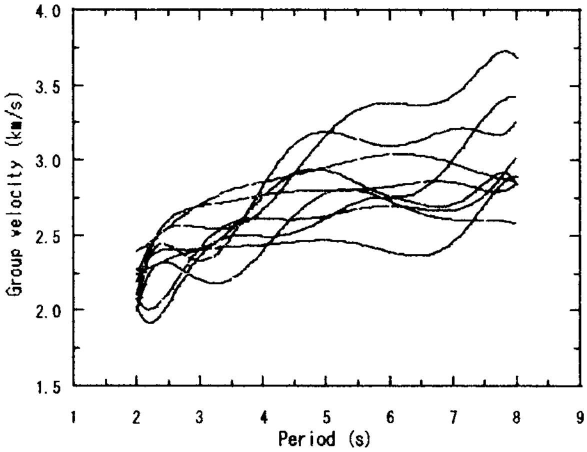

Fig. 4. Dispersion curves of Rayleigh wave group velocity for nine events listed in Table 1.

the method, the width of time window was set so as to be three times a given period. The group velocity or wave energy arrival time was determined at every 0.5 s in the period range of 2 to 8 s. The group velocity dispersion curve for #3 event is shown in Fig. 3, where the wave energy is displayed in dB as a time series at every 0.14 s for a given period and the contours corresponding to the isolines of the amplitude level are drawn with 20 dB interval. The group velocity dispersion curve is obtained by tracing the peak wave energy against period. Figure 4 shows the dispersion curves measured for the nine events. Although the group velocities are scattered and roughly change with respect to period, they systematically increase

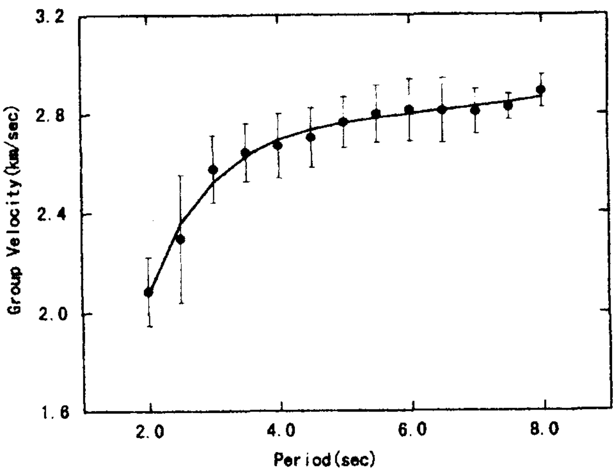

Fig. 5. Average group velocity of the dispersion curves of #03, #09, #13, and #16 events. Bar denotes the standard deviation. Theoretical group velocity calculated for the final velocity model in Table 2 is shown by the solid curve.

Table 2. Seismic velocity structure estimated by the group velocity inversion.

Layer No.	Vp (km/s)	Vs (km/s)	Density (g/cm^3)	Thickness (km)
1	3.10	2.03	2.15	0.9
2	5.50	3.17	2.27	1.1
3	5.90	3.45	2.65	4.0
4	6.25	3.54	2.70	7.0
5	6.86	3.88	2.90	19.0
6	8.10	4.20	3.10	-

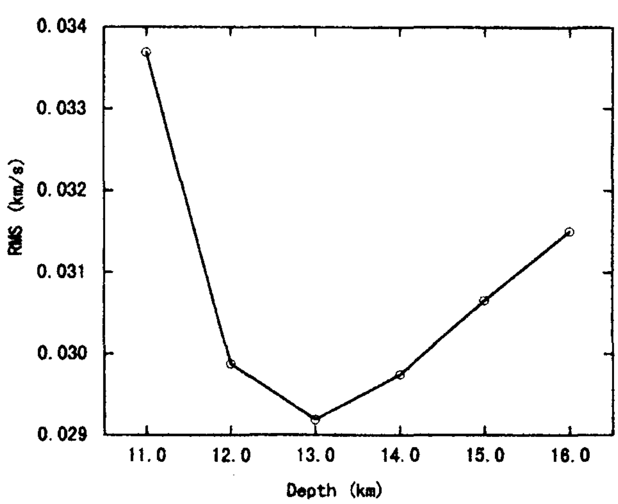

Fig. 6. RMS variation of group velocity residuals against depth of the Conrad discontinuity.

with an increase in period. The scatter of the group velocity may arise from lateral heterogeneity of seismic wave velocities. Further the rough change of the group velocity with respect to period may be due to contamination with refracted waves, reflected waves and short period coda waves.

3. Result and discussion

In order to smooth the group velocities roughly changing with respect to period, we further choose four dispersion curves that are between 85 and 115% of the mean value of group velocities measured for the nine events. The average group velocity of the four dispersion curves is plotted as a function of period (see Fig. 5). The group velocity drastically increases up to a period of 4 s and then slowly increases. The drastic increase at short periods is interpreted as being due to rapid increase in seismic wave velocities with increasing depth.

The group velocity data are inverted to investigate the S-wave velocity structure beneath the region between Okayama and Kobe. Because the Rayleigh wave group velocity is more sensitive to S-wave velocity than to P-wave velocity although it depends on both the velocity structures, we keep the P-wave velocities invariant during the group velocity inversion. We adopted the P-wave velocity model determined by explosion experiments in the Chugoku and Shikoku districts (e.g., Aoki and Muramatsu, 1974; Piao et al., 1996). Let U^c be the group velocity calculated for a given model of horizontally layered velocity structure. The difference of the calculated group velocity from observed one, U^o, at a period T_k is expressed by

$$U^o(T_k) - U^c(T_k) = \sum_{i=1}^{N} \frac{\partial U^c}{\partial \beta_i} \delta\beta_i \qquad (1)$$

where N is the number of layers and $\delta\beta_i$ is the correction for the S-wave velocity of the i-th layer. Since U^o, U^c and $\partial U^c / \partial\beta_i$ are known for the given velocity model, the unknown parameter $\delta\beta_i$ is determined by solving, in a least squares sense, Eqs. (1) formed for a set of group velocity data. The S-wave velocity structure is obtained by adding the velocity correction $\delta\beta_i$ to the initial value of the S-wave velocity in each layer. We used the computer program 'DISPER80' (Saito, 1988) in order to calculate the group velocity and its velocity derivatives. The group velocity data may have ability to estimate the velocity structure in the upper crust, because the wavelength of the surface wave is shorter than about 25 km in the present period range. The S-wave velocity model obtained by the group velocity inversion is shown in Table 2. Theoretical group velocities calculated for the velocity model is illustrated as a function of period in Fig. 5. Agreement between the calculated and observed group velocities is excellent. The boundary interface between the fourth and fifth layers corresponds to the Conrad discontinuity.

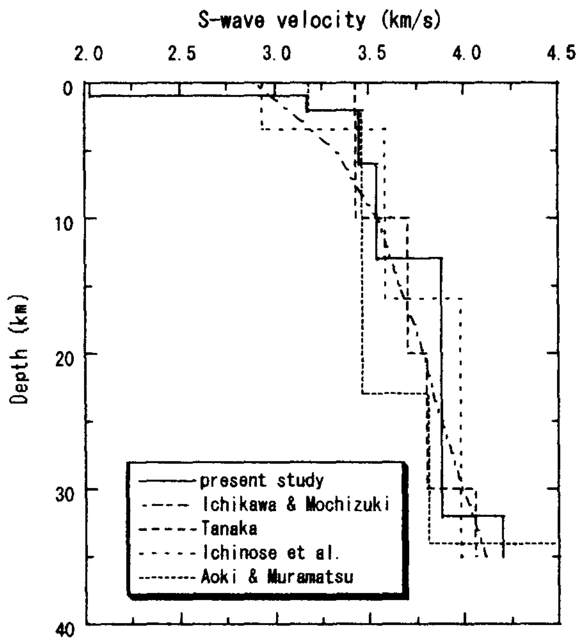

Fig. 7. Comparison between the S-wave velocity structures by group velocity inversion and travel time analysis (Ichikawa and Mochizuki ,1971; Tanaka ,1987). The S-wave velocities of Ichinose et al. (1982) and Aoki and Muramatsu (1974) are estimated from the P-wave velocity by assuming Vp/Vs=1.73.

Figure 6 shows the RMS variation of the group velocity residuals against depth of the discontinuity. Because the RMS has the minimum value when the Conrad discontinuity is at 13 km depth, we assign the depth to the Conrad discontinuity. The S-wave velocity structure by the group velocity inversion is compared with those of the previous studies using travel time analysis of P- and S-waves (see Fig. 7). The S-wave velocity falls within a range of scatter of those reported in the previous studies. This result indicates justification of our velocity model and identification of the long-period coda waves.

The effect of a liquid layer on Rayleigh wave group velocity is briefly examined, because the ray paths from most events cross over the Seto Inland Sea (see Fig. 1). The depth of sea bottom along the ray paths is less than 0.2 km. Assuming that a liquid layer with a thickness of 0.1 km is on the uppermost layer of the velocity model of Table 2, we calculated group velocity dispersion curve of the Rayleigh wave. The liquid layer decreases the group velocity by about 0.1 km/s or less at the period ranging from 2 to 8 s, and it is less influential on the group velocity dispersion curve. Thus the effect of the Seto Inland Sea does not have to be taken into account for the group velocity inversion for estimating the seismic velocity structure.

4. Conclusions

Long-period coda waves after the S-wave arrival were recorded on the vertical-component seismograms of aftershocks of the Hyogo-ken Nanbu earthquakes in 1995. We identified the coda waves as the Rayleigh wave, because they exhibited a characteristic of surface wave, normal dispersion, and appeared in the vertical component. By applying the moving window analysis to the long period coda waves, the Rayleigh wave group velocities were determined for nine aftershocks within the period range of 2 to 8 s. The group velocity dispersion data were inverted to investigate the S-wave velocity structure of the upper crust. The S-wave velocity drastically increases down to 2 km depth, and below the depth it gradually increases. The depth of the Conrad discontinuity was determined to be 13 km in the region between Okayama and Kobe. The S-wave velocity is 3.54 km/s above the discontinuity and 3.88 km/s below it. The obtained S-wave velocity structure was consistent with those obtained in previous studies using travel time analysis of body waves. The long period coda waves can be used for the study of the seismic velocity structure.

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