Jour. Fac. Sci., Hokkaido Univ., Ser. VII (Geophysics), Vol. 8, No. 5, 449-464, 1990.

Strong Ground Motions from Intermediate-Depth Earthquakes : A Study of Site Effects

Tsutomu Sasatani

Department of Geophysics, Faculty of Science,
Hokkaido University, Sapporo 060, Japan

(Received November 14, 1989)

Abstract

In this paper we study the site effects on long-period (1 to 10 sec) ground motions at various sites in Hokkaido, Japan, using JMA strong motion records from near-by, intermediate-depth earthquakes. First the source parameters of the events are determined from a study of teleseismic body waves. Then synthetic strong-motion seismograms are calculated for the source parameters obtained, assuming a plausible plane-layered earth model, and they are compared with the observed records to investigate the site effect. Body waves from intermediate-depth events can be explained fairly well by the synthetics. However, the long-period coda waves with a period of several seconds, which appear after S arrival at stations within the alluvial plains, cannot be explained by the simple synthetics. They are considered to be surface waves locally generated on the irregular, sediment-bedrock interface due to the incident S wave. We investigate the mode of excitation of the long-period coda waves by means of the envelope shape of the coda. The envelope shape varies from station to station and may represent an index of the site effect.

1. Introduction

The extensive damage to high-rise buildings in Mexico City during the 1985 Michoacan earthquake and oil-sloshing in the huge tanks at Niigata due to the 1983 Nihonkai-Chubu earthquake have been unprecedented earthquake disasters. They are considered to have been caused by the abnormal excitation of long-period (1 to 10 sec) strong motion within alluvial plains with thick sediment (Beck and Hall, 1986 ; Kudo and Sakaue, 1984). At present, one of the most urgent subjects in earthquake engineering is to establish the effects of surface geological conditions on long-period ground motions, that is, the site effects, in view of the recent increase of large structures such as high-rise buildings, oil tanks, and suspension bridges.

The factors affecting ground motions are mainly divided into three categories : source, propagation path, and site response. Usually the response of the alluvial site has been determined against that of the rock site by dividing the amplitude spectrum of the alluvial site ground motion for a given event by the amplitude spectrum of the rock site ground motion for the same event (e.g., Jarpe et al., 1988). This approach assumes similar source radiation and path properties to both sites. More directly, the response of sedimentary layers has been determined by means of a deep downhole seismometer array (Kinoshita et al., 1986; Hauksson et al., 1987). This uses seismic waves traveling almost vertically up the downhole array.

In this paper we apply the third method to investigate the site response, which can be used even at sites with only a single station and no downhole array. The method compares synthetic strong-motion seismograms, calculated taking the source and propagation effects into account, with the observed ones. This method requires that source and propagation path effects can actually be evaluated accurately. Thus we have to use seismic waves radiated from a simple source process and arriving at the basement through a simple structure. Strong motion records from near-by, moderate size, intermediate-depth earthquakes may be the most suitable data for this study.

We investigate the site effects at various sites in Hokkaido, Japan, by using

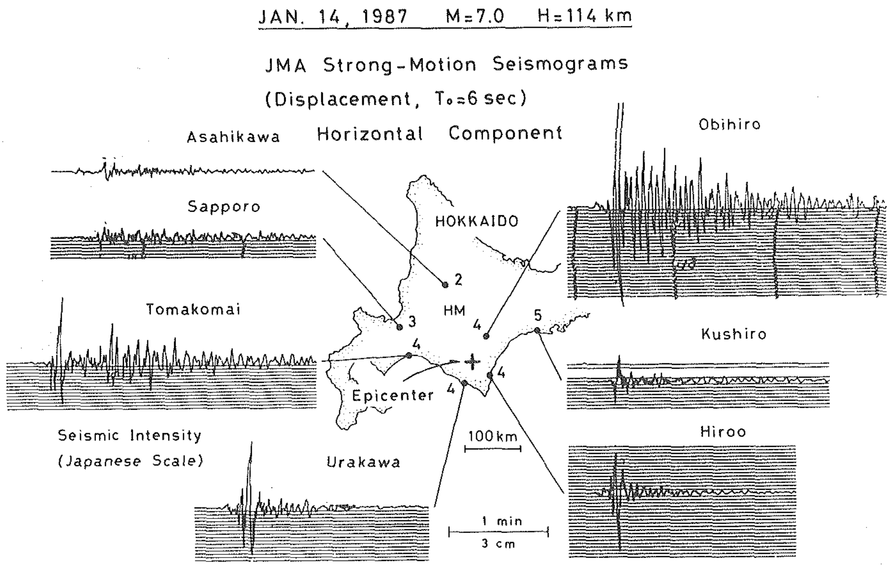

Fig. 1. An example of JMA strong-motion records of the 1987 event, showing strong variations in the ground motion characteristics from station to station. HM, Hidaka Mountains.

Table 1. Earthquake data and source parameters.

	JAN. 14, 1987	MAR. 06, 1984
Epicenter	42.54°N, 142.92°E	42.53°N, 142.93°E
Focal depth	114 km	118 km
Magnitude	7.0	6.0
Fault plane		
strike	265° (262°*)	261°*
dip	90° (83°*)	84°*
rake	−90° (−86°*)	−106°*
Source process	1.5 sec (the first shock)	1.5 sec
time	3.0 sec (the second shock)	
Seismic moment	1.26 (1.7*)×10^{26} dyne cm	3.83* ×10^{24} dyne cm

* =Centroid moment tensor solution by Dziewonski et al., (1984, 1988).

JMA strong motion records from two intermediate-depth earthquakes occurring in 1984 (M=6.0) and 1987 (M=7.0) beneath the Hidaka Mountains (Fig. 1 and Table 1). Figure 1 shows an example of JMA strong motion seismograms from the 1987 intermediate-depth earthquake. These exhibit strong variations in the ground motion characteristics from station to station, which may represent the various site effects. After comparison between the synthetic and observed seismograms, we investigate the ground motion characteristics through complex seismic trace analysis.

2. Source Process

Here we determine the source parameters such as seismic moment and source time function for the 1987 event, from forward modelling of teleseismic body waves. We use the simple ray approach described in Bouchon (1976). Crust-upper mantle models in Table 2 were assumed in the wave calculation. The model for the surce region was based on the crustal structure in the profile across the southern part of Hokkaido (Okada et al., 1973) and upper mantle structure for the northern Japan (Suzuki, 1978).

Figure 2 shows the focal mechanism of the 1987 event determined based on P wave polarities, S wave polarization angles, and waveform modelling. One nodal plane is vertical and the other, horizontal. This solution is approximately the same as that determined by Dziewonski et al. (1988).

Figure 3(a) shows an example of teleseismic long-period P waves. The detailed inspection of the direct P wave reveals that its amplitude suddenly

Table 2. Crust-upper mantle models used in the wave calculation.

Source Reigon				Teleseismic Receiver			
V_p km/s	V_s km/s	ρ g/cm^3	H km	V_p km/s	V_s km/s	ρ g/cm^3	H km
4.0	2.4	2.4	1	6.1	3.5	2.7	11
5.9	3.4	2.7	14	6.4	3.68	2.9	9
6.6	3.7	2.9	15	6.7	3.94	2.9	18
7.7	4.45	3.2	50	8.15	4.75	3.3	—
8.0	4.5	3.3	—				

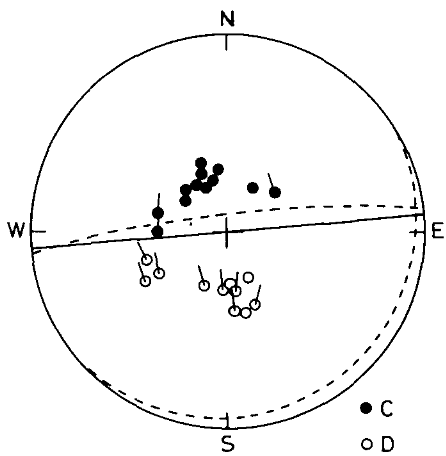

Fig. 2. Focal mechanism solution (solid curves) of the 1987 event. The first-motion polarities (open circles are dilatations and filled circles are compressions) are shown in the lower hemisphere of an equal area projection. Arrows indicate the S wave polarization angles. The dashed focal mechanism was determined by Dziewonski et al. (1988).

increases about 3 sec after the onset. This indicates that the event is a multiple shock, that is, a doublet (Sasatani, 1980). Taking this fact into account, we modelled observed P waveforms assuming a point source, and obtained the source time function (Fig. 3(b)) that provided optimal waveform fit. The two trapezoidal source functions have rise times, flat times and fall times of (0.5, 0.5, 0.5) and (1.0, 1.0, 1.0) in seconds respectively, with the monent of the second shock being about 5 times that of the first. The time separation between two shocks is 2 sec for IST (northern stations) and 3 sec for SNG (southern stations), showing a slight azimuthal dependence. In the next section we take the

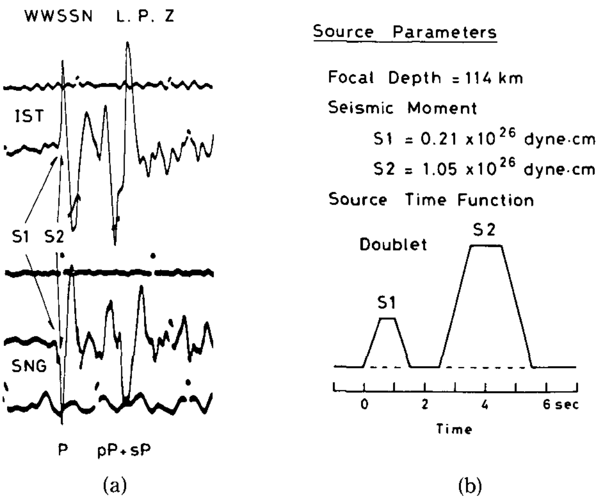

Fig. 3. (a) An example of the WWSSN long-period P seismograms of the 1987 event, showing the multiple shock (S1 and S2 events). (b) Source time function obtained from forward modelling of P waveform.

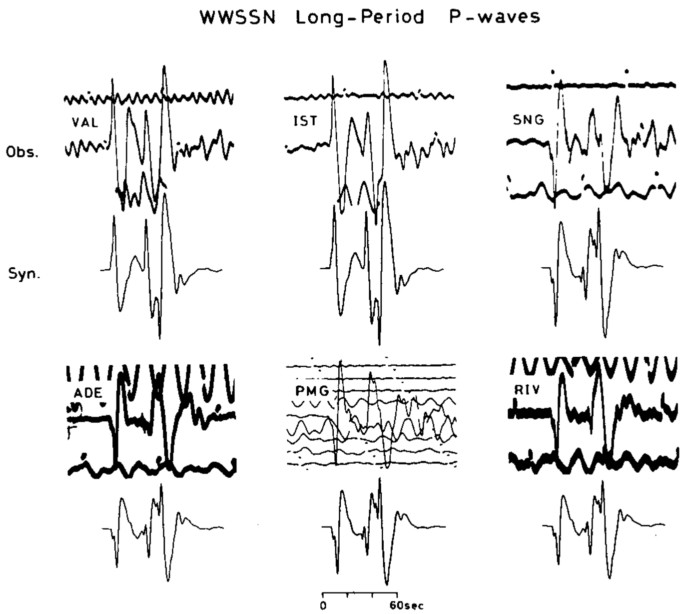

Fig. 4. Long-period P wave observations (upper traces) and matched synthetics (lower traces) for the 1987 event.

Table 3. WWSSN station data and seismic moment. Δ is the epicentral distance ; A_z, the epicenter-to-station azimuth ; and M_0, the seismic moment.

Station code	Δ deg.	A_z deg.	M_0, $\times 10^{26}$ dyne cm		
			P (UD)	S (NS)	S (EW)
ADE	77	184	1.50	1.18	1.07
AKU	71	352	1.31	—	—
DUG	73	51	1.68	1.13	1.70
HKC	31	239	1.28	—	—
HLW	84	306	2.18	—	—
HNR	54	159	—	0.76	1.04
IST	78	315	1.31	—	—
PMG	52	175	0.94	1.18	—
POO	62	270	—	1.08	1.43
QUE	61	285	1.02	0.96	1.09
RAB	47	167	1.00	—	—
RIV	76	173	1.27	0.55	0.98
SNG	51	239	1.82	1.14	1.36
VAL	83	344	1.19	—	—

average time separation of 2.5 sec in the calculation of synthetic strong motion seismograms. We compare the observed P wave seismograms with the synthetic ones for selected stations in Fig. 4. The station data are given in Table 3. Agreement between them is considerably good except for the depth phases. Figure 5 shows the comparison between the observed S wave seismograms and the synthetic ones for selected stations. The seismic moment was finally calculated by comparing the amplitude of the synthetics and the observations (Table 3). The average seismic moment is 1.26×10^{26} dyne cm. This value is consistent with that (1.7×10^{26} dyne cm) obtained by the centroid-moment tensor inversion (Dziewonski et al., 1988).

For the source parameters of the 1984 event, we adopted the centroid-moment tensor solution (Table 1) determined by Dziewonski et al. (1984), except the source time function. We assumed a symmetrical trapezoidal pulse with rise time, flat time and fall time of (0.5, 0.5, 0.5) sec for the source time function.

3. Strong Ground Motions

In this section we calculate the synthetic strong motion seismograms for the

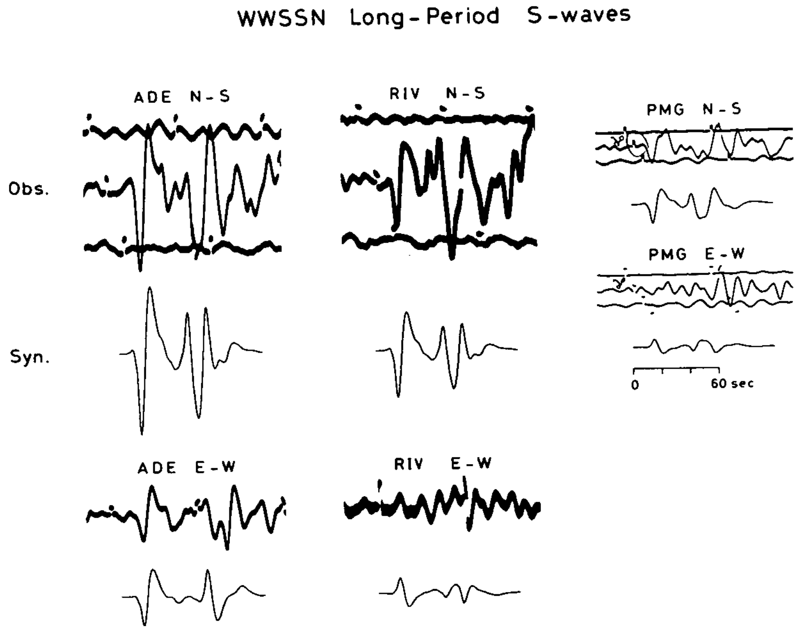

Fig. 5. Long-period S wave observations (upper traces) and matched synthetics (lower traces) for the 1987 event.

source parameters obtained above and compare them with the observed ones at seven stations in Hokkaido (Fig. 1). Surface displacements due to a dislocation source in a layered half-space structure were calculated by the method reported earlier (Sasatani, 1985), and the synthetic seismograms were obtained by convolving them with the impulse response of the seismograph used. The instrumental constants of JMA strong motion seismograph are: To (pendulum period) = 6 sec for the horizontal and To = 5 sec for the vertical component; ε (damping ratio) = 8; and v (magnification) = 1. The assumed structure (Table 2) has no surface layer with an extremely low velocity; thus the synthetic seismograms may be regarded as records on rock site.

Figure 6 shows a comparison between the synthetic seismograms and the raw records for the 1987 event. The observed records have no resolving power for short-period ground motions less than about 1 sec, because of the low paper speed (3 cm/min.). In addition to this, we have no detailed information about the source time function and the crustal structure; therefore, the precise comparison including short-period components is impossible at the present stage. Our comparison was derived by filtering out short-period components involved

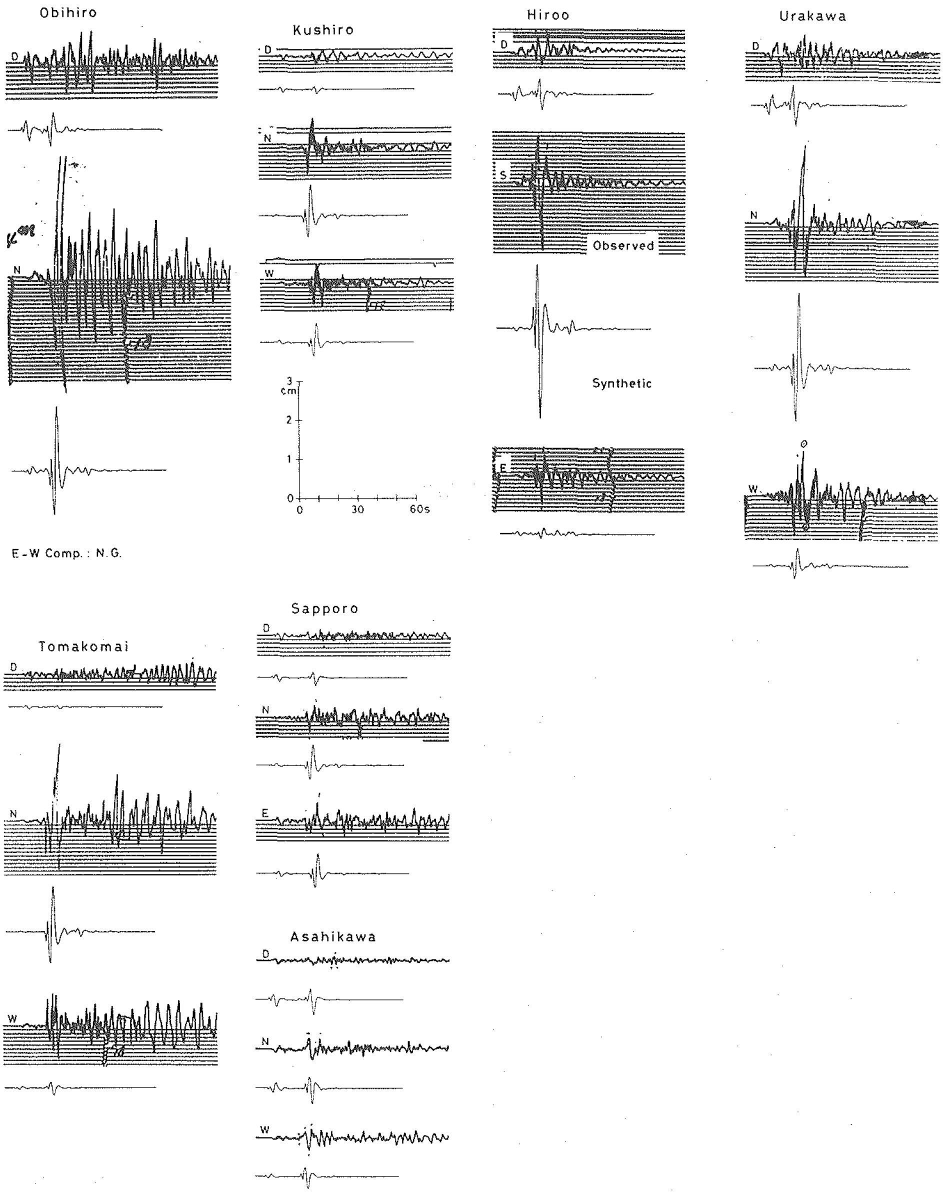

Fig. 6. JMA strong-motion seismograms (upper traces) and synthetic seismograms (lower traces) for the 1987 event.

in the records. The comparison showed that general features of the first 20 sec of the records including both direct P and S waves were explained by the synthetic seismograms. Especially we found the observed double pulse of S wave to be due to the source effect, that is, the doublet.

The synthetic seismograms calculated for the simple source and structure models show that the ground motion rapidly decays after the S arrival with the maximum amplitude, which can be theoretically predicted on the rock site. The records at Hiroo and Urakawa are very similar to this feature. The records at the other stations, however, show distinct later arrivals after S phases, resulting in a long duration of the ground shaking. These later arrivals

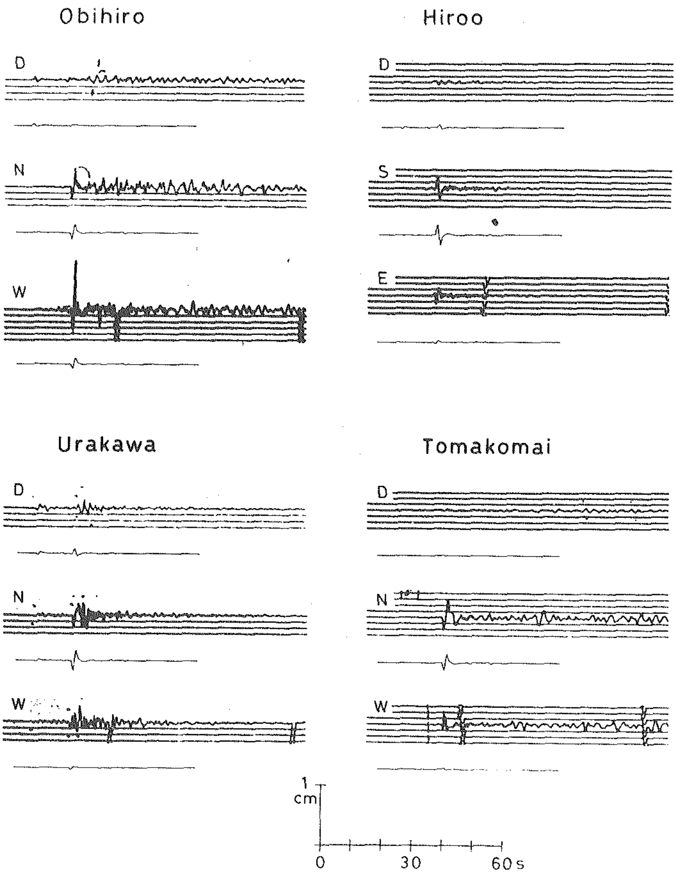

Fig. 7. JMA strong-motion seismograms (upper traces) and synthetic seismograms (lower traces) for the 1984 event.

cannot be explained by the simple synthetic seismograms and the degree of excitation of these arrivals differs from station to station. These facts demonstrate that the later arrivals are locally generated waves near the station and may represent an index of the site effect. For brevity, we hereafter call the later arrivals the long-period coda waves.

We show the comparison between the observed and synthetic seismograms for the 1984 event in Fig. 7. In this case, data from only 4 stations were available because of much smaller event (Table 1). This figure confirms the ground motion characteristics observed for the 1987 event: simple S phase, which can be fairly well explained by the simple synthetics, and strong excitation of the long-period coda waves at Obihiro and Tomakomai.

In conclusion, we found that body waves from intermediate-depth earthquakes can be explained by the synthetic seismograms calculated for the simple source and structure models, and that the excitation of the long-period coda waves represents an index of the site effect.

4. Long-Period Coda Waves

We investigated the nature of long-period coda waves through complex seismic trace analysis (Farnbach, 1985; Tarner et al., 1979). We used one horizontal component record of good quality at each station. After correction of the distortion of the original strong motion records due to the mechanism of the recording system, these records were digitized at a rate of 0.4 sec. We calculated the envelope shape and weighted average frequency for each record in terms of Fourier integrals (e.g. Tarner et al., 1979).

The envelope shapes for the 1987 event are shown in Fig. 8 together with a simple geological map of Hokkaido. We represent the envelope shapes being normalized on the S arrival in amplitude to illustrate how the coda waves are excited from the incident S wave at each site. The envelopes at Hiroo and Urakawa consist of an isolated S pulse and have almost no coda waves. These stations lie on a rock site or at the edge of an alluvial plain. Meanwhile, the envelopes at stations within the alluvial plains demonstrate strong excitation of the coda waves, which have amplitude as large as or larger than half of the S amplitude and duration longer than 2 minutes. The mode of excitation of the coda waves varies from station to station. For example, the coda waves appeared immediately after the S arrival at Obihiro, while they appeared about 30 sec after the S arrival at Tomakomai.

We compare the envelope shapes for the 1987 event with those for the 1984

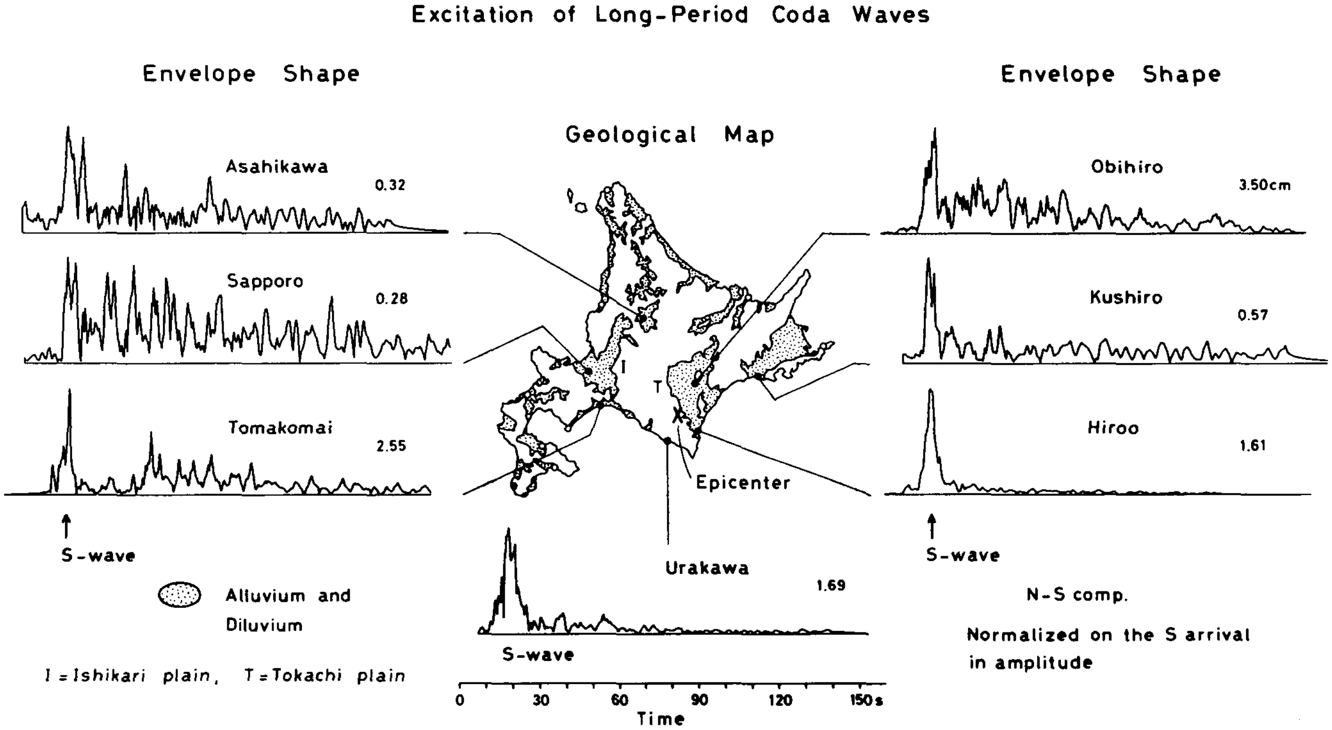

Fig. 8. Envelope shapes for the 1987 event and simplified geological map of Hokkaido. Only one side of the envelope is represented because of the symmetry of the shape. Numbers attached to each envelope indicate the peak amplitudes.

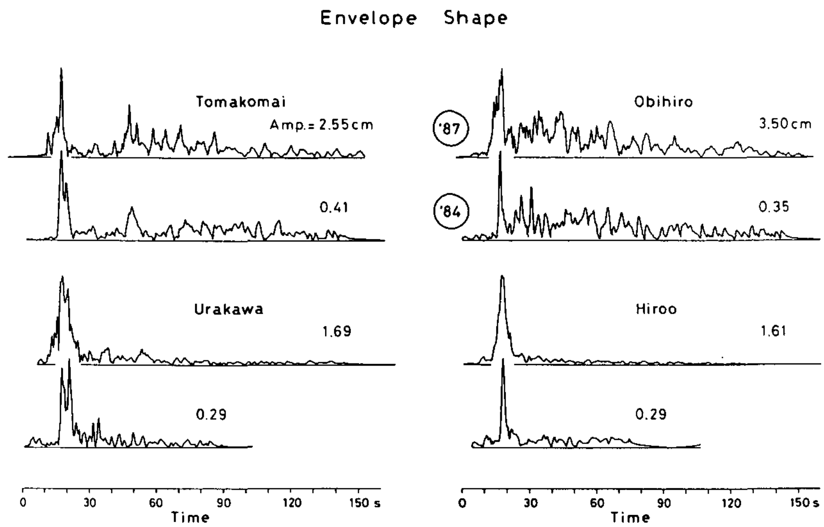

Fig. 9. Envelope shapes for the 1987 and 1984 events at four stations. Only one side of the envelope is represented because of the symmetry of the shape.

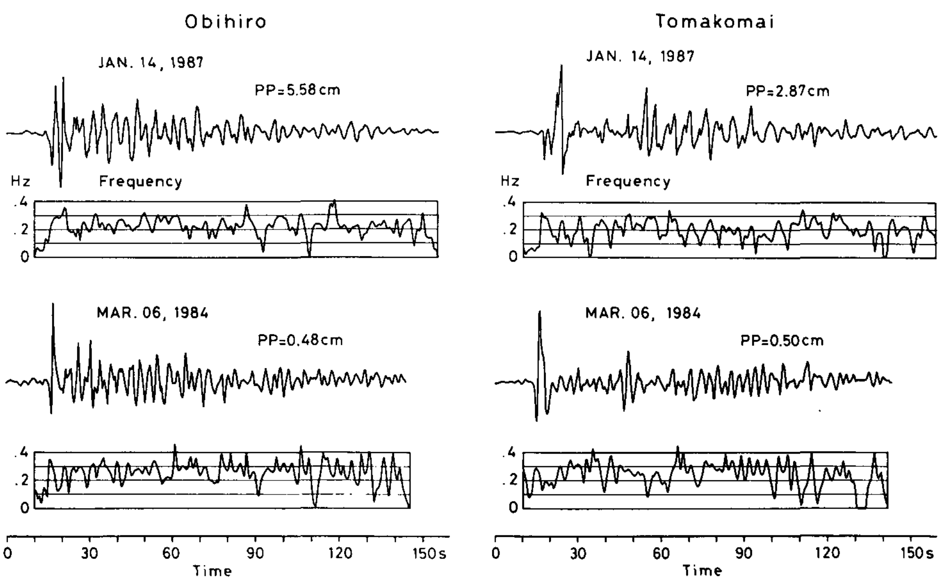

Fig. 10. Weighted average, time-dependent frequency of the long-period coda waves at Obihiro and Tomakomai.

event at 4 stations in Fig. 9. The S wave amplitude of the 1984 event is about 1/7 as large as that of the 1987 event, but two envelopes are very similar at a given station. This fact may indicate that the mode of the excitation of the long-period coda waves is inherent in an individual alluvial plain.

Finally we show the time-dependent frequency of the long-period coda waves at Obihiro and Tomakomai in Fig. 10. The coda waves have frequencies in the range of 0.2 to 0.4 Hz at Obihiro and 0.1 to 0.4 Hz at Tomakomai, but the predominant frequency for the 1984 event is slightly higher than for the 1987 event.

5. Discussion

Strong motion seismograms from near-by intermediate-depth earthquakes show the long-period coda waves at stations within the alluvial plains ; Obihiro in the Tokachi plain, and Sapporo and Tomakomai in the Ishikari plain. In these plains, the basement structures to depths of a few kilometers have been partly obtained by Matsushima and Okada (1989) using long-period microtremors. The structures obtained show the variation of the basement depths, that is, the irregularity of the sediment-basement interface. Recent numerical studies have revealed the strong influence of an irregular interface on seismic motion ; generation of surface waves, and resonance (for example, Aki and Larner, 1970 ; Bard and Bouchon, 1985 ; Horike, 1988 ; Sasatani, 1988). These studies may confirm that the long-period coda waves are surface waves locally generated on the irregular interface, and subsequently trapped inside the plain.

In Fig. 11 we show an example of the response of a sedimentary basin to a vertically incident plane SH wave, which may represent the incident wave field from a near-by deep earthquake. We assumed the simple, two-dimensional basin model for the Tokachi plain as shown in the figure, based on Matsushima and Okada (1989) and obtained the displacements at surface receivers by using the discrete wavenumber method (Sasatani, 1988). We can see the generation of coda waves and their subsequent trapping inside the basin, resulting in a long duration of the ground shaking at the sites within the basin. As pointed out by many authors (e.g., Horike, 1988 ; Sasatani, 1987, 1988), we cannot predict the long duration of ground shaking inside the basin by the Haskell method or the flat-layer approximation. The waveform agreement between the surface displacements in Fig. 11 and the observed seismogram at Obihiro in Fig. 1 is not good but is close enough to demonstrate that the basin model can probably

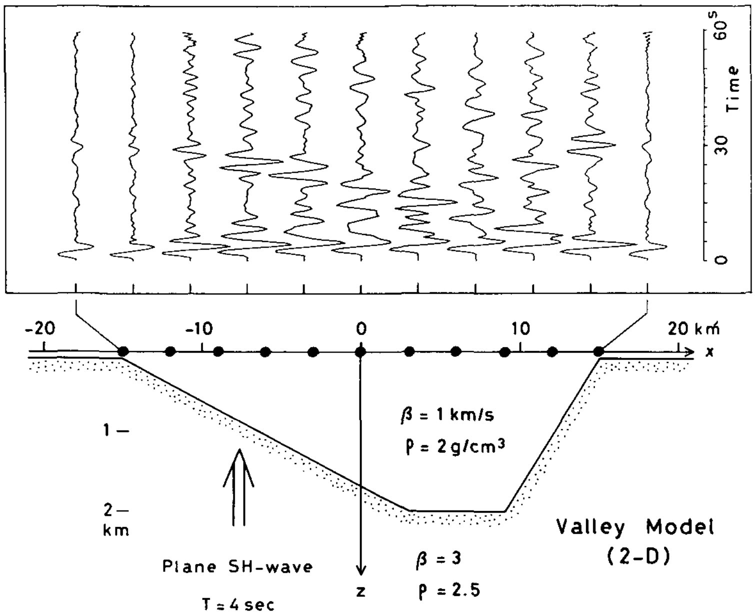

Fig. 11. Response of a sedimentary basin to vertically incident SH wave (one cycle sine wave with a period of 4 sec). The traces represent the displacement at surface receivers (filled circles in the lower figure).

explain the observed features. The waveform contrast between observed seismograms at Obihiro and Hiroo in Fig. 1 can be reproduced by synthetic seismograms within the basin and at the edge.

6. Conclusion

We obtained the following conclusions:

(1) Strong-motion seismograms from near-by, intermediate-depth earthquakes are clearly useful for study of the site effects.

(2) The alluvial plains excite long-period coda waves with the predominant period of several seconds.

(3) The mode of excitation of the coda waves is closely related to the character of the alluvial plain.

Acknowldgments

I thank Dr. K. Sudo of the Building Research Institute for supplying the

WWSSN records and Dr. K. Goto of the Meteorological Research Institute for supplying the JMA strong-motion records. I am grateful to Miss R. Takemori for her assistance in analyzing the strong motion records. The numerical calculations were carried out by HITAC M-682 H at the Hokkaido University Computing Center.

References

Aki, K. and K.L. Larner, 1970. Surface motion of a layered medium having an irregular interface due to incident plane SH waves. J. Geophys. Res., **75**, 933-954.

Bard, P.-Y. and M. Bouchon, 1980. The seismic response of sediment-filled valleys. Part 1. The case of incident SH waves. Bull. Seism. Soc. Am., **70**, 1263-1286.

Beck, J.L. and J.F. Hall, 1986. Factors contributing to the catastrophe in Mexico City during the earthquake of September 19, 1985. Geophys. Res. Letters, **13**, 593-596.

Bouchon, M., 1976. Teleseismic body wave radiation from a seismic source in a layered medium. Geophys. J.R. astr. Soc., **47**, 515-530.

Dziewonski, A.M., J.E. Franzen, and J.H. Woodhouse, 1984. Centroid-moment tensor solutions for January-March 1984. Phys. Earth Planet. Inter., **34**, 209-219.

Dziewonski, A.M., G. Ekstrom, J.H. Woodhouse and G. Zwart, 1988. Centroid-moment tensor solutions for January-March 1987. Phys. Earth Planet. Inter., **50**, 116-126.

Farnbach, J.S., 1975. The complex envelope in seismic signal analysis. Bull. Seism. Soc. Am., **65**, 951-962.

Hauksson, E., T.-L. Teng and T.L. Henyey, 1987. Results from a 1500 m deep, three-level downhole seismometer array : Site respose, low Q values, and fmax. Bull. Seism. Soc. Am., **77**, 1883-1904.

Horike, M., 1988. Analysis and simulation of seismic ground motions observed by an array in a sedimentary basin. J. Phys. Earth, **36**, 135-154.

Jarpe, S.P., C.H. Cramer, B.E. Tucker, and A.F. Shakal, 1988. A comparison of observations of ground response to weak and strong ground motion at Coalinga, California. Bull. Seism. Soc. Am., **78**, 421-435.

Kinoshita, S., T. Mikoshiba and T. Hoshino, 1986. Estimation of the average amplification characteristics of a sedimentary layer for short period S-waves. Zisin, Ser. 2, **39**, 67-80, (in Japanese).

Kudo, K. and M. Sakaue, 1984. Oil-sloshing in the huge tanks at Niigata due to the Nihonkai-Chubu earthquake of 1983, Bull. Earthq. Res. Inst., Univ. Tokyo, **59**, 361-382, (in Japanese).

Matsushima, T. and H. Okada, 1989. Determination of deep geological structures under urban areas using long-period micro-tremors. Submitted to Butsuri-Tansa (Geophys. Explor.).

Okada, H., S. Suzuki, T. Moriya and S. Asano, 1973. Crustal structure in the profile across the southern part of Hokkaido, Japan, as derived from explosion seismic observations. J. Phys. Earth, **21**, 329-354.

Sasatani, T., 1980. Source parameters and rupture mechanism of deep-focus earthquakes. J. Fac.Sci. Hokkaido Univ., Ser. VII (Geophys.), **6**, No. 2, 301-384.

Sasatani, T., 1985. Seismic wave propagation in a stratified half-space. J. Fac. Sci. Hokkaido Univ., Ser.VII (Geophys.), **7**, 401-420.

Sasatani, T. 1987. A study of the strong ground motion from intermediate-depth earthquakes. Geophys. Bull. Hokkaido Univ. **49**, 63-73, (in Japanese).

Sasatani, T., 1988. SH-waves in a laterally varying medium: (1) The seismic response of sediment-filled valleys to plane-incident SH-waves. J. Fac. Sci. Hokkaido Univ., Ser. VII (Geophys.), 8, 191-208.

Suzuki, S., 1978. Lateral variation of the upper mantle structure around Northern Japan and its application to hypocenter determination. J. Fac. Sci. Hokkaido Univ., Ser. VII (Geophys.) **5**, 79-120.

Tarner, M.T., F. Koehler, and R.E. Sheriff, 1979. Complex seismic trace analysis. Geophys., **44**, 1041-1063.