Strong Motion Seismology in Japan

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This paper reviews strong motion seismology in Japan. It describes advances in strong motion seismology in the last decade, the current state of the art, and unresolved issues in the following areas: (1) Observation and data processing of strong ground motion, (2) Local site effects, and (3) Simulation and prediction of strong ground motion. In the first area, it reviews the current state of strong motion observation and data processing: fundamental subjects in strong motion seismology. In the next area, it describes empirical evaluations of site amplification factors, effects of soil nonlinearity, effects of topographical and subsurface irregularity, and application of microtremors. In the last area, it summarizes the results of empirical, semi-empirical, and numerical techniques for the simulation and the prediction of strong ground motions due to actual earthquakes. Finally, future directions of research are suggested, based on the current state of the art in strong motion seismology.

1. Introduction

Observation of strong ground motion in Japan was initiated in 1953 with the installation of a strong motion accelerometer in Tokyo. This accelerometer is called SMAC, because it was designed by the Strong Motion Accelerometer Committee, established in 1951. It produced the first strong motion record from an earthquake of Feb. 14, 1956 (Takahashi, 1956). Because of the importance of recording strong motion in the development of strong motion seismology and earthquake engineering, eighty four SMAC-type accelerometers were installed throughout Japan in the ten years to 1962, and by 1987, the total had reached more than 1,500.

Large earthquakes have occurred frequently in Japan, such as the 1964 Niigata earthquake ($M = 7.5$), the 1968 Hyuga-nada earthquake ($M = 7.5$), the 1968 Tokachi-oki earthquake ($M = 7.9$), the 1973 Nemuro-oki earthquake ($M = 7.4$), the 1978 Miyagi-oki earthquake ($M = 7.4$), the 1983 Nihonkai-chubu earthquake ($M = 7.7$). Many strong-motion records have been obtained from these events by SMAC-type accelerometers.

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The responses of structures and the ground during large earthquakes have been studied from these records.

Another strong motion observation network has been established by the Japan Meteorological Agency (JMA), with more than hundred sites all over Japan. Displacement-type seismometers with a natural period of 5 or 6 s were mainly used. They were installed initially for seismological research. However, with the recent increase in the number of large-scale structures such as high-rise buildings, long-span bridges, and huge oil tanks, the JMA records of long-period strong ground motion, which cannot be sufficiently recorded by the SMAC-type accelerometers, have proven very useful.

In the last decade, strong motion instrument arrays have been newly established to elucidate the processes of seismic wave generation and propagation due to large earthquakes. Digital recording systems are typically used for these arrays, because of their large dynamic range and high resolution. High quality data sets, in addition to those from conventional observations, have provided us with the means to predict strong ground motions in future earthquakes.

Data from destructive events in the last decade; that is, the 1983 Nihonkai-chubu earthquake, the 1985 Michoacan earthquake and the 1989 Loma Prieta earthquake, have provided a new appreciation of the fact that ground conditions have a greater influence on structural damage than previously thought. Many studies have considered the relation between ground conditions and strong ground motions, and empirical and numerical techniques have rapidly advanced to evaluate local site effects.

Recently, two Japanese reviews related to strong motion seismology have been written by Iwata (1991) and Kudo and Higashi (1990). The former reviews studies the characteristics of strong ground motion in special relation to complex source processes of destructive earthquakes, and the latter review emphasized the role of subsurface structures. In the present paper, to review strong motion seismology in Japan more widely, the contents covers the studies during 10 years ending 1991 in accordance with the following three items:

1. Observation and processing of strong ground motion.
2. Local site effects.
3. Simulation and prediction of strong ground motion.

A review of studies on source and propagation processes of seismic waves, which include studies developing numerical techniques, will be left to other authors in this volume, though they are important factors in determining the characteristics of strong ground motion. Let us start our review with recent developments in observation and processing of strong ground motion.

2. Observation and Data Processing of Strong Ground Motion

2.1 Strong motion instruments and processing

The strong motion instrument most widely used in Japan is the SMAC type accelerometer. The distribution of SMAC accelerometers is shown in Fig. 1. The transfer function of the standard SMAC type seismometer is shown in Fig. 2. The SMAC accelerograph is mechanical and seismic motions are recorded on waxed paper or scratch film, while the SMA-1 accelerograph very commonly used all over the world is optical.
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Fig. 1. Distribution of SMAC type accelerometers in Japan (1987, March).

<table>
<thead>
<tr>
<th>District</th>
<th>Number</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hokkaido (HK)</td>
<td>86</td>
</tr>
<tr>
<td>Tohoku (TH)</td>
<td>129</td>
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<tr>
<td>Kanto (KT)</td>
<td>186</td>
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<td>Tokyo (TK)</td>
<td>236</td>
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<tr>
<td>Gunma (GB)</td>
<td>216</td>
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<tr>
<td>Aichi (AI)</td>
<td>63</td>
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<td>Kinki (KK)</td>
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<td>Osaka (OS)</td>
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<tr>
<td>Shikoku (SK)</td>
<td>60</td>
</tr>
<tr>
<td>Kyushu (KS)</td>
<td>64</td>
</tr>
<tr>
<td><strong>Total</strong></td>
<td><strong>1579</strong></td>
</tr>
</tbody>
</table>

Fig. 2. Magnification factors of SMAC-B2 and SMA-1 accelerometers (Mori and Crouse, 1981).

For comparison, the characteristic of the SMA-1 accelerograph is also displayed in the figure. As can be seen, records from the SMAC accelerograph reduce the high frequency components, while the records are affected by a base line shift in the long-period range (e.g., Toki et al., 1987). When long-period ground motion is sufficiently large, SMAC
Accelerographs can be used for estimation of long-period motion. For example, Kudo and Sakaue (1984) estimated long-period motion from a SMAC record in Niigata plain during the Nihonkai-chubu earthquake of 1983 and discussed the relation of the ground motion and sloshing of oil. However, various numerical errors may be generated in processing strong motion records and they should be properly corrected for by detailed analysis.

The accuracy of digitization of SMAC records has been investigated in several ways, because a large number of SMAC records have been accumulated. Kawashima et al. (1982a) discussed human-induced errors during the digitization procedure by using artificially generated data, and concluded that the digitization accuracy was high enough for the period range of 0.1 to 3 s. In processing the records, instrumental correction by a filter whose frequency characteristics are the inverse of that in Fig. 1 is needed. Goto et al. (1978) processed several actual records from the SMAC accelerometers using filters with different high and low cut-off frequencies and recommended a high cut-off frequency of 10 to 12 Hz.

The results of the above statistical analyses using actual records principally depend on the S/N ratio of records. It is therefore difficult to obtain a general rule for accuracy in the digitization procedure from an analysis of specific strong motion records. Accordingly, several investigations have been conducted on this subject by means of shaking table tests. Kawashima et al. (1982b) suggested a reliable frequency range of 0.33 to 12 Hz in the instrumental correction from a shaking test of a SMAC accelerograph. However, Goto et al. (1987) obtained an adequate low cut-off frequency of 0.1 Hz for SMAC records by comparing them to records from a seismometer that is more accurate than the SMAC. Toki et al. (1987) analyzed records produced by the SMAC and more accurate seismometers using shaking table tests and earthquake observations, and proposed a method to determine the low cut-off frequency of a filter assuming that the low-frequency noise due to base line shift has a characteristic with a spectral amplitude increasing in proportion to the inverse of frequency. Ohta et al. (1985) also used a similar procedure to process strong motion records.

As discussed above, SMAC accelerograms are less accurate in the long period range, while the displacement type strong motion instruments installed by JMA can accurately record long-period ground motions. The accuracy of the long-period component of the JMA records has been investigated by several researchers (e.g., Yamada et al., 1987; Katayama et al., 1990). Most of the results from shaking table tests and simultaneous observation of ground motions suggested that the JMA seismometers could record ground motion accurately in the periods from 2 to 20 s.

Although strong motion record from the JMA seismometers provide valuable data on long-period strong motion, they are sometimes saturated at an amplitude of about 3 cm by a stopper in the mechanical part of the seismometer. For example, strong motion records at Niigata during the 1983 Nihonkai-chubu earthquake were clipped because of the large long-period content due to amplification by deep sediments. Noda et al. (1988) proposed a method to recover data from a clipped record by assuming a dynamic model of the seismometer. The performance of this model was investigated using shaking table test data as shown in Fig. 3. The procedure was applied to process a saturated record at Tokyo during the Kanto earthquake of 1923 (Yokota et al., 1989).
Fig. 3. Results of applications of a method for mending saturated seismograms (Noda et al., 1988). (a) Saturated record to be corrected. (b) Seismogram mended by this method. (c) Seismogram corrected instrumentally. (d) Shaking table movement to be compared with trace (c).

To obtain reliable records in the long-period range, Muramatsu (1980) developed a velocity-type strong-motion seismometer which has a flat response for periods up to 50 s and a dynamic amplitude range from 0.01 to 100 cm/s. This type of seismometer has been used in several seismic arrays (e.g., Yamanaka et al., 1989; Zama, 1990)

Recently developed digital accelerographs have also been investigated with respect to accuracy in the long-period range. Kataoka et al. (1990) indicated, based on a shaking table test, that long-period strong motion including the DC content can be well recorded by the digital accelerographs typically used in Japan and the U.S.A. with the proper base line correction.

2.2 Array observation of strong ground motion

Strong motion records from SMAC- and JMA-type seismometers have provided valuable information on earthquake ground motion. However, such records are not sufficient for clear understanding of earthquake source processes, wave propagation,
and local site effects, because most of these instruments are single-isolated. Array observation of earthquake ground motion is required for recent detailed work in strong motion seismology and earthquake engineering. This was also emphasized during the International Workshop on Strong-Motion Earthquake Instrument Arrays in 1978 (Iwan, 1978). In accordance with a resolution adopted by this workshop, in 1980, the Science Council of Japan recommended the installation of dense strong motion arrays. This recommendation became an impetus to the development of new dense strong motion arrays in Japan.

As will be shown later, there have been many studies of local site effects in Japan. Accordingly, most of the strong motion arrays in Japan were designed for understanding
local site effects. In particular, many down-hole arrays have been operated for investigating amplification of seismic waves in soft soils. This is different from the situation in California, where many source-oriented strong motion arrays were installed. The locations and the dimensions of typical strong motion arrays in Japan are shown in Figs. 4 and 5. First, we look at seismic arrays having short interstation spacings of 10 to 100 m.

Several small-scale strong motion arrays were established in Japan (e.g., Katayama et al., 1984; Noda et al., 1982, 1988; Okubo et al., 1984). Most of them are installed on soft soils to determine local amplification factors. For example, a three dimensional array has been installed in Chiba by the Institute of Industrial Science, the University of Tokyo (Katayama et al., 1984). The location of boreholes in which seismometers are installed are shown in Fig. 6. Three component accelerographs having flat responses in the frequency range of 0.1 to 30 Hz are installed in the ground at 44 locations. The deepest borehole seismometer is 40 m deep and is installed in a layer with an S-wave velocity of 420 m/s. Various characteristics of ground motions, such as amplification,
phase velocity, and strain, have been investigated using array records (Katayama et al., 1990; Turker et al., 1990; Lu et al., 1990). One of the interesting findings of those studies is that the coherence of ground motion at two sites just 5 m apart is almost unity for frequencies lower than 5 Hz, while it decreases at higher frequencies. This result suggests the difficulty in predicting high-frequency strong motion in a deterministic way. In addition, a development of a database for the Chiba seismic array is worked out (Katayama et al., 1990).

To investigate the characteristics of ground motion which is not disturbed by low-velocity surface layers, four strong motion arrays have been deployed in the Kanto district since 1979 as a cooperative project among 10 Japanese electric power companies (Omote et al., 1980). Most of the seismometers are buried at 10 to 70 m depth in layers with an S-wave velocity of about 700 m/s.

A similar cooperative project of the electric power companies has been conducted in the Tohoku region, where two vertical arrays have been deployed (Omote et al., 1986) to determine amplification factors over the basement. The two arrays are 20 km apart and both boreholes penetrate a granitic layer with an S-wave velocity of 2.8 km/s at depths of 330 and 950 m.

In a large array, seismometers are usually set in strata of various geological conditions. Several large-scale arrays with station spacings of more than 1 km have been established, as shown in Fig. 4 (e.g., Kitagawa et al., 1988; Shimizu et al., 1988). For example, an array consisted of 11 stations with a few kilometers spacing have been installed in Sendai by the Building Research Institute. At each site, three seismometers have been installed at the surface, intermediate level (20–30 m) and lowest level (50–60 m) in the boreholes reaching a layer with an S-wave velocity of higher than 700 m/s. The area of the array is of special engineering interest, because severe damage was experienced during the 1978 Miyagi-oki earthquake, which had a magnitude of 7.4.

A strong motion instrument array in the near field is necessary to study the source process in detail. In Japan, a large earthquake with a magnitude of 8 is anticipated along the Nankai trough in the Suruga bay area. The Earthquake Research Institute, the University of Tokyo, has been conducting an array observation in the area, as well as around the Izu Peninsula since 1983 (Tanaka et al., 1984). Because their major interest is in investigating source characteristics, most of the accelerometers were installed on firm outcrops of rock with S-wave velocities of approximately 1.5 km/s.

The Earthquake Research Institute set up another strong motion accelerograph array in Ashigara valley (Kudo et al., 1988), to investigate local site effects and wave propagation nature in the sedimentary basin. Fifteen stations were sited in an area approximately 10 km wide and 15 km long. Some seismometers were located in the valley, while the others were located on outcrops of rock surrounding the valley. Since the observational system employed has a high dynamic range of 108 dB, they can observe weak as well as strong ground motions. Several kinds of geophysical studies have been carried out to determine the subsurface structure beneath Ashigara valley (Kudo and Shima, 1988).

In most of the above-mentioned strong motion instrument arrays, accelerometers were set at maximum depths of less than 1,000 m. However, several basins in Japan have sedimentary layers several kilometers thick. To investigate the response of thick
sedimentary layers, the National Institute for Earth Science and Disaster Prevention (NIED) has been performing array observation in the Kanto plain. Two types of arrays were established at four sites, as shown in Fig. 4 (Kinoshita et al., 1982, 1990). At these sites, vertical arrays have been employed using boreholes more than 2 km deep. Accelerometers have been installed at three different levels: the free surface, the intermediate level and the bottom, which is located in the basement with an S-wave velocity of 2.5 km/s. An example of the seismograms observed at the FCH array is displayed in Fig. 7. Up- and down-going waves can be clearly identified in the seismograms. The NIED also conducted a horizontal strong motion array observation in Tokyo lowland to clarify complex wave fields in the long-period range in a sedimentary basin (Kinoshita et al., 1990).

In the southern Kanto plain, there are several other seismic arrays with station spacings of several kilometers (e.g., Yamanaka et al., 1989; Zama, 1990). In this area, dense seismic refraction surveys have been conducted to clarify the structure of the deep sedimentary basin (e.g., Shima, 1980). Using the subsurface structural data and records from the array, the characteristics of long-period ground motion have been investigated (e.g., Yamanaka et al., 1989, 1991).

With recent advances in digital technology, highly intelligent systems such as telemetric monitoring have been employed in most of the newly developed seismic arrays. These systems allow us to obtain ground motion records of high quality without special difficulties. However, the quality of array data greatly depends on the locations of the seismometers. Therefore, one of the crucial factors in accomplishing array observation is to design appropriate array configurations. In addition to these efforts to acquire good data, the geological and geophysical conditions beneath the seismic array should also be determined to enable detailed investigation of strong ground motions.

3. Local Site Effects

3.1 Empirical evaluation of site amplification factors

In the last decade, many numerical techniques for estimating wave propagation in an irregular subsurface structures such as a sedimentary basin have been developed to

![Fig. 7. Examples of records observed at the seismic array in Fuchu (Kinoshita et al., 1982).](image)
explain the local amplification of seismic waves (Kohketsu and Takenaka, 1989). These techniques usually require the details of the subsurface structure over a basement to evaluate the local site effect, though this information is rarely available. Furthermore, the applicability of these methods is usually limited to two dimensional problems in the relatively long period range, because of restrictions in computational time and computer capacity. Therefore, an empirical approach is necessary to evaluate local site effects, especially in the high frequency range.

3.1 Conventional regression analysis

Aki (1988) summarized the state of the art concerning the local site effects on ground motion. According to his report, the characterization of site condition by the conventional broad classifications of soil and rock is inadequate to represent the real physical state of site effects, especially in the high frequency range. The site conditions must be characterized properly to capture the essence of the physical processes influencing site effects.

Site specific amplification factors have been obtained from regression analysis for accelerograms (e.g., Kamiyama and Yanagisawa, 1986; Kinoshita et al., 1986; Takemura et al., 1987) based on formulas of attenuation curves for the observed response spectra, which are usually described by earthquake magnitude and hypocentral or epicentral distance. Details of the attenuation curves will be described in Subsec. 4.1.

While many studies of the empirical evaluation of site amplification factors apply to strong ground motion in the period range shorter than 1 s, some studies apply to long-period strong ground motion. Okada and Kagami (1978) obtained the amplification factors for peak displacement in the period range around 5 s by applying the same kind of regression method to the data from the JMA network. Mamula et al. (1984) also obtained amplification factors in the period range from 3 to 15 s by comparing records observed by the JMA strong motion displacement meter from the 1961 Kita-Mino earthquake ($M = 7.0$) with the synthetic spectra computed utilizing the normal mode solution as shown in Fig. 8. Their results indicate that the amplification factors for the long-period strong ground motion are closely related to the depth to the pre-Tertiary basement. In the meantime, Sasatani et al. (1992) introduced the ratio of the area of the envelope shape for the scattered later phases to that for the primary S-waves using the JMA strong motion records. They indicate that the ratio becomes large on the sedimentary plains.

3.1.2 Separation of source, propagation-path, and site factors

The absolute values of site amplification factors cannot usually be obtained from a conventional regression analysis. If we have data on the stratum which can be assumed to be a homogeneous half-space, this problem may be easily solved (Kinoshita et al., 1986). Such strata are named seismic bedrock (Ohta, 1967). Midorikawa and Kobayashi (1978) indicated that basement with an shear-wave velocity of about 3 km/s, which corresponds to the upper boundary of the Earth's crust, is adequate as the seismic bedrock for engineering purposes.

Recently, some attempts have been done to evaluate site amplification factors by using inversion analysis based on the following point source solution in an infinite homogeneous space:
Fig. 8. Distribution of relative amplification factor of ground motion at different periods in Japan (Mamula et al., 1984). The relative values are determined so that their average over the corresponding stations is about zero.

\[ O_{ij}(f) = S_j(f)G_i(f)X_{ij}^{-1}\exp\left(\frac{-X_{ij}f}{Q_S(f)V_S}\right), \]

where \( O_{ij}(f) \) are the observed horizontal components for the Fourier amplitude spectra of S-waves from the \( j \)-th earthquake at the \( i \)-th station, \( S_j(f) \) is the source amplitude spectrum of the \( j \)-th earthquake, and \( G_i(f) \) is the site amplification factor at the \( i \)-th station. \( Q_S(f) \) and \( V_S \) are the average \( Q_S \) value and the S-wave velocity along the wave propagation path from source to station, respectively.

To obtain a unique solution for site amplification factor \( G_i(f) \), an additional constraint is also required for this approach (Iwata and Irikura, 1988; Kato et al., 1992; Kinoshita and Mikoshiba, 1988). Kato et al. (1992) proposed as a reasonable constraint
for data on outcrops of the granite stratum with a shear wave velocity of 2.2 km/s, that the site amplification factor must be 2. This site amplification factor is equivalent to the value on the homogeneous half space. This constraint condition must be revised in the frequency range higher than 5 Hz, taking into account the effect of surface topography around the station (Takemura et al., 1991).

Takemura et al. (1991) applied the revised constraint condition to the inversion analysis for S-wave portions of accelerograms observed at 19 stations for 22 events along the Pacific coast of the southern Tohoku and Kanto districts in the frequency range of 1 to 10 Hz. Figure 9 shows their results, which are classified into 5 groups by surface geology and shear wave velocity of the surface stratum. An amplification factors as high as 20 are observed in the frequency range of 1 to 5 Hz among sites with the
different geologies. The values of the site amplification factor strongly depend on the shear-wave velocity of the surface stratum. This indicates that the variations in impedance between sedimentary layers and basement are responsible for the differences at frequencies lower than 5 Hz. The amplification factors for most sites on sedimentary strata decrease as the frequency increases from 5 to 10 Hz. The decrease is largest for sites on the alluvium with a high damping factor and low shear wave velocity. The effect of anelasticity under the site is a possible reason for the decrease at high frequencies. Similar results have been obtained from data of coda waves from local earthquakes in California by Phillips and Aki (1986).

These results suggest that empirical approaches based on inversion analysis are effective in evaluating local site effects in the high frequency range without detailed information on the subsurface structure. However, these approaches usually cannot evaluate the difference in site amplifications for the events with various locations around the site due to irregularities of subsurface structure and topography. Furthermore, they cannot evaluate the effect of soil non-linearity for input waves with large amplitudes, especially at alluvial sites.

3.2 Effects of soil nonlinearity

In earthquake engineering, consideration of the nonlinear and hysteretic nature of soil behavior under large-amplitude cyclic loading is considered indispensable for estimation of the ground motion associated with sites containing local sediments, especially alluvial deposits. Although a large body of literature exists on laboratory and analytical studies on the nonlinear behavior of soils, few studies shows observational evidence.

The following describes evidence of nonlinearity including nonlinear soil amplification seen in the strong motion records.

3.2.1 Nonlinearity of the soil and strong motion records

There have been some studies on nonlinearity using strong motion records (e.g., Abdel-Ghaffer and Scott, 1979; Tokimatsu and Midorikawa, 1987), where the strain dependent shear modulus $G$ and damping factor $h$ were discussed. Tokimatsu and Midorikawa (1987) showed evidence of soil nonlinearity from observation records from 4 typical alluvial sites in Japan where many strong earthquake records with different amplitude levels were produced. They showed the variation of spectra due to different intensity levels at Shiogama (Fig. 10), where the impedance ratio of the surface layer to the basement is about 10. The predominant period becomes long and the damping factor estimated from the shape of the spectra becomes large with increasing maximum acceleration. They also summarized the strain dependency of the shear modulus from the results of the four sites as shown in Fig. 11. The tendency of decrease in the shear modulus agrees with that of the laboratory test. Their results indicate that the stress-strain relationship of the soil is linear when the peak acceleration of ground motion is less than about 20 gal and that the shear modulus decreases by half when the peak acceleration of the records is from 150 to 500 gal and the peak velocity is within a range from 10 to 20 cm/s.

Tazoh et al. (1988) also showed evidence of a significant shift of the predominant period during strong ground shaking. Ground motions from the 1983 Kanagawa-
Fig. 10. Variations of spectra of strong motion records with amplitude level (Tokimatsu and Midorikawa, 1987). The solid lines show the Fourier spectra for earthquakes with five different levels of ground motion at the Shiogama site. The broken lines indicate the amplification curves based on the SH wave propagation theory.

Fig. 11. Strain dependence of shear modulus ratio (Tokimatsu and Midorikawa, 1987).

Yamanashi border earthquake of $M=6.0$ were recorded at an epicentral distance of 18 km. The peak acceleration at the ground surface was 435 and 134 gal at the base layer during the main shock. The spectral ratio of the ground surface to the base layer showed a period shift from 0.33 to 0.5 s with increasing amplitude. Katayama et al. (1984) and Nozawa et al. (1988) investigated the same data. In these studies, nonlinear simulation analyses were performed.

Although evidence for the shift of predominant period is reported in these studies, other studies show different results. Darragh and Shakal (1991) showed the Fourier spectral ratios of ground motions observed at the soft soil deposits of Treasure Island and at the rock of Yerba Buena Island in the San Francisco Bay area for earthquakes with the same location and the different magnitudes of 7.0, 4.0, 4.1, 3.5, and 3.3, as shown in Fig. 12. In this figure, a shift in predominant period cannot be seen, although
the amplification becomes large with the decreasing intensity level of the ground motion. The phenomena in California and Japan appear to be contradictory. However, the difference in strain-dependent characteristics may explain this difference, because the bay mud shows less reduction in shear modulus with increase in shear strain compared with that of normal alluvial soil (Hryciw et al., 1991).

### 3.2.2 Nonlinear soil amplification

It is important to estimate the ground motion at the alluvial soil surface, considering the nonlinear effects on amplification factor. The shear wave velocity is usually 600–700 m/s where the nonlinear effect does not need to be considered. In some studies, nonlinear soil amplification characteristics are investigated using the observation records based not only on the site specific design but also on the microzonation purposes. Sugito and Kameda (1990) proposed a simple method for estimating amplification factors of maximum acceleration and velocity at the soil surface to those at the bedrock, based on an empirical approach. The amplification factor, "conversion factor," can be directly compared with the observed results for large earthquakes. They defined the conversion factors $\beta_a$ and $\beta_v$ as the ratios of peak acceleration and of peak velocity respectively, at the soil surface to those at bedrock, and estimated $\beta_a$ and $\beta_v$ for typical soil conditions specified by the following geotechnical parameters $S_n$ and $d_p$. The decrease of the conversion factors depends on these parameters $S_n$ and $d_p$, and acceleration $A$, at the rock surface. The parameter $d_p$ is depth to bedrock, where the shear velocity is 600–700 m/s and the parameter $S_n$ is calculated from the $N$-value profile obtained from the standard penetration test by the following formula

\[ S_n = \frac{N}{d_p} \]

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[Fig. 12. Fourier spectral ratios for Treasure Island and Yerba Buena Island for earthquakes with magnitudes of 7.0, 4.3, 4.1, 3.5, and 3.3 (Darragh and Shakal, 1991).]
where $N(x)$ is the $N$-value at depth $x$ (in m) and $d'_n$ is the depth (in m) of the $N$-value profile. Their results are shown in Fig. 13, where the flat part of the curve corresponds to the linear response region and the decrease from the flat part indicates the nonlinear region.

Sugito et al. (1991) showed the observed amplification factors from the Loma Prieta earthquake where the ratio of the observed peak acceleration at the soil surface to that at the rock outcrop is one or less. Their result indicates that the amplification factor decreases with increase in the peak acceleration at the bedrock surface and that the boundary of amplification and deamplification is about 200 gal at the basement. This result is consistent with that of Sugito and Kameda (1990).

The state of knowledge on amplification and deamplification of ground motion in terms of peak acceleration has been summarized by Seed et al. (1976). According to their results, deamplification of peak accelerations at soft clay sites begins at about 100 gal at rock site and becomes significant at 300 gal. Chin and Aki (1991) applied the weak motion amplification factors derived from coda waves to strong ground motions observed during the Loma Prieta earthquake. They indicated that a systematic overprediction is found for distances less than 50 km and concluded that, depending on the site and the level of strong ground motion, nonlinear effects become significant in the acceleration range 100–300 gal. The lower limit agrees with the acceleration level on bedrock proposed by Seed et al. (1976) as roughly the boundary between amplification and deamplification of seismic waves by the surface layers.

### 3.3 Effects of topographical and subsurface irregularities

Although the effect of irregularities on the ground motion is not a new topic in strong motion seismology, as reported in depth by Sato (1967), the observation record at Pacoima Dam of the 1971 San Fernando earthquake has triggered the study of topographical irregularities and the observation records in Mexico City from the 1985

![Fig. 13. Values of conversion factors, $\beta_a$ and $\beta_v$, for typical soil conditions (Sugito and Kameda, 1990). $\beta_a$ means amplification factor for peak acceleration and $\beta_v$ that for peak velocity. The soil parameters $S_n$ and $d_p$ are defined in the text.](image-url)
Michoacan earthquake have stimulated the study of subsurface irregularities. These studies have raised interesting engineering phenomena, including the recognition of coherent waves, which usually mean that the wave group propagated horizontally due to the irregularity and/or inhomogeneity of the soil. These waves result in the spatial variation of ground motion.

3.3.1 Surface topography and subsurface irregularity

The observation record at Pacoima Dam, which recorded a peak acceleration of 1,148 gal (the first to exceed gravity acceleration), has been investigated by some authors (e.g., Trifunac and Hudson, 1971) from the viewpoint of topographical irregularities. Although this research has stimulated many analytical studies, as summarized by Taga (1982), there are only a few examples where earthquake records have been analyzed to study the effects of topography (e.g., Irikura, 1980; Komaki and Toita, 1980; Jibson, 1987; Celebi, 1988). Komaki and Toita (1980) investigated the characteristics of ground motion at the edge of a cliff due to explosion tests, and microtremor and earthquake observations at the damaged site during the 1978 Izu Oshima-kinkai earthquake. They indicated that the amplitude decreases with increasing distance from the edge of the cliff and that the amplitude perpendicular to a cliff is larger than that along the cliff. The fact that the amplitude of ground motion at the edge of a cliff becomes large is consistent with the evidence of damage during this earthquake. Jibson (1987) showed the variation in amplification over a ridge structure based on the data from the Matsuzaki array in Japan (Okubo et al., 1984). The accelerations normalized to the crest acceleration are plotted in Fig. 14 as a function of elevation. The range of peak acceleration for the five earthquakes is limited from a low of a few gals at station

![Fig. 14. Relative distribution of accelerations along a ridge from the Matsuzaki array in Japan (Jibson, 1987). The solid circles show the mean values of peak accelerations at each observation station normalized to the crest acceleration at station No.1 for five earthquakes. The error bars indicate the standard deviations at each observation point. The solid line shows the smoothed distribution function.](image-url)
No. 5 at the base to a maximum of about 100 gal at the station No. 1 at the crest. The amplification at the crest relative to the base is about 2.5. The amplification factor increases rapidly as the crest of the ridge is approached. The amplification of motions at the crest of a ridge relative to motions at the base is also supported by damage patterns during the 1980 Friuli earthquake in Italy and in the 1985 Chilean earthquake (Finn, 1991). Celebi (1988) showed that a spectral ratio of 10 between the station in the canyon and the one on the ridge is not rare and that the damage distribution is consistent with the spectral ratio.

It should be noted that the effect of a topographical structure on ground motions depends on the size of the irregular structure compared to the wavelength of incident motions. Silva and Darragh (1989) show that the wavelength range is 40 m to 5 km over the period range of engineering interest, in the frequency range of 0.2 to 25 Hz by assuming a shear wave velocity in rock of 1 km/s. Topographical features with characteristic dimensions in this range can significantly affect ground motion, depending on the size of the irregular structure.

From the viewpoint of subsurface irregularity, perhaps more interesting from an engineering point of view is the response of a sediment-filled valley. As shown in Fig. 15, during the 1985 Michoacan earthquake, larger amplitudes and longer durations were recorded in the sediments of Mexico City than in the hilly zone (Celebi et al., 1987). These records suggest the limitation of interpretation of one-dimensional wave propagation theory. Many studies have tried to clarify the cause of the characteristic

Fig. 15. Ground motion characteristics due to the 1985 Michoacan earthquake (Celebi et al., 1987). Schematic section showing relative locations of the epicentral station at Caleta de Campos, Teacalco station (closest to Mexico City), and Mexico City stations, UNAM (hills zone) and SCT (lake zone). The seismograms are EW components of acceleration time histories and are plotted to the same scale.

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Fig. 16. Comparison of waveforms and nonstationary spectra of CDAO in Mexico City due to the 1985 Michoacan earthquake with those of Ogata village in Japan due to the 1983 Nihonkai-chubu earthquake (Motosaka et al., 1988).

It should be noted that the sediments in Mexico City and also in Ogata village are very soft deposits with very high water contents and that the later phase comprising the characteristic long duration cannot be explained without adopting a much lower damping factor than that predicted from the laboratory test, as far as the conventional elastic wave propagation theory is used. There may be a gravity wave, as indicated by Matsuzawa (1925) and recently by Lomnitz (1992).

It is also noted there has been discussion on the later phase, based on whether it is caused by deep structure or very shallow structure. This discussion could suggest that the effect is caused by the synergic effect of both the deep and shallow underground structures. Some other studies have been conducted in Japan on the later phase of ground motion. Toriumi (1975) indicated that the later phase of the ground motion was recognized in the strong ground motion record observed in the Osaka plain and performed a model test to examine the generation of later phases due to the surface wave. The test model comprised rubber sponge grains for the soft sediment and a cast-aluminium slab and concrete for base rock. A tapping shock was applied. He confirmed that the later phase propagating from the side concrete could be recognized. Yamanaka et al. (1989) showed that a significant Love wave is observed in the Kanto basin due to earthquakes occurring around the Izu Peninsula and that the surface wave is amplified in the basin. Many other authors have shown that there are some wave
propagation characteristics in the observation records which cannot be explained by one-dimensional wave propagation theory (e.g., Horike, 1988). Details of simulations of the complex wave forms due to subsurface irregularities are described in Subsec. 4.3.

3.3.2 Spatial variation of ground motion

The effect of topographical and subsurface irregularities cannot be explained without considering the seismic-wave propagation process. The high density array observations described above are useful for this purpose. Such array observations have helped to identify coherent waves in various frequency ranges depending on the array size, and have also yielded information on the propagation direction and the phase velocity of the contained surface waves.

Many authors have investigated the wave propagation characteristics of relatively long period waves. In order to determine the phase velocity, the phase difference method (Muto et al., 1982), the method of sums (Minamishima et al., 1986), the correlation method (Nagahashi, 1982) and so on, have been used. Where many observation points are available, the high resolution method proposed by Capon (1969), called the frequency-wavenumber spectral analysis (F-K spectral analysis), can be used to determine the propagation directions and apparent velocities of waves. Complex polarization analysis (Vidale, 1986) has also been used to determine the principal axis of the motion, which indicates the propagation direction. Many observations show that the propagation direction is not always coincident with the direction to the epicenter (e.g., Minamishima et al., 1986; Horike, 1988; Higashi and Kudo, 1992; Kinoshita et al., 1990). This suggests the effects of 3D irregularities in underground structure on wave propagation.

In the high frequency range, local horizontal array data have been used to evaluate the correlation among records at different points. Coherency, which indicates the degree of correlation, is usually modeled as a mathematical expression. Abrahamson (1985) calculated the coherency of P-waves and that of S-waves using the records obtained at the SMART-1 array. The results showed that the coherency of S-waves and P-waves decreases remarkably in the frequency range higher than 2 and 3 Hz, respectively. Horike (1988) found from the analysis of the temporal array records obtained at seven points about 100 m apart that no correlation between two points 70 m apart is seen in the frequency range higher than 6 Hz, and that when the distance becomes 150 m, there is no correlation at frequencies higher than 3 Hz.

These studies suggest that deterministic discussion of waveform is difficult at this stage for frequencies higher than a specific frequency which is dependent on the site, and that probabilistic or stochastic descriptions may be required at higher frequencies.

3.4 Application of microtremors

Subsection 3.2 discussed the differences and similarities between weak and strong ground motions. Although the effects of nonlinearity of soil is significant during strong shaking, analysis of weak motion data facilitate our understanding of local site effects. Many researchers in Japan have been trying, since the early stage of engineering seismology, to determine local site effects with very weak motions called microtremors. Microtremors are ambient ground motion that sometimes called microseisms, seismic noise, Earth noise and so on. In particular, the term “microtremors” is used when this...
kind of ambient motion is used for earthquake engineering purposes. Comprehensive reviews of application of microtremors can be found in the proceeding of the 17th Symposium on Ground Vibration sponsored by the Architectural Institute of Japan in 1989.

We can classify microtremors into two types according to the period range. One comprises short-period microtremors with periods less than 1 s, and is related to shallow subsurface structures several tens of meters thick. The other is long-period microtremors with periods longer than 1 s, which relate to deeper soil structure. In addition to the differences in period, Seo et al. (1990) noted that other different features of the two types of microtremors can be used to distinguish them. Short-period microtremors show daily variation of amplitudes, as initially pointed out by Kanai and Tanaka (1961), while amplitudes of long-period microtremors vary relatively slowly with time. These features of the two types of microtremors have been identified in several regions. For example, Fig. 17 clearly shows an example of different amplitude variations with time of short- and long-period microtremors in Los Angeles (Yamanaka et al., 1993). Short-period microtremors of 0.3 s have reduced amplitudes at night-time as well as on Saturdays and Sundays, when human activity is low. However, the above definitions of the two types of microtremors are not strict and some researchers use other definitions.

3.4.1 Short-period microtremors

We start with studies on short-period microtremors. Numerous investigations have been conducted to determine the nature of short-period microtremors. One of the possible sources of short-period microtremors can be human activity, such as traffic and industrial noises (e.g., Aki, 1957; Kanai and Tanaka, 1961). However, the wave type of microtremors is not still clear. Kanai et al. (1965) explained the characteristics of short-period microtremors observed at the surface and at a depth of 20 m by multiple reflections of S-waves in a sedimentary layer. However, Aki (1957) concluded from a correlation analysis of microtremors that short-period microtremors consist of surface waves. Allam and Shima (1967) pointed out that amplification due to both S- and

![Figure 17](image)

Fig. 17. Variation of amplitudes of microtremors at a period of 0.3 s (upper) and 6.5 s (lower) observed in the Los Angeles (Yamanaka et al., 1993). The amplitude at 0.3 s shows the daily variation in which the amplitude is larger in the day time and smaller at night. It is noted that the amplitudes are smaller on Saturday and Sunday, because of low human activities.
surface waves can elucidate the nature of short-period microtremors, when the subsurface structure consists of a surface layer over the half-space. In spite of the complexity of short-period microtremors, applications of short-period microtremors to evaluate the effects of local geology have been attempted in the earthquake engineering community.

Engineering application of microtremors was initiated by Kanai and his group. Kanai and Tanaka (1961) observed short-period microtremors at many sites. From these observations, they found good agreement between the period distribution curves of microtremors and those of earthquake ground motions. Thus, they proposed a procedure for classifying ground conditions using short-period microtremors. Since microtremors consist of seismic waves from many sources in various directions, their features can be addressed more or less as random processes. Therefore, Kanai's work is based on the assumption that SH-waves with the frequency characteristics of white noise are incident vertically at the base of soil, and that the observed spectral shape directly provides us with the transfer function of the soils. Although this assumption has still not been confirmed, this kind of procedure has been popularly applied in seismic microzonation work, because of the easy acquisition of short-period microtremor data (e.g., Kobayashi et al., 1986; Lermo et al., 1988; Osawa et al., 1988). Recent applications of the use of microtremors to seismic zonation can be seen in the review by Finn (1991).

To reduce the uncertainties of incident waves of microtremors, spectral ratios are sometimes used. Seo et al. (1989) attempted to estimate strong ground motion in sediments by multiplying strong motion spectra on firm rock with the spectral ratio of microtremors between sediment and rock sites. Similar idea concerning the spectral ratio of short-period microtremors between sediments and rock is often used to deduce the transfer function of the sediments (e.g., Field et al., 1990). Although the resonance peak period can be estimated from spectral ratios, the spectral ratio of microtremors sometimes reaches ten times the calculated amplification factors (Field et al., 1990). This suggests difficulties in estimating a transfer function of soils from only microtremor data. For more reliable application of the spectral ratio method for short-period microtremors, several questions, such as stability, assumption of the same input wave and wave type, should be clarified.

A procedure for removing source effects was also proposed by Nakamura (1989). He suggested that source effects could be removed from microtremor data by taking the spectral ratio of the horizontal record to the vertical record at a single site. He assumed that only horizontal microtremors are influenced by soil and that source spectral characteristics are maintained in vertical microtremors as well as in horizontal microtremors. However, the theoretical explanation is still not clear.

3.4.2 Long-period microtremors

The origins of microseisms or long-period microtremors were intensively discussed in the early stage of seismology. There is less doubt that microseisms are caused by meteorological effects, like oceanic disturbances (e.g., Santo, 1959; Longuet-Higgins, 1950). However, less attention was payed to the relation between the characteristics of microseisms and the subsurface structure.

The key to promoting research on long-period microtremors in earthquake engineering was finding a spectral peak at 2.5 s in the strong motion record at Hachinohe during the 1968 Tokachi-oki earthquake. Since then, investigations on long-period
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Microtremors at periods of more than 1 s have been conducted in Japan, because of the engineering requirement to estimate long-period strong ground motion for seismic design of large-scale man-made structures.

Naruse et al. (1976) carried out long-period microtremor measurements in the Hachinohe area and found two distinct spectral peaks in the long-period range in sites on sediments, as can be seen in Fig. 18. They interpreted this as implying that the common peaks at all the sites are due to the input wave to the sediments and another spectral peak, which differs from site to site, indicates the predominant period of soils beneath the sites. Naruse et al. (1976) estimated the thickness of deep soils by using the distribution of the predominant period. Similar results were observed from microtremor measurements in several regions in Japan (Taga, 1983; Kagami et al., 1976). This procedure for determining the effects of deep soils is basically the same as those for short-period microtremors. However, this approach cannot always work effectively. Kagami et al. (1982) could not find any systematic changes of predominant periods in microtremor spectra observed in the Niigata and Los Angeles basins. Instead, they found systematic variations in spectral amplitudes. They then extracted the amplification characteristics of the deep sedimentary layers by taking the ratios of spectra observed on sediments to that on basement, with the assumption that long-period microtremors consist of surface waves which are generated by the beating of oceanic waves on the nearby coast. As suggested by Aki (1988), this assumption is more reliable in the long-period range than in the short-period range. A similar procedure of spectral ratios was applied to long-period microtremor data in the Kyoto basin (Akamatsu, 1984), the San Fernando valley (Kagami et al., 1986) and the San Francisco Bay area.

![Fig. 18. Spectra of long-period microtremors observed at Hachinohe (left) and estimated spectral peak period and subsurface structure (Naruse et al., 1976). The spectral peak at 4 s is due to the input wave and the spectral peak which differs site by site indicates the predominant period of deep soils.](image-url)
(Dravinski et al., 1991; Seo et al., 1991; Akamatsu et al., 1991). According to Seo et al. (1990), the absence of a predominant period in long-period microtremor spectra reflecting the effects of deep soils is due to the absence of interfaces with high impedance ratios. The spectral ratio of vertical to horizontal records of microtremors at a site was also investigated by Shiono et al. (1979) and Kobayashi (1980). They proposed a method of deducing the subsurface structure by using spectral ratios, based on the assumption that long-period microtremors mainly consist of Rayleigh wave.

In the above studies, it was assumed that the subsurface structure consists of flat layers. However, it is usually not easy to verify this assumption for a sedimentary basin. Irikura and Kawanaka (1980) indicated that spatial variation of spectral amplitudes of microtremors is influenced by horizontal irregularities and can be used for mapping a fault.

The other approach to long-period microtremors is based on the use of array observation. Horike (1985) and Matsushima and Okada (1990) applied F-K spectrum analysis to long-period microtremors recorded by an array of vertical seismometers and estimated S-wave velocity profiles beneath the array by inversion of the phase velocity dispersion curve for the Rayleigh wave. The difficulty in determining the S-wave velocity in deep soils can be effectively resolved by such array measurements of long-period microtremors, particularly in urbanized area where artificial seismic sources, such as explosions, cannot easily be used.

In spite of efforts over many years, basic problems such as stability, wave type, and so on, still remain open questions. A better application of microtremors will be achieved when these problems are resolved. For this purpose, high-quality, dense, three-dimensional array observations of microtremors is necessary.

4. Simulation and Prediction of Strong Ground Motion

4.1 Empirical approach

Most empirical methods for predicting strong ground motion amount to evaluation of attenuation curves for the maximum amplitude and spectrum of seismic waves. Attenuation curves were originally used for determining the magnitude $M$ of an earthquake by correcting the observed maximum amplitudes for the effects of the distance from source to station. The variables of the attenuation curve for predicting the amplitudes of seismic waves are, therefore, usually magnitude and distance.

Kanai (1958) and Kanai et al. (1966) proposed an attenuation curve for peak velocity of seismic waves at the basement, and revised it for peak acceleration at the ground surface utilizing the multiple reflection theory, based on the assumption that the velocity spectrum of seismic waves is flat for the frequency at the basement. Midorikawa and Kobayashi (1978) evaluated the attenuation curve for the response spectrum at a basement with an shear wave velocity about 3 km/s, based on the assumption adopted by Kanai (1958).

As a large number of strong motion records have accumulated, a lot of attenuation curves have been derived from regression analysis. Tanaka and Fukushima (1987) summarized the results in Japan, while Cambell (1985) summarized mainly those in U.S.A. Most of the attenuation curves in Japan are represented by conventional
formulas as follows:

\[ \log A = aM - b \log X + c, \]
\[ \log A = aM - b \log (X + 30) + c, \]

where \( A \) is the peak acceleration, velocity, displacement, or response spectrum at a specific frequency, and \( a, b, \) and \( c \) are regression coefficients. \( X \) and \( D \) indicate hypocentral distance and epicentral distance, respectively. Dummy variables are often incorporated into the above equations to obtain amplification factors due to local site effects, as described in Subsec. 3.1. Kamiyama and Matsukawa (1990) adopted dummy variables not only to evaluate local site effects but also to estimate non-linear dependencies of spectral amplitude on earthquake magnitude and hypocentral distance. Fukushima and Tanaka (1990) adopted a two-step stratified regression analysis to avoid interaction between the coefficients \( a \) and \( b \). In the first step, \( aM + c \) in Eq. (3) are replaced by dummy variables for individual events to determine the distance coefficient \( b \).

Kinoshita et al. (1986) and Takemura et al. (1983) studied the regional variation of the distance coefficient \( b \) in Eq. (3) for the Tohoku and Kanto districts in Japan. According to their results, the value of \( b \) in the western part from the volcanic front is systematically larger than that in the eastern part. Takemura et al. (1983) suggested that the high attenuation of peak accelerations in the western part is due to the low-\( Q \) zone of the upper mantle under the volcanos.

Two difficult problems exist in evaluating attenuation curves. One is an ambiguity in the physical meanings of the formulation (e.g., Takemura et al., 1987), and the other is an ambiguity in the interpretation of distance when the station is located within a source dimension from the earthquake fault (e.g., Shakal and Bernreuter, 1980).

The first problem is a big obstacle in predicting the strength of ground motions for hypothetical earthquakes by extrapolation using the attenuation curve. To solve this problem, Takemura et al. (1987) proposed the following formula based on the point source solution of S-waves:

\[ \log A = aM - (\log X + bX) + c, \]

and the relation between the distance coefficient \( b \) and \( Q \)-value along the wave propagation path is as follows:

\[ b = \frac{\pi}{QsVsT \ln 10}, \]

where \( T \) is the period, and \( Qs \) and \( Vs \) are \( Q \)-value of S-waves and S-wave velocity, respectively. Ishida (1986), Kudo (1982), Ohta and Kagami (1976) also derived the same kind of formulas with coefficients determined semi-empirically.

Ikeura et al. (1991) determined the coefficients \( a, b, \) and \( c \) for Eq. (5) from response spectra of accelerograms observed at 19 stations for 22 events along the Pacific coast of the southern Tohoku and Kanto districts by regression analysis. \( Qs \)-values are estimated from the distance coefficient \( b \) using Eq. (6). These values are consistent with the results estimated from Fourier amplitude spectra of S-wave portions of the same data set by the inversion method for the formula of the point source solution (Takemura et al., 1987).
et al., 1991). Sato (1988) interpreted the linear relation between the logarithm of maximum amplitude $A$ and the hypocentral distance $X$, as expressed by Eq. (3), using theoretical results for wave propagation in a fractal structure. According to his result, attenuation of seismic waves due to scattering depends on the fractal dimension in the inhomogeneous media.

The second problem is strongly related to the applicability of the attenuation curve in the near source region. Shakal and Bernreuter (1980) indicated that the distance dependence of ground motion derived from close-in data is highly sensitive to the ambiguous definition of distance in the near source region and that the choice of the method for measuring distance can have a dramatic impact on the resulting attenuation curves. Fukushima et al. (1991) summarized attenuation formulas in Japan and U.S.A. for the near source region. Most of them, however, are inadequate in the choices of distance and definition of the distance, because they were decided arbitrarily. To solve this problem, Takemura et al. (1991) and Ohno et al. (1993) proposed a new distance measure, Equivalent Hypocentral Distance (EHD), on the assumption of randomness of seismic wave phases in the frequency range higher than the corner frequency of the source spectrum of the target large event. EHD is the distance from the virtual point source, which radiates the same seismic energy to the site as an actual finite-sized fault. If the fault plane is divided into $n$ small segments, EHD $X_{eq}$ is expressed as

\[ X_{eq} = \text{...} \]

Fig. 19. Contours of EHD for the 1979 Imperial valley earthquake fault utilizing the distribution of displacement on the fault plane estimated by Harzell and Helmberger (1982) (Takemura et al., 1991).
\[
X_{eq}^{-2}(f) = \frac{\sum_{i=1}^{n} M_{o_i}(f)X_i^{-2}}{\sum_{i=1}^{n} M_{o_i}(f)}, \tag{7}
\]

where \(X_i\) is the distance from the \(i\)-th small segment on the fault plane to the site, and \(M_{o_i}(f)\) is the seismic moment density from the \(i\)-th small segment. The seismic moment of each small segment may be substituted for the seismic moment density \(M_{o_i}(f)\), in Eq. (7), under the assumption that the source time function is the same form on time for all the small segments on the fault plane (Ohno et al., 1993). EHD includes the effects of fault size, fault geometry, distribution of displacement on the fault plane, and location of the site. Figure 19 shows the contour of EHD for the 1979 Imperial Valley earthquake fault, utilizing the distribution of displacement estimated by Harzell and Helmberger (1982). Figure 20 shows the Fourier amplitude spectra of accelerograms averaged over the frequency range from 1 to 10 Hz, which are plotted along the shortest distance \(X_{sh}\) from the fault plane and to EHD \(X_{eq}\). It is found that the Fourier amplitude spectra are inversely proportional to \(X_{eq}\) like the point source solution. This suggests the possibility that the attenuation relation determined in the far source region based on the point source solution can also be used in the near source region, if EHD is substituted for hypocentral distance in the attenuation relation.

Fig. 20. Relations among EHD \(X_{eq}\), shortest distance \(X_{sh}\) from the fault, and Fourier amplitude spectrum averaged over the frequencies from 1 to 10 Hz for the 1979 Imperial Valley earthquake (Takemura et al., 1991).
Studies of attenuation curves have been recently performed not only due to engineering demand but also as seismological research.

4.2 Semi-empirical approach

Harzell (1978) proposed a method of synthesizing the strong ground motions occurring during a large earthquake, utilizing observed seismograms from small events as empirical Green's functions. This approach is called the semi-empirical method and is useful for avoiding difficulties in evaluating the effects of heterogeneities in the earth's structure, because the observed seismograms from small events include the complex effects of the heterogeneous structure.

Irikura (1983) adopted a similarity law for different sized earthquakes utilizing Haskell's kinematic source model to improve Harzell's method. The formula for his method is as follows:

\[ S_L(t) = \sum_{i=1}^{n} \sum_{j=1}^{n} \sum_{k=1}^{n} \frac{r_{ij}}{r_{ij}} S_E(t-t_{ij}-\frac{k-1}{n} T_D), \]  

where \( S_L(t) \) and \( S_E(t) \) are the synthesized record of the target large event and the observed record of a subevent, respectively, \( r_{ij} \) is the hypocentral distance of the subevent, \( r_{ij} \) is the distance from the \( ij \)-th subfault on the fault plane of the target event to the site, \( t_{ij} \) is the delay time due to the rupture propagation and the travel time for the \( ij \)-th subfault, and \( T_D \) is the rise time of the displacement of the target event. The parameter \( n \) is a scaling parameter determined from the cube root of the seismic moment ratio of the target event to the subevent, assuming the similarity law.

Irikura (1983) applied Eq. (8) to a simulation of the near field velocity records for the 1980 Izu-toho-oki earthquake which had a magnitude of 6.7, using records from the foreshocks and aftershocks. The results of the simulation can explain the observed records for periods longer than 1 s. Yamada et al. (1988) also succeeded simulating long-period strong ground motion from the 1983 Nihonkai-chubu event using a similar method to that of Irikura (1983). However, Irikura's approach leads to an underestimate of the amplitudes for periods shorter than 1 s for the 1980 Izu-toho-oki event. This is because the spectrum of the synthesized record has a high-frequency fall-off due to the assumption of uniform displacement and rupture propagation over the fault plane.

Many revisions of the method by Irikura (1983) have been proposed for evaluating short-period strong ground motion. Almost all of the methods consider the effects of heterogeneity of faulting to increase the excitation of short-period components of seismic waves from the source. Dan et al. (1989) and Irikura (1986) presented revised formulas for deterministically evaluating synthetic motions having spectral contents predicted from the \( \omega^{-2} \) model. Iwata and Irikura (1990) also revised the formula of Irikura (1986) for heterogeneous faulting such as asperity model.

Other authors have adopted a stochastic approach to revising Eq. (8) by introducing some randomness for summing subevents. This randomness not only represents the degree of random heterogeneity of faulting but also prevents an artificial periodicity in the simulated motion (Joyner and Boore, 1988). Kanamori (1979), Muramatsu and Ohnuma (1988), and Yoshikawa et al. (1985) introduced random parameters to
superimpose the records of subevents concerning the slip function of the target event, while Iida and Hakuno (1983), Imagawa et al. (1984), Hadley and Helmberger (1980), and Houston and Kanamori (1986) introduced to the process of rupture propagation. Hadley and Helmberger (1980) and Houston and Kanamori (1986) also adopted this randomness for the slip function of the R-model by Kanamori (1979).

In the meantime, Takemura and Ikeura (1988) proposed a hybrid of stochastic and deterministic approaches. According to their method, the displacement on the fault plane of the target earthquake is divided into two parts. One is the average displacement over the fault plane and the other is the deviation from the average. The seismic radiation $S_L(t)$ due to the average displacement is evaluated from Eq. (8) by Irikura (1983), and the seismic radiation $S_S(t)$ due to the deviation is evaluated from the following formula:

$$S_S(t) = \sum_{i=1}^{n} \sum_{j=1}^{n} \frac{r_i^e}{r_{ij}} \kappa_{ij} S_B(t - t_{ij}), \quad (9)$$

where $\kappa_{ij} = x_{ij} - \mu$. $x_{ij}$ is chosen from a Gaussian distribution with an average of zero and a standard deviation of $S_D$, which shows the degree of heterogeneity of displacement on the fault plane. $\mu$ is an average of $x_{ij}$s. The final result of the synthesized strong ground motion is expressed as the sum of $S_L(t)$ and $S_S(t)$. A characteristic of this method is the stochastic presentation of the heterogeneous distribution of displacement by a single parameter, $S_D$.

The formula $S_L(t)$ will coherently sum the low-frequency end of the spectrum to $n^3$ times that of the subevent, which corresponds to the level of seismic moment, and the formula for $S_S(t)$ will sum incoherently the high frequency end of the spectrum to

![Fig. 21. Observed and synthesized accelerograms and their response spectra with a damping factor of 5% (Reproduced from Takemura and Ikeura, 1990). The observed values are from the 1983 Nihonkai-chubu earthquake and its aftershocks, which are used as the subevents in the semi-empirical synthesis. The synthesized values are $S_L$, $S_S$, and their sum, $S_L + S_S$ waves. The $S_L + S_S$ wave is in good agreement with the observed record from the mainshock.](image-url)
nSD times that of the subevent (Ikeura and Takemura, 1990). Therefore, the spectral ratio of the target event to the subevent derived from the formula of $S_L(t) + S_S(t)$ satisfies the scaling relation of the $\sigma^{-2}$ model (Aki, 1967; Boore, 1983) when $SD$ equals 1.0, and that of the bump models (Aki, 1972; Koyama et al., 1982; Gusev, 1983) when $SD$ is larger than 1.0.

Figure 21 shows a simulation of the accelerogram from the 1983 Nihonkai-chubu event of $M=7.7$, and its response spectrum with a damping factor of 5% (Takemura and Ikeura, 1990). It is found that a spectrum of $S_S(t)$ is larger than that of $S_L(t)$ for the periods shorter than 5 s, while $S_L(t)$ is larger than $S_S(t)$ for the periods longer than 5 s. These results indicate that the heterogeneity of the faulting influences the strong ground motions for periods shorter than 5 s.

The spectral ratios of target large events to subevents are derived analytically by Ikeura and Takemura (1990, 1991) under the assumption of bi-directional rupture propagation in the far field where the geometrical spreading factor $1/r_{ij}$ from each subfault is approximately equal to that from the subevent. Some typical formulas for superimposing the seismograms from the subevent are selected as shown in Table 1. Formula A is Eq. (8) by Irikura (1983); B the formula for the $\sigma^{-2}$ model by Irikura (1986); C and D the formulas by Yoshikawa et al. (1985), and by Kanamori (1979) and Muramatsu and Ohnuma (1988), which include randomness in the slip function; E is the formula for incoherent rupture propagation based on the method by Hadley and Helmberger (1980); and F is given by the formula by Takemura and Ikeura (1988).

Figure 22(a) shows the spectral ratios calculated from each formula where the synthesized motion of the target event of $M_o=1.6 \times 10^{28}$ dyn cm is made by superimposing seismograms of the subevent of $M_{oe}=3.2 \times 10^{25}$ dyn cm. A scaling parameter $n=8$ is obtained from the cube root of the seismic moment ratio $M_o/M_{oe}$. The high frequency ends of the ratios differ greatly among the formulas, while the low frequency ends of the ratios for all the formulas are identical with the seismic moment ratio of $n^2$. The spectral ratios obtained from various scaling models of source spectra,

Table 1. Typical formulas for superimposing seismograms from subevents (Reproduced from Ikeura and Takemura, 1991).

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<td>A : $S_{SV}(t) = \frac{1}{n} \sum_{i=1}^{n} \frac{r^2}{r_{ij}} \sum_{k=1}^{n} r_{ik} (t-t_{ik})(k-1) \frac{T_D}{n}$</td>
<td></td>
</tr>
<tr>
<td>B : $S_{SV}(t) = \frac{1}{n} \sum_{i=1}^{n} \frac{r^2}{r_{ij}} S_S(t-i_{ij}) + \frac{1}{n} \sum_{i=1}^{n} \frac{r^2}{r_{ij}} \sum_{k=1}^{n} r_{ik} (t-t_{ik})(k-1) \frac{T_D}{n}$</td>
<td></td>
</tr>
<tr>
<td>C : $S_{SV}(t) = \frac{1}{n} \sum_{i=1}^{n} \frac{r^2}{r_{ij}} S_S(t-i_{ij})(k-1) \frac{T_D}{n} + \Delta \tau_{ij}$</td>
<td></td>
</tr>
<tr>
<td>D : $S_{SV}(t) = \frac{1}{n} \sum_{i=1}^{n} \frac{r^2}{r_{ij}} S_S(t-i_{ij})(k-1) \frac{T_D}{n}$</td>
<td></td>
</tr>
<tr>
<td>E : $S_{SV}(t) = \frac{1}{n} \sum_{i=1}^{n} \frac{r^2}{r_{ij}} \sum_{k=1}^{n} r_{ik} S_S(t-i_{ij})(k-1) \frac{T_D}{n}$</td>
<td></td>
</tr>
<tr>
<td>F : $S_{SV}(t) = \frac{1}{n} \sum_{i=1}^{n} \frac{r^2}{r_{ij}} \sum_{k=1}^{n} r_{ik} S_S(t-i_{ij})(k-1) \frac{T_D}{n} + \frac{1}{n} \sum_{i=1}^{n} \sum_{j=1}^{n} \Delta \tau_{ij} S_S(t-i_{ij})$</td>
<td></td>
</tr>
</tbody>
</table>

The bold letters indicate random variables. The formulas are described in detail in the text.
Fig. 22. Comparison of spectral ratios (Reproduced from Ikeura and Takemura, 1991). (a) Spectral ratios between the target event and the subevent for some typical formulas in Table 1. The scaling parameter \( n \) is obtained from the cube root of the seismic moment ratio between the target event \( (M_0 = 1.6 \times 10^{28} \text{ dyn \cdot cm}) \) and the subevent \( (M_{oe} = 3.2 \times 10^{25} \text{ dyn \cdot cm}) \). (b) Spectral ratios from various scaling models of source spectra. The condition of seismic moments is the same as that in (a).

which are usually defined in the far field, are shown in Fig. 22(b) for comparison with the results shown in Fig. 22(a). The high frequency end of A has a value of 1.0, meeting the condition for the \( \omega \)-cube scaling law by Geller (1976). The formulas whose high frequency ends are \( n \), meeting the condition for the \( \omega \)-square scaling law by Aki (1967) and Boore (1983), are E and F for \( S_b = 1.0 \). The formulas C and D, which incorporate the randomness for the slip function, shows the value of \( n^{3/2} \), which is larger than the value from the \( \omega \)-squared scaling law. The value for the formula B is smaller than \( n \).

On the other hand, another study suggests that formula B gives a different result in the near field. Kamae et al. (1991) attempted to evaluate strong ground motion due to a large event of \( M=7 \) using formula B from small subevents whose source spectrum satisfies the \( \omega^{-2} \) model. Their result indicates that the source spectrum of the calculated motion for the large event also satisfies the \( \omega^{-2} \) model. It is concluded from their result that the high frequency end of the formula B is \( n \), meeting the condition for the \( \omega \)-square scaling law (Aki and Irikura, 1991) in the near field, where the strong ground motion is evaluated for the engineering purposes.

According to the original concept of the subevent as a Green's function, the corner frequency of the subevent should be higher than any frequency of interest in the simulation. In that case, the spectrum of the simulated event will depend only on how the seismograms of the subevents are superimposed. However, it may not be possible to use subevents so small that their corner frequencies are higher than any frequency.
of engineering interest. In the general case of predicting strong ground motion, the predicted result is strongly dependent on the source spectrum of the selected subevent (e.g., Joyner and Boore, 1988). Dan et al. (1990) examined the stability of synthesized ground motions for the 1980 Izu-toho-oki event, utilizing 17 foreshocks and aftershocks as subevents, and indicated that the variation of the ratios of the synthesized PGA, PGV, and SI to the observed ones are 40 to 80%. We have to pay attention to the source process of the selected subevents to apply the semi-empirical method. Recently, the source spectra of middle and small earthquakes have been examined actively using strong motion records in some regions in Japan (e.g., Iwata and Irikura, 1988; Kamae et al., 1990; Takemura et al., 1990).

4.3 Numerical approach

Instead of the empirical Green’s function described in the last section, the theoretical Green’s function can be used to simulate and predict ground motion, not only for a flat layered structure but also for an irregularly layered structure. Many studies utilizing the theoretical Green’s function have been performed by taking account the source process using mathematical descriptions that were originally proposed by Aki (1968), Haskell (1969), and Kanamori (1970). In discussing synthesis and/or simulation of ground motion generally in an irregular structure by a numerical approach, a methodological review of the wave propagation analysis should first be presented. However, a review is omitted to avoid duplication to other sections of this special issue. Review papers on methodology for wave propagation analysis have been published by Aki (1988), Kohketsu and Takenaka (1989), Motosaka (1990), and Kawase (1990). They discussed the features of various methods for wave propagation analysis by showing numerical examples.

This paper reviews only studies which discuss simulation analyses using strong motion records. First, some simulation analyses of ground motion in sedimentary basins for assumed plane wave incidence are introduced. Then, the synthesis of ground motion considering the dislocation source is described.

4.3.1 Simulation of ground motion of irregular site due to incident plane waves

Many simulation analyses for incident plane waves (SH and SV waves) have been performed using observation records as input motion. Although source information is not incorporated in this approach, it is important in earthquake engineering to solve the site response problem due to the incident plane wave for evaluating local site effects, or differences in ground motions affected by the underground structure to determine a ground motion at a specific site. Thus observation records at rock outcrops are usually used in the simulation analysis as input motions on the assumption of body wave propagation with an expected incident angle or surface wave propagation. Ohtsuki (1984) simulated the observation records at the small sediment-filled valley using the 2D finite difference method. Motosaka et al. (1988) simulated the observation records in Mexico City during the 1985 Michoacan earthquake using 2D FEM-BEM hybrid models (Fig. 23). In these examples the numerical results show relatively good agreement with the observation records. However, it should be noted that the observation records used as input motions sometimes include different wave groups with the different wave propagation paths which should be recognized beforehand, because the wave
propagation characteristics in irregular sites for body wave incidence are quite different from those for surface wave incidence. In this context, the later phases observed in a large sedimentary basin such as the Kanto basin suggest two possibilities for the surface waves. One is generated or converted surface wave due to body wave incidence. The other is transmitted and amplified surface waves due to surface wave incidence. Motosaka (1990) and Kamata et al. (1991) showed that the ground motion in basins is strongly dependent on the incident wave to the basin and suggested the observed record at Kohto in Tokyo due to the 1990 Izu Oshima kinkai earthquake is mainly composed of surface waves transmitted and amplified in the basin. The contribution of the surface waves generated by the body wave incidence is rather small. In these studies, the Aki-Lamer method (Aki and Lamer, 1970) is used for body wave incidence and the hyperelement method (Kausel and Roesset, 1977) is applied to the Love wave propagation. The simulation analysis by Yamanaka et al. (1989) based on the surface wave incidence yields better results.

4.3.2 Synthesis of ground motion considering the dislocation source

Simulation analyses considering the source and the propagation path as a flat layered structure have been performed since the latter half of the 1970's using various methods for calculating the Green’s functions. Kudo (1978) simulated observed records in Tokyo due to the 1974 Off Izu Peninsula earthquake using the normal mode method and interpreted the long period content as the lower modes of Love wave. These surface waves are influenced by the thick sedimentary layers in Tokyo which were estimated by Shima et al. (1976) based on an explosion test at Yumenoshima Island in Tokyo, and by Ohta et al. (1977) based on S-wave logging in a deep bore-hole. Yamada and
Noda (1986) used the discrete wavenumber FEM which can synthesize the full wavefield and applied the method to synthesize the ground motion for engineering use. Heaton and Helmberger (1977) simulated the records at El Centro due to the 1968 Borrego Mt. earthquake using generalized ray theory. The wavenumber integration method has also been used to synthesize the ground motion (e.g., Bouchon, 1981; Luco and Apsel, 1983).

Although these studies have made it possible to theoretically synthesize the ground motion on a flat layered structure, the observation records in and around the basin cannot be simulated by the same numerical model.

Numerical modeling has advanced from merely a flat layered structure to 2D/3D irregular structures. Kamiyama (1978) and Kudo (1980) applied Alsop’s method (Alsop, 1974).

![Fig. 24. Subsurface structural model of the Kanto basin. The profile from Enoshima (ENS) to Ryogoku (RGK) (Yamanaka, 1990).](image)

![Fig. 25. Comparison of the synthetic Love wave with the displacement records observed at Ohfuna (OFN), Kawasaki (KWS), and Ryogoku (RGK) (Yamanaka et al., 1991).](image)
which assumes a vertical discontinuity of two regions of horizontal flat layers. Yamanaka (1990) investigated the surface wave propagation characteristics of a sedimentary basin combining the normal mode method and 2D finite difference method, by simulating observation records in the Kanto basin due to the 1990 Izu Oshima earthquake. They used the fundamental Love mode solution due to a double couple source as the input motion for the 2D basin model as shown in Fig. 24. The results are shown in Fig. 25. Hisada et al. (1988) used the 2D Boundary Element Method considering a point source. A three-dimensional investigation was executed by Sato (1990), in which a hybrid model of thin layer elements and axisymmetric finite elements is used for simulating the observation records of Kobe and Osaka during the 1961 Kita-Mino earthquake. Figure 26 shows the simulation results.

A full 3D analysis considering the actual multi-layered irregular structure of the large sedimentary basin is still impossible, even though remarkable progress has been made in computational ability. However, a simplified 3D analysis may be useful at this stage. Toshinawa and Ohmachi (1990, 1991) proposed a simplified 3D FEM and make a 3D wave propagation analysis of the Kanto basin. Tamura and Suzuki (1988) proposed a quasi-3D lumped mass model for the wave propagation analysis of a sedimentary basin.

Finally, it should be mentioned that even though computational ability will become powerful enough to enable full 3D analysis in the future, it cannot be useful without adequate information on the actual 3D irregular structure.

5. Conclusions

The development of strong motion seismology in the last decade has been characterized by three points. One is the accumulation of array observation data of strong ground motion. Many vertical and/or horizontal array networks with digital recording systems have been established in several regions in Japan and high quality data have been obtained. Another is advances in numerical techniques for evaluating
local site effects due to subsurface irregularities. This has been supported not only by developments of computer abilities but also by the fact that ground conditions showed a greater influence on structural damage during some large earthquakes in the last decade than previously thought. The other is the introduction of source heterogeneities in models for evaluating strong ground motion during large earthquakes. This has been supported by the results of source processes of the large events deduced from observation records by using inversion techniques.

The deterministic approach for predicting strong ground motion has advanced with the development of numerical techniques in the last decade and deterministic simulation of low-frequency strong-motion records has been successfully performed in some regions where subsurface structures have been well known. However, there are few regions where the subsurface structures over the basement have been clarified. Most big cities in Japan are located on sedimentary basins where large amplification of low-frequency ground motion is expected. Seismic surveys of subsurface structures in and around city areas are urgently required for engineering seismology.

However, an essential limitation of the deterministic approach for high-frequency strong ground motions has been indicated by the results of array observations. This limitation is caused by the incoherency of the high-frequency waves. Some empirical or semi-empirical methods have been proposed to resolve this problem. These methods show sufficient results for the simulation of not only low-frequency but also high frequency ground motion at the observation sites where useful records are available. At present they are the most realistic approaches for predicting strong ground motions, especially in the high-frequency range.

Another method of solving this problem is the probabilistic or stochastic approach. These approaches have not produced satisfactory results in the last decade because there were few data on the spatial variation of physical parameters describing the heterogeneities of the underground structures. The development of a practical method for measuring the heterogeneities of the underground structures is indispensable to establish the probabilistic or stochastic approach.

Another problem in understanding high-frequency ground motions is to elucidate the phenomena of soil nonlinearity due to the strain dependence of the physical parameters of the ground, especially at alluvial sites. There are few cases in which the nonlinearity of the soil can be identified in strong motion records, though it is expected from many experimental results. The practical use of experimental results in evaluating site amplification of strong ground motions during large earthquakes is an important future topics for predicting strong ground motions at alluvial sites.

All the works described above will contribute to more precise evaluation of Green's functions for the earthquake. This is indispensable to facilitate our understanding of the source processes of past earthquakes from the observed records. It is also necessary in predicting strong ground motions to make a practical model of the heterogeneous source process of large earthquakes on the basis of the results of past events.

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REFERENCES

Abdel-Gaffer, A. M. and R. F. Scott, Shear moduli and sampling factors of earth dam., *J.

Abrahamson, N. A., Estimation of seismic wave coherency and rupture velocity using the

Akamatsu, J., Seismic amplification by soil deposits inferred from vibrational characteristics of

Akamatsu, J., M. Fujita, and H. Kameda, Long-period (1–10 s) microtremor measurement in
the area affected by the 1989 Loma Prieta earthquake, Proc. 4th International Conf. Seismic

Aki, K., Space and time spectra of stationarily stochastic wave with special reference to micro-


Aki, K., Seismic amplification by soil deposits inferred from vibrational characteristics of

1972.

Aki, K., Local site effects on strong ground motion, in *Earthquake Engineering and Soil Dynamics

Aki, K., and K. Irikura, Characterization and mapping of earthquake shaking for seismic zonation,

Aki, K. and K. L. Larner, Surface motion of a layered medium having an irregular interface due


Alsop, J. E., Transmission and reflection of Love waves at vertical discontinuity, *J. Geophys. Res.,*

Boore, D. M., Stochastic simulation of high-frequency ground motions based on seismological

Bouchon, M., Simple method to calculate Green’s functions for elastic layered media, *Bull.

Cambell, K. W., Strong motion attenuation relations: a ten years perspective, *Earthq. Spectra,

Capon, J., High-resolution frequency wavenumber spectrum analysis, Proc. IEEE, Vol. 57,

Celebi, M., Topographical amplification—a reality?, Proc. 9th World Conf. Earthquake

Celebi, M., J. Prince, C. Dietel, M. Onate, and G. Chaves, The culprit in Mexico City-amplification

Chin, B.-H. and K. Aki, Simultaneous determination of source, path, and recording site
effects on strong ground motion during the Loma Prieta earthquake—a preliminary result on


Irikura, K., Earthquake ground motions influenced by irregularities of sub-surface topographies, 7th World Conf. Earthquake Engineering, 175–182, 1980.


Ishida, K., Study on estimating the characteristics of the strong motion spectrum—Comparisons of the spectrum estimated by the simple semi-empirical formula and observed one on rock, Proc. 7th Japan Earthquake Engineering Symposium, 379–384, 1986 (in Japanese with English abstract).


Iwata, T. and K. Irikura, Simulation of wide-frequency band strong ground motion based on
the heterogeneous rupture process, 8th Japan Earthquake Engineering Symposium, Vol. 1, 205–210, 1990.


J. Phys. Earth


Vol. 43, No. 3, 1995


J. Phys. Earth


Noda, S., E. Kurata, S. Iai, and H. Tsuchida, Observation of earthquake motions by dense
instrument arrays in port area and preliminary analyses, Proc. 6th Japan Earthquake
Nozawa, Y., H. Ohki, and T. Annaka, Strain dependence of soil properties inferred from the
strong motion accelerograms recorded by a vertical array, Proc. 9th World Conf. Earthquake
Ohno, S., T. Ohta, T. Ikeura, and M. Takemura, Revision of attenuation formula considering
the effect of fault size to evaluate strong motion spectra in near field Tectonophysics, 69–81,
1993.
Ohta, T., K. Takahashi, H. Morishita, H. Koshida, and S. Hiehata, Characteristics of noise signals
Ohta, Y., Seismic bedrock and seismic waves, Part 5. Engineering seismology, special issue
Ohta, Y. and H. Kagami, Ultimate values of period and amplitude on seismic input motion in
with English abstract).
velocities in deep soil deposits—measurement in a borehole to the depth of 3500 meters and
English abstract).
Ohtsuki, A., Effects of lateral inhomogeneity on seismic waves—Observation and analysis,
Okada, N. and Y. Kagami, A point by point evaluation of amplification characteristics in Japan
Okubo, T., T. Arakawa, and K. Kawashima, Dense instrument array program by the Public
Works Research Institute and preliminary analysis of the records, Proc. 8th World Conf.
Omote, S., K. Ohmatsuzawa, and T. Ohta, Recently developed strong motion earthquake
instruments array in Japan., Proc. 7th World Conf. Earthquake Engineering, Vol. 2, 41–48,
1980.
Omote, S., B. Ohmura, S. Iizuka, T. Ohta, and K. Takahashi, Observation of earthquake ground
motion with boreholes, Proc. 7th Japan Earthquake Engineering Symposium, 481–486, 1986
based on the effect of local surface geology, Proc. 9th World Conf. Earthquake Engineering,
Phillips, W. S. and K. Aki, Site amplification of coda waves from local earthquakes in central
Santo, T., Investigation into microseisms using the observational data of many stations in Japan
1959.
Sasatani, T., M. Ikeda, and N. Sakajiri, A study of site effects by means of strong-motion seis-
Sato, H., Fractal interpretation of linear relation between logarithms of maximum amplitude

*J. Phys. Earth*


