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Short Notes

P-wave Amplification by Near-Surface Deposits at Different Excitation Levels

by Igor A. Beresnev[1] and Kuo-Liang Wen

Abstract Possible nonlinearity in the response of sedimentary layers to incident *P* waves is studied using the strong-motion data from a vertical array in Taiwan at the site where a significant nonlinear response to *S* waves had been found. Soil transfer functions are analyzed at different excitation levels with a maximum vertical acceleration at surface reaching $\sim 0.1\ g$, corresponding to the compressional strain of 7×10^{-5}. Weak-motion and strong-motion transfer functions are compared. No effects of reduced amplification or change in resonance frequencies with increasing strain level were observed, suggesting that the response remained linear in the range of strain under consideration.

Introduction

Recent seismological evidence shows that soft soil can respond nonlinearly to the *S*-wave excitation developing a surface acceleration greater than approximately 100 to 150 Gal (cm/sec^2) (Tokimatsu and Midorikawa, 1981; Jarpe *et al.*, 1988; Singh *et al.*, 1988; Chang *et al.*, 1989; Chin and Aki, 1991; Darragh and Shakal, 1991; Wen, 1994; Wen *et al.*, 1994; Beresnev *et al.*, 1995). Nonlinear soil response can be manifested in two ways. First, when elastic hysteresis takes place, it causes an extra damping in high-strain waves compared with low-strain waves (Erdik, 1987, pp. 513–516). This means that the amplification by soil is greater on the weak motion. Second, the nonlinearity of the stress-strain relationship implies that the material's elastic moduli and wave velocities become strain dependent. Because the fundamental frequency of the soil layer is proportional to the wave velocity,

$$f = V/4H, \tag{1}$$

where V is the velocity and H is the layer's thickness (Murphy *et al.*, 1971, p. 114), the resonance frequencies get strain dependent as well and can deviate in the strong ground motion from those in the weak motion.

Lately, laboratory ultrasonic modeling showed that compressional deformation in rock, carried by *P* waves, can also be nonlinear (Meegan *et al.*, 1993; Johnson and McCall, 1994). This is confirmed by the existing geotechnical experience in which the typical compressional stress-strain relationship in soil obeys a nonlinear power law (Finn, 1988, pp. 545–546). However, a few seismological investigations address this problem using the real strong-motion records. For instance, Mavko and Harp (1984) show that the linear elastic model of the saturated porous medium explains reasonably well the observed accelerograms and corecords of pore pressure for the maximum *P*-wave accelerations stretching to ~25 Gal. In the present article, we check the linearity of soil response to dilatational deformation up to acceleration of 105 Gal.

Data and Method

We use the data from a downhole accelerograph array deployed as part of the LSST (Lotung Large-Scale Seismic Test) project in the southwest quadrant of the SMART1 array in Taiwan (Chang *et al.*, 1989; Wen, 1994). By October 1985, two downscaled reinforced concrete nuclear reactor containment structures (1/4 and 1/12 scale) were built at the Lotung Power Substation by the Taiwan Power Company and the Electric Power Research Institute (U.S.A.). Two vertical arrays were installed to record earthquake ground motion at the boreholes drilled to a depth of 47 m. Accelerographs were placed at the surface and depths of 6, 11, 17, and 47 m. The first hole was 3 m away from the 1/4-scale containment structure, while the second was 46 m away. The data from the second borehole are used in this article.

The site is represented by alluvial silty sands and clays with gravel that are geotechnically classified as "deep cohesionless soil" (Abrahamson *et al.*, 1987). Its shallow low-strain *P*-wave velocities are given in Figure 1 (HCK Geophysical Company, 1986). The water table is within 0.5 m

[1]Present Address: 2935 Richmond Road #604, Ottawa, Ontario K2B 8C9, Canada.

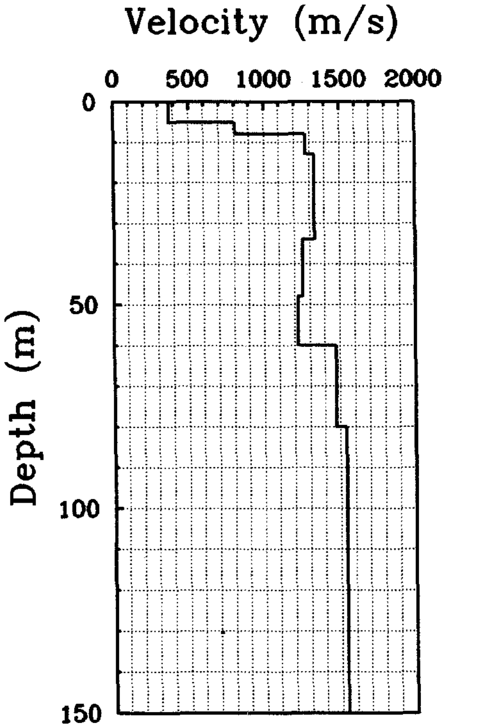

Figure 1. Low-strain P-wave velocities at the LSST borehole.

of the ground surface. There are no records of liquefaction during the history of seismological observations at the Lotung site and in northeastern Taiwan in general, and this site is not considered as potentially liquefiable. The downhole array was equipped with Kinemetrics FBA-13DH sensors connected to the recorders that digitized the data as 12-bit words with 2 g full scale at the rate of 200 samples per second. The pre-event memory was set to 5.1 sec. The preprocessing included baseline subtraction and high- and low-pass filtering with cut frequencies of 0.1 and 50 Hz, respectively.

Calculating transfer functions is a straightforward way to test soil nonlinearity. Indeed, the difference between the values of soil amplification and the frequencies of resonant peaks observed at different excitation levels may directly characterize the nonlinear response. In this study, transfer functions are calculated by taking the ratios of the Fourier amplitude spectra of acceleration at surface to borehole accelerometers. All spectral ratios are calculated as follows: (1) a window containing the longitudinal wave is identified; (2) the window is tapered on both ends using a cosine function; (3) the Fourier amplitude spectrum is calculated; (4) the spectrum is smoothed using a three-point running Hanning average; (5) the ratio of two smoothed spectra is then calculated. The smoothing operator was applied to the raw spectra 10 times. The vertical component of P-wave acceleration is considered in this study.

A signal-to-noise ratio was estimated from accelerograms having sufficiently long pre-event noise time history by dividing the smoothed amplitude spectra of the P wave and the pre-event noise. This analysis showed that the reliable frequency band where the signal-to-noise ratio was greater than 5 extended from approximately 1 to 30 Hz. All the results are plotted in this range.

The LSST array was operational in the 1985 to 1990 period, and it recorded a total of 30 earthquakes with local magnitudes ranging from 3.7 to 6.5 and hypocentral distances from 5 to 107 km. The all-time peak vertical and horizontal accelerations recorded by the surface accelerometer are 192 and 258 Gal, respectively. Because of the limited dynamic range of 12-bit instruments, not all of the earthquakes were well recorded, especially in the low-amplitude P-wave window. Typically, the motions having a peak acceleration larger than approximately 10 Gal can be considered well recorded. From the available data base, five earthquakes that could be attributed to the "weak motion" class satisfied this criterion. Event 12 with a peak vertical acceleration of 108.5 Gal was classified as the "strong motion." The parameters of all selected events are listed in Table 1, and their epicenters placed in Figure 2. Because the accelerometer at the depth of 47 m went unoperational for the earthquakes subsequent to number 11, we do not address its recordings in the present analysis.

The last column of Table 1 shows the peak dilatational strains calculated at surface. These estimates were made using the following technique. For a vertically propagating P-wave pulse, if its displacement is given by $u = f(t - z/V_p)$, then the strain is given by

$$|\varepsilon| \equiv \left|\frac{\partial u}{\partial z}\right| = \frac{f'(t - z/V_p)}{V_p} = \frac{v(t)}{V_p}, \qquad (2)$$

where $v(t)$ is ground velocity. In the calculation of maximum strain, the values of peak ground velocity listed in Table 1 and $V_p = 600$ m/sec were used. Table 1 shows that the maximum strain induced by the strong event 12 has the order of 7×10^{-5}, while it is below or equal to 10^{-5} in the weak motions. Using the only vertical component, however, we could underestimate the actual strain.

Previous studies revealed that significant nonlinear response to S waves in both deamplification and degradation of wave velocities took place at the LSST site in ground motions having peak horizontal acceleration at the surface over 160 Gal (Chang *et al.,* 1989; Wen, 1994; Wen *et al.,* 1994). The occurrence of nonlinear response to P waves representing a different kind of deformation remained an open question.

Results

Because of the proximity of the earthquake source to the recording site for the strong event 12 (hypocentral distance of 5 km), its P-wave window is restricted to approxi-

Table 1
Selected LSST Events

Event	Date (m/d/yr)	Depth (km)	M_L	Hypocentral Distance (km)	$A_0/A_6/A_{11}/A^*_{17}$ (Gal)	Peak Ground Velocity† (cm/sec)	Peak Strain
				Weak Motion			
5	03/29/86	10.3	3.9	13	32.5/16.8/21.1/18.9	0.59	1×10^{-5}
11	07/17/86	2.0	4.3	6	43.2/25.9/17.9/16.8	0.71	1×10^{-5}
14	07/30/86	2.3	4.2	5	20.9/14.1/12.3/11.2	0.40	7×10^{-6}
15	08/05/86	1.2	4.2	5	26.4/13.2/15.8/14.1	0.41	7×10^{-6}
25	11/10/87	34.4	4.9	44	16.4/12.7/12.2/12.5	0.80	1×10^{-5}
				Strong Motion			
12	07/30/86	1.6	5.8	5	108.5/73.0/67.1/58.3	3.90	7×10^{-5}

*Peak vertical acceleration at the surface, 6, 11, and 17 m, respectively.
†Peak ground velocity is taken from the high-pass filtered (cut frequency of 0.3 Hz) and integrated record of surface vertical acceleration.

mately 1.5 sec and is truncated by the arriving S wave; thus, the spectral amplitudes at the low frequencies cannot be assessed. We analyze the spectral ratios starting from 2 Hz.

Figure 3 exhibits the spectral ratios between the surface and three downhole accelerometers. Ratios for the strong event 12 are shown by the thick lines. On the other hand, the average ratios for five weak events are plotted by thin lines together with their standard deviations that characterize the error in weak-motion transfer function estimates. The soil amplification effect and the presence of resonance peaks are seen on all spectral ratios. The surface-to-6-m weak-motion transfer function shows a fundamental frequency around 17 to 18 Hz where the amplification reaches a factor of 6. The position of resonance is compatible with the velocity profile depicted in Figure 1. The first high velocity contrast occurs at the depth of 5 m where the P-wave velocity jumps from 370 to 810 m/sec. The theoretical resonance frequency of the upper stratum is then given by the formula (1) as $f = 370/20 = 18.5$ Hz that is close to the value observed from the surface-to-6-m spectral ratio. The fundamental frequency shifts to the lower values as ratios to the deeper accelerometers are considered.

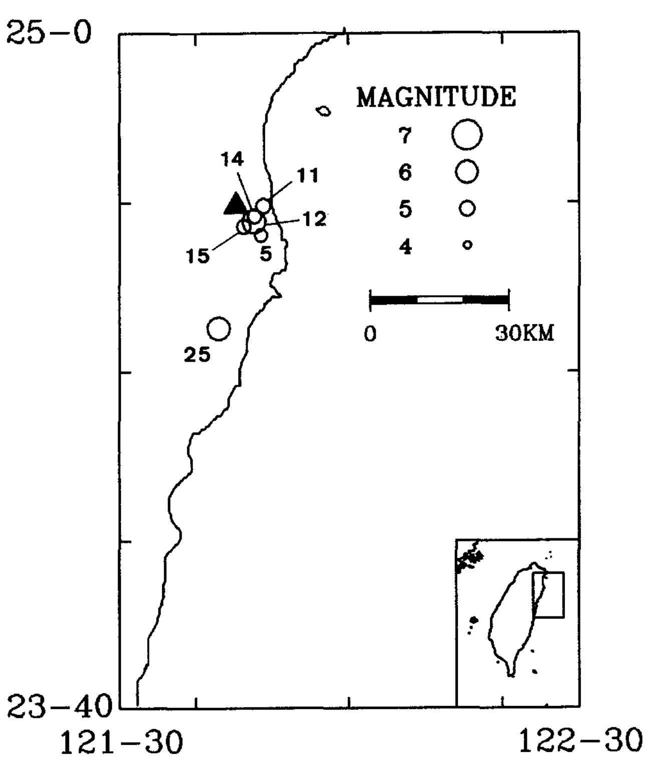

Figure 2. Epicenters of the selected earthquakes (circles). The circle size scales with magnitude, and the triangle stands for the location of the downhole array.

Ratios for all the depths in Figure 3 show no significant difference in the value of amplification or in the position of resonance frequencies between weak and strong motion, given the error margin. Thus, there are no indications of any symptoms of nonlinear response. The P-wave spectrum of event 12 is almost flat to the frequencies of approximately 20 Hz, showing that the strong event had enough energy around the first resonances where the possible nonlinear effect is sought for. The location of the weak-motion resonant peaks is uncertain within the bounds of roughly $\Delta f = \pm 0.5$ Hz. This corresponds to the uncertainty in the identification of the velocity shift effect of $\Delta V = 4H\Delta f = \pm 10$ m/sec for the upper 5-m-thick stratum, implying an error of about 3%. Thus, the response is linear within the uncertainty of a 3% velocity change. This conclusion is in drastic contrast to what was observed for the same layers for the case of the incident S waves, where the decrease in shear-wave velocities due to shearing nonlinearity was as large as 50% (Wen, 1994).

Conclusions

This study does not disclose any significant nonlinear soil response to the compressional deformation up to strains of approximately 7×10^{-5}. The similarity of the amplification values in the weak and strong ground motion presents no evidence of the hysteretic constitutive law, which coincides with the assumption adopted in geotechnical engi-

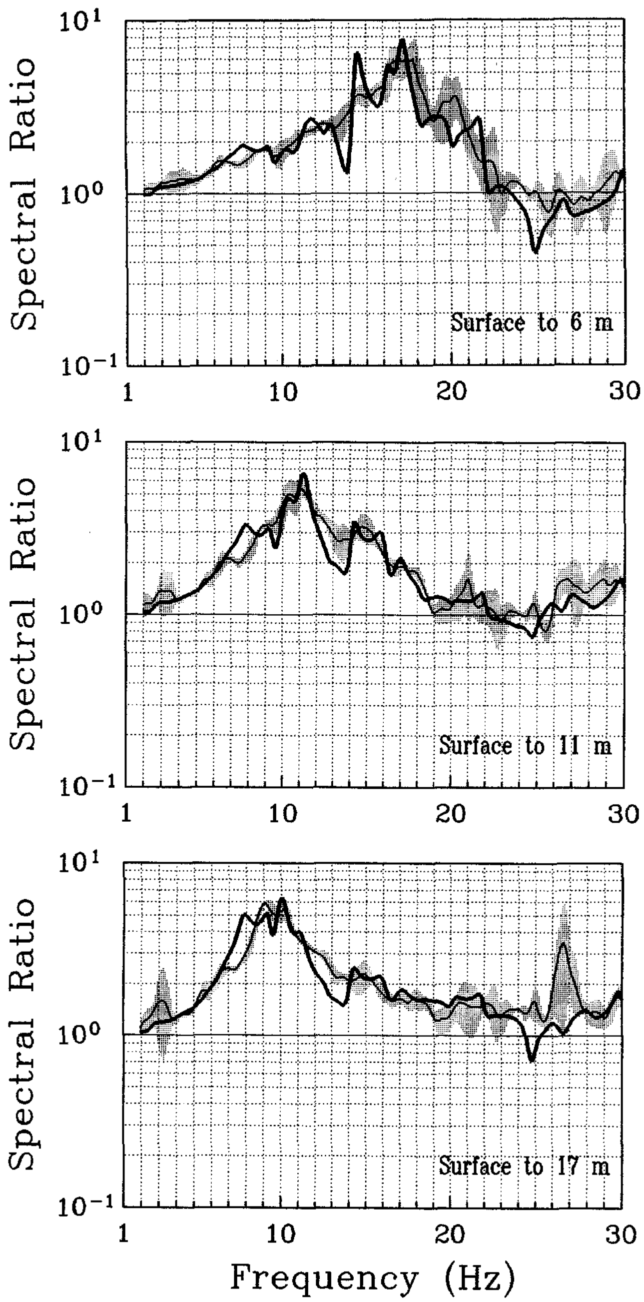

Figure 3. Spectral ratios between the accelerometers at surface and at depths of 6, 11, and 17 m at different excitation levels. The "strong motion" spectral ratio (event 12) is shown by the thick line, while the average for the five "weak motion" events is shown by the thin lines. The shaded bands represent ± 1 standard deviation around the average curves.

neering. However, the shift in resonance frequencies, which could be associated with the dependence of bulk modulus on strain as is also presumed in geotechnical practice, is not identified either. Johnson and McCall (1994) report that strains as low as 10^{-7} produced the observable nonlinear response in their ultrasound propagation experiments in sandstone. We did not observe such an effect within an accuracy of a few percent. Two possible explanations of such an inconsistency can be given at this time. First, for the particular soil profile considered, the level of *P*-wave acceleration ($\sim$0.1 *g*) may be not high enough to reach the nonlinear soil response range. Second, an inconsistency may lie in the methods of nonlinear effect detection. Johnson and McCall (1994) used an alteration of the spectra of ultrasonic pulses with distance as a criterion of nonlinear rock behavior. This was possible because the undistorted source spectrum was known. In our study, we compared the transfer functions of the sedimentary strata for a strong earthquake and a "reference" weak motion. The greater the contrast of accelerations in these compared data sets, the better the method works. However, this contrast is limited by the dynamic range of the strong-motion instrumentation.

We also realize that the record of only one strong earthquake having produced a *P*-wave acceleration of 108.5 Gal was used. Documenting an earthquake with a sufficiently strong *P* wave is a rare occasion. Further clarification of the degree of nonlinearity occurring in the compressional deformation is important to understand the mechanical properties of the *in situ* soils.

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Institute of Earth Sciences
Academia Sinica
Nankang, Taipei 11529
Taiwan

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