DETAILED STUDY OF SITE AMPLIFICATION AT EL CENTRO STRONG-MOTION ARRAY STATION #6

by

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Mueller, C. S., D. M. Boore, and R. L. Porcella, Detailed study of site amplification at El Centro strong-motion array station 6, *Proc. of the Third Int. Conf. on Earthquake Microzonation, 1*, Convened Seattle, Wash., June 28–July 1, 1982, 413–424, 1982.

Abstract

Large ground motions from the Oct. 15, 1979 Imperial Valley earthquake suggested the existence of a local amplification at station #6 of the U. S. Geological Survey's El Centro strong-motion array. The vertical-component accelerogram obtained at station #6 is of special engineering and seismological interest because the recorded peak acceleration is 1.74 g, the largest strong-motion acceleration ever measured and a factor of 2.5 greater than nearby stations #5 and #7. Using a spectral-ratio technique and vertical-component accelerograms from eight other Imperial Valley earthquakes, we found a consistent spectral-amplification peak at station #6: a factor of 2 to 5 near 10 Hz. We were surprised to find large site-response differences in an area of flat topography and homogeneous surface geology; we looked for an explanation in the transfer functions of the soil columns at stations #5, #6, and #7, calculated from borehole traveltimes. With increasing depth, P-wave velocities greater than 1500 m/sec (indicating water saturation) are first encountered within 2.5 m of the surface at stations #5 and #7 and between 5.0 and 10.0 m at station #6. Average P-wave velocities in the dry material above this depth at station #6 are 400 to 500 m/sec. We propose that the site amplification is due to this large P-wave velocity contrast. The transfer function of any reasonable velocity-Q model consistent with the traveltimes and driller's log at station #6 shows a spectral peak (amplitude, 3-6; frequency, 12-15 Hz) in agreement with observed spectra. At stations #5 and #7, the theoretical spectral peak is above 20 Hz and is not observed in the data. Horizontal ground accelerations at station #6 are also higher than stations #5 and #7, but in contrast to the vertical accelerations, they do not correspond to a simple peak in the observed or theoretical spectra.

Introduction

It is well known that ground motions from earthquakes and explosions can be quite variable at nearby sites and it is generally agreed that the local, near-surface geology can amplify or deamplify the ground motion and lead to local variations in damage. However, two questions concerning the amplification of strong ground motion at a site remain controversial. Can the amplification of strong ground motion be predicted using site amplifications determined from smaller ground motions? Will the amplification of strong ground motion at a site be consistent as the distance, azimuth, depth, and strength of the source vary?

Ground motions from the Oct. 15, 1979 Imperial Valley earthquake

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suggested the existence of a site amplification at station #6 of the U. S. Geological Survey's El Centro strong-motion array. A 1.74 g vertical acceleration (the largest strong-ground-motion acceleration ever measured) was recorded at station #6, a factor of 2.5 times greater than the motions recorded at nearby stations #5 and #7. In addition, a review of ground motions from past earthquakes revealed that station #6 has consistently recorded peak accelerations greater than nearby stations by a factor of up to seven on the vertical component and a factor of up to four on the horizontal components. For many earthquakes station #6 was the only station triggered; conversely stations #5 and #7 never triggered without station #6 (as long as station #6 was operational). If such a site amplification exists and is consistent at station #6, it implies that the mainshock ground acceleration, relative to stations #5 and #7, could have been predicted using motions from smaller earthquakes. Beyond the general seismological and engineering interest in the extremely high accelerations recorded at station #6, this observation has obvious implications for the siting of seismically-sensitive structures. Furthermore, it is of interest from a soil-mechanics viewpoint since it implies that the soil column behaved linearly - and in some sense predictably - during the strongest shaking.

Since the installation of the El Centro array, nine earthquakes have been recorded at stations #6 and #7 (seven of these were also recorded at station #5) providing an excellent opportunity to determine the existence and the consistency of the local amplification at station #6 for sources well distributed in azimuth, distance, and strength. In our method, we assume a common input motion below the three stations for each earthquake, dominated by common source and path effects. This input motion is modified at each station by the local site response that we wish to describe. We use a spectral technique because the local wave amplification leading to large ground accelerations should appear as a spectral peak. Furthermore, we take ratios of spectra because the source and path effects common to nearby stations will cancel, leaving the local site responses to dominate the ratios. We also test and confirm the suggestion of Tsujiura (1978) that the seismic coda can be used to delineate the site response. In fact we find that coda spectra are often undisturbed by radiation and path effects that sometimes obscure the site-response peak in body-wave spectra. Finally, we calculate synthetic spectral ratios from detailed knowledge of the near-surface sediment P- and S-wave velocities and compare these with the observed spectral ratios.

Our study differs from previous work in several ways. Both vertical and horizontal peak motions at station #6 are greater than at stations #5 and #7 , but we find a consistent spectral-amplification peak for the vertical component only. For this reason (and because of the importance of the very high mainshock vertical accelerations) we emphasize the vertical components in this work, although the horizontal components are usually considered more important in strong-motion work. Unlike most previous work, we do not compare a soil site with a bedrock site. Array stations #5, #6, and #7 would all be considered soil sites under the criteria of previous studies. We will show that the very shallow P-wave velocity structure responsible for the site amplification is not obvious from the surface geology.

Data

Locations of the El Centro array stations #5, #6, and #7, and the surface traces of the Imperial and the Brawley faults are shown in Figure 1. In this study, we have used all nine earthquakes that have triggered both stations #6 and #7 since the installation of the El Centro array in the mid-nineteen-seventies. (For simplicity we concentrate on a comparison of stations #6 and #7 in this paper, since the responses at stations #5 and #7 are similar.) Epicenters of eight of the nine earthquakes used in this study are plotted in Figure 1 and details of all nine earthquakes are given in Table 1. Locations and magnitudes (M_L) for three Nov. 14, 1977 swarm earthquakes (S1-S3) are from C. E. Johnson (written communication, 1977) who used data from the USGS/CIT Imperial Valley network. Peak accelerations for S1-S3 are from Porcella (1978). Information on the Oct. 15, 1979 mainshock (M) is from Archuleta (1982). We located three of its aftershocks (A2-A4) using strong-motion phase times supplemented by times from the USGS/CIT network (K. Hutton, written communication, 1981) and a velocity model suggested by Boore and Fletcher (1982). A1 was in the mainshock coda and was not located. Magnitudes (M_L) for A2-A4 are from P. German (oral communication, 1982). Peak accelerations for M and A1-A4 are from Porcella (1980). All data on the April 26, 1981 Westmorland earthquake are from Maley and Etheredge (1981).

The original film accelerograph records were automatically laser digitized at a nominal 600 samples-per-second (unequal sample interval); manual intervention by the digitizer operator was used on some traces during the strongest shaking but does not affect our results. Each trace was interpolated and evenly resampled at 200 samples-per-second, the skew of the film transport was removed using the digitized reference traces, and a DC trend was removed by subtracting the average DC level of the last 30% of each trace. No instrument correction was applied, since the instrument constants were nearly equal (A. G. Brady, written communication, 1980) and most of the results in this paper are presented in the form of spectral ratios. To calculate spectra, a smoothly tapered window (first and last 10%) was applied prior to fast-Fourier transformation.

Each amplitude spectrum was smoothed by a running triangle of effective width equal to 2 Hz. This width, determined by trial and error, was wide enough to reduce variance in the spectrum but not so wide as to significantly bias against the spectral feature we wished to delineate. As a general rule, we feel that the highest useful frequencies are 30 Hz for the vertical-component and 20 Hz for the horizontal-component seismograms; below these frequencies signal is at least a factor of ten above a high-frequency noise floor. Many spectra have tails at frequencies below 2 Hz due to long-period errors not removed by the processing, and since the effective width of the spectral smoothing window is 2 Hz, amplitude values below 2 Hz are not fully smoothed. For this reason, the spectra are not considered reliable below 2 Hz. Smoothing was done before spectral ratioing.

Vertical-component mainshock accelerograms showing the high accelerations at station #6 are shown in Figure 2a. Using the windows indicated above the accelerograms in Figure 2a, we calculated spectra and #6/#5 and #6/#7 spectral ratios for strong shaking (Figures 2b,2c) and coda (Figures 2d,2e). The aforementioned similarity of response at

TABLE 1

	ORIGIN (UTC) YR/DAY	HR:MN:SC	LOC LAT/LONG	DEPTH KM	M_L		PEAK ACCEL. #5	#6	#7
						230°	0.12	0.50	0.10
S1	77/318	00:11:35	32 49.6N	5.3	3.9	UP	0.05	0.13	0.04
			115 28.0W			140°	0.17	0.45	0.11
						230°	0.17	0.38	0.14
S2	77/318	02:05:47	32 49.5N	4.6	4.2	UP	0.04	0.11	0.03
			115 28.0W			140°	0.15	0.41	0.10
						230°		0.12	0.06
S3	77/318	05:18:02	32 49.8N	5.6	3.7	UP		0.06	0.01
			115 27.7W			140°		0.13	0.03
						230°	0.40	0.45	0.52
M	79/288	23:16:54	32 39.5N	8.0	6.6	UP	0.71	1.74	0.65
			115 19.8W			140°	0.56	0.72	0.36
						230°	0.04	0.10	0.04
A1	79/288	23:17:40				UP	0.04	0.14	0.04
						140°	0.04	0.09	0.04
						230°	0.05	0.10	0.03
A2	79/288	23:18:19	32 50.8N	8.8	3.4	UP	0.02	0.05	0.02
			115 30.9W			140°	0.05	0.08	0.05
						230°		0.17	0.03
A3	79/288	23:18:40	32 47.4N	6.7	3.8	UP		0.07	0.01
			115 27.9W			140°		0.11	0.02
						230°	0.29	0.26	0.23
A4	79/288	23:19:29	32 46.5N	10.3	5.2	UP	0.12	0.08	0.09
			115 25.6W			140°	0.24	0.18	0.15
						230°	0.06	0.05	0.05
W	81/116	12:09:28	33 07.8N	8	5.6	UP	0.01	0.03	0.01
			115 39.0W			140°	0.05	0.06	0.03

stations #5 and #7 can be seen in Figure 2. The site amplification corresponds to a peak in the station #6 spectra and spectral ratios near 9 Hz. For strong shaking (Figure 2c) the spectral ratio peak corresponding to the local amplification is a factor of 5 to 8 at 9 Hz. In the 2 to 5 Hz band, the ratio equals 1.5 to 3 and at frequencies above the peak, the ratio is near unity. For coda (Figure 2e) the spectral peak is a factor of 3 to 4 centered at 10 Hz. At 2 to 5 Hz the ratio is near unity and at frequencies above the peak, the ratio is generally less than unity. Seismograms and spectral ratios (#6/#7) for the aftershocks (A1-A4), swarm earthquakes (S1-S3), and Westmorland (W) are presented in Figures 3, 4, and 5, respectively.

Interpretation of Spectral Ratios

The observation that prompted this study is shown in Figure 2. Figure 2a shows the high mainshock accelerations and Figures 2b-2e the spectral expression of the site amplification. For the site response to be consistent, the spectra from the swarm earthquakes, aftershocks, and Westmorland earthquake must resemble the mainshock spectra. Figures 3, 4, and 5 show that in each case the spectral peak corresponding to the site amplification resembles that for the mainshock in amplitude and frequency. In Figures 3 and 4 the site effect is much less obvious in the individual spectral ratios (Figures 3b, 3d and 4b, 4d) than in the average ratios (Figures 3c, 3e and 4c, 4e). These aftershocks and swarm earthquakes are closer to stations #6 and #7 than the mainshock and Westmorland events and it appears that source and path differences have obscured the spectral peak corresponding to the site response. These differences tend to cancel when spectral ratios are averaged, leaving the site-response peak. This confirms the idea of Trifunac and Udwadia (1974) that averages of several sources should be used in site studies. In a different way, the coda also averages over source and path differences and proves, at least in this study, to be a robust indicator of the local response. The Westmorland spectral ratio clearly shows the site effect; this is an important earthquake because the waves approach the stations from a northern azimuth. It should be noted that despite reported peak horizontal accelerations a factor of up to four times greater at station #6 than at stations #5 and #7, we found no evidence in the horizontal-component spectra of a simple spectral peak corresponding to a site amplification.

The difference in the body-wave and coda spectra in the 2 to 5 Hz band is a general feature and may be significant. In the 2 to 5 Hz band, the body-wave ratios have values of 2 to 4 while the coda ratios are near unity. In both cases, the spectral peak corresponding to the site effect seems to ride on top of the 2 to 5 Hz level. Perhaps all direct body waves are amplified in the wedge between the Imperial and Brawley faults. At any rate, this broader-band body-wave amplification should be considered when interpreting absolute spectral-ratio levels. Taking this into account, we can say that the spectral peak corresponding to the site amplification has an amplitude of 2 to 5 at 10 Hz.

Synthetic Spectral Ratios

After the Oct. 15, 1979 mainshock, the U. S. Geological Survey conducted electronic cone-penetration tests and shallow borehole P- and S-wave velocity studies (using the methods of Warrick, 1974) at the El Centro strong-motion array sites. In the velocity surveys, a surface P-wave source (hammer), surface S-wave source (sliding hammer), and buried P-wave source (blasting cap buried 1 meter) were used with fixed uphole and moveable (2.5-meter spacing) downhole geophones. The borehole depth was 60 meters at stations #5, #6, and #7. Arrival times were picked from chart records and a simple correction was made for the effect of the geometry of the experiment on the traveltimes. P-wave traveltimes from the cap, reduced using a velocity of 1500 meter/second, are summarized in Figure 6. P-wave traveltimes from the hammer are not graphed, but match the cap results except for a uniform delay of 7 milliseconds. Times from the cap to the uphole geophone (not shown) indicate that P-wave velocities in the top 1 or 2 meters of material are as low as 250 to 300 meter/second. The hammer and blasting cap are, respectively, above and below this slow material, accounting for much of the uniformly delayed traveltime in the hammer survey. Reduced P-wave traveltimes at stations #5 and #7 are similar and show that 1500-meter/second (water-saturated) material lies just below the slow surface layer at these sites. 1500-meter/second material lies no shallower than 5-10 meters at station #6 and P-waves are considerably delayed above this depth, indicating a large velocity contrast. The cone-penetration test and driller's log show a large lithology contrast at a depth of 8 meters at station #6: clayey silts over dense sand and gravel. S-wave velocities increase smoothly with depth at all three sites from 150 meter/second at the surface to 300 meter/second at the bottom of the borehole.

We interpret these data as follows. The sand and gravel below 8 meters at station #6 form a water-saturated aquifer; this is consistent with P-wave velocities greater than 1500 meter/second below this depth. If the 5-millisecond delay at station #6 relative to stations #5 and #7 is distributed between 2 and 8 meters, the average P-wave velocity is roughly 600 meter/second. (It was difficult to determine accurate P-wave traveltimes using a 2.5-meter geophone spacing, especially in the shallowest part of the borehole which is crucial to our results. For this reason, absolute traveltimes and velocities are subject to error. However, the <u>relative</u> 5-millisecond delay at a depth of 8 meters at station #6 is well determined and is the key to our interpretation.) We conclude that the average P-wave velocity in the dry material between the surface and 8-meters depth at station #6 is 400 to 500 meter/second.

Using matrix wave-propagation techniques (Haskell, 1953, 1962) we have calculated synthetic spectral ratios (#6/#7) for comparison with the data. In this method, the soil column is modeled as a stack of horizontal viscoelastic layers bounded above by a free surface and excited from below by plane P waves of varying angles of incidence. The calculation does not permit nonlinear soil response. The simplest velocity models matching the essential features of the traveltime data are a 1500-meter/second layer at station #7 and an 8-meter-thick layer over a 1500-meter/second layer at station #6. Synthetic spectral ratios for two station #6 models (400-meter/second and 500-meter/second shallow P-wave velocity) and a 10^{o} angle of incidence are shown in Figure 7. Although peak frequencies are shifted, good agreement in the shapes and amplitudes of the synthetic and observed spectral ratios implies that the observed resonance is due to the shallow layer. The synthetic results are not sensitive to the choice of any reasonable S-wave velocity structure that matches the traveltime data. If two layers are included above 8 meters in the model at station #6, the shallowest layer (250 to 300 meter/second) has its own large resonance above 20 Hz that is not seen in the data. Since unrealistically small Q values are required to eliminate this peak, we chose to average the two layers as discussed above. We tried several gradient models and incidence angles without changing the basic results. In fact, any reasonable velocity model at station #6 that matches the relative 5-millisecond reduced-traveltime delay at 8 meters (this requires a large velocity contrast) must resonate near 12-15 Hz.

Discussion and Conclusions

We found that a local amplification exists at station #6 of the El Centro strong-motion array, which has a dominant influence on vertical ground motions at this site. The site amplification corresponds to a peak in Fourier spectral amplitude that we have shown to be constant in shape, amplitude, and frequency over a wide range of source locations and sizes (M_L = 3.4 to 6.6) and ground motion amplitude (0.01 g to 1.74 g for strong motion, less for coda).

Synthetic spectral ratios indicate a spectral peak near 12 to 15 Hz rather than the 10 Hz observed in the data. This peak matches one found by Boore and Fletcher (1982) in the spectral analysis of a small aftershock recorded on portable digital seismographs. Since these instruments were not exactly colocated with the strong-motion stations, it is not known if this frequency shift is due to lateral changes in the amplifying structure or if the peak in the strong-motion data is shifted to lower frequencies by sediment nonlinearity. Since the peak is near 10 Hz for low-amplitude coda, we feel that nonlinearity is not a major factor. Higher peak frequencies in the synthetic spectral ratios may be due to errors in the velocity determinations. We emphasize that the general results of this study should not be extended to shear-wave strong motions because nonlinearity might be significant. (Perhaps the relative equality of the mainshock peak-horizontal accelerations at stations #5, #6, and #7 is due to sediment nonlinearity. One could speculate that 0.7 g is a nonlinearity-imposed, maximum-horizontal acceleration at these stations.).

One general result of our study should be applicable to future work. Other workers (e.g. Trifunac and Udwadia, 1974) have emphasized the importance of source, path, and topographic effects on strong ground motion and stressed that a well-posed site study must average over a number of sources distributed in azimuth, distance, and strength. Our spectral-ratio method suggests a way in which these averages can be taken. We found that averages of spectral ratios are stable indicators of site response in cases where individual body-wave spectral ratios are dominated by source and path differences. We also confirmed the suggestion of Tsujiura (1978) that coda spectra (a different kind of average) can be used to delineate a site response. Studying coda spectral ratios and averages of body-wave spectral ratios should prove useful in future cases where site response has a dominant influence on strong ground motion.

Acknowledgements

We thank W. B. Joyner for useful discussions and for the use of his Haskell-method computer program. A. G. Brady and A. M. Converse assisted in the acquisition and processing of the strong-motion data.

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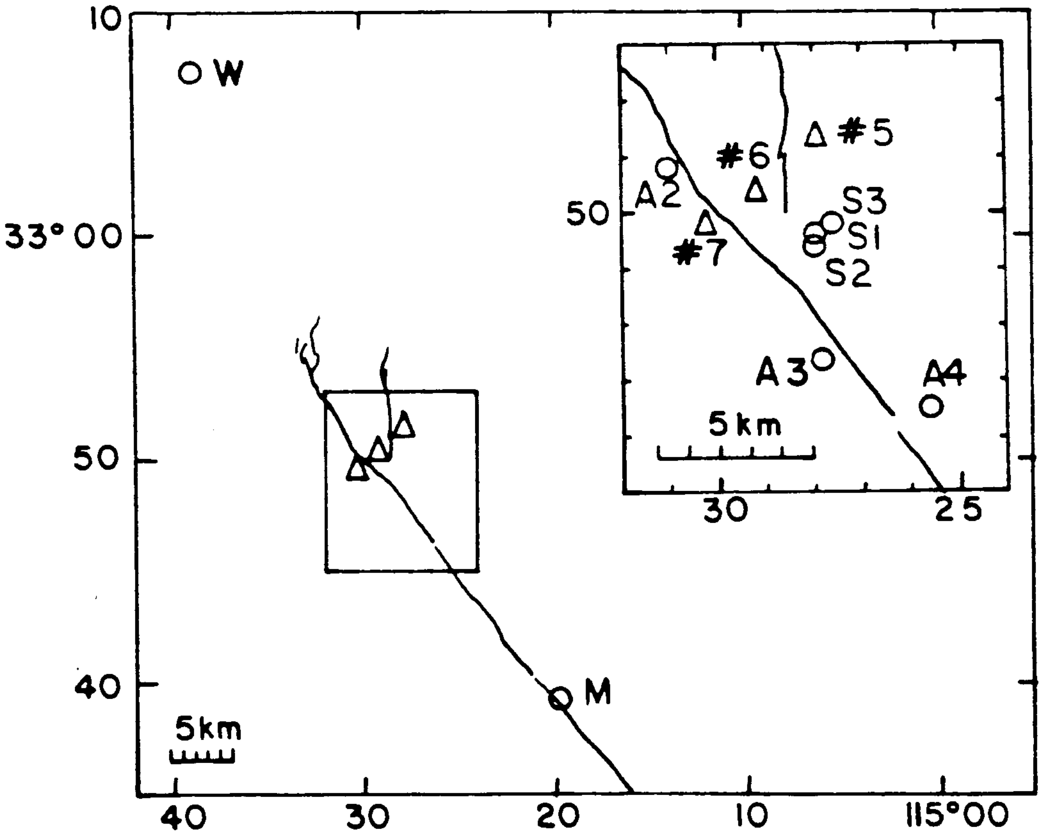

Figure 1. Map with faults, stations, and epicenters.

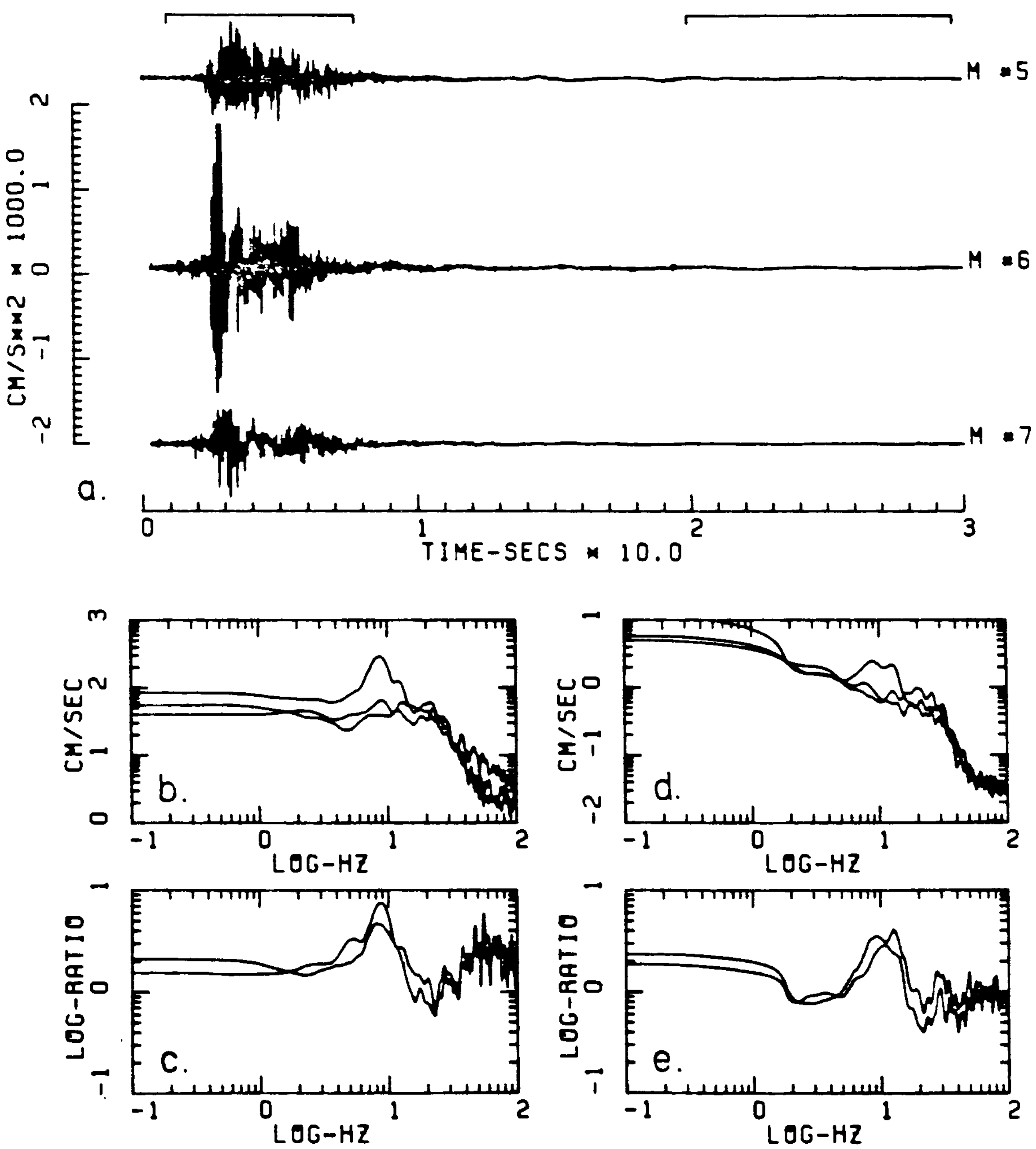

Figure 2. Mainshock vertical-component accelerograms (a), spectra (b,d), and spectral ratios #6/#5 and #6/#7 (c,e).

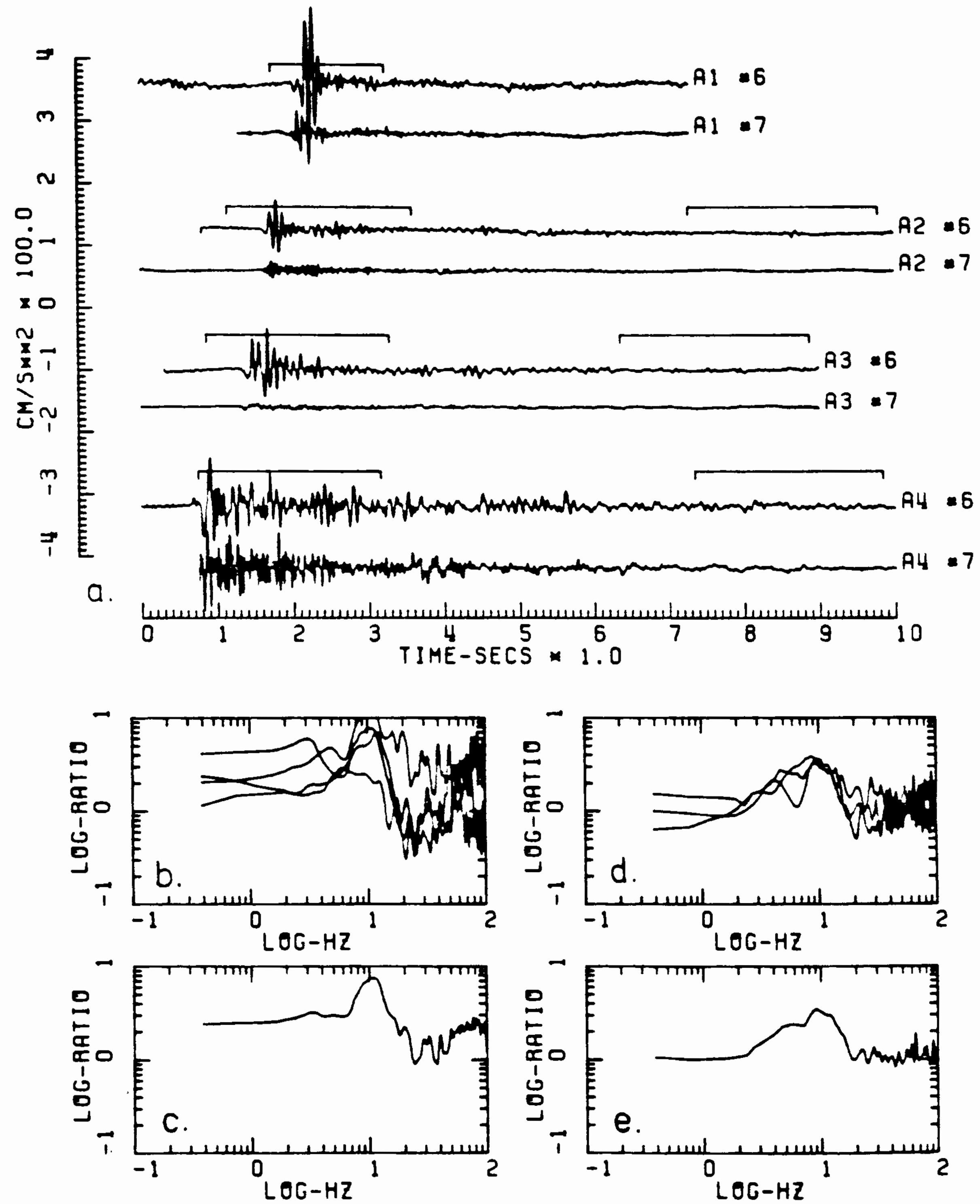

Figure 3. Aftershock vertical-component accelerograms (a), spectral ratios #6/#7 (b,d), and spectral ratio averages (c,e).

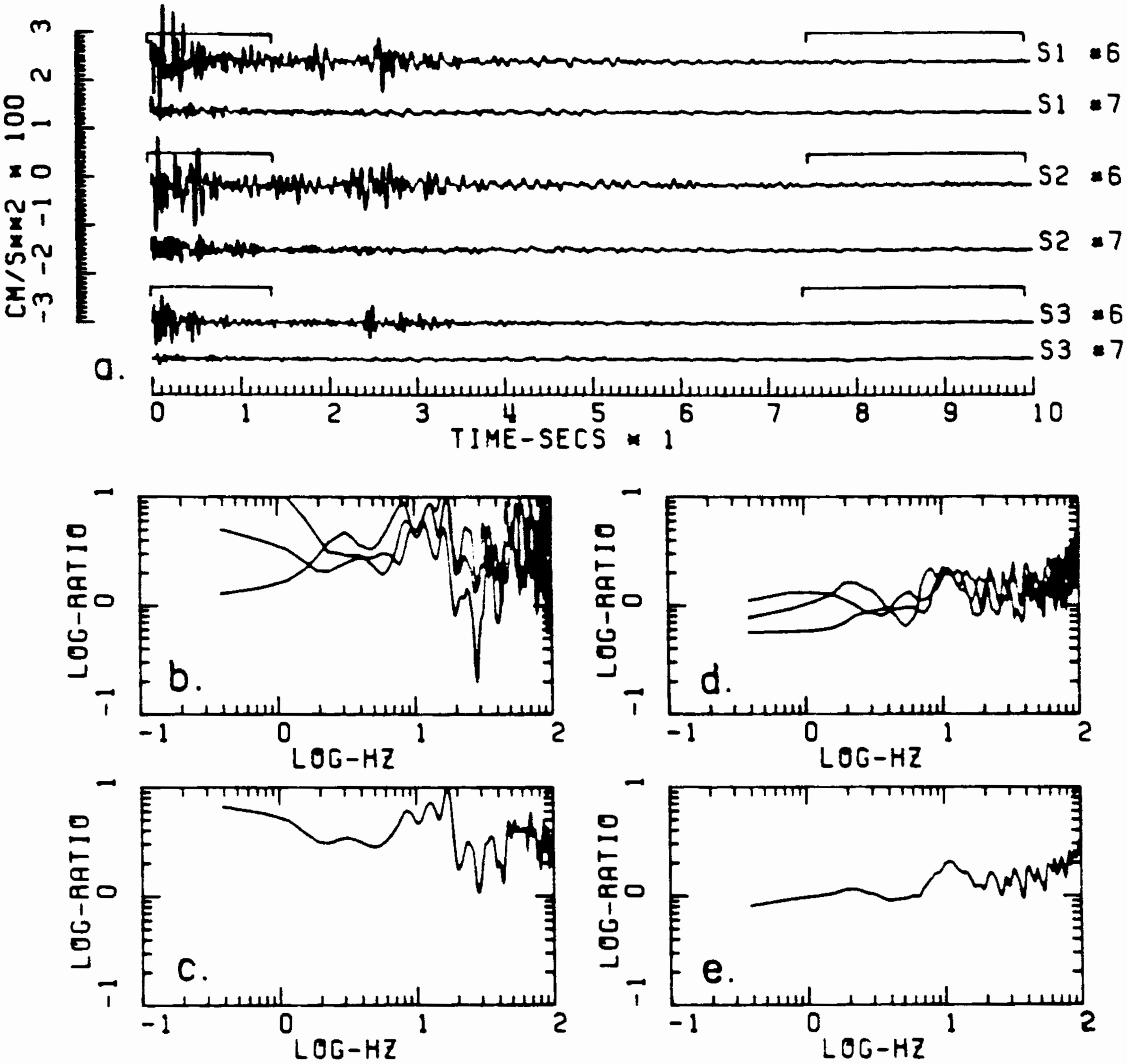

Figure 4. Swarm earthquake vertical-component accelerograms (a), spectral ratios #6/#7 (b,d), and spectral ratio averages (c,e).

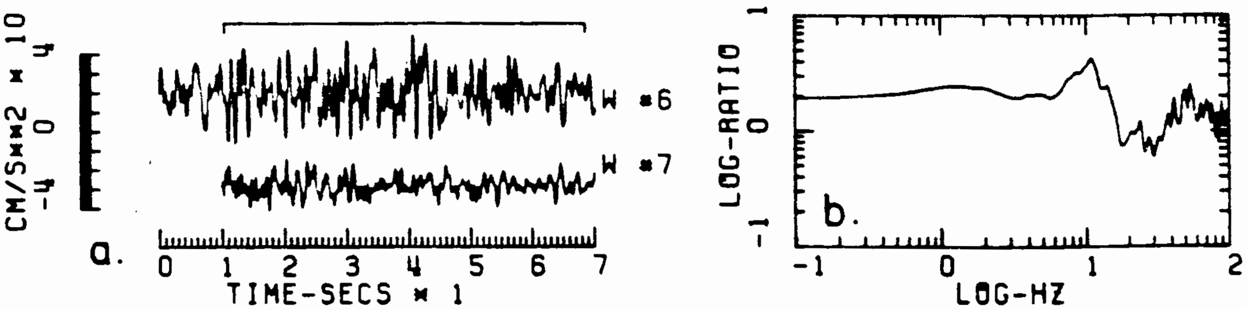

Figure 5. Westmorland vertical-component accelerograms (aligned using WWVB times) (a), and spectral ratio (b).

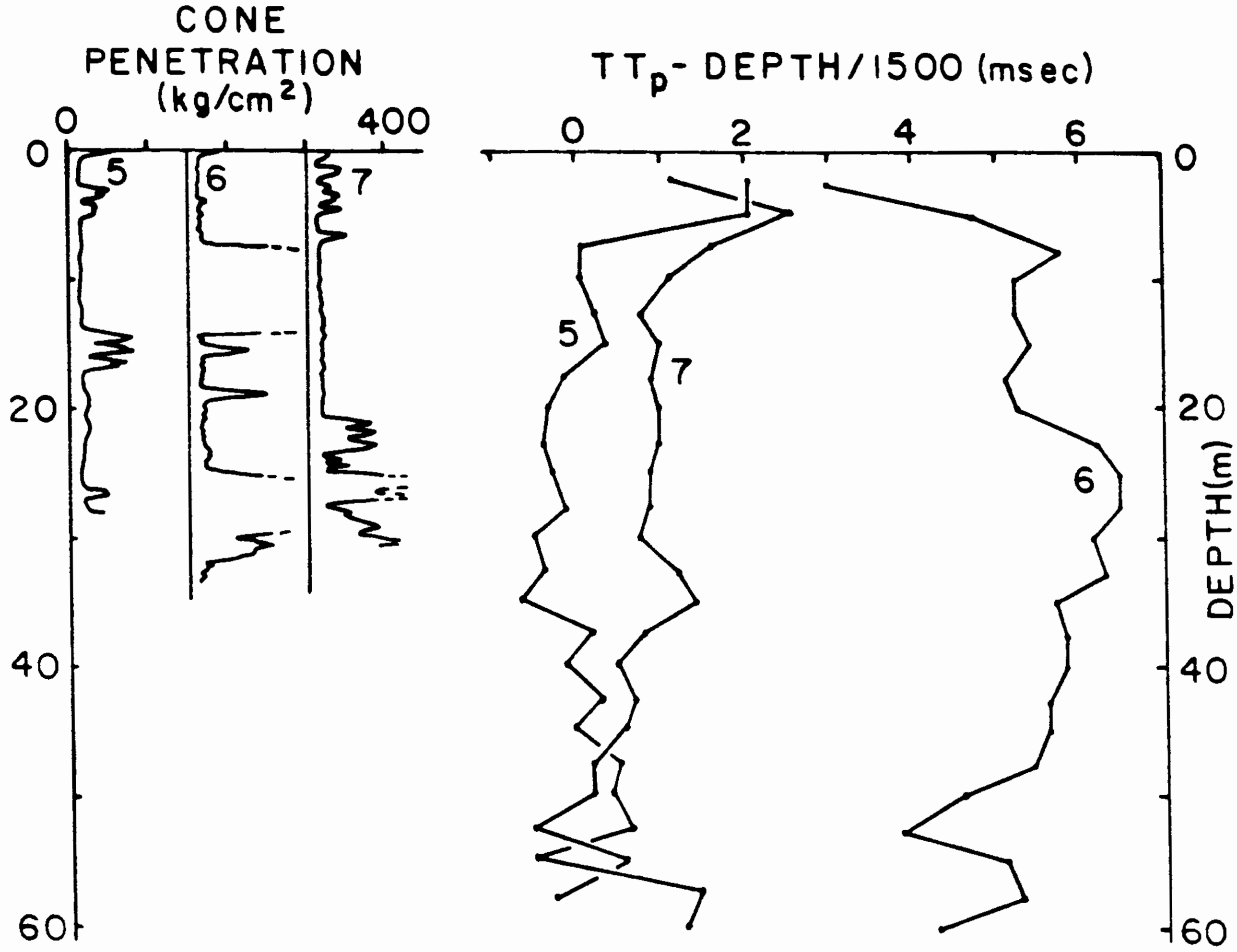

Figure 6. Shallow borehole data: cone penetration and reduced P-wave traveltimes.

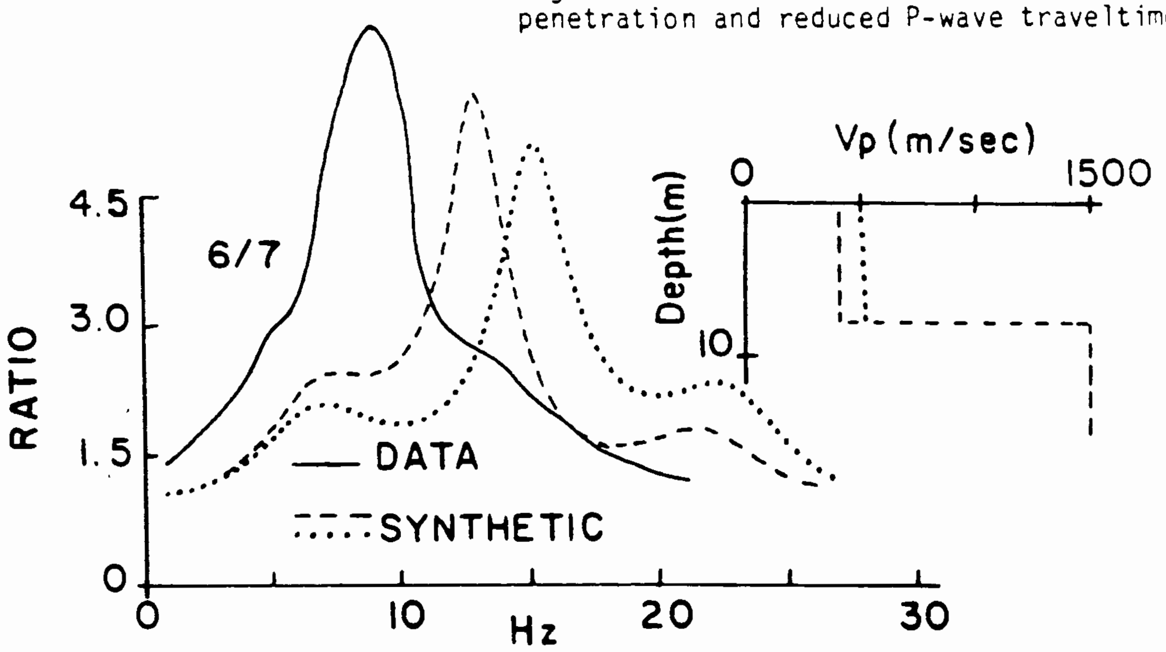

Figure 7. Comparison of synthetic and mainshock spectral ratios.