A Comparison of Coda and S-Wave Spectral Ratios as Estimates of Site Response in the Southern San Francisco Bay Area

by Lucia Margheriti, Leif Wennerberg, and John Boatwright

Abstract  We calculate spectral ratios for S waves and codas to evaluate the amplification of 14 sites in the southern San Francisco Bay area relative to a nearby bedrock site in the Coyote Hills. Our data are seismograms written by Loma Prieta aftershocks: the epicentral distances and azimuths to the stations are effectively the same. All the sites in the study are amplified with respect to the reference site at frequencies from 0.5 to 7 Hz. The shapes of the S-wave and coda spectral ratios are similar, but the coda ratios are greater than the S-wave ratios by as much as a factor of 4. The difference is larger for sites on alluvial and bay mud deposits, particularly at frequencies around 1 Hz, suggesting the presence of waves trapped in the alluvial basin. In general, the length of the analysis window affects the S-wave spectral ratios for alluvial sites. Longer windows give ratios similar to coda ratios, apparently because these windows include more of the phases that contribute to the coda.

We classify the sites according to their geological characteristics and surficial shear-wave velocities. For the S-wave ratios, the differences between the classes showed no systematic trend; the softest and hardest soil classes we consider have practically identical S-wave amplifications. The average coda ratios for the site classes clearly increase as the soil classes include slower and “softer” materials. After correction for differences in reference sites, the coda amplifications are very similar to the relative amplifications for these site classes estimated by Borcherdt and Glassmoyer (1992) from the strong-motion recordings of the 1989 Loma Prieta earthquake.

Introduction

Strong ground shaking and the concomitant damage to structures during an earthquake are critically determined by the geologic setting of a site. Since obtaining the first recordings of ground acceleration, seismologists have sought ways to characterize this dependence, in particular seeking methods by which they can reliably estimate the amplification or site response associated with specific sites. The most important methods that have been developed that consider seismic phase information compare the ground motion in either the direct S wave or in the S-wave coda as observed at different sites for the same earthquake (e.g., Borcherdt, 1970; Tsujiura, 1978; Tucker and King, 1984; Phillips and Aki, 1986). Site-response studies have grown more numerous with the increase in strong-motion accelerographs and digital recorders.

The results of these studies have been applied to the construction of buildings through the building code, and more recently, to the renovation of existing buildings. In these applications, a critical need has developed to understand and model the uncertainties in site-response estimates (e.g., Finn, 1991; Aki, 1988, 1993). Some of this uncertainty is derived from the variability in the response with source location and excitation, but we find that there may also be significant uncertainty derived from differences in amplification observed for different wave types; that is, direct or coda waves.

It has been assumed (e.g., Tsujiura, 1978; Phillips and Aki, 1986) that the relative amplifications derived from these two methods are consistent, but for the significant class of sites in alluvial basins, or for sites that are more widely separated than a few hundred meters, little work has been done that demonstrates their consistency. Some evidence suggests that the relative coda amplitudes are greater than the relative S-wave amplitudes for sites in sedimentary basins as a result of scattered energy trapped in the basin (Philips and Aki, 1986; Olsen et al., 1994; Frankel, 1994). Three-dimensional simulations of seismic-wave propagation in an alluvial basin by Frankel and Vidale (1992; see also Olsen et al.,
illustrate the large amplitude and long duration of ground motion in a basin compared to the surrounding rock, and show a larger factor of amplification at low frequency (<1 Hz) for codas than for direct S waves (Frankel and Vidale, 1992, Fig. 7).

The Loma Prieta aftershock sequence provided an extensive data set (Mueller and Glassmoyer, 1990) with which to compare methods of evaluating site response in the San Francisco Bay area. In this article, we calculate the ratios of S-wave and coda spectral amplitudes from a set of 27 aftershocks recorded at 14 sites in the southern San Francisco Peninsula and the southern East Bay to spectral amplitudes at a nearby rock site. The geometry of the stations and the epicentral distribution of the earthquakes permit a simple comparison between the S-wave and the coda spectral ratios: for each event, the epicentral distances and azimuths are similar for all of the stations. Our choice of this data set is also motivated by the range of geological conditions at the stations: these conditions range from the single hard-rock site used as a reference site for the study, to four soft-rock sites, eight alluvial sites, and two sites on Bay mud.

**Spectral Ratio Methods**

We compare the two methods of estimating site response by computing spectral ratios obtained from two different portions of the seismograms (Fig. 1). For both the direct S waves and the codas, we assume that the record spectrum, $R(f)$, for a given source and recording site,

$$R(f) = ES(f)P(f)SR(f),$$

is the product of a source spectrum, $ES(f)$, a spectrum characterizing the filtering effect of the propagation path, $P(f)$, and a spectrum describing the filtering effect of the site, $SR(f)$.

To estimate the relative response of two sites, the spectrum of a portion of the record from one site is divided by the spectrum of the corresponding portion of the record from another site for the same earthquake. The critical assumption is that the two sites share the same source spectrum, $ES(f)$, and have comparable propagation path effects for the phases included in the sample window. For the narrow range of azimuths and epicentral distances that our data span, any effects of radiation patterns should be minimal (Fig. 2).

For direct S waves, we do not correct for crustal attenuation, but assume that $P(f) \approx 1/\Delta$, where $\Delta$ is the hypocentral distance. This factor varies by less than 20% over the array for any given event, and is therefore a nominal correction. Thus, the spectral ratios of the direct S waves, normalized by the hypocentral distances should provide reliable estimates of the relative site response.

The analysis for the coda amplification differs in one critical way; that is, in the interpretation of $P(f)$. For a sufficiently large lapse time, seismic waves scatter in the Earth such that it is reasonable to assume that seismic energy becomes approximately uniformly distributed over an expanding volume around the earthquake source (Aki, 1969). For a given lapse time (that is, a given absolute time after the earthquake), the propagation functions, $P(f)$, are assumed to be the same for all the stations. Thus, for coda samples taken at the same elapsed time after an earthquake, the factor $ES(f)P(f)$ is the same for all stations, and the ratio of coda spectra should give the relative site response. In principle, the relative site response can be estimated from codas recorded even at widely separated sites.

These simple arguments indicate that the source and path effects should be similarly cancelled by taking spectral ratios of either the direct S waves or the codas, leav-
ing only the relative site responses. Any differences in the resulting estimates of relative site response must be derived from differences in the near-site propagation characteristics.

Data Analysis

The 27 earthquakes used in the study were recorded between November 1989 and January 1990 at sites located between the San Andreas and the Hayward faults in the southern San Francisco Bay area. We focused on this area because of the presence of a range of representative Bay area alluvial units and a conveniently located rock site. Extensive site response studies using aftershocks have not been done previously in this area.

The aftershocks were recorded on 16-bit GEOS digital seismographs (Borcherdt et al., 1985) at 200 samples/sec. The instruments recorded motion from three-component accelerometer and velocity transducers. We used only horizontal data in our analysis, combining the north–south and east–west components by taking the square root of the sum of the squares of the spectral amplitudes at each frequency. We used mostly velocity records, but included six high-quality accelerograms. See Mueller and Glassmoyer (1990) for more details on the instrumentation and recordings.

Locations of the aftershocks and the stations we used are shown in Figure 2 (USGS locations). Most of the events were located along the San Andreas fault in the epicentral zone of the Loma Prieta mainshock. We also included in our analysis events located on the Sargent fault, the Calaveras fault, and on the San Andreas fault south of the mainshock. The epicentral distances in the data set range from 25 to 110 km. For any given event, azimuths typically vary by less than 20°, and generally vary by less than 30°. Table 1 indicates which stations recorded which events; all the aftershocks included in the data set were recorded by two or more stations. Two of the stations, SRL and COY, have the most recordings.

Figure 2. Map of the location of the recording stations and epicenters of events used (U.S.G.S locations).
To assess the quality of the data, we first visually inspected the records to ensure that the codas were not contaminated by instrument noise. We tested the coda signal-to-noise ratio by taking the spectral ratio of a 1-sec pre-event window and a 1-sec window of the coda. These spectral ratios showed that for frequencies higher than 10 Hz, the spectral amplitude of the noise can be comparable to the spectral amplitude of the coda. Stations ESC, IVE, and STQ were the noisiest; for these stations, the upper frequency bound for reliable estimates of the coda amplitude is around 8 Hz. We also considered the signal-to-noise ratio at low frequencies by low-pass filtering the seismograms, since only a short pre-event noise record was available; some recordings of events around $M = 2$ had signal-to-noise ratios close to one at frequencies $<1$ Hz. We excluded these recordings from our analyses for relative site response.

We used 5.12-sec windows (1024 samples) in the spectral ratio analysis both for $S$ waves and codas and processed the data twice. First, we examined the spectral ratios of co-recorded events. In the second pass at the data, we used the generalized inversion method of Andrews (1982, 1986), as implemented by Boatwright et al. (1991). This inversion makes it possible to extract the relative response of a set of stations, all of which need not have recorded the same events. To save computation time, the spectra are sampled at logarithmic intervals where $f_{n+1} = 2^{1/4}f_n$. The noise is assumed to be half the amplitude of the signal, except for those velocity spectra that increase at low frequencies (possibly because of instrument noise). In these cases, the noise is constrained to exceed the signal at low frequency.

We compared the individual spectral ratios and their averages with the estimates from the generalized inversion technique and found good agreement. The results we present are obtained from the generalized inversion technique.

### Determining the Coda Window for Spectral Ratios

While identifying the $S$ wave for a spectral ratio is straightforward, some care must be taken in selecting the

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portion of the coda to be used for site response comparisons. A large enough lapse time must be chosen such that the seismic energy can be assumed to be uniformly distributed under the sites of interest. "Large enough" is typically taken to be at least twice the S-wave travel time to the farthest site. Rautian and Khalturin (1978) found that two to three S-wave travel times had to elapse before coda decays, and hence, the underlying scattering processes, stabilized. Mayeda et al. (1991) and Koyanagi et al. (1992) explicitly tested for the consistency of coda decay rates at stations of interest before computing spectral ratios. We follow their example.

To test the uniformity of the coda decay in our study area, we considered events recorded by most of the stations that had long enough records for a meaningful estimate of coda Q, the parameter defined by Aki (1969) and by Aki and Chouet (1975) to quantify the decay rate of the tail of a seismogram. The coda Q for a portion of a seismogram is determined by fitting the data with the equation $A(t) \exp(\frac{-t}{2Q_c})$, where $A$ is the initial amplitude, $t$ is the lapse time since the origin time, $\omega$ is the angular frequency, and $Q_c$ is the coda Q, which quantifies the decay rate. We used a maximum likelihood method that considers only uncorrelated data points to estimate $Q_c$ and its variance (Haar, 1989). The resulting uncertainties are significantly larger than those from a usual least-squares fit, but allow better-conditioned comparisons of $Q_c$ estimates.

For our coda Q measurements, we selected events with at least 20 sec of record after twice the travel time of the S wave (3300852, 3362002, and 3602246; Table 1). Two of these events were located on the San Andreas fault and the other on the Calaveras fault, Epicentral distances range between 30 and 40 km. We were not able to include any events recorded at sites MDC, ESC, and CNT for these observations; however, the results from nearby sites indicate that scattering processes had stabilized for all stations before the start time of our spectral ratio calculations.

We computed $Q_c$ for each component at each site for four frequency bands: 0.5 to 2, 2 to 4, 4 to 8 and 8 to 16 Hz. In Figure 3, we plot all the $Q_c$ values and the error bars we estimated for the stations that recorded each of the three events considered. We distinguish results for the three events we considered by plotting two of the sets of measurements offset by 1/2 Hz from the center frequency. The three components of motion for each site are plotted separately. The values of $Q_c$ that we found are consistent among the stations for all the frequency bands considered. In a few cases, for the lowest frequency band we found negative values for $Q_c$ with extremely large error bars, but this did not occur consistently for any station. We did not plot these points. These negative estimates seem to be due to late arrivals at low frequencies and not to low signal-to-noise ratios. We also show in Figure 3 the mean values of the $Q_c$'s estimated for each frequency band and their standard deviations. For comparison, we plot $Q_c$ values for central California found by Mayeda et al. (1992) (lower curve) and by Peng (1989) as summarized by Hartzell (1992; upper curve). Our $Q_c$ values are consistent with these previous observations, and show a frequency dependence, with $Q_c$ roughly proportional to frequency.

Figure 3 shows that there is no significant variation in $Q_c$ amongst the sites. We infer that the coda energy incident on the area occupied by the array is spatially homogeneous. Thus, we assume that the incident coda energy has the same amplitude spectrum at all the sites for a lapse time of twice the S-wave travel time, and therefore we assume that the ratios of the observed coda spectra reflect differences in the effects of the sites on the incident energy.

Spectral Shape of the Reference Site

We used COY as the reference station for our site comparisons because it was sited on reasonably hard rock, it recorded a large number of events (Table 1), and its coda decay does not show any anomalous behavior. The site was located in the Coyote Hills on the east margin of the Bay. The station was on a layer of soil directly overlaying the Franciscan formation. This formation is characterized by the highest near-surface shear-wave velocity of all the rock types occupied in the deployment. In this section, we examine the response of the reference site to consider how the site response at COY conditions the spectral ratios.

We estimate the spectral shape for the filtering effect of COY's site response by assuming that the source displacement spectra of microearthquakes are flat up to some relatively high corner frequency (Frankel and Wennerberg, 1989; Wennerberg et al., 1991). In this case, the observed spectral character at lower frequencies directly expresses the filtering effects on waves propagating to the receiver site. If the frequency dependence of crustal propagation at these distances is minimal, the resulting spectral shape will be largely determined by the site response of COY. This latter assumption implies that this approach is most appropriate for direct waves. In any case, a similar analysis for coda waves is difficult, because of the low signal-to-noise ratio in the codas of small events.

We perform this analysis by stacking the displacement (S wave) spectra of small earthquakes (events 3391207, 3560406, 3631132, and 3640935), with magnitudes $1.8 < M < 2.1$, that are not used in the spectral ratio analysis because they have low signal-to-noise ratios (below 1 Hz). We use them here to consider the high frequencies available in these small-event records. Figure 4 shows the average, normalized displacement spectrum for these small events plus or minus one standard deviation (shaded area); the spectrum is scaled for the
comparison with Hartzell's (1992) work (solid line), discussed below. There is a peak at 9 Hz, which shows an amplification of a factor of 4 relative to the longer-period levels, then the spectrum falls off rapidly at higher frequencies, suggesting an average corner frequency above 10 Hz for these four small events. The most striking feature of the stacked displacement spectrum is the peak at 9 Hz, which produces a strong fall-off around 8 Hz in the spectral ratios relative to COY (Figs. 5 through 9).

We interpret this peak as a local site resonance at COY. COY was located on a 2 to 3-m-thick layer of soil; if the S-wave velocity in this soil is 70 to 100 m/sec, the peak at 9 Hz corresponds to a quarter-wave resonance.

To determine more exactly the corner frequencies of the microearthquakes and clarify the frequency band for which the stacked displacement spectrum is a direct expression of the site filtering effect at COY, we divided acceleration spectra of relatively energetic events (M ~ 3.3) by the displacement spectra of co-located smaller events (distances between co-located pairs are on the order of 1 km). This ratio is \(\omega^2\) times the ratio usually calculated in empirical Green's function spectral analysis (e.g., Frankel and Wennerberg, 1989; Hough et al., 1991). The analysis is based on equation (1); for two co-located events recorded at a given site the ratio of the observed spectra is the ratio of the source spectra. The shape of the ratio we calculated is more useful for estimating corner frequencies since it increases up to the corner frequency of the bigger event, becomes flat (according to both observation and the \(\omega^2\) spectral model), and increases again above the corner frequency of the smaller event. We calculated ratios for a couple of microearthquakes for which we had a useful larger event, finding corner frequencies of 10 Hz or more for events with a magnitude of about 2.

Our concern here is to establish a reasonable frequency band over which we are able to observe the spectral shape of the site response of COY, but our estimated

![Figure 3](image-url)  
**Figure 3.** Coda Q values found for the events 3300852, 3362002, and 3602246. The values plotted represent each component for each site (smaller symbols and thinnest error bars); the big diamonds and heavily lined error bars represent the average value we found for each frequency band plus or minus one standard deviation. For comparison, we plot coda Q values for central California found by Mayeda et al. (1992) (lower curve) and by Peng (1989) (upper curve).
corner frequency has implications for microearthquake source parameters, and deserves a brief discussion. A corner frequency of 10 Hz implies a source radius of \( r = 0.21 f_c \approx 74 \text{ m} \) (Madariaga, 1976) and, if we use Bakun’s (1984) moment-magnitude relation for small earthquakes, \( \log M_0 = 1.2M_t + 17 \); we estimate for a magnitude 2 event a stress drop of about 28 bars \( \approx (7/16)M_0/r \) (Keiatis-Borok, 1957). This stress drop is well within the scatter of directly comparable borehole observations of earthquakes of this size (Abercrombie and Leary, 1993). For comparison with other Loma Prieta aftershock analyses, note that Hough et al. (1991), using an empirical Green’s function spectral analysis, found higher corner frequencies, by about a factor of 2, for other events of similar size. Fletcher and Boatwright (1991) found corner frequencies varying by a factor of 2 for small Loma Prieta aftershocks, including a corner frequency of 11 Hz for a magnitude 2 event.

Our corner-frequency estimate for the microearthquakes suggests that the spectral shape of the site response of COY is essentially flat up to 7 Hz. We will make a more systematic comparison of our results later, but we note that the site response of COY has also been analyzed by Hartzell (1992, Fig. 3). Hartzell obtained a spectral ratio that is relatively flat between 0.5 and 7 Hz and has a broad step starting at about 7 Hz that represents an increase of a factor of 3 in amplitude. Figure 4 shows how his result (solid line) compares with the spectral shape we found for frequencies from 2 to 11 Hz. The difference above 11 Hz is presumably due to the source spectral shape of the four small events differing from the ratios of high-frequency site response that Hartzell reported.

Direct S-Wave and Coda Amplification

The direct S-wave and coda segments produce spectral ratios with similar peaks and troughs, but there is a systematic difference in the amplitude between the average S wave and the coda ratios (Fig. 5). This difference is substantial in some cases and minimal in others. For three sites in Fremont and Newark (FLB, IVE, and MDC), the difference is a roughly constant factor of 2 to 4 for the entire band of frequencies analyzed. For eight stations (ACQ, BBP, CNT, DAS, DAW, DUV, ESC, and RAV), the difference varies up to a factor of 4, but is a factor of 2 on average, and generally increases at low frequency (<2 Hz). For the three sites outside the alluvial basin (AP7, SRL, and STQ), the differences between the two ratios are small, although the coda ratios are upper bounds for the S-wave ratios.

The error bars in Figure 5 represent \( \pm \) the standard error of the mean spectral ratios: they underestimate the ensemble standard deviations by the factor \((n - 1)^{1/2}\), where \( n \) is the number of earthquakes recorded by the station. The error bars are largest for stations CNT and MDC, which recorded the fewest events. To illustrate the relation between these error bars and the ensemble of spectral ratios, in Figure 6 we have plotted the entire set of S-wave and coda ratios for the station pair IVE and COY. The shaded areas approximate the 68% confidence intervals for these ensembles: these confidence intervals were estimated by using the generalized inversion on the same set of spectra and multiplying the standard errors of the mean spectral ratios by the factor \((n - 1)^{1/2}\).

Figure 6 shows that the assumption that the noise spectra are greater than or equal to half the signal spectra slightly overestimates the variability of the individual spectral ratios. Thus, our estimates of the spectral noise and the derived uncertainties for the relative amplifications are conservative. The scatter of the S-wave and coda ratios appears similar for frequencies from 2 to 7 Hz. The scatter among the S-wave ratios is greater at low frequencies, possibly due to the instrument noise associated with the turn-on transient, while the scatter among the coda ratios is greater at high frequencies, presumably due to the lower coda signal at high frequencies.

![Figure 4. Stack of S-wave displacement spectra at COY for events of magnitude 1.8 < M < 2.1. Shaded area is the logarithmic mean of the normalized spectral amplitudes plus or minus one standard deviation. An empirical Green’s function spectral analysis indicates that the spectral shape for these small earthquakes below 10 Hz is determined by the response of COY. The average spectrum is roughly flat between 1 and 7 Hz and has a peak around 9 Hz, attributed to a site resonance. For frequencies below 1 Hz, the signal-to-noise ratio for these small events is almost 1. The spectrum is vertically shifted for the following comparison. The solid black curve represents the site response for COY determined by Hartzell (1992). The two shapes of the curves agree for frequencies from 1.5 to 11 Hz. The difference at higher frequencies is probably due to the difference between the spectral shape of the microearthquake records above the corner frequency and the relative site amplifications computed by Hartzell (1992).](image-url)
Figure 5. Average spectral ratios for each site relative to COY. Vertical bars indicate plus or minus one standard deviation for estimates of spectral means. The results of the S-wave window (dashed curves) and coda window (solid curves) for all the stations are shown. The two different analysis windows have similar peaks and troughs, but the factors of amplification relative to the reference station are systematically larger for the coda ratios. The labels SC I, SC II, SC III, and SC IV indicate the site classes (hard rock, soft rock, alluvium, and bay mud) for the stations, discussed later. The shaded curves in the plots for AP7, DUV, SRL, and STQ show Hartzell's (1992) results. The dotted curve in the RAV plot is the site response model of Joyner et al. (1976).
Comparison of Coda and S-Wave Spectral Ratios as Estimates of Site Response in Southern San Francisco Bay Area

Site response relative to COY

Figure 6. Coda and S-wave spectral ratios of station IVE relative to COY for the individual events recorded. The scatter at low frequencies is greater for the S waves because of instrument noise, while the scatter at high frequencies is greater for the coda because of the attenuation of the coda. The shaded areas show estimates of the 68% confidence intervals for the distributions of individual spectral ratios, obtained by multiplying the standard errors from a generalized inversion of these recordings by the factor \((n - 1)^{1/2}\).

Figure 7. Spectral ratios relative to COY for the station IVE, event 3571120. The ratios for three different segments on the seismograms: the lower one is for the S-wave window, the middle one is for the succeeding 5.12 sec of the seismogram (early coda), and the third is for the coda window we used in our study. The amplification increases as the analysis window moves toward the coda window, which starts at a lapse time more than twice the S-wave arrival.

Figure 8. Average spectral ratio relative to COY at the station MDC for a 5.12-sec window and a 10.24-sec window on the direct S wave. A doubling in the size of the analysis windows leads to a near doubling in the amplification factor over the whole bandwidth, suggesting that particular care should be taken in comparing results that come from different seismogram segments.
Station IVE has a large difference between the direct $S$ wave and the coda ratios, which we now examine in some detail. The direct $S$ waves have very similar arrival times at IVE and COY. The spectral ratios increase while moving the analysis window along the trace from the $S$-wave arrival to the coda window used in our analysis. We repeated this observation for three events. In Figure 7, we show three windows for the event 3571120: the direct $S$-wave window, a window starting 5.12 sec later (early coda), and the coda window. Different signal windows in the early $S$-wave coda can give amplifications between the direct $S$ wave and the coda amplifications. This increase in relative amplitudes is apparently due to the build up of energy trapped in the surface layers under the basin sites, energy that is less significant for the relative direct-wave response.

This difference suggests that care must be taken in selecting a window size when calculating direct $S$-wave amplification: longer windows could give substantially larger amplifications as more trapped energy would apparently be included in the spectral window for the basin site, but no such energy would be included for the rock site. As an example, in Figure 8 we show average $S$-wave spectral ratios MDC/COY for 5.12- and 10.24-sec windows. This is an extreme case; the effect of doubling the window length for the other stations is not as large, but the amplifications generally increase if the window lengths are increased. The effect at MDC is possibly the result of its location near a hard-rock outcrop, where one might expect to see surface waves, converted from body waves at the rock-basin edge, earlier in the seismogram. Because of its limited resolution, the geologic map in Figure 10 does not show the southern extension of the Coyote Hills ridge that outcrops just west of MDC.

An important difference between the direct $S$ waves and the coda waves is that the direct $S$ waves are more likely to be affected by anomalous wave propagation, causing a significant dependence of direct-wave site-response estimates on source location. While this was generally not a problem for our stations, one station did manifest this effect. The $S$ waves recorded at station SRL from aftershocks in the Loma Prieta epicentral area exhibited a different amplification than $S$ waves recorded from events in the East Bay and in the southern portion of the San Andreas. Average spectral ratios relative to COY for the two sets of events are shown in Figure 9. The ratios for aftershocks located in the mainshock epicentral area have a peak around 3 Hz that is missing from the other ratios.

In the seismograms recorded at SRL, the events that contain this higher energy around 3 Hz have about twice as much energy on the E–W component than on the N–S component. The assumption that the propagation path to SRL is equivalent to the path to COY is apparently not valid. In fact, the $S$-wave amplification at 3 Hz, calculated by averaging all the events, is slightly larger than the coda amplification, which showed no dependence on source location. It is interesting to note that this variability in propagation characteristics, which is often regarded as a problem for using direct $S$ waves to estimate site response, is clearly evident in this data set only for a station sited next to the San Andreas fault.

### Amplification and Geology

The association of seismic site-response observations with near-surface geology is important for estimating site amplification in areas where seismic data are not available. The surficial geology of the Bay area has been studied in detail and a large amount of geotechnical data has been collected. The main tectonic features are the San Andrees and the Hayward faults. They bound the NW–SE-trending graben in which the San Francisco Bay is located. The schematic map in Figure 10 (Borcherdt and Glassmoyer, 1992) shows the general geological setting of our study area and the location of the sites. The lithologies at our sites (Table 2) represent the wide range of deposits exposed in the area (Helley and Lajoie, 1979). We classified the sites using four categories developed from the experience of several studies in the San Francisco and Los Angeles areas (Fumal, 1978; Borcherdt et al., 1979; Fumal and Tinsley, 1985). We use this classification to distinguish a priori groups of sites that should have similar site responses, and to compare our results to the spectral amplification levels found.
for the different classes by Borcherdt and Glassmoyer (1992) and Borcherdt (1992), who looked at strong-motion data.

The four site classes are defined on the basis of three criteria: geological description, shear-wave velocity, and minimum thickness of the soil units. The classification criteria are shown in Table 3 (from Borcherdt, 1992). We applied the classification using the geologic description and properties reported in Table 2. Geotechnical information on the sites is generally lacking: AP7 was the only site where the shear-wave velocity was measured directly (Gibbs et al., 1976). We classified our sites as follows: site class (SC) (hard rock)—COY; SC II (soft rock)—ACQ, AP7, SRL, STQ; SC III (alluvium)—CNT, DAS, DAW, DUV, ESC, FLB, IVE, MDC; SC IV (bay mud)—BBP, RAV.

All the sites considered in this study are amplified at low frequencies with respect to the site COY, which is the only hard-rock site. The band of amplification goes almost to 7 Hz. This frequency corresponds to the beginning of the inferred resonance frequency band at COY; for higher frequencies the ratios are generally below one. This transition is consistent with the interpretation as a transition from amplification controlled by the impedance contrasts of near-surface layers, to relatively high attenuation at higher frequencies for softer sites. Plots of the average spectral ratio of each class, both for the S-wave window and for the coda window, are presented in Figure 11. These average ratios do not include two sites, ACQ and MDC, that clearly show anomalous behavior compared to other sites in their class. Site ACQ, from its geological description, fits in site class II, but it is a few tens of meters from the edge of a vertical escarpment. This may account for the anomalous ratio that shows an amplification factor of almost 30 at 1.8 Hz (see Fig. 5) in both the S-wave and the coda windows. Station MDC has very low spectral ratios for S waves compared to the other class III sites, showing no effective amplification. The coda amplification factor is significantly lower than other sites in class III for frequencies below 3 Hz. The few recordings at this site makes interpretation difficult. We note again that this site is nearer a hard-rock outcrop than our other class III sites.

S-wave ratios for different classes show slightly different peaks below 7 Hz, being nearly flat for the soft bedrock sites and presenting different bands of amplification for alluvial and mud sites. There is no systematic trend, and the class II and IV S-wave ratios are quite similar. The coda ratios have average values that decrease with increasing soil stiffness; the difference between S-wave and coda spectral ratios is greater for mud

Figure 10. Simplified geologic map of the south San Francisco Bay area (Borcherdt and Glassmoyer, 1992) and location of the sites. Lithologic units used in the description of our site are shown. Qm = Holocene estuarine mud, Qal = Quaternary alluvium, QTs = Quaternary and Tertiary sedimentary rocks, TMzs = Tertiary and Mesozoic sedimentary rocks, and KJf = Franciscan Formation.
Table 2.
Site Location and Geologic Description, Shear-Wave Velocity Is Given Where Available

<table>
<thead>
<tr>
<th>Site</th>
<th>Locations</th>
<th>Geological Description and Soil Properties</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACQ</td>
<td>Alameda Creek, Quarry Lakes Dr., Fremont</td>
<td>Holocene gravel (Qal), near a 10- to 20-m cliff in a gravel quarry with the bottom covered by shallow water. Sensors buried in soil.</td>
</tr>
<tr>
<td>AP7</td>
<td>San Mateo County</td>
<td>Tertiary sedimentary rocks (TMzs): Butano sandstone. Firm to soft sandstone and shale. About 500 m from the San Andreas fault. Shear-wave velocity to a depth of 30 m of 435 m/sec (Gibbs et al., 1976). Sensors buried in soil.</td>
</tr>
<tr>
<td>CNT</td>
<td>Canterbury Ct., Fremont</td>
<td>Holocene coarse medium-grained alluvial deposits (Qal). Sensors buried in soil.</td>
</tr>
<tr>
<td>COY</td>
<td>Coyote Hills, R.P.</td>
<td>Franciscan formation (KJf). The shear-wave velocity to a depth of 30 m found for this formation is of 830 ± 350 m/sec (Fumal, personal comm.). Sensors buried in soil (the soil layer is about 2.5-m thick).</td>
</tr>
<tr>
<td>DAS, DAW, DUV</td>
<td>Palo Alto</td>
<td>Holocene alluvial medium-fine-grained deposits (Qal). The maximum distance between the instruments is about 500 m. The stations were on concrete pads in buildings or nearby them.</td>
</tr>
<tr>
<td>ESC</td>
<td>Newark</td>
<td>Holocene medium-grained alluvial deposits (Qal). Sensors buried in soil.</td>
</tr>
<tr>
<td>FLB</td>
<td>Fremont M. Library</td>
<td>Pleistocene alluvial deposits (Qal). On a pressure ridge of the Hayward fault that separates a deeper basin on the west from a smaller and shallower one on the east side (Kotlarmann and Gorelick, 1992). Shear-wave velocity measured nearby on the same formation has value of 289 m/sec (Fumal, personal comm.). Sensors on a concrete floor of a multistory building.</td>
</tr>
<tr>
<td>IVE</td>
<td>Fremont</td>
<td>Pleistocene alluvium (Qal). Shear-wave velocity measured nearby on the same formation has value of 289 m/sec (Fumal, personal comm.). Sensor buried in soil.</td>
</tr>
<tr>
<td>MDC</td>
<td>Newark</td>
<td>Holocene fine-grained alluvium (Qal) near the edge of the bay, but a few hundred meters from a bedrock exposure, the bedrock is about 127-m deep at the site (Hazlewood, 1976) and the contact with the alluvium is steeply sloping east. Sensors on a concrete floor.</td>
</tr>
<tr>
<td>RAV</td>
<td>Ravenswood, M. Park</td>
<td>Holocene water-saturated bay mud (Qm). Thickness of the mud strata, 11 m (Joyner et al., 1976). Sensors on a concrete pad.</td>
</tr>
<tr>
<td>SRL</td>
<td>Portola Valley</td>
<td>Santa Clara formation (QTs): conglomerate and sandstone, firm to soft; about 250 m from the San Andreas fault. Shear-wave velocity to a depth of 30 m of 440 ± 50 m/sec (Fumal, personal comm.). Sensors buried in soil.</td>
</tr>
<tr>
<td>STQ</td>
<td>Stanford Quad.</td>
<td>Pleistocene medium-fine-grained alluvium (Qal). Thickness of the alluvium, about 4 m (Pampeyan, 1970) overlaying the Santa Clara formation. Station on the concrete floor in the basement of a two-story building.</td>
</tr>
</tbody>
</table>

Table 3
Site Class Descriptions, from Borcherdt (1992)

<table>
<thead>
<tr>
<th>Site Class (SC) Names</th>
<th>General Description</th>
<th>Shear Velocity Mean to 30 m (m/s)</th>
<th>Minimum Thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SC I     FIRM AND HARD ROCKS hard rocks firm to hard rocks e.g., granite; Igneous rocks; conglomerate; hard sandstone, shale</td>
<td>1400 700</td>
<td>1400 10</td>
<td></td>
</tr>
<tr>
<td>SC II    GRAVELLY SOILS; SOFT TO FIRM ROCKS e.g., Igneous sedimentary soft rock; soft sandstone, shale; gravels (all degrees of coarseness)</td>
<td>375 700</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>SC III   STIFF CLAYS AND SANDY SOILS e.g., sand; silt loam and sandy clay; stiff clay and silty clay</td>
<td>200 375</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>SC IV    SOFT SOILS e.g., loose submerged fill; soft clay and silty clay; liquefiable soils, quick and highly sensitive clays, peats, very high plasticity clays, and soft soils</td>
<td>200</td>
<td>3</td>
<td></td>
</tr>
</tbody>
</table>
than for alluvial sites and tends to disappear for the soft bedrock sites.

Comparisons with Other Studies

First, we compare our results to Hartzell (1992), who also used aftershocks of the Loma Prieta earthquake, and examined five of the stations we used: AP7, COY, DUV, SRL, and STQ. Given the similar data sets, any differences between his results and ours must be due to different analysis methodologies. He evaluated site response using Andrews's (1982, 1986) generalized spectral ratio technique on "the S-wave and its larger amplitude coda waves." The average of two rock sites not included in our study defined his reference spectrum. He evaluated the relative site response of COY, so to make a direct comparison with our ratios we divide the relative site responses he found for the sites we studied by the response he estimated for COY. We show in Figure 5 (shadowed lines) the comparison of our results and Hartzell's results at the common sites. For all the sites, our results are consistent with his, especially for frequencies greater than 1 Hz. For three sites, our ratios are higher below 1 Hz. This discrepancy is found also in the otherwise favorable comparison that Hartzell made with the ratios for AP7 and SRL presented by Fletcher and Boatwright (1991). The discrepancy may be related to site characteristics of Hartzell's reference sites, or to differences in the treatment of low-frequency noise.

Joyner et al. (1976) studied site response relative to COY are similar to their model spectral ratio (Fig. 5, dotted line): the first three peaks at frequencies around 0.6, 1.2, and 2.2 Hz are similarly located. Their model fits our S-wave ratio best for frequencies above 0.8 Hz; the maximum amplification factor, around 8, is the same as we find at the same frequency in the S-wave ratio (Joyner et al., 1976, Fig. 12). The agreement between their model spectral ratio and our observations is consistent with our interpretation of COY's site response as approximately flat for frequencies below 7 Hz, and suggests that, below 10 Hz, the relative S-wave motions at COY were comparable to the bedrock input motions used in Joyner et al.'s modeling.

Finally, we compare the values of site amplifications we found with the average strong-motion ratios reported in Borcherdt (1992) for different site classes in the 0.5C- to 2.5-Hz frequency band. His amplification factors refer to an "average hard-rock site." To compare his results with ours, we used the relative amplification factor at the site AP7, which was common to the two studies. Averaging the S-wave and coda amplification of AP7/COY over the frequency band from 0.5 to 2.5 Hz, we obtain the factor 4.4. His corresponding estimate is

![Site response relative to COY](image)

Figure 11. Average spectral ratios relative to COY and their standard deviation for the three classes of sites. The results from the S-wave analysis (dashed curves) and the coda analysis (solid curves) are shown. Coda ratios decrease as the soil classes stiffen, but there is no systematic trend to the S-wave ratios. The difference in results for S waves and codas is significant at low frequency (0.5 to 2.0 Hz) for the alluvial and mud sites (class III and IV), suggesting the possibility of waves trapped in the alluvial basin. The difference becomes slightly bigger for class IV, which includes softer sites. The horizontal dotted lines from 0.5 to 2.5 Hz represent the amplifications, corrected for different reference spectra, predicted for each class by Borcherdt (1992) using strong-motion data. They are consistent with the amplifications we found using the coda windows.
1.41 (AP7/hard rock). We use the ratio 4.4/1.41 = 3.1 (hard rock/COY) to scale his estimates of amplification to correct for the difference in reference spectrum. This means that our reference site, COY, in this band of frequencies has an amplitude three times smaller than the average reference rock site used by Borcherdt and Glassmoyer (1992). COY also has a smaller amplitude than Hartzell's reference spectrum at low frequency; the value of his estimated relative site response is approximately one-half. Borcherdt's scaled average amplification factors are plotted in Figure 11 as horizontal dotted lines from 0.5 to 2.5 Hz; they are consistent with the amplifications we found using the coda windows. His spectral ratios are taken for 40.96-sec windows (Borcherdt and Glassmoyer, 1992), which include essentially the whole strong-motion record. Our results indicate that Borcherdt's definition of site response for this frequency band is controlled by the phases that determine coda amplitudes and implicitly includes the increased duration of shaking resulting from waves trapped in the alluvial basins.

Discussion

It has commonly been assumed that the relative amplification derived from direct S-wave and coda spectral ratios is consistent (e.g., Phillips and Aki, 1986). The often-cited evidence for this assumption is the study by Tsujiura (1978) of data from a subkilometer scale four-station array on rock. He found no significant differences in amplitude or shape for S-wave and coda spectral ratios. The largest spectral ratios he was able to compare had values near 3. He also reported a coda amplification factor as large as 20 for a rock-soil pair, but he did not compare this observation to an amplification factor for direct-wave data. Recently Del Pezzo et al. (1993) showed that the two estimates are coincident for frequencies of 3 and 6 Hz at sites located in an area underlain by recent volcanic deposits. These observations do not differ from our observations in that frequency band for soft-rock sites, which show comparable average amplification for the S-waves and codas.

Gutierrez and Singh (1992) reported that the amplification of S-waves and "codas" were very similar in their consideration of four stations located on soft sediments in the city of Acapulco and a granitic site located at distances of 5 to 15 km from the sediment sites. Their method was, however, quite different from, and inconsistent with, the underlying physical arguments for a conventional coda analysis. Their coda windows began immediately after their S-wave windows, at varying lapse times, and had durations that varied with site. In particular, the coda window for their reference site ended significantly before a lapse time of twice the S-wave travel time, 2T_s, but for the sediment sites 2T_s falls in the middle of coda windows (Gutierrez and Singh, 1992, Fig. 3).

On the other hand, Phillips and Aki (1986) noticed that the amplitudes at their Hollister sediment sites increased relative to their rock site with lapse time, an indication of a discrepancy between relative S-wave and coda amplitudes. They also reported high coda site amplification measurements at low frequencies are associated with sediments in other areas. Similarly, we observe at our alluvial and mud sites that the difference between S-wave and coda ratios at low frequencies is a factor of 2 to 3. Olsen et al. (1994) noted that while direct teleseismic waves of comparable amplitudes were observed in the Salt Lake Basin and in the mountains nearby, the amplitudes following the direct arrival were considerably larger in the basin. Their modeling suggested the effect was due to "reverberations in the near-surface unconsolidated sediments." Frankel (1994) also reported that for seismograms bandpassed between 1 and 2 Hz, the S-wave amplitudes for rock and basin sites in the San Bernardino area are comparable, but the coda amplitudes for the basin site were considerably larger.

Our observations of differences between S-wave and coda amplification factors imply that it is not always valid to assume that coda amplitudes can be used to infer direct shear-wave amplitudes. Inferences of physical phenomena could be affected. For example, discrepancies between S-wave and coda-wave amplifications as large as our larger observations (e.g., Fig. 5, MDC) would be adequate to explain all of Chin and Aki's (1991) reported disagreements between modeled ground motions and observed motions for sites relatively near the Loma Prieta mainshock epicenter. One would not then conclude that pervasive nonlinearity of soil response occurred. This comment does not address the systematic distance dependence of their results, nor possible systematic trends in the actual site amplification factors. It is worth recalling, however, that Hartzell (1992), using site-specific data, suggested that Chin and Aki's estimates of site amplification factors are too large over a significant frequency band for a pair of alluvial sites where nonlinear effects were inferred: Coralitos and Halls Valley. The difference appears adequate to remove the possibility of nonlinearity. These sites are not both basin sites, so the difference between S-wave and coda wave amplifications due to trapped energy may not be the mechanism responsible.

The observed differences between the S-wave and coda spectral ratios appear to be the result of seismic energy trapped in alluvial basins (see also Phillips and Aki, 1986). This possibility implies that the choice of the window length used for S-wave spectral ratios can influence estimates of the relative amplification; longer windows include more of the phases that contribute to the relative increase of the coda spectral ratios. We have
explicitly demonstrated this effect at stations IVE and MDC.

This effect could have biased the comparison made by Jarpe et al. (1989) of strong- and weak-motion site response at Treasure Island. Their estimates of relative amplification from aftershock data were significantly higher than their estimates from Loma Prieta mainshock recordings, especially at low frequencies. Because of the clear evidence of liquefaction on the mainshock recording, their spectral analysis of the mainshock included only the first 5 sec following the S-wave arrival, while they used 15-sec windows for the aftershock recordings. This difference in window length might have contributed to an overestimate of the nonlinear response of the soil before the onset of liquefaction at Treasure Island. We note that Chin and Aki (1991) found no evidence for nonlinear soil response at Treasure Island.

Conclusions

Relative site amplifications estimated from S waves and codas are not always the same. We found systematic differences between direct S-wave and coda amplifications that are correlated with the geological settings of the sites: these differences are most pronounced at low frequencies (0.5 to 2.0 Hz) and increase with decreasing soil stiffness. Among the sites on soft soils, there is some variability in the frequency range where the amplifications differ. Our observations are consistent with the hypothesis that the wave energy trapped by alluvial basins elevates coda amplitudes at soil sites relative to rock sites.

After correcting for differences in the reference site, our coda amplifications for the 0.5 to 2.5 frequency band are nearly identical to the amplification factors obtained by Borcherdt and Glassmoyer (1992), who analyzed 40 sec of the strong-motion accelerograms. This agreement between analyses of strong and weak motion strongly suggests that the soil response at the stations they analyzed was linear for the strain levels occurring during the Loma Prieta mainshock.

The differences in relative site responses estimated from the two techniques we have discussed vary significantly between soil classes; S-wave amplifications were poorly correlated with the geologic settings. Because coda amplifications are strongly correlated with the amplification factors derived from whole record spectra, they appear to be more useful for engineering purposes. Further, the coda amplifications generally yield upper bounds for the S-wave amplifications on the frequency band (0.5 to 10 Hz) analyzed in this article.

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