Seismic Wave-Field Observations at a Dense, Small-Aperture Array Located on a Landslide in the Santa Cruz Mountains, California

by Zhengyu Xu, Susan Y. Schwartz, and Thorne Lay

Abstract  A rectangular (4 by 5) array of short-period three-component seismometers with 15-m spacing was deployed to record several U.S. Geological Survey calibration explosions detonated around the Santa Cruz Mountains. The array was located at a site where an earlier station had recorded frequency-dependent polarized site resonances for aftershocks of the 1989 Loma Prieta earthquake. The site is on a hillside believed to be a landslide structure, with the near surface consisting of poorly sorted sediments and weathered rocks with dipping subsurface layers. The primary objective was to explore the site effects in this complex three-dimensional soft-rock environment, characteristic of much of the Loma Prieta source region. The direct P waves from four nearby (15 to 20 km) explosions at easterly azimuths from the array show counterclockwise arrival azimuth anomalies of 30° to 50°. These deflections are attributed to the presence of more than one dipping velocity contrast beneath the array, with dips of from 10° to 50° and dip directions generally toward the south. One such boundary may correspond to the landslide slip surface, and the presence of dipping velocity contrasts underlying the site is probably responsible for some of the observed directional site resonance. A slowness vector analysis demonstrates that arrivals early in the P coda have similar azimuthal anomalies, while later scattered arrivals come from many azimuths. Particle motions indicate that the more coherent arrivals in the coda are comprised of scattered P waves and Rayleigh waves, probably associated with scattering from the rough topography in the region. The coda displays greater spatial coherency along the hill strike than down the slope, consistent with a wedge-shaped landslide. The overall wave-field spatial coherence, \( \text{CCC}(f, \Delta x) \), decreases with increasing frequency, \( f \), and spatial offset, \( \Delta x \), and on average can be well represented by \( \text{CCC}(f, \Delta x) = e^{-cf \Delta x} \), with \( c = 0.6 \text{ km}^{-1} \text{ Hz}^{-1} \) for the vertical P wave in the first 1-sec window. This behavior is comparable to that found for previously studied hard-rock locations.

Introduction

High-frequency seismic wave propagation in the crust tends to be very complicated, and it is useful to analyze recordings from dense, small-aperture seismic arrays (station spacing from tens to hundreds of meters) to quantify the source and propagation effects that are relevant to earthquake engineering (e.g., Menke et al., 1990; Vernon et al., 1991; Frankel et al., 1991; Hartzell et al., 1994; Mori et al., 1994). Array recordings allow the deterministic features of the short-period wave field to be characterized, including the slowness vector and polarization of discrete plane-wave phases that sweep across the array (e.g., Dainty and Toksöz, 1990; Gupta et al., 1990; Bannister et al., 1990; Wagner and Owens, 1993). Localized scattering effects within the array must be treated differently than remote scattering effects, but the scatterers can still be analyzed deterministically (e.g., Hedlin et al., 1991). Identification of the coherent arrivals can constrain the large-scale structures and scattering heterogeneities in the crust between the source and the receivers and in the vicinity of the array. Seismic wave coda generally becomes progressively more complex with time, with many interfering arrivals scattered from different directions. Even for small-aperture arrays, there is a transition to an incoherent wave field in which individual arrivals cannot be reliably tracked across the array because of interference from multiple arrivals. Statistical measures of the wave field such as spatial coherence are then used to characterize the overall signal complexity and associated scattering properties of the medium, recognizing that the deterministic substructure of the coda is simply not fully resolvable.

The transition from deterministic to incoherent wave field is a fuzzy one, influenced by the signal frequency content, the limitations of the array response, and the local scat-
tering properties in the crust. For isolated three-component stations, little can be done other than to monitor the signal polarization as a function of time using the event backazimuth as a reference frame. When an array is located in a homogeneous hard-rock region, the short-period ($T < 1$ sec) signal coherence drops off significantly over station spacing of just tens of meters (e.g., Menke et al., 1990; Vernon et al., 1991), presumably due to intense scattering from shallow heterogeneity. Even more complex behavior is expected in heterogeneous soft-rock environments. For example, sediment-filled valleys cause frequency-dependent amplifications relative to hard-rock sites (e.g., King and Tucker, 1984; Tucker and King, 1984; Tucker et al., 1984), while rough-surface topography causes short-period signal amplifications (e.g., Geli et al., 1988; Hartzell et al., 1994; Pederson et al., 1994). Many earthquake engineering issues are associated with complex soft-rock sites in tectonically active regions.

This article considers three-component recordings of four shallow explosions from a dense, small-aperture array located in the Santa Cruz Mountains above the 1989 Loma Prieta earthquake rupture zone. The objective is to study the short-period wave field in a geologically complex region with rough-surface topography and dipping geological structures. This contrasts with earlier deployments of dense small-aperture arrays in geologically simple environments and provides insight into the possible site effects for isolated sensors deployed for aftershock recordings or earthquake engineering purposes.

We initially analyze the deterministic features of the explosion recordings, emphasizing the first arrivals. Many studies of direct $P$-wave backazimuths have revealed biases of up to tens of degrees (e.g., Niazi, 1966; Buchbinder, 1987, 1990; Magotra et al., 1987; Menke et al., 1990; Suteau-Henson, 1990, 1991; Zhang and Langston, 1992). These deflections are commonly attributed to lateral velocity gradients (e.g., Buchbinder, 1987), dipping surface geometry (e.g., Buchbinder and Haddon, 1990), or dipping velocity contrasts beneath the receiver (Niazi, 1966; Cassidy, 1992; Langston, 1977, 1979; Zhang and Langston, 1992). Unusually large backazimuth anomalies are found in our case, which we attribute to dipping subsurface boundaries. We extend the analysis of arrival direction to the $P$ coda using a plane-wave slowness vector procedure (analogous to frequency-wavenumber analysis), along with consideration of the polarization. The local crustal heterogeneity appears to cause significant scattering; longitudinally polarized arrivals impinge on the array from many azimuths.

Statistical characteristics of the coda are also considered for comparison with previous work using similar instrumentation and station spacing on hard-rock sites (Menke et al., 1990; Vernon et al., 1991). The influence of the deterministically constrained larger-scale structure on the incoherent signal is examined, along with a consideration of the likely causes of the preferred $S$-wave polarizations observed in the region. The latter issue will be discussed at greater length in another study (Bonamassa and Vidale, in preparation). The general conclusion is that shallow structure beneath the site exerts a profound affect on the short-period wave field that would be almost unrecognizable without the advantage of a dense array. This suggests caution in the analysis of isolated station signals for aftershock studies and earthquake engineering applications.

**The Zayante Dense Array**

The Santa Cruz Mountains are a tectonically active, geologically complex region with rough-surface topography, contorted sedimentary structures, and relatively low-velocity near-surface rock types. This environment offers many earthquake engineering challenges, including the variability of ground-motion amplification observed for hilltops (Geli et al., 1988; Hartzell et al., 1994) and in sediment-filled valleys (King and Tucker, 1984; Tucker and King, 1984; Tucker et al., 1984). Among the many interesting site-response characteristics observed in recordings of aftershocks of the 1989 Loma Prieta earthquake, frequency-dependent preferred resonance directions, independent of earthquake location and focal mechanism, were observed in ground shaking at several locations (Bonamassa and Vidale, 1991; Bonamassa et al., 1991). These resonances tend to be spatially highly variable, based on observations from nested tripartite mini-arrays with station spacing of 25 and 300 m.

One of the sites that displayed strong frequency-dependent preferred $S$-wave polarizations was selected as a study area (Fig. 1). The earlier ZAYA mini-array deployed in this region (Bonamassa and Vidale, 1991) exhibited directional resonances at four of six stations, with general consistency in frequency-dependent behavior (Fig. 1c), but the other two nearby stations showed no resonances, suggesting very shallow control on the ground motions. To improve our understanding of such site effects, we deployed a small-aperture array, which we call the Zayante array, near the location of the Z5 element of the ZAYA array. The Z5 recordings displayed westerly to northwesterly directions of preferred $S$-wave vibration, loosely paralleling the trend of a nearby ridge.

The Zayante array consisted of a 4 by 5 grid of short-period Mark Products L-22 three-component geophones with a grid spacing of 15 m (Fig. 1d). The signals from the 2-Hz seismometers were sampled at 200 samples per second. The performance of these seismometers is quite good in the 2- to 15-Hz range but deteriorates above this (Menke et al., 1991). The signal-to-noise ratio in our data is highest below 15 Hz, but we do consider the higher-frequency information recognizing that it may have some contamination from instrument nonlinearity. The array was situated on a landslide surface (Gerry Weber, personal comm., 1992) with the hillside dipping about 10° toward the south-southwest. The surrounding structure is the San Lorenzo syncline, a Tertiary structure, with the Lambert shale (Brabb and Dibblee, 1979) underlying the soil in the immediate vicinity of the array. The geometry and depth of the landslide slip surface are not
Figure 1. Maps of the ZAYA and Zayante arrays. (a) Base map indicating the location of the study area in the Santa Cruz Mountains of California. (b) Local topography near the San Andreas fault and locations of the ZAYA array and four explosion shot-points for the U.S. Geological Survey calibration shots. (c) Locations of the six elements of the ZAYA array along with frequency-dependent directions of preferred S-wave ground shaking (if any) indicated by the arrows (Bonamassa and Vidale, 1991). The small Zayante array was deployed near element Z5 of the ZAYA array. A nearby ridgetop is indicated. (d) Geometry of the Zayante array, with stations names given by row/column numbering with rows trending east–west and columns, north–south. The 20-element array has 15-m sensor spacing. The numbers in parentheses are the relative elevations of the array elements in meters.

known, but presumably, there is a southerly dip. A shallow refraction survey was subsequently conducted in the array vicinity (Bonamassa et al., 1993), finding that the near-surface layer varies in thickness from 1 to 20 m with a P velocity of about 550 m/sec, while the velocity below this layer is about 1250 m/sec. The quasi-planar velocity discontinuity dips about 20° toward the southwest.

The Zayante array was deployed to record several explosions detonated by the U.S. Geological Survey as part of an effort to calibrate shallow velocity structure in the Santa Cruz Mountains (Eberhart-Phillips and Michael, 1991). Four of the explosions (Table 1; Fig. 1b) were recorded with good signal-to-noise ratio. These events are at distances of 15 to 20 km from the array and have easterly backazimuths varying by about 63°. While some ambient noise time intervals were also recorded, no events at other azimuths were detected during the short deployment of the array. The array was primarily intended to explore variability of local ground motions in the vicinity of the site with polarized resonances and was not designed to include outlying elements that
would have enhanced the resolution of plane waves traversing the array, but the high sample rate and good signal-to-noise quality of the recordings still allow a limited use of the recordings as a dense small-aperture array.

### Table 1

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### Analysis of Direct-Arrival Backazimuth Anomalies

The signals recorded at the Zayante array provide a high-quality data set for four nearby explosions of precisely known location. Some of the waveforms for one of the explosions (event 2) are shown in Figure 2, with the station labeling defined in Figure 1d. Only minor instrument amplification variability is expected among the sensors, but the emplacement sites were all in pits in the sediments no deeper than 1 m, so ground coupling variability is definitely possible. Our analysis uses normalized amplitudes at each sensor, and we focus on relative waveform characteristics rather than absolute amplitudes.

As is the case for all four explosions, the seismograms for event 2 (Fig. 2) have good signal-to-noise ratios, especially on the vertical components, and the first P arrival on each trace is readily identifiable, allowing it to serve as a reference phase. The horizontal components tend to exhibit more waveform variability, but even the vertical components show some waveform changes over 15-m separations (compare traces 21 and 22). The early portion of the waveforms is enriched in higher frequencies. For this event, there is significant signal amplitude for about 9 sec after the P arrival, with longer-period energy arriving about 5 sec after P. It is worth noting that there is slightly greater visual coherence among the arrivals on a given row of the array than between rows, suggesting that local structure under the array varies down the hillside. We will quantify this tendency later.

The bandwidth of the signals is indicated by Figure 3, which shows the average spectra from all vertical-component stations using 9-sec windows for two separate shots, along with the average spectra from an eventless interval of microtremor. The explosion recordings have reasonable signal-to-noise ratios from 3 to 22 Hz, with little signal above 22 Hz, especially later in the recording. The stronger explosions, such as event 2, have good signal-to-noise ratios at frequencies as low as 1 Hz. The array-averaged microtremor spectra are fairly smooth, but individual spectra vary substantially.

We analyze two attributes of the direct P arrivals from each explosion to characterize any backazimuth anomaly. The particle motions of the horizontal components are used to estimate the backazimuth (Fig. 4b). The P-wave polarity measurement is made at each station using the unfiltered recordings, with the backazimuth estimates being averaged across the array for each event. An independent estimate of the backazimuth is provided by the relative arrival times of the P signals. Cross-correlation of a short time window (Fig. 4a) spanning the direct arrival is used to measure the tiny travel-time differences across the array, with the vertical
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components providing the more stable differential times due to their waveform coherences. These differential times are fit with a plane wave to estimate the backazimuth, but the angle of incidence of the wavefront cannot be separately resolved due to the planar array geometry. The relative timing also controls the backazimuths inferred from frequency-wave-number analysis. Results of this analysis (Fig. 4c) confirm the values of backazimuth obtained from fitting a plane wave for shot 2. The nearly linear motion in the first cycle of the P arrival is used to estimate the apparent backazimuth. (c) Broadband f-k analysis of the direct P-wave power spectrum for shot 2 in the window shown in (a). A maximum likelihood method is used, with the spectrum contoured in the wavenumber domain. The peak contour indicates the northeasterly backazimuth. (d) Summary of estimated (open circles) and actual (filled circles) backazimuths for the four shots. The average estimates from the particle motion and f-k methods are shown. Biases of 30° to 50° are found for all four events, with systematic counterclockwise rotations.

Effects of the Dipping Surface on Apparent Backazimuth Estimates

The surface topography at the Zayante array is approximately planar with a 10° dip and a dip direction of S14°W. This dip causes arrival-time variations that bias the backazimuth estimated by f-k or wave-front fitting procedures. By transforming the ray projections from the dipping plane to a horizontal plane, we can correct for the effects of the surface dip, without needing to know the shallow velocity structure. The geometry is such that the predicted bias in backazimuth is clockwise by about 20° to 30° for all four shots. This is opposite to the bias in the observations, indicating that structure beneath the array is responsible for the backazimuth deviations. Before we can determine subsurface structure consistent with our backazimuth observations, the large effects of the free surface must be corrected for.

A dipping surface also affects polarization estimates of backazimuth when the wave front sweeps obliquely across the surface, because the dipping surface defines the boundary controlling the receiver function that partitions amplitude onto the locally defined (i.e., relative to the dipping surface) vertical and horizontal components (Buchbinder and Had- don, 1990). The extent of backazimuth bias varies with the angle of incidence, dip of the surface, and ratio of S velocity to P velocity. We can estimate the angle of incidence from the particle motions, but this is also an apparent angle, influenced by the free-surface interaction. For a horizontal free surface, the relationship of the true incidence angle (relative to the vertical), \( \theta \), to the measured incidence angle, \( \theta' \), is given by

\[
\cos \theta' = 1 - 2V_s/V_p^2 \sin^2 \theta
\]

where \( V_s \) and \( V_p \) are the S-wave and P-wave velocities near the surface, respectively (Bullen, 1963). The velocity terms enter into the calculation through the receiver function. Using (1), it is possible to infer the true incidence angle based on the surface motions. An additional correction is needed to account for the coordinate rotation when the surface dips. For the two cleanest events, the measured values of \( \theta' \) are 11° to 19°. For a Poisson ratio of 0.25 (\( V_p = 1.732 V_s \)), this indicates true subsurface incidence angles that are a few degrees smaller. One can compute the effects of the dipping surface on polarization azimuths for a given incidence angle of the P-wave vector by rotating to the coordinate system orthogonal to the dipping surface, calculating the P-wave motions in that coordinate system, and then projecting the motions to the geographic coordinate system. Unlike the large arrival-time backazimuth bias of 20° to 30° caused by surface topography, the 10° dip of the Zayante array predicts a very small polarization backazimuth bias. For a typical soft-rock ratio of \( V_p/V_s = 2.0 \), we find that the surface effect on polarization azimuths actually goes to zero. Uncertainty

Figure 4. Procedures used to estimate the propagation direction of the initial P wave. (a) Vertical-component seismograms for shot 2, illustrating the time window used for frequency-wavenumber (f-k) analysis. A very short, highly coherent interval is used. (b) Horizontal particle motion for station 11 for shot 2. The nearly linear motion in the first cycle of the P arrival is used to estimate the apparent backazimuth. (c) Broadband f-k analysis of the direct P-wave power spectrum for shot 2 in the window shown in (a). A maximum likelihood method is used, with the spectrum contoured in the wavenumber domain. The peak contour indicates the northeasterly backazimuth. (d) Summary of estimated (open circles) and actual (filled circles) backazimuths for the four shots. The average estimates from the particle motion and f-k methods are shown. Biases of 30° to 50° are found for all four events, with systematic counterclockwise rotations.
in the precise near-surface $V_p/V_s$ ratio will always make it difficult to reliably correct polarizations for surface dip.

**Effects of Dipping Interfaces beneath the Array**

A likely cause of large backazimuth anomalies is the presence of a dipping interface with a strong velocity contrast below the array. Niazi (1966) considered the effects of a dipping Moho on array measurements at the surface and provided expressions for calculating the bias in backazimuth as a function of velocity contrast and dip of the interface. We use his results with minor corrections for inconsistencies in sign convention. Aside from plausibility, there are two lines of evidence to suggest that dipping interfaces below the array are present. The first is the geological structure, with the array lying on a southwestward dipping landslide deposit located in the southwestward dipping flank of a synclinal structure. The second is more direct, with the shallow reflection/refraction survey in the area, indicating that there is an interface separating layers with $P$ velocities of 550 m/sec and 1250 m/sec with a dip angle of 20° and a dip direction of S29°W (Bonamassa et al., 1993). A schematic of the shallow structure under the Zayante array is indicated in Figure 5. For specified true backazimuths and incidence angles, the biasing effect of the shallow dipping interface was calculated and found to cause an azimuthally varying counterclockwise bias for sources that lie eastward from the array. This is illustrated in Figure 6 for a generic dipping discontinuity. The near-surface dipping interface produces counterclockwise backazimuth biases on the order of 10° for the travel-time method. This has the correct sign but is much smaller than the observed biases and, at most, compensates the opposing bias induced by the surface dip. It appears that at least one more dipping interface, probably with a larger velocity contrast, exists below the array, as indicated in Figure 5. We seek simple models with an additional deeper interface that yields backazimuth biases for both travel-time and polarization methods consistent with the observed biases.

**Modeling Results**

Any modeling effort is complicated by our ignorance of the ray-path effects between the sources and receivers, with the ray paths not penetrating very deeply into the heterogeneous crust. We therefore calculated apparent backazimuths using arrival-time and polarity methods for a wide range of incident ray parameters below the modeled structure, assuming all the bias is due to local structure under the array. The limited class of models considered includes the known surface geometry; the geometry and velocities of the shallowest low-velocity layer; and one additional dipping interface below the array, separating material with a $P$ velocity of 1250 m/sec from deeper structure with $P$ velocity varying up to 40°.
around 4300 m/sec. We include corrections for the surface dip and the shallow-dipping interface.

We have nine reliable observations of angles: four apparent backazimuths from the arrival time method, two apparent incidence angles, and three apparent backazimuths from the polarity method. In order to stabilize the modeling, we impose two constraints. First, differences between predicted and observed incidence angles and backazimuths are not allowed to exceed the estimated measurement uncertainties of 3° and 7°, respectively. The second constraint is on the minimum velocity at the turning point of the ray path. This velocity must exceed the average velocity from source to receiver. The average velocity is close to 4 km/sec for shots 1, 2, and 5 (all with epicentral distances near 15 km) and close to 5.7 km/sec for shot 3 at a distance of 21 km.

We make a first-order approximation that the turning point velocity, \( V_t \), is directly proportional to the average velocity, \( V_a \). The higher the turning point velocity, the smaller the ray parameter and the steeper the incidence angle. Variations in \( V_a \) between events are used to scale differences in their incidence angles proportionally.

We perform a grid search on four parameters: wave incidence angles just below the deep interface for the four shots, velocity contrast across the deep interface, dip angle, and dip direction of the deep interface. For the polarization measurements, we must also consider the near-surface \( V_p/V_s \) ratio to correct for the effects of the dipping surface. We calculate the turning-point velocity and the contributions to each measurement from the deep discontinuity, the shallow discontinuity, and the dipping surface. The subsurface discontinuities contribute identically to the travel-time and polarization measurements in that they simply cause refraction of the ray paths. Thus, the key to finding a consistent model for the overall data set is the behavior of the surface contributions. But here we run into a problem; while our observations of backazimuth for the two methods agree, the dipping surface causes a large clockwise azimuth bias for the travel-time method, while it predicts either a counterclockwise or at most a small clockwise bias for the polarization method, depending on the \( V_p/V_s \) ratio. For hard-rock ratios such as \( V_p/V_s = 1.732 \), we obtain the maximum clockwise-valued bias, but it is only about 3° (compared to 20° to 30° for arrival times), thus a unique model compatible with all of the observations is not possible.

We found combinations of model parameters that satisfy the arrival-time backazimuths and incidence angles within their measurement bounds (Table 2), models that satisfy the polarization backazimuths and incidence angles (Table 3), but no models fitting the complete set of observations within their measurement bounds. Both data sets indicate that a deeper interface dips toward the S or SSE with poorly resolved dip ranging from 7° to 51°. A strong velocity contrast across a shallow-dipping (7° to 20°) deeper interface is preferred, as this makes the most sense geologically. The sets of models fitting each data set do not overlap, which may reflect our understanding of the measurement uncertainty or possibly an intrinsic failure of the ray method to account for finite-frequency effects. The fact that the raw backazimuth measurements from polarizations and arrival times are similar is fortuitous, as they actually are expected to differ due to the geometry of the problem. One possible factor is that the free-surface interaction controlling the polarizations involves finite-frequency arrivals that may effectively sense average near-surface properties rather than the very shallowest structure. We have higher confidence in the results based on relative arrival times, as these are insensitive

### Table 2

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<th>Deep Interface Geometry</th>
<th>Dip Angle</th>
<th>Dip Azimuth</th>
<th>Velocity Contrast</th>
<th>Incidence Angle</th>
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*Error is an uncertainty weighted variance of the misfit to the relative arrival time backazimuths and angle of incidence measurements.

### Table 3

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*Error measure is an uncertainty weighted variance of the misfit to the polarization backazimuths and angle of incidence measurements.
to the precise near-surface Poisson ratio. For $V_p/V_s$ ratios greater than 2.1, no model could be reconciled with the polarization-based observations, because the azimuthal pattern of backazimuth biases could not be matched. Results of the refraction experiment (Bonamassa et al., 1993) indicate $V_p/V_s$ ratios of up to 2.1, and we conclude that the polarization observations may have been contaminated in some manner. The models fitting the arrival-time data attribute large backazimuth anomalies on the order of 50° to 60° to a deeper interface, which may correspond to the slip surface of the landslide that underlies the Zayante array. Of course, several interfaces could actually be involved. While some ambiguity remains, and only simple models have been considered, the robust aspect of the analysis is that the very shallow subsurface effects produce large backazimuth biases that are difficult to predict a priori.

Array Analysis of P Coda

Analysis of secondary arrivals in the seismograms can further reveal the very near receiver effects of the site. First, we attempt to identify coherent secondary arrivals and their trajectories across the array. We do this using a slowness vector analysis (Xie and Lay, 1994) that is basically similar to f-k analysis. The procedure calculates the signal power distribution in the horizontal slowness plane in discrete time intervals. Figure 7 shows the summed power distributions for all three components of motion for shot 2, in 1-sec time windows (see Fig. 2 for time scale). The first window contains the onset of the P arrival, and the peak energy has a slowness vector pointing at an azimuth of 240° (i.e., incident at an azimuth of 60°, which is 40° counterclockwise of the true backazimuth of 100°), which is consistent with the corresponding f-k analysis shown in Figure 4c. The strongest coherent plane-wave energy is in the window from 1 to 2 sec (Fig. 7a) with a northeasterly arrival azimuth; this is the main P-wave motion. For later windows, the peak energy has highly variable propagation directions, although there is some tendency for energy to arrive from the northeast, clearly seen in the individually normalized plots (Fig. 7b). As the lag time increases to about 4 to 5 sec, the energy spreads more uniformly in the slowness plane, indicating increased levels of scattering from various directions. After 4 to 5 sec, there is some stability in the energy distribution over the slowness plane, with relatively well-defined peak arrivals that come from different directions. This corresponds to a transition to somewhat longer-period arrivals in the waveforms (Fig. 2), which presumably involve a larger percentage of scattered surface-wave energy. The patterns for the other events are similar but differ in detail. We can analyze the polarization of the coda to help identify the coda constituents.

The power distributions in the horizontal slowness plane for different particle motion directions for shot 2 are shown in Figure 8. For this analysis, the two horizontal components were projected onto the indicated particle motion azimuths, and the resulting signals were processed using the slowness method. By considering all polarization directions and the associated array slowness analysis, we can consider the correspondence between sense of ground motion and direction of wave propagation, which can help to identify the wave type. Of course, the presence of overlapping phases scattered from different directions can confuse this simple polarization analysis, but our impression is that much of the coda involves scattered, but relatively coherent, surface waves. In the first two time windows, the strongest energy has a polarization of N60°E, arriving from the same azimuth, and there is almost no signal in the orthogonal direction of N150°E. This reflects the nearly linear polarization of the direct P wave, along the biased backazimuth. The next few seconds of coda have little coherent polarized energy, perhaps due to strong near-receiver scattering of the direct P phase that does not produce coherent plane waves across the array. The 4- to 5-sec window shows an energy peak for east-west polarized motion propagating toward the west across the array. This is consistent with a Rayleigh-wave arrival, arriving about 20° off the true backazimuth. Polarization analysis of the full ground motion supports this interpretation. This is probably the onset of the direct surface wave from the source, arriving with a velocity near 2 km/sec, but may also involve P to Rayleigh scattered energy from the nearby topography. Later coda intervals, as well as similar polarization analysis for the other three shots, reveal similar behavior, azimuthally deflected direct P waves with coda that is dominated by P-SV-wave-type arrivals. More sophisticated methods to decompose the coda into distinct particle motion contributions exist (e.g., Wagner and Owens, 1993, 1995), but we do not pursue this given strong variability observed in the event-to-event behavior. This indicates that near-receiver scattering is very sensitive to the source azimuth, and our limited attempts to do any further deterministic imaging of the scattering structure were not encouraging. We turn our attention to the statistical aspects of the data.

Waveform Coherence

Spatial coherence of waveforms can be quantified using the cross-correlation coefficient (CCC) for pairs of seismograms as a function of the receiver spatial offset. The CCC can be computed in either the time or frequency domains. Similar to previous studies (Menke et al., 1990, Vernon et al., 1991), we use a conventional squared coherence measure. In the time domain, the CCC between two waveforms $s(x_1, t)$ and $s(x_2, t)$ at location $x_1$ and $x_2$ is

$$CCC(t, \Delta t, \Delta x) = \langle s(x_1, t) \cdot s(x_2, t) \rangle^2 \langle s(x_1, t) \rangle \cdot \langle s(x_2, t) \rangle$$

where $\langle \rangle$ denotes the average value of the enclosed function over a time interval $\Delta t$ centered on $t$; the offset $\Delta x = |x_1 - x_2|$.
Before calculating the CCC for a pair of waveforms, one must contend with the difficult issue of time shifts. With overlapping signals arriving from many possible azimuths, one cannot assume an appropriate move-out velocity for alignment, yet failing to shift the traces should bias the coherence measure low. For our alignments, we used relative time shifts among the array channels based on broadband cross-correlations of the traces. We select one of the stations near the center of the array as a reference trace. For each array element, a cross-correlation function is computed using a time window of 0.25 sec with sliding steps of 0.2 sec. The time lags for each window for all stations relative to the reference are applied to shift the data to common alignment when interstation correlations are computed. This procedure emphasizes the dominant frequencies in the signals and effectively shifts the signals to optimal coherence for the larger...
Azimuth of Particle Motion

Figure 8. Horizontal slowness vector power distributions for shot 2 (as in Fig. 7) using the horizontal components along different particle motion directions (obtained by rotating the north-south and east-west recordings onto various directions ranging from 0° (N-S) to 150° with 30° intervals. The relative power in all of the panels is normalized to the largest amplitude arrivals. The stronger (darker) arrivals have propagation directions roughly aligned with the particle motion, indicating that these are longitudinally polarized P- or Rayleigh-wave arrivals with backazimuth of 60° to 80°. The true backazimuth is 101°.

Each of our seismograms is filtered into a set of traces using Butterworth zero phase filters with a 5-Hz passband, moving steps of 3 Hz, and a total bandwidth from 3 to 33 Hz. The alignments found for the broadband signals are held fixed for correlation of the narrowband filtered traces, rather than optimizing the correlation in each passband, which could involve many cycle slips relative to the stronger arrivals. For each frequency band, we consider 0.5-sec portions of the waveforms with moving steps of 0.2 sec, aligned by the cross-correlation shifts. We compute the CCC between all record pairs for variable time lengths, \( T_w \). \( T_w \) depends on the center frequency of the band, \( F_c \):

\[
T_w = 0.5 \times \frac{5.5}{F_c} \text{ (sec)},
\]

where 5.5 is the \( F_c \) of the lowest passband of 3 to 8 Hz and...
0.5 is the time segment width for alignment. For example, $T_w(F_c = 5.5 \text{ Hz}) = 0.5 \text{ sec}; T_w(F_c = 30.5 \text{ Hz}) = 0.09 \text{ sec}.$ Thus, for a center frequency of 5.5 Hz, there is one window for calculating the CCC, while for a center frequency of 30.5 Hz, we calculate the average CCC from five windows.

Menke et al. (1990) have considered the possible bias due to frequency-dependent alignment decisions and find that it is relatively small for time windows like those we used, which we confirmed by following their exact procedure for some of our data. Our procedure estimates the frequency-dependent decrease in coherence relative to the statistical behavior. This approach is valid given the weak frequency-dependent behavior we found in the deterministic array analysis of the data, which did not reveal the presence of strong plane-wave arrivals with distinct frequency content and differing arrival directions in each time window.

Variations of Coherence Down the Hillside

We noted earlier that Figure 2 suggests stronger waveform similarity along rows of the array than down the columns. To test this quantitatively, we calculated the average CCC between elements on rows, CCC_row, and between elements on columns, CCC_col. The mean value for 2-sec time windows for the vertical components from all four shots was computed for the 3- to 9-Hz frequency band as a function of the receiver spatial offset. We then computed the fractional difference in coherence:

$$\text{CCC}_{\text{fkd}} = 2 \times \frac{\text{CCC}_{\text{row}} - \text{CCC}_{\text{col}}}{\text{CCC}_{\text{row}} + \text{CCC}_{\text{col}}},$$

which is shown in Figure 9. This effectively measures the anisotropy in waveform coherence, with positive values indicating stronger coherence along rows. For spatial offsets of 15 and 30 m, the CCC_{fkd} is positive, confirming the statistical tendency for receivers at relatively uniform elevations (Fig. 2d) to record more similar waveforms. This is consistent with the presence of the southward-dipping surface and subsurface interfaces discussed in the previous section. A wedge-shaped structure will cause more similar reverberations at stations overlying comparable depths to interfaces. The fact that the array is not perfectly aligned with the structure, and the irregularity of the structure itself, causes the 45-m offsets to be less coherent and to have less spatial pattern. This spatial pattern in the statistical measure indicates that we are again measuring right at the level of transition from deterministic control on the array signals to a stochastic wave field. While supporting the general expectations of the near-surface structural effects, it is difficult to further constrain the structure with such statistical measures.

Figure 9. The difference in signal coherence on vertical components between east–west and north–south sensor offsets averaged for four shots. The difference is normalized by the mean coherence at each offset. Results are shown for four time windows in the signals, and the frequency band used is 3 to 9 Hz.

Variation of Coherence with Frequency, Spatial Offset, and Time Lag

Generally, the spatial coherence across the Zayante array reduces with increasing frequency, sensor offset, and time lag for recordings of all four shots. Average coherence versus spatial offset in various frequency bands and time windows for vertical-component seismograms for the four shots, for microtremor intervals, and for synthetic Gaussian noise are shown in Figure 10. The Gaussian noise has an average coherence value of 0.1 and the coherence is never more than 0.2. Therefore, coherence values for the data of less than 0.2 are not considered significant. The microtremor coherence has an average value of about 0.3, and there is less variance in the 3-sec microtremor window than in the 1-sec window. The coherence is not very dependent on the frequency band or the spatial offset for microtremor and Gaussian noise. The microtremor spectrum is relatively flat (Fig. 3), and the weak coherence suggests that the wave field involves much localized scattering of the signals traversing the array. For the high-frequency energy (>15 Hz) in the four shots, especially in the 3 to 9-sec window, the coherence approaches the level of the microtremor signal. This probably reflects the low energy level (Fig. 3), and the scattered nature of the coda (Fig. 7), which is comparable to the microtremor signal. We infer that coherence greater than about 0.3 reflects the presence of more coherent signals traversing the array than is typical of microseism.

Following Menke et al. (1990), the array coherence can be expressed by a simple empirical equation, $e^{-c \delta x}$, where $c$ is a constant determined from the data. For the first 1-sec window for the vertical-component signals of the four shots $c = 0.6 \text{\ km}^{-1} \text{\ Hz}^{-1}$. This value is similar to the values of...
0.4 to 0.7 km$^{-1}$ Hz$^{-1}$ found for direct arrivals at hard-rock sites (Menke et al., 1990). This may imply that in the frequency range of 2 to 20 Hz, the coherence at wavelengths from tens to hundreds of meters may be controlled by similar heterogeneities below the surface for both soft- and hard-rock environments.

Averaging all $\text{CCC}(t, \Delta t, \Delta x)$ for a given offset, we obtain the average spatial coherence as a function of time, frequency, and spatial offset for the Zayante array (Fig. 11). Results are shown for both the vertical (Z) components and the north–south (N–S) components of motion for shot 2, the average of all shots, and the microtremor windows. The microtremor coherence on vertical and N–S components has similar patterns with time and frequency and defines the background signal level. The first 1 to 2 sec of the shot signals have the highest coherence in all frequency bands, up to frequencies of about 27 Hz. The N–S components are always somewhat less coherent than the Z component, but the explosion signals display a similar general decrease of coherence with time lag, frequency, and offset. The coherence values and trends with time, frequency (in the band 2 to 21 Hz, which has good signal to noise), and offset for the shots are similar to those at the hard-rock sites analyzed by Menke et al. (1990) and Vernon et al. (1991). While the Zayante array is located in a more complex geological environment, it appears that the small-scale near-surface heterogeneity induces comparable signal complexity to that found in other environments.

Discussion and Conclusions

The Zayante array recordings demonstrate the important role of site effects for the short-period seismic wave field. The site of this array is very typical of the geological environment in the Santa Cruz Mountains, and the host of complicated earthquake engineering issues raised by the 1989 Loma Prieta earthquake provide an important context for examining the high-frequency wave-field phenomena in this
Figure 11. Waveform coherence (indicated by graytone shading from black to white, which corresponds to values of 1 to 0.2) versus time and frequency for offsets of 15 and 45 m for various subsets of the data. The results for vertical components (left two columns) and north-south components (right two columns) are shown for just shot 2 (upper row), for the average of four shots (middle row), and for the microtremor window (lower row).

region. The array aperture is too small to provide high resolution of incoming wave fronts, but both P-wave polarization and relative arrival times indicate large backazimuth biases of from 30° to 50°. While we could not fully reconcile both sets of observations with a simple model, it appears that subsurface dipping layers with fairly strong velocity contrasts are responsible for the large azimuthal deflections. In a region like the Santa Cruz Mountains, with very low velocity near-surface sediments, direct seismic waves are steeply incident under the receivers, which enhances their
sensitivity to azimuthal deflection by moderately dipping structures. Combined with strong near-surface lateral-velocity gradients, produced by boulders and heterogeneities in the weathered layer, there is high potential for single-sensor and even small-aperture array measurements to be biased by site effects. We qualitatively account for the observations with a simple three-dimensional velocity model motivated by the shallow seismic refraction survey. The actual structure is undoubtedly more complex, but the integrated effects of the structure appear to be in the correct regime.

Given the strong effect of shallow-dipping interfaces on the direct $P$ wave, it is interesting to consider the effects on secondary arrivals. The receiver structure is such that one expects the entire wave field to be influenced, although variations in the incident angles of secondary arrivals should change the biases as a function of lag time into the coda. For both $P$ and $S$ waves, the effect of subsurface dipping interfaces with positive velocity contrasts is to refract the wave direction toward the normal to the interface, and toward the dip direction of the interface (Fig. 6). If we use the dipping interface to define the $SH$ and $SV$ components of motion, the amplification effect from refraction is most pronounced for the $SV$ component, and the $SV$ polarization is refracted to be more horizontal. The net effect of a dipping interface can thus be to cause some systematic polarization of the $S$-wave field, with amplification of motion in some directions. This will hold for complex wave fields with many $P$- and $S$-wave arrivals incident on the same structure from different azimuths. There will be a net refraction effect by the subsurface interface that can preferentially influence some components of motion.

The observation of preferred direction of $S$-wave shaking at the ZAYA array (Bonamassa et al., 1991) may be in part due to the presence of dipping interfaces under the stations. It is useful to consider a simple geometry. Along the axis of a V-shaped valley structure underlying low-velocity material, as shown in Figure 12, the direction of the $S$-wave ray paths from different azimuths will be refracted so that the horizontal component of shaking tends to align with the axis of the valley. For sites overlying one of the valley flanks, waves incident from the updip side of the valley will be similarly refracted, while waves incident from the down-dip side will refract so that horizontal shaking is along the dip direction. The precise pattern induced is a function of the incident wave parameters and the specific structural geometry. Frequency dependence of the preferred shaking directions is expected for complex layered structures, with finite-frequency effects differing in the extent of refraction and amplification for each passband. Higher frequencies will be more sensitive to very shallow structure, while lower frequencies average the structural effects over a larger depth extent. The development of preferred orientation of shaking is likely to be more pronounced in the coda than in the direct signal due to the fact that the coda involves arrivals from many azimuths, which allows the effect of refraction toward specific directions to accumulate. While simulation with three-dimensional wave propagation is needed to quantify the structural effect (and is being conducted by Bonamassa and Vidale, in preparation), the numerical simulations of Khair et al. (1991) show that preferred directions of ground shaking do develop along the valley axis as described, as a complex function of geometry and incidence angle. While simple dipping structures may not give rise to all of the directional resonances observed by Bonamassa et al. (1991), it is likely that they provide the gross structural controls, with superimposed small-scale heterogeneity. The wedge-like shape of this particular site effect may also contribute reverberation patterns that enhance directional interference.

The other clear manifestation of site effect in the Zayante data is the rapid spatial decay of high-frequency waveform coherence, which proves comparable to that observed at harder-rock sites. While some mild spatial systematics influenced by the gross structure were detected, the main feature is that high-frequency coherence drops off rapidly on the scale of tens of meters. This provides further evidence for caution in interpretation of high-frequency signal in terms of source effects, unless the site effect can be reliably corrected for.

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