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Study of seismic coda generation in crust
with non-uniform scattering and
anelastic properties

John E. Vidale

University of California, Santa Cruz
1156 High Street
Santa Cruz, CA 95064

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JAMES F. LEWKOWICZ
Contract Manager
Solid Earth Geophysics Branch
Earth Sciences Division

JAMES F. LEWKOWICZ
Branch Chief
Solid Earth Geophysics Branch
Earth Sciences Division

DONALD H. ECKHARDT, Director
Earth Sciences Division

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We have conducted a series of investigations to determine which structures affect the amplitude, duration, frequency content, and directional properties of ground motion response to explosions and earthquakes. Our investigations have shown that the shallowest tens of meters are crucial in these properties of the seismic wavefield. These shallow layers are marked by great lateral heterogeneity in velocity and attenuation, and consequently we observe both vertical and lateral resonances. The scrambling of the wavefield each time it reflects from the free surface is undoubtedly very important for seismic propagation to regional distances, but we are only beginning to understand the detail of this interaction.

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Table of Contents

Directional Site Resonances Observed from the 1 October 1987 Whittier Narrows, California Earthquake and the 4 October Aftershock 1

The strong influence of near-surface geology on the direction of ground motion above 1 Hertz in the Loma Prieta earthquake 19

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John E. Vidale
Ornella Bonamassa
Heidi Houston

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Directional Site Resonances Observed from the 1 October 1987 Whittier Narrows, California Earthquake and the 4 October Aftershock

John E. Vidale, Ornella Bonamassa, and Heidi Houston

Abstract

We present evidence that sites often resonate preferentially in a particular compass direction. The 1 October 1987 mainshock and 4 October 1987 aftershock in the Whittier Narrows, California, sequence had very different mechanisms. Nevertheless, at 8 of 11 strong motion stations for which digital records of both events are available, the direction of strongest shaking in the two events was much more similar than would be expected from their different focal mechanisms. The coincidence of the polarizations from the two events was greatest for the frequencies that showed the most amplification, suggesting that site amplification and directional resonances are linked. Knowledge of directional site resonances may aid in predicting the directions of damaging earthquake motions.

Introduction

The 1 October 1987 Whittier Narrows California (M_L = 5.9) earthquake and its 4 October aftershock (M_L = 5.3) created an excellent data set of strong motion recordings. These events produced nearly orthogonal radiation patterns. The mainshock has a thrust mechanism, the aftershock has a strike-slip mechanism. Observations from these two sources permits the isolation of site from source effects. Vidale (1989) showed that the 3 to 6 Hz peak accelerations of these two events are modulated by the focal mechanisms. In that study, only peak amplitudes from analog records were utilized. Since eleven records from the aftershock and numerous records from the main shock have now been digitized by the California Strong Motion Instrumentation Program (CSMIP) of the California Division of Mining and Geology, more quantitative analysis can be done.

Near-receiver geology is an important factor in determining the strength of shaking from an earthquake (Haukkson *et al.*, 1987, Malin *et al.*, 1988). Models of geology that assume homogeneous flat layers can explain some features of observed site amplification (e.g., Joyner *et al.*, 1976). The amplification of 2 sec energy in the lakebed deposits in Mexico City during the 1985 Michoacan earthquake is one of the more dramatic examples of the influence of thin, slow-velocity layers near the Earth's surface (e.g., Campillo *et al.*, 1989).

Patterns of amplification and duration of shaking that require lateral variations in geologic structure have also been documented (Vidale and Helmberger, 1987, 1988), and strong-motion effects of some simple large-scale structures have been investigated (Bard and Gariel, 1986, Kawase and Aki, 1989, Vidale *et al.*, 1985). However, the importance of near-receiver structures more realistic than horizontal layers has not been documented for high-frequency seismic energy. The data analyzed here indicate a need for an increased understanding of the effects of two- and three-dimensional structure near the receiver.

In this paper, to assess the prevalence of frequency- and directionally-dependent station resonances, we compare the particle motions of the S waves with those predicted from the earthquake focal mechanism. It is of interest to earthquake engineers whether particular sites have a preferred direction of ground motion in a given frequency range. Initial results from Loma Prieta aftershock recordings (Bonamassa *et al.*, 1990) suggest that such effects occur for more than half the sites investigated, and that the preferred direction does not depend on earthquake location. The results below indicate that directional resonances are a general feature.

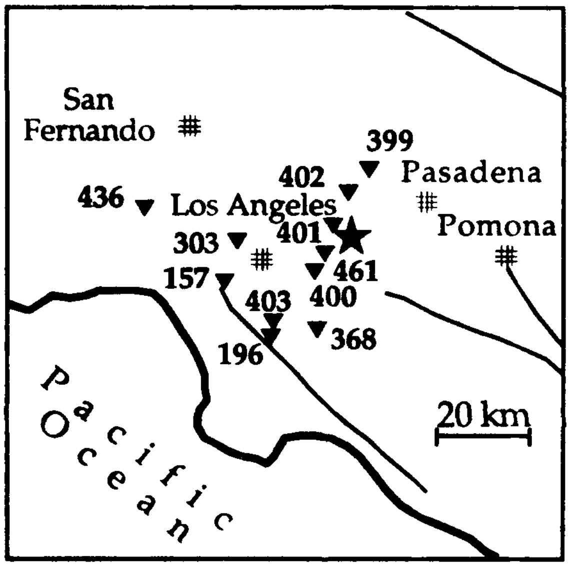

Figure 1. Map showing the location of the 11 CDMG stations. The star indicates the epicentral region of the mainshock and the aftershock, which are separated by only 2 km.

Polarization Analysis for 1 October and 4 October 1987 Whittier Narrows Events

The mainshock hypocenter was located at 14.6 km depth, and the mechanism is a gently dipping thrust (Haukkson and Jones, 1989). Numerous aftershocks filled the volume from 8 to 17 km depth extending about 4 km in all directions horizontally. Bent and Helmberger (1989) analyze the teleseismic body waves and propose a double source; their second source is 11 km deep and 5 times larger than the first with a slightly different mechanism. It is important to consider the location and mechanism of the largest patch of moment release to understand the strong ground motions. The double source they propose is best studied with teleseismic body-waves since the strong ground motions are more complicated by the Los Angeles

Whittier Mainshock and Aftershock

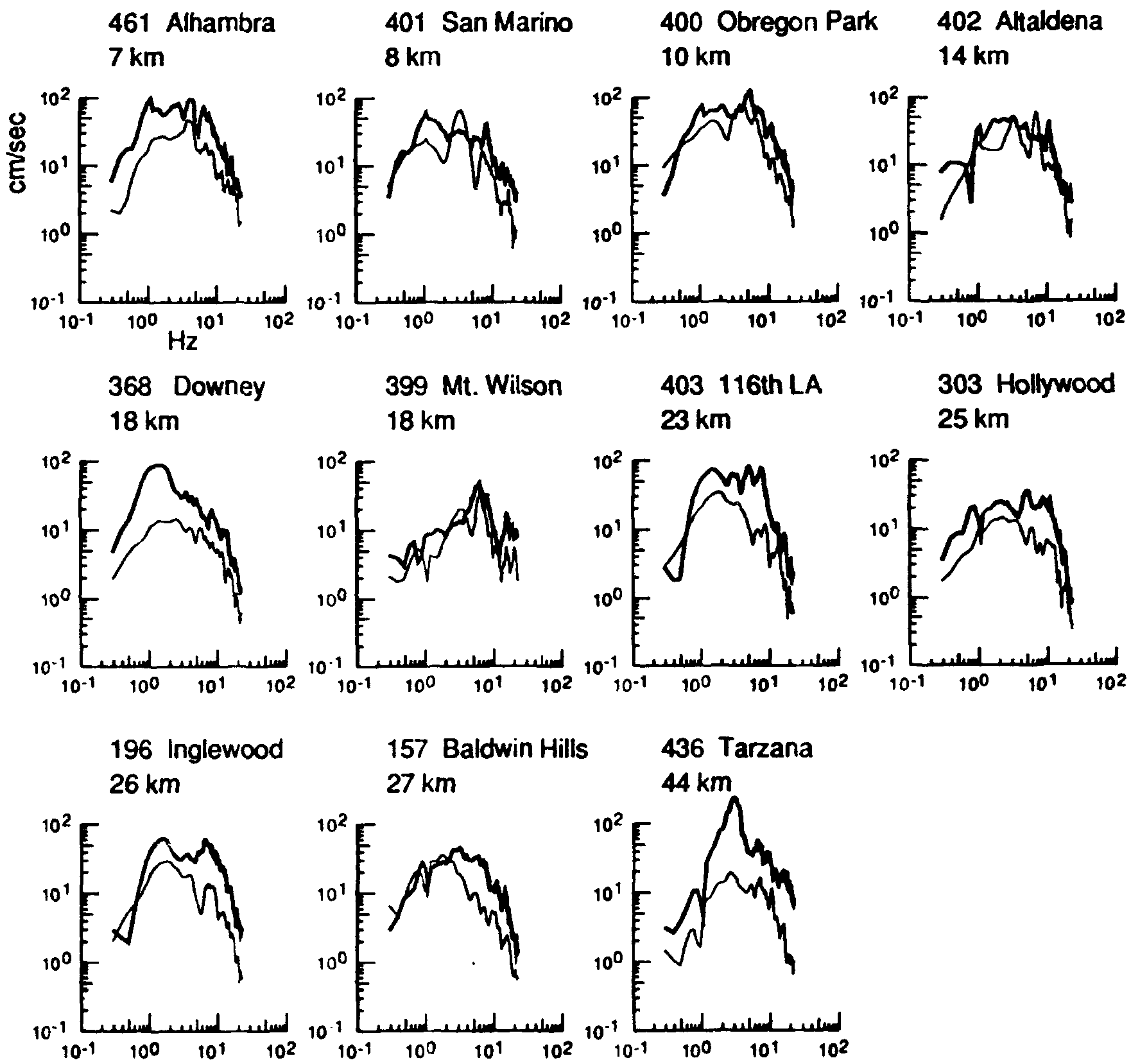

Figure 2 Acceleration spectra from the October 1 mainshock (heavy lines) and the October 4 aftershock (light lines) for 11 CSMIP stations. The distance in km from the mainshock epicenter is given beneath each station name. The rms sum of the SH and SV spectra is plotted.

basin near-surface structure. We use the depth and mechanism of their second and .argest source to represent the mainshock in this paper. The aftershock that occurred on October 4, 1987 was located 2 km northwest of the mainshock at a depth of 13.3 km, with a strike-slip mechanism on a vertical plane (Hauksson and Jones, 1989).

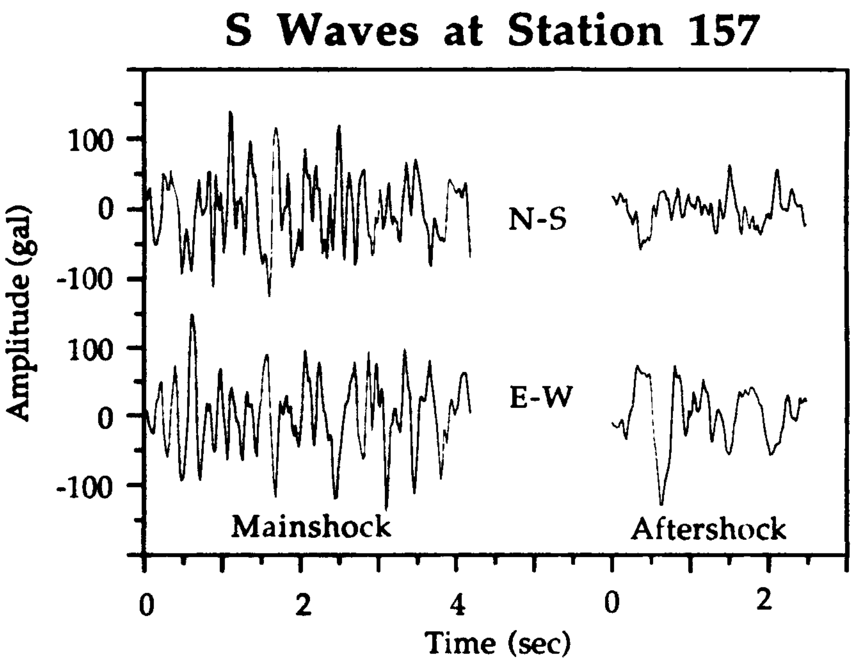

Figure 3. The windowed S waves for the mainshock and aftershock from station 157.

The 11 stations for which CDMG digitized strong motion records of both the mainshock and the aftershock are shown in Figure 1. The stations range from almost directly above the earthquakes to 50 km distant.

Acceleration spectra for the mainshock and aftershock are shown in Figure 2. The S waves are windowed into 4 sec and 2.5 sec segments for the mainshock and aftershock, respectively, and tapered, and then fourier transformed. The frequency of the peak acceleration ranges from 1 Hz for station 368 (Downey) to 5-6 Hz for stations 399 (Mt. Wilson). In general, the two spectra from each of the 11 stations are similar, although the mainshock produced a higher level of acceleration.

We processed the Whittier Narrows strong motions as follows: 1) All three components of motion were filtered with a central frequency of 1, 2, 4, 8 and 16 Hz. 2) The particle motion of each filtered record is characterized by its predominant azimuth of polarization (see Vidale, 1986, and Montalbetti and Kanasewich, 1970, for details of polarization analysis). This azimuth of polarization is the equivalent to the direction of the largest excursions in a particle motion diagram. The predominant azimuth of particle motion is compared with the azimuth expected from the source and receiver locations and the focal mechanism of the earthquake.

This procedure is first illustrated in detail for station 157 (Baldwin Hills), then applied to the other ten stations. The S waves from station 157 are shown in Figure 3.

Figure 4 shows the particle motions of the passband filtered horizontal motions, while Figure 5a shows the dominant direction of polarization in each passband for station 157. It is important to note that in Figure 5, the polarization is measured clockwise from north, while Figures 3 and 4 show rotation into radial and transverse components, where clockwise and outward are positive. Polarization has the usual two-fold ambiguity, for example, north-south vibration has a direction of either north or south. We therefor plot polarization in the 180° range centered about north.

The polarization directions agree between the records for the two events in each pass band, but do not agree particularly well with the directions predicted by the focal mechanisms. The directions of polarization of the broadband signals shown at the right of Figure 5a are similar for the two earthquakes because the azimuths agree fairly well in the 2 and 4 Hz windows where station 157 recorded the strongest accelerations as is apparent in the spectrum in Figure 2. The polarization directions also agree between the two events in other passbands, however the directions do not match well between different frequencies. These patterns can also be seen in Figure 4, particularly by comparing the 4 Hz and 8 Hz polarizations for the two events.

This pattern suggests that for station 157, 4 Hz shaking tends to be strong in the direction N30W, while 8 Hz shaking is strongest N75E, which is information that may be useful for earthquake engineers. This pattern also suggests that it is not the earthquake source that is controlling the polarization of the S waves. The other ten stations also show similar patterns to varying extents. They are presented in order of increasing station number (i.e., random order).

Station 196 (Inglewood) does not show agreement between the two events in broadband polarization direction, as seen in Figure 5b. Closer examination reveals some evidence for directional site effects, however. The strongest peaks in the two spectra for this station in Figure 2 lie at 2 Hz, and at this frequency the polarization directions agree between the mainshock and the aftershock. The broadband disagreement arises from stronger high frequency energy in the mainshock, but more energy at 1 Hz in the aftershock.

Figure 5c shows good agreement in the broadband polarization for station 303 (Hollywood), despite a large difference in the polarizations predicted by the focal mechanisms. The spectra for this station do not show prominent peaks.

Particle Motions for the Mainshock at Station 157

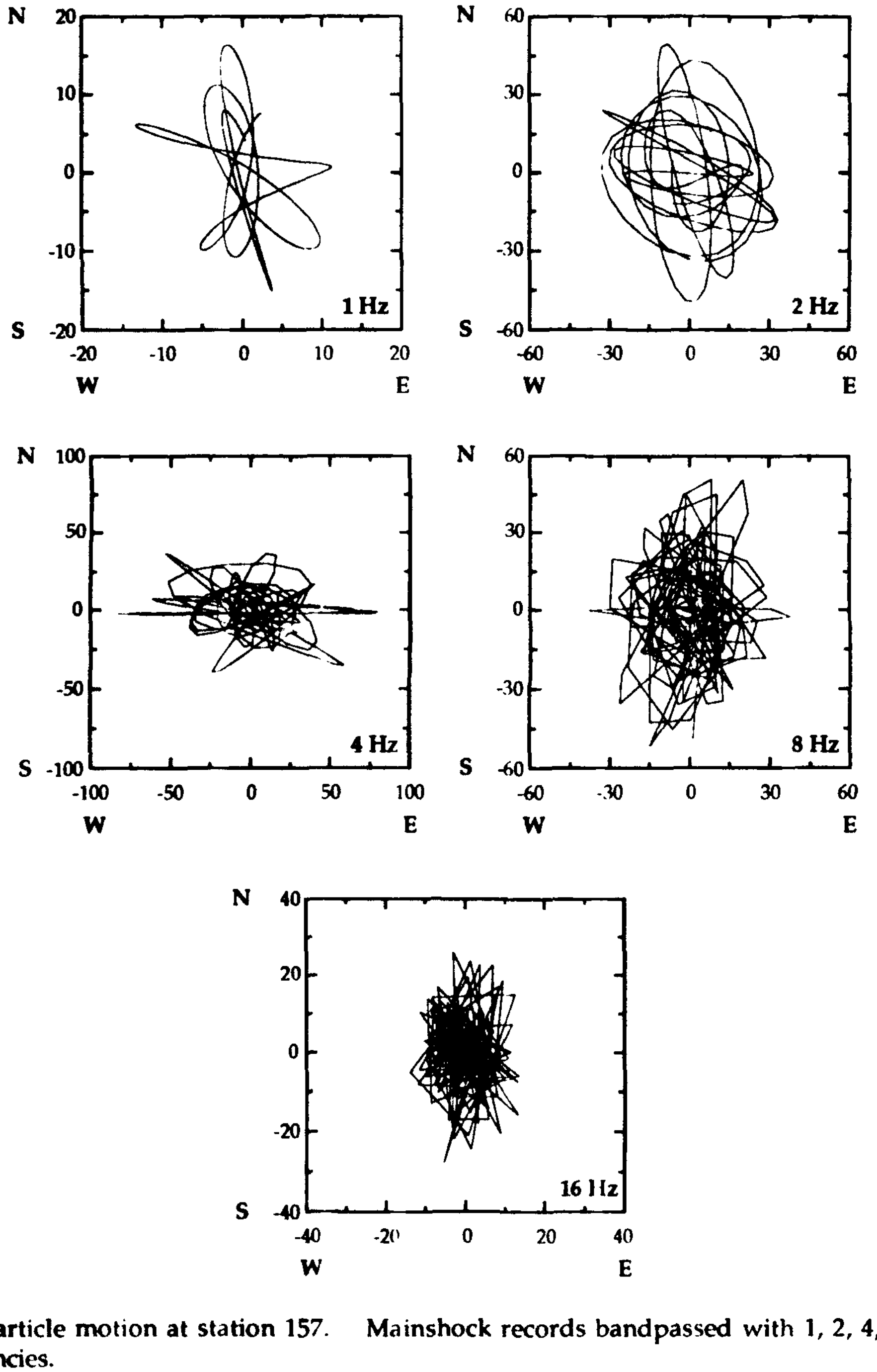

Figure 4a. Particle motion at station 157. Mainshock records bandpassed with 1, 2, 4, 8, and 16 Hz center frequencies.

Particle Motions for the 4 October Aftershock at Station 157

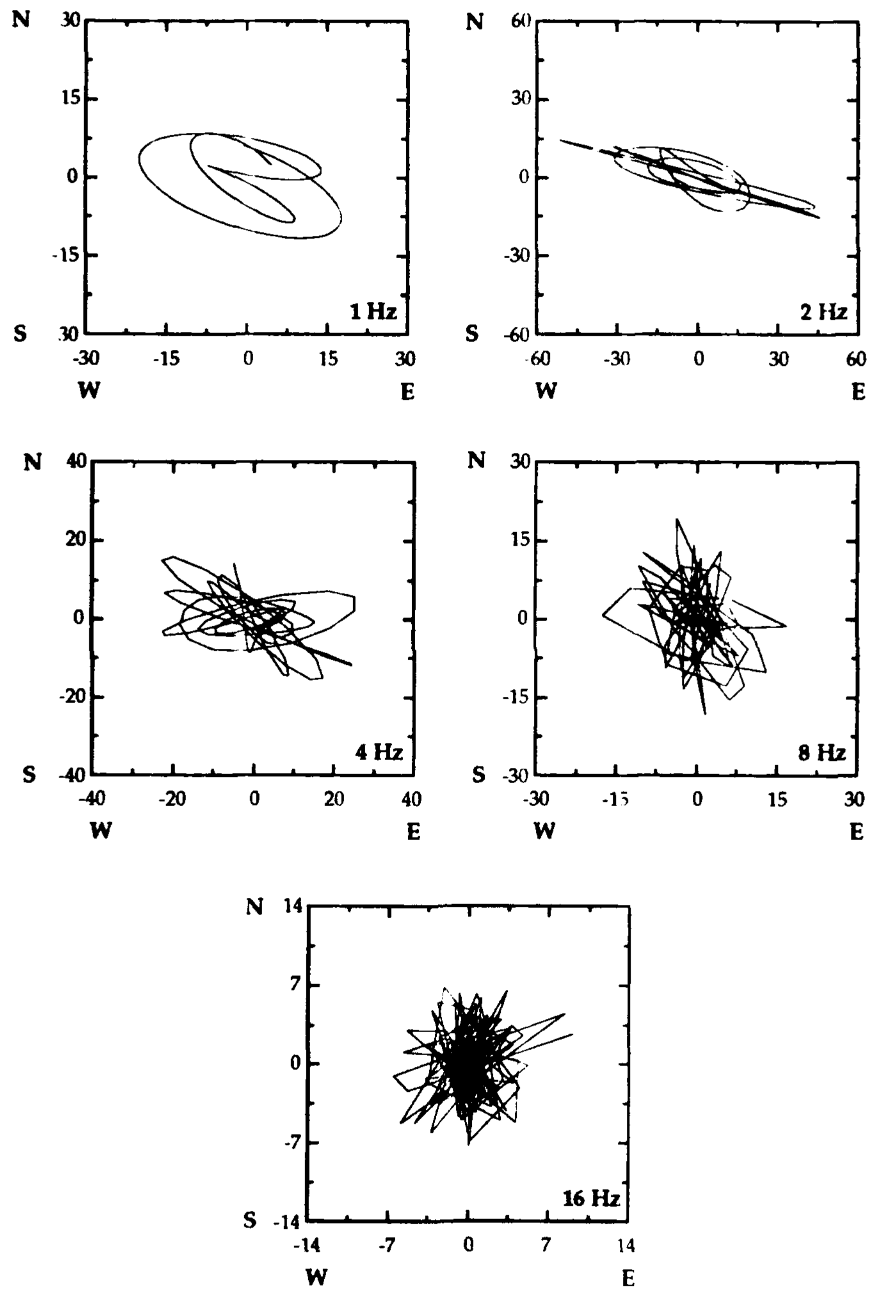

Figure 4b. Particle motion at station 157. Aftershock records bandpassed with 1, 2, 4, 8, and 16 Hz center frequencies.

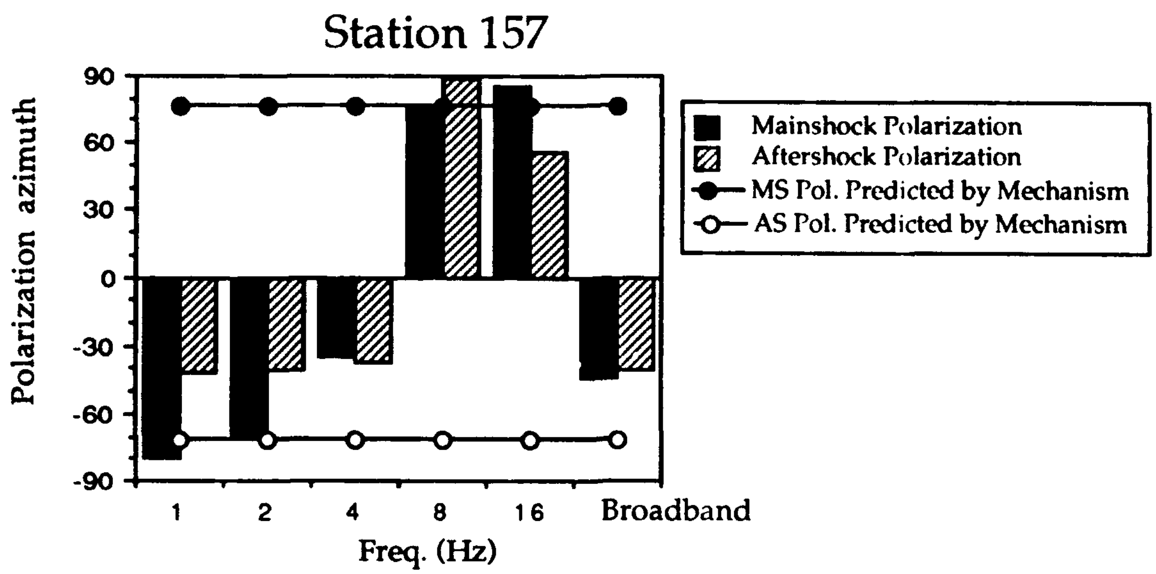

Figure 5a. The primary azimuth of particle motion during the mainshock and aftershock as a function of frequency for station 157. The circles indicate the azimuth of polarization expected from the focal mechanisms of the two events, and the bars indicate the polarization measured from the observations. Polarization is measured clockwise from north (north = 0°, northeast = 45°). The polarizations of the mainshock and aftershock are more similar to each other than to the predictions from their respective focal mechanisms.

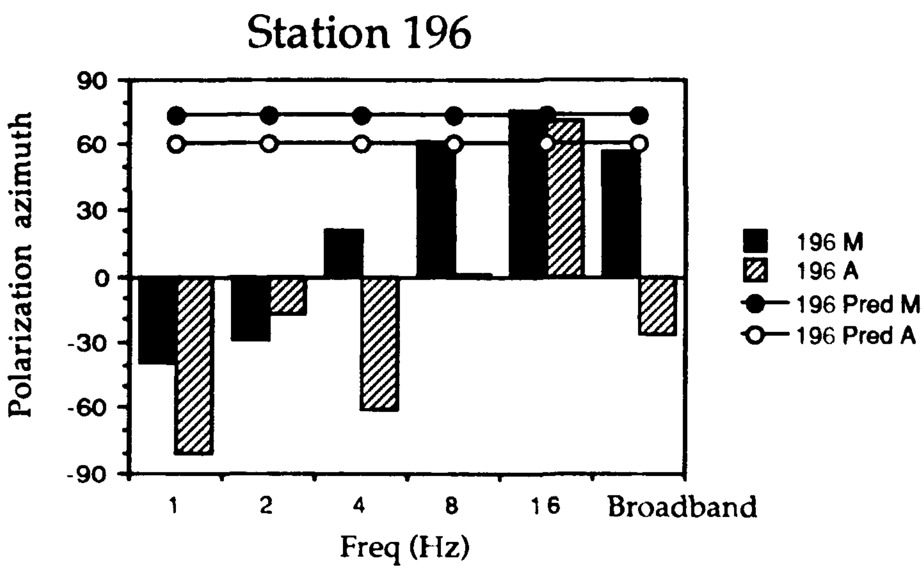

Figure 5b. The azimuth of particle motion of the mainshock and aftershock as a function of frequency for station 196. See caption for Figure 5a.

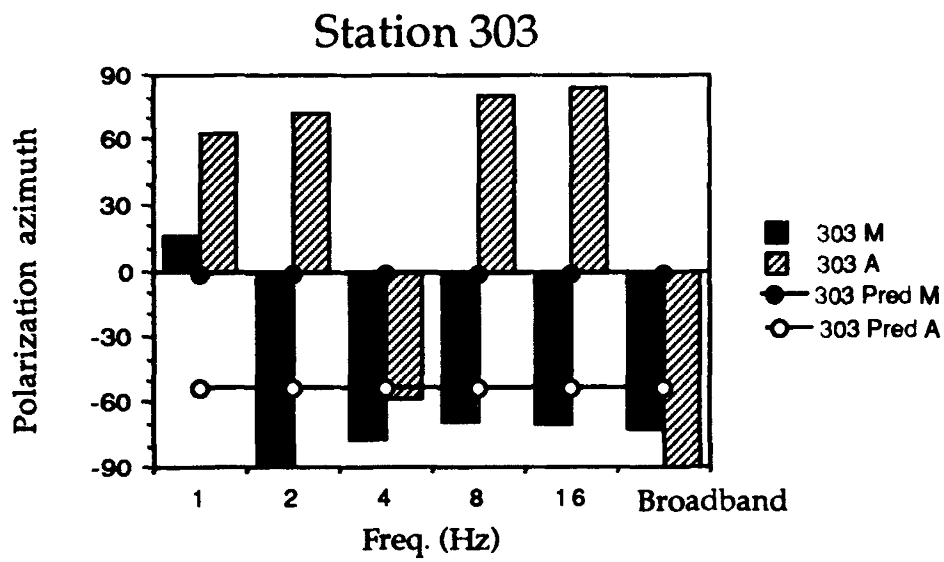

Figure 5c. The azimuth of particle motion of the mainshock and aftershock as a function of frequency for station 303. See caption for Figure 5a.

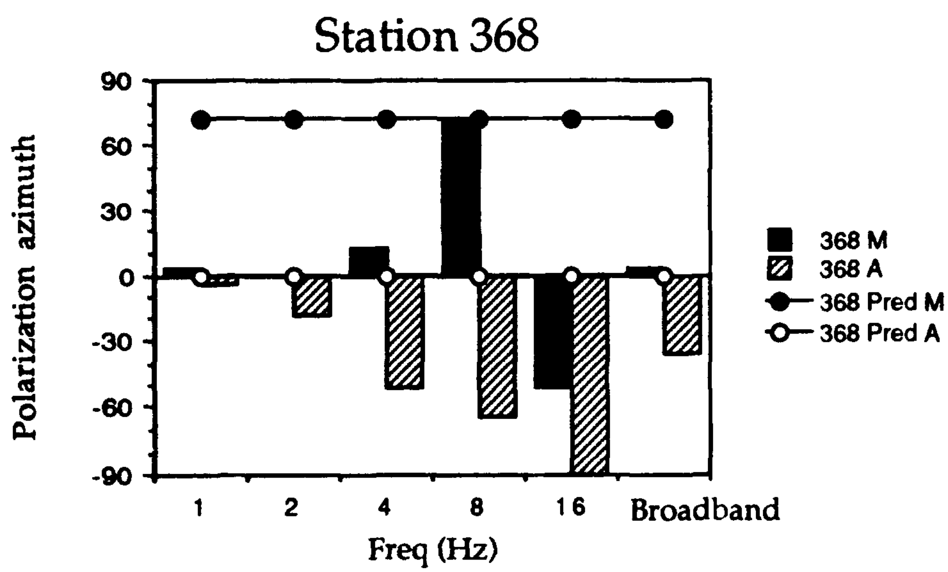

Figure 5d. The azimuth of particle motion of the mainshock and aftershock as a function of frequency for station 368. See caption for Figure 5a.

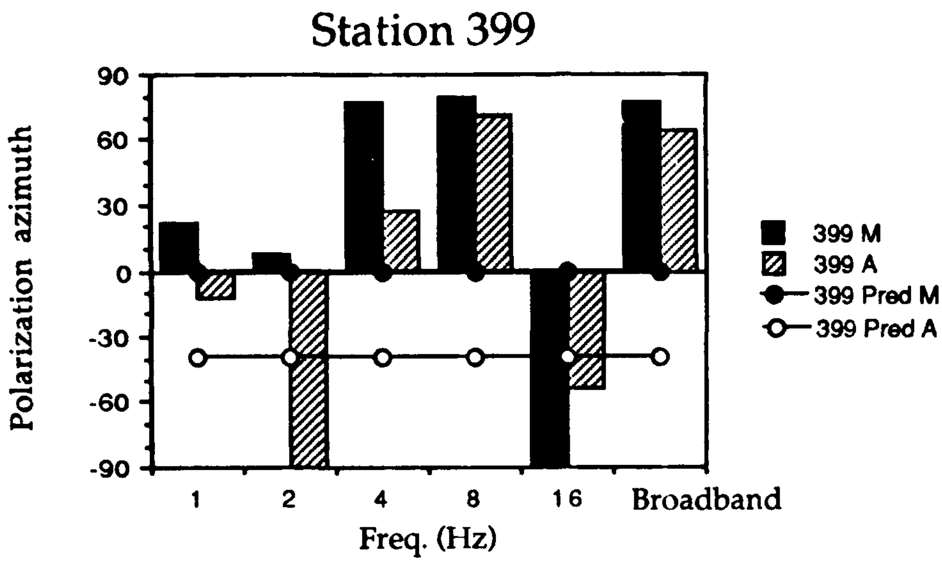

Figure 5e. The azimuth of particle motion of the mainshock and aftershock as a function of frequency for station 399. See caption for Figure 5a.

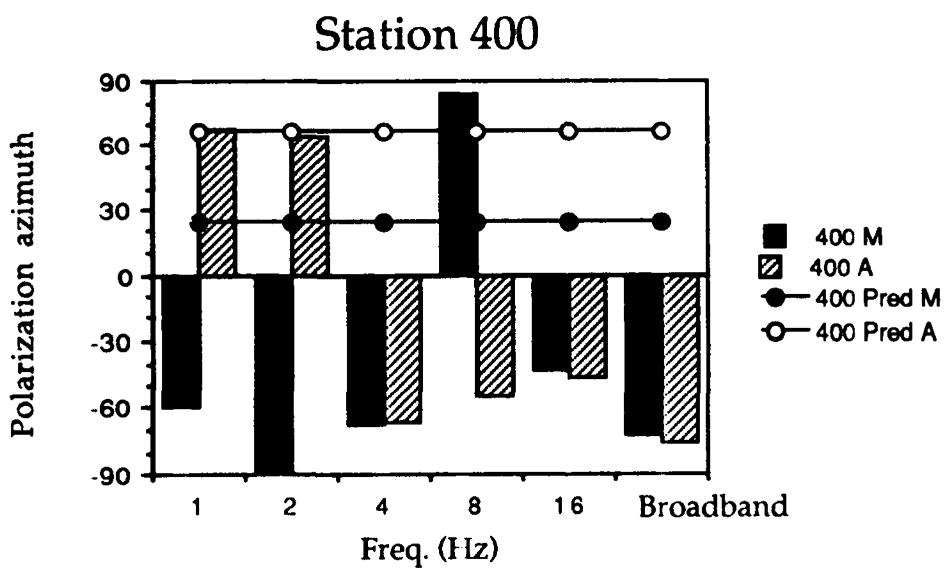

Figure 5f. The azimuth of particle motion of the mainshock and aftershock as a function of frequency for station 400. See caption for Figure 5a.

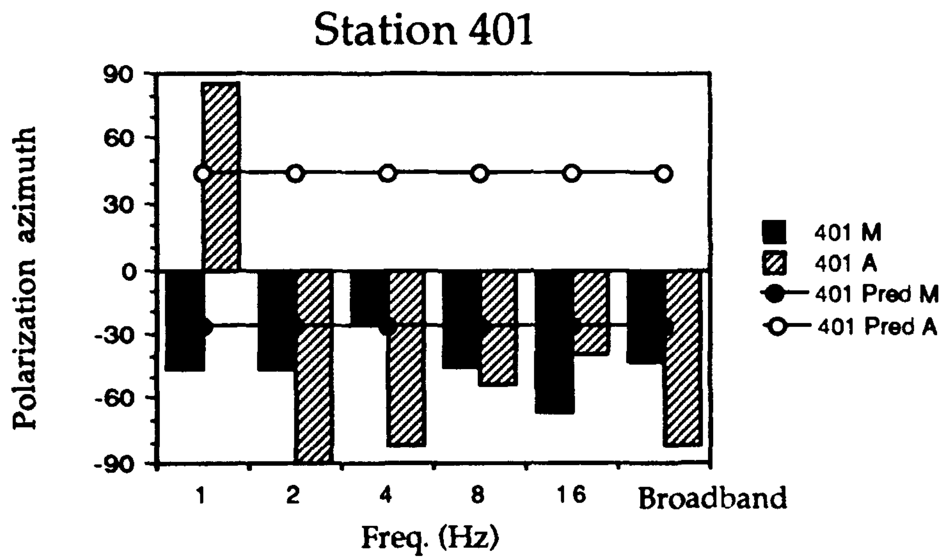

Figure 5g. The azimuth of particle motion of the mainshock and aftershock as a function of frequency for station 401. See caption for Figure 5a.

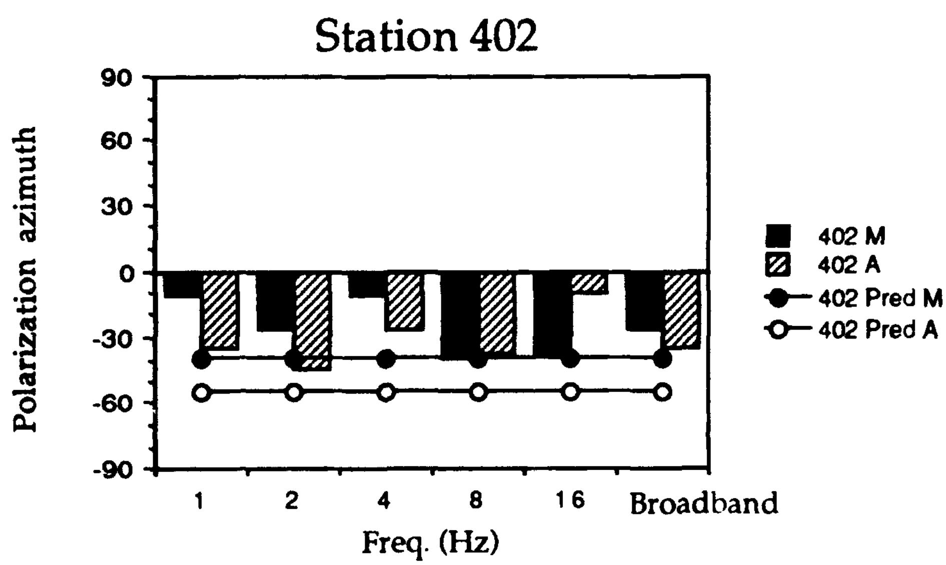

Figure 5h. The azimuth of particle motion of the mainshock and aftershock as a function of frequency for station 402. See caption for Figure 5a.

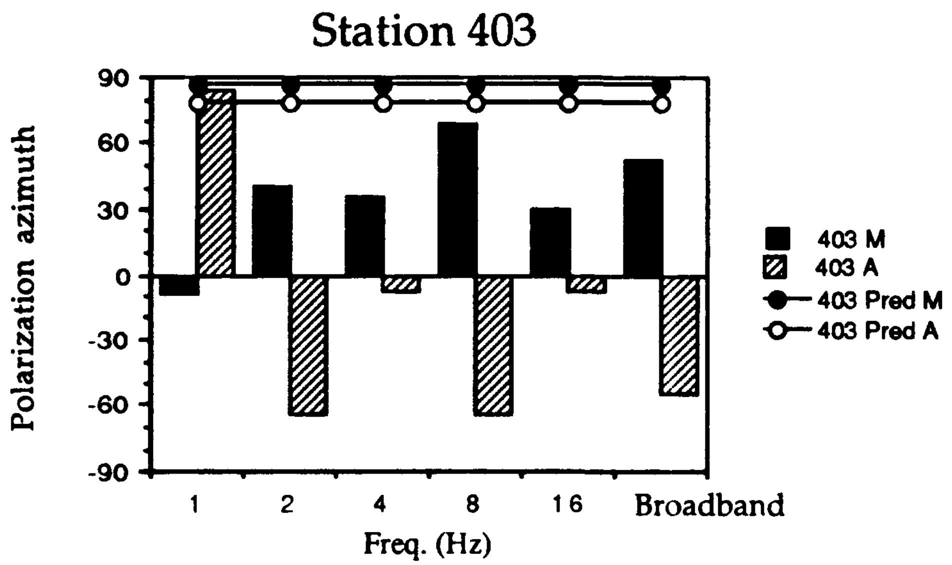

Figure 5i. The azimuth of particle motion of the mainshock and aftershock as a function of frequency for station 403. See caption for Figure 5a.

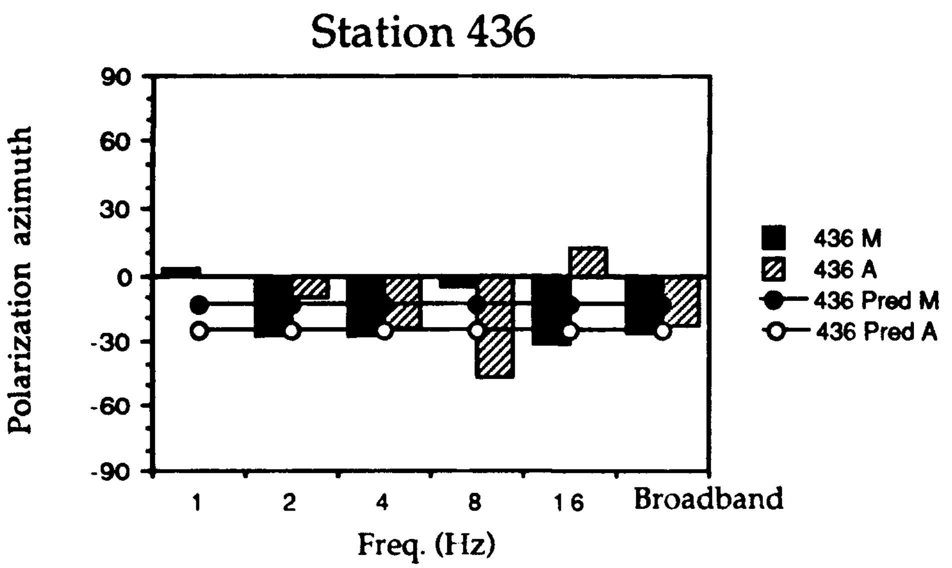

Figure 5j. The azimuth of particle motion of the mainshock and aftershock as a function of frequency for station 436. See caption for Figure 5a.

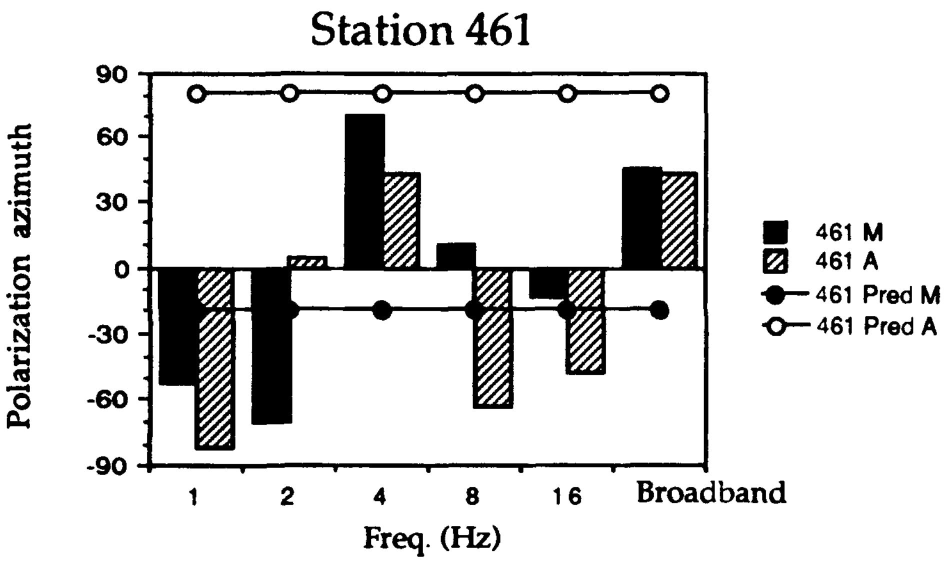

Figure 5k. The azimuth of particle motion of the mainshock and aftershock as a function of frequency for station 461. See caption for Figure 5a.

Figure 5d, for station 368 (Downey) shows that the directions of polarization are closer to each other than expected from the focal mechanisms. The 1 Hz passband, which has the largest spectral peak for the mainshock, shows excellent agreement in polarization.

Figure 5e shows the polarization for station 399 (Mt. Wilson). The polarization directions for the two events agree in most pass bands, but again do not agree particularly well with the directions predicted by the focal mechanisms. The azimuths agree fairly well in the 4 and 8 Hz windows where station 399 recorded the strongest accelerations (see Figure 2), and thus again the directions of polarization shown at the right of Figure 5e are very similar for the two earthquakes.

Station 400 (Figure 5f, Obregon Park) has good agreement in polarization direction at 4 Hz, where there is a peak in the spectra. The predicted directions from the focal mechanism are not similar to the observed directions.

Station 401 (Figure 5g, San Marino) does not have good agreement in polarization direction, although even greater disagreement in polarization direction is predicted by the focal mechanism. Different peaks appear in the spectra of the mainshock and aftershock, but no particular pattern is seen.

Station 402 (Figure 5h, Altadena) shows excellent agreement, even better than is predicted from the focal mechanisms. These signals seem coherently polarized at all frequencies, and close to the predicted directions.

Station 403 (Figure 5i, 116th St., Los Angeles) shows poor agreement in polarization direction despite the prediction of good agreement from the focal mechanisms.

Station 436 (Figure 5j, Tarzana) shows excellent agreement in the dominant direction of polarization, perhaps controlled by the sharp peak at 3 Hz. The mainshock and the aftershock are predicted to have a similar direction of polarization. This suggests that the discrepancy between the anomalous amplification seen in Tarzana from the mainshock and the more normal aftershock spectra is *not* due to a difference in the polarization of the incident S waves, which might have resulted in different levels of amplification at Tarzana in the two events. It may be due, however, to strong site amplification of 3 Hz energy coupled with weak excitation of 3 Hz energy in the aftershock, although evidence for this from the spectra in Figure 2 is equivocal.

Station 461 (Figure 5k, Alhambra) shows excellent agreement in broadband polarization direction despite the prediction of orthogonal motion from the focal mechanisms. The spectra are relatively flat.

Figure 6 shows that eight (157, 303, 368, 399, 400, 402, 436, 461) of the eleven stations have similar polarizations for the mainshock and the aftershock. This suggests that a majority of the stations may have a characteristic direction of polarization, which does not change from event to event.

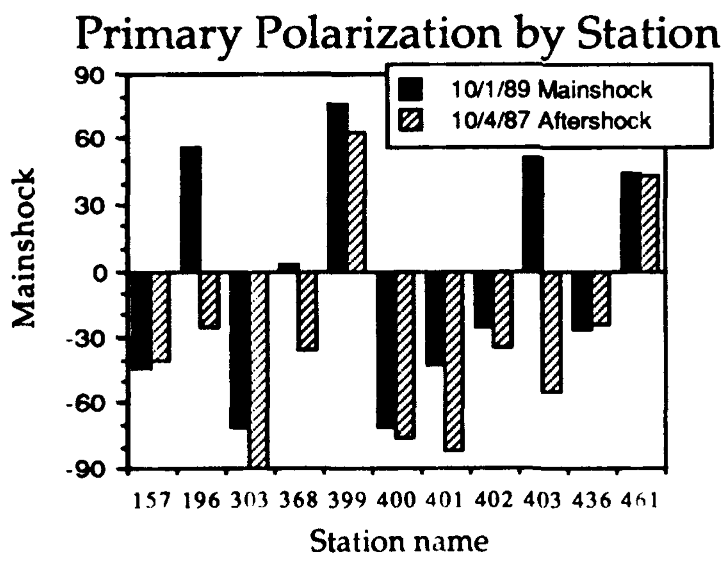

Figure 6. Observed directions of the strongest polarization of the broadband signal for acceleration records from the Whittier Narrows mainshock and aftershock for 11 stations. The predicted polarizations from the first-motion focal mechanisms do not agree as well. Polarizations are measured clockwise from north.

Previous work (Vidale, 1989) on the Whittier Narrows sequence suggests that the focal mechanism controls peak acceleration at a site, but the data presented here indicate that in many cases, the azimuth of polarization of the motion in the range 1-16 Hz depends on the site. In addition, in several cases, including stations 157, 196, 368, 399, 400, 436, the most similarity between the mainshock and aftershock polarizations is in the pass band where spectral peaks appear, suggesting that the geologic features that enhance amplitudes in a particular frequency band also have a preferred direction of particle motion. The 11 stations span a wide range of surficial geology from hard rock to soft sediment, summarized in Table 1, suggesting that these directional resonances are probably a common feature.

The present study provides results that complement those of Bonamassa *et al.* (1990). Bonamassa finds that S waves from 11 aftershocks of the Loma Prieta earthquake recorded at a dense 6-station 3-component array in the Santa Cruz mountains show directional resonances. The resonances vary within the array on a scale of 25 meters, but persist for a given station for a range of earthquake locations and expected S wave polarizations. The rapid variation across the array suggests very near-surface structure is causing the resonances. The present study has higher quality stations that are situated in a wider range of surficial geologies, suggesting that these resonances are a common occurrence.

The most likely explanation for these azimuthal patterns is that particle motion in one compass direction is amplified compared to the motion in other directions. The specific geological structures that cause this amplification are not yet known. Surface topography seems unlikely, as Buchbinder and Haddon (1990) estimate only small S-wave azimuthal anomalies due to topography. The most likely structures are strong lateral variation in the S-wave velocity in the top 10's of meters, where the velocity can be very low.

Conclusions

Three-component seismic recordings for 8 of 11 stations of the Whittier Narrows earthquake sequence show that in the frequency range from 1 to 16 Hertz, there is a preferred direction of ground motion, which we term "directional resonance" that does not depend on the polarization of the shear waves expected from the focal mechanism. The study of Bonamassa et al. (1990) also finds directional resonances and suggests that the preferred direction also does not depend on the earthquake location.

These preliminary conclusions drawn from good data for various sites in the Los Angeles area suggest that in-depth analysis of the processes that control directional resonances is necessary. Earthquake engineers as well as seismologists will benefit from knowledge of the strength of the characteristic resonance at a site and the area over which it is coherent.

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Table 1 : Stations co-ordinates and geology

Station	Latitude	Longitude	Directional Resonance	Surficial Geology
157	34.01	118.36	Yes	Fill over shale, sandstone
196	33.90	118.28	No	Terrace deposits
303	34.09	118.34	Yes	Alluvium (130 m) over sandstone, shale
368	33.92	118.17	Yes	Deep alluvium
399	34.22	118.06	Yes	Quartz Diorite
400	34.04	118.18	Yes	Alluvium
401	34.11	118.13	No	Alluvium
402	34.18	118.10	Yes	Alluvium
403	33.93	118.26	No	Terrace deposits
436	34.16	118.53	Yes	Shallow alluvium (10 m) over sandstone, shale
461	34.07	118.15	Yes	Alluvium

The strong influence of near-surface geology on the direction of ground motion above 1 Hertz in the Loma Prieta earthquake

John E. Vidale and Ornella Bonamassa

Abstract

The direction of strong shaking observed at 13 California Division of Mining and Geology sites across San Francisco and Oakland at frequencies less than one Hz roughly agrees with a prediction calculated from the well-determined long-period focal mechanism. The directions of shaking at frequencies higher than one Hz, however, show little resemblance to the simple prediction, suggesting that the near-surface geology interacts with the higher frequency seismic waves in a complicated way. This propagational complication, if a common feature, makes the recovery of source information from surface-recorded seismograms difficult and suggests that the focal mechanism does not determine the direction of strongest shaking in an earthquake at this range of about 100 km above about one Hz. It also shows that there is not a clean separation of P-SV and SH energy in regional wave propagation.

Introduction

The motion that an earthquake causes at the surface of the Earth is a combination of the details of the faulting at depth and the complications due to propagation through structures within the Earth of the seismic energy released by the faulting. Various ways of measuring the seismic source and propagational complications have been described. There exists considerable literature that documents the usefulness of the concept of a *site response*, where a particular site has a fixed set of frequencies which are amplified at that site no matter how the how the ground motion is induced (Joyner et al., 1976, Rogers et al., 1984, Borcherdt, 1970, Joyner et al., 1981). Seismic wave interaction with large-scale structures such as major sedimentary basins can be described deterministically; these structures can be shown to distort seismic waves in a semi-predictable way (Vidale and Helmberger, 1988, Kawase and Aki, 1989, Kagami et al., 1986). Bridging the gap between well-understood large-scale structures and fine structure where only the amplitude versus frequency behavior has been studied is the goal of considerable recent research.

This report will concentrate on empirically quantifying the distortion to the direction of strongest shaking caused by earth structures, since it has been suggested that the direction of shaking is sometimes a feature of the recording site (a *directional site resonance*) rather than the earthquake (Vidale et al., 1991, Bonamassa et al., 1991, Bonamassa and Vidale, 1991). These and other observations of horizontal ground motion above one Hz frequency (Chouet, 1989, Abrahamson et al., 1989) show very small lateral correlation distances, less than 10's of meters for frequencies above a few Hz. Also, comparisons of seismograms written by surface and borehole instruments have shown that propagation through the shallowest 10's of meters of the Earth can severely distort seismic pulses (Haukkson et al., 1987, Malin et al., 1988, Aster and Shearer, 1991).

While the near-surface layers of the earth appear to scramble high-frequency waves, at long periods the Earth often affects seismic waves in a predictable way that may be stripped off to study the earthquake source. This focuses on finding the transition frequency where earth structure, and mainly near-surface geology, begin .o obscure the signature of the seismic source at the range of 100 km. We will find a transition frequency near one Hz.

Data

The Loma Prieta earthquake of 18 October 1989, often and perhaps more appropriately called the Santa Cruz Mountains earthquake, was the largest to strike the San Francisco Bay area since 1906. It caused considerable damage and loss of life.

On the positive side, this earthquake was captured by more than a hundred strong motion seismometers, producing an unprecedented opportunity to investigate details of the earthquake and earthquake hazards in general. Numerous investigators are reconstructing the spatial and temporal patterns of fault movements during the 5 to 10 seconds it took for the earthquake to occur (Loma Prieta source references, 1990).

Study area and earthquake

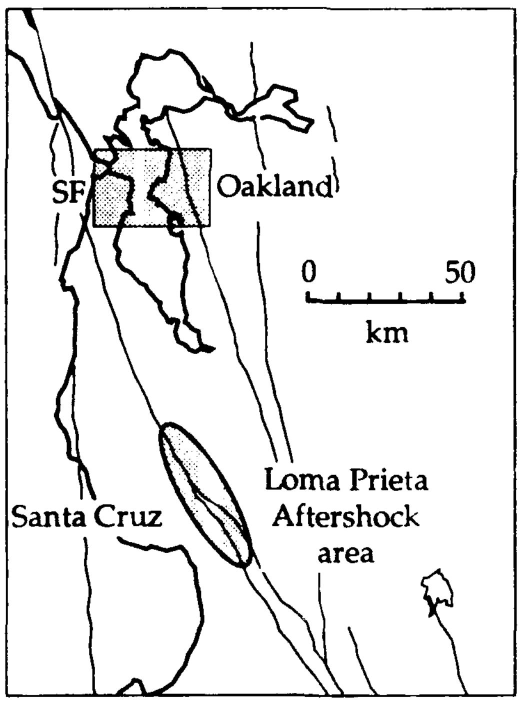

Figure 1. Location of the area of this study in relation to the aftershock zone of the 18 October 1989 earthquake. This study area is shown in more detail in Figure 2.

This report will concentrate on observations of the effect of earth structure on the seismic waves radiated by the earthquake. To achieve this end, we examine the Loma Prieta mainshock records of 13 accelerometers located near San Francisco. The regional setting is shown in Figure 1. The seismometers are spread across San

Francisco and Oakland, as shown in Figure 2. The station numbers, names, locations, and geologic settings are given in Table 1. These are all free-field, basement, or first floor installations, and all except station 480 are in two- or fewer story structures. The recordings, which were originally captured on film, have been digitized and disseminated by the California Strong Motion Instrumentation Project, managed by the California Division of Mines and Geology.

Map of CDMG Stations

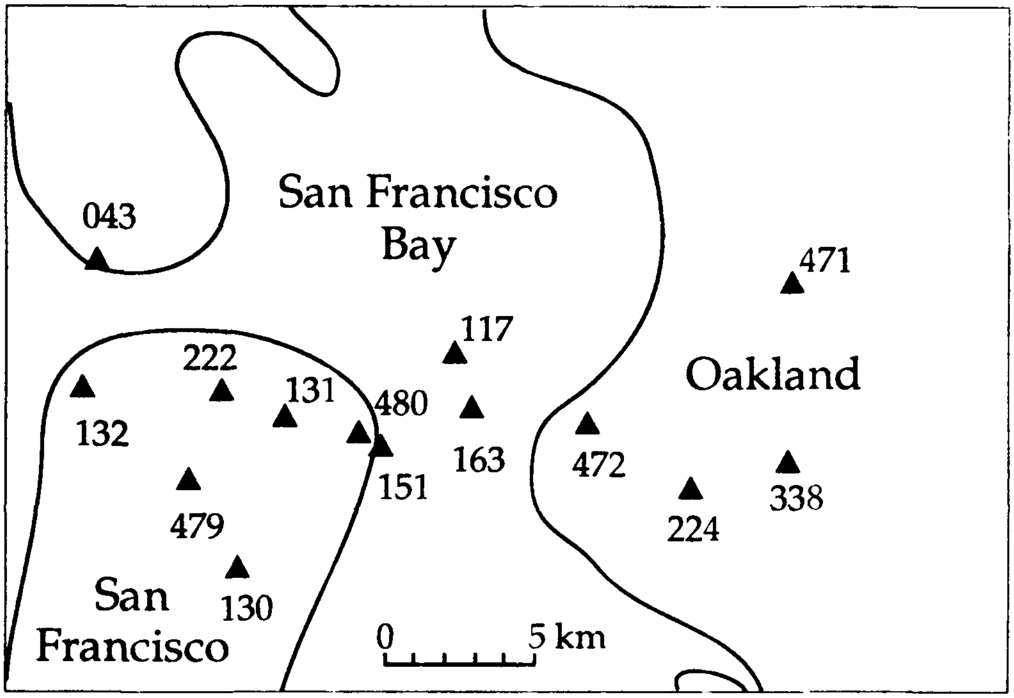

Figure 2. Location of the 13 California Strong Motion Instrumentation Program stations whose recordings are used in this report. The name and surficial geology of each station are given in Table 1.

This data is well-suited for analysis of seismic wave propagation because of the large size of the Loma Prieta mainshock and the close spacing of the instrument sites relative to the distance to the earthquake. The large size of the earthquake generated sufficient long-period seismic energy that ground motion at least down to 0.2 Hz (5 seconds period) were reliably measured by the strong motion instruments. Similar periods were recovered from the 1971 San Fernando earthquake, but fewer instruments were deployed at that time. The Whittier Narrows earthquake was recorded at a comparable number of stations, but strong motion recordings of the M_L 5.9 quake did not have recoverable energy at less that 0.5 Hz. This group of 13 stations lies within 20 km by 10 km area, but is 60 to 100 km from the fault plane that broke as estimated from aftershocks (Oppenheimer, 1990). Consequently, the stations only span about 10 to 15° in azimuth from the earthquake, and the signal that would be recorded across these stations would be similar in the absence of structural complications, particularly in their direction of shaking.

The motions expected for a simple, layered structure is nearly perfectly linear polarization in the WSW-ENE direction. We determined this with a reflectivity simulation (Muller, 1985) with a simple source. The oblique thrust mechanism of the Loma Prieta earthquake produces a large pulse of shear wave energy on the transverse component, and little motion on the radial and vertical components of motion at this azimuth and range, which is close to directly along the San Andreas fault. A simple two-layer crust over Moho model is assumed in this calculation. A more complicated earthquake, extending ten's of kilometers, as the Loma Prieta event probably did (Loma Prieta source references) would produce more complicated radiation, but it still would maintain a similar particle motion. The primary variation in particle motion predicted across the array is the rotation of the transverse direction of motion clockwise as one considers the stations more to the east. The large transverse pulse, which can also be thought of as an incipient Love wave, dominates all stations, and would appear for all frequencies considered in this report.

0.4 to 0.8 Hz Particle Motions

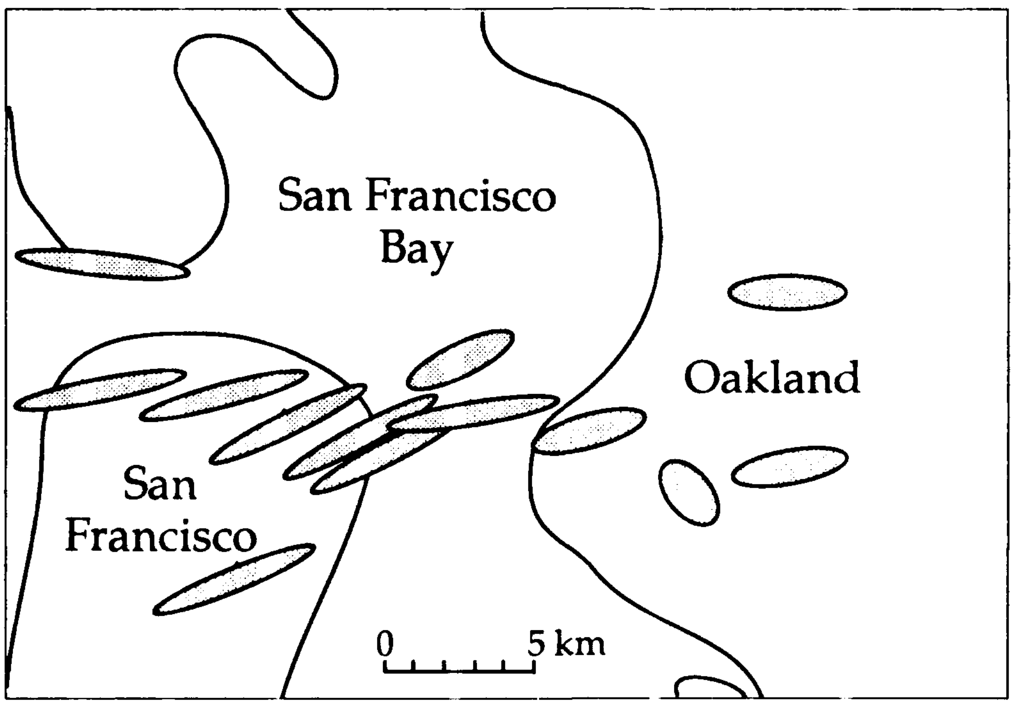

Figure 3. Particle motion diagrams for observations of the 13 stations listed in Table 1 of the 18 October 1989 Loma Prieta earthquake in the passband from 0.4 to 0.8 Hz (1.25 to 2.5 second period). Two passes of a three pole Butterworth filter were applied.

The observed ground motions show some resemblance to the prediction of the reflectivity simulation. Figures 3 through 6 show the particle motion directions seen in the acceleration records in six different passbands. Comparison of the synthetic particle motion directions with the observed directions shows a simple and unambiguous pattern: The directions of strongest motions at frequencies less than one Hz shown in Figure 3 generally agree with the direction expected from the simulation and the directions at frequencies above one Hz shown in Figures 4, 5 and 6 show directions of shaking that bear little relation to the direction expected.

Some caveats must be noted. Time domain information has been suppressed in this presentation, so it remains possible that the initial S wave arrivals exhibit the polarization direction expected from the focal mechanism even at high frequencies as has been observed by Bonamassa and Vidale (1991) and has also been observed for P waves by Menke (1990). Individual stations may record bizarre phenomenon; for example, Treasure Island (Station 117) underwent liquefaction near the CDMG site, Cliff House (Station 132) is situated on steep topography (Borcherdt, 1990), Oakland (Station 472) is located on a wharf that may be subject water waves in the bay, Oakland (Station 480) is placed in the basement of an 18-story building that may sway. Despite these potential outliers, the pattern is quite consistent across the array of strong motion stations.

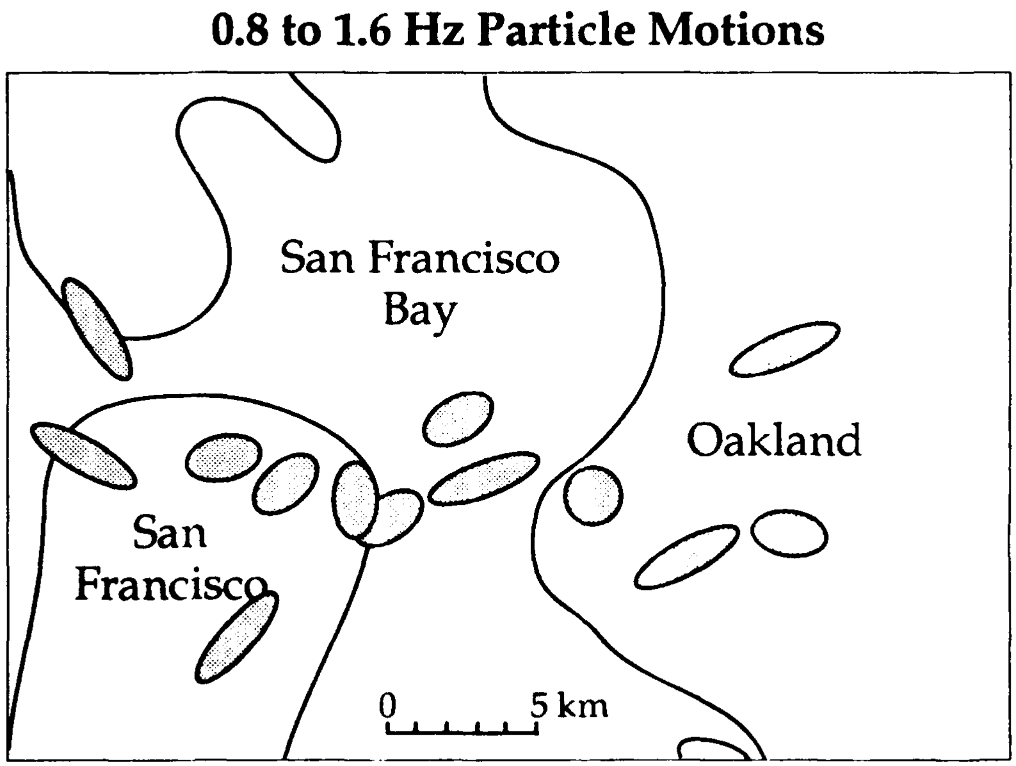

Figure 4. Particle motion diagrams for observations of the 13 stations listed in Table 1 of the 18 October 1989 Loma Prieta earthquake in the passband from 0.8 to 1.6 Hz.

Conclusions and Unanswered Questions

This relatively dense array of strong motion stations has provided one of the best fairly broadband (0.1 to 5.0 Hz) glimpses to date of the seismic wavefield generated by a large earthquake less than 100 km distant. The transition from particle motion that indicates source character to particle motion that is strongly affected by propagation through the Earth clearly takes place around 1 Hz. This highly variable high-frequency polarization is consistent with most previous studies; Vidale et al. (1991) and Bonamassa and Vidale (1991) see similar gross distortion of the 2 to 20 Hz seismic waves, probably by near-surface geology. As mentioned above, comparisons of seismic waves recorded on the surface and in boreholes have documented the scrambling effects of the near-surface at high frequencies (Aster and Shearer, 1991, Haukkson et al., 1987).

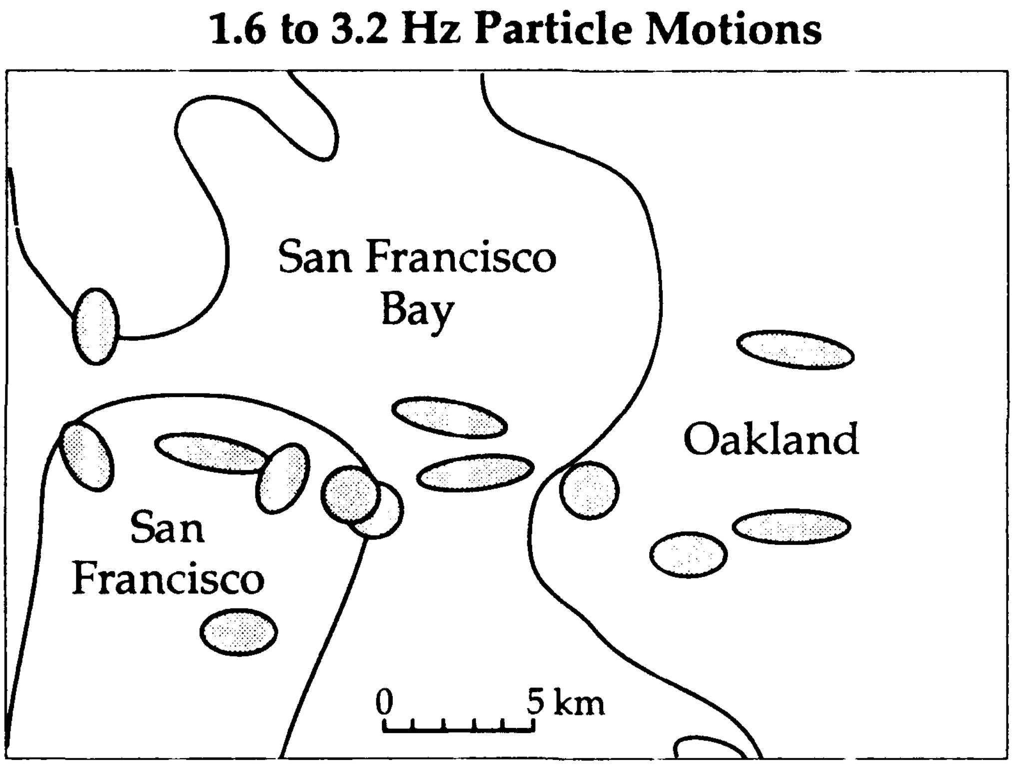

Figure 5. Particle motion diagrams for observations of the 13 stations listed in Table 1 of the 18 October 1989 Loma Prieta earthquake in the passband from 1.6 to 3.2 Hz.

The coherence below 1 Hz is also consistent with previous work. Trifunac (1988) showed that the 0.5 to 1.0 Hz seismic waves generated by the Whittier Narrows earthquake were fairly coherent. Helmberger and Liu (1985) have done a similar exercise at a closer range of 5 to 20 km, with a comparable answer of two Hz. Innumerable local, regional, and teleseismic earthquake source inversions have invariably proceeded on the assumption that seismic waves interact with Earth structure that is a stack of layers in the most complicated case, and source studies often interpret seismic energy up to frequencies near 1 Hz as indicating details of rupture propagation.

In contrast, Ebel (1988) finds that the 10 Hz particle motions of small earthquakes sometimes do reflect the focal mechanism. His study is in the stable craton in Germany, however, so perhaps the confounding effect of the near surface is strongest in active tectonic regions like California, where all the other cited works were sited.

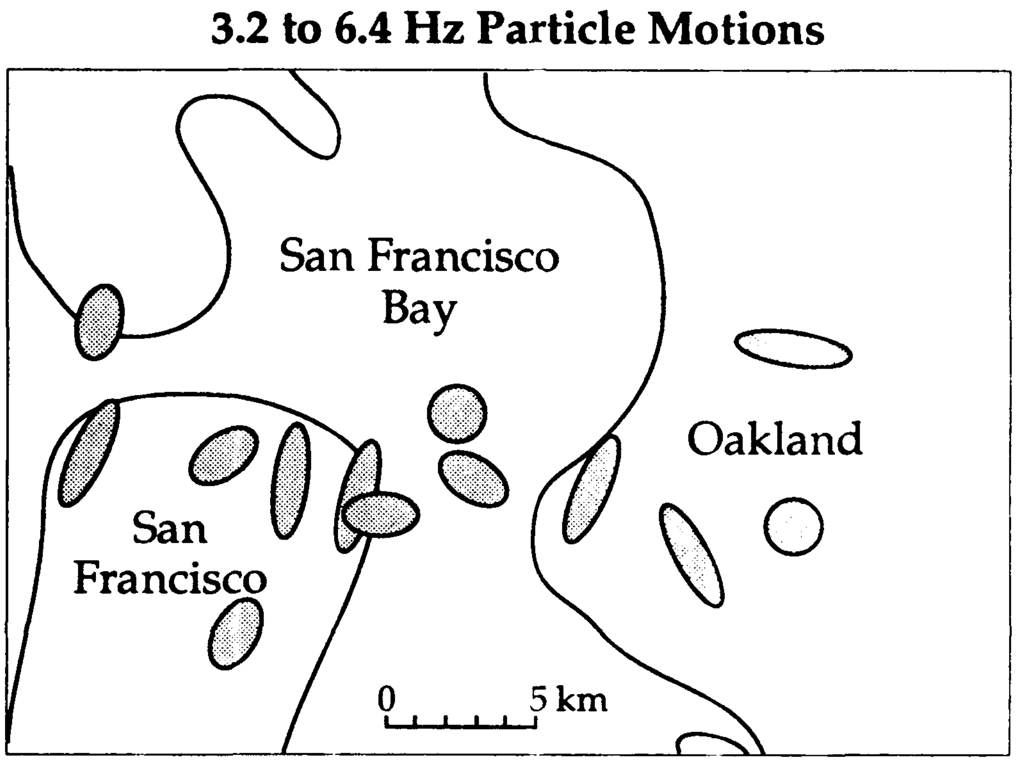

Figure 6. Particle motion diagrams for observations of the 13 stations listed in Table 1 of the 18 October 1989 Loma Prieta earthquake in the passband from 3.2 to 6.4 Hz.

The question that remains is: What geologic structures scramble the high-frequency polarization characteristics? The low spatial coherence suggests shallow structure, the borehole studies suggest shallow structure, and to the extent that common sense applies, the observation that the near surface is the least consolidated and most highly variable volume along the seismic ray path suggests that the structure lie near the surface. Candidates for these near surface structures include surface topography (Bard and Gariel, 1986, Kawase and Aki, 1989) and topography on the soil rock interface (Bard and Tucker, 1985). Candidate for wave interactions include focusing through seismic velocity gradients acting as lens (Rial, 1989, Langston and Lee, 1983), body-wave to surface-wave conversions at sharp, laterally heterogeneous velocity contrasts (Bard and Gariel, 1986, Vidale and Helmberger, 1988, Kawase and Aki, 1989), energy that becomes trapped and reverberates between high contrast interfaces (Novaro et al., 1990).

The task remaining is the construction of simulations with methods like three-dimensional finite differences (Frankel *et al.*, 1990) that reproduce to complexity we observe in the seismic wavefield using realistic velocity models. This task relies on the equally difficult task of accurately estimating realistic three dimensional velocity models of the near surface.

It is important to learn the processes affecting high-frequency seismic waves for earthquake hazard mitigation, earthquake source determination, and seismic array design goals. So far, 3-D site characterization is empirical, not based on an understanding of the physics involved. Clearly, optimal high frequency seismic arrays that aim to study seismic sources will minimize the interference from earth structures. Earthquake source inversions must allow sufficient variance in the solutions to withstand the randomization of the high-frequency waves by structure. Finally, site characterization for seismic hazard mitigation might become a more precise science when phase as well as amplitude information is incorporated into predictions of shaking in future earthquakes.

Table 1 - Location of thirteen CSMIP stations used

Name	Location	Near-surface geology
043	Point Bonita	2 m of broken rock, sandstone
117	Treasure Island	Fill
130	SF - Diamond Heights	Franciscan chert
131	SF - Pacific Heights	Franciscan sandstone, shale
132	SF - Cliff House	Franciscan sandstone, shale
151	SF - Rincon Hill	Franciscan sandstone, shale
163	Yerba Buena Island	Franciscan sandstone
222	SF - Presidio	Serpentine
224	Oakland - 2-story building	Alluvium
338	Piedmont - Jr. High	Weathered serpentine
471	Berkeley - LBL	Thin alluv. on shale, siltstone
472	Oakland - outer harbor wharf	Bay mud
480	Oakland - 18 story bldg.	Fill over bay mud

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Prof. Thomas Ahrens
Seismological Lab, 252-21
Division of Geological & Planetary Sciences
California Institute of Technology
Pasadena, CA 91125

Prof. Keiiti Aki
Center for Earth Sciences
University of Southern California
University Park
Los Angeles, CA 90089-0741

Prof. Shelton Alexander
Geosciences Department
403 Deike Building
The Pennsylvania State University
University Park, PA 16802

Dr. Ralph Alewine, III
DARPA/NMRO
3701 North Fairfax Drive
Arlington, VA 22203-1714

Prof. Charles B. Archambeau
CIRES
University of Colorado
Boulder, CO 80309

Dr. Thomas C. Bache, Jr.
Science Applications Int'l Corp.
10260 Campus Point Drive
San Diego, CA 92121 (2 copies)

Prof. Muawia Barazangi
Institute for the Study of the Continent
Cornell University
Ithaca, NY 14853

Dr. Jeff Barker
Department of Geological Sciences
State University of New York
at Binghamton
Vestal, NY 13901

Dr. Douglas R. Baumgardt
ENSCO, Inc
5400 Port Royal Road
Springfield, VA 22151-2388

Dr. Susan Beck
Department of Geosciences
Building #77
University of Arizona
Tuscon, AZ 85721

Dr. T.J. Bennett
S-CUBED
A Division of Maxwell Laboratories
11800 Sunrise Valley Drive, Suite 1212
Reston, VA 22091

Dr. Robert Blandford
AFTAC/TT, Center for Seismic Studies
1300 North 17th Street
Suite 1450
Arlington, VA 22209-2308

Dr. G.A. Bollinger
Department of Geological Sciences
Virginia Polytechnical Institute
21044 Derring Hall
Blacksburg, VA 24061

Dr. Stephen Bratt
Center for Seismic Studies
1300 North 17th Street
Suite 1450
Arlington, VA 22209-2308

Dr. Lawrence Burdick
Woodward-Clyde Consultants
566 El Dorado Street
Pasadena, CA 91109-3245

Dr. Robert Burridge
Schlumberger-Doll Research Center
Old Quarry Road
Ridgefield, CT 06877

Dr. Jerry Carter
Center for Seismic Studies
1300 North 17th Street
Suite 1450
Arlington, VA 22209-2308

Dr. Eric Chael
Division 9241
Sandia Laboratory
Albuquerque, NM 87185

Prof. Vernon F. Cormier
Department of Geology & Geophysics
U-45, Room 207
University of Connecticut
Storrs, CT 06268

Prof. Steven Day
Department of Geological Sciences
San Diego State University
San Diego, CA 92182

Marvin Denny
U.S. Department of Energy
Office of Arms Control
Washington, DC 20585

Dr. Zoltan Der
ENSCO, Inc.
5400 Port Royal Road
Springfield, VA 22151-2388

Prof. Adam Dziewonski
Hoffman Laboratory, Harvard University
Dept. of Earth Atmos. & Planetary Sciences
20 Oxford Street
Cambridge, MA 02138

Prof. John Ebel
Department of Geology & Geophysics
Boston College
Chestnut Hill, MA 02167

Eric Fielding
SNEE Hall
INSTOC
Cornell University
Ithaca, NY 14853

Dr. Mark D. Fisk
Mission Research Corporation
735 State Street
P.O. Drawer 719
Santa Barbara, CA 93102

Prof Stanley Flatte
Applied Sciences Building
University of California, Santa Cruz
Santa Cruz, CA 95064

Dr. John Foley
NER-Geo Sciences
1100 Crown Colony Drive
Quincy, MA 02169

Prof. Donald Forsyth
Department of Geological Sciences
Brown University
Providence, RI 02912

Dr. Art Frankel
U.S. Geological Survey
922 National Center
Reston, VA 22092

Dr. Cliff Frolich
Institute of Geophysics
8701 North Mopac
Austin, TX 78759

Dr. Holly Given
IGPP, A-025
Scripps Institute of Oceanography
University of California, San Diego
La Jolla, CA 92093

Dr. Jeffrey W. Given
SAIC
10260 Campus Point Drive
San Diego, CA 92121

Dr. Dale Glover
Defense Intelligence Agency
ATTN: ODT-1B
Washington, DC 20301

Dr. Indra Gupta
Teledyne Geotech
314 Montgomery Street
Alexanderia, VA 22314

Dan N. Hagedon
Pacific Northwest Laboratories
Battelle Boulevard
Richland, WA 99352

Dr. James Hannon
Lawrence Livermore National Laboratory
P.O. Box 808
L-205
Livermore, CA 94550

Dr. Roger Hansen
HQ AFTAC/TTR
Patrick AFB, FL 32925-6001

Prof. David G. Harkrider
Seismological Laboratory
Division of Geological & Planetary Sciences
California Institute of Technology
Pasadena, CA 91125

Prof. Danny Harvey
CIRES
University of Colorado
Boulder, CO 80309

Prof. Donald V. Helmberger
Seismological Laboratory
Division of Geological & Planetary Sciences
California Institute of Technology
Pasadena, CA 91125

Prof. Eugene Herrin
Institute for the Study of Earth and Man
Geophysical Laboratory
Southern Methodist University
Dallas, TX 75275

Prof. Robert B. Herrmann
Department of Earth & Atmospheric Sciences
St. Louis University
St. Louis, MO 63156

Prof. Lane R. Johnson
Seismographic Station
University of California
Berkeley, CA 94720

Prof. Thomas H. Jordan
Department of Earth, Atmospheric &
Planetary Sciences
Massachusetts Institute of Technology
Cambridge, MA 02139

Prof. Alan Kafka
Department of Geology & Geophysics
Boston College
Chestnut Hill, MA 02167

Robert C. Kemerait
ENSCO, Inc.
445 Pineda Court
Melbourne, FL 32940

Dr. Max Koontz
U.S. Dept. of Energy/DP 5
Forrestal Building
1000 Independence Avenue
Washington, DC 20585

Dr. Richard LaCoss
MIT Lincoln Laboratory, M-200B
P.O. Box 73
Lexington, MA 02173-0073

Dr. Fred K. Lamb
University of Illinois at Urbana-Champaign
Department of Physics
1110 West Green Street
Urbana, IL 61801

Prof. Charles A. Langston
Geosciences Department
403 Deike Building
The Pennsylvania State University
University Park, PA 16802

Jim Lawson, Chief Geophysicist
Oklahoma Geological Survey
Oklahoma Geophysical Observatory
P.O. Box 8
Leonard, OK 74043-0008

Prof. Thorne Lay
Institute of Tectonics
Earth Science Board
University of California, Santa Cruz
Santa Cruz, CA 95064

Dr. William Leith
U.S. Geological Survey
Mail Stop 928
Reston, VA 22092

Mr. James F. Lewkowicz
Phillips Laboratory/GPEH
Hanscom AFB, MA 01731-5000(2 copies)

Mr. Alfred Lieberman
ACDA/VI-OA State Department Building
Room 5726
320-21st Street, NW
Washington, DC 20451

Prof. L. Timothy Long
School of Geophysical Sciences
Georgia Institute of Technology
Atlanta, GA 30332

Dr. Randolph Martin, III
New England Research, Inc.
76 Olcott Drive
White River Junction, VT 05001

Dr. Robert Masse
Denver Federal Building
Box 25046, Mail Stop 967
Denver, CO 80225

Dr. Gary McCartor
Department of Physics
Southern Methodist University
Dallas, TX 75275

Prof. Thomas V. McEvilly
Seismographic Station
University of California
Berkeley, CA 94720

Dr. Art McGarr
U.S. Geological Survey
Mail Stop 977
U.S. Geological Survey
Menlo Park, CA 94025

Dr. Keith L. McLaughlin
S-CUBED
A Division of Maxwell Laboratory
P.O. Box 1620
La Jolla, CA 92038-1620

Stephen Miller & Dr. Alexander Florence
SRI International
333 Ravenswood Avenue
Box AF 116
Menlo Park, CA 94025-3493

Prof. Bernard Minster
IGPP, A-025
Scripps Institute of Oceanography
University of California, San Diego
La Jolla, CA 92093

Prof. Brian J. Mitchell
Department of Earth & Atmospheric Sciences
St. Louis University
St. Louis, MO 63156

Mr. Jack Murphy
S-CUBED
A Division of Maxwell Laboratory
11800 Sunrise Valley Drive, Suite 1212
Reston, VA 22091 (2 Copies)

Dr. Keith K. Nakanishi
Lawrence Livermore National Laboratory
L-025
P.O. Box 808
Livermore, CA 94550

Dr. Carl Newton
Los Alamos National Laboratory
P.O. Box 1663
Mail Stop C335, Group ESS-3
Los Alamos, NM 87545

Dr. Bao Nguyen
HQ AFTAC/TTR
Patrick AFB, FL 32925-6001

Prof. John A. Orcutt
IGPP, A-025
Scripps Institute of Oceanography
University of California, San Diego
La Jolla, CA 92093

Prof. Jeffrey Park
Kline Geology Laboratory
P.O. Box 6666
New Haven, CT 06511-8130

Dr. Howard Patton
Lawrence Livermore National Laboratory
L-025
P.O. Box 808
Livermore, CA 94550

Dr. Frank Pilotte
HQ AFTAC/TT
Patrick AFB, FL 32925-6001

Dr. Jay J. Pulli
Radix Systems, Inc.
2 Taft Court, Suite 203
Rockville, MD 20850

Dr. Robert Reinke
ATTN: FCTVTD
Field Command
Defense Nuclear Agency
Kirtland AFB, NM 87115

Prof. Paul G. Richards
Lamont-Doherty Geological Observatory
of Columbia University
Palisades, NY 10964

Mr. Wilmer Rivers
Teledyne Geotech
314 Montgomery Street
Alexandria, VA 22314

Dr. George Rothe
HQ AFTAC/TTR
Patrick AFB, FL 32925-6001

Dr. Alan S. Ryall, Jr.
DARPA/NMRO
3701 North Fairfax Drive
Arlington, VA 22209-1714

Dr. Richard Sailor
TASC, Inc.
55 Walkers Brook Drive
Reading, MA 01867

Prof. Charles G. Sammis
Center for Earth Sciences
University of Southern California
University Park
Los Angeles, CA 90089-0741

Prof. Christopher H. Scholz
Lamont-Doherty Geological Observatory
of Columbia University
Palisades, CA 10964

Dr. Susan Schwartz
Institute of Tectonics
1156 High Street
Santa Cruz, CA 95064

Secretary of the Air Force
(SAFRD)
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Thomas J. Sereno, Jr.
Science Application Int'l Corp.
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Dr. Michael Shore
Defense Nuclear Agency/SPSS
6801 Telegraph Road
Alexandria, VA 22310

Dr. Matthew Sibol
Virginia Tech
Seismological Observatory
4044 Derring Hall
Blacksburg, VA 24061-0420

Prof. David G. Simpson
IRIS, Inc.
1616 North Fort Myer Drive
Suite 1440
Arlington, VA 22209

Donald L. Springer
Lawrence Livermore National Laboratory
L-025
P.O. Box 808
Livermore, CA 94550

Dr. Jeffrey Stevens
S-CUBED
A Division of Maxwell Laboratory
P.O. Box 1620
La Jolla, CA 92038-1620

Lt. Col. Jim Stobie
ATTN: AFOSR/NL
Bolling AFB
Washington, DC 20332-6448

Prof. Brian Stump
Institute for the Study of Earth & Man
Geophysical Laboratory
Southern Methodist University
Dallas, TX 75275

Prof. Jeremiah Sullivan
University of Illinois at Urbana-Champaign
Department of Physics
1110 West Green Street
Urbana, IL 61801

Prof. L. Sykes
Lamont-Doherty Geological Observatory
of Columbia University
Palisades, NY 10964

Dr. David Taylor
ENSCO, Inc.
445 Pineda Court
Melbourne, FL 32940

Dr. Steven R. Taylor
Los Alamos National Laboratory
P.O. Box 1663
Mail Stop C335
Los Alamos, NM 87545

Prof. Clifford Thurber
University of Wisconsin-Madison
Department of Geology & Geophysics
1215 West Dayton Street
Madison, WS 53706

Prof. M. Nafi Toksoz
Earth Resources Lab
Massachusetts Institute of Technology
42 Carleton Street
Cambridge, MA 02142

Dr. Larry Turnbull
CIA-OSWR/NED
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Dr. Gregory van der Vink
IRIS, Inc.
1616 North Fort Myer Drive
Suite 1440
Arlington, VA 22209

Dr. Karl Veith
EG&G
5211 Auth Road
Suite 240
Suitland, MD 20746

Prof. Terry C. Wallace
Department of Geosciences
Building #77
University of Arizona
Tuscon, AZ 85721

Dr. Thomas Weaver
Los Alamos National Laboratory
P.O. Box 1663
Mail Stop C335
Los Alamos, NM 87545

Dr. William Wortman
Mission Research Corporation
8560 Cinderbed Road
Suite 700
Newington, VA 22122

Prof. Francis T. Wu
Department of Geological Sciences
State University of New York
at Binghamton
Vestal, NY 13901

AFTAC/CA
(STINFO)
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Dr. Michel Bouchon
I.R.I.G.M.-B.P. 68
38402 St. Martin D'Heres
Cedex, FRANCE

Dr. Michel Campilio
Observatoire de Grenoble
I.R.I.G.M.-B.P. 53
38041 Grenoble, FRANCE

Dr. Kin Yip Chun
Geophysics Division
Physics Department
University of Toronto
Ontario, CANADA

Prof. Hans-Peter Harjes
Institute for Geophysic
Ruhr University/Bochum
P.O. Box 102148
4630 Bochum 1, GERMANY

Prof. Eystein Husebye
NTNF/NORSAR
P.O. Box 51
N-2007 Kjeller, NORWAY

David Jepsen
Acting Head, Nuclear Monitoring Section
Bureau of Mineral Resources
Geology and Geophysics
G.P.O. Box 378, Canberra, AUSTRALIA

Ms. Eva Johannisson
Senior Research Officer
National Defense Research Inst.
P.O. Box 27322
S-102 54 Stockholm, SWEDEN

Dr. Peter Marshall
Procurement Executive
Ministry of Defense
Blacknest, Brimpton
Reading FG7-FRS, UNITED KINGDOM

Dr. Bernard Massinon, Dr. Pierre Mechler
Societe Radiomana
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Dr. Svein Mykkeltveit
NTNT/NORSAR
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Prof. Keith Priestley
University of Cambridge
Bullard Labs, Dept. of Earth Sciences
Madingley Rise, Madingley Road
Cambridge CB3 OEZ, ENGLAND

Dr. Jorg Schlittenhardt
Federal Institute for Geosciences & Nat'l Res.
Postfach 510153
D-3000 Hannover 51, GERMANY

Dr. Johannes Schweitzer
Institute of Geophysics
Ruhr University/Bochum
P.O. Box 1102148
4360 Bochum 1, GERMANY