Strong shaking directions from the 18 October 1989 Loma Prieta earthquake and aftershocks in San Francisco and Oakland

John E. Vidale
United States Geological Survey, Menlo Park

and

Ornella Bonamassa
University of California at Santa Cruz

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 interpretation is reinforced by the observation that aftershocks recorded at the CDMG sites show similar directional effects from earthquakes with a variety of mechanisms. This propagational complication suggests that the focal mechanism does not determine the direction of strongest shaking in an earthquake at this range of about 100 km at frequencies above about one Hz.

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 paper 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 paper focuses on finding the transition frequency where earth structure, and mainly near-surface geology, begin to 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).

This paper 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.

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 similarity of the motions at long period shown below agrees with 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 was assumed in our 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, simple particle motion. The primary variation in particle motion across the array seen in Figure 3a is the rotation of the transverse direction of motion clockwise as one considers the stations progressively towards 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 paper.

The observed ground motions show some resemblance to the prediction of the reflectivity simulation. Figure 3 shows the particle motion directions seen in the acceleration records in four 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 3a (as well as the passbands 0.1 to 0.2 and 0.2 to 0.4 Hz, which are not shown) generally agree with the direction expected from the simulation and the directions at frequencies above one Hz shown in Figures 3b, c and d show directions of shaking that bear little relation to the direction expected.

Aftershock recordings at the CDMG sites allow us to check whether the deviations from the particle motion predicted from the mechanism are repeatable and perhaps predictable, as the examples appeared to be in Vidale and Bonamassa (1991) and Bonamassa et al. (1991). The USGS deployed GEOS aftershock recorders at 18 CDMG sites, allowing us to compare strong and weak particle motions. We present only a comparison of the mainshock with one aftershock at station 131 in Figure 4. This aftershock is located underneath the peninsula, and thus has quite a different path and angle of incidence to the station than the mainshock. The primary direction of motion in each passband is similar, and the directions of the largest motions are in various directions. The results suggest that the polarization features seen in Figure 3 are persistent. We also draw a similar conclusion from examination of 5 more aftershocks and one additional station, but space limitations preclude presentation of that data here.

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 (199?). 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 to water waves in the bay, and 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.

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 caused 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 (Aster and Shearer, 1991, Haukkson et al., 1987).

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 transition from source to structural domination 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.

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 also relies on the equally difficult chore 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, our 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

Table 2 - The aftershock presented

Origin time	Lat.	Long.	Depth	Mag.
301 8:35	37° 42'	-122° 33'	9.6 km	2.5

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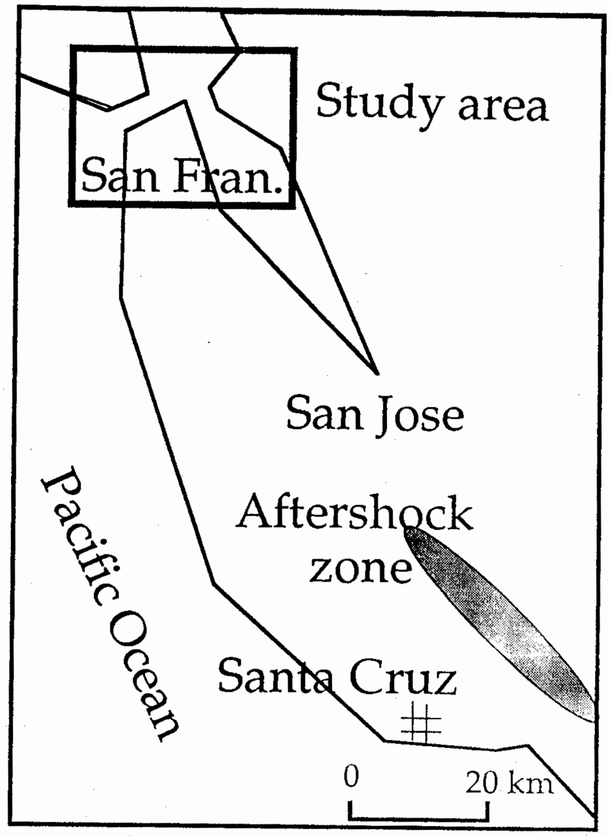

Figure 1. Location of the area of this study in relation to the aftershock zone of the 18 October 1989 earthquake. This study area is shown in more detail in Figure 2.

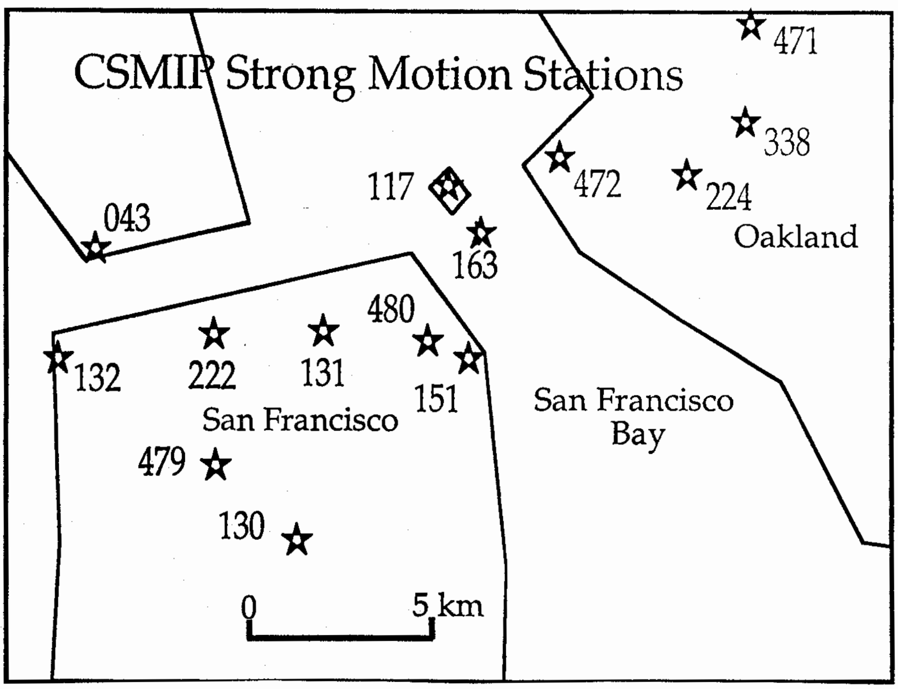

Figure 2. Location of the 13 California Strong Motion Instrumentation Program stations whose recordings are used in this paper. The name and surficial geology of each station are given in Table 1.

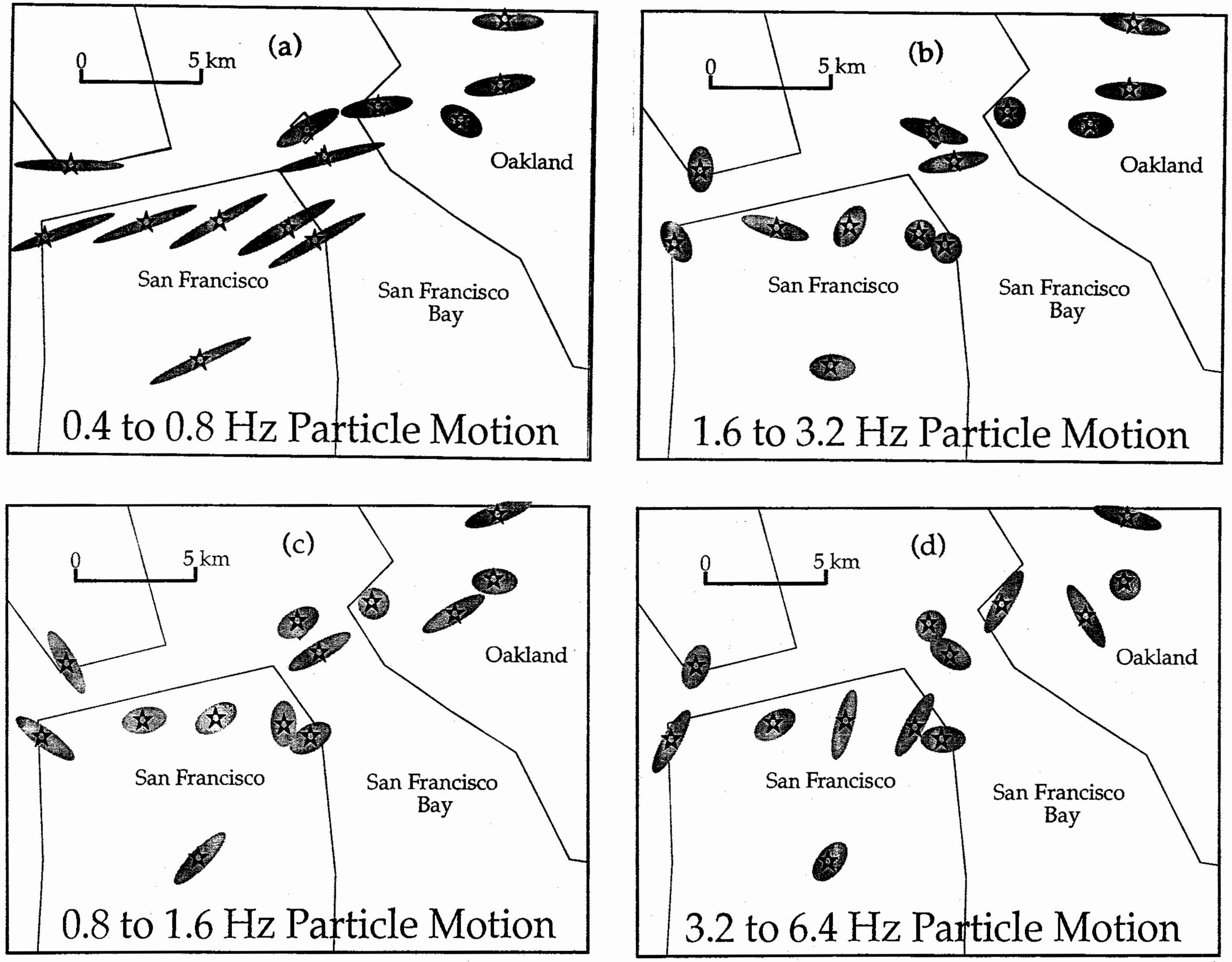

Figure 3. Particle motion diagrams for observations of the 13 stations listed in Table 1 of the 18 October 1989 Loma Prieta earthquake: (a) In the passband from 0.4 to 0.8 Hz (1.25 to 2.5 second period). (b) In the passband from 0.8 to 1.6 Hz. (c) In the passband from 1.6 to 3.2 Hz. (d) In the passband from 3.2 to 6.4 Hz. Two passes of a three pole Butterworth filter were applied.

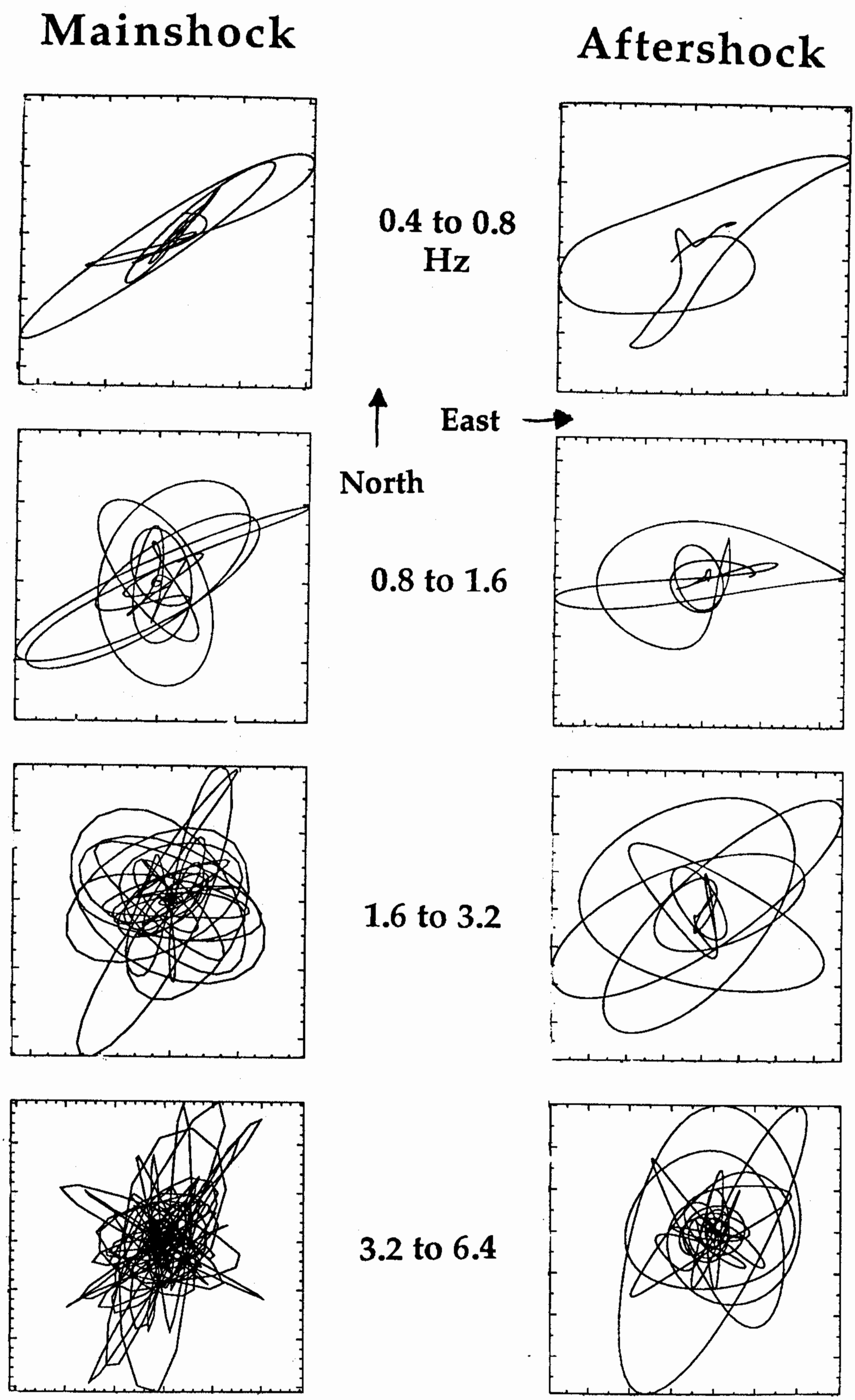

Figure 4. Comparison of weak and strong motions for mainshock and aftershock (Table 2) for all passbands at station 131.