WAVE PROPAGATION CHARACTERISTICS IN THE VICINITY OF THE SAN ANDREAS FAULT

by

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ABSTRACT

A preliminary analysis of seismograms from a number of earthquakes in the Bear Valley region recorded on linear arrays to the east and west of the San Andreas fault reveals that both Pg and Sg wave velocities in the material to the east of the fault are considerably slower than to the west (5.0 and 2.8 km/sec as opposed to 6.0 and 3.5 km/sec, respectively). Relocation of the earthquakes using a crustal model based on these results moves the events on to the fault from their routine locations, which were 2 to 3 km to the west.

DESCRIPTION OF EXPERIMENT

This study of wave propagation characteristics in the Bear Valley region was designed to take advantage of the seismic waves generated by the numerous aftershocks of the magnitude 5 earthquake of February 24, 1972. These events are within the permanent U.S. Geological Survey seismic array, and their hypo centers and origin times can be determined with considerable precision. Six 3-component portable seismic units of the type described by Eaton and others (1970) were used to form linear arrays perpendicular to the San Andreas fault in two phases. During the first phase, in March, 1972, the instruments were deployed west of the fault in a linear array extending from Bear Valley to Carmel Valley. After a week of recording, one of the station (SCR) was moved to a new site (RC2), so that a total of seven sites were occupied. During the second phase in November and December, 1972, five of the instruments were deployed in a linear array east of the fault between Bear Valley and the east ern edge of the Diablo Range. The sixth instrument was re-installed at the Parks Valley (PV) site used in the first phase to serve as a tie between the two profiles. (See Figure 1).

The superficial geology crossed by the profile shows granitic rocks to the west of the fault and Franciscan rocks to the east (Jennings and Strand, 1958). Complications of this basic pattern however, may have important influences on seismic wave propagation. The granites, for example, may not extend

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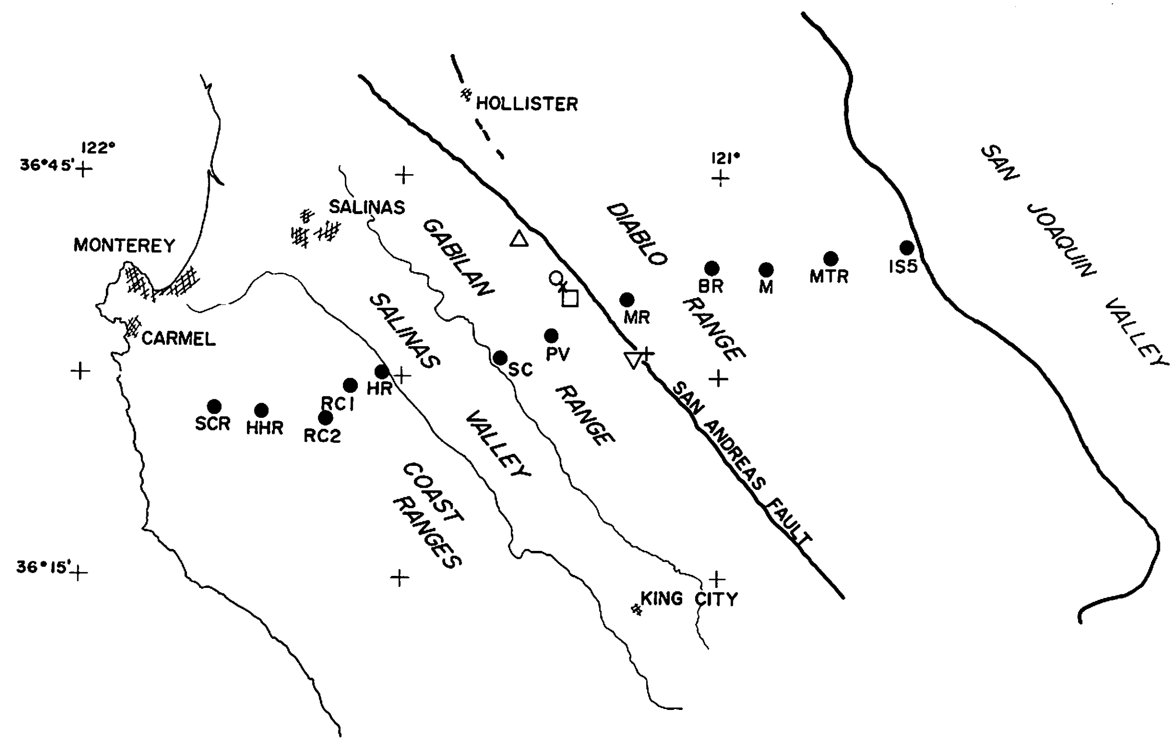

Fig. 1. Location map. Black dots represent recording locations; open symbols show routine locations of several of the events used in this study.

to a great depth (Yeats, 1968; also see discussion following Stewart, 1968). The boundary between the Mesozoic rocks of the Great Valley sequence and the Franciscan formation is complicated, and ultrabasic intrusions, present to the north and south in the Diablo Range, may be present at depth below the profile.

DATA

Record sections for events recorded to the west and east show some distinctive differences (Fig. 2 and 3). Some of these differences, such as the emergent character of P-waves on the west section or the S-waves on the east section are probably due to the radiation pattern of the two earthquakes from which these particular record sections were generated. The difference in the dominant frequency of waves recorded on either side of the fault, however, persists on records from all the events we have examined. That this is not a feature of the source alone can be easily seen by comparing seismograms recorded at the common station (PV) for the two events in Figs. 2 and 3. Preliminary calculations of spectra for selected seismograms confirm the differences in spectral content of waves propagating to the east and to the west of the fault. Whether this difference in frequency character can be explained wholly by material attenuation must wait for a careful study of the spectra at a number of sites.

Another distinguishing characteristic of the two profiles is the remarkable difference in the travel times for the P and S waves. This is best seen in Figs. 4 and 5, each of which are based on recordings of 8 to 10 events. To the west, P and S waves are clearly traveling at horizontal phase velocities of 6.0 and 3.5 km/sec respectively. The S-wave data to the east show more scatter, but travel time curves with P and S-wave velocities of 5.0 and 2.8 km/sec are well defined by the data beyond about 12 km.

The P and S velocities to the west are consistent with those found by Hamilton et al (1964) and Filson (1970). The P-wave velocities to the east, however, are significantly lower than reported in earlier studies, and in particular those found by Stewart (1968) from N-S refraction lines in the Diablo range to the north of our profile.

The east profile data between 5 and 12 km suggests that with respect to the trend defined by more distant stations, a consistent delay occurs in the waves arriving at MR. Although some of the apparent delay is caused by the use of epicentral rather than hypocentral distance in the plots, this cannot account for all of the observed difference. According to the routine locations, the earthquakes being recorded are from 5-10 km in depth, with epicenters to the west of the fault zone. MR is to the east of the fault zone and thus the delay suggests that the fault zone contains material of anoma-

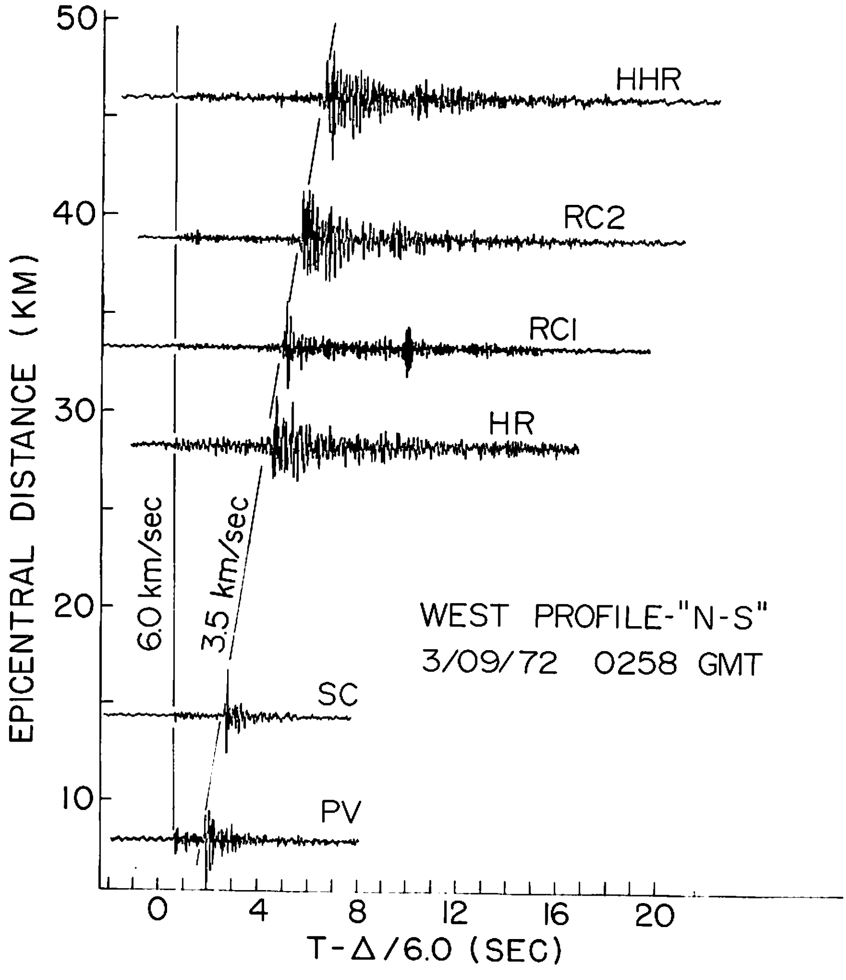

Fig. 2 Record section of north-south component-west profile. Amplitudes are normalized independently.

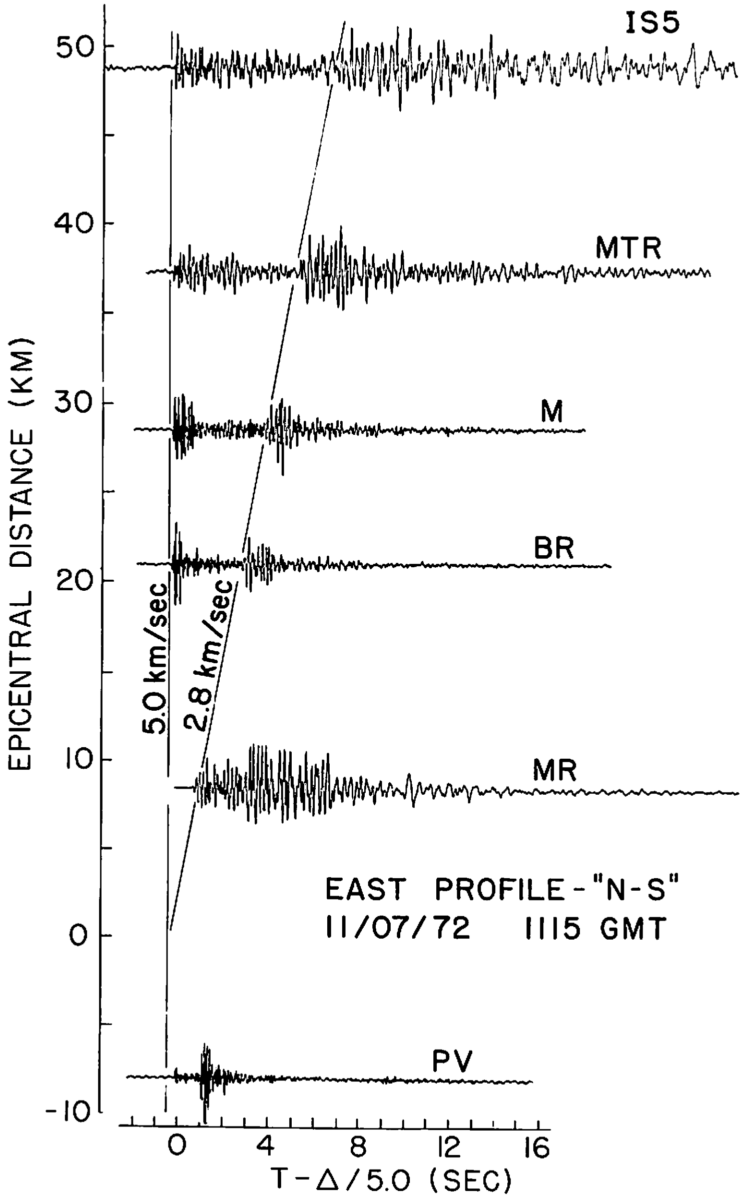

Fig. 3 Record section of north-south component-east profile. Amplitudes are normalized independently

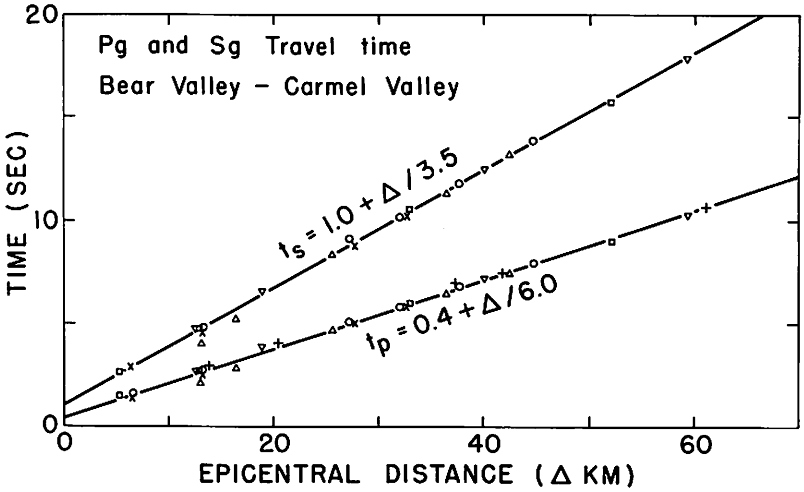

Fig. 4 Travel times of P and S waves for a number of events recorded on the west profile. Origin times and event locations were taken from the routine USGS locations. Poisson's ratio is 0.24 and $V_p/V_s = 1.71$.

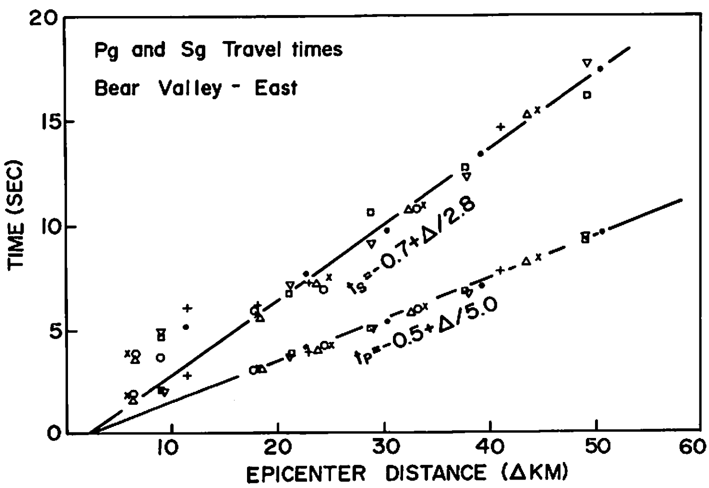

Fig. 5 Travel times for P and S waves recorded on the east profile. Poisson's ratio is 0.27 and $V_p/V_s = 1.79$.

lously low velocity which extends to a significant depth. Evidence for such material extending to at least 5 km was found by Mayer-Rosa (1973) associated with the Calaveras fault to the north of the present study area. Mayer-Rosa (1973) used travel time delays associated with the fault zone to explain the consistent bias of routine earthquake locations to the east of the mapped surface trace of the Calaveras fault.

RELOCATION OF EPICENTERS

Routine hypocenter determination of earthquakes occurring within the permanent USGS net show 2 to 5 km systematic displacement of epicenters on one side or the other of the surface traces of most of the major strike-slip faults in central California, and in particular this is shown by the events used to construct the travel time data in Figures 4 and 5. The negative intercept times for the P-wave travel time curve east of the fault suggest a systematic mislocation of the earthquakes. (Note that this suggestion is independent of the delay observed at MR). This observation, together with the difference in P-wave velocities described above, led us to try a relocation of the events assuming a crustal model with a velocity contrast across the fault. At this stage, we have not included a delay produced by waves travelling through the fault zone, and thus in this part of the fault system we are testing a different explanation for the observed bias than put forth by Mayer-Rosa (1973). In Mayer-Rosa's case the Franciscan formation was present on either side of the fault, and he found no obvious differences in crustal structure.

Epicentral locations and depths from routine processing of data from USGS permanent stations are shown by small dots at the tail of arrows in Fig. 6 for several events. The model used in these locations has three layers, and allows station corrections through the introduction of a top layer of variable thickness (Lee and Lahr, 1972; Wesson et al, 1973). The consistent location to the west of the fault is clearly shown in Fig. 6. On the basis of travel time data found in our study, a model with 25 km crustal thickness and P velocities of 6.0 and 5.0 km/sec on the west and east side of the fault was used to relocate the same events. We cannot tell from our data to what depth the 5.0 km/sec material extends; the 25 km is a numerical convenience, and does not affect the results. The combined data from both our temporary stations east of the fault and PV with those from the permanent network gave the location shown at the head of the arrow. Data from the permanent network alone gave the locations shown by open circles. No station corrections were used in these relocations. It is clear from Figure 6 that this model moves the earthquakes essentially to the surface expression of the San Andreas fault. This result fits the intuitive feeling of most people that the earthquakes must lie on the fault, although it must be emphasized that the model we used in the relocation is very simple. All we have shown in fact, is that if this model were an adequate representation of the real world, the

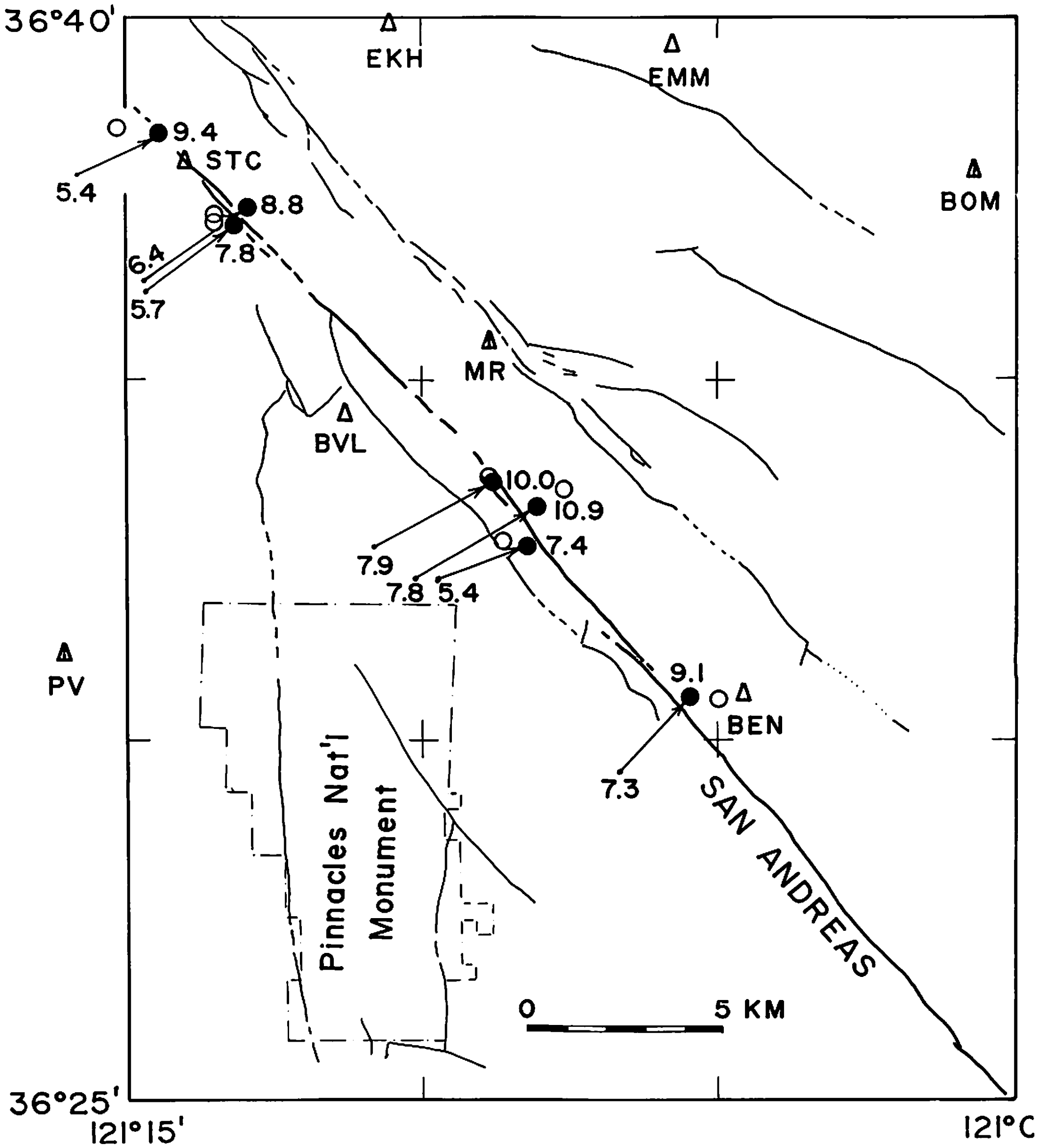

Fig. 6 Dependence of earthquake locations on velocity contrasts across the fault. The open triangles indicate the closest permanent stations; triangles with a vertical slash represent the closest portable stations. Depths of the location are shown by the numbers.

events would be located on the fault. The most objectionable feature of the model is the depth to which the 5.0 km/sec material is assumed to extend east of the fault. The earthquakes used in this study, however, are 5 to 10 km deep and the 5.0 km/sec P-wave velocity is well-defined to distances of at least 50 km. Thus, it seems difficult to avoid the conclusion that material of this velocity must extend to a significant depth. Waves propagating down a refractor dipping to the east would produce an anomously low apparent horizontal phase velocity, but such a refractor would need to be on the order of 10 km deep near the fault. We plan on performing a controlled experiment with artificial sources to investigate this problem.

Even if the systematic bias in the earthquake locations are a result of very different crustal structure on either side of the fault near Bear Valley we cannot appeal to this in other parts of central California. Both Stewart's refraction lines and Mayer-Rosa's work indicate that in the Diablo Range to the north the crustal velocities are similar to those we found on the west profile. Clearly, more work needs to be done in this very interesting area.

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