THE INFLUENCE OF CRITICAL MOHO REFLECTIONS ON STRONG GROUND MOTION ATTENUATION IN CALIFORNIA

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ABSTRACT

Using strong motion recordings from the 1987 Whittier Narrows and 1989 Loma Prieta earthquakes, we show evidence that critical reflections from the lower crust control the attenuation of peak acceleration at distances beyond the critical distance, causing a flattening of the attenuation curve. Depending on the focal depth of the earthquake and the crustal thickness, this flattening can begin at distances from the source as close as 40 km.

INTRODUCTION

Empirical attenuation relations assume a simple monotonic decrease of peak ground motions with distance. However, models of wave propagation in a layered crust predict that the attenuation relation for a specific earthquake will have a more complex form that is controlled by the focal depth of the earthquake and the crustal structure (Burger *et al.*, 1987), as shown schematically in Figure 1. At close distances, peak horizontal ground motions are controlled by direct upgoing shear waves. However, as distance increases, the reflections of shear waves from interfaces in the lower crust (such as the Conrad and Moho discontinuities) reach the critical angle and undergo total internal reflection, producing large amplitude arrivals (S_cS and S_mS) that cause a flattening of the attenuation relation. Strong evidence of the effect of critical reflections on the attenuation of strong ground motion from the 1988 Saguenay, Quebec earthquake was described by Somerville *et al.* (1990).

The present study was motivated by the hypothesis that critical reflections from the Moho caused peak ground motion amplitudes to remain approximately uniform in the epicentral distance range of about 50 to 100 km from the 1989 Loma Prieta earthquake (Somerville and Yoshimura, 1990). Strong ground motion levels on rock and alluvium in the central San Francisco Bay area during the Loma Prieta earthquake exceeded those of empirical attenuation relations. According to the hypothesis, these large ground motions were critical Moho reflections, and contributed about equally with impedance-contrast amplification effects in generating destructive ground motions at soft soil sites in the central San Francisco Bay area.

The principal objective of this paper is to examine evidence for the influence of critical reflections on strong ground motion attenuation in other regions of California. We have done this by examining the strong motion recordings of the following large earthquakes: 1952 Kern County; 1968 Borrego Mountain; 1971 San Fernando; 1985 Coalinga; 1986 North Palm Springs; and 1987 Whittier Narrows. In this paper, we concentrate on results from the 1987 Whittier Narrows and 1989 Loma Prieta earthquakes.

1989 LOMA PRIETA AFTERSHOCKS AND MAINSHOCK

We see the influence of critical reflections most clearly in recordings of aftershocks of the Loma Prieta earthquake, because aftershocks have briefer source processes than the mainshock and therefore provide a clearer representation of wave propagation effects. On the left side of Figure 2, we show a profile of the east component of velocity recorded at stations northwest of the 1:30 am aftershock of November 5, 1989. The data are described by Mueller and Glassmoyer (1990); all of the recordings have absolute times, allowing them to be aligned in time with a travel time reduction of 3.5 km/sec. Superimposed on the profile are travel times for the direct S, Conrad

critical reflection (S_cS), and Moho critical reflection (S_mS). On the right side of Figure 2, we show a profile of synthetic seismograms for the San Francisco Peninsula (along an azimuth of 320°) generated by the reflectivity method.

Except at the closest recording station on the left side of Figure 2, the direct arrival has a small amplitude relative to later arrivals, and the first conspicuous phase has the arrival time of the critical Conrad reflection S_cS. The largest phase at all stations except the closest station occurs soon after the expected arrival time of the critical Moho reflection S_mS. These aftershock recordings indicate that at distances beyond about 50 km, the recorded peak ground motion amplitudes are controlled not by the direct S wave but by the wave that is critically reflected from the Moho. This interpretation is confirmed by the synthetic seismograms shown on the right side of Figure 2: beyond about 45 km, the largest amplitudes are associated with the Moho reflection. Analyses of other aftershocks show the same effects. The large motions that follow the Moho reflection in the recorded data suggest that there may be strong lateral variations in crustal structure along the propagation path.

Evidence of the influence of critical Moho reflections on the attenuation of strong ground motion from the mainshock was described by Somerville and Yoshimura (1990). We have updated that study using a detailed rupture model of the Loma Prieta earthquake derived by Wald *et al*. (1990). A record section of recorded accelerograms was compiled using all accelerograms to the north of the epicenter (i.e. in the San Francisco Bay region) that have known trigger time (Figure 3, left side). The recordings are from a variety of site conditions, as annotated in the figure and discussed further below, and are copies of film records published by Shakal *et al*. (1989) and Maley *et al*. (1989). On the right side of Figure 3, we show a profile of accelerograms simulated using the procedure described by Wald *et al*. (1988). The simulated accelerograms were generated using a crustal structure model having a surface shear wave velocity of 1 km/sec, appropriate for soft rock or stiff soil conditions.

At distances beyond 50 km, the onset of the largest accelerations at each station coincides with the arrival time of the critical Moho reflection S_mS in both the recorded and simulated accelerograms. The moveout of this onset with distance follows the S_mS arrival time curve and not that of direct S. The duration of strong motion following the S_mS arrival time curve is about 5 seconds, which is compatible with the 6-second duration of the source observed teleseismically (Nabelek, 1990; Kanamori and Satake, 1990). Both the recorded and synthetic accelerogram profiles suggest that, beyond about 50 km, the peak accelerations are associated not with the direct upgoing shear wave (S) but with the shear wave that has been critically reflected from the Moho (S_mS).

The average horizontal peak accelerations of the simulated accelerograms show a trend similar to that of the recorded accelerograms, as demonstrated in Figure 4(a). In this figure, the northerly profile of stations of Figure 3 has been augmented by stations that lie within 30 km of the epicenter. The peak accelerations attenuate normally to about 40 km, but then do not attenuate further until reaching an epicentral distance of 80 km. While site conditions are presumably responsible for the larger amount of scatter in recorded peak accelerations (compared with the simulated ones) at a given distance, it does not appear that site conditions explain the overall lack of attenuation between 40 and 80 km in both the recorded and simulated values. Instead, it appears that the shape of the attenuation curve is due to critical reflections, with the scatter of recorded peak values about this shape attributable in part to local site effects. The standard error of the simulations in predicting the observed response spectra is shown in Figure 5 for near-source and distant stations.

1987 WHITTIER NARROWS EARTHQUAKE

A northerly profile of strong motion recordings of the 1987 Whittier Narrows earthquake having absolute times was compiled (Figure 6, left side), and a corresponding profile of synthetic seismograms was generated using the reflectivity method (Figure 6, right side). As seen teleseismically, the Whittier Narrows earthquake had a brief but complex source function consisting of a small event followed one second later by a larger event (Bent and Helmberger, 1988). In the

distance range of zero to 20 km, the peak velocity is associated with the time of the direct S arrival from this later event in both the recorded and synthetic seismograms. However, beyond about 37 km, large arrivals also occur in the data at times corresponding to reflections from the lower crust (the mafic and Moho layers); the synthetic seismograms show that the Moho reflection becomes critical at about 70 km. For example, at Lancaster, which is at 70 km, the reflections from the lower crust produce significant ground motion amplitudes in both the recorded and synthetic seismograms. The amplitudes of the reflected arrivals increase with distance, until at Rosamond (at a distance of 86 km) the largest arrival has the time of the critical Moho reflection, and the amplitude of the direct arrival is relatively small. Thus ground motion amplitudes beyond the critical distance are controlled by critically reflected waves, causing the peak acceleration to have very gradual attenuation in the distance range of 50 to 90 km (Figure 4(b)).

An analysis of the attenuation of the entire set of strong motion recordings of the Whittier Narrows earthquake was made by Campbell (1988). His plots of residuals in peak acceleration for deep soil and hard rock are shown in the top and middle panels respectively of Figure 7. For both data sets, the residuals are predominantly negative in the distance range of 30 to 60 km, become zero on average at about 70 km, and are positive between 80 and 110 km. This indicates that the functional form that he used to model the attenuation of the Whittier Narrows data does not fit the data. The derived attenuation relation overpredicts the data by about one to two standard deviations in the distance range of 30 to 60 km, and underpredicts the data by a similar amount in the distance range of 80 to 110 km. This consistent with the expected effect of critical reflections on the attenuation of strong ground motion. A similar pattern of residuals was obtained for rock site recordings of the 1989 Loma Prieta earthquake (Boore et al., 1989), as shown in the lower panel of Figure 7. Here, the crossover from negative to positive residuals occurs at a closer distance (about 45 km compared with 70 km). This is what we would expect from the shallower depth of the Moho beneath the Loma Prieta earthquake (25 km) compared with the Whittier Narrows earthquake (32 km); the two events have comparable centroid depths of about 13 km. In both cases, the crossover distance corresponds to the critical distance for the Moho reflection.

SUMMARY OF EFFECTS ON GROUND MOTION ATTENUATION IN CALIFORNIA

We have analyzed profiles of accelerograms from five additional large California earthquakes. For the older events, the strong motion recordings did not have absolute time, and so it is more difficult to identify critical reflections in these recordings. Also, for the larger events, the source duration is sufficiently long that it obscures the individual phases that we are trying to identify. Nevertheless, we find evidence of critical reflections (in large, late wave arrivals that may be S_mS, and in the flattening of attenuation curves along specific profiles) in most of the events studied: the 1968 Borrego Mountain earthquake; the 1971 San Fernando earthquake; the 1983 Coalinga earthquake; and the 1986 North Palm Springs earthquake. This suggests that critical reflections may influence the attenuation of strong ground motion throughout California.

A contour map of crustal thickness in California is shown in Figure 8 (Mooney and Weaver, 1990). In the region of the San Andreas fault system extending from north of the Transverse Ranges to north of the San Francisco Bay area, the depth to the Moho is 25 +/- 1 km. In the region south of the Transverse Ranges, including almost all of southern California, the crustal thickness is 30 +/- 2 km. The relatively uniform crustal thickness in each of these regions, and the significant contrast in crustal thickness between them, suggests that there could be differences in ground motion attenuation between these two regions. Specifically, for a given focal depth, the effect of critical reflections should occur at closer distances and be larger in northern California than in southern California, as suggested by the residuals for the Loma Prieta and Whittier Narrows earthquakes discussed above. As a rule of thumb, we expect that the distance at which the attenuation curve for a given earthquake begins to flatten can be estimated from the critical distance of the S_mS phase, which can easily be calculated from the crustal structure and the focal depth of the earthquake.

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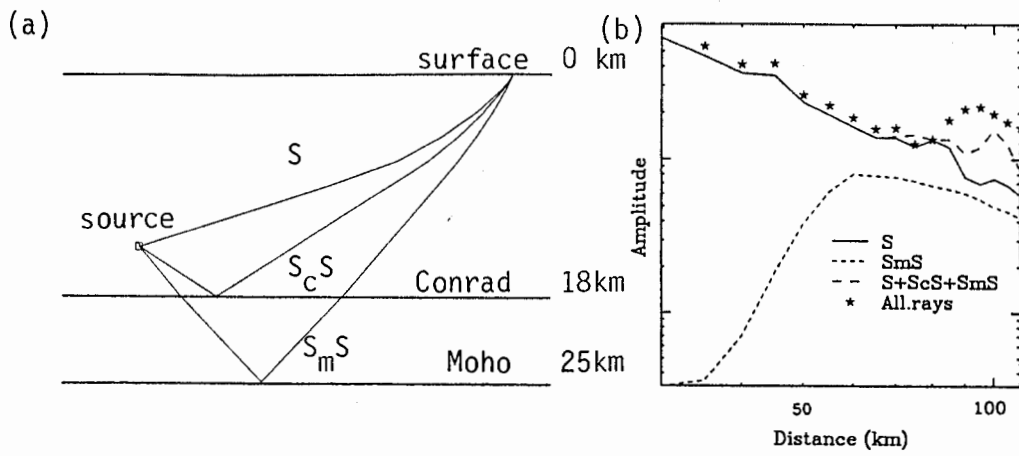
Figure 1. (a) Schematic model of wave propagation in central California showing the direct S wave and S waves reflected from the Conrad and Moho interfaces. (b) Schematic attenuation of individual rays in the wave propagation model shown in (a).
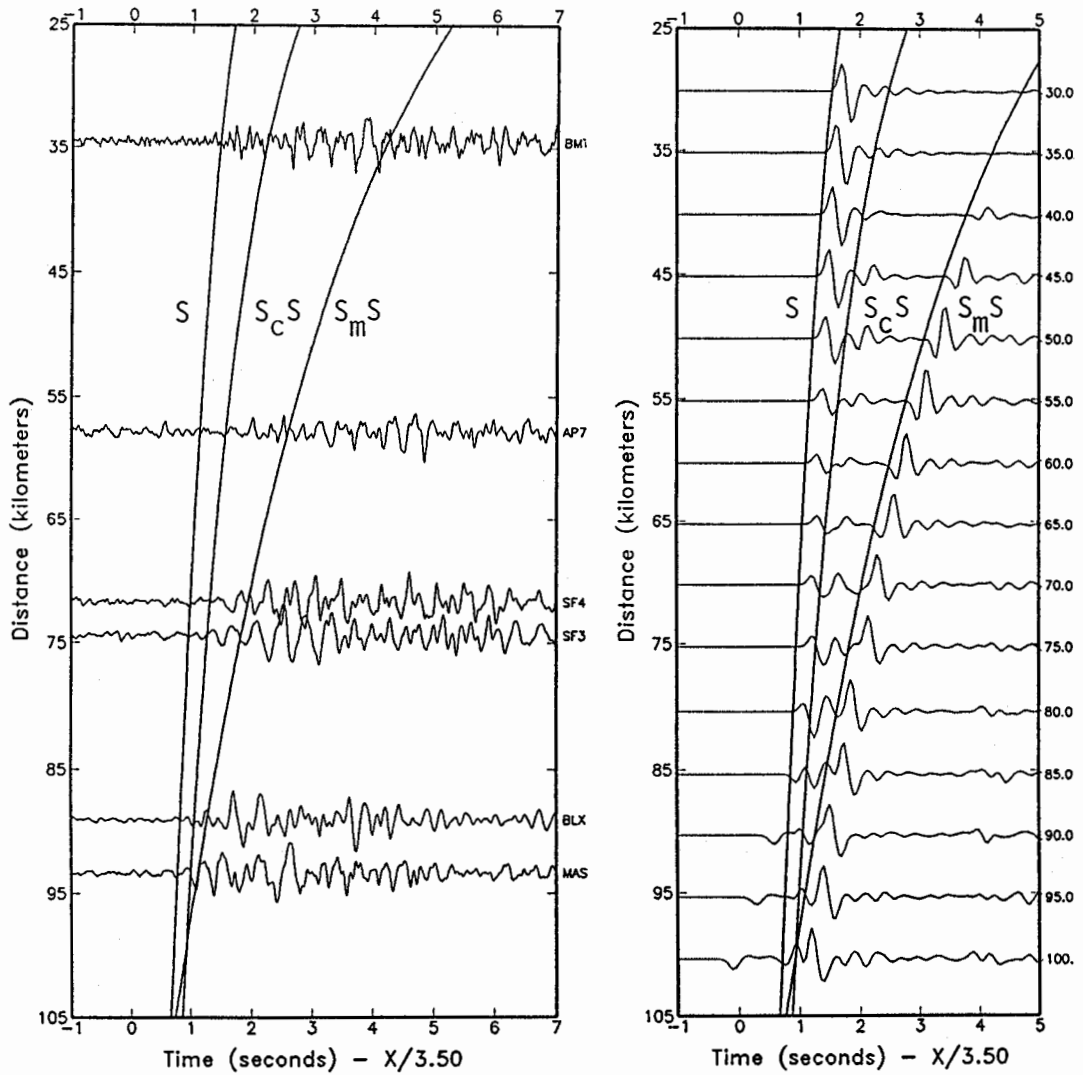

Figure 2. Profiles of recorded (left) and synthetic (right) tangential velocity of the 1:30 am, November 5 aftershock of the 1989 Loma Prieta earthquake, compiled using epicentral distance and a travel time reduction of 3.5 km/sec, and normalized to peak value. Travel time curves for the direct S, Conrad reflection (S_cS, 18 km depth) and Moho reflection (S_mS, 25 km depth) phases are shown. USGS station abbreviations are annotated on right side of traces.

2-5

RECORDED on various site conditions

SIMULATED for soft rock or stiff soil
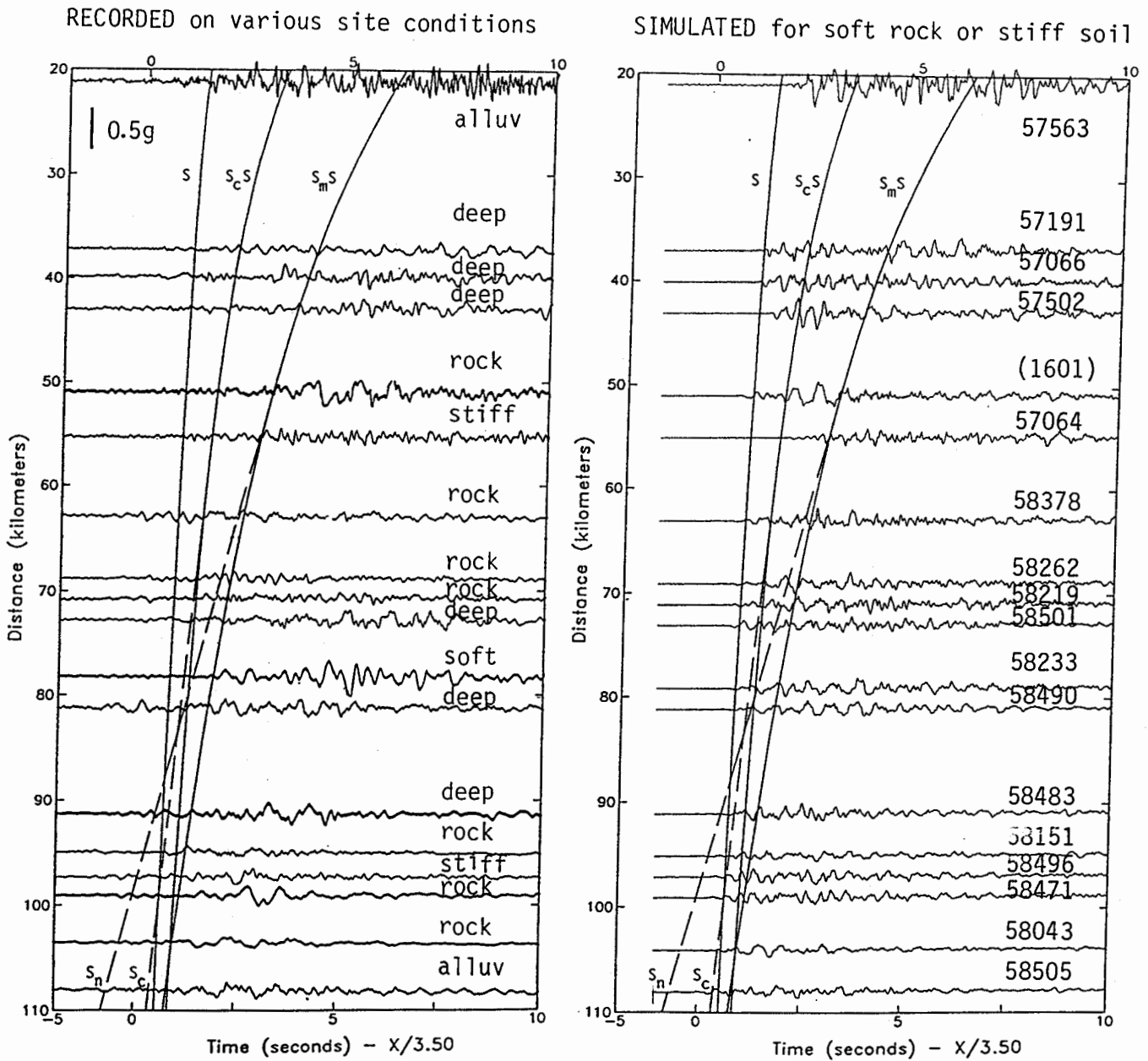
Figure 3. Profiles of recorded (left) and simulated (right) accelerograms of the 1989 Loma Prieta earthquake, compiled using epicentral distance and a travel time reduction of 3.5 km/sec, unnormalized. Travel time curves for the direct S, Conrad reflection (S_cS, 18 km depth) and Moho reflection (S_mS, 25 km depth) phases are shown. Soil conditions are annotated on recorded accelerograms, and CSMIP (and USGS in parentheses) station numbers are annotated on simulated accelerograms.

(a) Loma Prieta
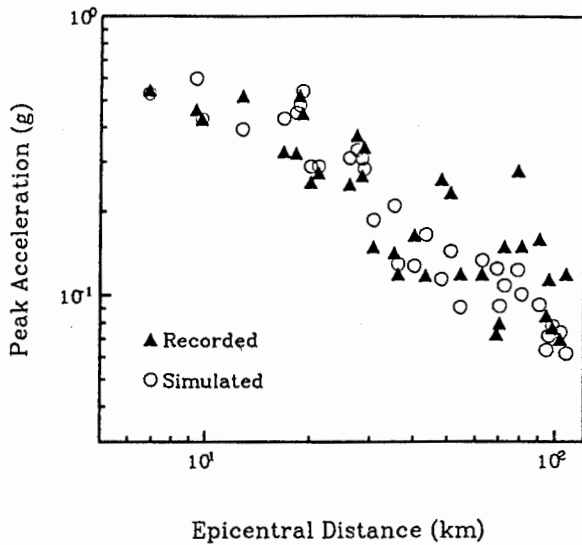
(b) Whittier
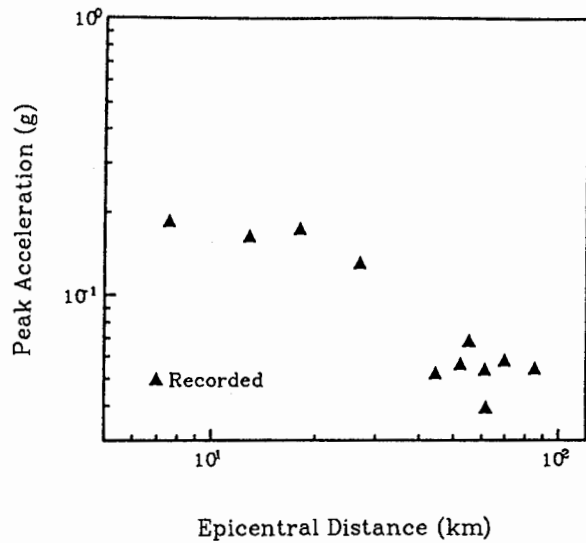
Figure 4. (a) Comparison of recorded and simulated peak accelerations as a function of distance from the Loma Prieta epicenter for recordings shown in Figure 3 and additional near-source recordings. (b) Peak tangential acceleration plotted as a function of distance from the Whittier Narrows earthquake for recording stations on the northerly profile shown in Figure 6.
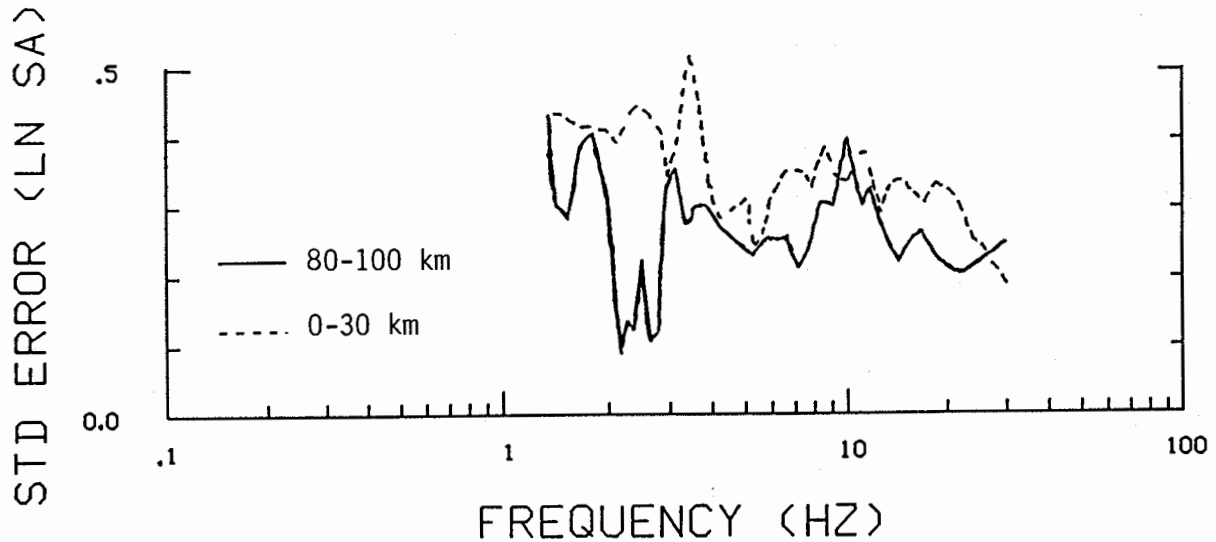

Figure 5. Natural logarithm of standard error of simulations in predicting the observed response spectrum of the Loma Prieta earthquake at near-source stations (dashed line) and at stations in San Francisco and Oakland (solid line).
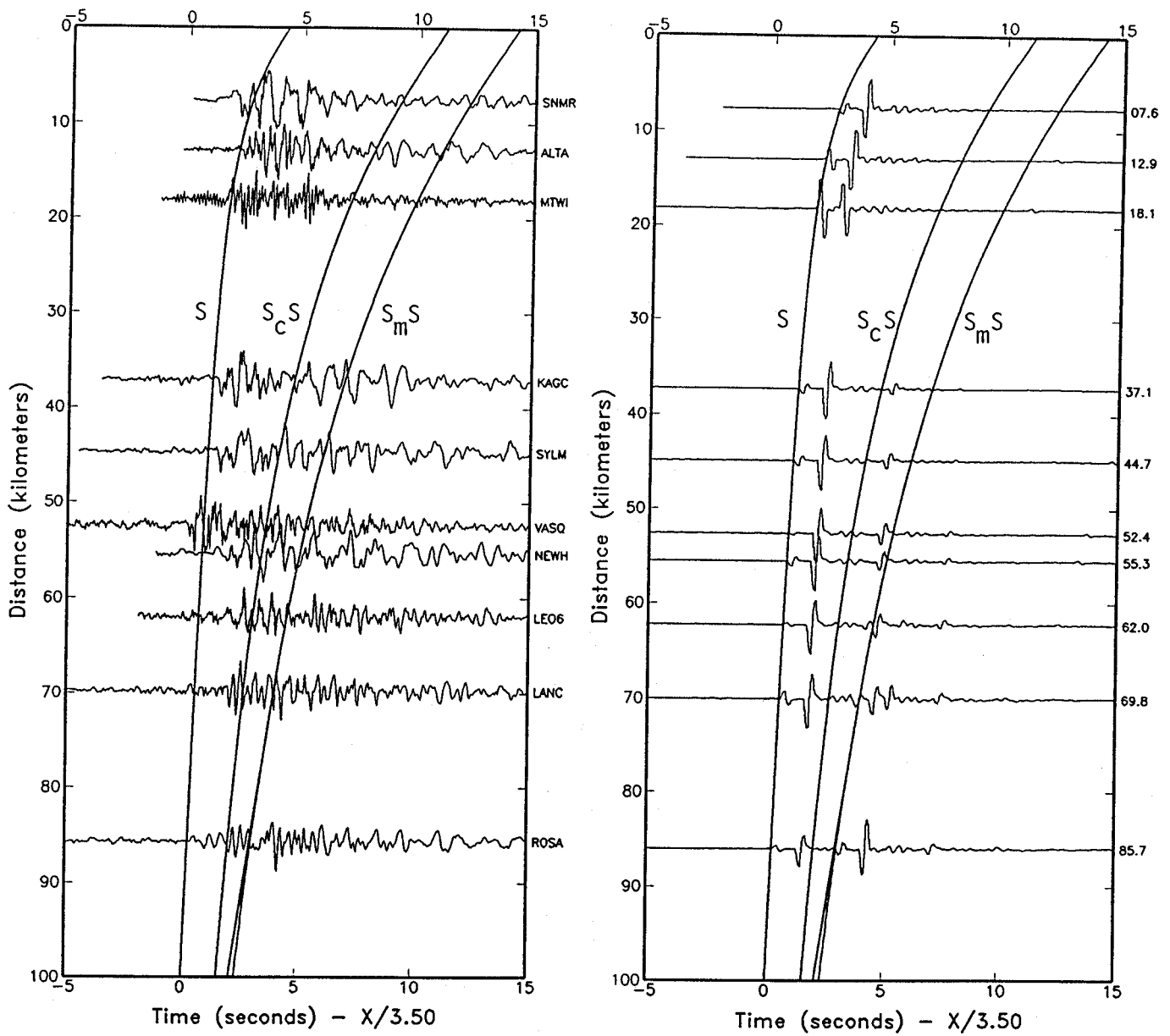
Figure 6. Profiles of recorded (left) and synthetic (right) tangential component ground velocity of the Whittier Narrows earthquake, compiled using epicentral distance and a travel time reduction of 3.5 km/sec, and normalized to the peak value. Travel time curves are shown for the direct S wave and waves reflected from the Conrad, mafic, and Moho interfaces.
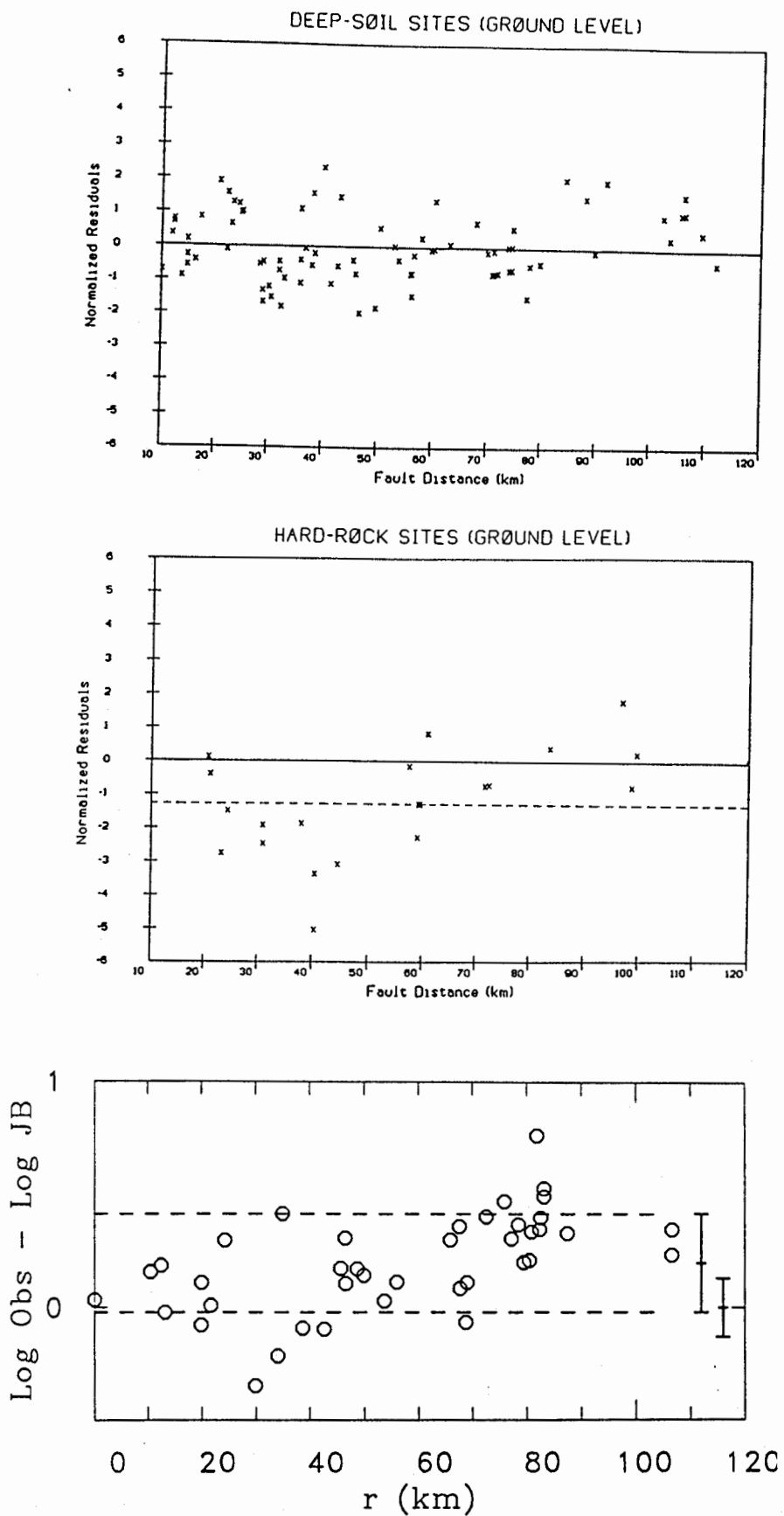
Figure 7. Top and Center: Residuals of recorded peak acceleration from the regression relations derived from deep soil and hard rock recordings of the 1987 Whittier Narrows earthquake, after Campbell (1988); Bottom: Residuals of peak acceleration from the regression relation of Joyner and Boore (1988) for rock site recordings of the 1989 Loma Prieta earthquake, after Boore et al. (1989).
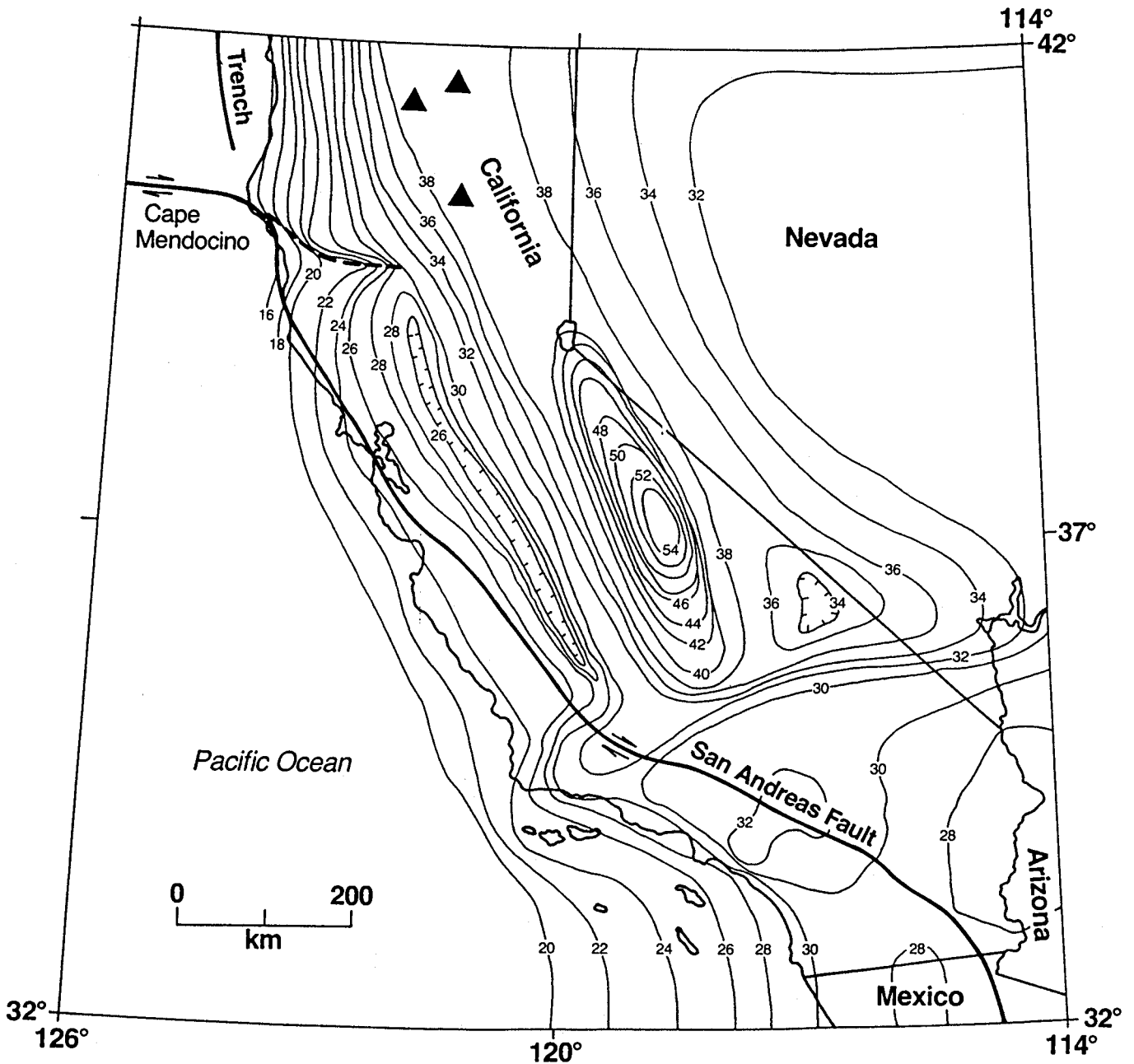
Figure 8. Contour map of crustal thickness (depth to Moho) for California. The thickness is 25 +/- 1 km in the San Francisco Bay region and 30 +/- 2 km in southern California. Source: Mooney and Weaver (1990).