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# THE TEMPORAL VARIATION OF TELESEISMIC *P*-RESIDUALS FOR STATIONS IN SOUTHERN CALIFORNIA

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ABSTRACT

**Teleseismic *P* residuals have been monitored as a function of time at 13 stations in southern California during the 5-year period 1972 to 1976. These residuals, when normalized to minimize common path and source effects, and corrected for the marked azimuthal dependence of residuals in southern California, show no significant variation. This indicates that no detectable velocity changes have occurred during this time in the vicinity of the stations monitored. It is estimated that a velocity change of ~9 per cent occurring over a path length of 10 km and lasting for at least 6 months should be resolvable. Either such changes have not taken place in the region monitored, or any velocity anomalies are confined to a small depth range in the crust, and are poorly sampled by teleseismic waves.**

## INTRODUCTION

The occurrence of seismic velocity changes prior to, and presumably associated with, earthquakes has been documented by a number of authors (see, for example, Savarensky, 1968; Semenov, 1969; Aggarwal *et al.,* 1973; Whitcomb *et al.,* 1973; Ohtake, 1973; and Wyss and Johnston, 1974). The investigation of temporal variations in velocity in a tectonically active area is thus of value, both as a possible means of predicting future earthquakes, and in order to establish a "background" level of fluctuations not associated with large earthquakes.

The USGS-Caltech Southern California Seismograph Network provides a good source of such data, particularly since 1972 when develocorder recording was introduced, allowing greater measurement accuracy. In addition, the presence of large changes in elevation in the vicinity of Palmdale (Castle *et al.,* 1976) makes this an area of special interest.

Previous searches for possible precursory velocity changes in California have met with limited success. Cramer (1976) was unable to detect any significant variations prior to the 1974 Thanksgiving Day (Hollister) earthquake, and Bolt (1977) and Cramer *et al.* (1977) found no changes in travel times from Nevada test site blasts to Oroville (ORV) prior to the 1975 Oroville earthquake. However, Cramer *et al.* (1977) reported a 0.1 sec delay in residuals at ORV for Novaya Zemlya explosions prior to the Oroville event. Small changes in *P* velocity have been resolved in studies of travel times from quarry blasts in southern California: Kanamori and Hadley (1975) reported changes of ~3 per cent, and Kanamori and Fuis (1976) observed variations of ~1 per cent prior to the 1975 Galway Lake and Goat Mountain earthquakes. However, Whitcomb *et al.* (1973) found a change of 10 per cent in $Vp/Vs$ and ~19 per cent in $Vp$ prior to the 1971 San Fernando earthquake, and Stewart (1973) concluded from data obtained from local and teleseismic events that a change of up to 30 per cent in $Vp$ may have occurred in the source region of the 1973 Point Mugu earthquake. Changes of 10 to 20 per cent were also observed in the vicinity of Riverside during the period 1964 to 1969 by Kanamori and Chung (1974), but they concluded that this was not obviously related to seismic activity in the area.

In this study, teleseismic residuals were monitored for 13 stations in southern

California during the period 1972 to 1976 in order to investigate the magnitude of any velocity fluctuations, and the possible existence of premonitory changes. Six of these stations are in the vicinity of the Palmdale uplift. The data can also be used to establish a baseline for comparison with any changes that may occur in the future.

## METHOD

Teleseismic residuals have been used to investigate premonitory velocity changes by a number of authors, including Wyss and Johnston (1974) and Cramer (1976). This method, which is discussed in detail by Engdahl *et al.* (1977), has the advantage that the locations of the sources used, and hence the theoretical travel times, do not depend on a detailed knowledge of local structure in the area of the study, as in the

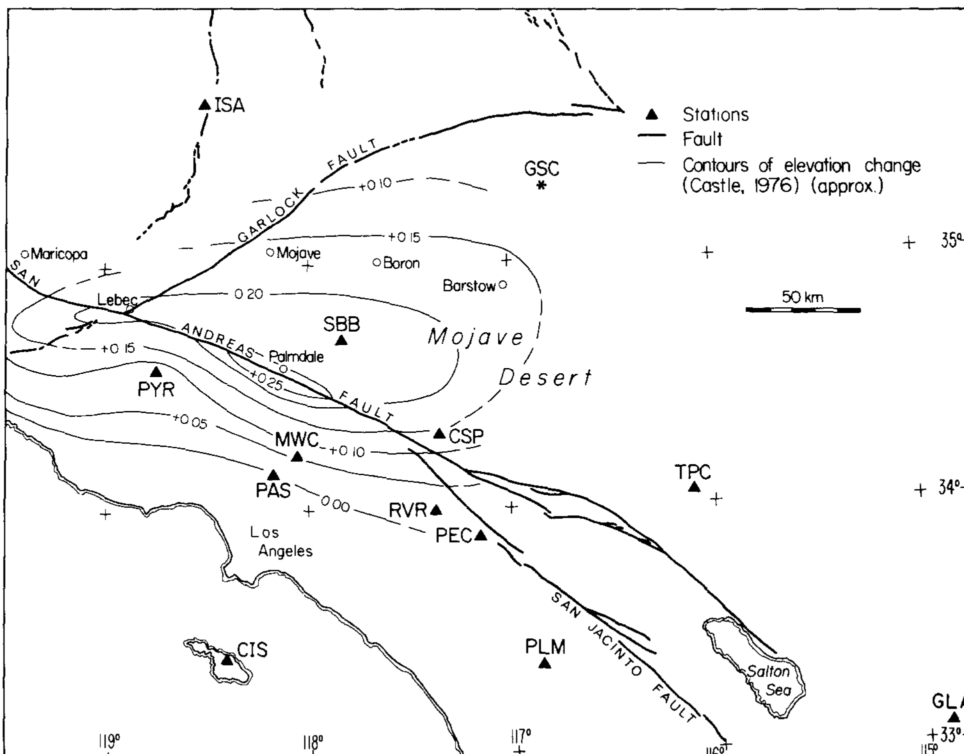


FIG. 1. Map showing the distribution of stations used in this study. Also plotted are the approximate contours of uplift, in meters, observed in the Palmdale region (Castle *et al.*, 1976).

case when local earthquakes are used. However, the effects of source mislocation and inhomogeneities along the travel path can cause considerable scatter, and must be minimized by normalization. This is usually achieved by taking relative residuals with respect to one or more of the stations studied, or by averaging.

The locations of the stations used in this study are shown in Figure 1, together with approximate contours of elevation in the vicinity of Palmdale, as reported by Castle *et al.* (1976). All the seismograms were recorded on develocorder film with WWVB time traces at the top and bottom, and were read at a scale of 1 cm/sec. The events used occurred in South America, the South, Central (i.e., Solomon Islands region), and North-Western Pacific, and Novaya Zemlya, at distances of greater than 45° from the stations. In order to ensure clear first arrivals, only events of magnitude 5.5 or greater, generally at depths of 50 km or more, were read, and only those having clear arrivals at 6 or more stations were retained in this study. (Earthquakes shallower than 50 km usually have emergent first arrivals, and increase the scatter of the data without significantly improving the temporal resolution.) In each case the first arrival was read to an estimated accuracy of 0.05 sec, and the residual, with respect to the Jeffreys-Bullen arrival time, calculated. (First arrivals for all the events read were unambiguous; the technique of reading first peaks or troughs by phase correlation was not used because the variation in instrument response between the different stations, and at a given station as a function of time, produced a greater amount of scatter in the residuals). The theoretical arrival times were determined using the USGS hypocentral data, and included corrections for the ellipticity of the earth and the station elevations. The residuals were then normalized by subtracting the delay observed at Goldstone (GSC); any changes at GSC would thus show up as changes at all the other stations.

Goldstone was chosen as the normalizing station because it records clear first arrivals for the majority of events, and because it is at a greater distance from major faults and local structural boundaries. The effects of normalization are discussed by Engdahl *et al.* (1977). In this case, it is estimated that the maximum error in relative residual for a source mislocation of 0.3° in latitude and longitude does not exceed 0.1 sec for the distance and azimuth range studied, and is, in general, 0.07 sec or less. Under certain conditions, the effects of plate structure in the source region on relative residuals can be quite large for earthquakes occurring at depths of less than 500 km (see, e.g., Engdahl, 1975). However, in this study the maximum distance of any station from the normalizing station (GSC) is 310 km, which is considerably less than that used by Engdahl, and no apparent variations in normalized residual were observed for events at different depths within a given source region.

A further problem arises in the use of teleseismic residuals for investigating temporal velocity changes in southern California, namely the effect of the marked azimuthal dependence of the residuals due to the local upper mantle structure (Raikes, 1976). This effect may be minimized by considering only events within a small distance and azimuth window (approximately 5° to 10° in each). However, if only a single source region is used, the temporal and spatial resolution may be seriously impaired. For consideration of overall velocity changes, it is convenient to correct the residuals from a number of windows by calculating a mean value for events within a single distance-azimuth window, which must contain at least 5 events, and subtracting this from the normalized residual. The residuals from a number of azimuths may then be plotted on a single graph; decreases in velocity show up as periods of increased corrected residuals. Values of the mean residuals for the principal azimuth-distance windows are listed in Table 1.

The corrected relative residuals for the stations studies are plotted in Figures 2 and 3. The $2\sigma$ bars indicate the average overall azimuths of the scatter observed in a single distance-azimuth window, and are similar to those expected from reading error alone. It is estimated, on the basis of these errors, that a change in residual of 0.15 sec lasting for at least 6 months would be clearly resolved by this method; this corresponds to a velocity change of 9 per cent over a path length of 10 km with a mean velocity of 6 km/sec. (A change of 0.1 sec, or 6 per cent, would be barely detectable within the scatter.)

During the period 1972 to 1976 no significant changes in residuals were observed at any of the stations monitored, although there may be small changes (~0.05 to 0.1 sec, or 3 to 6 per cent) similar to those reported by Kanamori and Hadley (1975) within the expected scatter.

## DISCUSSION

The absence of any significant anomalies during the period studied must now be considered in terms of the seismicity of the region and the origins and nature of possible velocity changes. Various formulas have been derived linking the duration of the anomalous period $T$ and the magnitude $M$ of the associated earthquake. Whitcomb *et al.* (1973) proposed the relationship

$$\log_{10} T(\text{days}) = 0.68M - 1.31 \tag{1}$$

and Rikitake (1975) deduced an average relationship from a variety of precursor data

$$\log_{10} T(\text{days}) = 0.76M - 1.83 \tag{2}$$

Since, in the previous section, it was estimated that a period of anomalous velocity must last at least 6 months to be clearly detectable, these relationships may be used to convert this into a lower magnitude detection limit, yielding values of 5.25 and 5.4 using equations (1) and (2), respectively.

TABLE 1

TABLE OF MEAN RESIDUALS FOR MAJOR AZIMUTH-DISTANCE WINDOWS

| | Novaya Zemlya | South America | | | South Pacific | | | Central Pacific | | | Mariana-Bonin | Japan-Kuril | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | North | | South | Kermadec | S. Fiji-Tonga | N. Fiji-Tonga | South | | North | | South | North |
| ISA | −0.96 | −0.15 | 0.03 | 0.24 | 0.11 | 0.20 | 0.30 | −0.10 | 0.03 | −0.19 | −0.60 | −0.80 | −0.89 |
| PYR | −0.01 | −0.65 | −0.42 | −0.46 | −0.24 | 0.36 | 0.19 | 0.08 | | 0.25 | −0.09 | 0.17 | 0.25 |
| SBB | −0.29 | −0.49 | −0.52 | −0.60 | −0.44 | −0.34 | −0.28 | 0.01 | | 0.01 | −0.16 | 0.02 | 0.02 |
| PAS | −0.53 | −0.29 | 0.12 | −0.12 | −0.38 | −0.25 | −0.10 | 0.01 | 0.16 | 0.03 | −0.22 | −0.10 | −0.32 |
| MWC | −0.28 | −0.33 | −0.13 | −0.30 | −0.34 | −0.26 | −0.13 | −0.23 | | −0.04 | −0.38 | −0.28 | −0.24 |
| CSP | −0.18 | 0.08 | −0.08 | −0.29 | −0.08 | −0.22 | −0.22 | −0.05 | | −0.17 | −0.40 | −0.10 | −0.11 |
| RVR | −0.95 | −0.21 | | −0.25 | −0.38 | −0.26 | −0.28 | −0.04 | | −0.09 | −0.65 | −0.59 | −0.75 |
| TPC | 0.24 | −0.22 | −0.14 | −0.05 | 0.12 | 0.05 | 0.14 | 0.15 | | −0.16 | −0.54 | −0.51 | −0.48 |
| PLM | −0.07 | 0.08 | −0.08 | −0.24 | 0.08 | −0.05 | 0.09 | 0.18 | | 0.08 | −0.56 | −0.29 | −0.36 |
| PEC | −0.81 | −0.14 | | −0.30 | −0.38 | −0.21 | −0.24 | −0.21 | | −0.09 | −0.50 | −0.52 | −0.58 |
| GLA | 0.19 | 0.13 | −0.29 | −0.38 | 0.53 | 0.71 | 0.60 | 0.60 | 0.70 | 0.62 | 0.26 | 0.28 | 0.32 |
| CIS | −0.26 | −0.07 | | −0.20 | −0.37 | −0.18 | −0.30 | 0.09 | | −0.06 | −0.51 | −0.35 | −0.45 |

Note: The actual azimuth-distance windows used vary according to the station; the values quoted here are for mean residuals in the windows having most events, and are described by the approximate location of the events.

However, these limits are in no sense absolute: relations (1) and (2) were derived empirically for earthquakes largely of the thrust type, and it has been suggested (Whitcomb *et al.*, 1974; Whitcomb, 1976) that the precursor duration may be longer for earthquakes not having a thrust mechanism. In particular, Whitcomb *et al.* (1974) reported an anomaly lasting 1.8 years that was apparently associated with a magnitude 4.0 earthquake.

In addition, Anderson and Whitcomb (1975) proposed a relationship between the size of the anomalous region $L$, and the magnitude $M$ of the ensuing earthquake, namely

$$\text{Log } L(\text{km}) = 0.26M + 0.46 \tag{3}$$

which for the magnitude limits derived above from equations (1) and (2) gives anomalous regions of diameter 73.1 and 66.8 km, respectively. The estimated detection limit may thus be re-expressed as a magnitude $5(\pm\sim 0.5)$ earthquake

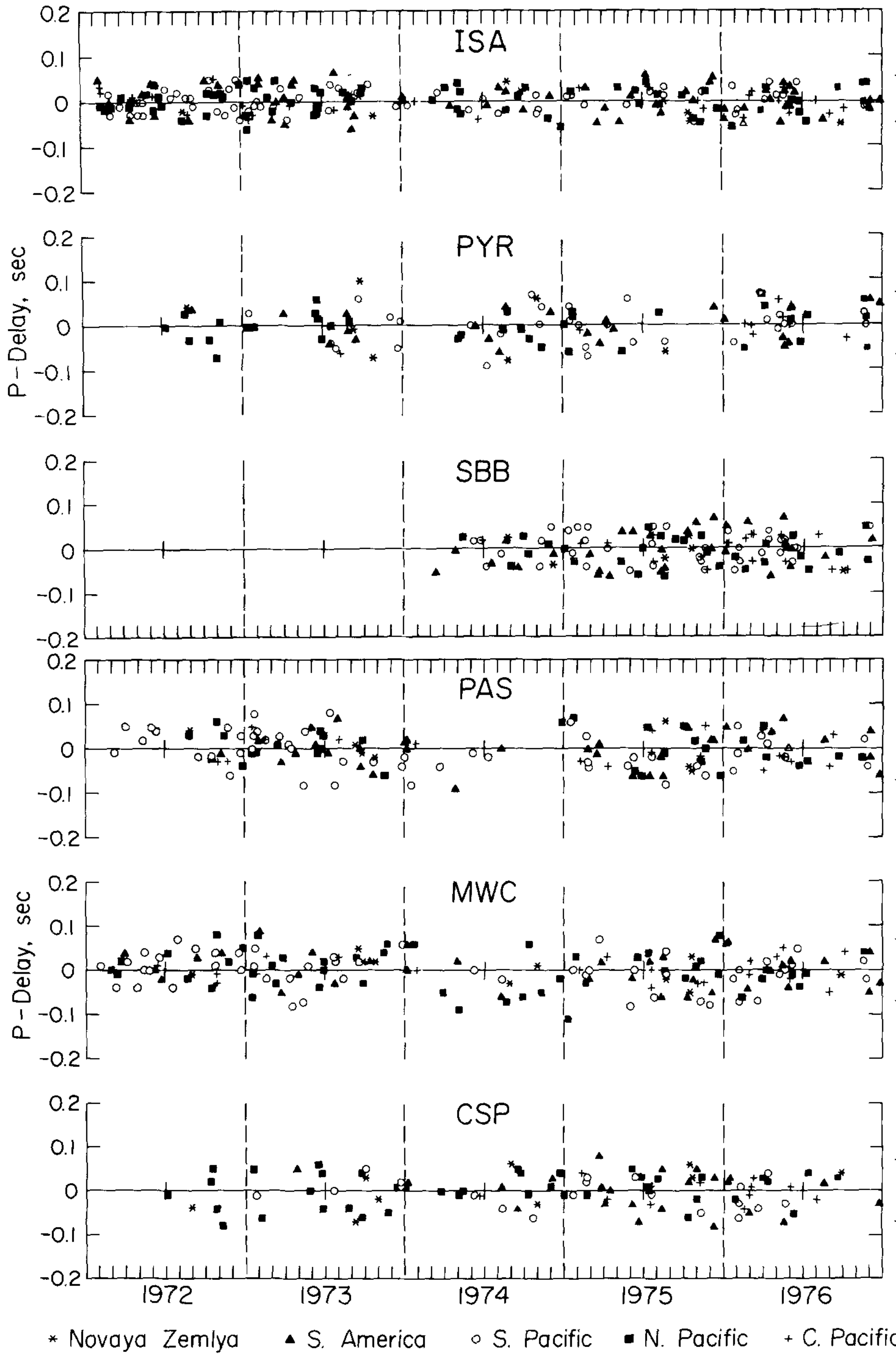


FIG. 2. Plots of corrected residuals, relative to GSC, as a function of time for stations ISA, PYR, SBB, PAS, MWC, and CSP.

occurring within 70 km of a station. Larger earthquakes will, of course, be resolvable at greater distances; for example, a magnitude 6 earthquake would have an associated anomalous area of diameter 105 km according to equation (3).

A limitation on the usefulness of teleseismic rays in investigation of crustal

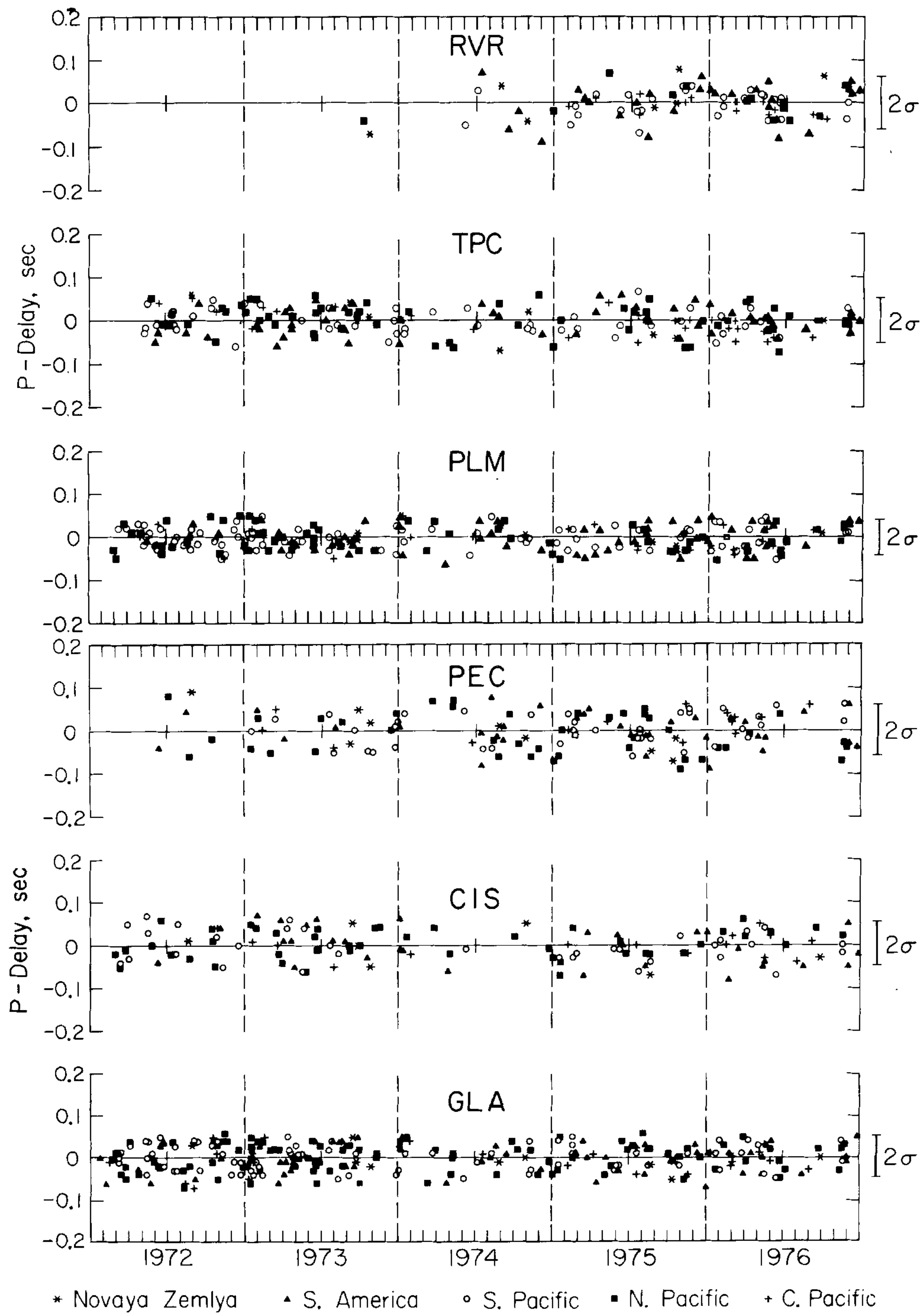


FIG. 3. Plots of corrected relative residuals as a function of time for RVR, TPC, PLM, PEC, CIS, and GLA.

velocity changes is imposed because of their steepness of incidence. In this study the rays enter the crust at, at most, 20 km from the station, and will only sample velocities within this radius. Consequently, for an earthquake occurring more than about 20 km from a station, teleseismic rays pass only through the edges of the anomalous region, where the velocity change may be lower, causing a smaller change in travel time. This would result in a smaller "detection radius" than the 70 km derived above.

Locations of earthquakes of magnitude 4.5 and greater occurring in southern California during the period 1972 to 1976 are plotted in Figure 4. Seismicity has been relatively low during this time, and there are few events sufficiently close to

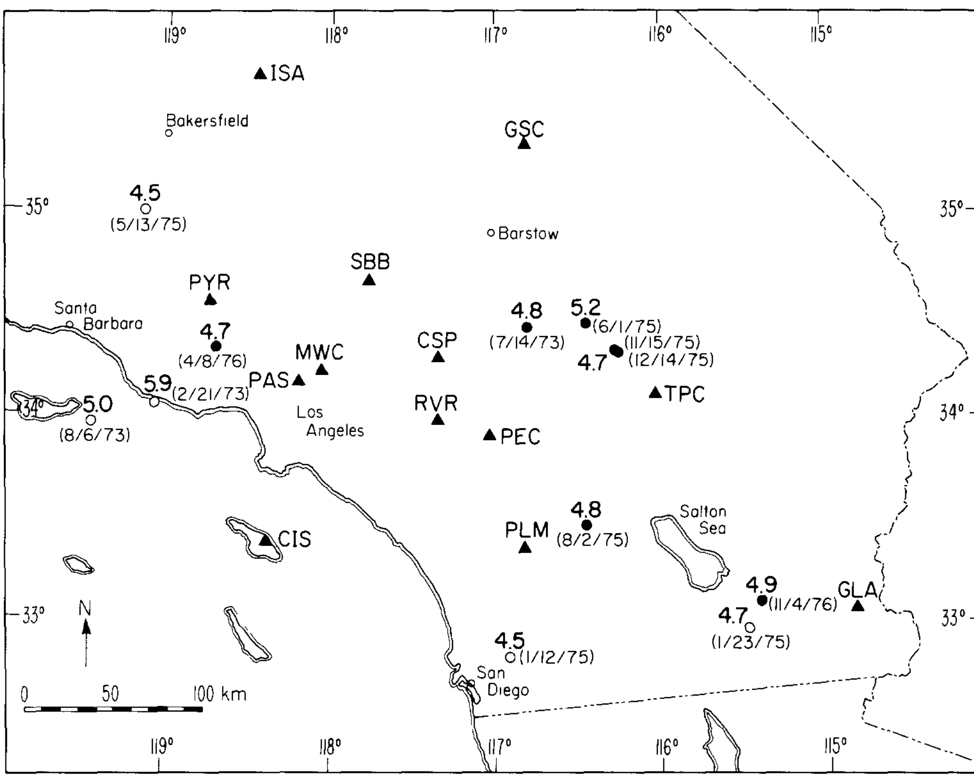


FIG. 4. Map showing locations of earthquakes of magnitude 4.5 or greater during the period 1972 to 1976. Open circles denote earthquakes outside the detection limit of the stations used.

any of the stations that they might be expected to give rise to detectable anomalies. The largest earthquake during this period, a magnitude 5.9 ($M_L$) at Point Mugu on February 21, 1973, occurs somewhat early to have been "predicted" using this data set, and is too far from PYR to produce a resolvable anomaly based on Stewart's (1973) estimate of 10 km for the size of the anomalous zone. The other earthquakes that might have been preceded by observable anomalies are the Galway Lake event of June 1, 1975, which is close to the detection limit for TPC, CSP, and PEC, and the Goat Mountain shocks of November 15 and December 14, 1975, which are rather small, but fairly close to TPC. None of these gave rise to a strongly visible anomaly, although there is a faint suggestion of a velocity increase in late 1975 for paths from the North Pacific to TPC. The Anza earthquake of August 1975 produced no resolvable change at PLM, and the April 1976 Newhall event produced no change at PYR, although there may be a small velocity increase near MWC in late 1975-

early 1976. In addition, there appear to be no changes associated with the spreading and deflation of the Palmdale Bulge reported by Castle *et al.* (1977).

Observations of local earthquake residuals are subject to much greater scatter, presumably due to complex local structure and difficulties of location, but appear to exhibit greater velocity fluctuations. Whitcomb (1976) observed a decrease in $Vp/Vs$ determined from local earthquakes during 1974 and early 1975, followed by a return to normal. He concluded that, based on current theories, a magnitude 5.5 to 6.5 earthquake should occur in the areas of southern California where the anomalous velocity ratio was observed. However, this conclusion later proved incorrect, as the velocity ratio decreased again after the occurrence of a magnitude 4.7 earthquake near Newhall in April 1976. Of the stations used in this study, SBB is on the northeastern boundary of Whitcomb's prediction area, PYR near the western margin, and MWC and PAS in the southeastern quadrant: No changes in teleseismic residuals clearly corresponding to his anomaly are seen at SBB or PYR, although there may be minor velocity increases at PAS and MWC for events from the South Pacific in early and late 1975, respectively. (However, the latter occurs too late to be associated with Whitcomb's anomaly.) A detailed analysis of the variation of residuals from local earthquakes for all stations of the USGS-Caltech Southern California Seismograph Network is presented by Powell and Whitcomb (in preparation). They conclude that during the time period 1972 to 1976 there are no significant variations in "local" residuals that can be related to the occurrence of earthquakes.

Discrepancies between local and teleseismic residual variations are not unexpected: the former depend strongly on the stations and local structural model used in the location of the earthquake. Furthermore, the teleseismic waves travel steeply through the crust, and if the anomaly is confined to a narrow depth range, it may be poorly sampled by telesismic rays whereas those from a local earthquake have a long travel path in the anomalous region. The orientation of the cracks that are thought to give rise to the velocity variations is also important. Not only does it give rise to horizontal anisotropy (Whitcomb, 1976), but if the cracks are vertical they will have little effect on the travel times of the nearly vertically incident teleseismic waves. An investigation of variations of seismic velocities in dilatant rock (Gupta, 1973a, b) showed that for areas of strike slip faulting the decrease in compressional velocity is greatest for near horizontal paths; this is also the case for normal faulting. However, for thrust faulting, which is common in the Transverse Ranges, the maximum velocity change is observed for near-vertical ray paths, and thus teleseismic residuals should be more affected than local ones. The size of the cracks may also contribute to their differing effect on teleseismic and local waves: small cracks will have a larger effect on the velocity of the higher frequency, shorter wavelength, $P$ waves from local earthquakes.

## Conclusion

Teleseismic residuals monitored at 13 stations in southern California from 1972 to 1976 show no significant variations when normalized to minimize common source and propagation effects, and corrected for their dependence on the azimuth of the event. The absence of clearly resolvable variations may be largely due to the low seismicity in the area during this period. Alternatively, it may imply that the only velocity changes taking place were small (such as might be associated with vertically oriented or small cracks) or confined to a narrow depth range in the crust and thus poorly sampled by teleseismic waves. The data presented in this study may provide a useful baseline for investigating any future variations.

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