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VELOCITY ANOMALIES: AN ALTERNATIVE EXPLANATION

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ABSTRACT

Material velocity changes have been reported to precede the last two magnitude 5 earthquakes along the San Andreas fault in central California. In both cases the anomalies were based on an increase of ~0.2 seconds in travel-time residuals from small regional earthquakes at one or more nearby seismic stations. A detailed reexamination of the data shows that the changes were more likely caused by differences in the depth and magnitude of the source earthquakes during the "anomalous" periods and were unrelated to any premonitory material property changes. Additional data are presented from sources chosen to minimize such problems. They show that travel times before the two magnitude 5 earthquakes were in fact stable to within a few hundredths of a second for rays that passed within a few kilometers of the hypocenters.

Given the great latitude that can be exercised in the selection of data after the fact to define premonitory changes, such anomalies may not be of any significance unless it is explicitly shown that they are not due to some other change in the sources used or signals measured.

"The ability of the human being to prove truth from coincidence is without bound" (Wrothall, 1977).

Introduction

Numerous studies have used small earthquakes to estimate *in situ* material velocities before larger earthquakes. The material velocity anomalies inferred by various adaptations of this technique (Aggarwal *et al.*, 1973; Whitcombe *et al.*, 1973; Robinson *et al.*, 1974) constitute much of the positive evidence for the dilatancy model of earthquake precursors (Nur, 1972; Scholz *et al.*, 1973). All these studies implicitly used some form of averaging to demonstrate a change in the mean value of some measured quantity in the presence of noise. Assessments of the significance of such changes usually assume, if only implicity, that the population has a time stationary Gaussian distribution, and that the samples selected consist of independent data points. When earthquakes are used as sources in a velocity study, the validity of these assumptions must be explicitly established for each experiment, since (1) arrival times based on emergent arrivals in the presence of noise may not have a Gaussian distribution (large later phases can be picked late as first arrivals, but never early and thus can have a skewed, rather than normal distribution), (2) earthquakes often do not constitute a stationary population with respect to depth, magnitude, or fault-plane solution, all of which can systematically bias velocity estimates (indirectly through changes in the shape or amplitude of the first arrivals, or, in the case of depth, directly when the actual and assumed velocity structures do not agree), and (3) earthquakes often cluster in space and time (aftershocks and swarms, for instance) and thus often cannot be considered independent events.

In this paper we show that travel-time residual changes preceding two magnitude 5 earthquakes along the San Andreas fault in central California are more likely due to sampling problems than to premonitory velocity changes. We use other travel

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times to show that when data are chosen so as to minimize these biases, a much higher precision is attainable and that a large velocity change preceding either earthquake is extremely unlikely unless confined to a very small volume.

THE THANKSGIVING DAY EARTHQUAKE "ANOMALY"

The hypocenter of the Thanksgiving Day earthquake (TGDEQ) (November 28, 1974, $M_L = 5.1$) was 8 km northwest of Hollister, California, at 6 km depth (Figure 1a). Fourteen aftershocks with $M_L > 2$ were located during the 36 hr following the main event (Figure 1b). Cross sections and fault-plane solutions indicate left-lateral motion on a vertical strike-slip fault striking N40°E. The main shock had a local magnitude of 5.1 ± 0.1 based on arrivals at nine Wood-Anderson stations (Lee, unpublished data) and a moment of about 10^{24} dyne-cm (Wayne Thatcher, personal communication, 1977). A longitudinal cross section of the early aftershocks outlines a zone roughly 3 km square centered at 5 km depth.

Preliminary evidence for a velocity anomaly before the TGDEQ was cited by Lee and Healy (1975), based on an analysis of travel-time residuals at two nearby seismic stations, CAN and FEL (Figure 1a). Residuals at these stations increased by about 0.2 sec during a 4-week period from mid-October to mid-November; the TGDEQ occurred November 28. The interpretation of these large residuals as a premonitory velocity change was strengthened by the apparent correlation with nearby tilt (Mortensen and Johnston, 1976) and magnetic (Smith and Johnston, 1976) changes that also preceded the earthquake (Figure 2).

The travel-time residuals shown are from small earthquakes ($M_L < 3.5$) on the San Andreas fault near Bear Valley, some 50 km to the southeast of the TGDEQ (Figure 1c). The earthquakes were located and the residuals calculated with a least-squares location program (HYPO-71, Lee and Lahr, 1975) modified to allow different velocity models to be used for stations situated on either side of the San Andreas fault. The arrival times used were those read by U.S. Geological Survey personnel for use in routine hypocentral locations.

As the residuals shown are the difference of an observed and a calculated travel time, they are sensitive to changes in the location of the source earthquakes, to the degree that the model used for travel-time calculations does not accurately reflect the actual crustal structure. Thus a systematic change in the location of the source earthquakes at Bear Valley could produce a change in observed residuals in the vicinity of the TGDEQ which would have nothing to do with a material velocity change. That this might in fact have occurred was suggested by the three anomalous residuals at station HER (also shown in Figure 2), which, because of the station's location some 50 km southeast of Bear Valley (Figure 1a), are improbably related to velocity changes near the TGDEQ. To explore this possibility we examined the hypocentral parameters of the source earthquakes for evidence of a residual correlation that might account for the apparent anomalies; the depth and magnitude of the source earthquakes are shown in Figure 2 for comparison with the residuals.

From the source depths plotted in Figure 2, it is clear that one event (16) is much shallower than most other events used. An examination of residuals from similar Bear Valley earthquakes during the previous 3 years (1972 to 1974) turned up only seven events with $M_L > 2.5$ and depths less than 3 km; they had a mean residual at CAN of 0.2 ± 0.1 sec. The 0.29-sec residual at CAN for event 16 is thus not anomalous at any level of significance when compared to events at a similar depth. This is not a new problem. In one of the original works reporting premonitory velocity changes, Nesersov *et al.* (1971) routinely corrected for the effect of the depth of the source earthquakes. Allen and Helmberger (1973) pointed out that the high "normal" velocities reported by Whitcomb *et al.* (1973) were consistent with relatively deep crustal sources; they suggested that the anomalous residuals reported might be due to shallow source earthquakes. More recently, Steppe *et al.* (1977) noted a correlation between travel-time residual and source depth for about one-third of the stations they studied in central California.

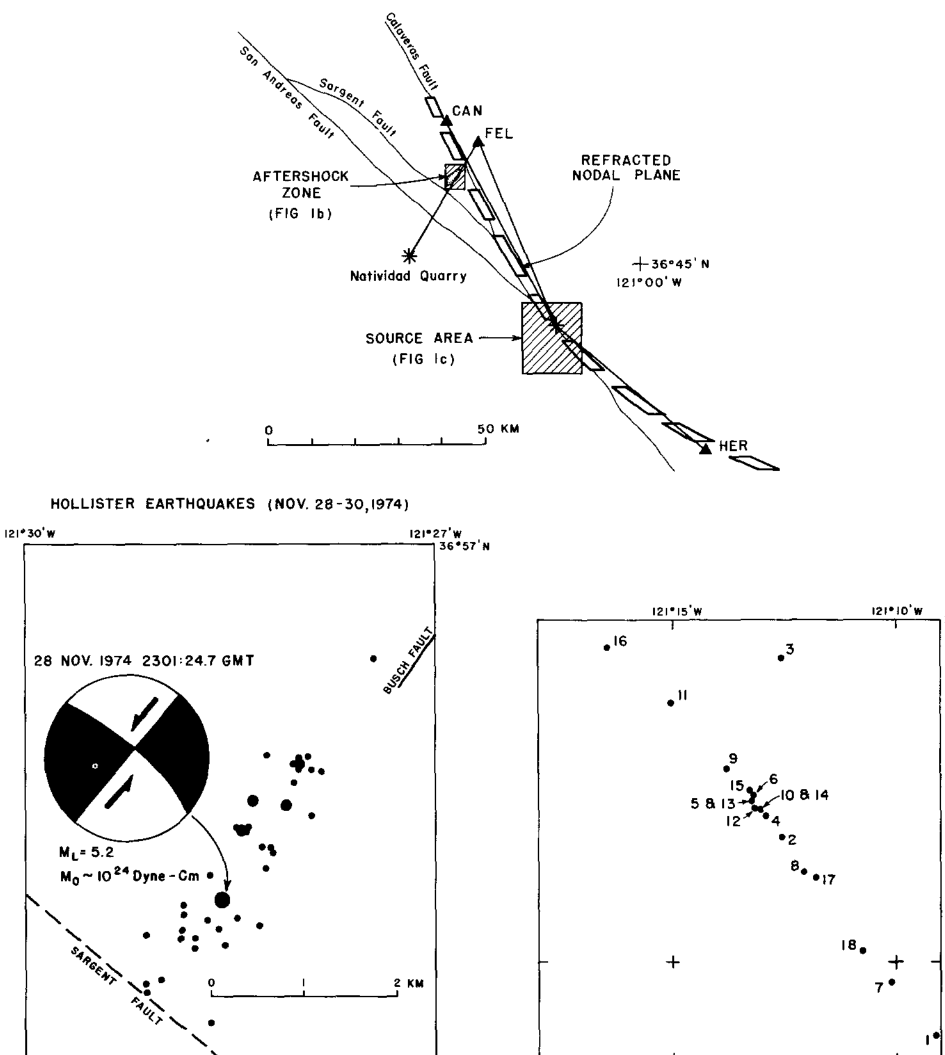

FIG. 1. (a) Map showing locations of earthquake source area, quarry, aftershock zone, and stations used in velocity study. The stations used to estimate the origin times of the Natividad Quarry blasts are indicated by open triangles. (b) Detail of Thanksgiving Day Earthquake aftershock zone. (c) Detail of earthquake source locations at Bear Valley.

With respect to the remaining anomalous points, examination of the magnitudes plotted in Figure 2 suggests an inverse correlation between magnitude and residual; residuals at CAN are plotted against source magnitude in Figure 3. These plots confirm the commonsense expectation that as the magnitude of the source event decreases (and the amplitude of the first arrival at a distant station decreases) the probability increases of a later arrival being picked as the first arrival. This will only be true, of course, for emergent first arrivals, where the signal amplitude increases with time. Steppe *et al.* (1977) also noted a correlation between residual and magnitude at a few stations in this area, although CAN was not one of the stations they studied.

Two points should be emphasized here. First, as mentioned above, the strong correlation between magnitude and residual shown in Figure 3 was not observed at

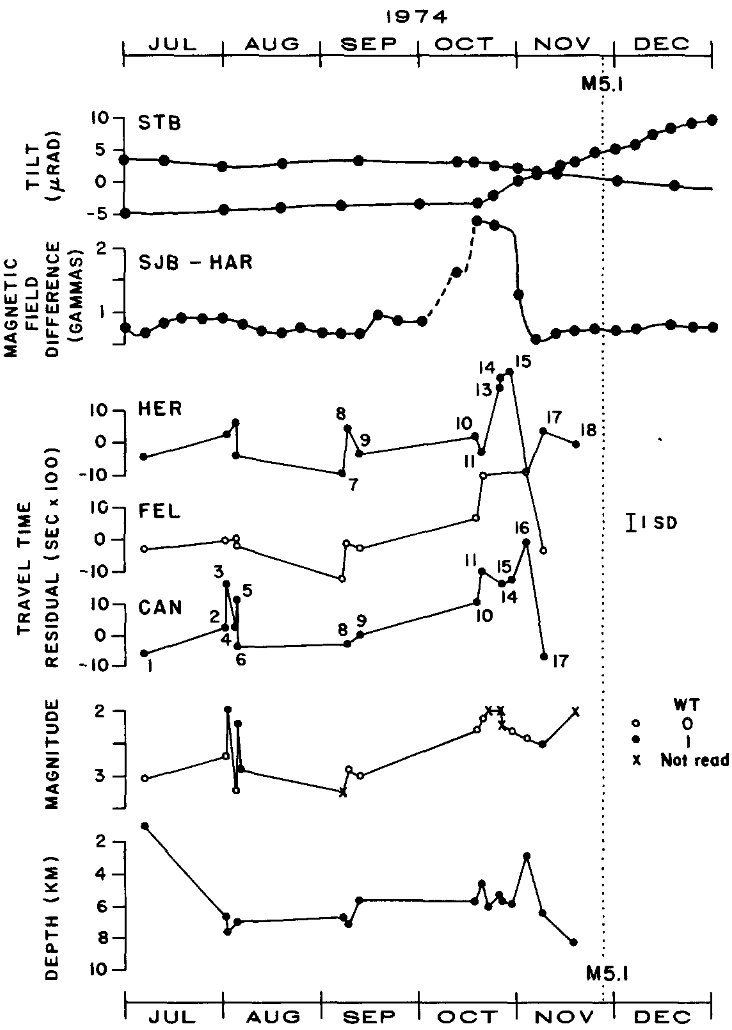

FIG. 2. *Top,* tilt (Mortensen and Johnston, 1976) and magnetic (Smith and Johnston, 1976) measurements near TGDEQ. *Middle,* travel-time residuals from earthquake sources. *Bottom,* depth and magnitude of earthquake sources.

any stations besides the three discussed. One possible explanation is that these three stations lie near a laterally refracted nodal plane of the source earthquakes' radiation pattern. This is shown in Figure 1a where the location of this refracted nodal plane is indicated with a broad dashed line. Such distortion of the first motion pattern due to lateral refraction has previously been noted along the San Andreas in the Bear Valley area (McNally and McEvilly, 1977) and at Parkfield (Lindh and Boore, 1974). The emergent character of the first arrivals at nodal stations might then result in the magnitude-residual correlation.

Second, the arrival times discussed here were almost entirely routine readings, made before the TGDEQ. (We reread all events in September, October, and November and found very few arrivals for which we disagreed with the original readings by more than 0.05 sec.) It is interesting to note that events of magnitude less than 2.5 are not normally read at stations as distant as those discussed here. However, due to an experiment being conducted at Bear Valley, USGS personnel processing the data during mid-1974 chose to read at those stations a number of smaller events. The residual "anomaly" was not noted, however, until after the TGDEQ.

While we find no evidence in the earthquake travel times discussed above for a temporal velocity change, the travel paths involved lie rather far from the aftershock zone of the TGDEQ (see Figure 1a), and the possibility remains of a more modest change, or one restricted to a smaller area.

Table 1 lists residuals for the time period April 1974 to September 1975 at station FEL, for blasts at the Natividad Limestone Quarry near Salinas (Figure 1a). These

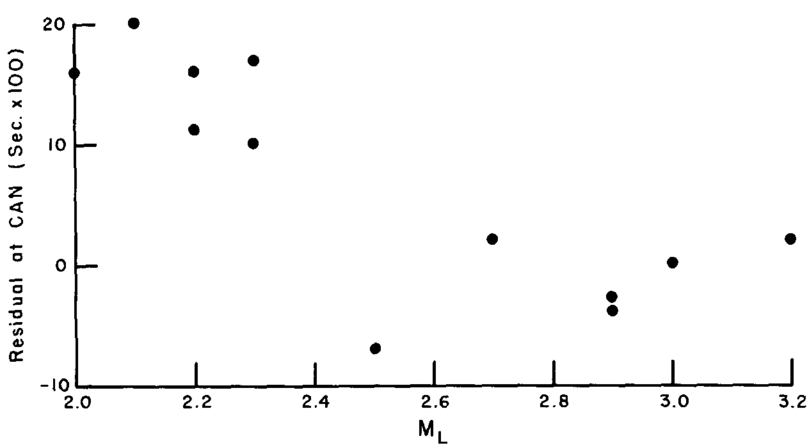

FIG. 3. Travel-time residuals at station CAN from Figure 2 plotted versus magnitude of the source events.

residuals were obtained by a method similar to that employed by McEvilly and Johnson (1974), except that arrivals at six nearby stations surrounding the quarry were used to estimate the origin time. (The six stations used are shown in Figure 1a). As can be seen in Figure 1a, the straight-line travel path passes within about 2 km of the aftershock zone. While exact routes taken by first arrivals in this geologically complex region are not known, it seems likely that the ray path did pass close to the main shock source volume. The mean of five residuals outside the period of the anomaly is −0.016 (±0.03 sec). The mean of three residuals during the anomaly is −0.013 (±0.05 sec). Given the irregular event spacing (station FEL was inoperative at times), these data do not preclude a velocity anomaly before mid-October. However, the hypothesis of a large velocity decrease throughout October (as is suggested by the earthquake travel times) can be rejected.

In an earlier study, Cramer (1976a) examined teleseismic residuals at 30 seismic stations in this area, and found no evidence for travel-time changes as large as 0.1 sec. This included arrivals at station OCR (Figure 1a) from South American and Central American sources that he estimated to have passed within 2.5 km of the hypocentral region of the TGDEQ. A similar conclusion was reached by Robinson and Iyer (1976) in a study of teleseismic arrivals at OCR from nuclear explosions in the Soviet Union. While teleseismic data are inherently less precise than higher frequency arrivals from local sources, they have the advantage of sampling deeper in the crust.

We conclude, then, that no significant evidence exists for a temporal velocity change before the TGDEQ, and that the data are inconsistent with a regional travel-time change of several tenths of a second. The observed "anomaly" is more likely attributable to the changes in the source earthquakes described above.

THE BEAR VALLEY EARTHQUAKE "ANOMALY"

The hypocenter of the Bear Valley earthquake (BVEQ) (February 24, 1972, M_L = 5.0) was located at a depth of 6 km on the San Andreas fault, 10 km north of San Benito, California (Ellsworth, 1975). The epicenter and outline of the aftershock zone are shown in Figure 4.

A travel-time residual anomaly before the BVEQ was originally reported by

TABLE 1

NATIVIDAD QUARRY RESIDUALS AT STATION FEL

Date (y.m.d.)	Origin Time* (GMT)	Location*		M_L	Δt_{FEL}†
		Lat.	Long.		
74 04 05	0047 29.54	36 45.33	121 35.61	2.5	−0.01
74 04 25	1551 24.49	36 45.28	121 35.55	2.1	+0.01
74 10 18	2259 50.77	36 45.34	121 35.83	2.1	−0.07‡
74 10 22	2259 16.58	36 45.34	121 35.68	2.2	+0.02‡
74 10 24	2300 01.69	36 45.19	121 35.59	2.1	+0.01‡
74 11 06	0000 53.66	36 45.12	121 35.62	2.2	−0.01
74 12 12	0000 49.73	36 45.33	121 35.61	2.0	−0.04
75 05 21	2300 32.91	36 45.27	121 35.53	2.4	−0.06
75 08 20	2300 10.67	36 45.30	121 35.52	2.4	+0.02

* Origin time and location are calculated. This means that, while of relative significance, they may not correspond to actual locations or shot times.

† Mean Δt_{FEL} ± its S.D. = 0.016 ± 0.015 for five normal events. Mean Δt_{FEL} ± its S.D. = 0.013 ± 0.028 for three during "anomalous" period.

‡ Residuals during the time of the "anomaly" preceding the TGDEQ. Event on November 6 not used since it falls on edge of anomaly; it is not clear if it should be included in normal or anomalous times.

Robinson *et al.* (1974), who used small regional earthquakes on the Calaveras fault as sources (Figure 4). They corrected for a weak epicentral distance-residual correlation, otherwise the method of analysis used was much as that described above for the TGDEQ study. The resulting residuals at seismic stations BVL, EKH, and JHC are shown in Figure 5; the large residuals at station BVL were taken as evidence for a velocity decrease preceding the BVEQ. These data were examined for possible source effects by Wesson *et al.* (1977); they concluded that the possibility of such an effect could not be excluded. This question will be considered here in somewhat more detail.

We have included in Figure 5 the depth and magnitude of the source earthquakes; the anomalous points are numbered for ease of comparison. An additional problem with this data set is that not all of the residuals at BVL are based on routine readings made before the BVEQ; to obtain enough data points during the anomalous time period it was necessary for Robinson *et al.* (1974) to read weak arrivals at the BVL station that had been passed over in the original processing of the data. The symbol plotted in Figure 5 indicates whether the residual is based on an original reading or was added after the BVEQ. Figure 5 shows several interesting features

1. Four of the anomalous events are shallower than any of the normal ones.
2. Of the four remaining anomalous events, three are among the smallest magnitude earthquakes used. The two events large enough to produce unambiguous arrivals at stations as distant as BVL (events 4 and 5) are among the shallow group.
3. Of the eight anomalous points, five were not read at station BVL by the USGS personnel originally processing the data, but were added after the BVEQ by Robinson *et al.* (1974). In addition, one of three points (8) was originally read as a "normal" value and is anomalous only on the basis of a rereading by Robinson *et al.* (1974).

With respect to the first item above, a search of the USGS catalog for the period 1971 to 1975 turned up only one event that had been read at the BVL station with source depth as shallow as the four shallow anomalous points; its hypocentral

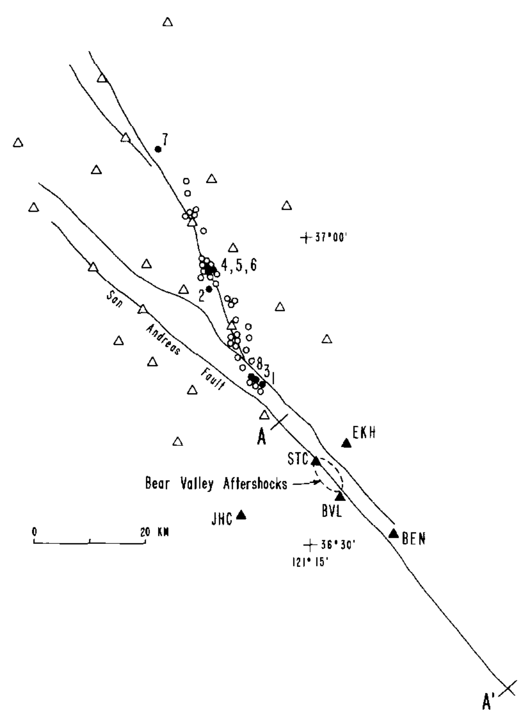

FIG. 4. Map showing location of Bear Valley earthquake epicenter and aftershock zone. Also shown are the locations of the source earthquakes along the Calaveras fault used by Robinson *et al.* (1974).

parameters and residuals are listed in Table 2, along with the same data for the two large shallow anomalous events (events 4 and 5). (The two small shallow events, 2 and 6, are not included since they were not read originally at BVL.) The crustal model and station corrections used are those of Robinson *et al.* (1974). The arrival times used both in calculating the locations and in determining the residuals are from the USGS catalog. The residual at BVL for the control event is within 0.05 sec of the average for the two anomalous events, and seems to be very strong evidence against a large travel-time anomaly at BVL in January of 1972. Unfortunately, the control event has a smaller magnitude than the two anomalous ones (see Table 2), and the arrival was thus assigned a greater uncertainty. We have attempted to confirm the arrival time by further work, and we believe that it is accurate to within ±0.05 sec. However, given the arguments we have made above for the TGDEQ concerning the possible bias involved when residuals are compared for events of differing magnitude, it would hardly be consistent to do so here. It is thus not possible to establish conclusively for these shallow events what constitutes a normal residual. We can note, however, the similarity to the anomaly preceding the TGDEQ discussed above. The travel path is in fact almost the reverse of that used for the TGDEQ anomaly, and it is reasonable to hypothesize that the anomalous residuals from shallow events are due to a similar effect at the source end. But regardless of the cause, we conclude that the four large residuals from shallow source events do

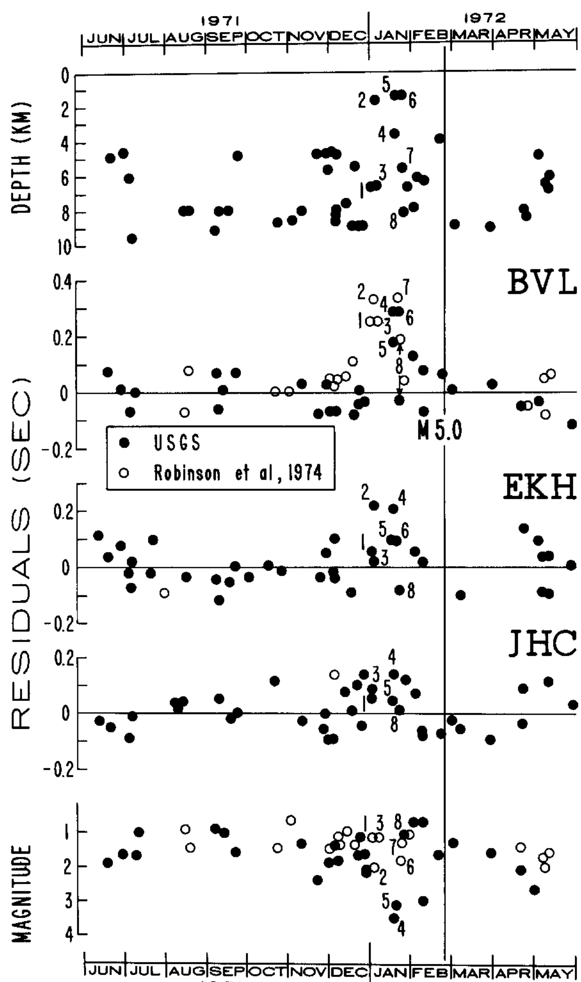

FIG. 5. Travel-time residuals at stations BVL, EKH, and JHC (from Robinson *et al.*, 1974). Symbol plotted indicates if residual was based on arrival time routinely read by USGS personnel or was added by Robinson *et al.* (1974). Also plotted are depth and magnitude of source events.

not constitute significant evidence for a premonitory velocity change.

With respect to the second item above, concerning the magnitude of the source events, again a clear resolution is not possible. The effect of magnitude is more difficult to assess than for the TGDEQ since the source events are distributed along a 50-km stretch of the Calaveras fault; the shape and amplitude of the first arrivals is thus not a simple function of magnitude. Of the four remaining anomalous residuals, one (7) is from a magnitude 1.3 earthquake over 70 km from the BVL station; it is an exceedingly weak and emergent arrival and can, we feel, be dismissed from further consideration. The three remaining anomalous events (1, 3, and 8) are not easily dismissed. Their magnitudes are all 1.1 or less, but they are located less than 25 km from the BVL station at a depth of about 8 km. Tracings of the three anomalous arrivals and those from three normal events from the same hypocentral region are shown in Figure 6. The wave forms are aligned by the calculated arrival times; positive residuals correspond to offsets to the right. The first arrival picks

TABLE 2

RESIDUALS FROM THREE SHALLOW EARTHQUAKES ON CALAVERAS FAULT

No.	Date (y.m.d.)	Origin Time (GMT)	Lat.	Long.	Depth	Mag.	Δt_{BVL}	Δt_{EKH}	Δt_{JHC}
4	72 01 18	1322 05.56	36 56.81	121 26.43	1.79	3.4	.31	.31	.13
5	72 01 18	1331 07.30	36 56.71	121 26.31	1.95	3.1	.34	.33	.14
—	75 07 28	0748 21.17	36 56.74	121 26.28	1.83	2.2	.27	.28	.14

corresponding to the residuals in Figure 5 are indicated with small vertical arrows; for the bottom trace the first arrow is the original pick, the second is that of Robinson *et al.* (1974). While the station is rather noisy, and all but one of the arrivals are emergent, we agree that the arrivals for two of the anomalous events,

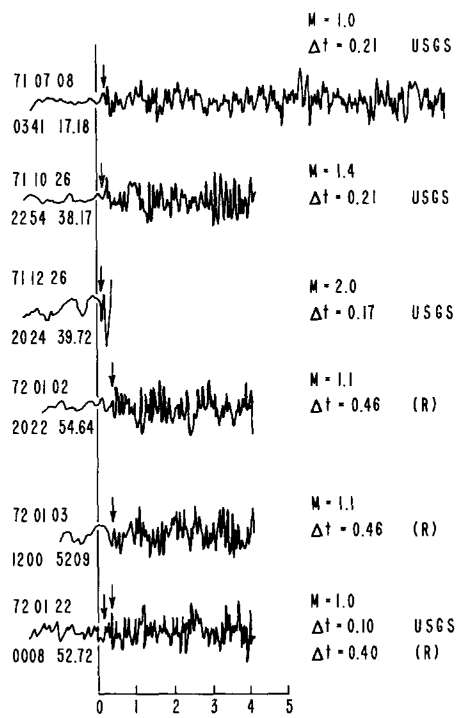

FIG. 6. Wave forms at BVL of 3 "normal" and 3 "anomalous" arrivals, wave forms aligned by calculated arrival times. Actual arrivals picked indicated by small vertical arrows.

numbers 1 and 3, may be late by 0.2 to 0.3 sec. In the case of all three anomalous arrivals, however, the signal-to-noise ratio is perilously close to 1. In addition, the fact that the apparent first motions in Figure 6 differ from event to event suggests that the BVL station is on a node for the *P*-wave radiation pattern, similar to the stations used to show the TGDEQ "anomaly" discussed above.

The question of adding and changing events on rereading the record is even more subjective but is crucial, we believe, to studies of this sort. Five of the eight "anomalous" arrivals at BVL were not read until after the "predicted" earthquake, and a sixth was changed at that time. A search of the USGS data set shows that in four of these cases, other stations recorded on the same Develocorder film were read. The USGS personnel originally reading the data thus made an implicit decision that the arrivals in question were not readable. We conclude, then, that in a critical sense, six of the "anomalous" points were drawn from a different population than most of the normal points. Given the subjectivity inherent in timing arrivals as emergent as those shown in Figure 6, it is simply not possible after the fact to assess objectively what kind of evidence for a velocity change they comprise.

We conclude that while serious questions can be raised concerning the significance of the data used by Robinson *et al.* (1974) in arguing for a velocity anomaly before the BVEQ, the data are not adequate to allow any resolution of the matter. Fortunately, though, other data are available that do shed light on the question.

Cramer and Kovach (1975) studied teleseismic residuals at stations BVL and BEN during the period July 1971 to April 1972. By differencing residuals at nearby stations and applying an azimuthal correction for source events from the southwest, they were able to achieve a standard deviation of 0.07 sec at BVL and 0.14 at BEN (see Figure 4 for station locations). They did observe two residuals at BVL during the "anomalous period" greater than 2 S.D. from the mean; however, the mean of seven residuals between January 1 to 20, 1972 was 0.04 sec with a S.D. of 0.1. The hypothesis that a 0.2-sec travel-time change occurred during that period is rejected. The mean residual at BEN during the same time period was 0.09; the scatter during normal times at that station precludes assigning it any significance. Similar conclusions were reached by Cramer (1976b).

Bakun *et al.* (1973) used *P* and *S* arrivals from earthquakes south of Bear Valley recorded at station STC to estimate the ratio of compressional and shear velocities (Vp/Vs) during 1972 in the Bear Valley area. They had two data points during the period of the purported anomaly, which differ from the mean Vp/Vs value by less than 0.2. As their data cannot be interpreted in terms of *P*-velocity change (simultaneous *P*- and *S*-velocity changes could go undetected), we repeated their experiment, using sources both north and south of the BVEQ aftershock zone along the San Andreas fault and examining only residuals from *P*-arrival times.

The crustal model and station corrections of Ellsworth (1975) were used to relocate the earthquakes. Arrivals at stations BVL and STC were not used in the solutions, to reduce the possibility of an anomaly being masked by hypocentral biasing. (The strong negative results obtained by Steppe *et al.* (1977) in their search for an anisotropic velocity change before the BVEQ effectively eliminate this possibility anyway.) All arrival times used in the locations and those used to calculate residuals at BVL are routine arrival times from the USGS catalog. Arrival times at STC were read for this study, as that station was not read by USGS personnel in 1971 to 1972. Station STC has somewhat lower gain than the other stations used, making identification of first arrivals difficult for smaller events. There are, however, very clear large second arrivals (see Bakun *et al.*, 1973 for illustration), one of which was used here (STC_2 in Table 3). Events were carefully selected with respect to location and magnitude to ensure uniform arrivals.

Figure 7 is a map and cross section (along the fault) of the Bear Valley area, showing the location of the source events used, the BVEQ aftershock zone, and stations BVL and STC. The source data and residuals at BVL and STC are listed in Table 3. Mean residuals during "normal" and "anomalous" times are also listed;

within the resolution (±0.015 sec at BVL, ±0.04 at STC) there is no evidence for a velocity change. The data are sufficient to rule out any travel-time changes larger than 0.05 sec. It should be noted that both vertical and horizontal velocity gradients

TABLE 3

EARTHQUAKE TRAVEL TIME RESIDUALS AT BVL AND STC

Area 1
Sources South of Bear Valley

Date (y.m.d.)	Origin Time (GMT)	Lat.	Long.	Z	M_L	Δt_{BVL}*	Δt_{STC1}*	Δt_{STC2}*
72 01 18	0454 44.18	36 25.6	121 00.9	5.9	1.6	−0.04†	+0.06†	−0.04†
72 01 19	0408 22.51	36 25.6	121 1.0	6.3	1.3	−0.07†		−0.04†
72 01 20	0345 42.97	36 22.5	120 56.0	6.3	1.7	−0.09†		+0.05†
72 01 27	1038 50.95	36 27.6	121 2.8	5.8	1.5	−0.04	+0.13	−0.03
72 03 19	1812 38.49	36 27.2	121 2.5	5.9	1.5	−0.05		
72 04 02	1737 0.69	36 27.6	121 3.0	5.7	1.6	−0.04	+0.04	+0.07
72 04 12	0152 24.0	36 27.1	121 02.3	5.2	1.8	−0.07		
72 05 28	1343 3.77	36 22.3	120 56.9	6.0	1.9	−0.14		−0.07
72 06 09	0915 57.2	36 23.0	120 58.0	6.5	1.3	−0.15		
72 06 14	0709 20.8	36 22.7	120 57.6	6.7	1.6	−0.09		
72 06 23	1957 48.0	36 22.9	120 57.9	5.3	2.0	−0.13		
72 07 09	0917 15.7	36 24.4	120 59.5	6.9	1.5	−0.05		
72 07 24	1053 12.8	36 24.8	120 59.6	7.4	1.7	−0.04	+0.20	
72 07 24	1639 49.5	36 24.3	120 59.4	6.2	1.4	−0.10		−0.01
72 09 20	1848 22.2	36 25.5	120 00.6	6.5	1.6	0.0		
72 11 25	0927 12.9	36 25.9	120 01.2	6.1	1.5	−0.07		

Area 2
Sources North of Bear Valley

Date (y.m.d.)	Origin Time (GMT)	Lat.	Long.	Z	M_L	Δt_{BVL}‡
71 03 12	1135 5.2	36 42.2	121 20.7	3.8	1.9	−0.07
71 09 03	0 8 1.1	36 40.1	121 18.2	3.1	2.4	−0.14
71 10 25	616 42.9	36 40.3	121 18.4	5.9	2.1	−0.03
71 12 19	1045 44.6	36 40.9	121 19.3	5.6	2.8	−0.01
72 01 01	951 49.7	36 41.6	121 20.0	5.5	2.5	−0.01†
72 01 01	1027 22.5	36 41.2	121 19.5	5.2	2.4	−0.03†
72 01 04	2317 28.8	36 40.6	121 18.8	4.0	2.4	−0.04†
72 01 07	2242 34.1	36 40.7	121 19.0	3.3	2.0	−0.07†
72 01 12	1808 43.9	36 41.4	121 20.4	3.7	2.2	−0.09†
72 01 19	2004 55.2	36 40.6	121 19.6	4.5	1.9	−0.12†
72 02 02	240 22.0	36 41.1	121 19.6	3.3	1.9	−0.09
72 02 12	1825 11.6	36 41.2	121 19.8	5.0	2.2	0.0
72 03 14	0925 16.0	36 40.1	121 18.1	5.3	2.0	−0.07
72 04 18	1557 23.5	36 41.7	121 20.7	4.1	2.2	−0.08
72 07 17	2020 28.9	36 40.5	121 18.4	3.2	2.2	−0.15
72 07 25	2022 17.9	36 41.3	121 20.0	4.2	2.7	−0.09

* Mean Δt_{BVL} ± its S.D. = −0.08 ± 0.014 sec for 12 normal events, −0.07 ± 0.017 sec for 3 events during anomalous period. Mean Δt_{STC2} ± its S.D. = −0.01 ± 0.03 sec for 3 normal events, −0.003 ± 0.04 for 3 events during anomalous period.

† Residuals during the time of the "anomaly" preceding the BVEQ.

‡ Mean Δt_{BVL} ± its S.D. = −0.07 ± 0.016 for 10 normal events, −0.06 ± 0.016 for 6 events during anomalous period.

are known to exist in this area (Healy and Peake, 1975; Stierman *et al.,* 1976), but they are not so well determined as to permit detailed calculation of ray paths in this case. Thus, while the travel paths sketched in Figure 7 are somewhat conjectural it is likely that a positive vertical gradient does exist in the fault zone, and that rays with curvature convex downward are roughly correct. In particular it is difficult to see how the ray paths to STC from area I or BVL from area II could avoid passing through (or very near) the BVEQ aftershock zone. The important point is that because the travel paths are short and directly along the fault, they stand a much better chance of passing through (or near) the source volume than those of Robinson *et al.* (1974). The total absence of any evidence for a velocity change in Table 3 is difficult to reconcile with their claims of a premonitory velocity change.

The data in Table 3 further illustrate that by careful selection of sources to ensure unambiguous arrivals, and the use of a small source volume to minimize modeling difficulties, travel times from local earthquakes can provide velocity estimates whose precision is limited only by the intrinsic reading error. Conversely, as we have attempted to illustrate in this paper, if such care is not taken with the source earthquakes used, travel-time anomalies will be found which have no relation to material property changes. We are not in a position to assess how much of this criticism might apply to other published velocity anomalies, as none of the published reports contain sufficient detail concerning the source earthquakes used. In central

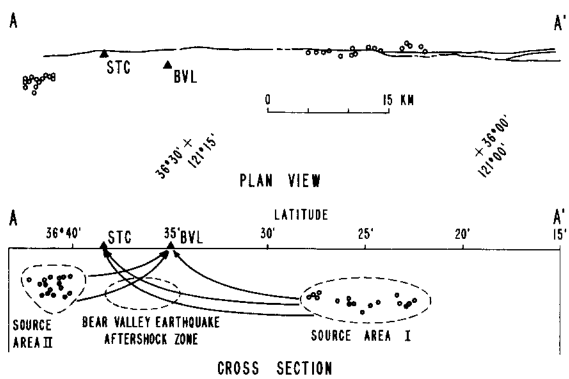

FIG. 7. *Top,* map of Bear Valley area, showing source event south of Bear Valley, and BVEQ aftershock zone. *Bottom,* cross section along A to A′, showing hypothetical travel paths from source earthquakes to stations BVL and STC.

California, however, we know of no published evidence for a velocity change to which it does not apply.

We conclude, that the bulk of the available data argue strongly against a travel-time increase of 0.2 sec preceding the BVEQ. The only data supporting such a change are those of Robinson *et al.* (1974). Given the weaknesses of that data set discussed above and the quantity and strength of the negative evidence, it seems very likely the changes they reported are due entirely to source effects.

Discussion

We have reexamined data from studies along the San Andreas fault in central California in which temporal changes in travel-time residuals have been interpreted as evidence for material velocity changes premonitory to two magnitude 5 earthquakes. We find that

1. Some of the anomalous earthquakes are shallower than the normal ones. In the case of the TGDEQ we establish that depth is the cause of one of the large residuals; for the BVEQ it is likely the source of four of the anomalous residuals.

2. In each case the remaining anomalous earthquakes are smaller than the average normal event. In the case of the TGDEQ this clearly contributed to the appearance of an anomaly. The same cannot be established with certainty for the BVEQ, but in light of the other evidence cited, it is very likely.

3. The BVEQ anomaly was based largely on arrival times which were originally not read in routine processing of the data, but were added after the "predicted" earthquake, in a search for a velocity anomaly. Since the earthquakes added include the small ones mentioned above, they contribute an element of subjectivity to the problem which is inherently unresolvable.

Thus we conclude that while both data sets are weakly consistent with a premonitory velocity change, neither constitute significant evidence for one. In fact, the anomalous populations differ so fundamentally from the normal ones that no meaningful test of significance is possible.

We also examined other data to bear on the question of premonitory velocity changes. We found that by careful selection of earthquake and quarry sources with respect to location, depth, and magnitude, travel-time residual stability of better than 0.05 sec was obtainable. For rays passing closer to the source volumes of the magnitude 5 earthquakes than those along which the purported anomalies were observed, we found no significant residual fluctuations. During the time periods of the reported anomalies, our data preclude travel-time changes exceeding 0.05 sec. Our findings are substantiated by studies of teleseismic residuals at nearby stations; in a series of papers summarized by Cramer (1976b), no evidence was found for any travel-time variations preceding either earthquake.

Given that earthquake arrival time residuals can be a function of location, magnitude, and fault-plane solution, and that these parameters cannot be assumed to be drawn from a stationary population, velocity changes inferred from such data must be explicitly shown not to be due to such causes before they constitute evidence for an *in situ* material property change. This is particularly true in light of the fact that residual distributions can easily be skewed to the late side and current theories of premonitory velocity changes predict only lowered velocities.

If the observed fluctuations in travel-time residuals are due primarily to random fluctuations in noisy data, one might ask whether it is nonetheless a remarkable coincidence that in each case the observed changes occurred just before large earthquakes, and at neaby stations. When we consider the degrees of freedom available in searching after the fact for such anomalies, we think the coincidence is not remarkable. Given that many data sets are often available to draw from, and that the magnitude, location, onset time, duration, and character of the anomaly can be chosen with considerable latitude, it is in fact predictable that changes in *something* will be found preceding any earthquake of interest. This is not to say that identifying coincidences is not a legitimate approach in a field like earthquake prediction, where little observational or theoretical basis exists to guide research, but only that the analysis must not stop at that point if the work is to be of any scientific value.

Finally, while we dealt here with changes in P-arrival times, the same considerations could apply to Vp/Vs values inferred from P and S arrivals. For instance, the slope on a Wadati plot is often controlled by the most distant point(s), and S arrivals are often larger in amplitude than P arrivals. If smaller events were read during an "anomalous" time, and if P-arrival times at distant stations were as a result sometimes picked late, the result would be an apparent decrease of the Vp/Vs ratio.

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