by the thin reddish North Temperate Belt (NTeB). While the NTrZ shows detailed features, especially on the southern edge, the NTeB appears rather uniform and so does the North Temperate Zone above it. Figure 3 shows that there is a swift current at the NTeB. The northern North Temperate region shows intricate detail with cell patterns, although these do not appear as marked as in the South Temperate regions.

The North Polar Region begins with a striking, dark, elongated feature (see Figs. 1, 2, and 5). Belts at high latitudes often occur in such discontinuous pieces. Here, some foreshortening must be allowed for because of the southern latitude (−10°) of the subspacecraft point; less than a full hemisphere is observed this close to the planet. Little detail is seen in the North Polar Region. There is a variety of evidence from Earth-based polarimetry (*7*) and photometry (*8*) for greater effective penetration of visible radiation at latitudes higher than about 45° compared with equatorial regions. The deeper penetration of a thick atmosphere appears to be responsible for the relative lack of features in the polar regions.

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Variation of P-Wave Velocity before the Bear Valley, California, Earthquake of 24 February 1972

Abstract. *Residuals for* P-*wave traveltimes at a seismograph station near Bear Valley, California, for small, precisely located local earthquakes at distances of 20 to 70 kilometers show a sharp increase of nearly 0.3 second about 2 months before a magnitude 5.0 earthquake that occurred within a few kilometers of the station. This indicates that velocity changes observed elsewhere premonitory to earthquakes, possibly related to dilatancy, occur along the central section of the San Andreas fault system.*

Premonitory changes in seismic-wave velocity in the source regions of earthquakes in the Soviet Union, New York State, the Transverse Ranges of southern California, and Japan have been reported (*1*, *2*). In particular, Wyss and Holcomb (*2*) have shown that observations of teleseismic *P*-wave traveltime residuals (observed minus expected arrival times) at a seismograph station near an approaching shock can be used to detect such changes in velocity. Along the active central section of the San Andreas fault, however, several attempts to observe such changes in *P*-wave velocity have produced generally negative results (*3–5*), and it has been suggested (*6*) that a popular explanation for velocity changes (dilatancy) may not be applicable to strike-slip faults in general.

The Bear Valley earthquake [magnitude (*M*), 5.0] of 24 February 1972 provides an ideal opportunity to search for variations in traveltime residuals for an earthquake along the San Andreas fault, as a U.S. Geological Survey (USGS) seismograph station (BVL) lies nearly above the hypocenter of the event (Fig. 1). Direct application of the method of Wyss and Holcomb, however, is not possible because teleseismic events are poorly recorded at USGS stations that are designed for the detection of small, local events. Moreover, the relatively large errors involved in determining teleseismic residuals require that many separate observations be averaged over long periods of time (*2*). It would be difficult, then, to resolve changes in the traveltime residuals (expected to be less than about 0.4 second) occurring over the 3-month precursor time predicted by the dilatancy model (*7*) for an event with a magnitude of 5.0.

We have made use of abundant local earthquakes to observe *P*-wave traveltime residuals. In particular, we have used small events ($M = 0.7$ to 3.6) located along the Calaveras fault 20 to 70 km northwest of the Bear Valley region during the 1-year period between 1 June 1971 and 30 May 1972. These events (a total of 49 observed sufficiently well at BVL) provide an adequately dense sample in time. And as they can be precisely located, by making use of the network of USGS seismograph stations in the region (*8*), the error in any one observation of the traveltime residual is less than the expected variation due to velocity changes. Figure 1 shows the epicenters of these events and the seismograph stations used to locate them. Each event was located by using only stations within 40 km of the epicenter, and excluding BVL and the nearby stations EKH and JHC (at Elk Horn Ranch and Johnson Canyon), at which we wished to observe traveltime residuals. Under this restriction, all events, large or small, are located with the same station geometry. The location procedure involved adjustments of station delays from those normally used for locating earthquakes throughout central California (*8*) to values more appropriate to the set of earthquakes occurring within this restricted region (*9*). The relative location of events should be significantly improved by this relocation procedure (*10*).

The *P*-wave traveltime residuals so obtained at BVL show a weak dependence on distance from the source, probably due to minor errors in the simplified velocity model for paths from the source region along the Calaveras fault south to the Bear Valley area. This distance trend at BVL was estimated by the relation (determined by least squares)

$$R(\Delta) = 0.300 - 0.0038\Delta$$

where R is the traveltime residual (in seconds) estimated as a function of the distance Δ (in kilometers). Only residuals from the time period preceding the anomaly were used in estimating the distance trend. A correction for the appropriate distance was then removed from each residual. Residuals at JHC and EKH were estimated by the relations

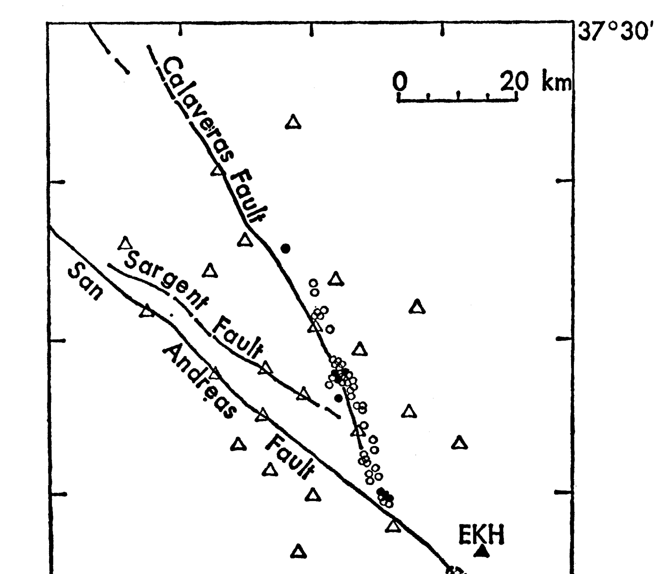
$$\text{(JHC)} \quad R(\Delta) = -0.031 + 0.024\Delta$$
$$\text{(EKH)} \quad R(\Delta) = 1.112 - 0.0341\Delta + 0.00026\Delta^2$$

and were corrected in a similar manner. Alternatively, slightly different velocity models could be used to accomplish the same thing, but three separate models would be necessary. This correction procedure significantly reduces the scatter in the observed residuals (for example, by a factor of ½ at EKH).

The corrected *P*-wave traveltime residuals are shown in Fig. 2, where it can be seen that there is a period between late December 1971 and late January 1972 when the residuals at BVL are anomalously high, jumping nearly 0.3 second above normal values. The mean of the eight observations within this period differs significantly from the mean for the preceding 7 months at a very high confidence level (99 percent). The events occurring during this anomalous period cover a wide range in distance (20 to 70 km; Fig. 1), magnitude (0.7 to 3.4), and depth (2 to 9 km) and, on the average, are similar in these regards to events occurring during the nonanomalous periods. Only solutions resulting in high-quality locations [USGS class A or B (*8*)] are used. Results at EKH are somewhat ambiguous, two events producing high residuals during the anomalous period at BVL while the remaining five are within normal bounds. The residuals at JHC show no indication of exceeding the normal bounds; this can be taken to rule out any systematic mislocation of events observed at both JHC and BVL during the anomalous period at BVL, and also serves to limit the possible extent of any temporary low-velocity region that could explain the other observations. Although the average residual at JHC is slightly higher during the anomalous period, statistically it is not significantly so. It appears, then, that there is a clear indication of a decrease in *P*-wave velocity near the location of the Bear Valley earthquake beginning 53 to 60 days before the event. The velocity returns to normal several weeks before the event, but the scatter in the data is such that this could occur either rapidly or gradually. Residuals during the aftershock sequence are within normal bounds. The magnitude of the change in residual at BVL could be explained by a 10 to 15 percent decrease in *P*-wave velocity within a volume of the same radius as the observed aftershock zone (7 km).

Arrival times used for locating the source events were routine readings made by the staff of the USGS and have an estimated error of 0.05 second. We have reread arrival times for all events in December 1971 and January and February 1972 in order to test the variability of the resulting locations and residuals. We find that although individual residuals may change within about ± 0.1 second, the pattern of anomalously high residuals at BVL remains essentially unchanged. From this, we have confidence that the routine readings of arrival times are fully adequate to resolve variations in traveltime residuals and that the observed period of high residuals does not result from a series of reading errors.

The results presented here may appear to be in conflict with those of McEvilly and Johnson (*4*), who investigated the stability of *P*-wave traveltimes along a path that passes about 10 km north of Bear Valley. Using quarry blasts as sources and seismograph stations operated by the University of California, they found no significant variation in traveltime preceding the Bear Valley earthquake. However, no quarry blast actually occurred during the period of clearly anomalous residuals at BVL. And the ray path they investigated could well pass to the north of or beneath a zone of low velocity near Bear Valley. The negative results of Bakun *et al.* (*3*) are inconclusive for several reasons, including poorly defined phase arrivals, and especially biasing of source-event locations because the stations were within or near the zone of anomalous velocity. McEvilly and Johnson (*4*) also report apparently stable velocities for other paths across seismically active areas in central California. The combination of their results with those presented here suggests that if strike-slip earthquakes in central California are preceded by



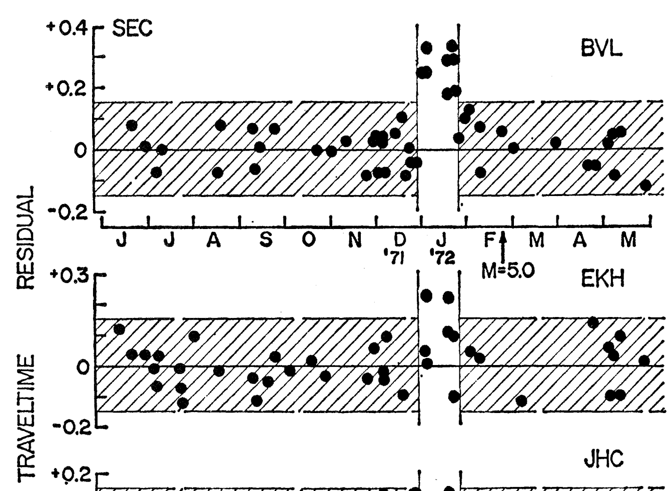


Fig. 1 (left). (Hatching) Location of the Bear Valley earthquake and its approximate aftershock zone. (△) Seismograph stations used to locate events; (▲) stations at which traveltime residuals are examined; (○) locations of events occurring during periods of normal residuals at BVL; (●) locations of events occurring during the anomalous period. Fig. 2 (right). Variation in *P*-wave traveltime residuals at stations BVL, EKH, and JHC over a 12-month period. The hatched areas represent the estimated normal range of residuals (± 0.15 second). Not all the events used produced usable first arrivals at all stations so that the data sets are not identical.

a localized zone of low P-wave velocity, then the velocity change is confined to a volume in the crust with a characteristic dimension of, at most, two or three times the dimension of the main shock rupture. The temporal duration of the period of low P-wave velocity before the Bear Valley earthquake, measured from the onset of the anomaly to the earthquake, is consistent with the precursory time scale that has been suggested by other workers (*1, 7*), although the period of clearly anomalous traveltime residuals is somewhat short.

Attempts to extend our observations of traveltime residuals to other earthquakes along the San Andreas fault system and to define the details of the anomaly around Bear Valley are under way. Our present observations indicate to us a clear velocity change premonitory to the Bear Valley earthquake and give us hope that the monitoring of such changes, or related ones, may aid in predicting future earthquakes along the San Andreas fault (*11*).

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9. The regional USGS crustal-velocity model for central California was modified to more accurately model earthquake traveltimes along the southern end of the Calaveras fault by adjusting the thickness of the surficial layer (sedimentary section) at each seismograph station used for our locations. Initial estimates of the first-layer thickness, based on crustal refraction data, were used to locate a suite of earthquakes distributed along the Calaveras fault. The average traveltime residual at each station for the suite of earthquakes was then used to adjust these thicknesses, and new locations were computed. This process was repeated until no significant improvement in the mean station residuals could be obtained (all were less than 0.05 second).
10. For each individual earthquake, a root-mean-square (r.m.s.) traveltime error can be calculated. The average r.m.s. error for the suite of events considered here was reduced from 0.12 to 0.07 second. W. L. Ellsworth, in data being prepared for publication, estimates that a similar relocation process reduced by one-half the relative epicentral errors for a suite of earthquakes near Bear Valley.
11. Periods of anomalously high traveltime residuals have now been observed preceding two other moderate earthquakes along the central San Andreas fault system; this will be reported on elsewhere (R. Robinson, R. L. Wesson, W. L. Ellsworth, A. Steppe, in preparation).

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Strain Release Mechanism of the 1906 San Francisco Earthquake

Abstract. *Reexamination of geodetic data has shown that aseismic slip occurred on or near the San Andreas fault in the period of about 20 years after 1906. The inferred displacements are comparable to but at greater depths than the sudden slip that occurred at the time of the earthquake. The postseismic slip is constrained only between late 1906 and 1925, and data are insufficient to determine the movements, if any, below about 20 kilometers on the fault. Two independent observations also indicate substantial anomalous crustal deformation away from the fault at least 30 years before the earthquake.*

Geodetic surveys made before and after 1906 have provided considerable information on the crustal deformation that accompanied the great San Francisco earthquake (magnitude, 8.3). These data led Reid (*1*) to propose in 1910 the now widely accepted elastic rebound mechanism for earthquake occurrence. Reid postulated that strain accumulated elastically by continuous slow motions at great distances from the San Andreas fault, and was released abruptly by sudden slip on the fault surface at the time of the earthquake. Using modern model-fitting analytical techniques, I have reexamined these data as well as some later observations. The principal new result of this work is the conclusion that substantial aseismic movements occurred along the San Andreas both before and after the earthquake, producing elastic strains within the adjacent blocks and measurable displacements of permanent survey monuments. These motions occurred on or near the fault plane at depths greater than those at which most of the seismic slip took place.

Table 1. Angle changes at triangulation stations in the Fort Ross net (see Fig. 1b) surveyed in 1875, 1906, 1930, and 1969. For each angle change, the station number refers to the vertex of the triangle from which the other two stations of that triangle are observed; N.S., not surveyed.

Station	Angle change (arc seconds)		
	1876–1906	1906–1930	1906–1969
2	+44.1	− 8.5	− 7.1
3	+13.5	− 2.3	− 3.9
4	−52.5	+11.3	+11.1
Closure	+ 5.1	+ 0.5	+ 0.1
3	−17.8	+ 9.6	N.S.
4	+17.9	−13.4	N.S.
5	N.S	N.S.	N.S.
1	−48.6	+ 2.0	− 3.9
2	+96.2	+ 4.7	+ 9.5
4	−50.4	− 0.7	− 1.9
Closure	− 2.8	+ 6.0	+ 3.7

Observed surface ground breakage in 1906 was extensive (Fig. 1a). Surface fault slippage averaged 4 m north of San Francisco, was nearly 3 m in the San Francisco Bay area, and decreased abruptly to 1 m or less south of San Jose. The basic data used in this study are repeated measurements of the angular separations of permanent survey markers (triangulation) (*2*). The uncertainties in these measurements are reliably obtained by the requirement that the angles of each triangle sum to 180° plus the excess due to the earth's sphericity (triangle closure).