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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References and Notes

1. See, for example, E. F. Savarensky, *Tectonophysics* **6**, 17 (1968); I. L. Nersesov, A. N. Semenov, I. G. Simbireva, in *The Physical Basis of Foreshocks* (Nauka, Moscow, 1969); Y. P. Aggarwal, L. R. Sykes, J. Ambruster, M. L. Sbar, *Nature* (*Lond.*) **241**, 101 (1973). A. Nur [*Bull. Seismol. Soc. Am.* **62**, 1217 (1972)] proposed that variations in the traveltime ratio of P waves and S waves (t_s/t_p) are due to dilatancy and that the principal change is in the P-wave velocity. J. H. Whitcomb, J. D. Garmany, and D. L. Anderson [*Science* **180**, 632 (1973)] and P. G. Richards and Y. P. Aggarwal [*EOS Trans. Am. Geophys. Union* **54**, 1134 (1973)] have corroborated Nur's laboratory data with field data, and along with Scholz *et al.* (*7*) have argued that the region of anomalous velocity may be much larger than the source region of the earthquake involved.
2. M. Wyss and D. J. Holcomb, *Nature* (*Lond.*) **249**, 139 (1973).
3. W. H. Bakun, R. M. Stewart, D. Tocher, *Stanford Univ. Publ. Geol. Sci.* **13**, 453 (1973).
4. T. V. McEvilly and L. R. Johnson, *Science* **182**, 588 (1973). These authors have also prepared for publication more extensive data, including information on S-wave velocities.
5. R. L. Wesson, D. P. Hill, J. W. Hillhouse, D. M. Boore, *EOS Trans. Am. Geophys. Union* **53**, 1041 (1973). Diffculties in reading the arrival time of S waves precluded definite conclusions. Work on recording S waves and intepreting t_s/t_p ratios in central California is continuing.
6. A. Nur, M. L. Bell, P. Talwani, *Stanford Univ. Publ. Geol. Sci.* **13**, 391 (1973).
7. C. H. Scholz, L. R. Sykes, Y. P. Aggarwal, *Science* **181**, 803 (1973).
8. A description of the USGS seismograph network in central California, the velocity model used for earthquake locations, and the routine locations of the earthquakes used in this study can be found in: W. H. K. Lee, K. L. Meagher, R. E. Bennett, E. E. Matamoros, *Catalog of Earthquakes along the San Andreas Fault System in Central California for the Year 1971* (Open-file report, U.S. Geological Survey, Menlo Park, Calif., 1972); R. L. Wesson, R. E. Bennett, K. L. Meagher, *Catalog of Earthquakes along the San Andreas Fault System in Central California, January–March 1972* (Open-file report, U.S. Geological Survey, Menlo Park, Calif., 1972); R. L. Wesson, R. E. Bennett, F. W. Lester, *Catalog of Earthquakes along the San Andreas Fault in Central California, April–June, 1972* (Open-file report, U.S. Geological Survey, Menlo Park, Calif., 1972).
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 events considered 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).

25 February 1974; revised 1 April 1974

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.

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).

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

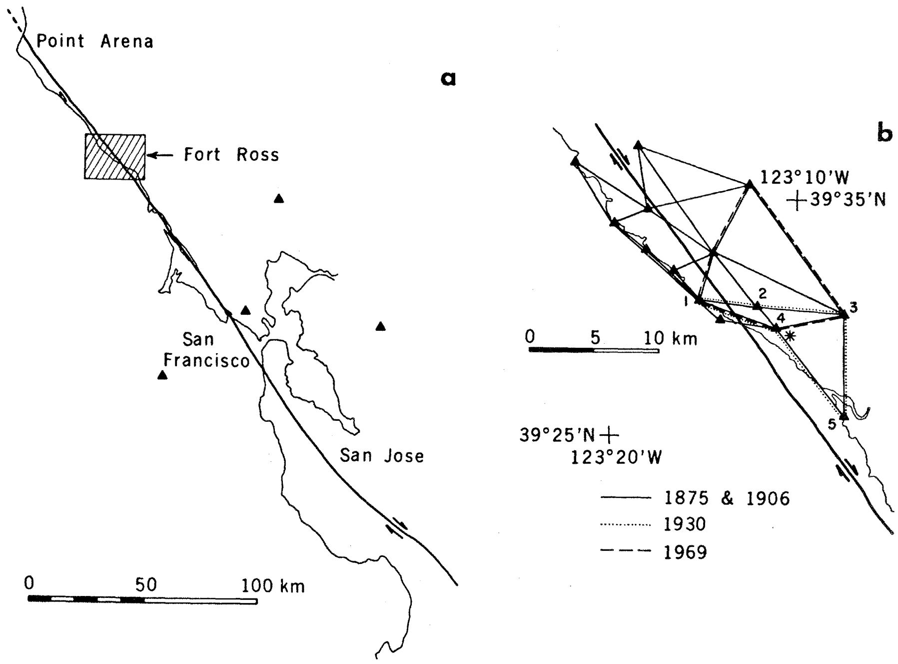

Fig. 1. (a) Map of the surface rupture of the 1906 San Francisco earthquake, showing the locations of primary arc triangulation stations (triangles) and the position of the Fort Ross network (hatching). (b) Fort Ross triangulation network, showing portions surveyed between 1875 and 1969. (*) Angle referred to in Fig 2a.

Here I consider only data in the zone of largest displacements, north of San Francisco. Figure 1a shows several long-line primary arc triangulation stations, and Fig. 1b shows a small-scale tertiary network that spans the fault in the vicinity of Fort Ross. The primary arc has been surveyed at roughly 20-year intervals since 1880, and the Fort Ross net was observed in 1875 and 1906, with partial resurveys in 1930 and 1969.

Primary evidence for postseismic movements comes from the Fort Ross data, and is best demonstrated by considering angle changes in triangles which do not span the fault, such as 3–4–5 in Fig. 1b. Angles changed by as much as 18 arc seconds in the time interval which includes the earthquake (1875 to 1906), and, with opposite sign, by more than 10 arc seconds during 1906 to 1930. Triangle closures indicate that the standard deviation of a single angle change is about 2 arc seconds for these data.

Consider now the effect of fault slip on observed angles for a fault model in which slippage varies only with depth; that is, motion is constant at a given depth along an infinitely long fault. Figure 2a shows, for the angle subtended by stations 5 and 3 at station 4, the angle change for 1 m of slip in successive 3-km increments of depth down to 21 km. For 1 m of slip below that depth, the angle changes have been combined. Changes are positive only for slip from 0 to 3 km in depth, and are significantly negative only below 6 km. Note that the observed angle changes for 1875 to 1906 require no less than 3 m of slip in the upper 3 km, and changes for 1906 to 1930 require about 3 m of slip below 6 km.

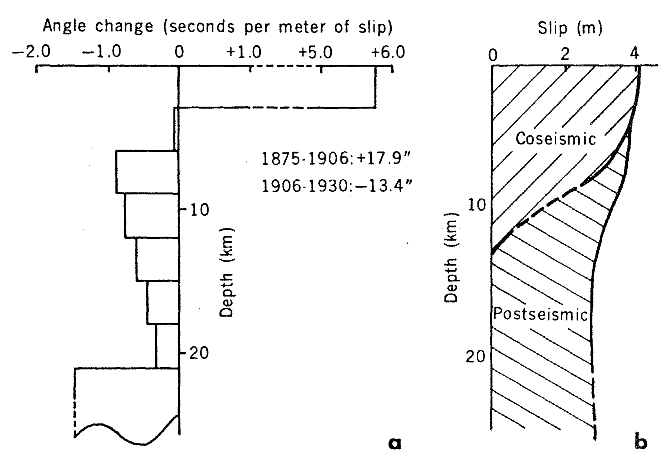

Fig. 2. (a) Angle changes for 1 m of slip in successive 3-km depth intervals on a vertical strike-slip fault, for the angle marked by * in Fig. 1b. Slip is constant at a particular depth along the strike of the fault, and for 1 m of slip below 21 km the total angle change has been combined here. (b) Schematic plot of a particular model of fault displacement versus depth which adequately satisfies the coseismic (1875 to 1906) and postseismic (1906–1930 and 1906–1969) triangulation data. Other admissible models are possible, as discussed in the text.

Other data from the Fort Ross net show similar features. Angle changes for triangulation stations occupied or observed in all four surveys (stations 1 to 5 in Fig. 1b) are given in Table 1. Where common data exist, changes for 1906 to 1930 agree with those for 1906 to 1969 within their uncertainties, indicating little, if any, displacement from 1930 to 1969. In all, 22 independent angle changes were measured for 1875 to 1906, 6 for 1906 to 1930, and 14 for 1906 to 1969. Observations have been fit to both two- and three-dimensional faults by using displacement dislocation models and linear inversion techniques (3). Schematically shown in Fig. 2b is a two-dimensional solution which adequately satisfies the data. It shows about 4 m of displacement to 10 km at the time of the earthquake (coseismic), and at least 3 m below 10 km sometime between 1906 and 1930 (postseismic). It is shown here in cartoon fashion to emphasize that other acceptable models are possible. For example, if postseismic slip extended only to about 25 km, then about 4 m of slip would be required. If it were as shallow as 6 km, 3 m of slip to 10 km and 2 m of slip below 10 km would also be satisfactory. Other approximate models can be constructed by referring to Fig. 2a.

Surveys from triangulation stations up to 40 km from the fault (Fig. 1a) can potentially provide constraints on possible postseismic fault slip below about 20 km. Such evidence examined to date is not conclusive, although block motions across the fault greater than about a meter are excluded. Some of the relevant data have already been published (4).

The time constant for the postseismic displacements is not well determined. The Fort Ross surveys were carried out about 8 months after the 1906 earthquake and again in early 1930. At Point Arena (Fig. 1) a similar small-scale network was surveyed in late 1906, and a few angles were reobserved in 1925. These data, although fragmentary, are also consistent with 2 to 3 m of slip below 10 km. Most of the movements must then have occurred sometime between 1907 and 1925, but no further refinements on this time scale can be provided. The slip must have been predominantly aseismic, since seismic motion of 3 m over a 10-km depth interval and at least 70 km of fault (the distance from Point Arena to Fort Ross) would correspond (*5*) to an earthquake of at least magnitude 7, and no such event has occurred in this region since 1906.

The post-1906 observations suggest that strains for seismic faulting are not accumulated primarily by relatively continuous aseismic motion on the fault plane immediately below the seismic zone. Strain accumulation might be due to either steady or irregular motions on the fault plane at greater depths (say below about 30 km), in which case shear strains would be greatest at the fault trace and decrease away from it. Alternatively, the fault may be locked throughout the entire thickness of the lithosphere, with the San Andreas system loaded regionally by shear tractions applied far from the fault at the base or edges of the lithospheric plates. In this case, shear strains measured at the earth's surface by geodetic means would be approximately constant over a wide region on either side of the fault trace. A third alternative is that a "locked" section of the fault is loaded entirely by seismic or aseismic slip at its ends. It is also possible that strain is accumulated and released along many subsidiary subparallel faults as well as on the San Andreas itself. Each of these mechanisms produces a different pattern of surface deformation, and geodetic observations are potentially useful in differentiating between them. I consider that the data which have been examined to date are inconclusive in this regard, and thus the precise mode of strain accumulation remains an open question.

Some limited evidence from northern California does suggest irregular strain accumulation across the San Andreas fault in the half century before the 1906 earthquake. Astronomical azimuths, accurate to about 0.5 arc second, were observed in 1859, 1882, and 1907 between the primary triangulation stations Mount Diablo and Tamalpais (the second and third from north to south plotted in Fig. 1a) (*6*). This azimuth increased by 7.84 arc seconds in 1859 to 1882 and decreased by 0.55 arc second in 1882 to 1907, corresponding to relative right lateral motions parallel to the fault of +2.8 and −0.2 m, respectively. Reduction of the independent (though less accurate) triangulation network data gives changes of 5.38 and −1.57 arc seconds for essentially the same time intervals (*6*). The smaller change in the interval 1882 to 1907 is, at least in part, due to the elastic rebound that occurred in 1906, since station Tamalpais is only 15 km from the fault trace. The relative motion of 2.8 m during 1859 to 1882 far exceeds the rate since 1906 inferred from triangulation (*4*). An earthquake in 1868 on the Hayward fault, about 25 km from either station, appears too small to have contributed significantly to this large relative motion (*7*). Hence, these data, although limited, indicate some anomalous large crustal deformation away from the fault in the approximately 30 or more years preceding the 1906 earthquake.

The actual postseismic motions are probably more complex than indicated by this model of dislocation sources in a perfectly elastic half space. They should be understood as simple analog models which are convenient for discussion, and not necessarily realistic physical models of the actual time-dependent nonelastic processes. For example, a viscoelastic relaxation mechanism such as that proposed (*8*) to account for the vertical crustal movements which followed the thrust-type earthquake (magnitude, 8.2) in Nankaido, Japan, may be appropriate here.

Finally, the work reported here may have some relevance for the question of whether large earthquakes cause secular shifts of the earth's rotational pole or excite the earth's Chandler wobble (*9*). Should large postseismic displacements be a common feature of strain release by great earthquakes, it seems likely that these aseismic movements would be more important than the earthquakes themselves in exciting the earth's polar motions.

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References and Notes

1. H. F. Reid, in *The California Earthquake of April 18, 1906, Report of the State Earthquake Investigation Commission* (Carnegie Institution of Washington, Washington, D.C., 1910), vol. 2, pp. 16–28.
2. Triangulation was carried out by the U.S. Coast and Geodetic Survey, now the National Geodetic Survey (NGS), and the observed direction lists have been made available to me by the courtesy of B. K. Meade of NGS.
3. G. Backus and F. Gilbert, *Geophys. J. R. Astron. Soc.* **16**, 196 (1968); *Phil. Trans. R. Soc. Lond. Ser. A Math. Phys. Sci.* **266**, 123 (1970); R. Wiggins, *Rev. Geophys. Space Phys.* **10**, 251 (1972); D. D. Jackson, *Geophys. J. R. Astron. Soc.* **28**, 97 (1972). The inversion methods described in these papers have been applied to geodetic data by W. Thatcher [*EOS Trans. Am. Geophys. Union* **54**, 1132 (1973)].
4. J. C. Savage and R. O. Burford, *J. Geophys. Res.* **78**, 832 (1973).
5. W. Thatcher and T. C. Hanks, *ibid.*, p. 8547.
6. J. F. Hayford and A. L. Baldwin, *The California Earthquake of April 18, 1906, Report of the State Earthquake Investigation Commission* (Carnegie Institution of Washington, Washington, DC., 1908), vol. 1, pp. 114–145.
7. Surface faulting was observed along 32 km of the Hayward fault, with right-lateral offsets varying from 25 to 90 cm (A. C. Lawson *et al.*, *ibid.*, p. 435).
8. A. Nur and G. Mavko, *Science* **183**, 204 (1974); H. Kanamori, *Annu. Rev. Earth Planet. Sci.* **1**, 213 (1973).
9. The question remains a controversial one; the spectrum of views is represented by: D. E. Smylie and L. Mansinha, *Geophys. J. R. Astron. Soc.* **23**, 329 (1971); M. Israel, A. Ben-Menahem, S. J. Singh, *ibid.* **32**, 219 (1973), F. A. Dahlen, *ibid.*, p. 203.
10. I thank J. C. Savage and A. H. Lachenbruch for helpful discussion.

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Earthquake Mechanics in the Central United States

Abstract. *Focal mechanism solutions of earthquakes in the central United States suggest that local stress fields are important in determining the type and orientation of faulting. The implied stress system is considerably more complicated than that which would be produced by east-west trending compressive stresses, as previously suggested for this region.*

Sbar and Sykes (*1*) proposed a relatively simple stress model for the eastern portion of the North American continent. Using data obtained from geological observations, in situ stress measurements, and fault plane solutions, they concluded that the central United States is presently experiencing a predominantly horizontal compressive stress whose axis tends east-northeast. A consequence of this model is that ongoing earthquake activity in this region should be mostly of the high-angle, thrust-type faulting with a strike in the north-south direction. However, a detailed investigation of moderate size earthquakes that occurred during the last 13 years within the area of Missouri, Kentucky, Tennessee, Illinois, Mississippi, and Arkansas indi-