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PRELIMINARY RESULTS FROM A SEARCH FOR REGIONAL TECTONO-MAGNETIC EFFECTS IN CALIFORNIA AND WESTERN NEVADA

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ABSTRACT

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Physical theory and laboratory experiments both indicate that tectonomagnetic effects in seismically active areas should be detected with highly sensitive drift-free differential magnetometers. By using a pair of synchronized 0.25 γ absolute magnetometers to measure precisely field differences between 70 adjacent sites with a 10—15 km separation, more than 1,000 km of faults in California and western Nevada have been monitored for anomalous changes in the local magnetic field. Over a nine-month period, four sets of measurements have been completed along 350 km of the San Andreas and two sets along the Excelsior Mountains, Mono Lake, and Owens Valley. Preliminary results show that significant changes have occurred between each subsequent data set and that these changes appear to be related to tectonic structure and seismicity. This method looks promising as a simple inexpensive scheme for indicating some hazardous sections of active faults, particularly in countries where extensive geophysics programs are not feasible.

INTRODUCTION

Of the potential techniques for predicting seismic activity, few have proved more promising in theory, yet more disappointing in practice, than that based on tectonomagnetism. Tectonomagnetism (Nagata, 1969) is the field observation accompanying geologic activity, of piezomagnetic effects in rocks. Piezomagnetism and its thermodynamic inverse, magnetostriction, can be clearly demonstrated under laboratory conditions for stresses and temperatures comparable to those occurring naturally in the upper 10 km of the earth's crust (Kapitsa, 1955; Ohnaka and Kinoshita, 1968). The physical principles are well understood and have been interpreted in terms of multidomain theories of domain structure in rocks containing magnetic minerals (Kern, 1961; Stacey, 1963, Stacey and Johnston, 1972) and in terms of single domain theories (Nagata, 1966, 1970; Kinoshita, 1969). It is surprising, therefore, that no clear observations of this effect have been made in earthquake-prone areas where crustal stress variations are known to occur.

Records of magnetic field effects attributed to earthquakes date back to the eighteenth century. Observations prior to the early 1900's have all been found to be spurious by Reid (1914), who showed that they were produced by mechanical, not magnetic disturbances during the passage of seismic waves under the sensor. The results to 1968 have been reviewed by Rikitake (1968), who found the majority unconvincing with amplitudes radically diminishing with time as improved measurement techniques became available, particularly with the introduction of absolute magnetometers.

Some recent observations are encouraging. These include:

(1) Local magnetic anomalies that resulted from the crustal load as a large dam was filled in southern Australia (Davis and Stacey, 1972).

(2) Magnetic anomalies sensed on volcanoes in New Zealand prior to eruption (Johnston and Stacey, 1969a, 1969b).

(3) Permanent local magnetic field changes that occurred after the Cannikin nuclear blast on Amchitka Island (Hasbrouck and Allen, 1972). The effects of the shock wave on magnetization are difficult to interpret in this case.

The preliminary results from a program designed to search for tectonomagnetic effects on the San Andreas and other active faults in California and western Nevada are outlined in this paper.

BRIEF OUTLINE OF THEORY OF THE PIEZOMAGNETIC EFFECT AND CALCULATION OF TECTONOMAGNETIC EFFECTS

Rocks containing grains of magnetic materials exhibit both an induced and a remanent magnetization that result from a vector sum of the magnetizations from each grain. The contribution from a particular grain is a result of the individual domain structure within that grain. For grain sizes exceeding the single-domain limit, the domain structure is sensitive to stress, temperature, and magnetic field. Processes controlling rock magnetization have been reviewed by Kittel (1949) and Stacey (1963).

Within the upper 10 km of the earth's crust, the range of deviatoric stress variation, magnetic field, and temperature expected is generally less than 100 bars, 0.5 oersteds and 800°C, respectively. Under these conditions the simple theory of piezomagnetism and laboratory observations (Kapitsa, 1955; Ohnaka and Kinoshita, 1968) indicate a reversible change in magnetization by a process of individual domain rotation with stress. This process can include small, reversible movements of 71° and 109° domain walls but do not major readjustments. The range considered appears to include all tectonically interesting deviatoric stresses.

The calculated change in susceptibility both parallel χ_p and normal χ_n to a stress σ_s (Stacey and Johnston, 1972) is:

$$\chi_p = \chi_p(0)[1 - S_\chi \sigma_s] \qquad (1)$$

$$\chi_n = \chi_n(0)[1 + 0.5\, S_\chi \sigma_s] \qquad (2)$$

where S_χ, the stress sensitivity has values in the range 0.1 to $1.0 \cdot 10^{-3}$ cm^2/kgm for reasonable compositions of titanomagnetites. The corresponding expressions for the stress-dependence of remanence both parallel $I_\parallel$ and perpendicular $I_\perp$ to a stress σ_s are:

$$I_\parallel = I_\parallel(0)\ [1 - S_R \sigma_s] \qquad (3)$$

$$I_\perp = I_\perp(0)\ [1 + 0.5\, S_R \sigma_s] \qquad (4)$$

where S_R has values in the range 0.1 to $2 \cdot 10^{-3}$ cm^2/kgm, again for varying compositions of titanomagnetites.

A number of calculations of expected tectonomagnetic anomalies for typical fault models have been made, notably by Stacey (1963); Stacey et al. (1965); Yuketake and Tachinaka (1967); and Shamsi and Stacey (1969). Because of the heterogeneity of both rock properties and the likely stress fields, these models can at best provide only crude estimates of the amplitude (~ few gammas) and spatial extent of such anomalies.

Critical factors in detecting real tectonomagnetic effects are:

(1) *Noise reduction.* In order to observe local field changes of a gamma or so, it is necessary to discriminate against the normal geomagnetic field variations whose primary source is in the ionosphere and magnetosphere and whose amplitude can be 50—100 times greater. To do this, it is necessary to operate magnetometers in a differential mode and to obtain good estimates of the background noise and its periodicity for various sites. Several detailed experiments have been done on this problem (Stacey and Westcott, 1965; Rikitake, 1966; Rikitake et al., 1968; Johnston, 1970; Mori and Yoshino, 1970).

(2) *Site selection.* Little is known of the location and spatial extent of the stress fields and the quantity and shape of sufficiently magnetic rocks in seismically active regions.

(3) *Instrument selection.* Although the primary instrumental requirements are optimum stability and sensitivity, a difficult choice must be made, whether to use absolute total field instruments when the chance of a spurious instrumental effect is small or to use a component measuring magnetometer. There is some indication that for strike-slip faults such as the San Andreas the horizontal components of local magnetic anomalies may be slightly more sensitive to stress than the total field (Shamsi and Stacey, 1969).

DETECTION PROGRAM

Recognizing that the most critical problems in detecting tectonomagnetic effects are in obtaining clear unambiguous observations and in being able to monitor large regions of the California and Nevada seismic zone, quickly and cheaply, we chose to use 0.25 γ total field proton magnetometers in a three-stage detection program. The first stage has been to search for broad-scale long-period effects with a pair of absolute 0.25 γ proton magnetometers, operated synchronously in "leap frog" fashion at more than 70 sites along the San Andreas fault and other active faults in California and western Nevada. Over a nine-month period four runs have been completed along one 350 km section of the San Andreas fault system south from San Francisco, two along Mono Lake, the Excelsior Mountains, the White Mountains and Owens Valley and one each along the San Jacinto fault, the Hayward—Calaveras and Rodgers Creek faults, and the section of the San Andreas from Cholame to the Tehachapi Mountains. The location of the sites from San Francisco to south of Cholame are shown on the geologic map of central California in Fig. 1. As tectonic features are likely to be important in an experiment of this type, site selection was always made with this in mind. For example, sites 37 and 34 are on outcrops of Miocene volcanic rocks at each end of a 40 km long Mesozoic granite batholith that forms the Gabilan Range.

At each pair of adjacent sites, a set of about 75 total field values were recorded. The magnetometers are synchronized by radio, and the data later used to generate mean differences and standard deviations. The largest standard deviation for these single samples was 0.5 γ. The standard deviation in 10-minute averages of local field differences for this site separation will be determined precisely when the installation of an array of permanent stations is completed. However, preliminary data from a rubidium magnetometer array (Smith et al., 1973) indicates a value of 1.8 γ. The time changes in 10-minute means of local field differences are therefore only considered to be significant (at the 95% confidence limit) if they exceed 3.5 γ.

Although surface samples were collected at most of the sites, extrapolation from these to an average regional magnetization is not entirely satisfactory. Some idea of the regional distribution of magnetization can also be obtained from the detailed aeromagnetic map of the fault (Hanna et al., 1972). It appears, therefore, that if local magnetic field changes are produced by stress changes along the fault, then development of a relation between earthquake magnitude and anomalies in a particular region will be largely empirical.

The variations with time of local magnetic field changes that have occurred from October 1972 to June 1973, are shown in Fig. 2. This represents four complete runs along the section of the San Andreas fault system shown in Fig. 1. The first three plots show the changes that occurred between October 1972 and February, March, and June, respectively in 1973. They are obtained simply by subtracting the local field differences for each pair of stations at each subsequent survey from the values obtained in the original survey. Similarly the next two plots show the February to March and the February to June variation while the final plot shows the March to June variation. Two standard deviations (2σ) correspond to 3.6 γ.

Although the largest changes are only marginally above the expected noise, it is encouraging that anomalous changes for independent sets of measurements occurred at the same sites and that the scale of the anomalies exceeded, in some cases, the site separation.

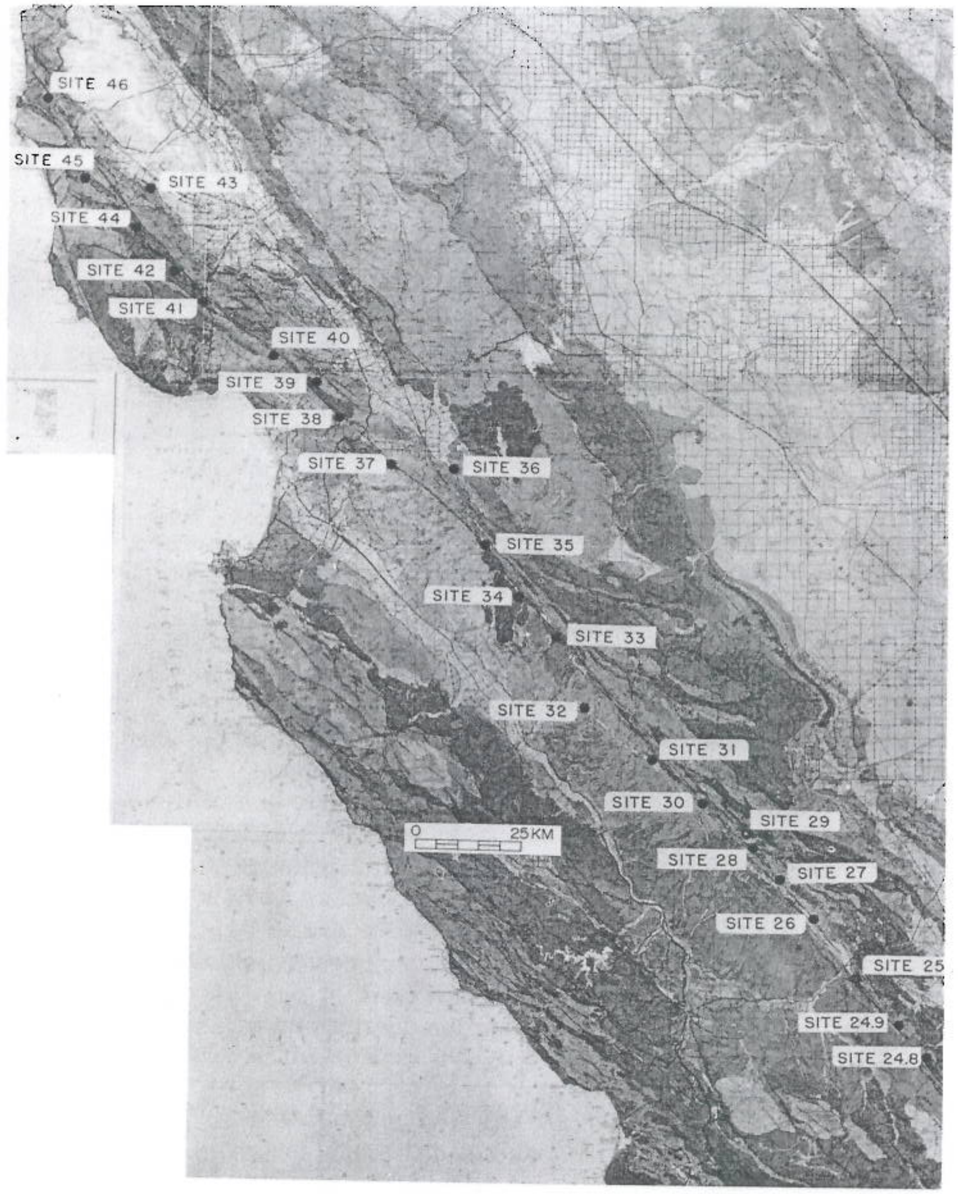

Fig. 1. Map of central California showing geology and the location of the magnetometer sites most frequently resurveyed along the San Andreas fault. Site 27 is now not used and has been replaced by site 28.

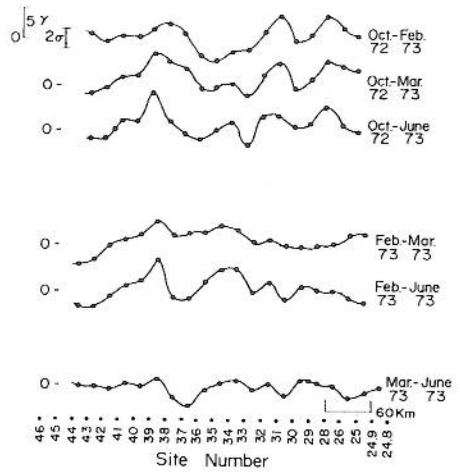

Fig. 2. Time variations of local magnetic field differences as a function of site position for the sites along the San Andreas fault shown in Fig. 1. The standard deviation, σ, for 10-minute averages of the difference field is equal to 1.8 γ. The average site separation is 15 km.

The features of interest concerning these results and their relation to the tectonic features are:

(1) A consistent decrease, over the nine-month period, of the local magnetic field at the southern end of the Santa Cruz Mountains.

(2) A decrease in local magnetic field at the northern end of the Gabilan Range (site 37) until March 1973, but an increase since March.

(3) An increase at the southern end of the Gabilan Range (site 33 and 34).

(4) A decrease between October 1972 and February 1973 at Monarch Peak (site 31) and Slack Canyon to Parkfield (site 26 and 28) that has remained generally steady, perhaps increasing slightly since then. Site 28 has been moved to near site 27 and site 27 is now not used.

A continuing check has been made on the distribution of central Californian seismicity in the region covered by this experiment. Earthquakes with magnitudes greater than 3.0 that have occurred since the experiment was begun together with their location and occurrence time are listed in Table I. The relation between these earthquakes and the local magnetic field variations is shown in Fig. 3. There appears to be some correspondence between the location of these earthquakes and the largest field variations. It is worth pointing out that the dimensions of these earthquakes is small (~ few kilometers) compared with the scale of the magnetic anomalies.

Time-dependent local magnetic field changes have also been found across the Excelsior Mountains in Nevada (Johnston et al., 1973). This area is both seismically interesting (Ryall et al., 1970) and highly magnetic. No significant changes were obtained from here to the southern end of Owens Valley.

TABLE I

Local earthquakes with magnitudes greater than 3.0 that have occurred in central California since October 16, 1972, their locations and time of occurrence (Karen Meagher, U.S. Geological Survey, unpublished data, 1973)

No.	Date	G.M.T.	Lat. (deg)	N	Long. (deg)	W	Depth (km)	Magn.
E31	10-19-72	0028	36	06.6	120	41.6	5.0	3.0
E21	11-13-72	2310	36	30.1	121	8.6	5.7	3.3
E7	1-15-73	0943	36	40.5	121	20.0	5.6	4.0
E8	1-15-73	1008	36	40.3	121	19.9	6.3	3.0
E9	1-15-73	1013	36	40.6	121	20.4	6.8	3.2
E10	1-15-73	1023	36	40.2	121	19.7	5.7	3.7
E11	1-15-73	1441	36	40.2	121	19.8	5.7	3.7
E12	1-15-73	2013	36	39.6	121	19.0	6.7	3.3
E13	1-15-73	2114	36	39.4	121	18.9	5.9	3.0
E14	1-20-73	1230	36	39.3	121	18.7	6.2	3.4
E15	1-21-73	1823	36	38.9	121	18.7	6.2	3.0
E16	1-21-73	2221	36	38.8	121	18.7	8.0	3.2
E33	1-23-73	1305	35	55.7	120	31.7	10.5	3.6
E17	1-26-73	0037	36	41.4	121	21.4	5.3	3.0
E22	1-28-73	2225	36	31.1	121	07.6	9.7	3.0
E23	1-29-73	0746	36	46.3	121	28.8	4.2	3.0
E18	2-13-73	1741	36	38.7	121	17.6	4.7	3.3
E24	3-14-73	0824	36	32.4	121	05.0	13.4	3.2
E25	3-14-73	1545	36	33.1	121	05.7	12.0	3.0
E26	3-19-73	0037	36	33.3	121	05.9	13.0	3.0
E4	5-24-73	0945	36	50.1	121	35.1	5.8	3.1
E27	6-22-73	0129	36	33.8	121	12.3	9.5	3.9
E34	6-30-73	0823	35	52.6	120	25.5	4.7	3.0
E19	7- 4-73	1702	36	39.2	121	18.1	5.5	3.0
E29	7- 8-73	0259	36	09.1	120	45.0	10.5	3.0
E30	7- 8-73	0340	36	09.0	120	45.1	9.6	3.1
E5	7- 9-73	2100	36	48.7	121	32.9	6.2	3.3
E1	8- 2-73	0426	36	52.9	121	37.3	5.8	3.0
E2	8- 2-73	0544	36	52.8	121	37.2	5.7	3.1
E28	8- 2-73	1610	36	32.7	121	10.5	7.9	3.3
E6	8- 7-73	0417	36	46.8	121	29.3	5.0	3.0
E20	8-10-73	1135	36	35.9	121	13.5	3.8	3.1
E32	8-20-73	1319	36	01.3	120	37.8	5.2	3.0
E3	8-26-73	0734	36	51.6	121	35.9	5.7	3.0

Other aspects of this program include using the filling of a dam in southern California in a tectonomagnetic control experiment for a typical Californian geologic environment and the installation of an array of permanent magnetometer stations along the San Andreas (Johnston et al., 1973).

CONCLUSION

If the changes observed are indeed related to subsurface stress variations,

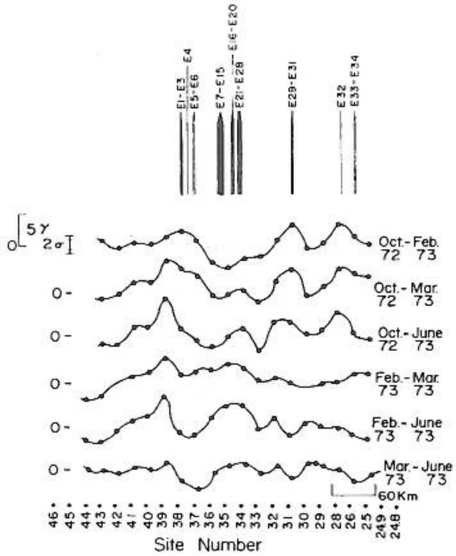

Fig. 3. Relation between the locations of earthquakes with magnitudes equal to or greater than 3.0 and time-dependent magnetic anomalies along the San Andreas fault. *E* refers to earthquake number listed in Table I.

the results appear to support a model of the San Andreas fault system where higher seismicity occurs in regions where changes in stress are relatively rapid. If similar behavior occurs in other fault systems then this method looks promising as a simple inexpensive scheme for indicating some hazardous sections of active faults, particularly in countries where extensive geophysics programs are not possible. However, since no large earthquakes have occurred within the surveyed region since the start of the experiment, such general conclusions may be premature.

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