J. Geomag. Geoelectr., **37**, 1051–1061, 1985

A Comparison of Proton and Self-Calibrating Rubidium Magnetometers for Tectonomagnetic Studies

Randolph H. WARE[1], Malcolm J. S. JOHNSTON[2], and Robert J. MUELLER[2]

[1]*Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, CO 80309 U.S.A.*
[2]*U.S. Geological Survey, Menlo Park, CA 94025, U.S.A.*

(Received May 7, 1985; Revised August 5, 1985)

A network of 27 proton magnetometers (PM's), designed to detect stress related magnetic events of crustal origin, has been operating near active faults in California for the past decade. We present here comparative magnetic difference field data from PM's used in this net with that obtained from new self-calibrating rubidium magnetometers (SCR's). The instruments were first compared over a 50-m baseline in an aseismic and magnetically quiet region in Colorado. For PM's having either a 0.25-nT or a 0.125-nT least count, the observed difference variations were 0.2-nT and 0.17-nT rms, respectively. For SCR's having a 0.001-nT least count and a 100-second averaging interval, the difference variation was 0.002-nT rms. Power spectra of these data indicate that the noise for the PM's is close to their least count limit. However, the SCR noise decreases at about 20 dB per decade until it approaches its least count limit 40 dB below the PM limit, for periods less than 30 minutes. A similar experiment was conducted using collocated SCR and PM pairs separated by 13 km along the San Andreas fault. Power spectra indicate that both systems are equivalent and are dominated by external noise at periods greater than 4 minutes. Below 4 minutes the PM noise approaches its least count limit while the SCR noise continues to decrease at about 20 dB per decade until it is 20 dB below the PM limit at a period of 30 seconds. Improved discrimination of magnetic transients caused by fault activity with periods of several minutes to perhaps an hour appears to be possible with higher sensitivity magnetometers.

1. Introduction

Changes in crustal stress are expected both to precede and to occur simultaneously with earthquakes. The fact that the magnetic properties of various crustal rocks are sensitive to stress leads to the possibility of using magnetic measurements to monitor crustal stress. Attempts to observe such phenomena, known as tectonomagnetic events, have met with some success (BREINER and KOVACH, 1967; SMITH and JOHNSTON, 1976; DAVIS *et al.*, 1980; RIKITAKE *et al.*, 1980; SHAPIRO and ABDULLABEKOV, 1982). However, the most easily identified tectonomagnetic event (i.e., the coseismic change expected to accompany rupture), has not yet been unambiguously recorded at sites near moderate to large earthquakes. Improved resolution, particularly at high frequencies, may allow these measurements to be made.

Extensive efforts to observe tectonomagnetic events have been concentrated along active faults in California where the U.S. Geological Survey has monitored local magnetic fields at several hundred sites since 1973, mostly with proton magnetometers (PM's). Recently, several new magnetometers with improved sensitivity and high accuracy have become available. We compare here the performance of new self-calibrating rubidium magnetometers (SCR's) collocated with PM's at typical sites in seismically active regions of the San Andreas fault, and at sites in a seismically quiet region in Colorado.

2. Instrumentation

The high-accuracy instruments used in this comparison are self-calibrating rubidium magnetometers accurate to 0.01-nT (WARE, 1983). In the SCR, Rb^{87} atoms are polarized by optical pumping. The precession frequency of the aligned atoms, which is proportional to the total magnetic field, is counted to give a field average during the counting interval. A 10-second counting interval gives a 0.014-nT least count uncertainty. Although the least count can be arbitrarily reduced by increasing the averaging interval, the SCR is limited to an absolute accuracy of ± 0.007-nT by the uncertainty in atomic constants (ALLEN and BENDER, 1972).

The USGS proton magnetometers (Geometrics model G-816 or G-826) have been modified by substituting a more accurate reference oscillator having a temperature sensitivity of less than 10^{-8} per °C, in order to increase stability. The least count uncertainty is normally 0.25-nT, but it was reduced to 0.125-nT in several USGS PM's. The local field is measured by counting the precession frequency of protons that have been polarized by a pulsed, impressed field. The USGS magnetometer network consists of 27 PM's located along the San Andreas fault from just south of San Francisco to the Salton Sea (MUELLER *et al.*, 1981). All instruments sample synchronously every 10 minutes and the data are transmitted by digital telemetry to the USGS laboratory in Menlo Park for routine analysis and display.

It is important to note that for some measurement intervals the SCR and the PM will give inherently different results. The SCR measures the field continuously, and averages or integrates for any arbitrary period. The PM measures the field during 1.5-second periods separated by arbitrary intervals greater than 7 seconds. Therefore, an exact comparison of the two types of instruments is possible only during the 1.5-second PM sampling interval. For this sampling interval, the SCR precision is limited to 0.09-nT rms by least count noise. In general, with no averaging, the PM instrument noise ranges between 0.2 and 0.3-nT rms. For hourly averages of typical USGS PM difference data taken at 10-minute intervals, the residual variation is 0.12-nT rms (JOHNSTON *et al.*, 1984).

3. Aseismic Region

In order to compare the instrument precision and noise levels of SCR's and USGS PM's, a short-baseline comparison was carried out in seismically inactive, geologically simple, and magnetically quiet region on the grounds of the Boulder Magnetic Observatory in Colorado. The PM's were separated by a 50-m east–west baseline, and the two SCR's were separated by 15 m from the PM's, to avoid detection of the PM polarizing field by the SCR. The results of this experiment, using 0.25-nT PM's, are shown in Fig. 1. The SCR data show that the local noise is 0.002-nT rms or less for this period of 16 hours. In contrast, the PM data in this quiet site are limited by instrument noise at 0.2-nT rms.

A similar experiment was performed using 0.125-nT PM's (Fig. 2). The PM noise is 0.17-nT rms, which is considerably higher than the 0.06-nT rms least count and digitization noise expected for the difference between two perfect 0.125-nT least count instruments (JOHNSTON *et al.*, 1984). The PM data, processed by a 30-minute filter, are also shown in Fig. 2. The additional filtering reduces the high-frequency noise to 0.05-nT rms. The background noise level obtained by the SCR is 0.008-nT rms.

The power spectral density for these short-baseline PM data (1.5-second averages) and SCR data (100-second averages) and their theoretical least count noise levels are shown in Fig. 3. Over the entire frequency range the PM spec-

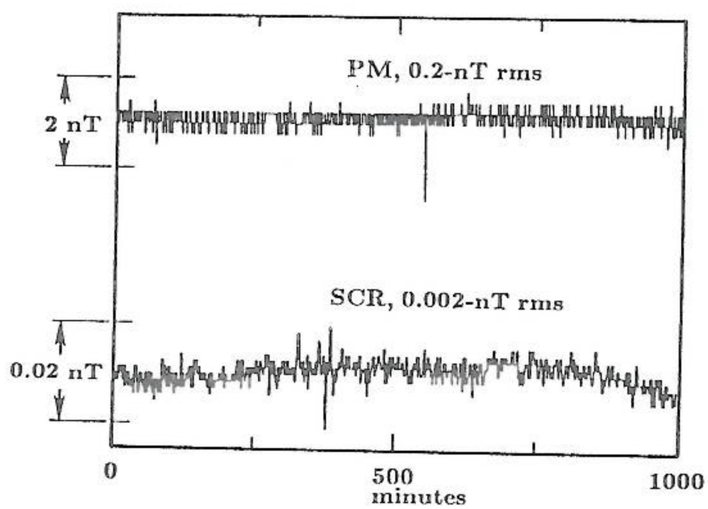

Fig. 1. Short-baseline (50 m) total field differences observed in Colorado. The PM data are 1.5-second field averages observed every minute with a 0.25-nT least count. The SCR data are consecutive 100-second field averages (measurements) with a 0.0014-nT least count. Note the different scale factors.

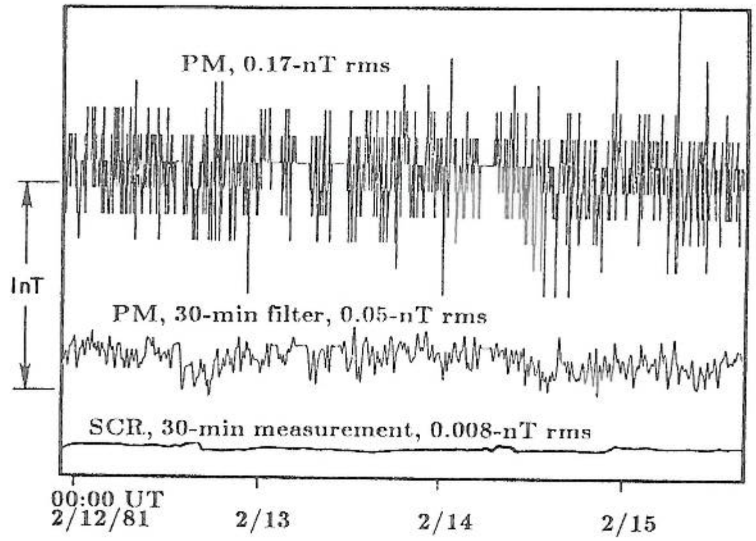

Fig. 2. Short-baseline (50 m) total field differences observed in Colorado. The PM data are 1.5-second averages observed every minute with a 0.125-nT least count. The SCR data are consecutive 30-min measurements with a 0.0001-nT least count.

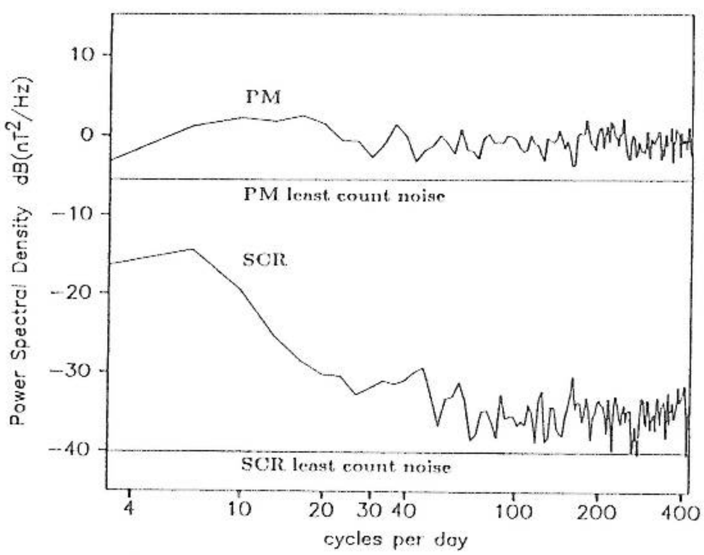

Fig. 3. Power spectral density of short-baseline (50 m) differences observed in Colorado. The 0.125-nT least count PM data are 1.5-second averages observed every minute. The SCR data are consecutive 100-second measurements with a 0.0014-nT least count.

trum is close to its least count noise limit. At periods less than about one hour the SCR curve approaches its least count noise limit, 40 dB below that of the PM.

Observations were also made over longer baselines in Colorado using PM's and SCR's. Hourly averages of more than 4 days of 0.25-nT PM differences recorded at 1-minute intervals on two different 12-km baselines varied by 0.17-nT rms (JOHNSTON *et al.*, 1984). Hourly averages of more than one month of SCR differences on the same two baselines varied by 0.11-nT rms (WARE, 1979). These results suggest an instrument noise level of 0.13-nT rms for hourly averages of 0.25-nT PM differences. This agrees with the 0.9 to 0.15-nT rms found by JOHNSTON *et al.* (1984) during short baseline tests in California and Colorado.

4. Seismic Region

In order to determine the effects of instrument precision and averaging interval in typical USGS data, two SCR's were operated at PM sites separated by 13 km in central California, 150 km southeast of San Francisco on the San Andreas fault. The location of these two sites, San Juan Bautista (SN) and Harris Ranch (HA), are indicated by JOHNSTON *et al.* (1983; 1984). The SCR frequency at HA was changed to audio frequency and transmitted by telephone to SN, where the difference field frequency was digitally recorded. The PM differences from these sites were the usual 0.25-nT, 1.5-second field averages taken at 10-minute intervals and the SCR differences were continuous 100-second measurements (Fig. 4). The rms value for the PM data was 0.47-nT and for the SCR data was 0.36-nT. Higher resolution as a result of higher sensitivity and smoothing by the inherent averaging is clearly evident in the SCR differences for periods less than several hours. However, longer period fluctuations dominate both data sets.

An additional test using a portable version of the standard USGS PM's, modified to have a least count sensitivity of 0.125-nT, was carried out at SN and HA. The PM's recorded 1.5-second field averages at 15-second intervals on site using digital printers, while the SCR's recorded continuous 10-second averages (Fig. 5). The rms values for the PM and SCR data were both 0.33-nT. Although both data sets are dominated by variations with periods greater than 10 minutes, least count and digitization noise is apparent in the PM data. Of course, aliasing of short-term fluctuations in the local geomagnetic field could also contribute to the apparent high-frequency PM noise. However, inspection of SCR differences recorded at one-second intervals indicates that variations over a few tens of seconds did not exceed one least count (0.14-nT). The 1.5-second PM noise level during the same time period was about four least counts (0.5-nT). Noise levels of similar magnitude are also present in other 0.125-nT PM data (Figs. 2 and 5). This result rules out aliasing of short-term variations in the local magnetic field as the source of the observed high-frequency PM noise.

The noise limits of the PM's at high frequency are shown in comparative plots of power spectral density (Fig. 6). The SCR's exceed the resolution of the

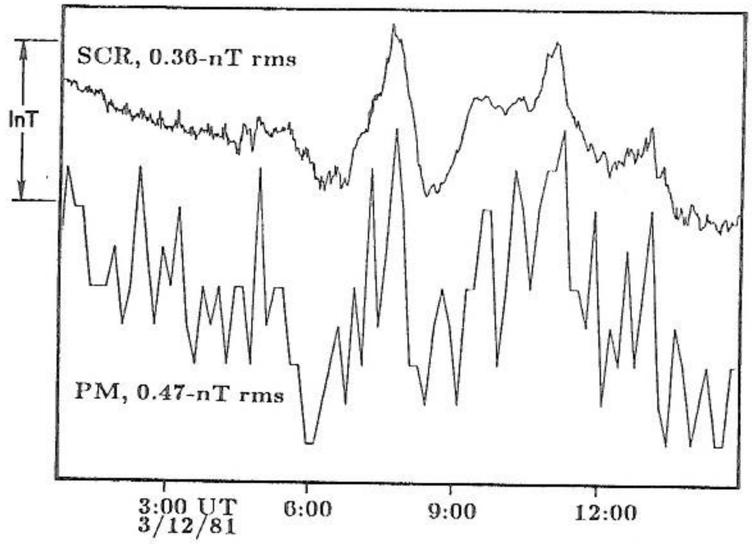

Fig. 4. Simultaneous differences between SN and HA. The SCR data are consecutive 100-second measurements with a 0.0014-nT least count. The 0.25-nT least count PM data are 1.5-second averages observed every 10 minutes.

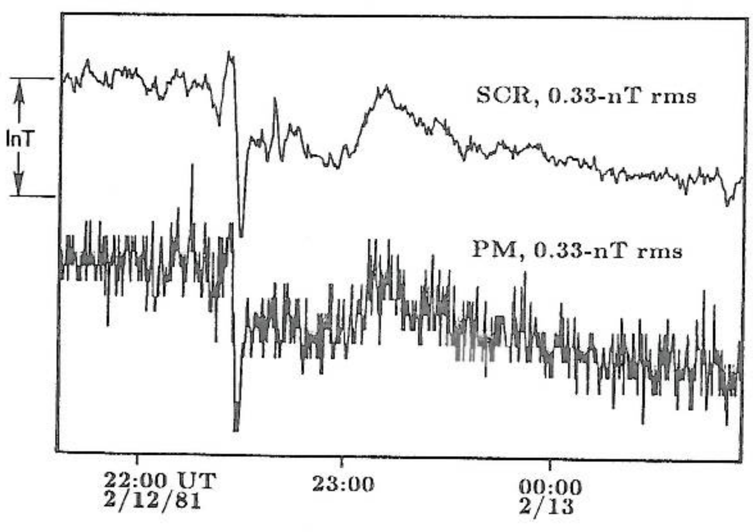

Fig. 5. Simultaneous differences between SN and HA. The SCR data are consecutive 10-second measurements with a 0.014-nT least count. The 0.125-nT least count PM data are 1.5-second averages observed every 15 seconds.

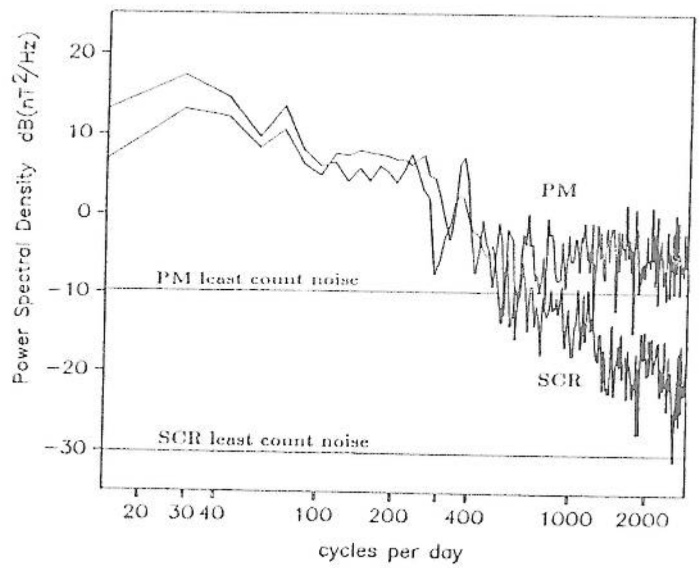

Fig. 6. Power spectral density of the simultaneous differences between SN and HA shown in Fig. 5. The SCR data are consecutive 10-second measurements with a 0.014-nT least count. The 0.125-nT least count PM data are 1.5-second averages observed every 15 seconds. The PM and SCR power spectral densities are essentially equal for frequencies less than 400 cycles per day.

0.125-nT PM's by up to order-of magnitude for periods less than 2 minutes (720 cycles/day). For periods greater than 3.5 minutes (420 cycles/day) both systems are equivalent and both are limited by noise of external origin which decreases, as shown previously for larger data sets (JOHNSTON *et al.*, 1984), by about 3 dB per decade of frequency. For shorter periods the PM spectrum approaches its least count limit but the SCR spectrum shows that the external or non-tectonic geomagnetic noise continues to decrease with increasing frequency at about the same rate of 3 dB per decade.

5. Reduction of Non-Tectonic Noise

The tectonomagnetic detection threshold can be improved by removing noise resulting from non-tectonic sources. A transfer function technique reduced the noise from 0.11-nT to 0.03-nT rms for more than a month of hourly averages between SCR sites in Colorado separated by 12 km (WARE, 1979). A second technique using Wiener filters removes coherencies between vector field components and total field differences. The Wiener filter reduced the noise of hourly averages from 0.3-nT rms to 0.1-nT rms for baselines under 10 km in California

(DAVIS *et al.*, 1981), and from 0.7-nT rms or less to 0.3-nT rms for baselines ranging from 8 km to 100 km (DAVIS and JOHNSTON, 1983). A third technique filtered magnetic signals at tidal frequencies that were found in the USGS data, reducing the residual noise from 0.66-nT rms to 0.26-nT rms (JOHNSTON *et al.*, 1983).

Noise reduction results have not been reported for periods less than 1 hour. However, the authors are now analyzing simultaneous flux-gate vector component and SCR data recorded in California. Wiener filters will be applied to these 10-second interval data, providing a test of that approach to noise reduction for frequencies less than 1 hour.

6. Transient Events

Improved differential magnetic measurements are possible at short periods using high-frequency, high-sensitivity instruments. For example, several rubidium magnetometers were operated between 1965 and 1967 on the San Andreas fault by BREINER (1967). Although the accuracy of these magnetometers was limited by long-period diurnal drifts of several nT, the measurement precision for periods of several minutes was better than 0.3-nT rms. A number of local magnetic events with amplitudes of 1-nT or less and durations of 1 hour or less, similar in form to the event shown in Fig. 4, were observed simultaneously at two or three sites separated by tens of km but not at other sites in the 5-station network. These events were followed some tens of hours later by local creep events at site 4.5 km away. BREINER and KOVACH (1967) suggested that the magnetic and creep events were both generated by stress changes at depths of about 10 km.

According to MCHUGH and JOHNSTON (1976), deformation measurements show that creep events are primarily near-surface phenomena with dimensions of not more than a few km, and are probably triggered by deeper but slower slip on the fault. JOHNSTON *et al.* (1980) argued that the events detected by Breiner, if real, and if correlated to observed creep events, must have been generated by sources close to such creep events, and would therefore be much larger for nearby sites, considering the inverse-cube attenuation of a magnetic dipole source with distance. A search was carried out for related creep and magnetic events from collocated creep meter and magnetometer pairs 270 m to 9.3 km apart, including sites occupied by Breiner. No transient magnetic events with durations of about an hour, suggested by Breiner to be typical, were seen with a 1-nT detection threshold for the 60 observed creep events that occurred. However, if the magnetic signals detected by Breiner at a distance of 4.5 km from the location of creep events were actually generated by slip occurring at depths of four km or greater, the magnetic signal within 300 m of the creepmeter would be 1 nT or less. Such transient signals would be difficult to detect in typical PM data such as that shown in Fig. 4. For example, the 1-nT event seen in the 0.125-nT PM data in Fig. 5 would not be resolved in the 0.25-nT PM data shown in Fig. 4.

Transient magnetic fields are expected to accompany the post seismic crustal readjustments in aftershock zones. Observations by rapid sampling magnetometers with high resolution in such zones could produce a unique data set. In the aftershock zone of the August 6, 1979, $M=5.7$ Coyote Lake earthquake, DAVIS and SEARLS (1981) deployed 1-nT PM's sampling every minute. The detection threshold was 2-nT peak to peak. JOHNSTON *et al.* (1981) deployed 0.25-nT PM's sampling at 10-minute intervals in the same aftershock zone and reported a 0.25-nT peak to peak detection threshold for daily averages. However, in the data of Johnston *et al.*, the detection threshold for short-term transient events appears to be about 1-nT peak to peak or 0.2-nT rms, which is equal to the unaveraged 0.25-nT PM instrument noise level seen in Fig. 1.

7. Discussion

Using current skills and technology, the observation of magnetic signals that are generated by earthquake processes has proven to be difficult. This is true because such signals appear to be modest, even for large earthquakes, and we do not know well enough when and where large earthquakes will occur. A common approach to this problem has been to operate a number of sensitive magnetometers in areas where earthquakes are expected to occur, in an attempt to obtain magnetic records before, during, and after any earthquakes that happen to occur nearby. Of course, there is a tradeoff between the cost, reliability, and sensitivity of each instrument and the total number of instruments that are deployed.

Proton magnetometers are sufficiently reliable, sensitive, and inexpensive to allow the deployment of tens of PM's in earthquake zones by research groups in several nations. Comparison of the performance of high sensitivity SCR's, and PM's which are used in some of these studies by the U.S. Geological Survey, has shown:

(1) For short baselines the PM's are limited by instrument noise that is several times larger than their least count and digitization noise. In contrast, the SCR's on the same short baseline show a noise level that decreases with increasing frequency by 20 dB per decade until the least count noise limit, 40 dB below the PM's, is approached at periods of less than 30 minutes.

(2) For collocated site pairs with a 13-km separation along the San Andreas fault both the PM's and the SCR's record the same noise power spectrum at periods greater than 4 minutes for 0.125-nT PM's sampling at 15-second intervals. At periods less than 4 minutes the PM's are limited by least count noise. For 0.25-nT PM's sampling every 10 minutes instrument noise limits the resolution for periods less than 50 minutes. In contrast, the SCR data indicate that noise power continues to decrease with increasing frequency until the SCR least count limit is approached at a few tens of seconds.

If noise reduction techniques are not used, not much improvement can be expected in the detection threshold for tectonomagnetic events in active fault zones by decreasing the PM least count sensitivity below 0.25-nT or by decreasing the sampling interval below 10 minutes. However, if noise reduction techniques are used, some improvement is expected. For example, noise was reduced from 0.1-nT rms to 0.03-nT rms for five weeks of hourly SCR averages from a 12-km baseline in Colorado (WARE, 1979). This level is lower than the 0.12-nT rms instrument noise level that has been observed in hourly averages of 0.25-nT PM differences (JOHNSTON *et al.*, 1984). It is not known whether this result is widely applicable, since similar SCR studies have not been carried out in other areas.

At periods less than an hour in seismically active areas improved sensitivity and more rapid sampling could be utilized in attempts to record transient signals associated with both seismic and aseismic fault activity. Preliminary experiments indicate that commercial PM's can be modified to reduce the instrument noise level. The authors operated, on a 50-m baseline, two PM's (Geometrics 856) with circuit modifications that included (1) temperature controlled oscillators, (2) provided for a 0.05-nT least count, and (3) used polarization currents of 3 amps compared to the factory value of 1 amp. The instrument noise level was observed to be less than 0.08-nT rms, a significant improvement over the 0.2-nT rms observed for 0.25-nT PM's reported in this paper and by JOHNSTON *et al.* (1984). If noise reduction techniques are used, particularly in the case of aftershock studies, magnetometers having lower instrument noise levels and increased sampling rates could be effective in reducing the detection threshold for tectonomagnetic events.

REFERENCES

ALLEN, J. H. and P. L. BENDER, Narrow line rubidium magnetometer for high accuracy field measurements, *J. Geomag. Geoelectr.*, **24**, 105–125, 1972.

BREINER, S., The piezomagnetic effect in seismically active areas, Ph. D. Thesis, Stanford University, 1967.

BREINER, S. and R. L. KOVACH, Local geomagnetic events associated with displacements of the San Andreas fault, *Science*, **158**, 116–118, 1967.

DAVIS, P. M., and C. A. SEARLS, Magnetic field measurements in the aftershock region of the Coyote Lake earthquake, *J. Geophys. Res.*, **86**, 927–930, 1981.

DAVIS, P. M. and M. J. S. JOHNSTON, Localized geomagnetic field changes near active faults in California 1974–1980, *J. Geophys. Res.*, **88**, 9452–9460, 1983.

DAVIS, P. M., D. D. JACKSON, and M. J. S. JOHNSTON, Further evidence of localized geomagnetic field changes before the 1974 Thanksgiving Day earthquake, Hollister, California, *Geophys. Res. Lett.*, **7**, 513–516, 1980.

DAVIS, P. M., D. D. JACKSON, C. A. SEARLS, and R. L. MCPHERRON, Detection of tectonomagnetic events using multichannel predictive filtering, *J. Geophys. Res.*, **86**, 1731–1737, 1981.

JOHNSTON, M. J. S., R. J. MUELLER, and V. KELLER, Preseismic and coseismic magnetic field measurements near the Coyote Lake, California, earthquake of August 6, 1979, *J. Geophys. Res.*, **86**, 921–926, 1981.

JOHNSTON, M. J. S., B. E. SMITH, and R. O. BURFORD, Local magnetic field measurements and fault creep observations on the San Andreas fault, *Tectonophysics*, **64**, 47–57, 1980.

JOHNSTON, M. J. S., R. H. WARE, and R. J. MUELLER, Tidal current channeling in the San Andreas fault, California, *Geophys. Res. Lett.*, **10**, 51–54, 1983.

JOHNSTON, M. J. S., R. J. MUELLER, R. H. WARE, and P. M. DAVIS, Precision of geomagnetic field measurements in a tectonically active region, *J. Geomag. Geoelectr.*, **36**, 83–95, 1984.

MCHUGH, S. and M. J. S. JOHNSTON, Short period non-seismic tilt perturbations and their relation to episodic slip on the San Andreas fault. *J. Geophys. Res.*, **81**, 6341–6346, 1976.

MUELLER, R. J., M. J. S. JOHNSTON, B. E. SMITH, and V. KELLER, U.S. Geological Survey magnetometer network and measurement technique in western USA, USGS Open File Report #81-1346, Menlo Park, CA, 1981.

RIKITAKE, T., Y. HONKURA, H. TANAKA, N. OHSHIMAN, Y. SASAI, Y. ISHIKAWA, S. KOYAMA, M. KAWAMURA, and K. OHCHI, Changes in the geomagnetic field associated with earthquakes in the Izu Peninsula, Japan, *J. Geomag. Geoelectr.*, **32**, 721–740, 1980.

SHAPIRO, V. A. and K. N. ABDULLABEKOV, Anomalous variations of the geomagnetic field in East Fergana-magnetic precursor of the Alay earthquake with $m=7.0$ (1978 November 2), *Geophys. J. R. Astron. Soc.*, **68**, 1–5, 1982.

SMITH, B. E. and M. J. S. JOHNSTON, A tectonomagnetic effect observed before a magnitude 5.2 earthquake near Hollister, California, *J. Geophys. Res.*, **81**, 3556–3560, 1976.

WARE, R. H., High-accuracy magnetic field difference measurements and improved noise reduction techniques for use in tectonomagnetic studies, *J. Geophys. Res.*, **84**, 6291–6296, 1979.

WARE, R. H., An improved self-calibrating rubidium magnetometer accurate to 0.01-nT rms, *Rev. Sci. Inst.*, **54**, 1739–1743, 1983.