ELECTROMAGNETIC PRECURSORS TO EARTHQUAKES IN THE ULF BAND: A REVIEW OF OBSERVATIONS AND MECHANISMS

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Abstract. Despite over 2 decades of international and national monitoring of electrical signals with the hope of detecting precursors to earthquakes, the scientific community is no closer to understanding why precursors are observed only in some cases. Laboratory measurements have demonstrated conclusively that self potentials develop owing to fluid flow and that both resistivity and magnetization change when rocks are stressed. However, field experiments have had much less success. Many purported observations of low-frequency electrical precursors are much larger than expectations based on laboratory results. In some cases, no precursors occurred prior to earthquakes, or precursory signals were reported with no corresponding coseismic signals. Nonetheless, the field experiments are in approximate agreement with laboratory measurements. Maximum resistivity changes of a few percent have been observed prior to some earthquakes in China, but the mechanism causing those changes is still unknown. Anomalous electric and magnetic fields associated with fluid flow prior to earthquakes may have been observed. Finally, piezomagnetic signals associated with stress release in earthquakes have been documented in measurements of magnetic fields.

INTRODUCTION

Anomalous electromagnetic (em) fields and changes of electrical properties of the Earth have been reported before earthquakes by many researchers over the years. The observations cover the entire electromagnetic spectrum from visible light to quasi-dc, and viewed collectively, they present a bewildering, and to some improbable, array of cause and effect relationships. A review by *Parrot and Johnston* [1989] concentrated on magnetic fields and radio frequency effects and did not include the many reports dealing with low-frequency to quasi-dc electric phenomena or with the many reports of changes in electrical conductivity in the Earth. In order to address the nature of the phenomena at lower frequencies, an international workshop was held at Lake Arrowhead, California, during June 14–17, 1992. The comprehensive review presented here of proposed mechanisms and observations at ULF (0.01–10 Hz) and lower frequencies began at this workshop.

This review is not intended to be an exhaustive analysis of all of the literature ever published in this field. Instead, we have tried to provide representative examples of all reported phenomena and mechanisms from throughout the world. This paper begins with a summary of field observations and possible explanatory mechanisms observed in the laboratory. These results are then discussed in terms of stress and strain. While the focus of this paper is on precursory phenomena, we will also discuss coseismic signals because these are sometimes observed while monitoring for precursors. Precursory signals are those that occur before an earthquake, while coseismic signals are directly related to the stress or strain release at the time of the earthquake.

SUMMARY OF OBSERVATIONS

Observations can generally be classified into two categories: changes of material properties, and varia-

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tions in signal amplitudes. Changes of resistivity or magnetization are detected with systems measuring natural fields, and resistivity can also be measured by means of imposed fields. Signal amplitudes are monitored with systems designed to measure natural electric and magnetic fields. The electromagnetic spectrum is nearly continuous, and both electric and magnetic fields are always present. However, we will arbitrarily distinguish between anomalous electric fields, magnetic fields, and em fields based on the type of measurement made. Anomalous em fields are detected with systems designed to detect both electric and magnetic fields [*Honkura et al.*, 1976], while anomalous electric fields and magnetic fields are detected with systems designed to measure only one type of field [*Fraser-Smith et al.*, 1990]. We are not suggesting that anomalous magnetic fields are not accompanied by electric fields and vice-versa, but simply that some of the measurement systems do not routinely measure both. As will be shown below, the presence or absence of one or the other anomalous field can be critical in determining the causative mechanism for anomalous variations.

Resistivity Variations

Resistivity is the physical property most often monitored in earthquake prediction studies because changes in response to stress and strain have been demonstrated in the laboratory. Changes of resistivity have been detected with both active measurements and passive telluric and magnetotelluric (MT) arrays. Active measurements typically inject 1–100 A of low-frequency (1 Hz) current with a dipole transmitter and detect the induced potential difference across a receiving dipole. In a homogeneous half-space the resistivity is related to the imposed current I and observed voltage V:

$$\rho = G(V/I), \tag{1}$$

where G is a geometric factor that depends on the configuration of the transmitting and receiving dipoles. The Earth is not homogenous, however, and the response to the imposed current is a complicated combination of geometry and resistivity. Equation (1) is still used in these cases, but the resistivity ρ is replaced by "apparent resistivity" ρ_a. The experiments described below all monitor changes in apparent resistivity, and only a few have attempted to discuss the change in true resistivity. All of these experiments have fixed geometries (and thus fixed geometric factors), correct for fluctuations in current, and therefore monitor changes in observed voltage. Commercial resistivity instruments for exploration typically regulate the current to within 1%; the instruments described below with precisions of much less than 1% are specially built to provide precise current and maintain the apparatus at constant temperature.

Passive arrays are also capable of detecting relative changes of resistivity which are inferred from changes in the telluric or MT transfer functions. In the MT method, fluctuations of the magnetic field induce electric currents in the Earth. The electric fields set up by these currents and the magnetic fields are simultaneously recorded, and a frequency-dependent complex impedance transfer function is computed via spectral analysis:

$$Z(\omega) = \frac{E(\omega)}{H(\omega)}, \tag{2}$$

where ω is angular frequency. The relationship of the impedance Z to resistivity is simple if the Earth is a homogenous half-space:

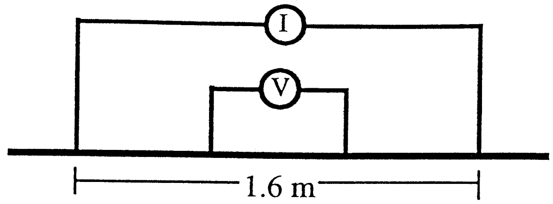
$$\rho = \frac{1}{\omega\mu}|Z(\omega)|^2, \tag{3}$$

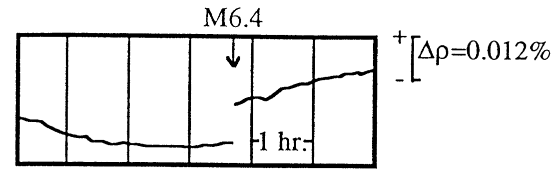
where μ is the magnetic permeability of free space. Again, ρ is replaced by ρ_a in (2) because the Earth is structurally complicated. The apparent resistivity (or phase because the impedance in (2) is now complex) is monitored over time for possible changes. Electric fields at a reference site and at field sites are recorded in the telluric method, and the magnetic fields inducing the currents at all sites are assumed to be the same. Under these assumptions, the relationship between the electric field at the base site (E_B) and that at the field site (E_F) is

$$E_F(\omega) = T(\omega)\ E_B(\omega), \tag{4}$$

where T is the telluric transfer function. Unlike the active resistivity measurements or the MT method, this telluric transfer function cannot be directly related to resistivity.

Changes of resistivity have often been reported in active measurements. *Sadovsky et al.* [1972] and *Barsukov* [1970] reported large (10%) changes of resistivity in the Garm region of the former USSR prior to earthquakes, but measurement errors were not given. *Yamazaki* [1967, 1974] used Wenner arrays with electrode spacings of less than 1 m to monitor resistivity in a cave (Figure 1*a*). Correlations of the daily variations of resistivity with strain measurements in the cave (Figure 1*b*) show that the instrument had a long-term precision of 0.01% [*Yamazaki*, 1968]. This instrument used a highly precise current source and a very stable bridge circuit to achieve this precision. Clear coseismic steps in resistivity of 0.005% due to strain changes were recorded (Figure 1*c*), but *Yamazaki* [1974] concluded that precursory changes were absent. *Mazella and Morrison* [1974] reported changes of 15% prior to a M 4 earthquake along the San Andreas fault, but *Morrison et al.* [1979] later observed no precursory or coseismic changes accompanying an earthquake of similar magnitude and location when resistivity was





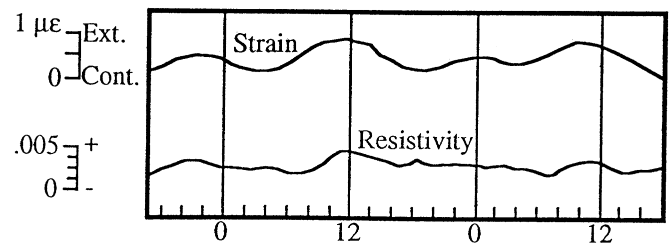


Figure 1. (*a*) Diagram of array used to measure resistivity by *Yamazaki* [1967] in a cave near the ocean in Japan. (*b*) Simultaneous records of resistivity and strain [after *Yamazaki*, 1967] for a 36-hour period. Strain was measured parallel to the array in Figure 1*a*, and both records clearly exhibit effects of tidal loading. The horizontal axis is scaled in hours. (*c*) Coseismic resistivity step observed by *Yamazaki* [1974]. This earthquake was 96 km from the resistivity instrument.

measured with a system accurate to 0.1%. *Fitterman and Madden* [1977] also monitored shallow resistivity using a Schlumberger array with a 200-m-long current dipole adjacent to a creep meter on the San Andreas fault (Figure 2*a*). This instrument was compensated for temperature and also monitored the current continuously. They observed no changes of resistivity larger than 0.005% associated with creep events (Figure 2*b*). In contrast, highly suggestive evidence for resistivity variations prior to earthquakes comes from China. The Chinese have monitored shallow resistivity

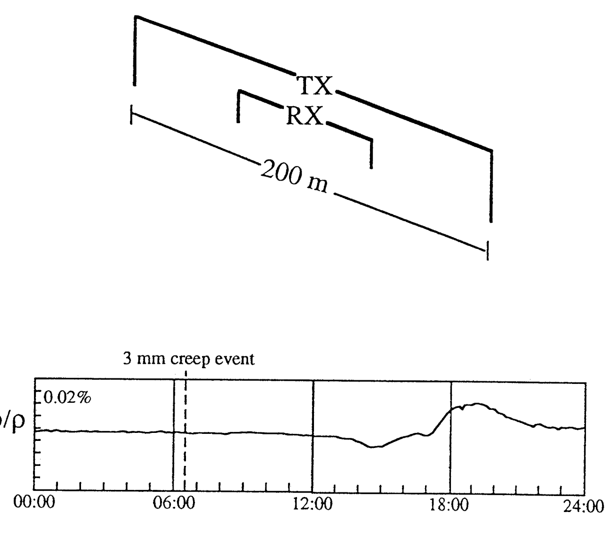


Figure 2. (*a*) Electrodes in a fixed Schlumberger configuration for *Fitterman and Madden*'s [1977] experiment to detect resistivity changes associated with a creep event on the San Andreas fault. (*b*) Percent variations of resistivity recorded over a 24-hour period during which a creep event of 3 mm occurred [after *Fitterman and Madden*, 1977]. Note that individual divisions on the vertical axis are 0.02% and no fluctuation of resistivity was seen at the time of the creep event. The variation centered on 1800 hours was due to a change in the temperature of the instrument.

continuously since 1966 (Figure 3) and observed decreases of several percent beginning approximately 4 years prior to the 1976 *M* 7.8 Tangshan earthquake [*Qian et al.*, 1983; *Zhao et al.*, 1991*b*]. These measurements are made with fixed dipoles in a Schlumberger configuration with a 1000-m-long current dipole. Six to twelve measurements are made daily, and each measurement consists of several readings [*Qian et al.*, 1983]. Currents of 1–5 A are used to generate a signal 100 times larger than the noise level. Precisions of $O(0.1\%)$ are obtained by averaging the readings into monthly means. While these measurements are sensitive only to the shallow crust (upper 500–1000 m), coherent changes in resistivity can be seen over a large region (Figure 4) which also correspond to variations in strain (Figure 5). *Zhao et al.* [1991*a*] attribute these changes to variations in the pore structure of the shallow rocks. Drops in water levels were recorded in wells during this same period of time [*Jin*, 1985], so the resistivity change cannot simply be the result of changing saturation. In fact, a decrease in the water level should lead to an increase in resistivity and not the observed decrease.

Relative resistivity changes can also be inferred from passive measurement of the telluric and magnetotelluric transfer functions. Early attempts by *Reddy et al.* [1976] and *Honkura et al.* [1976] with the MT method showed possible precursors, but *Gamble et al.* [1979] later showed that MT measurements at a single site can be affected severely by local noise. *Gamble et al.* [1979] also showed that accuracies of 1.0% in apparent resistivity measurements could be achieved with the MT method if a remote magnetic field reference was used and a sufficiently long time series was analyzed. No one has reported using MT transfer functions to monitor at these error levels. The telluric method is in use, however. The daily averages of the telluric transfer function (equation (4)) have substan-

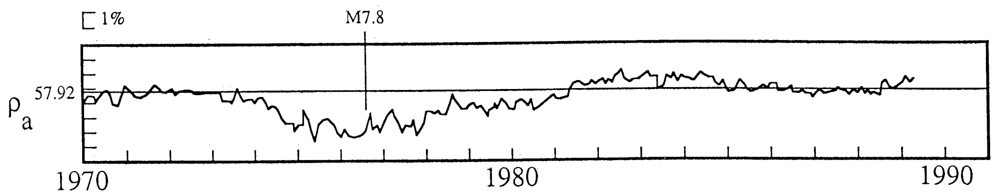


Figure 3. Resistivity variations measured with electrodes in a fixed Schlumberger configuration in China [after *Zhao et al.*, 1991*b*]. Note the gradual decrease several years prior to the 1976 Tangshan earthquake (*M* 7.8) and the gradual increase afterward. These data were taken at a distance of 80 km from the epicenter, and error bars are approximately 0.4% for the monthly means of resistivity.

tial variations [*Park*, 1991]. *Park* [1991] showed that the projections of the daily fluctuations of the transfer function on the eigenvectors for the mean transfer function are stable to within $O(0.1\%)$. *Madden et al.* [1992] used telluric arrays to measure changes in these projections with a precision of 0.1% and reported a long-term decrease of 0.2% yr^{-1} along the Palmdale section of the San Andreas fault zone beginning in 1986. They attributed this decrease to a 50% reduction in the average resistivity of a 1-km-wide zone in the lower crust along the San Andreas fault. No earthquake has yet occurred, so the significance of this observation is unknown. *Park* [1991] observed no significant changes of resistivity in Parkfield using a similar array except for a possible precursor to a *M* 3.7 earthquake there. The projection parallel to the San Andreas fault varied by 7% over a 3-week period ending approximately 1 month prior to the earthquake (Figure 6). *Park* [1991] showed that this fluctuation could have been caused by a 10–17% change in resistivity beneath the array. While this earthquake is the largest one to occur since 1988, when monitoring began, no conclusions can yet be reached about the significance of this possible precursor.

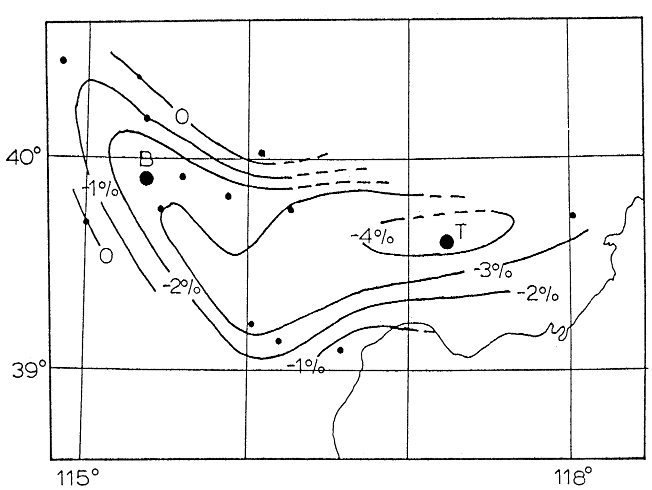


Figure 4. Contour map of resistivity decreases in eastern China prior to the 1976 Tangshan earthquake [after *Qian et al.*, 1983]. Although these data were obtained with shallow soundings, the pattern is coherent over a wide region. Degrees of latitude and longitude are shown (1° of latitude is approximately 110 km). The contours are based partially on the stations identified with solid circles and partially on additional stations not shown. Symbols used are B, Beijing, and T, Tangshan.

Electric Field Fluctuations

Anomalous electric signals have been observed prior to earthquakes in frequency bands from dc to ULF. Hourly or daily means of fields from horizontal, grounded dipoles are usually measured. These means still contain telluric signals with periods from hours to days which are unrelated to earthquakes but can be mistaken for precursory phenomena unless removed. Methods for removal are discussed later. *Myachkin et al.* [1972] reported variations in electric field amplitudes of 100–300 mV km^{-1} which began 3–16 days prior to earthquakes, but not all earthquakes were accompanied by anomalous fields. *Sobolev* [1975] monitored hourly means of electric fields and also reported decreases prior to earthquakes. In Japan, *Miyakoshi* [1985] observed an increase in the electric field strength on just one of two orthogonal dipoles. *Corwin and Morrison* [1977] observed clear decreases of up to 12 mV km^{-1} in the electric field prior to a *M* 2.4 earthquake located 2.5 km from the array (Figure

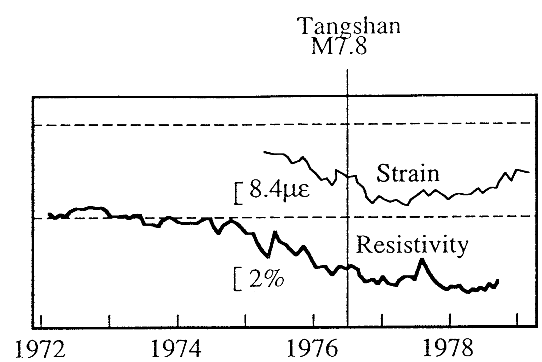


Figure 5. Simultaneous records of resistivity and strain in China [after *Zhao et al.*, 1991*a*]. Strain was measured on a 120-m baseline at a site 30 km from the resistivity array. Both the array and the baseline were over 150 km from the epicenter of the Tangshan earthquake. The resistivity decrease was accompanied by a shortening of the base line, and the amplification factor $(\Delta\rho/\rho)/\epsilon$ at this site was 1800.

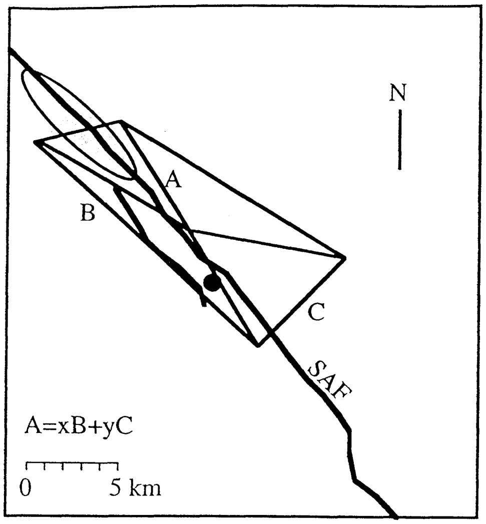
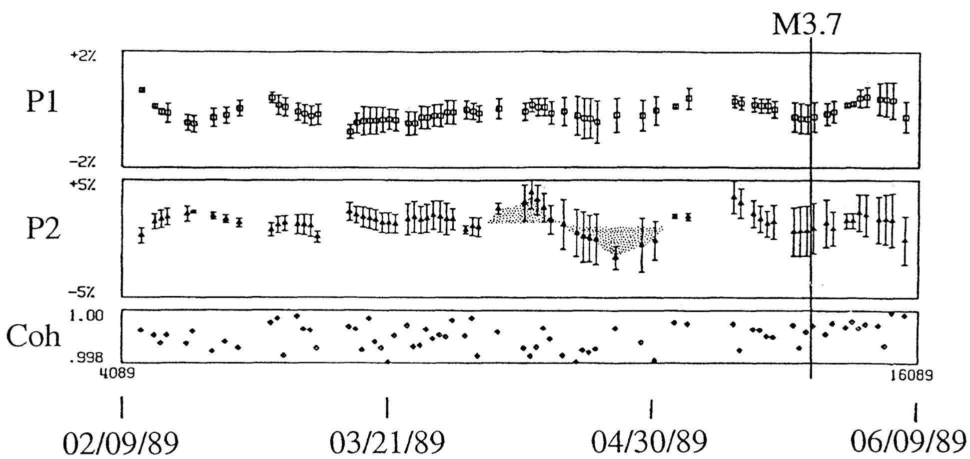


Figure 6. Telluric array in Parkfield [after Park [1991]). Dipoles B and C are references for the array, and the fluctuation shown in Figure 6*b* was seen on dipole A. Dipole A is represented as a linear combination of dipoles B and C with telluric coefficients x and y, and analysis is done for periods between 300 and 7200 s in order to eliminate effects from the frequency dependence of the telluric transfer function. Segments of the San Andreas fault (SAF) are denoted with heavy black lines, and the epicenter of the M 3.7 earthquake is shown with the dot. Modeling by *Park and Fitterman* [1990] showed that changes of resistivity in the shaded region could have generated the fluctuation in Figure 6*b*. (*b*) Projections of the daily fluctuations of the telluric transfer coefficients (x and y in Figure 6*a*) on average eigenvectors perpendicular (P1) and parallel (P2) to the San Andreas fault. Error bars are obtained from a weighted standard deviation computed for a sliding 9-day averaging window [*Park*, 1991]. The fluctuation is shaded and occurred approximately 1 month before the earthquake. The coherence between the observed signal on dipole A and that predicted from $A_{\text{pred}} = xB + yC$ is also shown.



7). These fields were measured with nonpolarizing Cu-$CuSO_4$ electrodes at the ends of a 300-m-long dipole. They attributed this variation to fluid flow into a dilatant region prior to the earthquake.

The variations discussed above have been observed over periods of hours to days; attempts to monitor electric fields at higher frequencies have had less success. *Honkura et al.* [1976] observed daily variations of over 50% in electric field amplitude at periods of 60–7200 s during a year of monitoring when no moderate or large earthquakes occurred. *Sims and Bostick* [1969] observed natural variations of over 100% in both the electric and magnetic fields at mid-latitudes in Texas, so we conclude that changing source fields are capable of generating substantial variations that are unrelated to earthquakes.

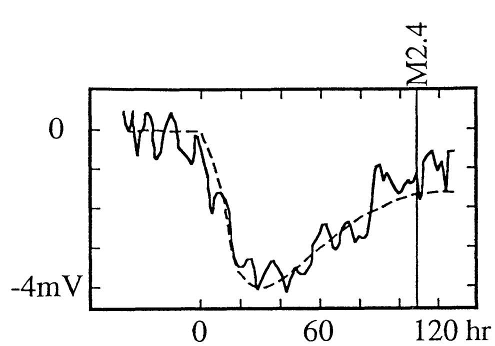


Figure 7. Variation of electric field prior to a M 2.4 earthquake. The solid line is the signal, and the dashed line is the signal predicted with a model of fluid flow into a spherical dilatant zone [*Corwin and Morrison*, 1977].

One of the better documented experiments in which variations of anomalous electric signals are used for earthquake predictions is in Greece. *Varotsos and Alexopoulos* [1984*a*, *b*, 1987] and *Varotsos and Lazaridou* [1991] observed pulses (similar to square waves) in electric field strength of up to 250 mV km^{-1} which preceded earthquakes in certain regions by 7–260 hours (Figure 8), and they report some success in using these seismic electric signals (SES) to predict earthquakes. Multiple temporary dipoles are set up at electrically quiet sites and monitored for SES. The permanent dipoles are those that have exhibited SES prior to earthquakes and are deemed sensitive sites. Future predictions are based on the repeated occurrence of SES at those sites. Scaling of signals on dipoles of unequal length are used to distinguish SES

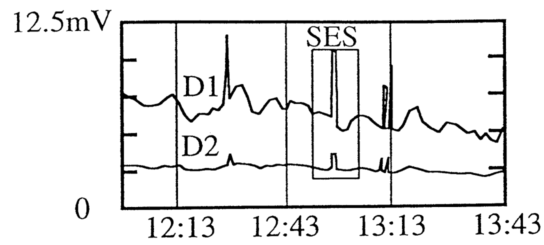


Figure 8. Example of SES observed on two parallel dipoles in Greece [after *Varotsos and Lazaridou*, 1991]. D1 is 181 m long, while D2 is only 47.5 m long. Note how the signal (within labeled box) scales with dipole length. Record length is approximately 105 min.

from noise or local sources (Figure 8). The relationship between the SES at sensitive sites and earthquakes in specific regions thus appears to be well defined and repeatable. *Varotsos and Alexopoulos* [1986] attribute these steps to alignment of polarized crystal defects at the hypocentral region due to increased stress (discussed in the next section). Similar attempts in Japan by *Uyeda et al.* [1992] have not yet met with success because of high noise levels at the measurement sites or because the signals are not present.

Other attempts have recently been made to monitor the vertical electric field (E_z) either in the atmosphere or in the Earth. *Battalino* [1992] monitors E_z in the atmosphere with field mills and may have seen earthquake-related signals at ULF and lower frequencies. *Fujinawa et al.* [1992] show the appearance of signal spikes in the ULF band prior to earthquakes related to a volcanic eruption but are cautious about claiming that these are precursory phenomena. Spikes are observed in the voltage between a long, buried steel pipe and a ground plane at the surface. This sensor rejects signals from the atmosphere and detects variations of E_z in the Earth.

Magnetic Field Variations

Variations of magnetic field amplitude prior to earthquakes have been cited by many authors (see *Johnston et al.* [1973] for review), but the amplitudes of reported anomalies decreased rapidly after the mid-1960s (Figure 9), when high-quality instrumentation became available and spurious signals from ionospheric and magnetospheric disturbances were removed from seismomagnetic signals by referencing data to nearby instruments outside the seismogenic region [*Johnston et al.*, 1984]. Most modern experiments use proton recession magnetometers to monitor total field and difference signals between magnetometers to reject natural signals [*Johnston et al.*, 1984]. Not all of the reported anomalies are attributable to noise or lack of referencing, however. *Mizutani and Ishido* [1976] correlated changes of 5–10 nT with an earthquake swarm and attributed the magnetic field change to fluid flow. *Shapiro and Abdullabekov* [1982] issued a warning for a M 7.0 earthquake on the basis of a 23-nT anomaly 3 days prior to the earthquake which was based on differencing fields from two observatories. *Johnston and Mueller* [1987] and *Mueller and Johnston* [1990] reported dc offsets of the total magnetic field of about 1 nT which were associated with both the M 5.9 North Palm Springs and M 7.1 Loma Prieta earthquakes (Figure 10*a*). They concluded that this was a piezomagnetic effect due to stress removal (Figure 10*b*). Most of the magnetic field anomalies mentioned above are quasi-dc with periods of days or longer. Stable measurements of magnetic field strength at frequencies above quasi-dc are plagued by fluctuations of the natural field which have a $1/f$ type spectrum at frequencies below 0.1 Hz [*Honkura et al.*, 1976]. However, *Fraser-Smith et al.* [1990] observed that signal levels in the ULF band (0.01–10 Hz) increased up to 1 month prior to the Loma Prieta earthquake on an instrument near the epicenter. No such increase was observed at a site distant from the epicenter, although that site was monitoring a slightly different frequency band. Only one component of the magnetic field is monitored with an induction coil, but signals of 0.43 nT at 0.1 Hz were seen approximately 1 month prior to the earthquake. Wideband noise levels increased steadily prior to the earthquake with maximum activity of 60 nT $Hz^{-1/2}$ in the 0.01- to 0.5-Hz band 3 hours before the earthquake. *Draganov et al.* [1991] suggested that these changes are due to magnetic fields induced by fluid flow at seismogenic depths.

Electromagnetic Field Variations

Numerous high-frequency em emissions have been reported before earthquakes [*Gokhberg et al.*, 1982; *Yoshino et al.*, 1985], but these lie above the ULF and ELF (0.01–3 kHz) bands. Within these bands, few anomalous signals have been reported in association with earthquakes. Part of the problem may be that the waveguide in the atmosphere between the Earth and the ionosphere attenuates waves with frequencies below 1.0 Hz and measurements at these frequencies

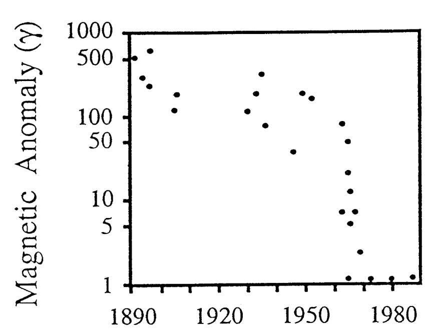


Figure 9. Reported tectonomagnetic anomalies as a function of time [after *Johnston et al.*, 1973]. The magnitude of the reported anomaly decreased abruptly as more accurate instrumentation became available in the 1960s and referencing techniques were used to cancel magnetospheric and ionospheric signals.

must therefore be taken close to the source. Fields with frequencies above 1.0 Hz have been measured both from satellite and on the ground. *Serebryakova et al.* [1992] observed anomalies below 1 kHz in satellite orbits over the epicenter of the 1989 Armenian earthquake, and these anomalies were stronger at the lower frequencies. *Matthews and Lebreton* [1985] and *Henderson et al.* [1991] found no correlation of ULF fields measured with a satellite and earthquakes, however. Henderson et al.'s study was particularly thorough, with comparisons of 62 orbits passing over recent or imminent earthquakes with another 62 orbits over seismically quiet regions. *Dea et al.* [1992] report ULF signals at a ground observatory which they correlate to 29 M >3.5 earthquakes located throughout California and Nevada in an 18-month period; *Van Bise et al.* [1992] have observed similar signals. However, Dea et al. detected signals for only 7 of the 67 M >3.5 earthquakes within southern California during that same 18-month period.

Geophysical Constraints

The purported precursors discussed above are attributed to changing stress or strain prior to earthquakes or perhaps to fluid flow triggered by mechanical changes. Therefore it is relevant to review observations of stress, strain, and water level in regions that are prone to earthquakes. High-quality regional strain measurements in seismically active regions have now been available for more than 20 years in both California and Japan [*Savage*, 1983]. The dominant characteristic of the data obtained in both the transform margin in California and the convergent margin in Japan is the uniformity of the strain rate with time. The mean strain accumulation rate in California is 0.3 $\mu\epsilon$ (microstrain) yr^{-1} [*Savage*, 1983; *Savage et al.*, 1986], and similar results are obtained in Japan. The state of strain in the intermediate period (months to seconds) prior to the earthquake might be lost in the regional strain measurements because of the infrequent sampling and poor sensitivity of these data (1 $\mu\epsilon$) compared with the regional strain rate given above. Fortunately, arrays of high-quality borehole tensor strain meters have recently been installed in California and Japan that provide continuous strain measurements at sensitivities of 0.0001 ppm (0.1 nϵ). Data from these instruments provide a very detailed view of the epicentral region prior to large earthquakes. While variations in the near-field strain immediately prior to earthquakes may occur as a result of nonlinear fault failure processes [*Johnston et al.*, 1987] and minor movement of pore fluids, the near-surface strain changes associated with these variations appear to be at or below the nanostrain level. No precursory strain changes above the noise level of 1 nϵ were seen on instruments at a distance of approximately 40 km from the 1989 M 7.1 Loma Prieta earthquake [*Johnston et al.*, 1990]. In other words, changes in

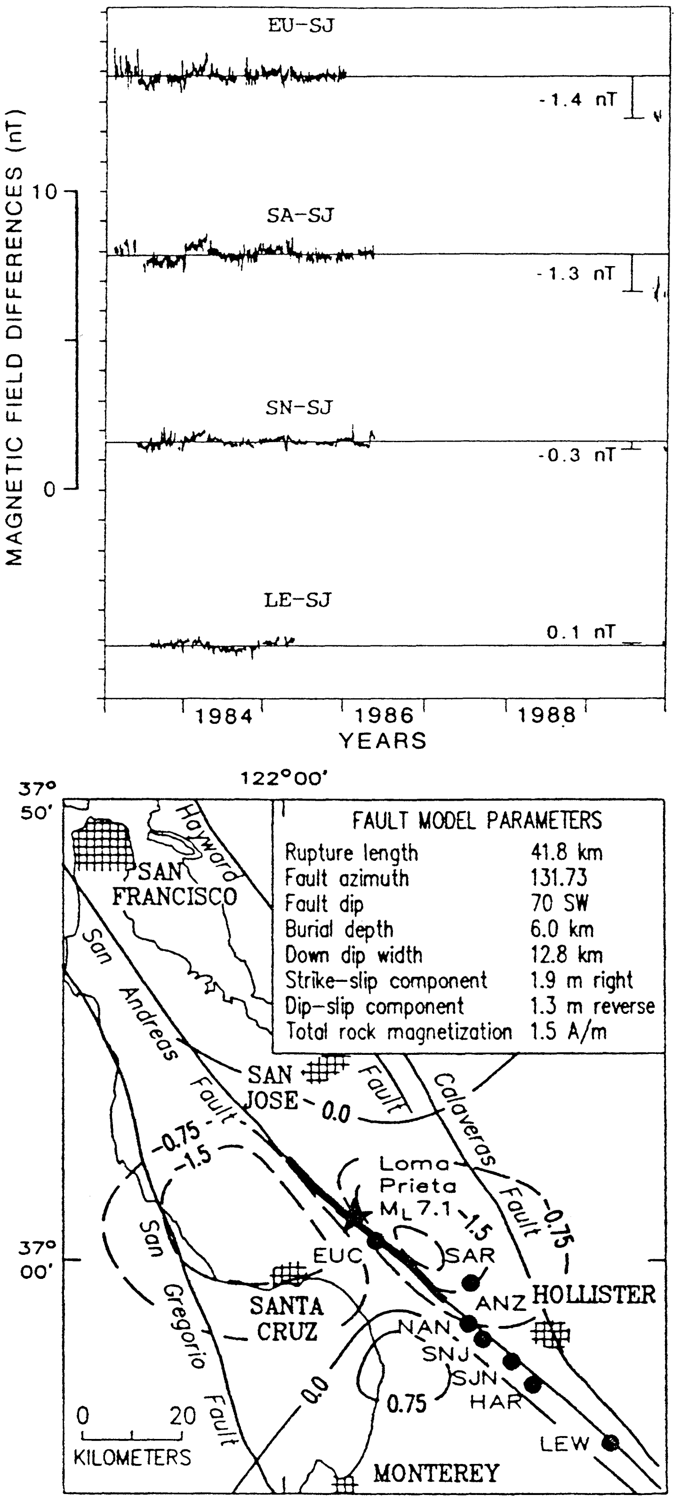


Figure 10. (*a*) Records of magnetic field at sites along the San Andreas fault near Loma Prieta [*Mueller and Johnston*, 1990]. While data are missing from mid-1985 to late 1989, the offsets are clear owing to the stability of the data. Fields at stations EUC (EU-SJ), SAR (SA-SJ), SNJ (SN-SJ), and LEW (LE-SJ) are referenced to station SJN (see Figure 10*b* for station locations). (*b*) Changes of magnetic field predicted by *Mueller and Johnston* [1990] due to release of stress in the 1989 Loma Prieta earthquake. The piezomagnetic mechanism explains both the magnitude and the distribution of magnetic field changes in Figure 10*a*. Model and fault parameters for the dislocation model are shown in the upper right corner.

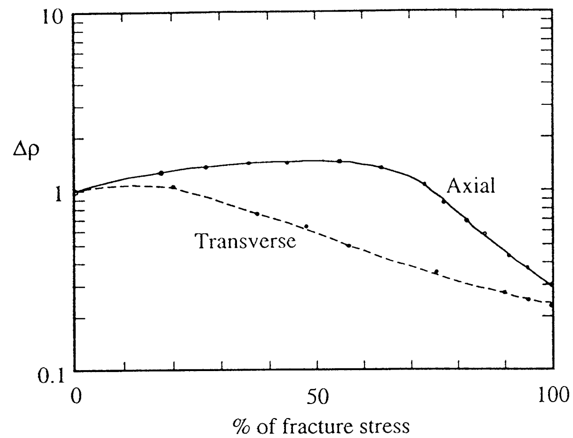


Figure 11. Normalized stress-resistivity curves for granite measured parallel (axial) and perpendicular (transverse) to the applied stress [after *Brace and Orange*, 1968*a*]. Note how the resistivity drops as the percent of fracture stress increases to 100%, and substantial changes are seen in the transverse direction at moderate values of fracture stress.

regional strain and stress in the near field of, and prior to, moderate earthquakes are apparently below 1–10 nε and one μbar, respectively. Additionally, observations of strain fields of earthquakes can be fit with simple dislocation models which predict several hundred nanostrains within 50 km of moderate earthquakes [*Linde and Johnston*, 1989]. If the observed precursors are linearly proportional to stress or strain, then coseismic changes should be larger than precursory ones; this is not usually observed, however.

Water levels are also routinely monitored in seismogenic regions. *Roeloffs et al.* [1989] have shown that changes in water levels are associated with creep events along the San Andreas fault. Maximum changes of 16 cm were observed, although most fluctuations were less than 5 cm. Similarly, water level changes prior to the 1976 Tangshan earthquake were in the range of 5–10 cm [*Jin*, 1985]. The strain and water level measurements provide constraints for possible mechanisms causing both resistivity changes and variations of electric and magnetic fields. In the upper crust, rock resistivity is dependent primarily on the mobility of ions and the distribution of the fluid within a rock. Deformation of the pore space, which would affect the distribution of the fluid, can be constrained by strain measurements. Variations of saturation, and thus resistivity, may be reflected in the water level fluctuations. Transient changes of water level may also generate self-potentials.

SUMMARY OF LABORATORY MEASUREMENTS AND MECHANISMS

Scientists have tried to explain all of the observations reported above with mechanisms which have produced signals in laboratory experiments. The existing laboratory data base in the areas of resistivity, electrokinetic phenomena, piezoelectricity, piezomagnetism, and solid-state defect mechanisms is briefly summarized here. Only resistivity and magnetization have shown definite precursory behavior in the laboratory, although other mechanisms such as electrokinetic phenomena and solid-state changes cannot presently be ruled out.

Resistivity

The electrical resistivity of a rock in the crust is largely controlled by its porosity and pore fluid resistivity. *Archie* [1942] gave an approximate empirical expression for this behavior as

$$\rho_R \approx \rho_F \phi^{-m} \tag{5}$$

where ρ_R and ρ_F are the resistivities of the rock and pore fluid, respectively, ϕ is the porosity, and m is an index between 1.5 and 2.0 for most rocks. When the rock is partially saturated, (5) is usually modified to

$$\rho_R \approx \rho_F \phi^{-m} S^{-n}, \tag{6}$$

where S is the fractional saturation and n, the saturation index, is approximately 2.

In the dilatancy models of the earthquake source [*Anderson and Whitcomb*, 1973; *Hanks*, 1974; *Nur*, 1974] a rock begins to create new cracks or pore space prior to fracture as the effective stress is increased. The effective stress is the difference between the confining or lithostatic pressure and the pore pressure. Increasing effective stress for fully saturated, drained rocks reduces the porosity and, as (5) shows, increases the resistivity [*Brace et al.*, 1965]. As the failure stress is approached and dilatancy begins, the porosity increases, and the resistivity accordingly decreases (Figure 11) [*Brace and Orange*, 1968*a*]. For partially saturated rocks the behavior is generally the same except that at low stress levels, the resistivity decreases as the volume of the nonconducting (air) phase is reduced [*Brace and Orange*, 1968*b*]. This decrease in resistivity for an increase in stress may be the explanation for the data in Figure 5. A decrease in the shallow resistivity in China was accompanied by compression indicated by shortening of a baseline. If the shallow rocks were partially saturated, then such a decrease would be expected in light of the results of *Brace and Orange* [1968*b*].

Equation (5) predicts that the fractional change in rock resistivity $\Delta\rho/\rho$ is related to the volumetric strain ($\epsilon = \Delta V/V$) by

$$\frac{\Delta\rho}{\rho} = -\frac{m}{\phi}\epsilon, \tag{7}$$

or that the ratio of the relative change in resistivity ($\Delta\rho/\rho$) to the relative change in volume (ϵ) is given by

$$\left(\frac{\Delta\rho}{\rho}\right)\bigg/\epsilon = -\frac{m}{\phi}; \tag{8}$$

$|(\Delta\rho/\rho)/\epsilon|$ is usually termed the amplification factor because it indicates the factor by which the strain is "amplified" in a resistivity measurement.

At confining pressures characteristic of those near the Earth's surface, "flat" cracks are easily closed, and large decreases in pore space occur for small increases in confining pressure. All rock properties that depend on the pore space thus show nonlinear behavior at small strains. Amplification factors (Figure 12) of up to 500 were noted for strains down to 10^{-4} [*Yamazaki*, 1965], but factors were as high as 10^4 for smaller strains in partially saturated tuffs [*Morrow and Brace*, 1981]. *Zhao et al.* [1991*a*] have computed amplification factors of over 10^3 from field observations of shallow resistivity and strain.

Equation (8) predicts that the resistivity should increase as the porosity decreases in response to increasing confining pressure. The opposite behavior is frequently seen in field measurements of resistivity, however, and *Fitterman* [1976] has shown that such behavior can occur if a rock is subjected to shear stress. *Brace and Orange* [1968*a*] have observed decreases in resistivity at shear stresses as small as 15% of the failure stress which are due to opening of cracks in the σ_3 direction (Figure 11, dashed curve).

Lockner and Byerlee [1986] have shown that a rock possesses both real and imaginary parts to its frequency-dependent conductivity and that the imaginary component is sensitive to the onset of dilatancy. The imaginary component of the resistivity of rocks characteristic of fault zones was less than 2.5% of the real component of the resistivity at all frequencies below 1 kHz [*Lockner and Byerlee*, 1986]. While the imaginary component of the resistivity may be sensitive to the onset of dilatancy, monitoring this component in the field will be even more difficult than is recording the real component. *Lockner and Byerlee* [1985] propose that induced polarization at low frequencies (i.e., the imaginary component) could permit sufficiently slow decay of charge after an earthquake that phenomena such as earthquake lights could result. Residual charge after a few seconds is only 0.5–1% of the initial charge, so large initial charge generation would be necessary for a detectable signal. Necessary initial conditions could be achieved for coseismic phenomena but are unlikely for precursors.

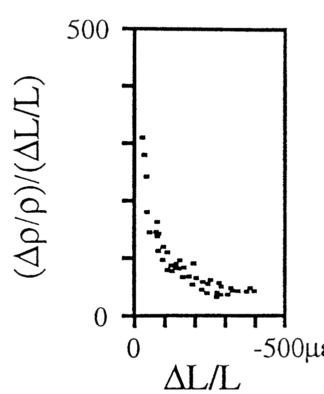


Figure 12. Laboratory experimental data showing variation of strain amplification factor $(\Delta\rho/\rho)/\epsilon$ with linear strain $\Delta L/L$. Measurements were made on a specimen of lapilli tuff with 18% water content [*Yamazaki*, 1965].

Most of the laboratory experiments reviewed here were run over a period of hours and do not simulate the lower strain and stress rates typical of the Earth. Very few experiments have been run over time scales of months or years. Rocks under constant stress for long periods exhibit pressure dissolution or precipitation and stress corrosion or healing [*Weyl*, 1959; *Scholz*, 1972]. While the resistivity versus time as pressure solution and stress corrosion progressed has not been measured, the expected result for a drained rock is that porosity should decrease for high confining stresses unless the rock becomes dilatant. According to equation (5), this will result in an increased resistivity. In contrast, T. Madden (personal communication, 1986) has observed an annual decrease of 3% in resistivity in a sample of Westerly granite held at a constant uniaxial stress of 400 bars. This experiment is preliminary, however, and no general conclusions can as yet be drawn. More experiments of this nature are needed to help us understand the behavior of rocks in the Earth.

Electrokinetic Phenomena

Natural electric potentials in the Earth are caused by subsurface thermal or concentration gradients and by the flow of pore fluids. A likely candidate for the source of precursory self-potential anomalies is the streaming potential mechanism [*Morgan et al.*, 1989]. Streaming (electric) potentials are generated across rocks through which electrolytic fluids are moving [*Johnson*, 1983; *Ishido and Mizutani*, 1981; *Morgan et al.*, 1989]. The cross-coupling coefficient, which is the voltage generated per unit driving pressure difference, is dependent on the mineral type, the electrolyte type and concentration, and the temperature of the mineral-solution interface. Cross-coupling properties are also dependent on flow rates if the flow becomes turbulent; laminar or fully established flow is assumed in the case of precursory phenomena because of the small flow rates. To first order, the cross coupling is independent of permeability. There is actually an apparent dependence, but this is due to varying surface conductivity effects and not directly to permeability. For rocks at room temperature a good average value of the cross-coupling coefficient for 1:1 electrolytes such as NaCl is -4 mV (atm Ω^{-1} m^{-1})$^{-1}$ and -2mV (atm Ω^{-1} m^{-1})$^{-1}$ for 2:1 electrolytes such as $CaCl_2$.

The effects of temperature are complicated and not well understood mainly because of a lack of experimental data [*Morgan et al.*, 1989]. Existing data over limited temperatures below 100°C are contradictory and indicate that the cross-coupling coefficients both increase and decrease [*Morgan et al.*, 1989; *Ishido and Mizutani*, 1981]. The major problem seems to be associated with the establishment of equilibrium conditions at elevated temperatures. Ishido and Mizutani's experiments are the closest to these equilibrium conditions, and they show appreciable increases in the magnitude of the cross-coupling coefficient with temperature. Additional measurements at pressures and temperatures representative of in situ conditions are needed.

Morgan et al. [1989] and *Morgan and Nur* [1986] point to a new mechanism that may have significant relevance to electrical precursory signals such as earthquake lightning and self-potential anomalies. Two-phase fluid flow enhances cross-coupling properties. The enhancement comes mainly from the increased resistivity at partial saturation and can account for a factor of 2 or more in the cross-coupling coefficient. Partial saturation may result at the source region either by dilatancy [*Brace*, 1975] or boiling [*Lockner et al.*, 1983]. However, boiling is unlikely except as a coseismic phenomenon.

Piezoelectricity

A quartz crystal stressed in the appropriate direction will produce a voltage. This well-established physical phenomenon has been observed for rocks in the laboratory. The effect can occur in the Earth for regions over which there is some alignment or long-range ordering of quartz grains [*Finkelstein et al.*, 1973; *Dmowska*, 1977; *Baird and Kennan*, 1985], but self-cancellation precludes the development of large potentials [*Tuck et al.*, 1977]. Piezoelectric signals from quartz-bearing rocks are less than 0.1% of those observed for single crystals of quartz [*Tuck et al.*, 1977], even if the crystals are aligned as in a quartzite. Additionally, the generated electric field is stress rate dependent for rocks. *Dmowska* [1977] discusses the correlation of the laboratory effect and its relation to precursory and coseismic effects. The conclusion is that the stress rates can be high enough only during the earthquake and may therefore contribute to coseismic, but not to precursory, electric phenomena.

Piezomagnetism

Much experimental work has been carried out on the stress sensitivities of both the magnetic susceptibility and remanent magnetization [*Nagata*, 1970*a*, *b*; *Martin and Wyss*, 1975; *Wyss and Martin*, 1977]. The general result is that rocks change their magnetization as a function of stress. Changes of the order of 2×10^{-4} bar^{-1} are expected for igneous rocks [*Stacey and Johnston*, 1972]. Typical magnetization for rocks are 10^{-2} to 10^{-3} emu cm^{-3}. Stress levels of approximately 10 bars will result in magnetic fields of a few nanoteslas, which are consistent with the field observations discussed earlier. Although magnetization effects decrease with temperature [*Pozzi*, 1977], the overall effect of high pressure and temperature in the Earth is unclear [*Carmichael*, 1977]. Grain size also determines the degree of magnetization [*Kean et al.*, 1976]. *Martin et al.*, [1978] showed that magnetic effects decreased as dilatancy progressed. Therefore magnetization is another physical property for which precursors to rock failure have been demonstrated in the laboratory.

Solid State Defect Mechanisms

A different phenomenon has been described by *Dologlou-Revelioti and Varotsos* [1986] for the generation of electric currents in rocks. Rocks that are thermally stimulated by first cooling them to −80°C and then heating to 100°C to produce an internal pyroelectric field develop transient electric currents of 10^{-9} to 10^{-11} A in the range 6° to 23°C. The suggested mechanism involves the polarization and depolarization of impurity or defect dipole complexes, and the relaxation times for these dipoles are sufficiently short near room temperature to generate currents upon relaxation. *Dogloglou-Revelioti and Varotsos* [1986] propose that these necessary internal electric fields could alternately be generated by application of stress to piezoelectric materials in rocks and that changes of stress prior to earthquakes could result in transient currents. Varotsos and coworkers are presently attempting to obtain similar current effects by pressure stimulation and at more realistic in situ conditions of temperature. *Freund et al.* [1992] suggest the solid-state generation and hole-type transport of O^- in dry minerals. The effects seem well founded in solid-state physics, but further experiments with wet rocks need to be attempted before this can be viewed as a cause of electrical signals in the Earth.

DISCUSSION

The observations summarized in the first section provide a representative cross section of the anomalies reported as earthquake precursors. We now discuss these anomalies in terms of the probable causative mechanisms and how these mechanisms relate to other geophysical constraints. Combination of the stress and strain data with the mechanisms deduced from laboratory measurements leads to the conclusions that coseismic signals should be larger than precursory ones if the deformation and mechanisms are linearly related and that precursory phenomena will probably be detected only with very sensitive instruments. The possibility exists that deformation is nonlinearly related to changes in material properties, however; the amplification factors ($|(\Delta\rho/\rho)/\epsilon|$) discussed above are an example of nonlinear behavior. Regardless, we reiterate that precursor experiments need very sensitive instruments.

Resistivity Changes

Mechanisms for perturbing resistivity in response to changes of stress and strain are well known and involve changing the distribution of the fluid within a rock by deformation of pore space or by partial desaturation. Resistivity can reasonably be expected to vary from a few tenths of a percent to several percent if the stress and strain sensitivities cited above are combined with typical coseismic stress drops of 10–100 bars and strain changes of a few microstrains. *Fitterman* [1976] concluded that at least 20 years of stress accumulation would be necessary to produce a change in resistivity of 1% along the San Andreas fault. *Yamazaki* [1967, 1968, 1974] observed coseismic resistivity steps of $O(0.01\%)$ associated with distant earthquakes. Field observations [*Madden et al.*, 1992; *Qian et al.*, 1983] agree approximately with laboratory measurements; maximum changes of a few percent are reported. Thus only experiments with precisions of $O(0.1\%)$ or better can reasonably be expected to yield results. *Qian et al.* [1983] achieve these accuracies by averaging multiple estimates of apparent resistivity obtained with an active source. *Morrison et al.* [1977] used a deep active sounding technique with this precision. *Yamazaki* [1967, 1968] and *Chevalier and Morgan* [1992] have instruments that utilize a current source and are precise to within $O(0.001\%)$ but sample only a few cubic meters at the surface. *Madden et al.* [1992] and *Park* [1991] have achieved this precision with passive telluric arrays by considerable averaging of the telluric transfer functions.

Most of these accurate measurements are designed to detect local resistivity changes only and are usually focused on the earthquake focal zone; *Morrison et al.* [1977] and *Park and Fitterman* [1990] both showed that the effect of the perturbed resistivity rapidly decreased away from the perturbed zone, as is illustrated in Figure 13. *Qian et al.* [1983] have a sufficiently accurate active system, but they are focused on local changes that occur at distances of 100–200 km from the earthquake. None of the mechanisms discussed here are capable of producing the observed local changes in response to stress or strain in a distant seismogenic zone, however. Only one of the observations consisted of distant sensing of resistivity changes. *Madden et al.* [1992] infer from models of the observed decrease of 0.2% yr^{-1} in the telluric transfer function that their telluric array was detecting a change along the San Andreas fault with dipoles approximately 10 km from the fault. While this model is nonunique, it does show that perturbation of the resistivity along the fault can change the distribution of electric fields away from the fault.

Electric Field Changes

Most of the variations in electric field described above are at such low frequencies that induction is an unlikely cause for the anomalous signals. Possible causes include electrokinetic effects and solid-state mechanisms. For earthquake processes, the most likely candidate for the source of precursory anomalies is the streaming potential mechanism. Streaming potentials can be created by fluid flow resulting from changing stress or strain and can produce local anomalies near the seismogenic zone or local anomalies that occur in response to a strain field from a distant seismogenic zone. Enhancement of streaming potentials by two-phase fluid flow [*Morgan et al.*, 1989] may be important in the generation of self-potential anomalies. Several experiments have probably directly detected streaming potentials in the seismogenic zone [*Corwin and Morrison*, 1977]. However, distant sensing of these signals is difficult because of the rapid decrease of the dc current from a discrete source and because of rapid attenuation of an em field component in a generally conductive Earth (Figure 14) [*Honkura*, 1992]. Anomalies measured at distances greater than 10 km from earthquake hypocenters may instead be due to local responses to a strain field from a distant seismogenic zone [*Varotsos and Alexopoulos*, 1984*a*]; *Yamazaki* [1974] demonstrated that such a strain field was capable of producing measurable electric signals, but the signals are very small. However, a confined aquifer could act as a natural amplifier for such strains and generate signals as large as those reported by *Myachkin et al.* [1972], for example.

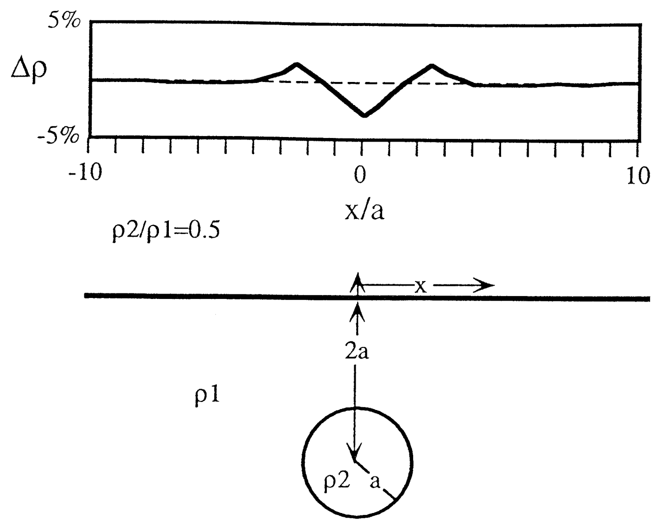


Figure 13. Changes of apparent resistivity observed with Schlumberger configuration centered over a spherical target. The target is buried twice as deep as its radius, and the resistivity contract is 0.5. Current electrodes are located far from the sphere. Note how the perturbation in apparent resistivity is negligible at horizontal distances beyond 4 times the sphere's radius. Calculations are based on formulae from *Telford et al.* [1990].

Dologlou-Revelioti and Varotsos [1986] describe a solid-state mechanism in which electric dipoles formed from defects in crystal lattices are aligned at the seismogenic zone by increasing the stress. A net current

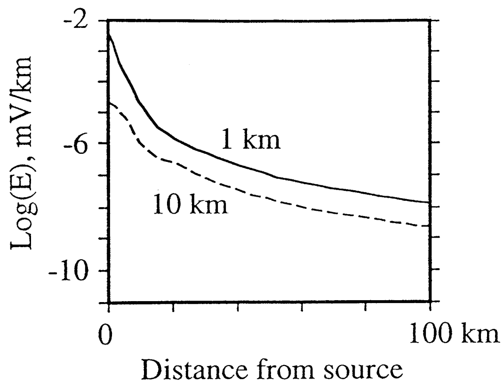


Figure 14. Decay of electric field at 1 Hz away from dipolar sources buried at 1 km and at 10 km [after *Honkura*, 1992]. Dipole strength is 1 A-m and the resistivity of the Earth is 100 ohm-m. *Honkura* [1992] concludes that a minimum of 10^5 A-m is needed to produce measurable ULF signals.

generated by this alignment creates a source of electric field which is detected at distances of up to 100 km at sensitive sites. While this mechanism has been shown to work under laboratory conditions, it has not been extended to conditions more typical of seismogenic zones and to rocks with realistic crustal conductivities. *Bernard* [1992] argues that it is unlikely that specialized current channeling conditions exist that could produce electric fields of millivolts per kilometer due to a distant current source and instead proposes that the SES are the result of unstable fluid reservoirs at the measurement site responding nonlinearly to precursory strain changes. A local source would eliminate the problem of transmission of a diffusive electric signal from a distant source without negating the validity of the SES. In any case, measurements of local strain and water levels at sensitive sites could help identify the mechanism causing SES.

Piezoelectricity is also invoked as an explanation of electric field fluctuations. Rocks containing quartz emit em signals when stressed, but *Dmowska* [1977] concludes that these phenomena are measurable only with coseismic stress changes because only then are the stress rates high enough. Additionally, laboratory measurements are usually made on dry rocks and not on rocks with the usual upper crustal conductivities of 0.001–1 S m^{-1}; the more conductive rocks will probably damp out any piezoelectric signals. *Brady* [1992] attributed the lack of em emissions upon fracturing pyrite to the inability of a conductive mineral to maintain a charge separation, for example.

Observed changes in the vertical electric field in the atmosphere may be due to surface charges or to changes in the conductivity of the atmosphere. Generation of ions at the surface by earthquake-related mechanisms is problematic; the mechanism must not only generate ions but also decrease the conductivity of the Earth to prevent recombination of the ions. *Lockner et al.* [1983] propose a coseismic mechanism involving vaporization of water by frictional heating, which produces ions and lower conductivities, but the heating involved is much too large for precursory changes. *Freund et al.* [1992] suggest that O^- charge carriers released at seismogenic depths could migrate to the surface, but these mobile ions are likely to combine with ions already present in fluids which cause the high upper crustal conductivities.

Magnetic Field Changes

Magnetic field changes are usually attributed to piezomagnetic effects or to electrokinetic effects of fluid flow. On the basis of laboratory measurements of the stress sensitivity of magnetization for igneous rocks [*Stacey and Johnston*, 1972], coseismic changes of a few nanoteslas are to be expected for a 10-bar stress drop. These magnetic field changes are consistent with those measured by *Mueller and Johnston* [1990]. However, piezomagnetic effects cannot explain the larger signals observed by *Mizutani and Ishido* [1976]. *Mizutani and Ishido* [1976] invoke electrokinetic effects to generate signals of 10–20 nT, but *Fitterman* [1981] showed that magnetic fields generated by fluid flow would be much smaller in rocks with conductivities typical of the upper crust. The ULF anomalies observed by *Fraser-Smith et al.* [1990] have yet to be satisfactorily explained.

Electromagnetic Field Changes

Mechanisms for generating ULF em fields have not been discussed extensively, but fields at higher frequencies are usually attributed to piezoelectric effects [*Warwick et al.*, 1982] or redistribution of electric charges [*Gokhberg et al.*, 1982]. Both mechanisms have been observed in laboratory experiments on dry rocks [*Baird and Kennan*, 1985; *Brady*, 1992], but similar experiments on wet rocks have not been performed. *Brady* [1992] observed no em emissions upon fracturing a conductive mineral (pyrite) because the conductor could not maintain a charge separation, so em signals from crustal rocks may be unlikely. Interaction of gravity waves with the ionosphere may be capable of generating em signals, but these would likely be observed as coseismic or postseismic signals. *Mueller and Johnston* [1989] may have observed such a signal associated with the eruption of Mount St. Helens; the magnetic signal propagated at a velocity less than that of sound. There is at present little evidence of the precursory ground motion required to generate the gravity waves.

Experimental Design

Most of the phenomena discussed above are small and require highly precise systems for detection. Complete characterization of the system response and of the noise is essential. Additionally, either continuous or periodic checks of the system response are necessary for precursor monitoring systems because long-term drift and temperature effects can bias the results from these systems. These checks are currently in use in several of the experiments discussed above. *Fraser-Smith et al.* [1990] and *Chevalier and Morgan* [1992] continuously monitor system response to a reference, while others check the system response periodically [*Park*, 1991; *Madden et al.*, 1992].

Systems designed to monitor natural fields must distinguish between possible precursors and normal fluctuations of the fields. One way to identify normal field changes is through the use of the remote reference technique. Common signals on parallel sensors at different locations are attributed to natural field variations, while signals present at one location only are inferred to be tectonic. This technique is currently employed by several of the experiments discussed above [*Johnston et al.*, 1984; *Park*, 1991; *Madden et al.*, 1992]. The appearance of a signal on two parallel electric dipoles of unequal length is used by *Varotsos and Alexopoulos* [1987] to distinguish between SES and noise. The remote reference technique is not currently in widespread use in precursory monitoring at frequencies above 1 Hz, but it would significantly enhance our ability to confirm anomalous fields if it were employed.

Several of the experiments have also established objective criteria by which the anomaly is identified. An important factor for this step is continuous monitoring in order to provide adequate characterization of background levels. *Fraser-Smith et al.* [1990] characterized his background levels with over 2 years of monitoring prior to the Loma Prieta earthquake and thus were able to identify a statistically significant increase in signal levels in the ULF band which preceded the earthquake. *Zhao et al.* [1991*b*] present resistivity monitoring data for over 20 years and show that the decrease prior to the Tangshan earthquake was larger than the background fluctuations and hence may be statistically significant. *Park* [1991] analyzes all of the telluric array data from Parkfield and establishes criteria for identifying significant fluctuations. Finally, *Varotsos and Alexopoulos* [1987] have established objective criteria for recognizing SES.

After fully characterizing the system and carefully defining the nature of the anomaly, a statistically significant correlation between precursory signals and earthquakes must be shown. Additionally, magnitude and distance from the earthquake will be important factors. Every experiment is designed to detect changes resulting from a mechanism. On the basis of that mechanism, a region of sensitivity can be established and all earthquakes in that region be examined. The experiments reviewed above are at different levels of maturity, so this step cannot yet be applied in all cases. Many of the precursors have been seen only once in association with one earthquake [*Fraser-Smith et al.*, 1990; *Park*, 1991], and statistical analysis of one event would not be very useful. Other precursors have had repeated associations and are amenable to statistical treatment [*Varotsos and Alexopoulos*, 1987]. Such attempts have been applied to SES [*Meyer et al.*, 1985; *Mulargia and Gasperini*, 1992], but no consensus has yet been reached among researchers. The success rate of SES as precursors apparently increases for larger magnitude earthquakes [*Hamada*, 1992]; this may be an artifact of statistical analysis on a smaller data set (the number of larger earthquakes) or may be because only larger earthquakes generate detectable signals. The precursor is useful only if it results in a statistically better chance of prediction than a random sampling of the appropriate earthquake catalog.

CONCLUSIONS

Mechanisms such as resistivity changes in response to stress or strain and stress demagnetization have been proven in the laboratory and observed in the field. Thus experiments designed to detect these precursory changes have a solid basis in laboratory data. However, further laboratory measurements of these changes at stress and temperature conditions found in the Earth are needed. Solid-state mechanisms may also be capable of generating electrical precursors, but these first need to be observed in rocks with realistic crustal conductivities. A promising mechanism for explaining a variety of electric and magnetic field changes is the generation of streaming potentials by fluid flow. However, measurements of these potentials at temperatures, pressures, and pressure gradients similar to those found in the crust are needed. Cross-coupling coefficients are especially enhanced with two-phase flow, suggesting that precursory signals may be generated in the partially saturated vadose zone. Indeed, the SES observed in Greece may be due to shallow changes.

Anomalous signals in the field have almost always been larger than would be predicted from laboratory measurements; piezomagnetic effects are the one exception to this generalization. Some of the experiments in which these anomalous signals have been observed are carefully designed and have high precision; simply attributing the signals to error would be a mistake. For example, the resistivity measurements in China clearly show changes prior to the Tangshan earthquake and recovery afterward. However, we know of no mechanism capable of affecting the shallow structure over such a large area. The observations of magnetic field changes prior to the Loma Prieta earthquake are equally convincing, but again, we have no viable explanation for these signals. Some laboratory data indicate that the mechanisms presumably responsible for precursors are nonlinearly related to strain at the low strain levels expected in the Earth prior to earthquakes. It would be a mistake either to blindly attribute all of the anomalous signals to nonlinear mechanisms or to dismiss them; further study is needed. At the present, none of the mechanisms causing the reported precursors is known well enough so that the these precursors could be used reliably for earthquake prediction.

The experiments described above have revealed that precise systems are needed to measure precursors. Resistivity must be measured with precisions of less than 1%, and magnetic fields must have precisions of 0.1 nT. Measurements of electric and magnetic fields without remote referencing are almost useless. Complete measurements of both electric and magnetic fields are needed in many of the existing experiments. Thorough characterization of background noise levels for experiments monitoring changes in field strength requires continuous recording. Complementary geophysical data such as strain measurements and records of water level and perhaps chemistry will help identify the cause of precursory signals.

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