# PRECURSORS TO EARTHQUAKES: SEISMOELECTROMAGNETIC SIGNALS

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**Abstract.** Field measurements in the past several years have documented electromagnetic signals which are attributed to precursory stress and strain changes which were followed ultimately by earthquakes. Precursory electric field changes observed in Greece on multiple dipoles have been used to issue earthquake predictions. While the source of these signals is still unknown, a sufficient number of predictions has been issued to allow some, but not all, statistical analyses to show this method is better than randomly sampling the earthquake catalog. Ongoing efforts to identify the sources of both these signals and the magnetic field variations prior to the Loma Prieta earthquake are focusing on electrokinetic coupling of fluid flow and transient electric fields. A mechanism related to local fluid flow appears to be best suited at this time of explaining the variety of purported precursors. However, much more work is needed to improve the observations and refine the models of precursor generation. Efforts to monitor magnetotelluric transfer functions at longer periods ($T > 10$ s) have been hampered by variability of the functions. The use of modern noise reduction techniques such as remote referencing should reduce this variability, but may not reduce errors to a level needed for monitoring. Monitoring of high frequency (81 kHz) seismoelectric emissions may be promising, but lack of simultaneous observations on multiple instruments hinders the utility of this technique.

## 1. Introduction

Electromagnetic signals from earthquakes cover an especially broad range of phenomena from earthquake lights to changes in resistivity over periods of years. Signals may be precursory (prior to earthquakes) or coseismic (coincident with earthquakes). Frequencies range from megahertz to quasi-dc. A comprehensive review of observations and mechanisms for this broad field would be impractical, so I will instead limit this review to a few promising topics. The current review will build on previous reviews by Parrot and Johnston (1989) on magnetic fields and radio frequency effects and by Park *et al.* (1993) on electromagnetic phenomena in the ULF band. In particular, I will attempt to update topics in these previous reviews with new, published results since 1992. The focus here will also be on precursory phenomena; coseismic phenomena will not be mentioned except as they relate to mechanisms. Finally, the discussion will be limited to phenomena at ULF frequencies (0.01–10 Hz) and below, except for a section on seismoelectric emissions at kilohertz frequencies.

This review will discuss specifically the following topics: seismic electric signals; ULF magnetic signals; monitoring electromagnetic fields; and seismoelectric emissions. A brief review of the observations will be followed by evaluation of the proposed mechanisms for each topic. The review will conclude with a summary

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of constraints on observations and mechanisms from other geophysical data and speculation on a possible model to explain a number of observations.

## 2. Seismic Electric Signals

Seismic electric signals (SES) are anomalous perturbations of the potential measured on electric dipoles and have no accompanying magnetic field change of sufficient size to be observed on existing sensors. Some experiments have monitored electric and magnetic fields simultaneously (Hadjioannou *et al.*, 1993) while others have observed only electric fields (Fujinawa and Takahashi, 1993). Anomalous magnetic fields may have been present in the latter type of experiment, but this cannot be confirmed. I therefore treat these restricted observations as electrical signals only, following the convention of Park *et al.* (1993).

The potential difference on an electric dipole can change because the potential near one of the electrodes changed or because the electric field parallel to the dipole changed. In the following discussion, I differentiate between electric fields with scale lengths greater than several dipole dimensions and electric potentials generated by sources with dimensions smaller than one dipole. While this distinction may seem artificial, potential differences caused by electric fields can be confirmed with multiple dipoles and those caused by smaller sources may be seen on only one dipole. As discussed below, these latter changes may be the result of small scale variations caused by tectonic shifts or the result of electrochemical changes in the immediate vicinity of the electrode.

Miyakoshi (1986) provides an excellent example of changes in electric potential (Figure 1). Potential differences on two short, parallel dipoles with a common carbon rod electrode were monitored continuously. Although the difference in length between the two dipoles was only 10 m, one saw potential changes prior to a $M = 5.6$ earthquake and the other did not. The dipole that saw the changes terminated in the fault zone; the other spanned the fault. These changes (Figure 1) cannot result from changes in an electric field (as defined above) or signals would have been seen on both dipoles which were proportional to dipole length ($\Delta V = E * \text{length}$). A tectonic cause for these signals was preferred because no rainfall occurred at the time of the signals and because the author concluded that the electrode had aged sufficiently (Miyakoshi, 1986). Miyakoshi (1986) attributed these signals to changes of the self-potential in the surrounding rock and proposed a fluid flow model to explain the changes.

It seems very unlikely that two electrodes only 10 m apart would detect such different self-potential signals from a source at any appreciable distance from the electrodes. Models of such mechanisms (Miyakoshi, 1986; Fitterman, 1981) show broad anomalies. A more likely explanation is that a very local change in self-potential occurred, whether in the fault zone or at the electrode surface. While this correlation between self-potential and earthquakes is intriguing, the observation

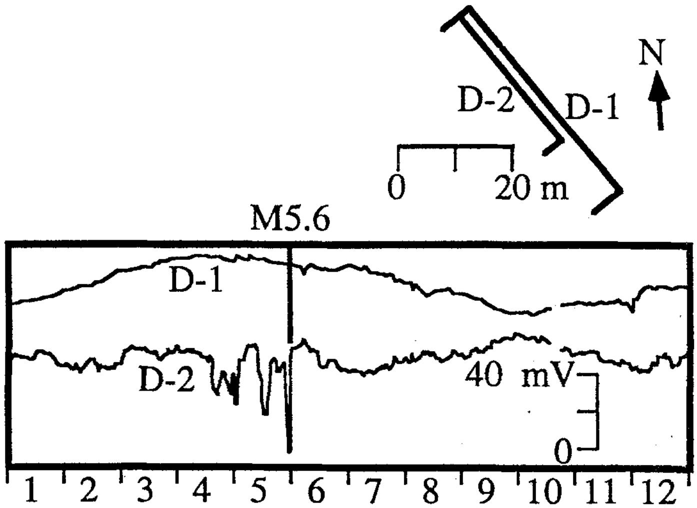


*Figure 1.* Measurements of self potential on two parallel dipoles (D-1 and D-2) of unequal length (after Miyakoshi (1986)). Time in months is displayed, as well as time of $M5.6$ earthquake at an epicentral distance of 3 km and a depth of 17 km. Note that shorter dipole records large signals and longer dipole shows nothing.

on only one sensor is disturbing because large variations in the local self-potential can result from nontectonic causes. Morat and Le Mouel (1992) have shown that signals in excess of 0.05 V $m^{-1}$ are observed on short dipoles (1 m) in partially saturated rocks in response to variations of atmospheric pressure. Such signals have durations of hours to days. Morat *et al.* (1994) have also shown that self-potentials of up to 0.1 V $m^{-1}$ can arise from sap flow in trees. They suggest that similar signals will arise due to fluid circulation in roots thus complicating correlation of electric signals with tectonic events. Localized variations in temperature, atmospheric pressure, fluid content, and fluid chemistry can thus all result in changes of self-potential. Changes in the contact potential of the electrode may also occur. Efforts to reliably identify changes of self-potential due to tectonic events need to include multiple sensors, corroborating measurements such as water level, and thorough documentation of nontectonic events. Low noise, stable nonpolarizable electrodes should be used for all measurements. Petiau and Dupis (1980) have shown that the two quietest electrodes are Ag–AgCl and $Pb–PbCl_2$, but that the latter is more stable over long periods of time. Observations on multiple sensors are needed to verify the existence of signals, and corroborating measurements are necessary for determination of the source of the signal.

Changes in the potential difference on an electric dipole can also be caused by changes in electric fields. The best known example of these types of changes are the SES reported from Greece by Varotsos, Alexoupoulos, and Nicomos (VAN method; Varotsos *et al.*, 1993a). Anomalous electric signals are observed on both short (10–200 m) and long (1–3 km) dipoles with electric field amplitudes of up

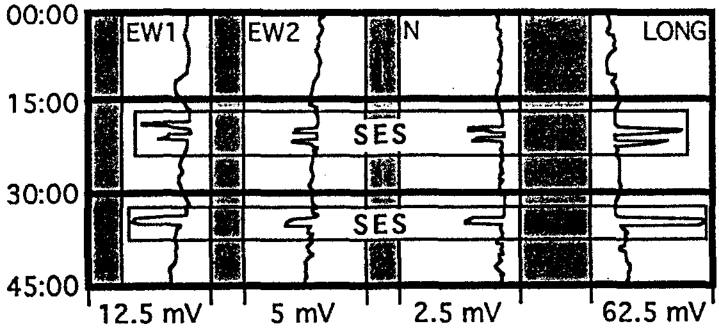


*Figure 2.* Examples of seismoelectric signals (SES) from station in Greece (after Varotsos *et al.* (1993a)). Signals from four different dipoles at three different orientations are shown. Dipole lengths are EW1, 181 m; EW2, 47.5 m; N, 48 m; and LONG, 2.5 km. Note how signal scales with dipole length for parallel dipoles EW1 and EW2.

to 20 $\mu$V m$^{-1}$ and durations of a few minutes (Figure 2). The signals are seen on multiple dipoles located over an area of 1 km$^2$ (Varotsos *et al.*, 1993b). Multiple dipoles are used to distinguish between SES, induced telluric signals (seen on all dipoles and at multiple sites), noise from a single electrode (seen on only one dipole), and noise from known sources such as villages and towns (polarity patterns in signals on short and long dipoles). Based on correlations of SES with earthquakes in Greece, Varotsos *et al.* (1993a) have developed maps (Figure 3) which show the sensitivity of stations to different regions of Greece. The SES appear to precede earthquakes by as much as 22 days. While the selective sensitivity of the VAN stations is a controversial topic, Varotsos *et al.* (1993a) have issued numerous predictions via telegram of earthquakes since 1984 using these sensitivity maps. These predictions include a location within 100 km and a magnitude determined from size of the electric signal.

Recent attempts have been made in Japan to detect SES (Nagao *et al.*, 1993), but cultural noise has forced the use of longer dipoles (1–10 km). SES-like signals have been detected prior to several earthquakes, but these have been traced to potential variations of one electrode. Nagao *et al.* (1993) are cautious about labeling these signals as precursors because similar signals were observed which were unrelated to earthquakes. SES have also been detected in boreholes with measurements of vertical electric fields (Antonopoulos *et al.*, 1993). Simultaneous signals have been seen in two boreholes in association with a few of the SES at one of the nearby VAN stations (Ioannina), so it is unlikely that these signals are electrode noise. Other efforts to monitor vertical electric field have encountered difficulties with lack of correlated signals between parallel sensors (Fujinawa and Takahashi, 1993).

The VAN predictions constitute the only published data set with multiple tests of the methodology and are therefore amenable to statistical analysis. Statistical tests published so far have analyzed only whether the VAN method does significantly better than randomly sampling the earthquake catalog. Most authors (Hamada,

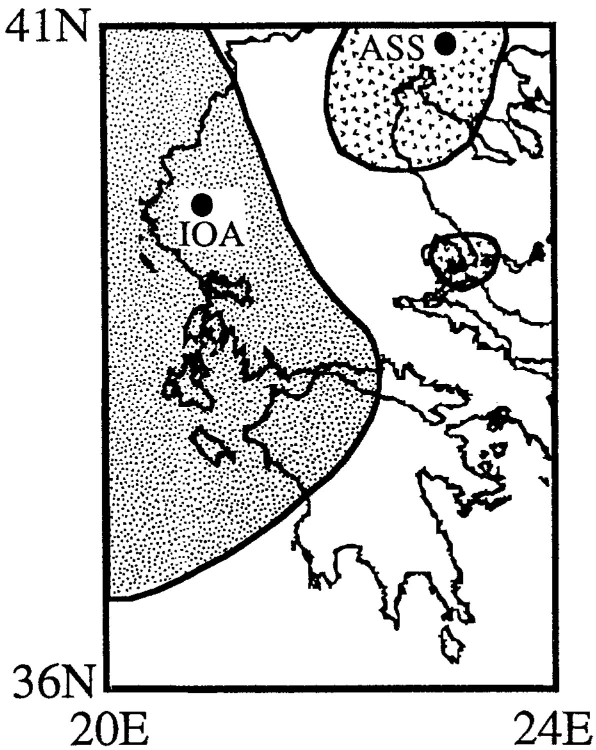


*Figure 3.* Maps of sensitive areas for stations IOA and ASS (after Varotsos *et al.* (1993a)). Earthquakes within sensitive areas are preceded by SES at stations. Note lack of overlap between areas.

1993; Mulgaria and Gasperini, 1992) have randomly sampled a uniform spatial distribution, although Hamada (1993) did attempt to incorporate a nonuniform spatial distribution by making earthquakes three times more likely in one region. Aceves *et al.* (1996) used the earthquake catalog from 1960 to 1985 to create a nonuniform probability density map for their comparison of the VAN method to a random sample. Additionally, they used the observed magnitude-frequency relation for the catalog in the random predictions. Most other studies (Hamada, 1993; Shnirman *et al.*, 1993) have simply used overlapping distributions of magnitudes (e.g., $M_s \geq 5.0$, $M_s \geq 5.5$, etc.). The statistical tests have reached different conclusions. Hamada (1993), Aceves *et al.* (1996), Shnirman *et al.* (1993) and others have concluded that the VAN method does significantly better than either a uniform or nonuniform random model (Figure 4). Mulgaria and Gasperini (1992) concluded that a random model did better than the VAN method and raised the intriguing possibility that the VAN predictions were better correlated with earthquakes occuring *before* the SES. Thus, the SES could be postseismic signals. All of these tests were performed with predictions issued between 1987–1989 (Varotsos and Lazaridou, 1991). Based on my analysis of the 16 predictions issued between August 10, 1989 and May 31, 1992 (Varotsos *et al.*, 1993b; Varotsos and Lazaridou, 1991), the

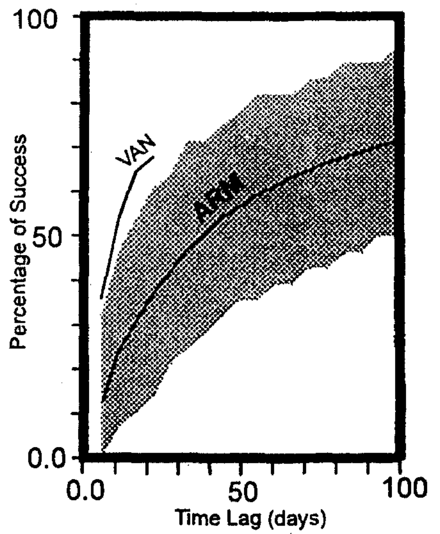


*Figure 4.* Comparison of prediction rates (number of successful predictions over total number of predictions issued) for VAN method and random model for different time lags between prediction and the earthquake (after Aceves *et al.* (1994)). Random model consisted of sampling a probability density surface created from earthquakes between 1960 and 1985 (Aceves *et al.*, 1994). The lower bound of the shaded region is the average prediction rate and the upper bound is the 99% confidence limit. The VAN method has a prediction rate much higher than the random model, even at the 99% confidence limit.

VAN method has a prediction rate of 69% (11 out of 16) which is comparable to a prediction rate of 79% for the period between 1987 and 1989 (Hamada, 1993).

These seemingly contradictory conclusions result from differences in the definition of success. Authors such as Hamada (1993) have concluded that the VAN method works better than random chance because they have compared prediction rates (number of successfully predicted earthquakes divided by total number of predictions). The VAN method is comparable to or worse than random chance if success rate (number of successfully predicted earthquakes divided by total number of earthquakes) is compared because there were approximately 10 times as many earthquakes as predictions. Future analyses of success rate should incorporate the selective sensitivity that is a central feature of the VAN method. Earthquakes outside these sensitive areas should not be included in an analysis of the success rate. Nonetheless, I conclude that the VAN method has a prediction rate greater than would be estimated from randomly sampling the earthquake catalog and that the VAN experiment warrants close scrutiny in order to determine the cause of SES. (A debate on the statistical validity of the VAN method is scheduled for publication in *Geophysical Research Letters* in 1996).

Several different mechanisms have been proposed for the causes of SES. These mechanisms can be grouped broadly into two types: local and hypocentral. The local mechanisms require generation of SES near the recording sites in response to precursoly strain or stress (Dobrovolsky *et al.*, 1989). SES result from electrokinetic phenomena associated with possible fluid flow across lithologic boundaries, and dipoles spanning these boundaries are very sensitive to SES. The hypocentral mechanisms require generation of current in the preparation zone of the earthquake and diffusion of electromagnetic energy to the recording site (Varotsos *et al.*, 1993a). Current generation can result from stress-induced reorientation of impurity-vacancy dipoles (Dologlou and Varotsos, 1986), electrokinetic effects, or displacement of charged edge dislocations (Slifkin 1993). Each of these mechanisms has aspects which are supported by the observations; none are completely comistent with all of the data however.

Varotsos *et al.* (1993a) discussed objections to a local electrokinetic effect, so I will only summarize them here. One objection is that pressure varies according to the diffusion law, and a pressure variation due to a rapid strain change may be much slower. The electric field is proportional to the gradient of the fluid pressure across heterogeneities in streaming potential coefficients (Nourbehecht, 1963) and should have similar time behavior as the pressure gradient. Varotsos *et al.* (1993a) conclude that starting times and durations of SES should be dependent on placements of electric dipoles with respect to the heterogeneities and that time lags of several hours to days should be expected. Such behavior is not observed in geothermal areas where streaming potential coefficients have been measured in situ by varying flow rates from wells, however (Ishido *et al.*, 1983). Potentials over a 0.25 km$^2$ area were monitored as an injection well was turned off and then on. Changes in potential were established with the same time scales as the changes in water level (few minutes), and there was no observable delay over the region monitored (Ishido *et al.*, 1983). Such behavior has also been seen at Beowawe Geothermal Area in Nevada (USA) over an area of roughly 6 km$^2$. While the fluid flow conditions in a geothermal area are probably inapplicable to the VAN measurement sites, these results show that SES-like signals can result from electrokinetic effects.

Another objection to the electrokinetic mechanism is that SES signals are proportional to the length of the observation dipole, and this could only result if the dipoles are located on one side of a heterogeneity (Varotsos *et al.*, 1993a). Additionally, the variation in dipole lengths from less than 100 m to a few kilometers suggests that the source of the SES cannot be at the measurement site or the proportionality to length would not be preserved. A source close to the site possibly causes the signals, but this must be verified. Special geologic and/or geometric conditions could exist at each site; this would match the general observation of selectivity of sites. Invoking unusual conditions at and near each site is unsatisfying however, and must be verified with detailed site surveys in any case.

The principal advantage of a local electrokinetic effect is the elimination of the need to focus electromagnetic energy from the hypocenter at the sensitive sites. Without this focusing, Honkura and Kuwata (1993) have shown that electric fields of $10^{-13}$ V m$^{-1}$ result from a current source of 1 A-m at a depth of 5 km in a homogeneous earth with realistic crustal resistivities (100 ohm-m). In order to generate signals of $2 \times 10^{-5}$ V m$^{-1}$, a current source of $2 \times 10^{8}$ A-m would be needed. Currents of $10^{5}$ A or more would be needed if source regions were of the order of kilometers, and these are unrealistically high. Varotsos *et al.* (1993a) propose the existence of conductive channels connecting the source region to sensitive sites in order to focus the electromagnetic energy and thus drop the necessary current levels. Such channels could also explain why some sites are insensitive; they lie outside the channels. Bernard (1992) and Utada (1993) have both presented models which would channel currents, but Bernard concludes that these are unlikely and prefers local electrokinetic effects as an explanation. Utada's model is again very specialized, with a resistive body in an otherwise conductive zone channeling current from the source. This model seems as unlikely as the specialized models needed for local electrokinetic effects.

A critical constraint for any model is the observation by Hadjioannou *et al.* (1993) that transient magnetic signals are not associated with SES recorded at station IOA (Figure 5). Earlier attempts also showed that there were no observable magnetic fields observed with SES and that a magnetotelluric (MT) cancellation scheme showed that the SES were preserved in the residual electric fields (Chouliaras and Rasmussen, 1988). Hadjioannou *et al.* (1993) also removed the natural, induced electric field from the recorded electric fields by using the MT transfer function and the observed magnetic fields. The SES is preserved in the residual electric fields almost perfectly (Figure 5). This result leads me to conclude that the SES mechanism generates no observable magnetic fields for the following reason. Goldstein and Strangway (1975) showed that the ratio of the electric and magnetic fields in the far field is identical to that for a plane wave source and that conventional magnetotelluric (MT) analyses are valid for the fields from the grounded dipole. Far field is deffned as beyond three skin depths ($\delta = \sqrt{(2/\omega\mu\sigma)}$). With a rise time of less than 60 seconds and an assumed earth conductivity of 0.01 S m$^{-1}$, the skin depth for the transient part of the SES is 38 km and the transition to far field behavior occurs at 114 km. Under the assumption that SES are generated at the hypocentral location, the MT cancellation scheme of Hadjioannou *et al.* (1993) should have reduced the magnitude of the SES if there was an anomalous magnetic field. If the SES had been generated in the near field and had an associated magnetic field, then the cancellation scheme should have altered the shape of the SES because the MT transfer function would not have correctly related the anomalous electric and magnetic fields. Additionally, Park and Fitterman (1990) have shown that low frequency EM effects were clearly present in the received electric fields from a current dipole, even over distances of several kilometers. I conclude that the mechanism generating the SES does not result in observable magnetic fields, regardless of location.

One mechanism capable of generating electric fields without significant magnetic fields is electrokinetic effects from fluid flow. Fitterman (1979) and Mizutani *et al.* (1976) have shown that magnetic fields can arise from fluid flow across

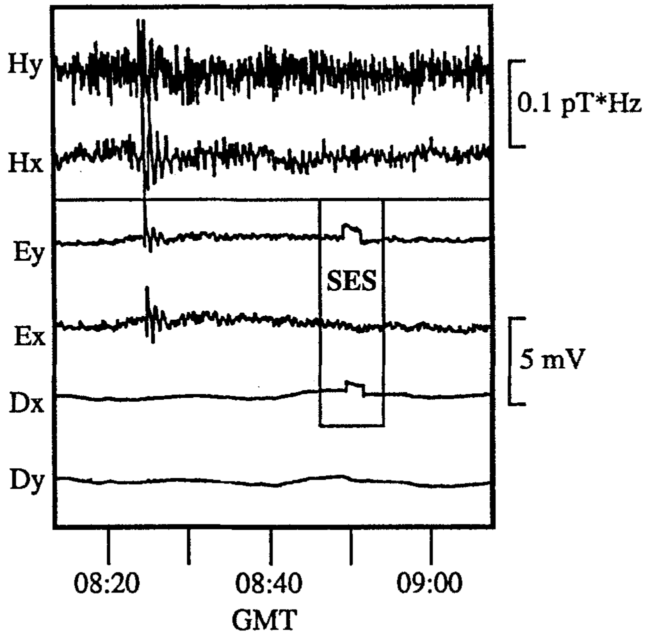


*Figure 5.* Records of electric, magnetic, and differential electric fields from Greece (after Hadjioannou *et al.* (1993)). The differential fields ($Dx$ and $Dy$) are generated by subtracting the inductive component of the MT field from the electric dipoles. Note that the SES is preserved almost completely, indicating that there is no corresponding magnetic field associated with the SES.

discontinuities in electrokinetic properties, but Fitterman (1979) argues that the magnetic field will be small and difficult to detect. Mizutani *et al.* (1976) also argue that the relaxation time for the electric field will be very short and a steady state with a net zero current flow will be achieved quickly. Under these conditions, a detectable magnetic field may not be generated and the lack of an observable magnetic field may be consistent with an electrokinetic mechanism.

The SES observed in Greece are repeatable phenomena that are observed on multiple dipoles with nonpolarizable electrodes. SES magnitudes observed on parallel, independent dipoles scales approximately with the length of a dipole (exact scaling is assured only for a homogeneous earth). They simply cannot be attributed to electrode noise. SES are apparently not accompanied by fluctuations in the magnetic field. Discrimination between SES and geomagnetic micropulsations is possible because SES are observed at only a few stations, while micropulsations are seen simultaneously across the network. Repeatable correlations between SES at individual sites and earthquakes in the same region have permitted Varotsos *et al.* (1993a) to develop a scheme to issue predictions via telegram. While statistical studies have shown both that the VAN method is no better than chance and that it is much better than chance, I conclude that the method is much better than a random sample of the earthquake catalog and thus warrants additional study.

Future efforts should be directed at identification of the mechanism causing SES. Generation of SES by local electrokinetic effects and by currents at the hypocentral region have been proposed, but both hypotheses are inconsistent with some of the observations of SES. Measurements of ground water level and strain are needed at the VAN observation sites. In addition, the geoelectrical structure of the sensitive sites should be characterized with resistivity measurements. Gershenzon and Gokhberg (1993) showed that, within an area of approximately 4 km$^2$, only three of six dipoles recorded SES due to an earthquake 200 km from the station. It seems unlikely that conductive channels could focus the electromagnetic energy from a distant earthquake so efficiently that only some of the dipoles in a restricted area would record signals. In any case, more complete characterization of the sensitive sites is needed. The regional geoelectrical structure of Greece also needs to be characterized and models for the distribution of the sensitive regions developed. Such models are essential to tests of the hypothesis that the signals are generated at the hypocentral region and focused at the sensitive sites.

## 3. Seismomagnetic signals

Fraser-Smith *et al.* (1990) observed increases of the magnetic field amplitude at ULF frequencies up to one month prior to the 1989 $M_s7.1$ earthquake at Loma Prieta south of San Francisco. The station recording these changes was located only 7 km from the epicenter, and a similar station recording ELF/VLF data 52 km from the earthquake saw no changes. These increases in magnetic field amplitude were most prominent at frequencies below 0.2 Hz and began with a 1500 pT increase approximately one month prior to the earthquake. Two additional increases were observed two weeks prior and a few hours prior to the earthquake (Figure 6), and the higher amplitudes continued for almost six weeks after the main shock (Fenoglio *et al.*, 1993). Fenoglio *et al.* (1993) have shown that there is no significant correlation between the amplitude of the magnetic field and either the frequency or magnitude of the aftershocks. Similar ULF sensors have subsequently been installed in central and southern California and have recorded no signals associated with either the $M7.4$ Landers or the $M_s6.7$ Northridge earthquakes in 1992 and 1994, respectively (Fraser-Smith *et al.*, 1994).

This evidence for seismomagnetic signals indicates that a low frequency electromagnetic signal is not being generated at the hypocenter or epicenter and then propagating through the atmosphere or earth to the recording sites. The lack of signal from the Landers and Northridge earthquakes and the large signal from the Loma Prieta earthquake on a proximal sensor indicates that the source is localized to the hypocentral region and that the signals decay rapidly in the earth away from the source. Such conclusions have been the basis of models of magnetic field generation by fluid flow (Draganov *et al.*, 1991). A crucial constraint on any model of the source is the oscillatory nature of the magnetic field. Fraser-Smith *et al.* (1990)

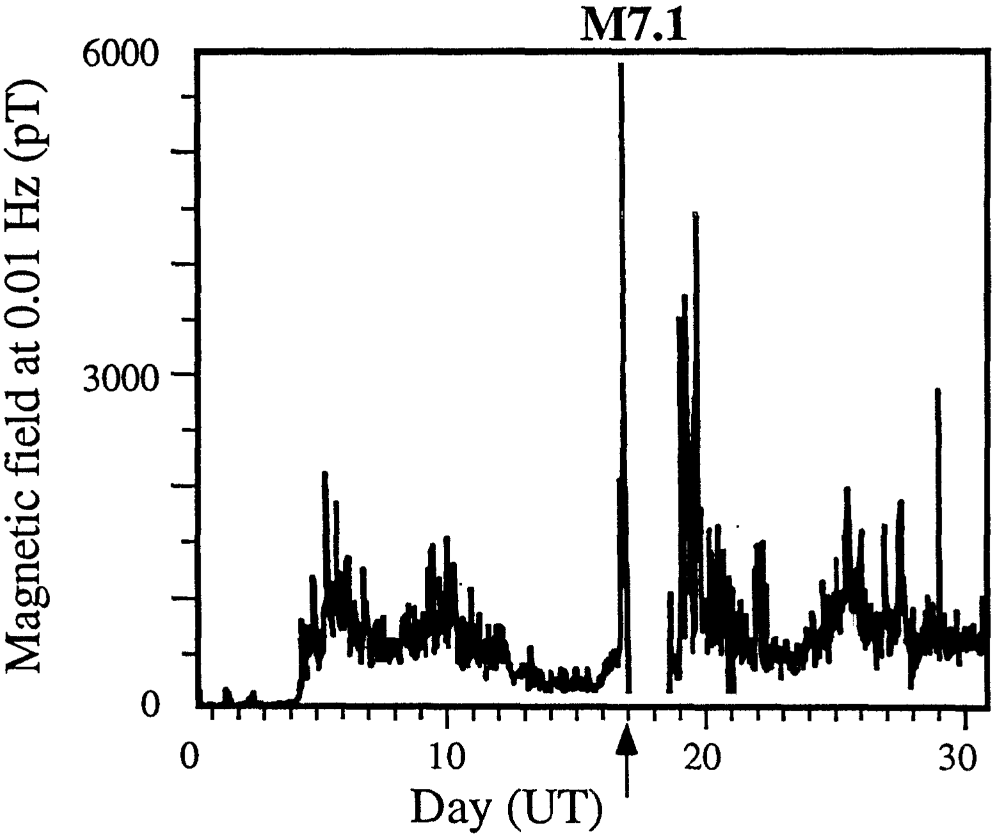


*Figure 6.* Magnetic field at 0.01 Hz recorded on a single induction coil prior to and after the 1989 Loma Prieta earthquake (after Fraser-Smith *et al.* (1990)). Note the increases in magnetic field almost 2 weeks prior to the earthquake and the large increase a few hours prior to it. No field was recorded immediately after the earthquake because of power loss.

observed these changes in a narrow band between 0.01 and 0.2 Hz. Fenoglio *et al.* (1993) have proposed a model whereby isolated fluid reservoirs in the fault zone are ruptured, causing electrokinetically generated transient magnetic fields. The oscillatory nature results from the continued rupture of these isolated reservoirs. Electric and magnetic fields would be generated locally along the fault zone, thus explaining why sites close to earthquakes detect these signals while distant sites detect nothing. However, Fenoglio *et al.* (1993) assume that the compensating conduction current does not flow through the same path as the streaming current. If this unproven assumption is incorrect, then the dipole moment of the current source is greatly reduced (if not reduced to zero) and only small magnetic fields could possibly be generated. This latter alternative would be more consistent with the results of Fitterman (1979) and Miyakoshi (1986). This mechanism is thus speculative, and verification must await additional observations of strain, fluid flow, and electric field in association with the magnetic field. The Parkfield Prediction Experiment could provide these observations prior to the next characteristic earthquake at Parkfield (Park *et al.*, 1993).

## 4. Seismoelectromagnetic Signals

Long period ($T > 10$ s; $f < 0.1$ Hz) electric and magnetic fields have been monitored simultaneously in several experiments for earthquake prediction (see Niblett and Honkura, 1980 for review of earlier work). These experiments have been performed to detect two types of precursory events: resistivity changes (Ernst *et al.*, 1993) and electromagnetic (EM) emissions from the earth (Rozluski and Yukutake, 1993). Ernst *et al.* (1993) attempts to correlate 10–40% changes in apparent resistivity with a series of small ($M < 4.1$) earthquakes, but the authors note that considerable artificial signals are caused by electric trains at one of the stations. As a result, the authors attempted to employ a differencing scheme to cancel out noise at one of the stations. Insufficient detail about the noise reduction process is presented to permit assessment of its efficacy, and no confidence limits on the results are presented in any case.

A much more successful method for eliminating noise and obtaining stable MT impedances is the remote referencing scheme (Gamble *et al.*, 1978). This method of MT analysis is in nearly universal usage because of its noise reduction capability. The usual analysis for the MT impedance involves Fourier transformation of the time series followed by computation of auto- and crosspower spectra of the electric and magnetic fields. Noise biases the autospectra and correlated noise between field components can bias crosspower spectra, resulting in impedance estimates that are severely affected (Figure 7). Gamble *et al.* (1978) showed that use of a remote measurement of the magnetic field eliminates the need for autopowers and yields much more stable impedances as well as much smaller errors in the estimates (Figure 7). The remote magnetic sensors are placed sufficiently far from the MT station that noise between the local and remote magnetic fields is uncorrelated. Then, the MT impedance is estimated using crosspower spectra between the local electric and magnetic fields and the remote magnetic fields. Although errors as small as 0.4% are reported for apparent resistivities from MT impedances at high frequencies ($f > 1$ Hz), errors are typically several percent for excellent MT soundings and orders of magnitude for poor soundings. Abundant laboratory and field measurements show that the expected changes of resistivity prior to earthquakes are on the order of a few percent (Park *et al.*, 1993). Once the region of perturbed resistivity is embedded in a much larger invariant medium, the changes of a few percent will be diluted (Park *et al.*, 1993). Detection of these diluted changes by monitoring an MT transfer function that is accurate only to within a percent is probably unlikely.

Another problem with the use of MT methods for monitoring precursors is that the apparent resistivities determined from the impedance tensor are dependent on

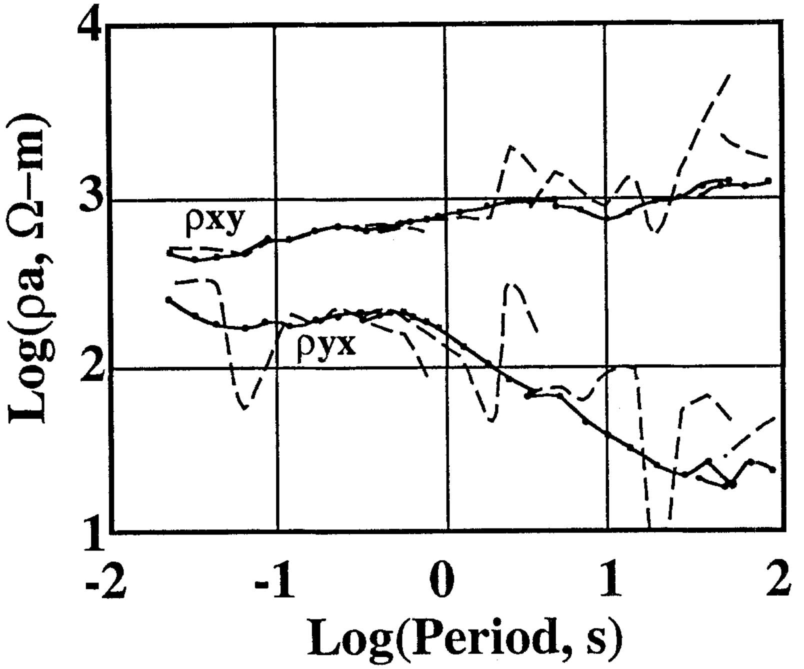


*Figure 7.* Comparison of MT sounding curves using a local reference (dashed curve) and a remote reference (solid curve with data points). The remote reference data are much smoother and more stable (after Gamble *et al.* (1978)).

the directions of the principal axes of the tensor. These directions are normally determined from the singular value decomposition of the impedance tensor and usually correspond to the directions of the maximum and minimum electric field eigenvectors (LaTorraca *et al.*, 1986). The apparent resistivities and phases are determined from the complex eigenvalues associated with these eigenvectors. Park (1991) showed that the principal directions of a telluric tensor relating electric fields measured at two different locations can vary by as much as 15° between adjacent days. The electric fields are much more coherent with each other than with the magnetic field, so angular variability is even more important for the MT tensor. Changes of up to 3% per degree of rotation can result in the maximum and minimum apparent resistivities (Figure 8). This problem is well-known in the MT field, and several approaches have been suggested for the removal of this angular dependence. Rotationally invariant parameters such as the arithmetic mean ($Z_1 = 0.5(Z_{xy} - Z_{yx})$) and the determinant of the tensor ($Z_2 = Z_{xx}Z_{yy} - Z_{xy}Z_{yx}$) (Berdichevskiy and Dmitriev, 1976) have been proposed. Alternatively, the rotation directions for $Z_{xy}$ and $Z_{yx}$ could be fixed for each frequency as proposed by Park (1991) for telluric monitoring. Once a true perturbation in apparent resistivity is identified, the final problem is that the MT response is global and not local. Conductivity structure hundreds of kilometers away can contribute to the MT

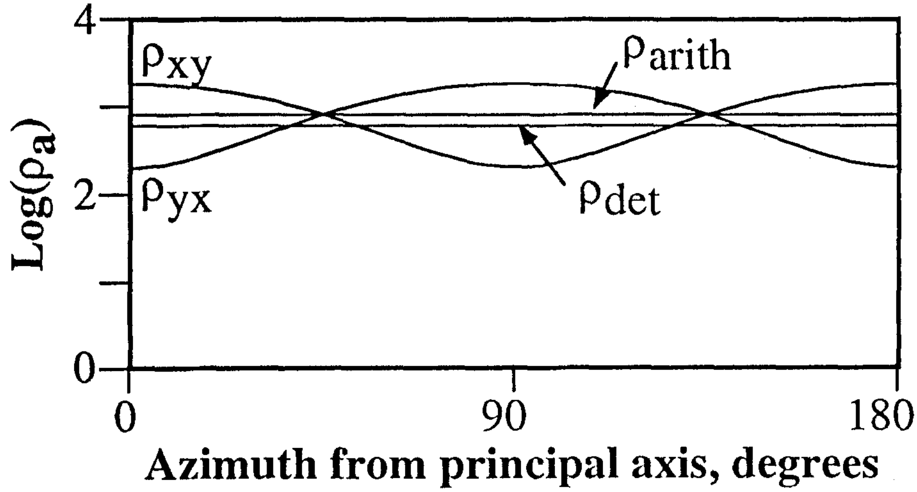


*Figure 8.* Variation of apparent resistivities versus direction of assumed principal axes. Also shown are the arithmetic mean and determinant. Note that the impedance tensor in this test case was 2-D, so the determinant and geometric means are identical.

response measured at a site (Ranganakyaki and Madden, 1980), and identifying the cause of a perturbation seen at one site will be very difficult. One solution, albeit expensive, to this problem is to use a network of sites. Such a network would help locate the source of the perturbation and also provide remote references for processing data.

MT measurements have also been used in attempts to detect EM emissions from seismogenic zones. Rozluski and Yukutake (1993) derive a formula for the transmission coefficient for radiated power:

$$T = 4c\frac{Re[G_{xy}(E_x, B_y)] - Re[G_{yx}(E_y, B_x)]}{G(E_x + cB_y) + G(E_y - cB_x)}, \tag{1}$$

where $G$ is the power spectrum. The transmission coefficent is usually positive, indicating energy radiating into the earth. It is occasionally negative however, and Rozluski and Yukutake (1993) attempt to correlate the negative values with earthquake activity although no uncertainties are provided for the transmission coefficients. Translation of the transmission coefficent into more conventional MT impedances shows that it should always be positive (See Appendix), so it is not clear what is the signficance of a negative transmission coefficient.

Attempts to monitor magnetotelluric fields for earthquake precursors have not advanced much past the initial attempts by Honkura *et al.* (1976). Standard techniques of noise reduction via remote referencing of multiple stations are not currently employed, although Ernst *et al.* (1993) attempted to cancel out some identifiable noise. There is some question whether the accuracy of impedance determination will reach the low levels needed to detect resistivity changes. In any case, analysis of multiple stations is necessary for confirmation of any observed changes and for remote referencing. Additional work is also needed to evaluate the uncertainties in the transmission coefficients.

## 5. Seismoelectric Emissions

Seismoelectric emissions (SEE) are a subset of seismoelectromagnetic signals, but are treated separately here because they occur at much higher frequencies (81 kHz) than discussed above. Increases in signal amplitude at 81 kHz occur from a few minutes to over an hour prior to earthquakes and volcanic eruptions (Gokhberg *et al.*, 1982; Yoshino 1991). The increases are detected at distances of tens to hundreds of kilometers from epicenters, and are observed on, although not correlated between, multiple instruments (Yoshino *et al.*, 1985). These signals are attributed to generation of electric currents by fracturing rocks at seismogenic depths and subsequent radiation of EM waves to the surface and beyond (Yoshino, 1991). A basic difficulty with this mechanism is that the attenuation of the EM radiation as it passes through a conductive earth. Skin depths at 81 kHz range from 5.6–56 m for resistivities of 10–1000 ohm-m. This may no longer be a difficulty however, because the magnitudes of the signals are found to correlate with distance from the source but not with the magnitude or depth of the earthquake (Yoshino and Sato, 1993). Thus, the signals may be generated at the surface of the earth. Current generation from crushed rock may be the source of the emissions, but no in situ measurements of these currents have been reported. Alternatively, small amplitude precursory ground motion may couple with the ionosphere to generate the signals. Parrot *et al.* (1993) have shown that the amplitude of an acoustic wave generated at the earth's surface grows exponentially with height because of the exponential decrease of atmospheric density. Vibration of the ionosphere by the acoustic wave would result in generation of EM waves. Additional work is needed on this causative mechanism, however.

## 6. Discussion of Constraints on Mechanisms

The most important constraint on any mechanism of generating electromagnetic earthquake precursors is the fact that the earth is essentially a conductor. Resistivity values from magnetotelluric soundings are rarely as high as 10,000 ohm-m and are usually 10–1,000 ohm-m (Jones, 1992). These values are observed in a variety of continental geological settings from shield areas to convergent margins. Thus, EM radiation from seismogenic depths will be severely attenuated by the time it reaches the surface. Honkura and Kuwata (1993) have calculated the electric field strength at various distances from a variety of sources buried at depths from 1 to 30 km (Figure 9). The electric field at ULF frequencies from a 1 A-m source

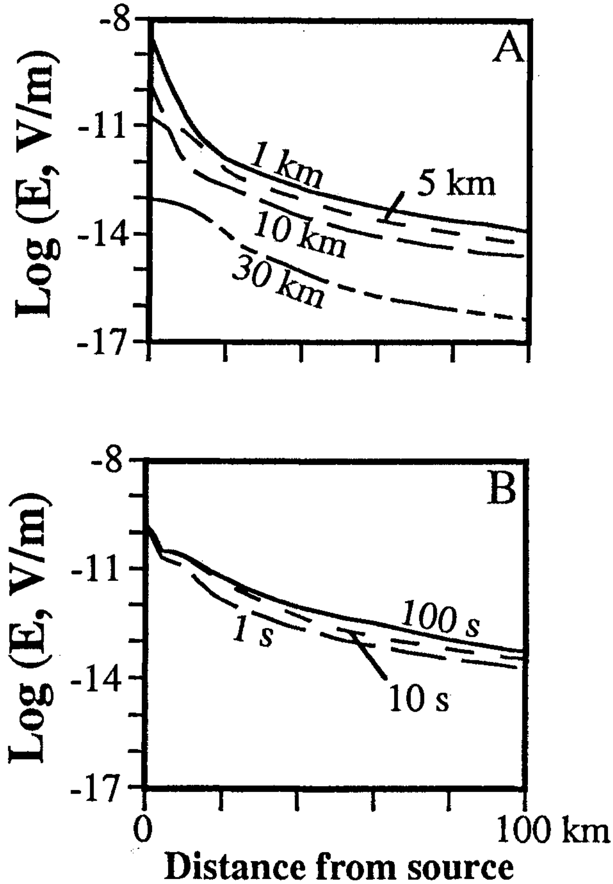


*Figure 9.* Calculated electric fields at a depth of 10 m at varying distances from a parallel electric dipole source (after Honkura and Kuwata (1993)). Values for different source depths are shown in A, and field strengths for different periods of oscillation are shown in B. Note that far field behavior is achieved in these examples at a distance of 20–30 km. All calculations were done for a halfspace with $\sigma = 0.01$ S m$^{-1}$ for a source strength of 1 A-m.

dipole is generally at a level of $10^{-12}$ V m$^{-1}$ directly above the source and decays approximately as $1/r^3$ in the far field (Figure 9). Honkura and Kuwata (1993) conclude that large current sources are needed to generate observable electric fields at distances of 10–100 km. With appropriate geometry, the EM radiation could be focused and therefore decay as $1/r^2$ with distance from the source. This focusing would reduce the size of the current source needed.

A general observation from all of the experiments designed to detect precursors is that signals are observed before earthquakes, but rarely are signals observed at the time of earthquakes. The only exceptions to this observation appear to be piezomagnetic changes in the total magnetic field related to stress release (Mueller and Johnston, 1990), coseismic resistivity changes (Yamazaki, 1974), and resis-

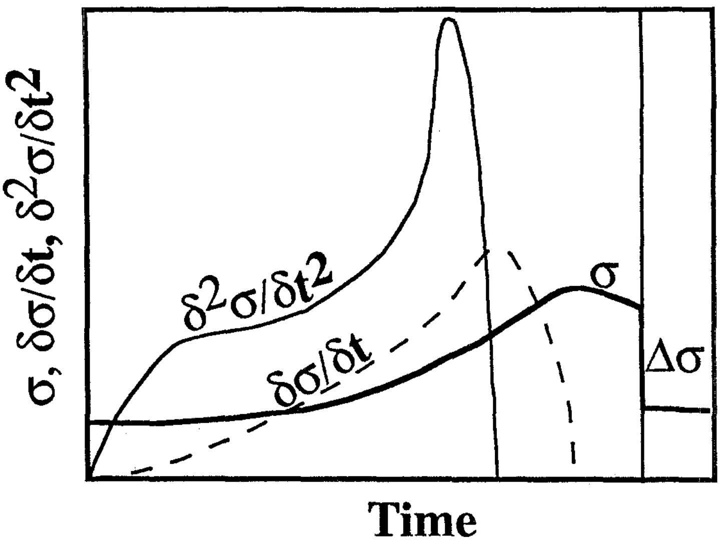


*Figure 10.* Variation of stress and its temporal derivatives as stress is inaeased to failure (after Teisseyre (1992)). Note that the second derivative of stress achieves a positive peak prior to failure. However, the behavior of this derivative at the time of failure is not shown.

tivity decreases seen at the time of rockbursts in mines (Stopinski and Teisseyre, 1982). This general observation of precursory signals without coseismic signals is important because strain changes are largest at the time of an earthquake (Johnston *et al.*, 1987). Any mechanism must explain why electrical signals are associated with small precursory ($O(10^{-9}\epsilon)$) strain changes but not with much larger coseismic strains ($O(10^{-6}\epsilon)$). Teisseyre (1992) has proposed a model of formation of charged dislocations and subsequent reestablishment of charge balance in which the curent density is proportional to the second time derivative of the stress (Figure 10). Such a model predicts maximum current generation as the stress is building. However, the step change in stress at the time of failure should then result in a large electrical signal because of its short duration and Teisseyre (1992) does not discuss the coseismic stress behavior and generation of signals. This large electric signal will contain high frequency components which may decay rapidly with distance, however.

A further constraint on precursory strains and stresses, and thus electrical precursors, comes from the accumulating evidence that the strike-slip faults are weak and thus unable to accumulate much shear stress before failure. In classical fault theory, the amount of shear stress ($\tau$) needed to rupture a fault must exceed the coefficient of friction ($\mu$) for the fault times the effective normal stress:

$$\tau = \mu(\sigma_n - P), \tag{2}$$

where $\sigma_n$ is the normal stress and $P$ is the pore pressure (Jaeger and Cook, 1969). The maximum shear stress in a rock is generally observed at an angle of $\theta$ = 30–60°, and this angle is related to the coefficient of friction:

$$\tan(2\theta) = \frac{1}{\mu}. \tag{3}$$

Mount and Suppe (1992) showed that the maximum principal stresses in central California and in Sumatra are aligned almost perpendicular to the strike-slip faults (Figure 11). They argue that this is evidence for near-lithostatic fluid pressures ($P \approx \sigma_n$) or abnormally low coefficients of friction. Low coefficients of friction ($\mu$ = 0.1–0.2) are also required in models of the deformation along faults in California (Bird and Kong, 1994). Byerlee (1990, 1992) proposes a model whereby the apparent low coefficient of friction is explained with high fluid pressures along faults, and extends the model to rupture of discontinuous pockets of high pore pressure to produce episodic flow before earthquakes (Byerlee, 1993). Finally, Stein *et al.* (1992) showed using elastic models that a pattern of stress increases in a series of earthquakes leading up to the 1992 $M7.4$ Landers earthquake could explain the locations of the large earthquakes in the sequence. The surprising result was that the change in the Coulomb failure stress (a weighted combination of the increase in shear stress and the decrease in normal stress) was on the order of a few bars. If there is a causative relationship between the stress changes calculated by Stein *et al.* (1992) and earthquakes, then this is a strong argument for the crust being maintained at a stress level close to failure most of the time. The stress changes leading to earthquakes are then small, and precursory changes of physical properties and/or generation of EM signals may be negligible. A consequence of this model may be the requirement to monitor the fault zones directly.

The stress model of Stein *et al.* (1992) does provide a mechanism for focusing stress changes over distances in excess of 100 km, however. Changes in Coulomb failure stress at distances of 100 km are almost as large as near the epicenter. Such focusing may also explain the existence of the sensitive sites observed by Varotsos *et al.* (1993). Sensitive sites occur at the lobes of the maximum stress changes, although I emphasize again that the maximum stress change is only 1–2 bars. Such stress changes may trigger failure of overpressured fluid reservoirs, leading to electrokinetic generation of electric and magnetic fields. Fenoglio *et al.* (1993) have proposed this mechanism for the generation of the ULF signals prior to the Loma Prieta earthquake. Rupturing of small, isolated fluid reservoirs in the fault zone would lead to unsteady fluid flow and transient electromagnetic signals due to the fluid flow. Dobrolovsky *et al.* (1989) have suggested that the VAN signals are generated in a similar manner, but with a fluid reservoir close to the sensitive sites.

A number of other geophysical constraints must be considered when developing models of precursory electromagnetic signals. Precursory stress and strain changes are smaller than coseismic changes, and any model of the observed EM precursors

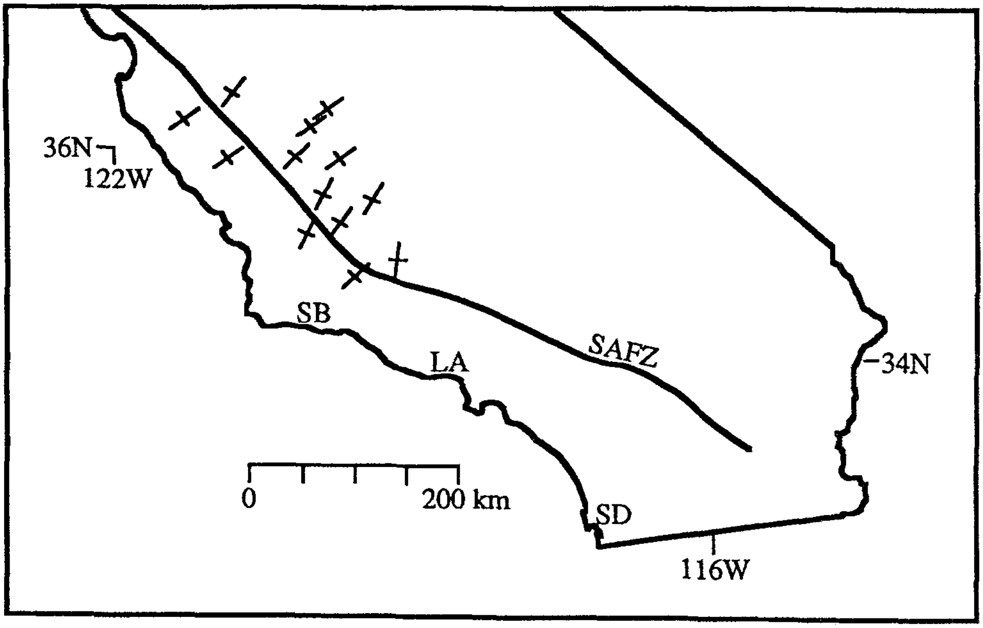


*Figure 11.* Orientation of maximum principal stress in central California (after Mount and Suppe (1992)). Note that it is almost perpendicular to the San Andreas fault (SAFZ). Symbols used are: SB, Santa Barbara; LA Los Angeles; and SD, San Diego.

must explain how to get a measurable signal before the earthquake and no coseismic signal. Faults appear to be weak surfaces with low apparent coefficients of friction. Thus, the amount of stress accumulation before failure must be small. This inference is corroborated by models of Coulomb failure stress in which it appears that stress changes of 1–2 bars are sufficient to trigger earthquakes over a period of years (Stein *et al.*, 1992). Finally, fluids probably have a key role in the generation of EM precursors as well as producing changes of resistivity prior to earthquakes.

## 7. Conclusions

Successful experiments to detect precursory variations of earth potential or electric fields have been conducted with multiple dipoles of varying length. Nontectonic causes for the fluctuation of the electrode potential require the use of multiple dipoles and a scheme for distinguishing between electrode noise and dipole signal. Such an experiment is being conducted in Greece. SES has been observed repeatedly on multiple dipoles and correlated with subsequent earthquake activity. Some statistical evaluations of these observations show that the method succeeds at prediction more often than random sampling of the earthquake catalog, while others do not. Future efforts should be directed at identifying the cause of the SES, as well as reconciling the statistical analyses. Ancillary measurements of water table fluctuations, magnetic field, DC resistivity, and strain should be made, as well as thoroughly characterizing the geoelectric setting of both the individual sites at a detailed scale and of Greece at a regional scale.

Continuing efforts to explain the seismomagnetic signals observed with the Loma Prieta earthquake are focusing on electrokinetic models, but much more work is needed here. Attempts to monitor electric and magnetic signals have not yet yielded useful results, with the exception of an experiment to monitor the magnetic field simultaneously with the SES. Modern techniques of magnetotelluric data acquisition and analysis must be employed in these experiments. Further work is also needed to better document seismoelectric emissions at radio frequencies as well as identifying the causative mechanism.

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## Appendix

The transmission coefficient defined by Rozluski and Yukutake (1993) is the ratio of the outgoing Poynting vector to the ingoing one (Equation 1). By using a simple 2-D model with the measurement axes coincident with the structural axes, this coefficient can be expressed in terms of more familiar MT parameters. For a 2-D earth,

$$\begin{vmatrix} E_x \\ E_y \end{vmatrix} = \begin{vmatrix} 0 & Z_{xy} \\ Z_{yx} & 0 \end{vmatrix} \begin{vmatrix} H_x \\ H_y \end{vmatrix}. \tag{A-1}$$

In terms of crosspower spectra between $E$ and $H$, the two impedance terms can be written as

$$Z_{xy} = \frac{\langle E_x H_y \rangle}{\langle H_y H_y \rangle}, \quad Z_{yx} = \frac{\langle E_y H_x \rangle}{\langle H_x H_x \rangle}. \tag{A-2}$$

The $\langle \ \rangle$ denotes an ensemble average over a frequency band of finite width, and the second field in the average is conjugated. The constitutive relation between the magnetic field and the magnetic induction is $B = \mu H$. The crosspower spectra in (1) in the main text can now be written in terms of impedances and the power spectra between the individual fields:

$$G(E_x + cB_y) = \langle E_x E_x \rangle + [c\mu(Z_{xy} + Z_{xy}^*) + c^2\mu^2]\langle H_y H_y \rangle, \tag{A-3}$$

$$G(E_y - cB_x) = \langle E_y E_y \rangle + [-c\mu(Z_{yx} + Z_{yx}^*) + c^2\mu^2]\langle H_x H_x \rangle, \tag{A-4}$$

$$Re(G_{xy}(E_x, B_y)) = \mu\, Re(Z_{xy}\langle H_y H_y \rangle), \tag{A-5}$$

and

$$Re(G_{yx}(E_y, B_x)) = \mu\, Re(Z_{yx}\langle H_x H_x \rangle), \tag{A-6}$$

where $c$ = speed of light. With the substitutions in (A-3) through (A-6), the transmission coefficient in (1) can be rewritten as

$$T = \frac{2}{1 + \dfrac{\langle E_x E_x \rangle + \langle E_y E_y \rangle + c^2\mu^2(\langle H_x H_x \rangle + \langle H_y H_y \rangle)}{2c\mu[\langle H_y H_y \rangle\, Re(Z_{xy}) - \langle H_x H_x \rangle\, Re(Z_{yx})]}}. \tag{A-7}$$

For a time dependence of $\exp(-j\omega t)$, the real part of $Z_{xy}$ is greater than zero and the real part of $Z_{yx}$ is less than zero. Therefore, all terms in (A-7) are positive and the transmission coefficient should never be less than zero. The analysis done for the MT impedance makes no assumptions about the direction in which the fields are propagating, so any upward propagating fields would be included in the derivation. Without further analysis of the errors involved in each of the terms in (A-7), it is difficult to explain why the transmission coefficient becomes negative. Noise in either the electric or magnetic fields is additive to the autopowers, so the noise cannot be the direct cause of negative transmission coefficients. Noise may affect the estimation of the impedances, however.

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