REVIEW OF ELECTRIC AND MAGNETIC FIELDS ACCOMPANYING SEISMIC AND VOLCANIC ACTIVITY

M. J. S. JOHNSTON
U.S. Geological Survey, Menlo Park, CA 94025, USA

Abstract. New observations of magnetic, electric and electromagnetic field variations, possibly related to recent volcanic and seismic events, have been obtained on Mt. Unzen in Japan, Reunion Island in Indian Ocean, the Long Valley volcanic caldera in California, and for faults in China and Russia, California and several other locations. For volcanic events, contributions from different physical processes can be identified during the various eruption stages. Slow processes (weeks to months) include near-surface thermal demagnetization effects, piezomagnetic effects, and effects from rotation/displacement of magnetized material. Rapid processes (seconds to days) include piezomagnetic effects from instantaneous stress redistribution with explosive eruptions and electrokinetic effects from rupture of high pressure fluid compartments commonly encountered in volcanic regions. For seismic events, the observed coseismic offsets are instantaneous, provided care has been taken to ensure sensors are insensitive to seismic shaking and are in regions of low magnetic field gradient. Simple piezomagnetic dislocation models based on geodetically and seismically determined fault parameters generally match the observed signals in size and sign. Electrokinetic effects resulting from rupture of fluid filled compartments at hydrostatic to lithostatic pore pressures can generate transient signals in the frequency band 100 Hz to 0.01 Hz. However, large-scale fluid driven processes are not evident in near-field measurements in the epicentral region minutes to weeks before large earthquakes. The subset of ionospheric disturbances generated by trapped atmospheric pressure waves (also termed gravity waves and/or acoustic waves, traveling ionospheric disturbances or TID's) that are excited by earthquakes and volcanic eruptions are common and propagate to great distances. These are known and expected consequences of earthquakes, volcanic explosions (and other atmospheric disturbances), that must be identified and their effects removed from VLF/ULF electromagnetic field records before associating new observations of ionospheric disturbances with earthquake activity.

Key words: earthquakes, volcanoes, electric fields, magnetic fields, mechanics, prediction

1. Introduction

Independent knowledge of the physical mechanisms driving seismic and volcanic activity can be obtained from observations of electric and magnetic fields generated by these complex processes. During the past few decades, we have seen a remarkable increase in the quality and quantity of electromagnetic data recorded before and during eruptions and earthquakes. In this paper we review the most significant recent data and the implications these data have for different generating mechanisms. We note that, despite several decades of relatively high quality monitoring, it is still not clear that precursory EM signals occur, although causal relations between magnetic field changes and earthquake stress drops or volcanic eruptions are no longer in question. Earlier reviews (Johnston, 1989; Park et al., 1993) and special journal issues (Johnston and Parrot, 1989; Parrot and Johnston, 1993)

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and books (Hayakawa and Fujinawa, 1994) cover various aspects of tectonomagnetism, volcanomagnetism and tectonoelectricity and concentrate on various parts of the electromagnetic spectrum from radio frequencies (RF) to sub-microhertz frequencies. An impressive improvement in sensors, sensor reliability, data collection techniques, analysis, and international cooperation is apparent during the past few years. As these techniques and instruments have improved, the amplitudes of signals observed with earthquakes (Figure 1a) have decreased with time but the data have become very convincing. The data during volcanic eruptions (Figure 1b) are also less ambiguous. More work needs to be done however, to convincingly demonstrate to the international geophysical community, causality, or lack thereof, between EM signals and earthquakes and volcanic eruptions. In particular, further improvement is needed in areas such as use of multiple detectors, use of reference sites, noise identification/reduction procedures, and demonstration of consistency with other geophysical data that independently reflect the state of stress, strain, material properties, fluid content, and approach to failure of the earth's crust in volcanic and seismically active regions.

We review here recent results of magnetic, electric and electromagnetic disturbances apparently associated with earthquakes and volcanic eruptions together with the physical mechanisms likely to have produced them. While some of these observations are larger than expected, the best field observations are in general agreement with calculations. Some are suggested as precursors yet have no corresponding co-event signals and some have co-event signals yet no precursory signals. More field observations are clearly needed.

2. Summary of Physical Mechanisms Involved

The loading and rupture of water-saturated crustal rocks during earthquakes and volcanoes, together with fluid/gas movement, stress redistribution, change in material properties, has long been expected to generate associated magnetic and electric field perturbations. The detection of related perturbations prior to fault rupture or volcanic eruptions has thus often been proposed as a simple and inexpensive method for monitoring the state of crustal stress and perhaps providing tools for predicting crustal failure. (Wilson, 1922; Kalashnikov, 1954; Stacey, 1964; Stacey et al., 1965; Yamazaki, 1965; Brace and Orange, 1968a, 1968b; Nagata, 1969; Barsukov, 1972; Rikitake, 1968, 1976; Honkura, 1976; Fitterman, 1979, 1981; Ishido and Mizutani, 1981; Varotsos and Alexopoulos, 1987; Dobrovolsky et al., 1989; Sasai, 1980, 1991a, 1991b; Park, 1991; Fujinawa et al., 1992; Fenoglio et al., 1995; Utada, 1993). The primary mechanisms for generation of electric and magnetic effects with crustal deformation and earthquakes effects include piezomagnetism, stress/conductivity, electrokinetic effects, charge generation and dispersion, and magnetohydrodynamic effects. Discussion of the different mechanisms will be roughly in order of the degree of attention that has been accorded them.

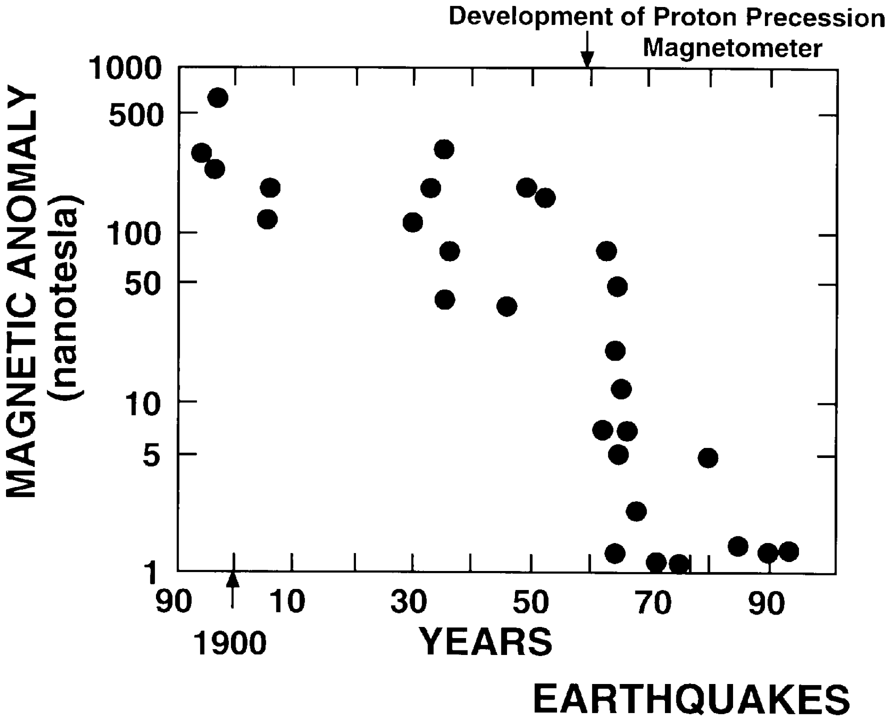

Figure 1a. Reported tectonomagnetic anomalies as a function of time. "P" indicates the time at which absolute proton precession magnetometers and noise reduction techniques were introduced (updated from Rikitake, 1968).

2.1. Piezomagnetism

The magnetic properties of rocks have been shown under laboratory conditions to depend on the state of applied stress (Wilson, 1922; Kalashnikov and Kapitsa, 1952; Kapitsa, 1955; Ohnaka and Kinoshita, 1968; Kean et al., 1976; Revol et al., 1977; Martin, 1980; Pike et al., 1981) and theoretical models have been developed in terms of single domain and pseudo-single domain rotation (Stacey, 1962; Nagata, 1969; Stacey and Johnston, 1972) and multi-domain wall translation (Kern, 1961; Kean et al., 1976; Revol et al., 1977). The fractional change in magnetization per unit volume as a function of stress, can be expressed in the form;

$$\Delta \mathbf{I} \approx K\sigma \cdot \mathbf{I} \tag{1}$$

where $\Delta \mathbf{I}$ is the change in magnetization in a body with net magnetization $\mathbf{I}$ due to a deviatoric stress σ. K, the stress sensitivity, typically has values of about $3^{*}10^{-3}$ MPa^{-1}. The stress sensitivity of induced and remanent magnetization from theoretical and experimental studies has been combined with stress estimates from dislocation models of fault rupture and elastic pressure loading in active volcanoes

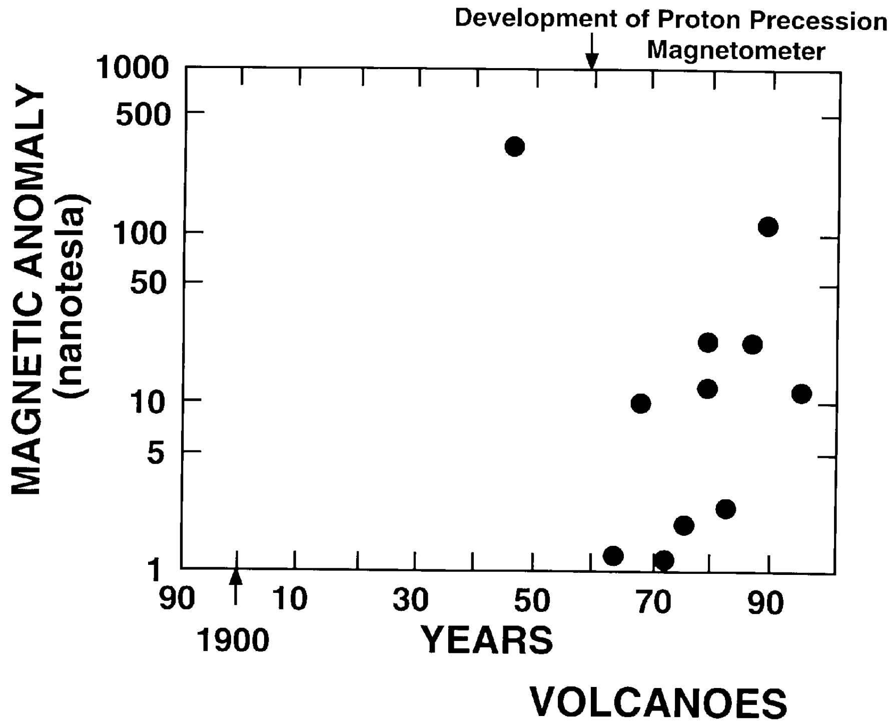

Figure 1b. Reported volcanomagnetic anomalies as a function of time. "P" indicates the time at which absolute proton precession magnetometers and noise reduction techniques were introduced.

to calculate magnetic field changes expected to accompany earthquakes and volcanoes (Stacey, 1964; Stacey et al., 1965; Shamsi and Stacey, 1969; Johnston, 1978; Davis et al., 1979; Sasai, 1980, 1983, 1991a, 1991b; Davis et al., 1984; Johnston et al., 1994; Banks et al., 1991). The surface fields ($\Delta \mathbf{B}_P$) at a point, P, can be calculated by either integrating the change in magnetization $\Delta \mathbf{I}_Q$ in a unit volume, $\mathrm{d}v$, at a point Q where the stress is σ_{ij}, and r is the distance between P and Q, according to,

$$\Delta \mathbf{B}_P = -\frac{\mu}{4\pi} \nabla \int_V \Delta \mathbf{I}_{\mathbf{Q}} \cdot \frac{\mathbf{r}}{r^3} \mathrm{d}v, \tag{2}$$

as originally done by Stacey (1964), or by using a simpler method pioneered by Sasai (1980, 1994) in which analytic expressions of the surface piezomagnetic potential, W, produced by a known stress distribution in a magnetoelastic half-space are obtained by transforming the stress matrix and integrating over the magnetized region. The surface field can be found from

$$\Delta \mathbf{B}_P = -\nabla W. \tag{3}$$

These models show that magnetic anomalies of a few nanoteslas (nT) should be expected to accompany earthquakes and volcanic eruptions for rock magnetizations and stress sensitivities of 1 Ampere/meter (A/m) and 10^{-3} MPa^{-1}, respectively. As shown below, these signals are readily observed with the correct sign and amplitude.

2.2. Stress/resistivity and strain/resistivity effects

In a like manner, the stress dependence of electrical resistivity of rocks has been demonstrated in the laboratory. Resistivity in low porosity crystalline rock increases with compression as a result of crack closure at about 0.2%/bar (Brace et al., 1965) and decreases with shear due to crack opening at about 0.1%/bar (Yamazaki, 1965; Brace and Orange, 1968a,b; Brace, 1975). More porous rocks have even lower stress sensitivity. The situation is further complicated by the fact that non-linear strain can also produce resistivity changes (Lockner and Byerlee, 1986). A stress/resistivity relation equivalent to Equation (1) above has the form

$$\frac{\Delta\rho}{\rho} \approx K_r\sigma \tag{4}$$

for homogeneous material, where ρ is resistivity, K_r is a constant, and σ is the stress. Unfortunately, the earth is not homogeneous and many factors including rock type, crack distribution, degree of saturation, porosity, strain level, etc., can localize or attenuate current flow. Nevertheless, this equation provides a starting point for calculating resistivity changes near active faults. Measurements of resistivity change are being made with both active experiments (where low frequency currents are injected into the ground and potential differences, V, are measured on receiver dipoles), or passive telluric and magnetotelluric (MT) experiments where changes in resistivity are inferred from changes in telluric or MT transfer functions. These transfer functions are given by:

$$Z(\omega) = \frac{E(\omega)}{H(\omega)}, \tag{5}$$

where ω is angular frequency, $E(\omega)$ and $H(\omega)$ are observed electric and magnetic fields. For active experiments (Park et al., 1993),

$$\frac{\delta\rho}{\rho} = G\frac{\delta V}{V}, \tag{6}$$

where δV is the change in potential difference and G is a constant. For MT experiments,

$$\delta\rho = \frac{\delta|Z(\omega)|^2}{\omega\mu}. \tag{7}$$

Based on the field observations of stress changes accompanying earthquakes ($\approx$ 1 MPa), resistivity changes of at least one percent might be expected to accompany crustal failure. Field experiments for detection of resistivity changes thus need to have a measurement precision of better than 0.1% (Fitterman and Madden, 1977; Park, 1991; Park et al., 1993). This may be difficult with MT measurements unless remote magnetic field reference measurements are used (Gamble et al., 1979) although measurement precision for telluric electric fields can be made at the 0.1% level (Madden et al., 1992).

2.3. Electrokinetic effects

The role of active fluid flow in the earth's crust as a result of either fault failure or volcanic activity can generate electric and magnetic fields (Mizutani et al., 1976; Fitterman, 1978, 1979; Ishido and Mizutani, 1981; Dobrovolsky et al., 1989; Fenoglio et al., 1995). Electrokinetic electric and magnetic fields result from fluid flow through the crust in the presence of an electric double layer at the solid-liquid interfaces. This double layer consists of ions anchored to the solid phase, with equivalent ionic charge of opposite sign distributed in the liquid phase near the interface. Fluid flow in this system transports the ions in the fluid in the direction of flow, and electric currents result. Conservation of mass arguments (Fenoglio et al., 1995) supported by surface strain observations (Johnston et al., 1987) constrain this process to be limited in extent and transient in nature since large-scale fluid flow cannot continue for very long before generating surface deformation.

The current density **j** and fluid flow **v** are found from coupled equations (Nourbehecht, 1963; Fitterman, 1979) given by

$$\mathbf{j} = -s\nabla E - \frac{\xi\zeta\nabla P}{\eta}, \tag{8}$$

where E is streaming potential, s is the electrical conductivity of the fluid, ξ is the dielectric constant of water, η is fluid viscosity, ζ is the zeta potential, E is electric field potential, and P is pore pressure. The flow rate of water **v**, is given by

$$\mathbf{v} = \frac{\phi\xi\zeta\nabla E}{\eta} - \frac{\kappa\nabla P}{\eta}, \tag{9}$$

where κ is the permeability.

The current density in equation (8) has two components. The second term represents electric current resulting from mechanical energy being applied to the system and is sometimes called the "impressed" current (Williamson and Kaufman, 1981). This term describes current generated by fluid flow in fractures. The first term of Equation (8) represents "back" currents resulting from the electric field generated by fluid flow. The distribution of electrical conductivity determines the net far-field magnetic and electric fields resulting from these effects. In an extreme case, if the fluid is extremely conducting and the surrounding region is not, current flow in the fluid cancels the potential generated by fluid flow (Ahmad, 1964). At the other extreme, if the fluid is poorly conducting, "back" currents, usually termed "volume currents" (Williams and Kaufman, 1981) flow in the surrounding region. If the region were homogeneous, magnetic fields would be generated by impressed currents only since the volume currents generate no net field (Fenoglio et al., 1995; Fitterman, 1979).

The situation for finite flow in limited fault fractures or limited volcanic dykes more closely approximates the second case where the surface magnetic field is approximately given by

$$\mathbf{B} = \frac{\mu_0}{4\pi} \int_A \frac{\mathbf{j}_i \times \mathbf{r}}{r^2} \mathrm{d}A. \tag{10}$$

Note that, the physics describing the electric and magnetic fields generated in the human body as blood is pumped through in arteries provides a very good analog to those generated in fault zones (Williamson and Kaufman, 1981), since the electrical conductivities of bone (0.001 S/m), muscle (0.1 S/m) and blood (1 S/m) and likely velocities are similar to those of rock, fault gouge and fault zone fluids, Reasonable fault models, in which fluid flows into a 200 m long rupturing fracture at a depth of 17 km, indicate that transient surface electric fields of several tens of millivolts/km and transient magnetic fields of a few nT can be generated (Fenoglio et al., 1995).

2.4. Charge generation processes

Numerous charge generation mechanisms have been suggested as potential current sources for electric and magnetic fields before and during earthquakes and volcanic eruptions. These mechanisms include piezoelectric effects (Finkelstein et al., 1973; Baird and Kennan, 1985), rock shearing/triboelectricity (Lowell and Rose-Innes, 1980; Gokhberg et al., 1982; Brady, 1992), fluid disruption/vaporization (Chalmers, 1976; Matteson, 1971; Blanchard, 1964), solid state mechanisms (Dologlou-Revelioti and Varotsos, 1986; Freund et al., 1992). Each of these mechanisms has a solid physical basis with support by laboratory experiments on dry rocks in insulating environments or single crystals of dry quartz and each is capable of producing substantial charge under the right conditions. The problems regarding demonstration of their applicability to EM field generation in the earth concern, firstly, the amplitude of these effects in wet rocks at temperatures and pressures expected in the earth's crust, and, secondly, how to maintain charge for any appreciable length of time in the conducting crust.

Regarding the first problem, experiments clearly need to be done for each mechanism to quantify the effects expected in wet rocks at temperatures of at least 100 °C and at pressures of 100 MPa expected at earthquake hypocenters. Experiments on dry rocks at atmospheric pressure are not very relevant to this issue. Piezoelectric effects in dry quartz bearing rocks are less than 0.1% of those observed for single crystals of quartz due to self cancelling effects (Tuck et al., 1977), and effects in wet rocks would be expected to be smaller still and transient at best. EM generation by fracturing dry rocks (Warwick et al., 1982; Brady,

1992) needs to be extended to wet rocks under confining pressure. Experiments on hole transport of O^- in dry rocks (Freund et al., 1992) and stress charging of dry non-piezoelectric rocks (Dologlou-Revelioti and Varotsos, 1986) need also to be repeated with wet rocks under confining pressure so that these effects can be quantified. Brady (1992) observed no EM emission during fracture of conductive rocks since the conductor could not maintain charge separation.

The second problem concerns the discharge time for these processes and just how far EM signals generated by them might propagate. The charge relaxation time τ for electrostatic processes is given by the product of permittivity (ϵ) and resistivity (ρ). ϵ is 0.5–1.0 $\times$ 10^{-10} F/m for crustal rocks. If, $\rho \approx 10^3$ ohm m (typical upper value for near fault crustal rock) then,

$$\tau \approx 10^{-6}\text{sec.} \tag{11}$$

While polarization effects (Lockner and Byerlee, 1985) may generate somewhat longer timescales, EM signal generation by charge generation processes must necessarily be very rapid unless mechanisms can be found for isolating and maintaining large charge densities in a conducting earth. Furthermore, dispersion precludes EM fields propagating far in a conducting earth (Honkura and Kuwata, 1993).

Attenuation of the magnetic field, B, of a plane electromagnetic wave generated at depth, d km, by charge generation/cancellation processes as a function of penetration distance through a conductive medium is given by:

$$B = B_0 e^{-\gamma z} \tag{12}$$

where B_0 is the initial field strength, z is the penetration distance into the medium, and γ is the complex propagation coefficient given by:

$$\gamma = \sqrt{\omega^2 \mu\epsilon + j\omega\mu s}, \tag{13}$$

where ω is the angular frequency of the radiation, μ is the magnetic permeability of the earth, ϵ is the permittivity, and s is the conductivity of the medium.

If s is 0.1 S/m, the frequency is 0.01 Hz, the “skin depth” is 10 km, so fields generated at this depth could be observable at the earth’s surface.

If these fields are generated by rock cracking and fracturing, then acoustic (seismic) signals should also be generated (see Lockner et al., 1991). Seismic wave attenuation with distance z has the form

$$A(z) = A_o \exp^{-\omega z/2cQ}, \tag{14}$$

where ω is the angular frequency, c is the phase velocity, and Q is the quality factor. Taking observed values of 3 km s^{-1} and 30 for c and Q, it can easily be shown that seismic waves in the frequency band 1 Hz to 0.01 Hz are not attenuated significantly at all instruments in the epicentral area. At higher frequencies, both seismic and EM signals are heavily attenuated. For example at 10 Hz the EM "skin depth" is 493 m in material with conductivity of 0.1 S m^{-1} while the seismic equivalent "penetration depth" is 2864 m. At 100 Hz the comparative depths are 156 m, 286 m and at 1 KHz 29 m, 49 m, respectively.

So, both high frequency seismic and EM waves are heavily attenuated in the earth's crust. EM sources at 10 Hz should have an acoustic component that is more easily detected over a greater area. In fact, for all EM sources at seismogenic depths capable of propagating to the earth's surface (i.e., with frequencies less than 0.1 Hz), acoustic/seismic consequences of these sources propagate more effectively to the surface and might be used to verify their existence.

2.5. Thermal remagnetization and demagnetization

Crustal rocks lose their magnetization when temperatures exceed the Curie Point ($\approx$ 580 °C) and become remagnetized again as the temperature drops below this value (see Stacey and Banerjee (1974), for a more complete description of this process). In crustal rocks at seismogenic depths near active faults, this process is unlikely to contribute to rapid changes in local magnetic fields since the thermal diffusivity of rock is typically about 10^{-6} m^2 s^{-1} and migration of the Curie Point isotherm by conduction cannot be as much as a meter in a year (Stacey, 1992). At shallow depths in volcanic regions, particularly in recently emplaced extrusions and intrusions, thermal cracking with gas and fluid movement can transport heat rapidly and large local anomalies can be quickly generated (Rikitake and Yokoyama, 1955; Hurst and Christoffel, 1973; Emeleus, 1977; Zlotnicki and Le Mouel, 1988; Hamano et al., 1990; Dzurisin et al., 1990; Zlotnicki et al., 1990; Tanaka, 1993, 1995). These anomalies can be modeled as a magnetized slab in a half-space. Good examples of magnetic modeling of anomalies generated by cooling of extrusions can be found in Dzurisin et al. (1990) for Mount St. Helens and in Tanaka (1995) for Mt Unzen in Japan.

2.6. Magnetohydrodynamic (MHD) effects

The induced magnetic field $\mathbf{B}_i$ generated by the motion $\mathbf{v}$ of a fluid with conductivity s in a magnetic field $\mathbf{B}_o$, is governed by the equation:

$$\frac{\partial \mathbf{B}}{\partial t} = \nabla \times \mathbf{v} \times \mathbf{B} + \frac{\nabla^2 \mathbf{B}}{\mu_o s} + \frac{\nabla s \times \nabla \times \mathbf{B}}{\mu_o s^2}, \tag{15}$$

where μ_o is the permeability in a vacuum (Shercliff, 1965). For low magnetic fields and low electrical conductivities in the earth's crust where the fluid motion is not affected by the induced fields, the induced field is given approximately by the product of the magnetic Reynolds number R_m and the imposed field $\mathbf{B}_o$, i.e.

$$B_i \approx R_m \times B_o \approx \mu s v d B_o, \tag{16}$$

where d is the length scale of the flow. Critical parameters here are the likely flow velocities and fluid electrical conductivities in the crust. Flow velocity is determined by rock permeability and fluid pressure gradients according to Darcy's Law. Permeability of fractured rock is not less than 10^{-12} m^2 (Brace, 1980) and pore pressure gradients cannot exceed the lithospheric gradient. It is difficult to achieve widespread flow velocities of even a few millimeters/sec with this mechanism. Furthermore, fluid conductivities are unlikely to exceed that of seawater ($\approx$1 S m^{-1}). Using these numbers, fluid flow in fractured fault zones at seismogenic depths ($\approx$ 5 km) with a length scale of 1 km could generated transient fields of about 0.01 nT. This is far too small to be observed at the earth's surface. As a check on these calculations, we note that fields of a few nT are observed with waves in the ocean where the conductivity is 1 S m^{-1} and wave velocities exceed 100 cm/sec (Fraser, 1966).

3. Experimental design and measurement precision

The precision of local magnetic and electric field measurements on active faults and volcanoes varies as a function of frequency, spatial scale, instrument type, and site location. Most measurement systems on the earth's surface are limited more by noise generated by the ionosphere, the magnetosphere, and by cultural noise than by instrumental noise. Thus, systems for quantifying these noise source are of crucial importance if these measurements are to be generally accepted in the geophysical community. At the lower frequencies (microhertz to hertz) for both electric and magnetic field measurements, the use of reference sites to isolate external noise is common (Rikitake, 1966; Johnston et al., 1984; Park, 1991; Park and Fitterman, 1990; Varotsos and Lazaridou, 1991). On spatial scales comparable to moderate magnitude earthquake sources (tens of kilometers) for typical sites on active faults far from cultural noise sources, electric and magnetic noise power has a $1/f^2$ or "red" spectrum (Johnston et al., 1984). Against this background, magnetic fields can be measured to several nanotesla over months, to 1 nanotesla over days, to 0.1 nanotesla over minutes, and 0.01 nanotesla over seconds (Ware et al., 1985). Comparable electric field noise limits are 10 mv/km over months, several mv/km over days, 1 mv/km over minutes and 0.1 mv/km over seconds (derived from Park, 1991). EM noise increases approximately linearly with site separation (Johnston et al., 1984).

Fewer measurements of the scale and temporal variation of noise and signal sources exist for electromagnetic field measurements at higher frequencies. These measurements are essential if we are to fully understand the generating mechanisms and to set up tests to verify that EM systems are truly measuring signals generated in the earth. Some recent reports of EM signals that perhaps preceded earthquakes are questionable because they were obtained using single detectors and the possibility that they result from spurious noise cannot be precluded.

Unambiguous observation of EM crustal fields thus requires discrimination against disturbances in the ionosphere, the magnetosphere, and from cultural noise, a careful experimental design and well-determined noise characterization (measurement precision). Techniques for further noise reduction such as adaptive filtering (Davis et al., 1981; Davis and Johnston, 1983), use of multiple variable-length sensors in the same and nearby locations (Mori et al., 1993; Varotsos and Alexopoulos, 1987), use of reference sites with synchronized data sampling and with site spacing at least comparable to the source size (Rikitake, 1966; Johnston et al., 1984; Park, 1991). The latter technique allows discrimination of local sources and as much as a 30 dB reduction in noise (Johnston et al. 1984; Park, 1991).

4. Recent Results

While both electric and magnetic fields are expected to accompany dynamic physical processes in the earth's crust, simultaneous measurements of both fields are not routinely made. We will therefore discuss electric fields, magnetic fields, and electromagnetic fields separately during, and preceding, earthquakes and volcanic eruptions. Magnetic and electric fields generated by earthquakes and volcanic eruptions are termed "seismomagnetic (SM)", "seismoelectric (SE)", "volcanomagnetic (VM)" and "volcanoelectric (VE)" effects. Those preceding earthquakes and eruptions, or occurring at other times, are termed "tectonomagnetic (TM)" and "tectonoelectric (TE)" effects.

Clearly, since primary energy release occurs at the time of earthquakes and eruptions, reliable magnetic and electric field observations at these times (i.e., those unaffected by seismic shaking) should relate to the seismic or volcanic source and should be readily observable if these observations are source related. In fact, co-event observations provide a calibration of stress sensitivity since the stress redistribution and source geometry of earthquakes and volcanic sources are often well determined (Aki and Richards, 1980). With this calibration, tectonomagnetic and tectonoelectric effects can be quantified and spurious effects identified.

The following list of examples will be restricted to the strongest data. That is, data recorded independently on more than one instrument, data that are independently supported by other stable geophysical measurement systems, and data for which noise levels have been quantified. Reported measurements made with single instruments or time histories of measurements showing data only for a short period before earthquakes with some "precursive" feature but no coseismic signals, are generally suspect and will not be included here.

4.1. Seismomagnetic effects

Surprisingly, while a number of magnetic field networks have been in operation around the world, only a few observations of coseismic magnetic field changes have been made in the near-field of large earthquakes during the past five years.

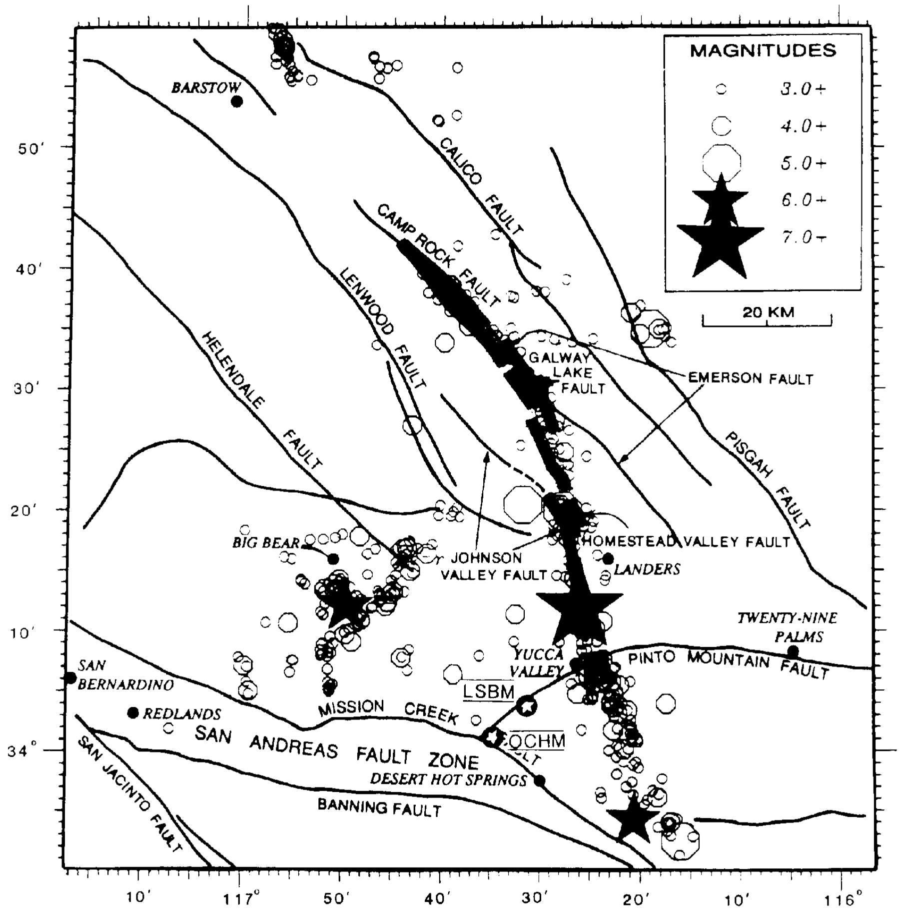

Figure 2. Locations of magnetometers (stars inside circles) relative to the epicenter (largest star) of the June 28, 1992 M7.3 Landers earthquake. (from Johnston et al., 1994).

4.1.1. *M7.3 Landers earthquake (June 28, 1992)*

An important set of observations of SM effects were made during the July 28, 1992 Landers earthquake (Johnston et al., 1994). This earthquake had a moment of $1.1 * 10^{27}$ dyne-cm and a magnitude of 7.3. Two total field proton magnetometers were in operation at distances of 17.1 km and 24.2 km from the earthquake and had synchronously sampled local magnetic fields every 10 minutes since early 1979 using satellite digital telemetry (Mueller et al., 1981). The locations are shown in Figure 2.

The local magnetic field at the magnetometer closest to the earthquake decreased by 1.2 nT while that at the second, 24.2 km from the epicenter, decreased by 0.7 nT. These values are consistent with a simple SM model of the earthquake in which the fault geometry and slip used are derived from geodetic and seismic inversions of the earthquake (Johnston et al., 1994). Figure 3 (upper) shows the differences between

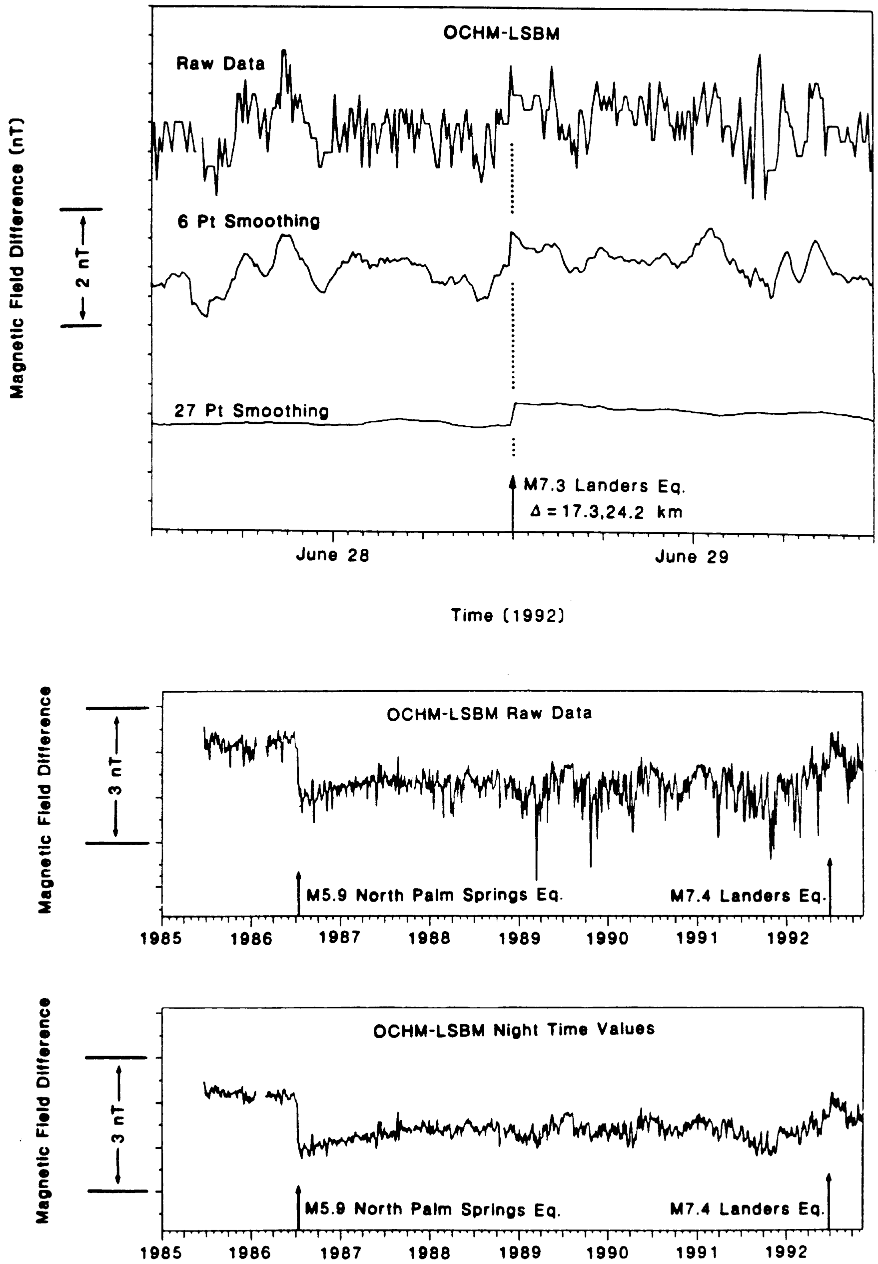

Figure 3. (Upper) Magnetic field differences between OCHM and LSBM on the day before and after the Landers earthquake. (Lower) Similar magnetic field differences from 1985 through 1992 showing the occurrence times of the July, 1986 M6 North Palm Springs earthquake and the June, 1992 Landers earthquake (from Johnston, 1992).

data obtained at the two sites OCHM and LSBM for the period 1 day before and after the earthquake. Note that there is no indication of diffusion like character in the magnetic field offsets that might indicate these effects were generated by fluid flow, nor are there any indications of enhanced low-frequency magnetic noise preceding the earthquake or indications of changing magnetic fields outside the noise in the hours to days before the earthquake. The lower plot shows the longer term data for the previous 7 years. The SM effect from the 1986 M6 North Palm Springs which occurred beneath these same two instruments is clearly evident (Johnston and Mueller, 1987). Similar coseismic results were found for the 1989 M7.1 Loma Prieta earthquake (Mueller et al., 1990).

4.2. Seismoelectric effects

Observations of seismoelectric effects which show expected scaling with both earthquake moment release and inverse distance cubed are difficult to make because of the sensitivity of electrode contact potential to earthquake shaking. Earlier work by Yamazaki (1974) showed clear correspondence between local coseismic strain steps and coseismic resistivity steps obtained using a Wenner array with a measurement precision of 0.01%. Strain measurements obtained with deep borehole strainmeters (Johnston et al., 1987) or with geodetic techniques (e.g. Lisowski et al, 1990) do reflect the strain/stress expected from seismically determined models of earthquakes. SE effects should be expected to do likewise. Unfortunately, the effects of shaking on contact potential and on self-potential as a result of changes in fluid content and fluid chemistry, conspire against making these data reliable reflectors of earthquake stress changes (Ozima et al., 1989). Clearly, co-located strain measurements (at tidal sensitivity) and electric field measurements are needed to demonstrate the sensitivity of the electric field measurements to changes in stress/strain in the earth's crust. Until this is done, it is difficult to have real confidence in measurements reported as earthquake precursors. Observations on multiple sensors with stable electrodes (Petiau and Dupis, 1980) are necessary prerequisites before the geophysical community will accept the reality of these data. Miyakoshi et al. (1994) provide some good examples of SE signals obtained with a multi-channeled system at the times of local earthquakes (see Figure 4), but it is not clear how these changes relate to the earthquake source.

Indirect observations of possible SE effects are obtained using the magnetotelluric (MT) technique to monitor apparent resistivity in seismically active regions. Even with the best designed systems (Gamble et al., 1978) using remote referencing systems to reduce noise and obtain stable impedence tensors, it is difficult to reduce errors below 5% for good soundings and 10–40% for poor soundings (Ernst et al., 1993). Resistivity changes associated with earthquakes are expected, and are observed, to be only a few percent at best. Thus, it is unlikely that this technique will be used generally for detection of resistivity changes. The pioneering work of Honkura et al. (1976) still largely defines the limits of observability for MT

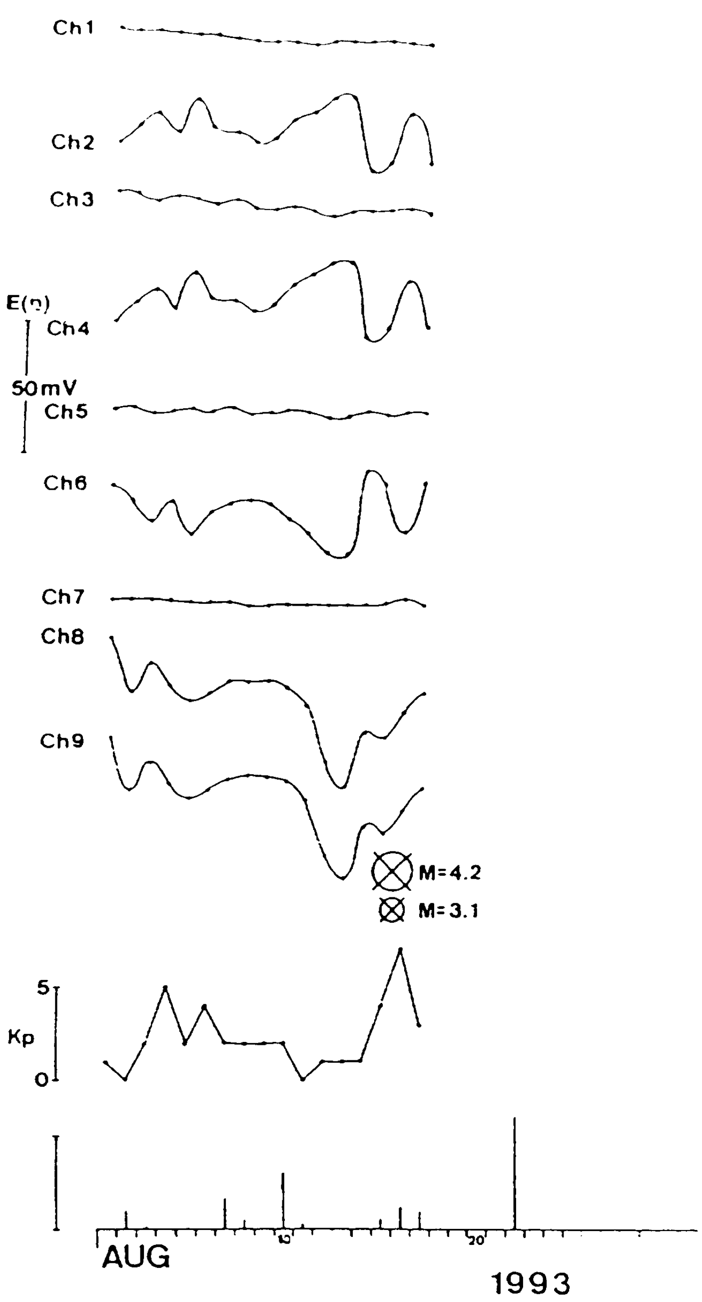

Figure 4. Plots of night-time electric potential difference in August, 1993, from an array of electrodes at Wakayama City during a M4.2 earthquake. Rainfall is shown on the lower plot. (from Miyakoshi et al., 1994).

observations, though some interesting ways of using MT to detect EM emissions with earthquakes have been explored by Rozluski and Yukutake (1993).

4.3. Tectonomagnetic and possible precursory effects

During the past 5 years, few new indications of convincing longer-term tectonomagnetic events (i.e., durations greater than minutes to weeks) are apparent in multiple near-field magnetometer records obtained along active faults as a result perhaps of strain redistribution prior to moderate/large earthquakes (see, for example, Figure 3 (Lower) above). These signals either are rare or do not generally have amplitudes greater than a nanotesla or so. If signal amplitudes are less than a nanotesla, it is unlikely that any reported precursive signals at great distances from these earthquakes are truly earthquake related. Long term relatively uniform changes, apparently related to crustal loading have been previously reported (Johnston, 1989; Oshiman et al., 1983).

The situation may be different at higher frequencies (i.e., $>$ 0.01 Hz). Recent efforts have been concentrated at these higher ULF frequencies as a result, primarily, of fortuitous observation of elevated ULF noise power near the epicenter of the M7.1 Loma Prieta earthquake of October 18, 1989 (Fraser-Smith et al., 1990). The magnetometer was located only 7 km from the epicenter (Figure 5) and recorded increased ULF noise reaching 1.5 nT in amplitude during the month before, two weeks before, and a few hours before the earthquake (Figure 6). A similar system recording ULF/VLF data 52 km from the earthquake showed no similar changes. Nor were similar records obtained during the M6.7 Northridge earthquake at a distance of 81 km from the epicenter (Fraser-Smith et al., 1994) or the M7.3 Landers earthquake.

For the Loma Prieta observations, Fenoglio et al. (1993) have shown that there is no correlation between the ULF anomalies and either the magnitudes or the rates of aftershocks. The absence of signals for these smaller magnitude earthquakes, or for larger earthquakes at greater distances, indicates a localized source for these oscillatory signals. Draganov et al. (1991) have suggested a magnetohydrodynamic origin, whereas Fenoglio et al. (1995) have suggested an electrokinetic source generated by fluid flow following rupture of high-pressure fluid filled inclusions in the fault zone.

4.4. Tectonoelectric and possible precursory effects

Interest in tectonoelectric (TE) phenomena related to earthquakes has experienced a resurgence of interest during the past five years, primarily as a result of suggestions in Greece and Japan that short-term geoelectric field transients (SES) of particular form and character precede earthquakes with magnitudes greater than 5 at distances up to several hundreds of kilometers (Varotsos et al., 1993a; Varotsos et al., 1993b; Nagao et al., 1996). These transients are recorded on multiple dipoles

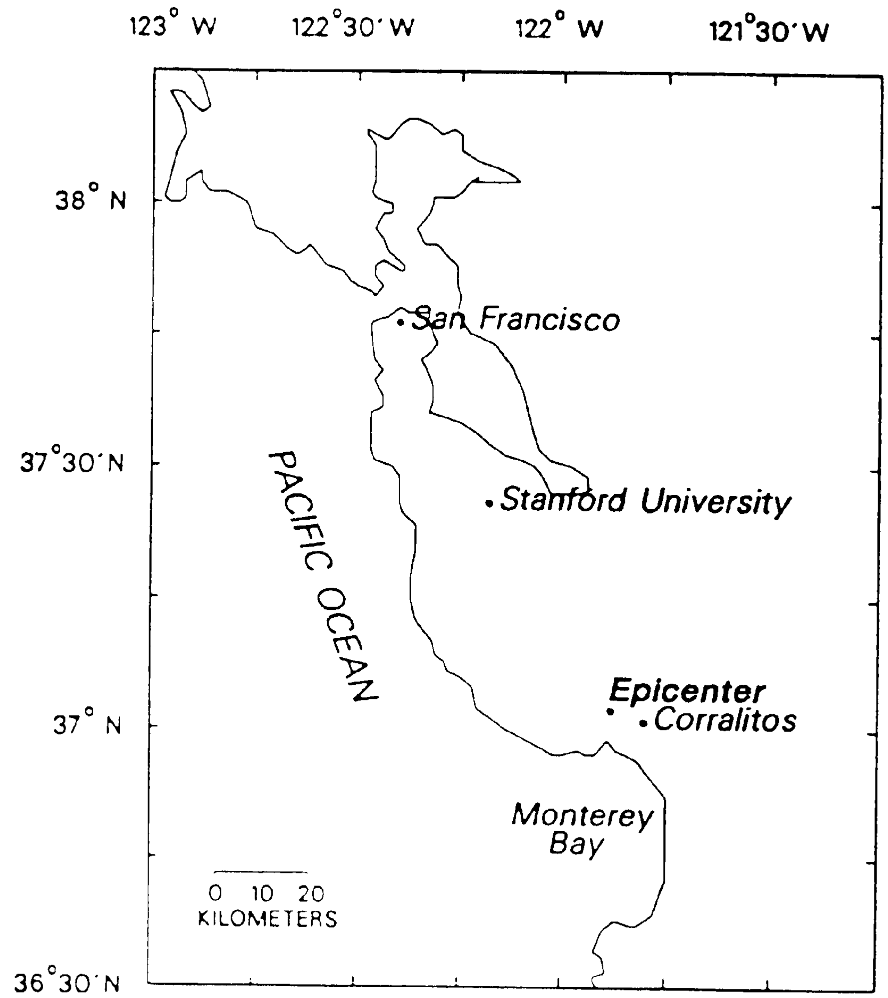

Figure 5. Location of ULF receiver at Corralitos, 7 km from the epicenter of the October 18, 1989 M7.1 Loma Prieta earthquake. A second receiver was located at Stanford University (from Fraser-Smith et al., 1990).

with different lengths (10–200 m for short arrays and 1–3 km for longer ones) with signal amplitudes of 20 mv/km and durations of several minutes. The observation of consistent electric field amplitudes independent of dipole length indicates a spatially uniform source field. There are no corresponding magnetic field transients and no apparent coseismic effects. The SES are empirically associated with subsequent earthquakes in "sensitive" areas. An example of an SES recorded on multiple orthogonal dipoles together with parallel recordings of magnetic field rate at Ioannina in north-west Greece on April 18, 1995, is shown in Figure 7 (from Varotsos et al., 1996). This SES was suggested to have preceded a M6.6 on May

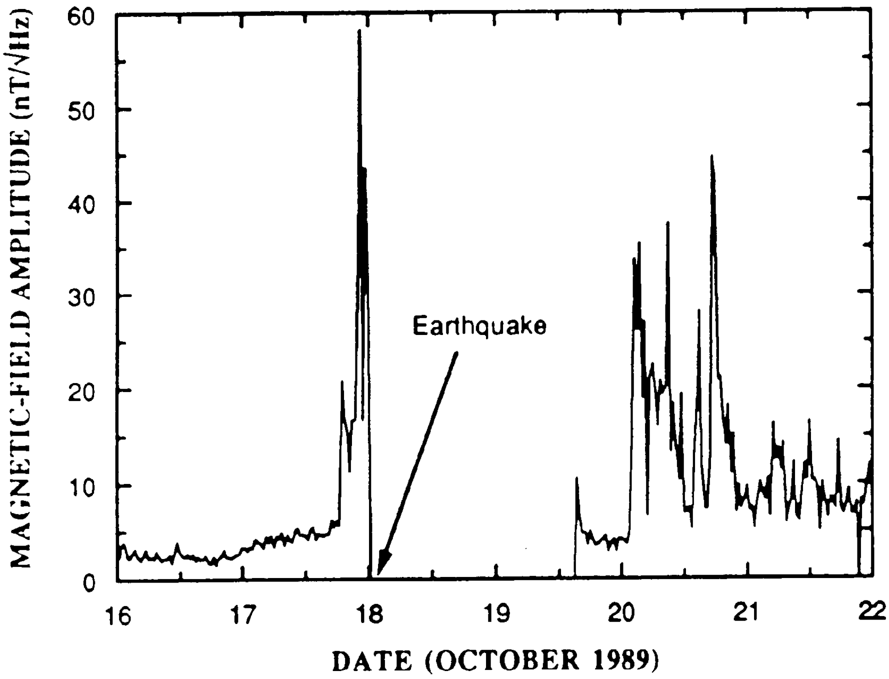

Figure 6. Magnetic field amplitude as a function of time during the 2 days before and 4 days after the Loma Prieta earthquake (from Fraser-Smith et al., 1990).

13, 1995, some 83 km to the north-east. Two other large earthquakes, the M6.6 on May 4 at Chalkidiki and the June 15 M6.5 Eratini earthquake, also were suggested to have been predicted by SES on distant stations several weeks before (Varotsos et al., 1996). Similar experiments have been run in Japan, France and Italy with various levels of claimed success. Nagao et al. (1996) suggest that anomalous SES may have been recorded prior to the M7.8 Hokkaido earthquake in June 1993.

Careful study of the SES recordings indicates that the SES signals, whatever their cause, do appear to have been generated in the earth's crust at the observation sites. It is not at all clear how these signals relate to earthquakes occurring sometimes hundreds of kilometers away (Bernard, 1992) while sites closer to the earthquake do not record SES's. Without a clear causal relation, demonstration of statistical significance is controversial (Mulargia and Gasperini, 1992; Hamada, 1993; Shnirman et al., 1993; Aceves et al., 1996; Varotsos et al., 1996). Better physical understanding is certainly needed. This could be obtained by careful study of the electrical conductivity structure around sites where SES's are recorded, measurement of high precision crustal strain, fluid levels in wells, and pore pressure at depth to determine whether these effects are local fluid-driven sources or large scale source effects.

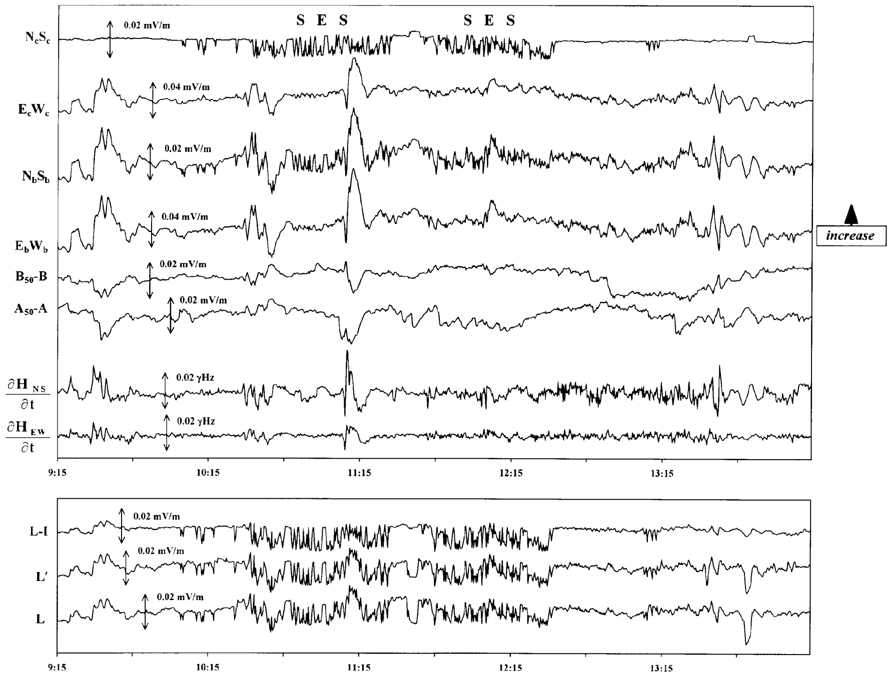

Figure 7. Observed SES recorded on multiple dipoles on April 19, 1995 at the Ioannina Station. Simultaneous measurements of magnetic field gradient are shown in the lower two plots (from Varotsos et al., 1996).

4.5. Electromagnetic Effects

Another enigma concerns the generation of high frequency ($>$ 1 KHz) electromagnetic emissions prior to moderate earthquakes. Such emissions have been reported detected at great distances from these earthquakes (Gokhberg et al., 1982; Oike and Ogawa, 1982; Yoshino et al., 1985; Yoshino, 1991; Parrot et al., 1993; Fujinawa and Takahashi, 1994; Hayakawa and Fujinawa, 1994) and are reported detected also in satellite data (Molchanov et al., 1993; Parrot, 1994), although the statistical significance of these observations is under dispute (Henderson et al., 1993; Molchanov et al., 1993; Parrot, 1994). This area of research has received attention following reports of changes in the level of 81-kHz electromagnetic radiation around the time of a magnitude 6.1 earthquake at a depth of 81 km beneath the receiver (Gokhberg et al., 1982). Radio emissions at 18 MHz were recorded at widely separated receivers in the northern hemisphere for about 15 minutes before the May 16, 1960 great Chilean earthquake (Warwick et al., 1982).

While generation of high frequency electromagnetic radiation can be easily demonstrated in controlled laboratory experiments involving rock fracture (Warwick et al., 1982; Brady and Rowell, 1986; Brady, 1992), the physical mechanism for generation and the method of propagation of very high frequency (VHF) electromagnetic waves through many tens to hundreds of kilometers of conducting crust (and through the ocean) is not at all clear. As discussed above, simple “skin depth” attenuation arguments would preclude sources generating fields in the kilohertz to megahertz range tens to hundreds of kilometers deep in a conducting crust (1 S m^{-1} to 0.001 S/m) that would be detectable at the earth’s surface. Appeal to secondary sources at the earth’s surface (Yoshino and Sato, 1993) may avoid this difficulty but the implied large surface fields are not observed.

High frequency disturbances are, of course, generated in the ionosphere as a result of coupled infrasonic waves generated by earthquakes. Essentially, the displacement of the earth’s surface by an earthquake acts like a huge piston generating propagating waves in the atmosphere/ionosphere waveguide (see review by Francis, 1975). Traveling waves in the ionosphere (traveling ionospheric disturbances or TID’s) are thus a consequence of earthquakes (and volcanic eruptions). EM data at VHF frequencies recorded on ground receivers or by satellite thus require correction for TID disturbances (and disturbances from other sources) before identifying these as direct electromagnetic precursors to earthquakes, or consequences of earthquakes.

4.6. Volcanomagnetic effects

Several important eruptions for which electromagnetic, magnetic, and electric field data have been obtained occurred during the past six years. Critical data sets have been obtained during the eruptions of Mt Unzen in Japan (Tanaka, 1995), Merapi in Indonesia (Zlotnicki and Bof, 1996), Etna in Italy (Rees et al., 1995), and during rapid deformation in the Long Valley in California (Mueller et al., 1991). Earlier reports of observations of volcanomagnetic effects can be found in Johnston (1989) for Mount St. Helens and Zlotnicki et al. (1990) for Piton de la Fournaise. An important summary of data obtained during the Izu-Oshima eruption in 1986–87 can be found in Yukutake (1990a).

It is important to realize that different physical processes are likely to give rise to magnetic and electric field perturbations during volcanic activity as changing physical conditions (stress, temperature, fluid flow, etc) and changing geometry of magnetic bodies and the volcano occur. Prior to eruptions, larger-scale stress-generated contributions (indicated by coherence with independent deformation measurements) are most likely (e.g. Mueller et al., 1991). Some thermally generated contributions may result if large volumes of high temperature fluids and gases transport heat from magma reservoirs (Yukutake et al., 1990b, 1990c; Sasai et al., 1990). During eruptions, large localized changes occur near active vents from mass displacement, massive remagnetization/demagnetization effects, stress change and, of course, rapid changes from massive charge separation generated in the rising ash cloud usually with spectacular lightning displays (Johnston et al., 1981). Dramatic post-eruption effects primarily arise from remagnetization/demagnetization effects accelerated by penetration of rainwater in opening cracks and fissures in and

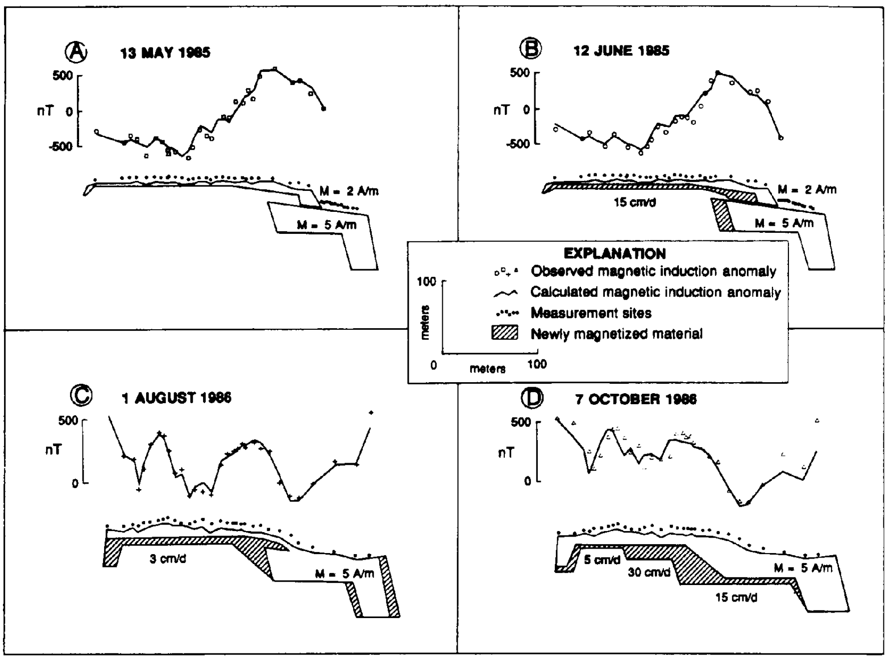

Figure 8. Observed and modeled thermomagnetic anomaly profiles on Mount St. Helens during four time periods from 13 May, 1985 to October 7, 1985 (from Dzurisin et al., 1990).

near eruptive materials (Dzurisin et al., 1990; Tanaka, 1995). Mount St. Helens provided a good example of how this process works. Figure 8 from Dzurisin et al. (1990) illustrates the size of the region acquiring magnetization in a matter of months and, in doing so, generated magnetic anomalies of hundreds of nanotesla. In places where rainfall is seasonal, acquired magnetization would be expected, and is observed (Zlotnicki et al., 1993), to have an annual variation, although Zlotnicki et al. (1993) attribute this to electrokinetic effects.

4.6.1. *Mt Unzen, Japan, 1990–91*

A summit eruption from Unzen volcano in Japan started on November 17, 1990, after nearly 200 years of quiescence. Magnetic field monitoring during the four-year period prior to the eruption indicated a systematic change of 20 nT (Tanaka, 1995). Concern about this change triggered installation of a small array of closely spaced continuously recording magnetometers. During the eruption, changes of from +18 nT to −5 nT were observed followed by changes of hundreds of nanoteslas during the post eruption period until mid-1992. Locations of the recording sites are shown in Figure 9a and records collected during the eruption in February, 1991 are shown in Figure 9b. (Tanaka, 1995). Figure 10 shows the longer term data covering the dome building episode in May-June, 1991 (Tanaka, 1995). Tanaka

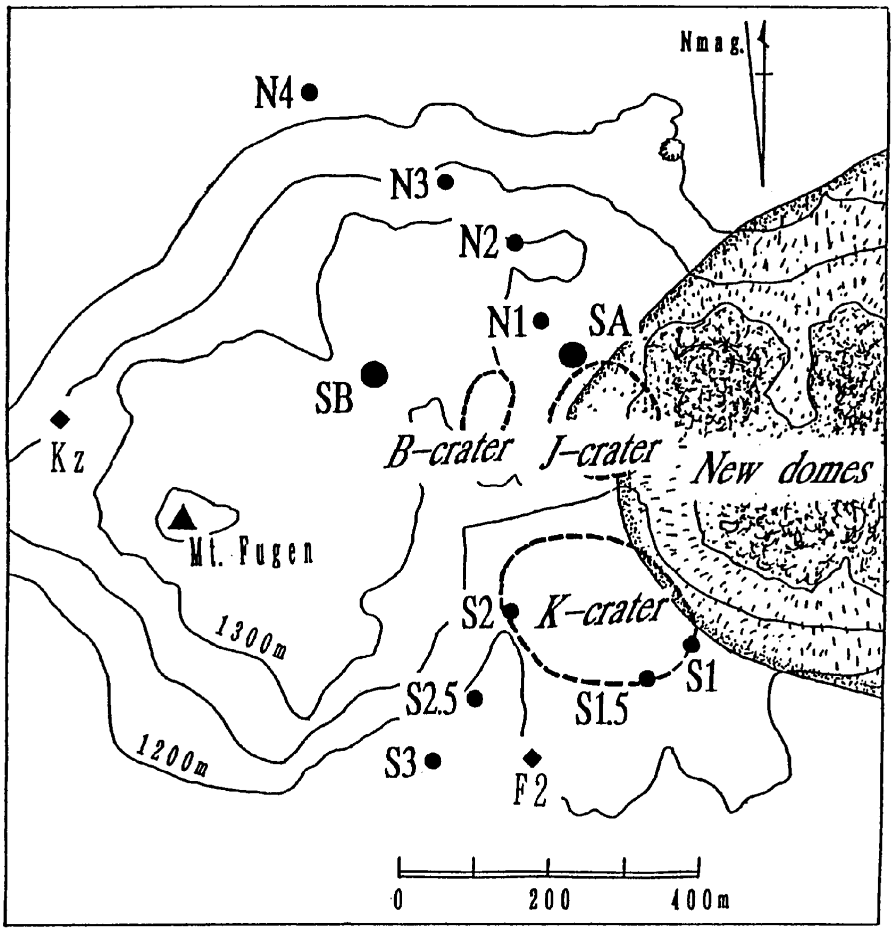

Figure 9a. Magnetometer sites on Unzen volcano (from Tanaka, 1995).

(1995) attributes these signals primarily to magnetization/demagnetization effects. Other electrical and VHF features are discussed below.

4.6.2. *Merapi, Indonesia (1992–95)*

The January 20, 1992, the November 22, 1994, and the May, 1995 eruptions of Merapi volcano in Indonesia were monitored by a small network of magnetometers (Figure 11a from Zlotnicki and Bof, 1996). The observed changes for each eruption are relatively coherent across the three monitoring points and amount to only a few nT at sites between 2.5 and 5 km from the growing dome in the caldera (Figure 11b). At this distance, the coherence between the observed changes indicates they result from stress magnetization changes due to a pressure source at depths greater than station separation, beneath the dome (Zlotnicki and Bof, 1996). Observations on or near the dome would presumably have indicated larger localized co-eruptive

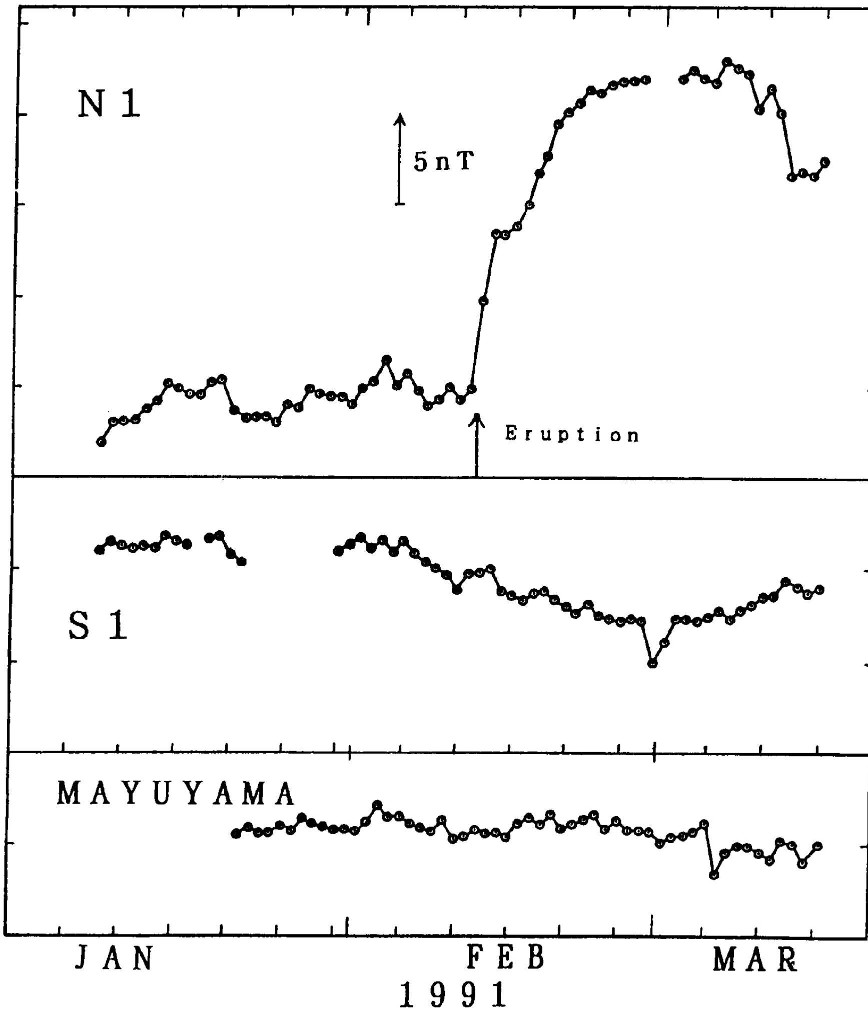

Figure 9b. Magnetometer data recorded on Unzen Volcano during the eruption on February 12, 1991 (from Tanaka, 1995).

and post-eruptive changes but the frequent collapses and pyroclastic flows probably would have destroyed any such instruments.

4.6.3. *Etna, Italy, 93–present*

Observations during the recent eruptions of Mount Etna in Sicily (Rees et al., 1995) indicate results similar to those obtained on Oshima, Japan (Yukutake, 1990a), Unzen (Tanaka, 1995) and Mount St. Helens (Johnston et al., 1981; Dzurisin et al., 1990), with only small amplitude long term changes seen at distances of 5 km to 10 km from the volcano, but large changes of several hundred nanotesla near the areas of recent activity.

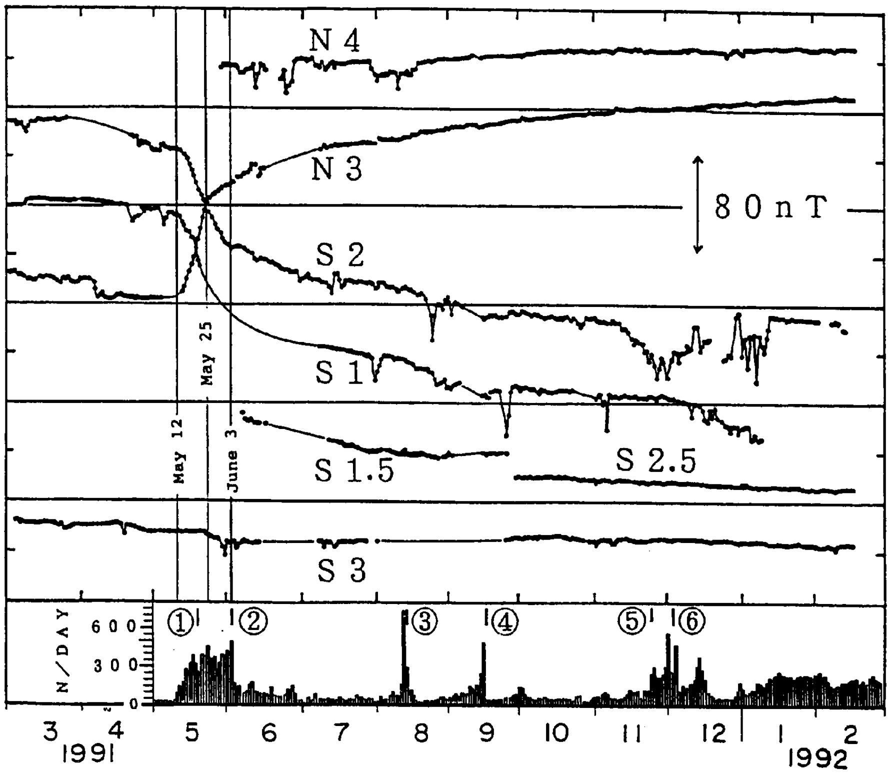

Figure 10. Magnetic field data recorded on Unzen volcano from March 1, 1991 to February 29, 1992 through the dome building sequence in May, 1991 (from Tanaka, 1995).

4.6.4. *Long Valley, California, 1989–present*

One of the more spectacular examples of volcanomagnetic changes clearly related to deformation measurements on a rapidly inflating volcanic resurgent dome, are currently being obtained in the Long Valley Caldera in eastern California. Swarms of earthquakes and gas emissions continue, but no surface eruption has yet occurred. Figure 12 shows comparative records of magnetic field at two sites on the resurgent dome, cumulative seismicity, extension as measured on one of the many geodetic lines monitored continuously in the caldera (see Mueller et al. (1981) for a description of experimental details) and the earthquake magnitude time history. Inflation and uplift accelerated in late 1989 due apparently to magma injection at about 7 km beneath the dome (Langbein et al., 1993). As a result of this inflation, this region is undergoing rapid continuous broad-scale straining of about 3 ppm/year and the magnetic field is decreasing at the center of the resurgent dome at 2 nT/yr. A simple piezomagnetic source model derived from the geodetic strain data generates the observed signals and their time scale (Mueller and Johnston, 1996).

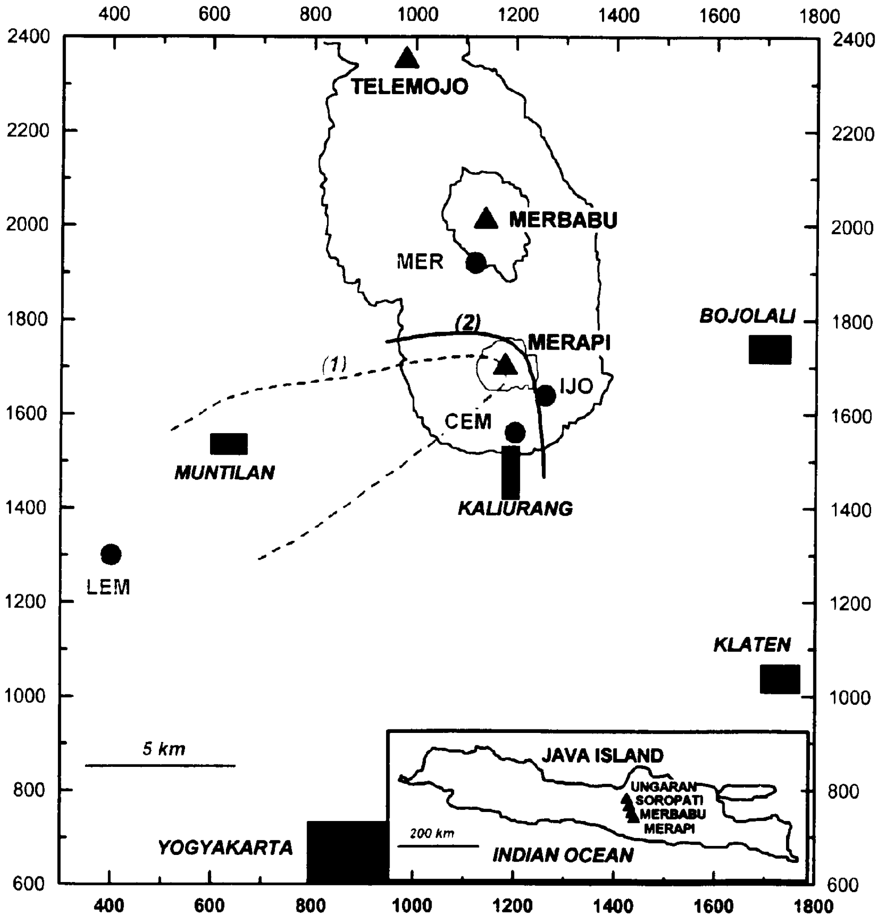

Figure 11a. Locations of recording magnetometer sites on Merapi volcano, Indonesia (from Zlotnicki and Bof, 1996).

4.7. Volcanoelectric Effects

A number of important new experiments have focused on electric field measurements on, or near, active volcanoes. In particular, impressive self-potential observations have been made that delineate the changing electric potentials around erupting volcanoes (Zlotnicki et al., 1994; Hashimoto and Tanaka, 1995). Self-potential (SP) anomalies do show apparent correlation to episodes of extrusive activity and eruptions on Kilauea volcano in Hawaii (Jackson and Kauahikaua, 1987). The pertinent physical mechanisms in this case are electrokinetic effects, generated by thermally driven or strain driven fluid flow within the volcano, or thermoelectric effects, generated by injection of hot material into and through the volcano. A second important experiment type concerns continuous ULF/VLF vertical electric field

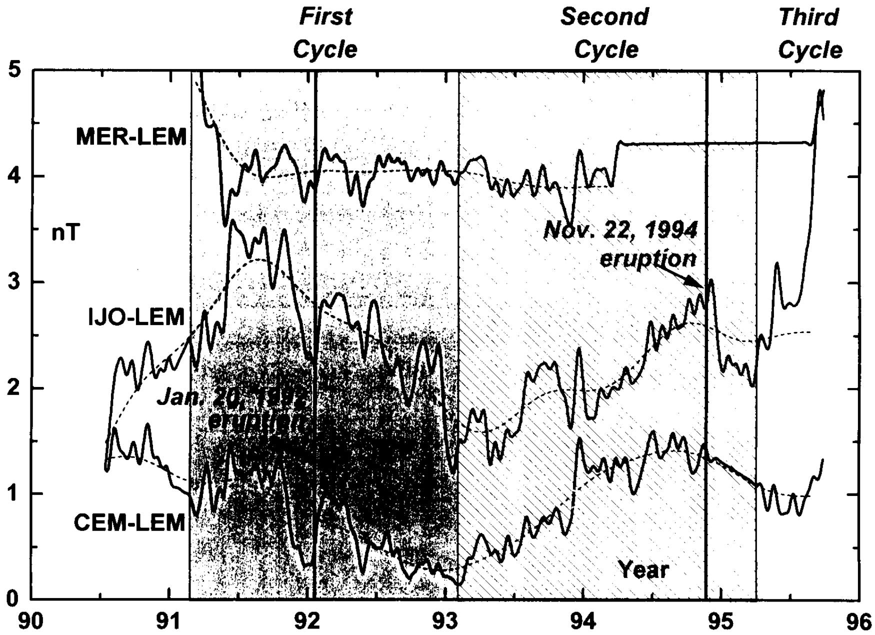

Figure 11b. Magnetic field data recorded on Merapi volcano from 1990 to late 1995 though eruptions on Jan 20, 1992, and November 22, 1994 (from Zlotnicki and Bof, 1996).

measurements that were made during a minor eruption of Mt Mihara on Oshima, Island (Fujinawa et al., 1992.

4.7.1. *Unzen, Japan, 1990–92*

A large positive self-potential anomaly has been identified by Hashimoto and Tanaka (1995) in the vicinity if the new lava dome on Unzen volcano. Figure 13 shows this distribution in relation to the most distant south-western site. Potentials generated by fluid flow in the hydrothermal system beneath the array is suggested as the most likely cause for this anomaly. Similar self-potential anomalies have been observed by Zlotnicki et al. (1994) on Piton de la Fournaise volcano.

4.7.2. *Oshima, Japan, 1990*

Continuous measurements of vertical electric fields in the frequency range 0–3 KHz before, during and after a minor eruption from Mt Mihara on Oshima Island show a series of rapidly-rising slow-decaying pulses (Fujinawa et al., 1992). Fujinawa et al. (1992) attribute these effects to electrokinetic phenomena in the hydrothermal system.

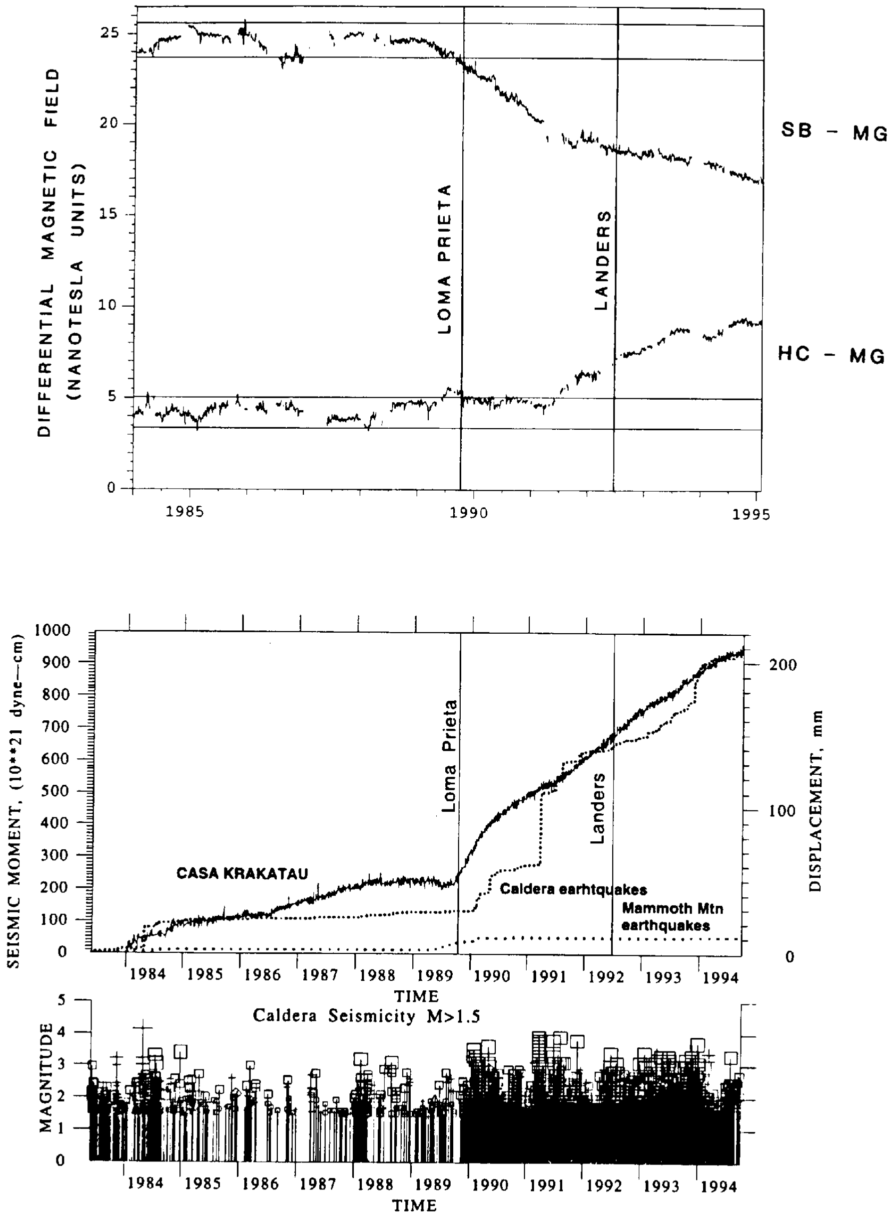

Figure 12. Comparative time-series plots of local magnetic field (top plot), geodetic line length (upper time history – middle plot), cumulative seismicity within the caldera (close dots – middle plot) and under Mammoth Mountain (open dots – middle plot) and earthquake magnitudes as a function of time (bottom plot) during the current inflation episode in Long Valley caldera (from Mueller and Johnston, 1996).

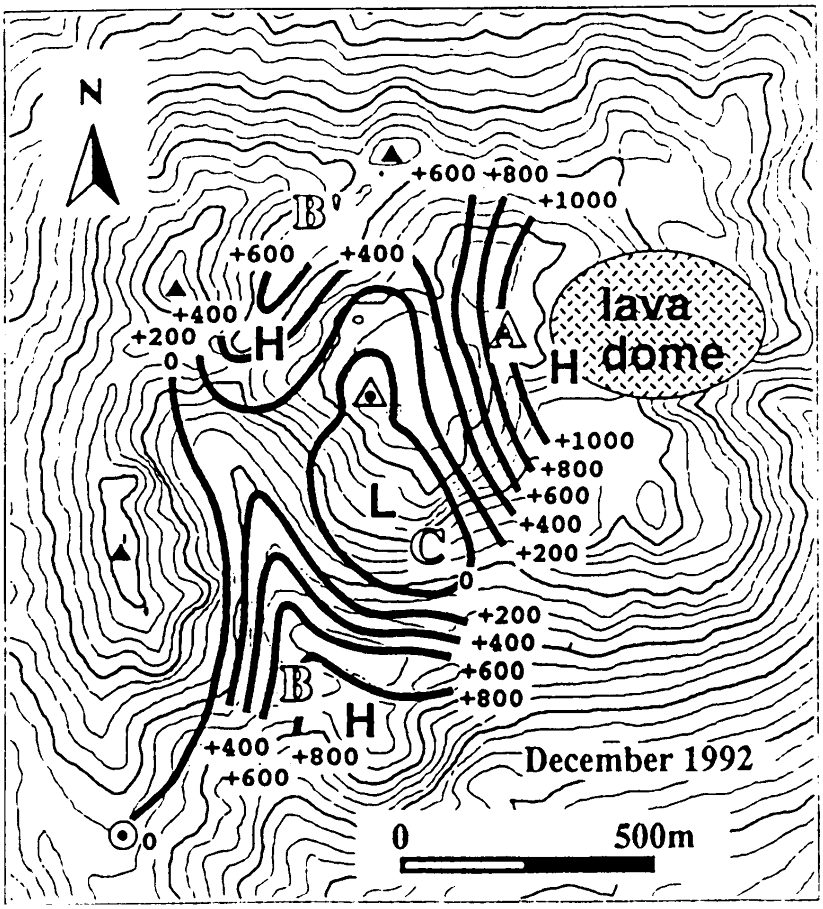

Figure 13. Contours of self-potential (in millivolts) on Unzen volcano in December 1992, (from Hashimoto and Tanaka, 1995).

5. Conclusions

In summary, as more and better quality measurements of magnetic, electric and electromagnetic fields are being made in the epicentral regions of earthquakes and on volcanoes around the world, it is clear that observed field perturbations are generated by a variety of source processes. It is also clear from this review that there are many problems which, if not addressed, will soon reflect on the credibility of this field of geophysics. Particular problems needing work include:

(1) Inclusion of constraints on the various physical mechanisms and models of various processes that are imposed by data from other disciplines such as seismology, geodesy, etc.

(2) Ensuring reports of observations demonstrate self consistency, adequate signal-to-noise, adequate noise quantification or consistency with other geophys-

ical data obtained in the area. Some unusual records are claimed as precursors just because they precede some local or distant event without adequate demonstration of consistency with coevent source models or demonstration of significance against the background noise. Since high-precision deformation data obtained in the near-field of earthquakes show precursive strain in the weeks to seconds before earthquakes is less than 1% of that occurring during the earthquake, acceptance by the geophysical community of "precursive" EM signals without believable coseismic signals, should not be expected.

(3) The need to use reference stations to quantify and remove common-mode noise generated in the ionosphere/magnetosphere and to isolate the the most likely location of signal sources in the earth's crust. This appears to be particularly neglected in recent associations of ULF/VLF data with earthquakes and volcanic eruptions.

On the positive side, better understanding has been obtained from well designed experiments with multiple independent instrument types recording in the same frequency band. This has allowed quantification of at least some of the mechanisms generating electromagnetic fields with tectonic and volcanic activity. It now seems likely that:

(1) Thermal demagnetization and remagnetization effects occurred during volcanic activity on Mount St. Helens, Oshima Island (Japan), Piton de la Fournaise (Reunion Is), Unzen (Japan), and Etna (Italy).

(2) Stress generated magnetic effects (piezomagnetic effects) occurred during volcanic activity at Long Valley, California, Oshima Island, Japan, Mount St. Helens, and during the Loma Prieta, North Palm Springs, and Landers earthquakes.

(3) Electrokinetic effects occurred during volcanic activity at Piton de la Fournaise, Unzen and Oshima Island (Japan), and perhaps during the Loma Prieta earthquake.

Finally, it is obvious that more complete observational arrays are needed in critical locations to provide a sound data base for this field. Earthquakes and eruptions occur infrequently and great skill is need to "catch" these events with adequate arrays so we can understand better the physical processes involved.

Acknowledgements

I would like to thank my many colleagues for stimulating discussions.

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Address for correspondence:
Dr. M.J.S. Johnston, USGS ms977, 345 Middlefield Rd., Menlo Park, CA 94035, U.S.A.
Phone: 1(415)3294812; Fax: 1(415)3295163; Email: mal@thebeach.wr.usgs.gov