S

SEISMO-ELECTROMAGNETIC EFFECTS

Seismo-electromagnetic effects refer to electromagnetic (EM) signals generated by fault failure processes in the Earth's crust. These may occur slowly (when associated with plate tectonic loading, slow earthquakes, postseismic slip, etc.) or rapidly preceding, during and following earthquakes. Several different physical processes related to crustal failure can contribute to the generation of seismo-electromagnetic (SEM) effects. Unambiguous observations of SEM effects provide new independent information about the physics of fault failure. Causal relations between co-seismic magnetic field changes and earthquake stress drops have been clearly documented. However, despite several decades of high quality monitoring, clear demonstration of the existence of precursory EM signals has not been achieved.

Brief history

Suggestions that electromagnetic field disturbances are a consequence of crustal failure processes have been made throughout recorded history. Unfortunately, much of the earliest work was recognized as spurious by Reid (1914) who showed that transients recorded by magnetographs located close to earthquake epicenters resulted from earthquake shaking, not earthquake source processes. This invalidated earlier reports from magnetic variometers (suspended magnets) and other instruments sensitive to ground displacement, acceleration, and rotation common in epicentral regions during the propagation of seismic waves. Other early problems resulted from inadequate rejection of ionospheric, magnetospheric, and man-made noise (Rikitake *et al.*, 1966).

Since the mid-1960s, these problems have been avoided through the use of absolute magnetometers installed in regions of low magnetic field gradient to reduce sensitivity to earthquake shaking and by the application of new noise reduction techniques. As a consequence, unambiguous observations of EM variations related to earthquakes and tectonic stress/strain loading, have now been obtained near active faults in many countries (Japan, China, Russia, USA, and other locations). However, careful work still needs to be done to convincingly demonstrate causality between "precursory" EM signals and earthquakes and consistency with other geophysical data reflecting the state of stress, strain, material properties, fluid content, and approach to failure of the Earth's crust in seismically active regions.

Physical mechanisms involved

The loading and rupture of water-saturated crustal rocks during earthquakes, together with fluid/gas movement, stress redistribution, change in material properties, has long been expected to generate associated magnetic and electric field perturbations. The primary mechanisms for generation of electric and magnetic fields with crustal deformation and earthquake related fault failure include piezomagnetism, stress/conductivity, electrokinetic effects, charge generation processes, charge dispersion, magnetohydrodynamic effects, and thermal remagnetization and demagnetization effects. Physical limitations, constraints, and frequency limitations placed on these processes are discussed in Johnston (2002).

Basic measurement limitations

The precision of local magnetic and electric field measurements on active faults 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 ionosphere, magnetosphere, and cultural noise than by instrumental noise. Thus, systems for quantifying these noise sources are of crucial importance if changes in electromagnetic fields are to be uniquely identified. For spatial scales of a few kilometers to a few tens of kilometers comparable to moderate magnitude earthquake sources, geomagnetic, and electric noise power decreases with frequency as $1/f^2$, similar to the "red" spectrum behavior of most geophysical parameters. Against this background noise, transient magnetic fields can be measured to several nanotesla over months, to 1 nT over days, to 0.1 nT over minutes, and 0.01 nT over seconds. Long term changes and field offsets can be determined if their amplitudes exceed about a nanotesla. Comparable electric field noise limits are 10 mv km^{-1} over months, several mv km^{-1} over days, 1 mv km^{-1} over minutes and 0.1 mv km^{-1} over seconds. EM noise increases approximately linearly with site separation. Cultural noise further complicates measurement capability because of its inherent unpredictability. This largely precludes measurements in urban areas. At lower frequencies (microhertz to hertz) for both electric and magnetic field measurements, the most common technique involves the use of reference sites with synchronized data sampling in arrays using site spacing comparable to the expected source sizes of a few tens of kilometers. Adaptive filtering, use of multiple variable-length sensors in the same and nearby locations further reduce noise by about a factor of 3.

These same techniques can be applied to electromagnetic field measurements at higher frequencies (100 Hz to MHz) but much less is known about the scale and temporal variation of noise. These frequencies may be less important since basic physics precludes simple propagation of high-frequency EM signals from seismogenic depths (5–100 km) on active faults in the Earth's crust where the electrical conductivity is more than 0.1 S m^{-1}.

Recent results: general constraints

If reliable magnetic and electric field observations are indeed source related, clear signals should occur at the time of large local earthquakes because the primary energy release occurs at this time. These signals should scale with the earthquake moment (size) and source geometry. In fact, co-event observations provide a determination of stress sensitivity since the stress redistribution and the source geometry of earthquakes are well determined. With this "calibration," SEM effects can be quantified and spurious signals identified. Observations without consistent and physically sensible co-seismic effects are generally considered suspect.

High-resolution strain data at the epicenters of moderate to large earthquakes show that precursive moment release during the months to minutes before rupture is less than 0.1% of that occurring coseismically (Johnston and Linde, 2002). This strongly limits the scale of precursive failure and the expected "size" of precursive effects.

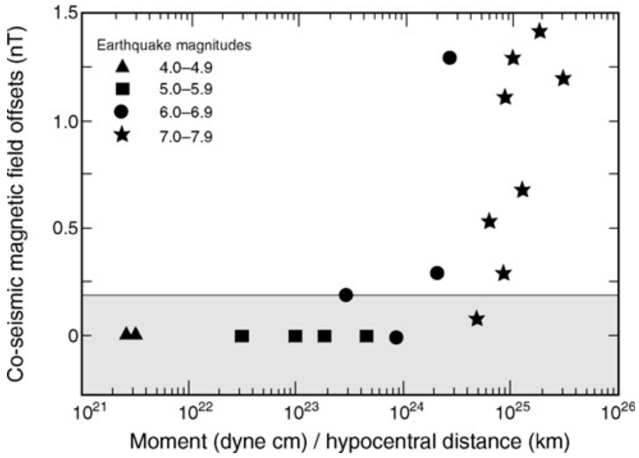

Figure S26 Co-seismic magnetic field offsets as a function of seismic moment scaled by hypocentral distance. The shaded region shows the 2-sigma measurement resolution (from Mueller and Johnston, 1998). Geodetically based seismomagnetic models (Sasai, 1991) fit each offset.

Examples of seismomagnetic effects

The primary features of seismomagnetic effects are shown in Figure S26 from Mueller and Johnston (1998). It is apparent that maximum signals are not more than a nanotesla or so and these signals occur only for larger earthquakes ($M > 6$) for which corresponding strain changes are about a microstrain or so. An example of a magnetic record observed at the epicenter of the 1986 M5.9 North Palm Springs earthquake and 17 km from the 1992 M7.4 Landers earthquake is shown in Figure S27. For this, and some 40 other earthquakes with magnitude between 5.5 and 7.4, no significant precursory magnetic signals were observed.

Seismoelectric effects

Seismoelectric observations that 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. Measurements of electrical resistivity to better than 1% have been made since 1988 in a well designed experiment installed near Parkfield, California (Park, 1997). An expected M6 earthquake together with several M5 earthquakes have occurred beneath this array since 1990. None of these earthquakes generated any observable changes in resistivity above the measurement resolution (Park, 1997; Langbein *et al.*, 2005).

Indirect observations of possible SE signals might be obtained using the magnetotelluric (MT) technique to monitor apparent resistivity in seismically active regions. Even with the best designed systems using remote referencing systems to reduce noise and obtain stable impedance tensors, it is difficult to reduce errors below 5% for good soundings and 10–40% for poor soundings.

Possible high-frequency precursory effects

A number of observations purported to be high-frequency SEM effects have been recently reported (Hayakawa, 1999). Interest in these higher ULF frequencies primarily resulted from the fortuitous observation of elevated ULF noise power on a single 3-component magnetometer near the epicenter of the M7.1 Loma Prieta earthquake of October 18, 1989. However, similar records were not obtained with the 1992

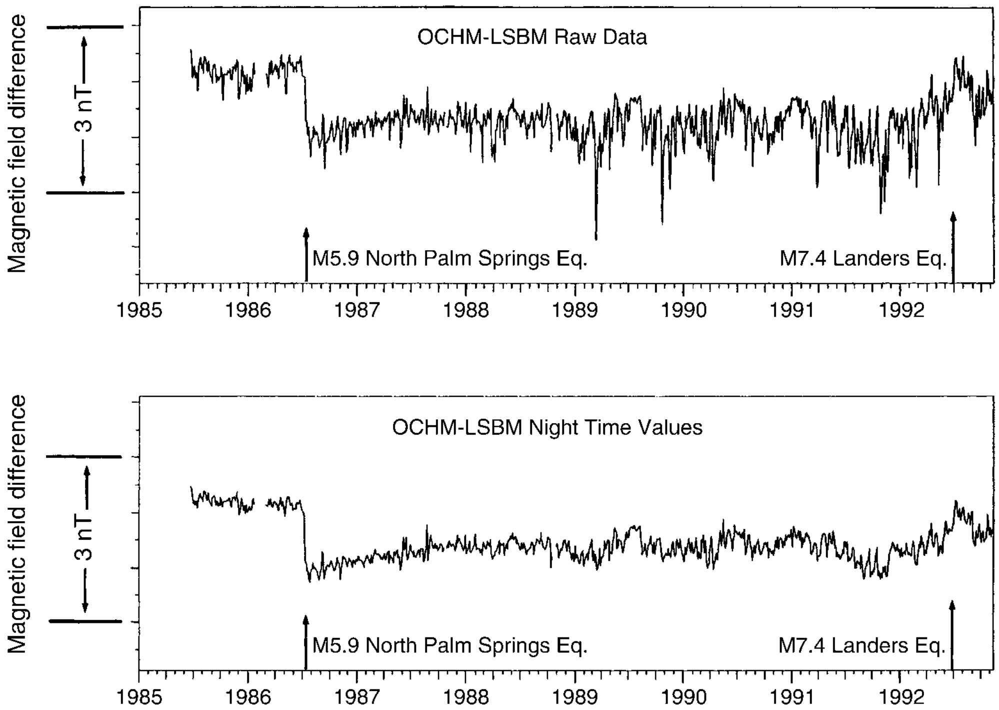

Figure S27 Magnetic field differences between stations OCHM and LSBM before, during and after the July, 1986 M5.9 North Palm Springs and the June, 1992 M7.4 Landers earthquakes (from Johnston *et al.*, 1994).

M7.4 Landers earthquake, the 1994 M6.7 Northridge earthquake, the 1999 M7.1 Hector Mine earthquake or the 1999 M7.4 Izhmet, Turkey earthquake.

Though controversial, increased interest in tectonoelectric (TE) phenomena related to earthquakes has resulted from suggestions in Greece and Japan that short-term geoelectric field transients (SES) of particular form and character precede earthquakes with $M > 5$ at distances up to several hundreds of kilometers. These transients appear to have a spatially uniform source field on the scale of the array but no clear corresponding magnetic field transients and no sensible coseismic effects. The SES have been empirically associated with subsequent distant earthquakes and claimed as precursors (Varotsos *et al.*, 1996).

Careful study of the SES recordings indicates that the SES signals have the form expected from rectification/saturation effects of local radio transmissions from high-power transmitters on nearby military bases. Without any clear physical explanations describing how the SES signals are earthquake generated yet coseismic effects related to the much larger earthquake source are not observed, these observations have been extremely controversial (Debate on VAN, 1996).

Another enigma concerns the generation of high-frequency (>1 kHz) electromagnetic emissions associated with subsequent moderate earthquakes but, again, with no coseismic effects. Such emissions are reported to have been detected at great distances from these earthquakes (see summary by Hayakawa and Fujinawa, 1994) and by magnetometers onboard satellites. However, the statistical significance of these observations is under dispute.

The generation of high-frequency electromagnetic radiation can be easily demonstrated in controlled laboratory experiments involving rock fracture in dry rocks. However, the Earth's crust in seismically active areas is quite conducting (0.3–0.001 S m^{-1}) and propagation of very high-frequency (VHF) electromagnetic waves even short distances through the crust is difficult to justify physically. Propagation from earthquake source regions (10–100 km in depth), and in some cases through oceans with conductivities of 1 S m^{-1}, is physically implausible. More significantly, the amount of allowable rock fracturing prior to earthquakes is strongly constrained by high sensitivity crustal strain measurements in the near-field of many earthquakes. These measurements indicate moment release (μslip × area) prior to earthquakes is at least three orders of magnitude smaller than that released at the time of an earthquake (Johnston and Linde, 2002). Appeal to secondary sources at the Earth's surface may avoid this difficulty but the expected associated near-field crustal strain and displacement fields are not observed.

High-frequency disturbances are generated in the ionosphere as a result of coupled infrasonic waves generated by earthquakes and are readily detected with routine ionospheric monitoring techniques and global position system (GPS) measurements. In essence, displacement of the Earth's surface by an earthquake acts like a huge piston, generating propagating pressure waves in the atmosphere/ionosphere waveguide. Thus, traveling waves in the ionosphere (traveling ionospheric disturbances or TIDs) are a consequence of earthquakes (and volcanic eruptions). EM data at VHF frequencies recorded on ground receivers or by satellite require correction for TID and other disturbances before any association can be made to source processes or earthquake precursors.

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Cross-references

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