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Electromagnetic Disturbances Associated With Earthquakes: An Analysis of Ground-Based and Satellite Data

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Abstract—Several observations were made of Very Low Frequency (VLF) emissions apparently associated with earthquakes, which were recorded independently at ground-based stations and on satellites. The observations at the Kerguelen station (49°26′S, 70°25′E) were made using magnetic antennae, on April 24 and 25, 1980, during a period when three earthquakes with magnitude $M_s > 4.7$ took place near the station. Several increases of electromagnetic waves at the time of earthquakes were recorded on the polar-orbiting satellite AUREOL-3. The observations on the geostationary GEOS-2 satellite were made using magnetic and electric antennae during the period 1977–1981. Data were analysed for those cases when both intense ($M_s > 5$) earthquakes occurred in the region close to the satellite longitude and the satellite was operating in the VLF mode. A statistical analysis, based on the enhancement of wave intensity at the time of earthquakes and using GEOS-2 data, seems to indicate that there is a (possibly indirect) association between seismic activity and some of the VLF emissions observed at the satellite. Ionospheric measurements made from the ground also showed an increase of the critical frequency f_0E_s of the sporadic layer E_s when earthquakes occurred nearby. Some aspects of the relation between the VLF emissions and the seismic activity are discussed.

Introduction

When considering the problem of electromagnetic emissions correlated with earthquakes, one is often faced with an overwhelming problem of complexity, not only because the mechanism of wave generation is not entirely understood, but also because earthquakes without any emissions are known to occur. However, there is a great interest in the subject for the simple reason that when the precursor event occurs it normally does so less than a few hours before the shock. A further possibility is that this phenomenon could perhaps be related to other intriguing processes already observed, such as the earthquake lights (Derr, 1973) and/or anomalous animal behaviour (Rikitake, 1981). Since the first papers of Gokhberg, Morgunov, Yoshino, and Tomizawa (1982), and Warwick, Stoker, and Meyer (1982), a great deal of theoretical work, and active and passive experiments has been done to provide evidence of this short-term precursor.

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The next chapter will present the observations made from the geophysical data available in France. The unresolved problems raised by such phenomena are discussed in the last chapter.

The Data

This chapter presents the electromagnetic perturbations observed with different experiments at different altitudes, from ground level up to the geostationary orbit. For approximately 30 years, a French geophysical station has been operational in the Kerguelen Islands (49°26′S, 70°25′E), and in April 1980 three earthquakes occurred very close by. Their magnitude was moderate ($M_s \approx 4.7$) and the distance between their epicenters and the station was about 100 kilometers. In two cases a change of the wave intensity was observed, $\approx$1.5 hour before the earthquake, in three filters centered around 800, 1,700, and 3,600 Hz (Parrot, Lefeuvre, Corcuff, & Godefroy, 1985). The main difference with the natural emissions already observed at Kerguelen was that the starting time of the increase was different for the three filters, the lowest frequencies detecting the earthquake event well before the shock. The signal variations were similar to those reported by Ralchovsky and Komarov (1988) at 5 and 10 kHz, and the increase did not stop abruptly at the time of the shock, as was observed by Gokhberg et al. (1982) at higher frequencies (81 kHz).

The first report on an ionospheric anomaly occurring before an earthquake was made by Nestorov (1979) and, as a sounder was functioning in the Kerguelen Islands, the ionospheric data were checked. It was shown (Parrot et al., 1985) that the critical frequency of the E_s layer (f_0E_s) increases at the time of the shocks, indicating that the ionosphere was disturbed at this time. The same quantitative measurement was made at the station of Djibouti when earthquakes occurred close by (Parrot & Mogilevsky, 1989), where the results also indicated a disturbed ionosphere just prior to the earthquakes. But, there are many other reasons to increase the electron density, and very commonly the f_0E_s curves show other peaks, in particular during the daytime. Thus, the results are more convincing when the earthquake occurs during the night.

Concerning the upper ionosphere, examples of ELF emissions observed with a low-orbiting satellite above the epicenter of an earthquake about to happen, are given by Larkina, Nalivayko, Gershenzon, Gokhberg, Liperovskiy, and Shalimov (1983) and by Parrot and Mogilevsky (1989). But, the search for correlations between seismic activity and electromagnetic emissions is also restricted by the natural noise. The maximum intensity of the low-altitude ELF hiss occurs when the invariant latitude is larger than 50°, see Figure 7 of Parrot (1990), and then only in the region between 40° N and S of the equator is used, where on average no natural emission occurs. Another example of such observations made by the low-orbiting satellite AUREOL-3 (apogee 2012 km, perigee 408 km, inclination 82°5′, period

109.5 minutes) is given in Figure 1. It concerns an earthquake of magnitude 5.5 that occurred on March 19, 1982 at 00.17.52 UT, whose epicenter was in West Irian at latitude 02.80°S, longitude 138.81°E. The focal depth was 48 km. A part of the Aureol-3 orbit is plotted in Figure 1A with an indication of the time. The star indicates the epicenter of the earthquake, which occurred roughly 3 hours after the pass of the satellite over the epicentral area. The signals of the electric component E_H in the 72 Hz filter are plotted in Figure 1B as a function of the time. Regular blanks are a result of the on-board calibrations, which were removed from the data. A signal increase is seen around 20.50 UT when the satellite passed over the same latitude as the epicenter. The increase is also observed on the B_z component in the 140 Hz filter, and on the same component E_H in the 150 and 325 Hz filters, but the amplitude of the peak decreases as the frequency increases. No signal is seen at frequencies > 800 Hz in this case. As was explained before, the level increase observed after 21.00 UT is not related to the seismic activity, as it can be observed each time the satellite enters in the midlatitude zone.

The case studies have shown increases of the signal at the time of earthquakes, but ELF emissions are very common and can result from many other phenomena. Thus, the only way to know if those increases are coincidental or not, is by using statistics.

A statistical study was carried out with the data recorded by the geostationary satellite GEOS-2 (Parrot & Lefeuvre, 1985). Using a geostationary satellite made it possible to prevent the effects due to the time–space ambiguity. Earthquakes with magnitudes greater than 4.7 (the magnitude of the earthquakes in the Kerguelen Islands) and with epicenters located near the magnetic field line of the satellite were selected. With a rough signature of an earthquake based on an ELF increase a quarter of an hour before or after the earthquake, a positive correlation of 44% was obtained. The same analysis performed on a random data set taken during the life of GEOS-2 gave a percentage of 41%, which is very similar. However, the important point is that, when we decrease the distance between the longitude of the epicenter and the longitude of the satellite to less than 20°, the percentage goes up to 51%. When only low-magnetic-activity periods are considered, the percentage of positive correlation is 46% against 31% for the random data set. Another interesting point is the relation with the frequencies. Parrot and Lefeuvre (1985) showed that for the random data set, the maximum of positive correlations occurred around 1 kHz, which is the frequency where most of the natural noise was usually observed on GEOS-2, whereas, for the earthquake data set, this maximum moved to lower frequencies.

Discussion

A multitude of wave emissions that were observed prior to earthquakes have been described by many authors and obtained in different countries.

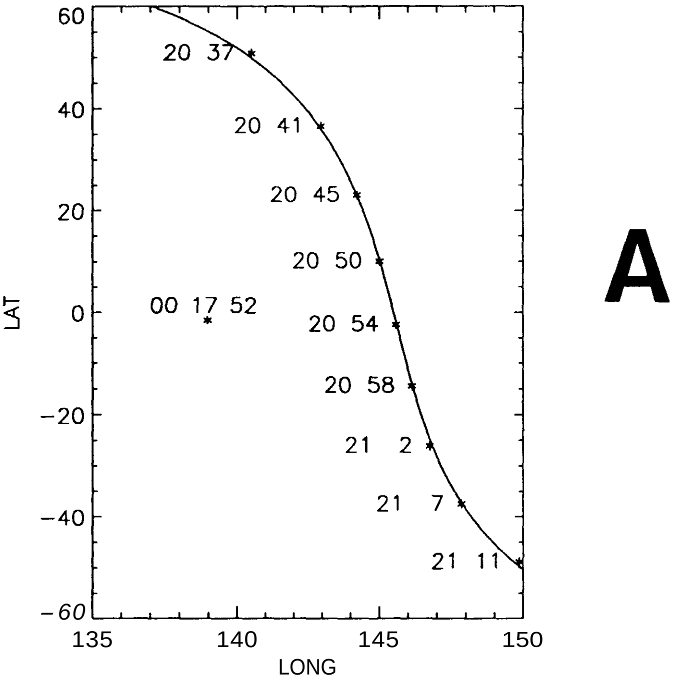

Fig. 1. (A) Orbit of the satellite AUREOL-3 between 20.37 and 21.11 UT on March 19, 1982. The star indicates the epicenter of an earthquake that occurred at 00.17.52 UT. (B) Time variation of the signal recorded by the E_H component in the 72 Hz filter. The units are in $V.m^{-1}.Hz^{-1/2}$.

However the main question that still remains to be answered is: Where is the origin of such electromagnetic waves? The possible locations of this origin are:

- At the focal point, which is the place where the mechanical energy of the earthquake is strongest.
- At the surface or in the higher levels of the crust, where microfractures and friction between constituants rocks occur.
- In the atmosphere or in the ionosphere, where a variation of the DC electric field at the ground can either produce an instability, or change the conditions of propagation of the natural waves (in this last case, the phenomena will contribute only to amplify the natural waves).

Much of the research was also aimed at understanding what causes the electromagnetic signals observed prior to an earthquake. It is virtually certain that these phenomena are due to stress variations before the shock, but the precise physical mechanisms involved are not yet understood. Several hypotheses have been put forward in the literature, describing mechanisms

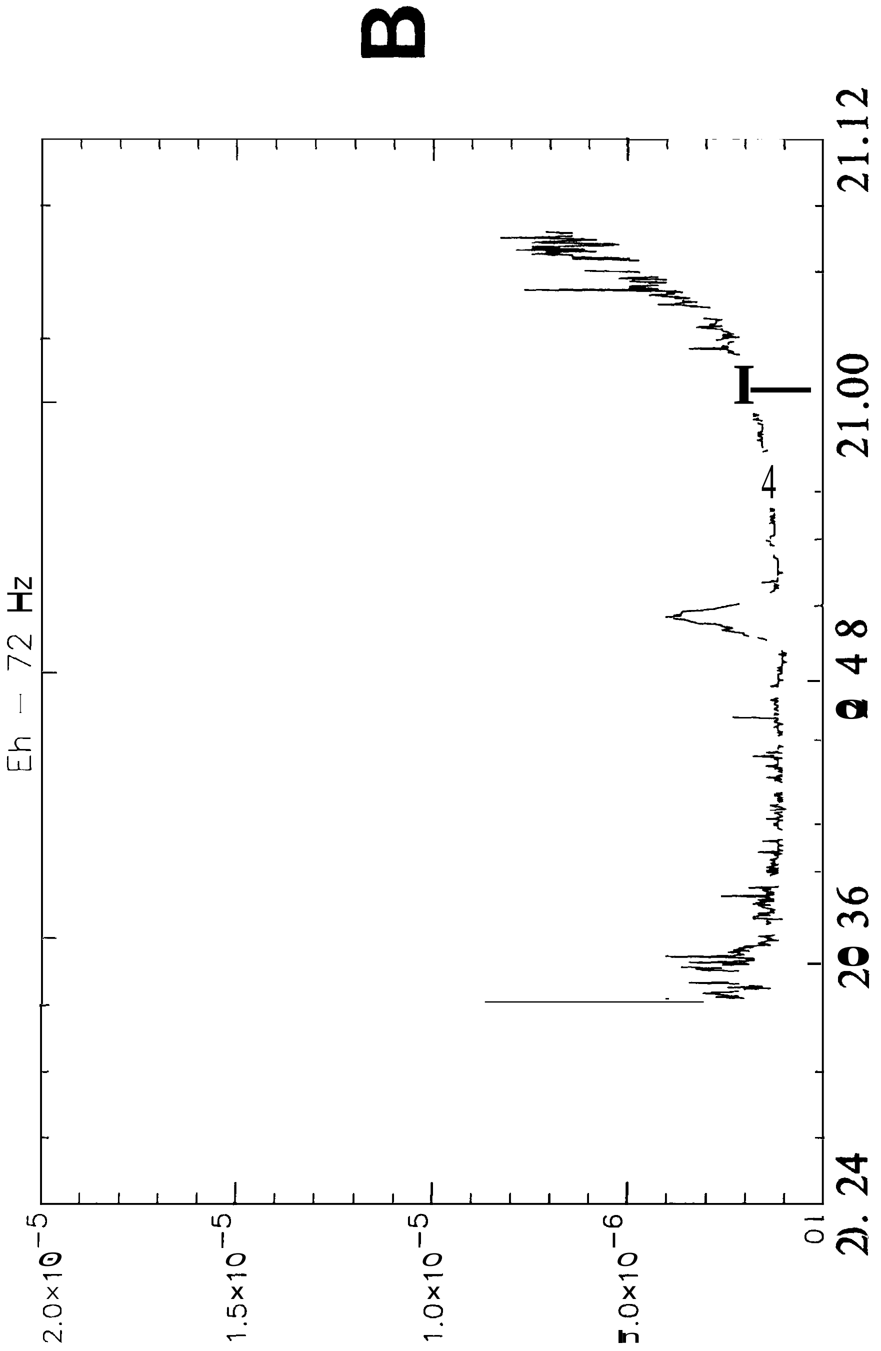

Fig. (continued)

by which electromagnetic emissions can be produced by earthquakes, and two main mechanisms are generally invoked.

The first concerns direct wave production by compression of rocks near the focal point. During laboratory experiments on rock fracturing, several significant emissions have been obtained in the following frequency range:

- 1–10 MHz (Nitsan, 1977),
- 500 kHz–10 MHz (Warwick et al., 1982),
- 0.3–300 kHz (Martelli & Cerroni, 1985),
- 10 Hz–30 kHz (Ogawa, Oike, & Miura, 1985),
- 0–5 kHz (Cress, Brady, & Rowell, 1987).

However, if it is assumed that the source of the electromagnetic waves is located at the focal point of an earthquake, a major problem is posed by wave attenuation, which is proportional to

$$\exp[-(\pi f \mu \sigma)^{1/2} x], \tag{1}$$

where x is the distance from the generation zone, f is the wave frequency, μ is the magnetic permeability, and a the electric conductivity. Such attenuation in dB is shown in Figure 2 as a function of f and a, and for $x = 10$ km. It can be observed that for common values of σ (for the rocks, from 10^{-2}–10^{-5} S/m), only waves with very low frequencies (a few Hz) can reach the surface.

The second mechanism for the wave generation by earthquakes is related to a redistribution of the electric charges in the Earth's atmospheric system, which can produce electrical discharges. An earthquake can generate electric charges in different ways:

- *By compression of the rocks:* it is known that the movement and compression of small fragments of rocks produce electric charges by piezoelectric and/or triboelectric effects (Finkelstein & Powell, 1970; Finkelstein, Hill, & Powell, 1973).
- *By the diffusion of water inside the ground:* in the focal zone, under compression, the groundwater flow through rocks can produce electrokinetic interactions between the fluid and the rock pores (Mizutani, Ishido, Yokokura, & Ohnishi, 1976).
- *By the diffusion of radon:* with a simplified model of the atmospheric electric field, Pierce (1976) has shown that an increase of radon in the atmosphere can change its conductivity.

However, charge accumulation as a mechanism that could explain electromagnetic emissions from earthquakes is bedevilled by the problem that the electric field generated by the electrostatic charges decreases as

$$\exp(-\sigma t/\epsilon), \tag{2}$$

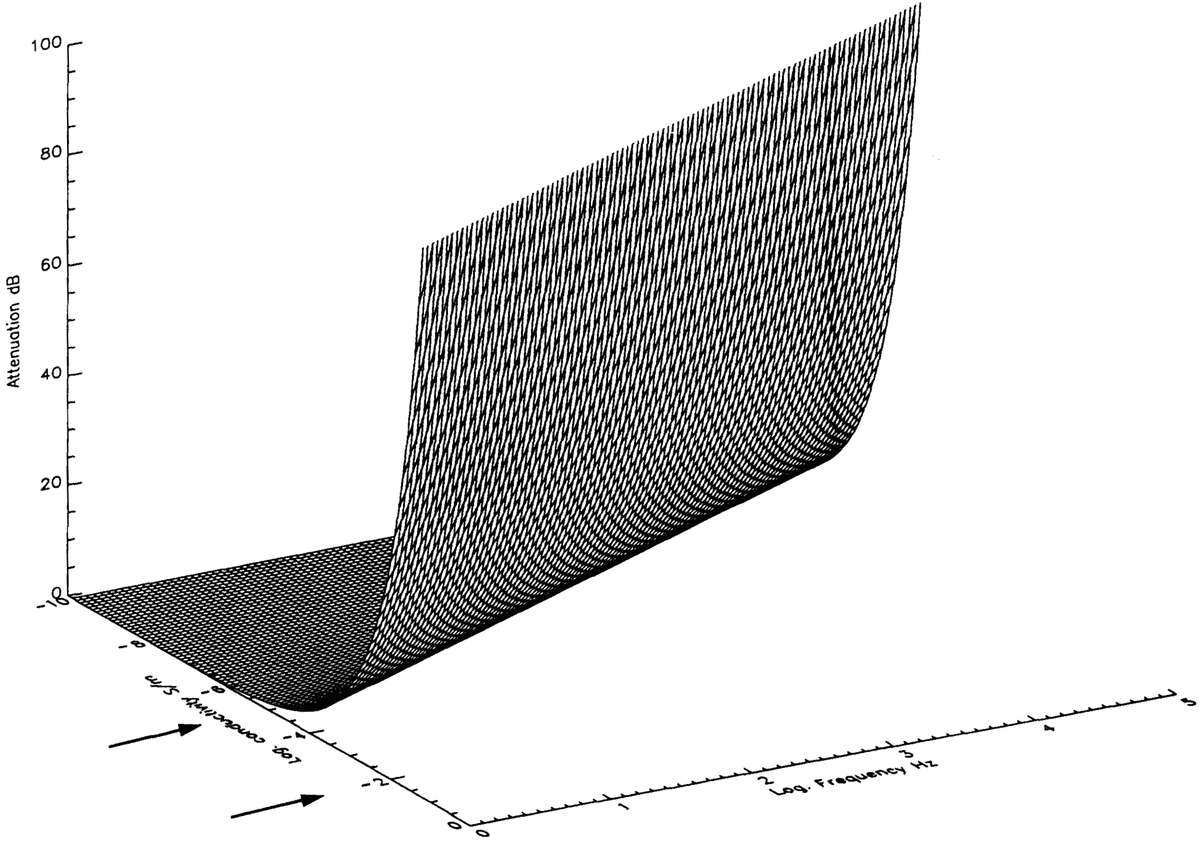

Fig. 2. Attenuation in dB of a wave emitted at a depth of 10 km. **as** a function of the wave frequency and the ground conductivity.

(where ϵ is the electric permittivity) and that the time constant ϵ/σ is too short (1 μs) to allow a charge accumulation.

A possible answer to the problems raised by the Equations (1) and (2) has been proposed-by Lockner, Johnston, and Byerlee (1983). They claimed that, under certain conditions, there can be strong heating along a fault due to the compression, and that in a very small zone around this, a can decrease by several orders of magnitude. These results have been discussed by Lee and Delaney (1987).

Up to now, most of the correlations of electromagnetic waves with earthquakes have been observed with experiments not dedicated to this study; for a better understanding of such phenomena, more observations are needed, but above all observations with more parameters measured at the same time. The EM spectra of the electric and the magnetic fields must be measured over a large frequency range, because, until now, only a restricted range was observed, such as the ULF range (Fraser-Smith, Bernardi, McGill, Ladd, Helliwell, & Villard, 1990), or the MHz range (Warwick et al., 1982). The other parameters to measure are: the ground resistivity (Rikitake & Yamazaki, 1985), the telluric current (Varotsos & Alexopoulos, 1984), the ionospheric density (Alimov, Gokhberg, Liperovskaia, Gufeld, Liperovsky, & Roubtov, 1989), the night airglow (Fishkova, Gokhberg, & Pilipenko, 1985), and the radon concentration (Pierce, 1976).

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