

# Non-inductive components of electromagnetic signals associated with L'Aquila earthquake sequences estimated by means of inter-station impulse response functions

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**Abstract.** On 6 April 2009 at 01:32:39 UT a strong earthquake occurred west of L'Aquila at the very shallow depth of 9 km. The main shock local magnitude was  $M_l = 5.8$  ( $M_w = 6.3$ ). Several powerful aftershocks occurred the following days. The epicentre of the main shock occurred 6 km away from the Geomagnetic Observatory of L'Aquila, on a fault 15 km long having a NW-SE strike, about  $140^\circ$ , and a SW dip of about  $42^\circ$ . For this reason, L'Aquila seismic events offered very favourable conditions to detect possible electromagnetic emissions related to the earthquake. The data used in this work come from the permanent geomagnetic Observatories of L'Aquila and Duronio. Here the results concerning the analysis of the residual magnetic field estimated by means of the inter-station impulse response functions in the frequency band from 0.3 Hz to 3 Hz are shown.

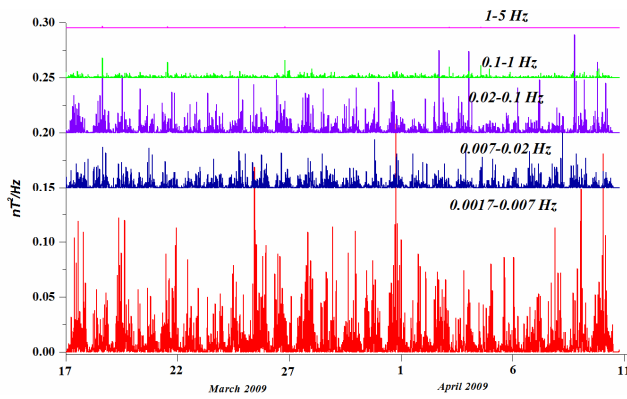
in which the signal-to-noise ratio is more favorable for the observation of magnetic signals of tectonic origin. These investigations are mainly concentrated during the main phase of the earthquake when the seismogenic signals are able to reach maximum amplitude. The analysis presented in this paper uses data sampled at higher frequency (10 Hz) measured at the magnetic station of the European MEM Project installed close to L'Aquila observatory in 2006 (Palangio et al., 2009), located inside the seismogenic area. The reference station of Duronio (Karakelian et al., 2000) is located outside the seismogenic region, 130 km away from L'Aquila. Our analysis is based on differential measurements between the two permanent observatories of L'Aquila and Duronio in order to minimize the contamination from multiple sources, such as the local background noise and the magnetic field of external origin. Many experimental and theoretical studies on the electromagnetic phenomena associated with earthquakes in the frequency range from ULF to HF have been reported, see Varotsos et al., 1984a, b, Bernardi et al., 1991; Fenoglio et al., 1995; Fraser-Smith et al., 1990, 1993; Gershenson et al., 1989; Gokhberg et al., 1982; Hayakawa et al., 1996; Johnston et al., 1987, 1989, 1994, 1997; Merzer et al., 1997; Molchanov et al., 1992, 1995; Nagano et al., 1975; Parrot et al., 1989; Park et al., 1991, 1993; Palangio et al., 2007, 2008, 2009. These studies focus on different aspects including variations in quasi-static electric and magnetic fields, telluric potentials, ULF magnetic fields, alternating electric fields in the ULF, ELF and VLF bands, and variations in the ground resistivity. Laboratory experiments performed by several scientists, in order to better understand the mechanism producing electromagnetic anomalies, showed that the rocks emit electromagnetic radiation when crushed (Ogawa et al., 1985; Sasaoka et al., 1998; Cress et

## 1 Introduction

Extensive investigations were conducted by the University of L'Aquila in ULF band (0.001 Hz–0.2 Hz) to search for magnetic anomalies associated with the earthquake of L'Aquila (Villante et al., 2009). The authors have studied the magnetic signals recorded in the ULF station of L'Aquila University located near the INGV geomagnetic Observatory. The results of these studies do not support the existence of any magnetic anomalies associated with the main shock and aftershocks. The present work aims to extend the investigation by identifying both the temporal and spectral windows



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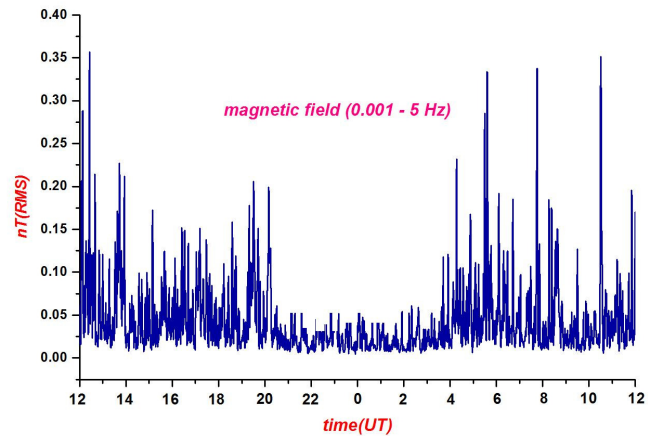


**Fig. 1.** Continuum magnetic background noise at L'Aquila Geomagnetic Observatory.

al., 1987; Frid et al., 2000). In these papers the authors show that in the tectonic motion of faults responsible for earthquakes, the Earth's crust responds with impulsive EM events which span a broad range of frequencies. These general aspects of the dynamics of the crust, irrespective of the physical mechanism and details of the system, when perturbed by a slowly varying stress, are always present in the proximity of and during the earthquake. An earthquake can generate electric charges in different ways: by compression of the rocks through the piezoelectric or triboelectric effects and by the diffusion of fluids inside the ground. The groundwater flowing through the rocks could produce electrokinetic interactions between the fluid and the rock pores. Another generation mechanism of signal emission proposed by Varotsos and Alexopoulos are the PSPC (Pressure Stimulated polarization Currents), (Varotsos et al., 1993, 1998), (Uyeda et al., 2009). Another interesting model of generation of electric current was proposed by Freund and his co-workers (Freund, 2002, 2003, 2007) named P-Holes theory which took strong low frequency electromagnetic emissions reported in other published papers into account. Kopytenko et al. (1993) and Johnston (1997) show that the detection of seismogenic signals in the extreme frequency band would require however surface measurement systems to be very close to the epicenter of the earthquake.

## 2 Local electromagnetic background noise

The magnetic noise from both natural and man-made sources is the main source of the interference limiting the discrimination of signals of seismogenic origin. To achieve the distinction between true precursory signals and noise, a procedure based on the natural time concept has been recently proposed (Varotsos et al., 2005; 2006a, b). Here we make use of the conventional inter-station impulse response functions method. It is common knowledge that there exists a frequency band for which a compromise between the sig-



**Fig. 2.** Time window, diurnal variation of background noise.

nal attenuation through the earth and the background noise level in the frequency and time domain is reached (Dea et al., 1993). So the role of background noise is crucial in the research of seismogenic signals. The anthropogenic electromagnetic noise, such as power lines, DC railways, factories, etc., generates signals whose amplitude is often higher than those of tectonic origin and in the same frequency band (Lanzerotti et al., 1990; Fraser-Smith et al., 1975, 1978). These sources of noise which vary in frequency and time, are local in nature, so they could be difficult to distinguish from anomalous signals of tectonic origin (Fraser-Smith et al., 1978). Both background local noise and the signals of external origin are characterized by a large diurnal periodicity with a remarkable consistency of phase. A partial discrimination between the two contributions can be made only when there is a change in daylight saving time because the noise goes with the local time while the external signals are related to UT time. In order to explore the possibility to detect seismogenic magnetic signals emitted from an earthquake source, it is very important to identify the most suitable time and frequency window by means of long and continuous records of the field in the frequency band of interest. Figure 1 shows a typical feature of amplitude distribution of ULF geomagnetic activity, the signals are filtered into five frequency bands from 0.0017 Hz to 5 Hz. This background noise arises from several contributions. Each contribution increases with the decreasing frequency as shown in Fig. 1. At L'Aquila Geomagnetic Observatory the weakest background noise level occurs between 21:00 and 03:00 UT (Fig. 2). In this time interval the noise is much lower than during the daytime, about ten times, instead in the lowest frequency band, around 1 mHz, the day-night ratio is about 50. The spectral density of the noise reaches the minimum in the frequency band from 0.3 Hz to 3 Hz. In this frequency band the noise level is of the same order of magnitude as instrumental noise, furthermore the spectral properties take on the characteristics of white noise (Fig. 3). This frequency range is

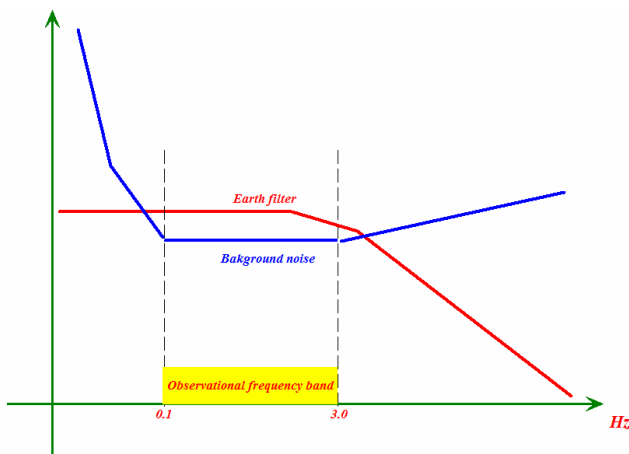


Fig. 3. Spectral window.

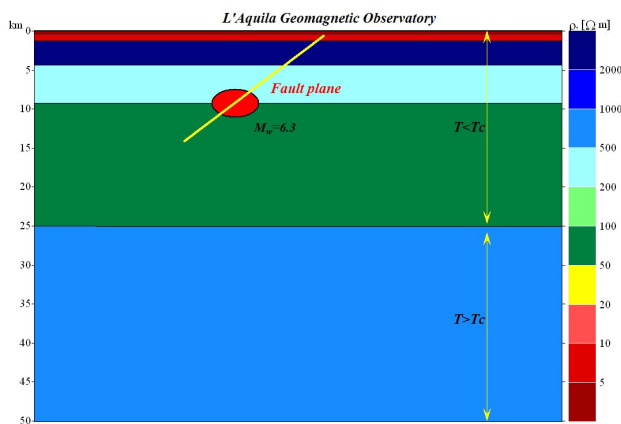


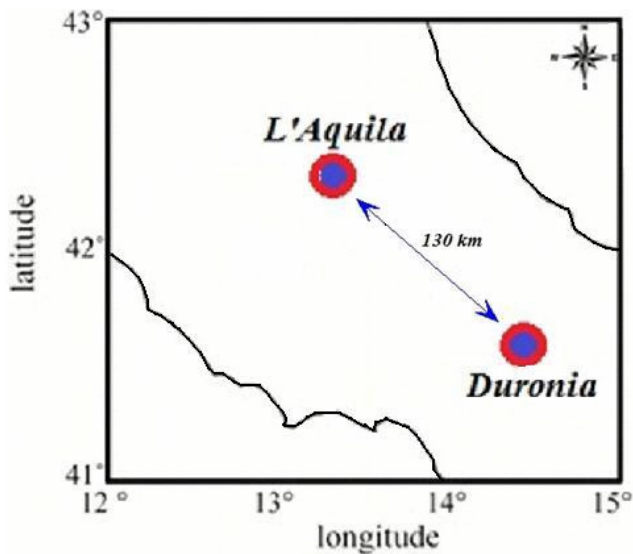
Fig. 4. Local earth resistivity structure around L'Aquila Geomagnetic Observatory.

dominated by PC1 pulsations of magnetospheric origin and by IAR signals (Ionospheric Alfvén Resonator). Both frequency and time windows are influenced by the global magnetic activity. For  $K_p < 2$  the frequency “window” occupies the frequency band 0.02–5 Hz and the interval of local time around 24:00 UT. During daytime, the frequency range of the “window” is narrower and occupies the 0.05–5 Hz frequency band. For  $K_p > 2$ , the lower frequency boundary of the “window” increases up to 0.1–0.2 Hz and becomes independent on local time. Therefore the frequency window 0.3 Hz–3 Hz is almost completely independent on  $K_p$  index.

### 3 Conductivity structure of the earthquake area

The knowledge of the underground resistivity structure is an essential requirement for the study of the electromagnetic manifestations linked to earthquakes. The starting point of these studies is based on the knowledge of the electric prop-

erties of the materials present in the focal zone of the earthquake by means of a stable estimate of the Earth’s conductivity structure. Assuming that the source of the geogenic field is located at hypocentral depth and that this source can be represented by a magnetic dipole, we evaluated the cut-off frequency below which seismogenic signals are not attenuated, employing the magnetotelluric method in order to build a simple model of the subsurface resistivity. In Central Italy the major tectonic activity is concentrated in the underground depth range of about 5–15 km. This electromagnetic skin depth sets the scale for the useful depth of exploration. Our magnetotelluric investigations were extended to a depth of 50 km. In the area of Central Italy the direction of the active faults is roughly NW-SE. By means of a permanent magnetotelluric station located close to the L’Aquila Observatory, we performed continuous measurements from 2004. Figure 4 shows the ground electric resistivity profile calculated for L’Aquila station by means of the single station magnetotelluric tensor evaluation. The 1-D profile is obtained using a conventional magnetotelluric approach. To obtain the resistivity profile we used a standard Occam 1-D inversion code (Constable et al., 1987), using both the apparent resistivity and the phase for the inversion. The profile shown in Fig. 4 was obtained using only the two horizontal components of the magnetic field and the two horizontal components of the telluric field. The earth resistivity structure model shown in Fig. 4 represents the average elaborations calculated over several years from 2004 to 2008. It is assumed that the medium is formed of multiple layers of horizontally stratified materials. The hypocenter of the earthquake is located at a depth of 9 km between two layers with different conductivity properties. The layer which contains the earthquake source has a resistivity of 500  $\Omega \cdot m$  located on the top of the lower resistivity layer of 100  $\Omega \cdot m$ . This low resistivity layer, below 10 km from surface, extends till 25 km depth. Based on this model, we have estimated the integrated resistivity in the zone extending from the surface to the hypocentral depth of 9 km. We calculated the expected attenuation of the magnetic signals generated in the hypocentral area in the frequency band where the local background noise is lower. We used a simplified three layer conductivity model for the region, which includes the observatory and the earthquake area consisting of a top layer 2 km thick with a resistivity of about 5  $\Omega \cdot m$ , a lower layer 3 km thick with a resistivity of 3000  $\Omega \cdot m$  and a bottom layer with a resistivity of 500  $\Omega \cdot m$ , so the soil was considered as a homogeneous isotropic medium characterized by an integrated resistivity  $\sum \rho \approx 1200 \Omega \cdot m$ . The cut-off frequency of the earth filter modelled in this way lies around 3 Hz, so that the energy of seismogenic emission will be able to be transmitted from the source depth to the Earth’s surface with very little attenuation for frequencies below 3 Hz (Fig. 3). The time response  $\tau \approx \mu \zeta^2 \sum \sigma$  of the source due to the diffusion time within the crustal medium is of the order of 0.8 s. ( $\zeta$  is the crustal depth and  $\sum \sigma$  is the integrated conductivity)



**Fig. 5.** Geographical location of the two observatories AQU and DUR.

Another relevant aspect is the temperature. In the upper portions of the crust it is well below the Curie temperature. It is expected that in the focal zone where the temperature is less than 300 °C, rocks tend to behave as brittle bodies, especially when they are nearly dry. Therefore we cannot exclude that piezomagnetic phenomena have developed in the focus of this earthquake and have contributed to the genesis of the observed signals.

#### 4 Data analysis

The vector components of the geomagnetic field were measured continuously at the two Italian permanent geomagnetic observatories (Fig. 5), situated at L'Aquila (42°23' N, 13°19' E, 682 m a.s.l.) and Duronia (41°39' N, 14°28' E, 910 m a.s.l.). Duronia is located outside the seismogenic region, 130 km away from L'Aquila. The magnetic signals were sampled at 10 Hz. The peculiarity of the of Duronia Observatory is the low electromagnetic background noise of the site and the low noise of the instrumentation used for the measurements. For example, in the frequency band from 0.1 Hz to 40 Hz the background magnetic noise level is particularly low, less than  $20\text{fT}/\sqrt{\text{Hz}}$ . In the study of seismogenic fields, the necessity arises to separate the weak inhomogeneous magnetic fields produced by local sources from the background noise, from the inductive signals, and from the geomagnetic field of external origin, largely governed by the activity of the sun. The problem of separating a signal from noise when the noise level is tens of times higher than the signal level, can be solved only with differential measurements. It is important to evaluate the scale length and the distance from the measurement points of the var-

ious sources involved. External sources have a large spatial extent and thus produce uniform fields on spatial extensions up to about 100 km. However, over these distances there are small gradients that vary during the day. Therefore the simple differences between AQU and DUR are not enough to extract the weak seismogenic signals measured at L'Aquila station efficiently because the spatial gradient between AQU and DUR is of the same order of magnitude as the signal of internal origin. The scale length of the noise signals is of the order of tens of km while the scale length of seismogenic signals for the L'Aquila earthquake is less than 10 km and the distance between the measurement station and the hypocentral point is of the same order. So in terms of electromagnetic induction of the three regions (noise, external and geogenic), we can identify three spatial regions:  $L_{\text{ext}} > 2\delta$ ,  $1/2\delta < L_{\text{noise}} < 2\delta$ ,  $L_{\text{seis}} < 1/2\delta$ . Where  $\delta = \left(\frac{2\Sigma\rho}{\omega\mu}\right)^{1/2}$  in the first region the area of the earthquake is in the far field zone: therefore the measured signal is due to superposition of signals coming from external sources and signals of inductive origin.

In the second region, considering the artificial noise, the area of the earthquake is located in the transition zone in which the measured signal is given by the overlap of the signals generated by the noise sources and signals partially induced.

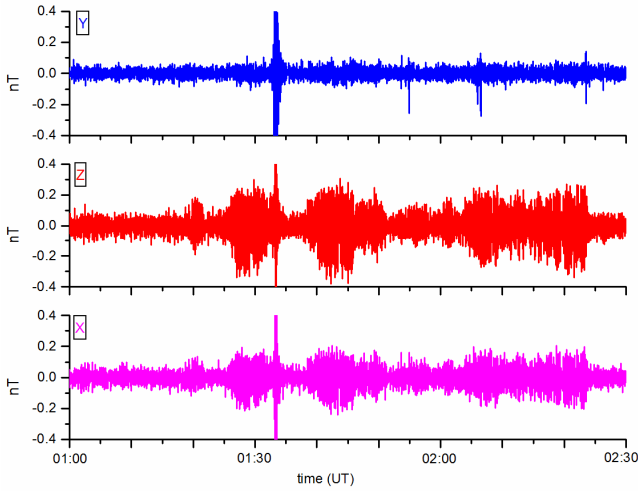
In the third region, considering the seismogenic signals, the area of the earthquake is in the near field zone where the seismogenic signals do not lead to significant inductive effects. The measured signals are those emitted from the source without any contamination of the inductive nature. In this case it should be possible to study the internal structure of the seismogenic source by means of a measurement array around the seismic area.

In order to discriminate the seismogenic signals we have used the standard interstation impulse response functions. This methodology takes into account the following circumstances:

1. inductive effect;
2. differences between the transfer functions of the magnetic sensors;
3. non collinearity of the sensor axes;
4. instrumentation noise;
5. local background noise;
6. differences between the resistivity structure of DUR and AQU

Luckily, the geomagnetic activity during the earthquake was rather low. Figure 4 shows the Kp index of geomagnetic activity, and it is clear that the value of this index was not larger than 6 nT during the day 6 April 2009. We can therefore calculate the magnetic signals at L'Aquila from the recorded





**Fig. 6.** Residual field deperated from inductive and external fields.

signals at Durlonia Geomagnetic Observatory regarded as inputs. In the present case, we used the inter station transfer function approach and found that it is effective in removing the known, predictable magnetic signals from the observed data at L'Aquila.

$$X_A = \int_0^\infty I_{xx}(\tau) X_D(t - \tau) d\tau + \int_0^\infty I_{xy}(\tau) Y_D(t - \tau) d\tau + \int_0^\infty I_{xz}(\tau) Z_D(t - \tau) d\tau \quad (1)$$

similar expressions for the other two components, which in discrete terms becomes:

$$X_{Ai} = \sum I_{xxj} X_{i-j} + \sum I_{xyj} Y_{i-j} + \sum I_{xzz} Z_{i-j} \quad (2)$$

from which we can calculate the nine impulse functions  $I_{kl}$  using linear least squares method (Swanson et al., 1997).

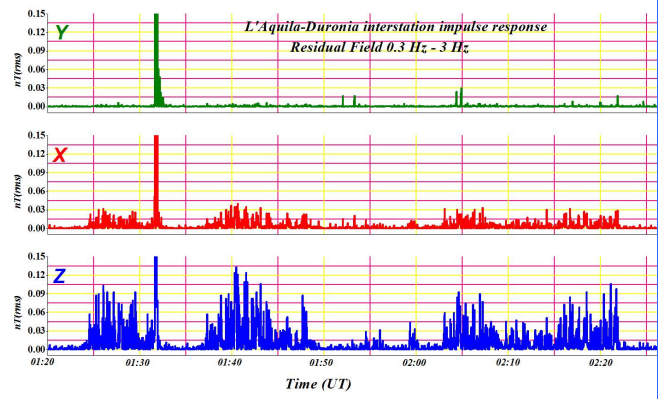
The full expression is:

$$\begin{pmatrix} X_a(t) \\ Y_a(t) \\ Z_a(t) \end{pmatrix} = \begin{pmatrix} I_{xx}(\tau) & I_{xy}(\tau) & I_{xz}(\tau) \\ I_{yx}(\tau) & I_{yy}(\tau) & I_{yz}(\tau) \\ I_{zx}(\tau) & I_{zy}(\tau) & I_{zz}(\tau) \end{pmatrix} \otimes \begin{pmatrix} X_d(t) \\ Y_d(t) \\ Z_d(t) \end{pmatrix} \quad (3)$$

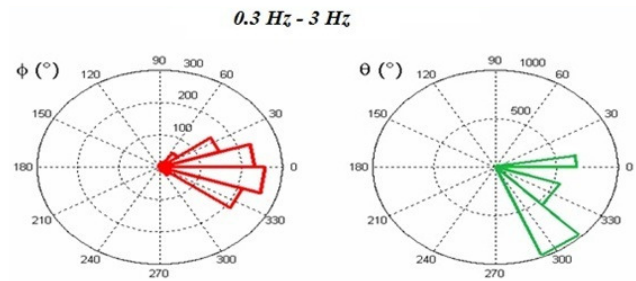
So the residual field is:

$$\begin{aligned} X_r(t) &= X_{ma}(t) - X_a(t) \\ Y_r(t) &= Y_{ma}(t) - Y_a(t) \\ Z_r(t) &= Z_{ma}(t) - Z_a(t) \end{aligned} \quad (4)$$

Where  $X_{ma}$ ,  $Y_{ma}$  and  $Z_{ma}$  is the field measured at AQU.  $X_d$ ,  $Y_d$ , and  $Z_d$  is the field measured at DUR.  $X_a$ ,  $Y_a$  and  $Z_a$  is the field predicted at AQU.  $X_r$ ,  $Y_r$  and  $Z_r$ , are the residual field components. We have estimated with high accuracy the nine elements of the impulse matrix  $\mathbf{I}_{ij}$  before the earthquake as a mean over several months before. These functions are considered to be invariant in time, Fig. 6 shows the residual field at L'Aquila geomagnetic Observatory located 6 km



**Fig. 7.** RMS representation of the residual field in the frequency band from 0.3 to 3 Hz.



**Fig. 8.** The arrival direction of the anomalous signal.

away from the epicenter. The signals are emitted 10 min before the earthquake and during the event till 50 min after. The maximum amplitude of the signals before the earthquake is about 100–200 pT rms, in the frequency band from 0.3 Hz to 3 Hz. In this frequency band our calculations are based on simple diffusion of the signals through the ground. The directions of incidence of the signals (Fig. 8) are well focused in the direction of the hypocenter. We believe that it is unlikely that these signals may have been generated by piezoelectric or triboelectric phenomena. Because of the heterogeneity of the rocks in the Earth's crust, the quartz crystals are randomly oriented, the dipole fields cancel each other partially, so that a long-range field is not generated. Simple consideration suggests that  $n$  aligned dipoles generate a total dipole moment  $nM_i$ , assuming  $M_i$  all equal, while  $n$  randomly oriented dipole generate a total dipole moment  $M_i \sqrt{n}$  (thermal approximation). Figure 11 shows the power spectrum of the residual field. Signals were selected before and after the co-seismic signals in order to isolate the signals of possible tectonic origin from the signals produced by ground motion. From this figure it is clear that the energy of the signals is concentrated in the spectral region close to the Nyquist frequency. We believe that this could be due to an impulsive character in the magnetic field source signals. The observed signal is the convolution of the source-time-functions and the

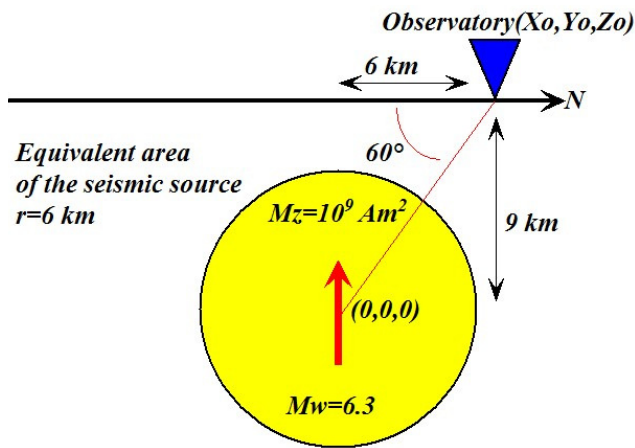


Fig. 9. Simple model of magnetic source generating the observed signals.



Fig. 10. Distribution of Earthquakes from 1 April to 20 April 2009 around L'Aquila.

Earth's impulse response functions. The knowledge of the Earth's response functions is a fundamental point for these kind of investigations.

### 5 Modeling of the source

In this simple model it is assumed that a magnetic dipole should be placed at a depth of 9 km below the Earth's surface on the top of high conductive layer ( $100 \Omega \cdot \text{m}$ ) and inside a low conductive layer ( $500 \Omega \cdot \text{m}$ ), whose relative permittivity is about 5 and relative permeability is 1. The overall size of the underground electromagnetic source might be related to the seismic source, which in terms of equivalent radius of the source is  $r_e = \sqrt{\frac{M_w}{\pi \mu D}} \approx 6 \text{ km}$  where  $M_w$  is the seismic moment,  $\mu$  is the rigidity modulus of the rocks involved in the earthquake and  $D$  is the average displacement along the

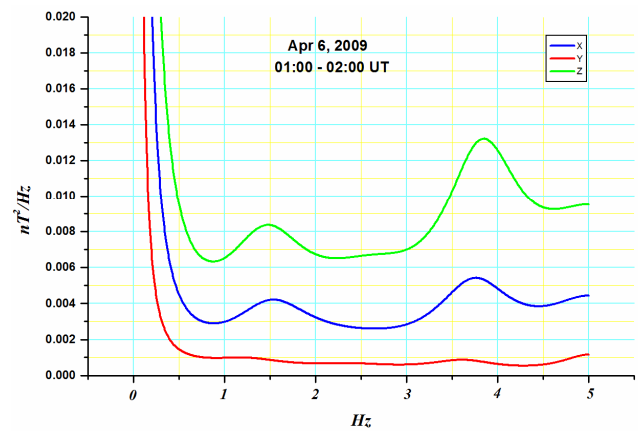


Fig. 11. Power spectrum of the residual field of seismogenic origin.

seismogenic fault. The size of the source is of the same order of magnitude as the distance between the source and the measuring station. Because the wavelength of the signals in the frequency band 0.3 Hz–3 Hz, which is related to the skin depth, extending from 8 km to 24 km.

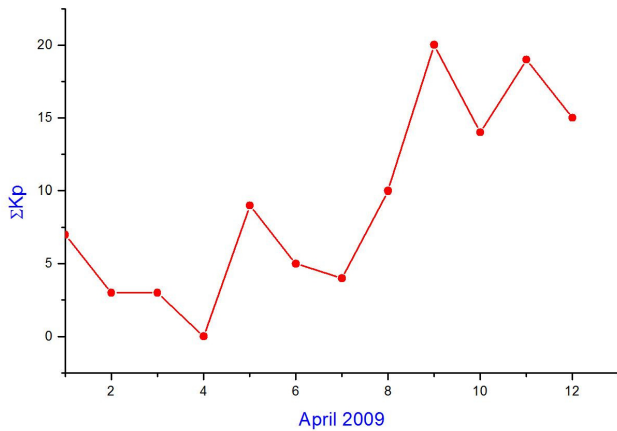
Irrespective of the different mechanisms of the electro-mechanic energy conversion which could generate the observed fields, the ipogeic EM source can be considered as a complex system containing both toroidal and poloidal components of current and fields; inside the source, the fields are described by the well known relation:

$$\nabla \times \mathbf{J}^T \sum \rho = -\frac{\partial \mathbf{B}^P}{\partial t} \quad \nabla \times \mathbf{J}^P \sum \rho = -\frac{\partial \mathbf{B}^T}{\partial t} \quad (5)$$

While on the surface of the Earth, at the measurement station:

$$\mathbf{J}^{P,T} = 0, \quad \nabla \times \mathbf{B}^P = 0 \quad \text{and} \quad \nabla \cdot \mathbf{B}^P = 0 \quad (6)$$

Where  $\mathbf{J}^P$  and  $\mathbf{B}^P$  are the poloidal components of the fields,  $\mathbf{B}^T$  and  $\mathbf{J}^T$  are the toroidal components.  $\mathbf{J}$  is the current density and  $\sum \rho$  is the integrated resistivity. Of course we do not have any knowledge on the spatial distribution of the source density current  $\mathbf{J}(x, y, z, t)$  in the surrounding ground, nor on the mechanism of currents generation. This simple model is based on the measured magnetic field and on the knowledge of the resistivity structure of the earth. The model calculation is performed in the frequency band from 0.3 Hz to 3 Hz in which the wavelength is larger than the characteristic size of the earthquake, so the measurements are essentially being made of diffusive fields in the so-called “near field” or “quasi-static”. Indeed the size of the magnetic diffusion zone  $\Delta S = Df^{-1/2}$ , which is related to the magnetic diffusion coefficient  $D$  ( $D = (\mu\sigma)^{-1/2} \approx 10^5 \text{ m} \times (\text{s})^{-1/2}$ ), is of the order of 50–100 km, this is much larger than the hypocentral depth. Therefore magnetic diffusion is the dominant factor.



**Fig. 12.** Diagram of the Kp index during the L’Aquila earthquake sequences.

In general the magnetic field generated by a magnetic dipole with  $M_x$ ,  $M_y$  and  $M_z$  components is:

$$\begin{aligned}
 B_x^p &= \left[ \frac{3(x_0M_x + y_0M_y + z_0M_z)x_0}{L^5} - \frac{M_x}{L^3} \right] \frac{\mu}{4\pi} \\
 B_y^p &= \left[ \frac{3(x_0M_x + y_0M_y + z_0M_z)y_0}{L^5} - \frac{M_y}{L^3} \right] \frac{\mu}{4\pi} \\
 B_z^p &= \left[ \frac{3(x_0M_x + y_0M_y + z_0M_z)z_0}{L^5} - \frac{M_z}{L^3} \right] \frac{\mu}{4\pi}
 \end{aligned} \tag{7}$$

Where  $B_x^p$ ,  $B_y^p$ ,  $B_z^p$  are the poloidal components of the measured field,  $L$  is the distance between the measuring station and the hypocenter.  $x_0$ ,  $y_0$  and  $z_0$  are the coordinates of the measurement station, the hypocenter is located at the origin of the coordinates. In general, the possible orientation of the equivalent dipole vector is determined by casual circumstance, such as the medium heterogeneity in the conductivity distribution, asymmetry of the crackness development and so on. In our case, the direction of the magnetic dipole to produce the measured magnetic field is approximately in the vertical direction. From simple calculation as  $B_y \approx 0$ ,  $M_z > M_x > M_y$ , so the direction of the total magnetic moment is approximately vertical with a small component in the north-south direction. Therefore the possible source of the magnetic signals observed on the earth surface should be due to an electric current flowing around the focal volume mainly in the horizontal plane. The horizontal component of these electric currents is fed into the high conductive substrate and partly in the fracture plane tilted about  $42^\circ$  from the horizontal plane. Assuming that the equivalent diameter of the source is of about 12 km, the density of this electric current which flows around the focal area should be less than  $10 \text{ mA m}^{-2}$ . The total magnetic moment of the dipole is of the order of  $10^9 \text{ Am}^2$ . The total magnetic energy observed is of the order of 5–6 MJoule, so only few parts in a billion of the total energy of the earthquake have turned into magnetic energy.

**Table 1.** List of earthquakes from 30 March 2009 to 23 June 2009 with local magnitudes of  $M_l > 3.9$ .

Time (UTC)	Lat.	Long.	Depth	MI
23/06/2009 00:41:56.180	42.444	13.369	14.9	4.0
22/06/2009 20:58:40.270	42.445	13.354	13.8	4.6
23/04/2009 21:49:00.840	42.228	13.486	9.7	4.2
23/04/2009 15:14:08.310	42.247	13.484	10.3	4.0
18/04/2009 09:05:56.280	42.436	13.359	14.5	4.0
16/04/2009 17:49:30.180	42.535	13.291	11.5	4.1
15/04/2009 22:53:07.560	42.515	13.330	9.8	4.0
14/04/2009 20:17:27.160	42.526	13.298	10.3	4.1
14/04/2009 13:56:21.210	42.542	13.320	9.9	4.0
13/04/2009 21:14:24.470	42.498	13.377	9.0	5.0
09/04/2009 19:38:16.960	42.504	13.350	9.3	5.0
09/04/2009 13:19:33.830	42.341	13.259	9.7	4.1
09/04/2009 04:43:09.600	42.502	13.373	9.6	4.0
09/04/2009 04:32:45.050	42.445	13.434	9.8	4.2
09/04/2009 03:14:52.260	42.335	13.444	17.1	4.6
09/04/2009 00:52:59.690	42.489	13.351	11.0	5.1
08/04/2009 22:56:50.190	42.497	13.367	10.8	4.2
07/04/2009 21:34:29.770	42.364	13.365	9.6	4.3
07/04/2009 17:47:37.340	42.303	13.486	17.1	5.4
07/04/2009 09:26:28.610	42.336	13.387	9.6	4.8
06/04/2009 23:15:36.760	42.463	13.385	9.7	5.0
06/04/2009 16:38:09.730	42.363	13.339	10.0	4.1
06/04/2009 07:17:10.140	42.356	13.383	9.0	4.0
06/04/2009 03:56:45.700	42.335	13.386	9.3	4.1
06/04/2009 02:37:04.250	42.360	13.328	8.7	4.6
06/04/2009 01:42:49.970	42.300	13.429	10.5	4.2
06/04/2009 01:41:37.770	42.364	13.456	8.7	4.3
06/04/2009 01:41:32.690	42.377	13.319	8.5	4.0
06/04/2009 01:40:50.650	42.417	13.402	11.0	4.1
06/04/2009 01:36:29.190	42.352	13.346	9.7	4.7
06/04/2009 01:32:40.400	42.342	13.380	8.3	5.9
30/03/2009 13:38:38.960	42.321	13.376	9.8	4.1

## 6 Conclusions

The primary purpose of this work was to discriminate the extremely feeble magnetic signals originating during the earthquake of L’Aquila from those coming from other magnetic sources by means of the inter-station impulse response functions between L’Aquila Geomagnetic Observatory and Durnia Observatory 130 km away from L’Aquila. In order to explore the possibility to detect seismogenic magnetic signals emitted from earthquake source and avoid contamination from other sources, we limited our analysis to a well-defined temporal and spectral window. We have shown that in these time and frequency domains, there is a maximum chance of detecting magnetic signals of seismogenic origin even if their amplitude is very small, consistent with the sensitivity of the instrumentation used. The weakest background noise level occurs between 21:00 and 03:00 UT, the

earthquake occurred just during this time interval (01:32), therefore during the earthquake we should have the highest probability of measuring the seismogenic signals. (Johnston, 1997). Data analyzed from L'Aquila earthquake suggest that the magnetic field is about an order of a magnitude smaller than that reported in other published papers, considering the magnitude, the hypocentral depth and distance from the measuring station. During the entire duration of the earthquake, the emissions were only observed just a few minutes before and during the arrival of the first P seismic waves of the earthquake, and another burst was observed after the main phase, which lasted about 50 min. The amplitude of the signals is of the order of 100–200 pT. Figure 8 shows the arrival direction of the anomalous signal calculated from the spectral magnetic tensor.  $\Phi$  is the azimuthal angle,  $\Phi = 0$  indicates the North direction.  $\theta$  is the zenithal angle,  $\theta = 0$  indicates the surface of the Earth in the South direction. This figure shows the existence of a group of signals coming from the direction of the epicenter that are clearly separated from signals from other sources.

These signals observed at L'Aquila Geomagnetic Observatory in association with the earthquake come predominantly from the direction of the hypocenter-measurement station. This estimate is based on diffusion of the signals through the ground. The results of our analysis do not support the existence of any magnetic signals associated with the foreshocks and aftershocks listed in Table 1 with local magnitudes  $M_l$  less than 5.3, that emerge clearly from the noise. In summary, these emissions do not give enough warning because they are too short in time. However these results do not preclude the possibility that the electromagnetic monitoring of seismogenic areas may help to understand the physical processes associated with earthquakes, especially those preceding the seismic activity in the preparatory phase. However, the reliability of these results is limited by the fact that the observations come from a single measurement station. We have no information about the spatial variation of the observed anomaly.

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