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Magnetic transfer function entropy and the 2009 $M_w = 6.3$ L'Aquila earthquake (Central Italy)

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Abstract. With the aim of obtaining a deeper knowledge of the physical phenomena associated with the 2009 L'Aquila (Central Italy) seismic sequence, culminating with a $M_w = 6.3$ earthquake on 6 April 2009, and possibly of identifying some kind of earthquake-related magnetic or geoelectric anomaly, we analyse the geomagnetic field components measured at the magnetic observatory of L'Aquila and their variations in time. In particular, trends of magnetic transfer functions in the years 2006–2010 are inspected. They are calculated from the horizontal to vertical magnetic component ratio in the frequency domain, and are very sensitive to deep and lateral geoelectric characteristics of the measurement site. Entropy analysis, carried out from the transfer functions with the so called transfer function entropy, points out clear temporal burst regimes of a few distinct harmonics preceding the main shock of the seismic sequence. A possible explanation is that they could be related to deep fluid migrations and/or to variations in the micro-/meso-fracturing that affected significantly the conductivity (ordered/disordered) distribution in a large lithospheric volume under the seismogenic layer below L'Aquila area. This interpretation is also supported by the analysis of hypocentres depths before the main shock occurrence.

1 Introduction

After a long seismic sequence which started in the first half of December 2008 and continued for several months, on 6 April 2009 at 01:32:40 UTC a large earthquake ($M_w = 6.3$) struck central Italy (42.34° N, 13.38° E, depth 9.5 km). The approximately 300 people who died and the severe damage suffered by L'Aquila, the major city very close to the epicentre and site of many precious medieval art masterpieces, shocked public opinion which thus questioned the scientific community about the present capability of scientific researchers to "predict" such a harmful event.

Actually, many efforts have been made by geophysicists to better understand the "earthquake" phenomenon, in particular the physics behind it, and many attempts have been made to make reliable diagnoses of the corresponding physical system, and to possibly uncover prior indications of an impending large earthquake. In doing so, many techniques have been used, sometimes taking advantage of knowledge acquired in fields far from seismology, but progress has been much slower than expected (Wyss, 2001).

Unfortunately, no deterministic predictions (exact time, place and magnitude of the impending earthquake) can be made at the present time (e.g. Geller et al., 1997), in the same way that no medical doctor can precisely "predict" the time of the death of his/her patient, whereas a reliable diagnosis (with some uncertainty) can be made. Analogously, it is possible to issue a more or less detailed "diagnosis" of the state of the Earth crust in a given area, and to assess therefore whether that area may be eventually more or less prone to suffer a seismic shock, and what the largest magnitude (with some uncertainty) might be. It is now generally established that there exist some preparatory stages that may culminate with an earthquake (Scholz, 1988; De Santis et al., 2011; Pulinets, 2011). Each stage is believed to be characterised by changes involving the physical parameters of large volumes of rocks around the sites where the earthquakes will occur. To mention only a few of the most prominent, changes in seismic velocities, electrical conductivity and crustal deformations may occur (Wyss, 1997). In order to explain such observed phenomena, many models have been proposed and among those most often referred to, there is the dilatancy model which represents essentially an increase in porosity (Brace et al., 1966; Brace, 1975).

In this work, we wish to test whether some changes in electromagnetic (e.m.) properties (e.g. the electrical conductivity, or its reciprocal, the resistivity) of the fault structure close to L'Aquila can be recognised, so that they can be indicated as a possible pattern for expecting future impending earthquakes in that area with higher probability than usual. Indeed, it is known that increasing tectonic stress before an event causes diffuse micro-fractures and that their propagation into the volume of possibly fluid saturated rock may alter the resistivity distribution of the medium (Niblett and Honkura, 1980). So, in principle, magnetovariational methods, such as geomagnetic deep sounding (GDS) and transfer function analysis (e.g. Everett and Hyndman, 1967; Gregori and Lanzerotti, 1982; Gregori et al., 1982; Banks, 2007) should be able to detect changes in the conductivity distribution of local crustal features related to earthquakes, such as seismogenic faults and other associated tectonic and geological structures.

Recently, some studies have been published highlighting the fundamental role of fluids in the space-time evolution of the L'Aquila seismic sequence. Lucente et al. (2010) reported a sudden change of the $V_\mathrm{p}/V_\mathrm{s}$ ratio between the seismic wave velocities and of the S-wave polarisation just after the onset of the large $M_\mathrm{L} = 4.1$ earthquake on 30 March 2009 that preceded the main shock by a week. These authors interpreted their results as having been caused by migration of fluids moving from the fault footwall, where they had previously been sealed, towards the hanging wall. From the analysis of the foreshock sequence (from October 2008 to 6 April 2009) and of more than 2000 aftershocks, all relocated by means of the double-difference method, Di Luccio et al. (2010) found an increase in the $V_\mathrm{p}/V_\mathrm{s}$ ratio due to the rise in "pore fluid pressure along the fault planes". In addition, Terakawa et al. (2010) applied the so-called focal mechanism tomography to the earthquakes in the same source region in order to infer the 3-D fluid pressure field at depth by analysing the focal mechanisms and the fault orientation relative to the regional tectonic stress pattern. They found that the fluid pressure was "near-hydrostatic" at the onset of the main shock, whereas it showed "significantly" higher values when the foreshocks and larger aftershocks occurred.

With this scenario in mind, it is reasonably expected that a change of conductivity structure beneath the L'Aquila region may have occurred. Thus, since the geomagnetic field at the Earth's surface is particularly sensitive to the change of conductivity, we apply the transfer function analysis to magnetic data in both "conventional" and "unconventional" ways to study the recent 2008–2009 seismic sequence that affected the L'Aquila area. In particular, to calculate the magnetic transfer functions together with their variations in time and to verify whether this electromagnetic method enables us to recognise a clear anomalous pattern preceding L'Aquila main shock, we analyse the vector magnetic data in the period 2006–2010 from L'Aquila observatory: this is run by the Istituto Nazionale di Geofisica e Vulcanologia (INGV) and has the not-so-common feature of being so close to the fault system area (only 6 km from the epicentre; Fig. 1), practically above the rupture area where such a large event struck after a long lasting sequence. Hence, for us it represented a sort of "benchmark" to begin such studies with.

The next section will provide a brief description of the geomagnetic transfer functions. In Sect. 3 we will describe the geomagnetic data and the applied "conventional" and "unconventional" methods used. Section 4 shows the main results with their possible interpretation. In Sect. 5 we will compare our results with those obtained from a seismic data analysis in order to have a confirmation of the proposed interpretation. The last section will present the main conclusions.

2 Transfer functions

Geomagnetic deep sounding is a method used in geophysics to gain information about the lateral and vertical electrical structure of the Earth's crust by means of the response of the medium, i.e. the crustal rocks, to the fluctuations of penetrating natural magnetic signals (e.g. Banks, 2007). From the spectral analysis of the time variations of the three components of the geomagnetic field (X,Y,Z) and their coupling (see below) it is possible to deduce and reconstruct an appropriate resistivity profile. High frequency time variations in the geomagnetic field components are mainly attributed to sources in the high-altitude ionosphere and magnetosphere, where complex large current systems result from the interactions between solar emissions (electromagnetic radiation and plasma fluxes) and the Earth's environment (atmosphere and planetary magnetic field) (e.g. Cowley, 2007; Richmond, 2007).

Electromagnetic (e.m.) field variations, generated by such large current systems, penetrate downward into the conducting layers of the crust and induce currents which, in turn, generate secondary magnetic fields that add to the original primary geomagnetic field. Both the conductivity σ of the crustal layers and the angular frequency ω of the inducing e.m. waves determine the typical depth δ at which the signal attenuates by $1/e$, in agreement with the following relation

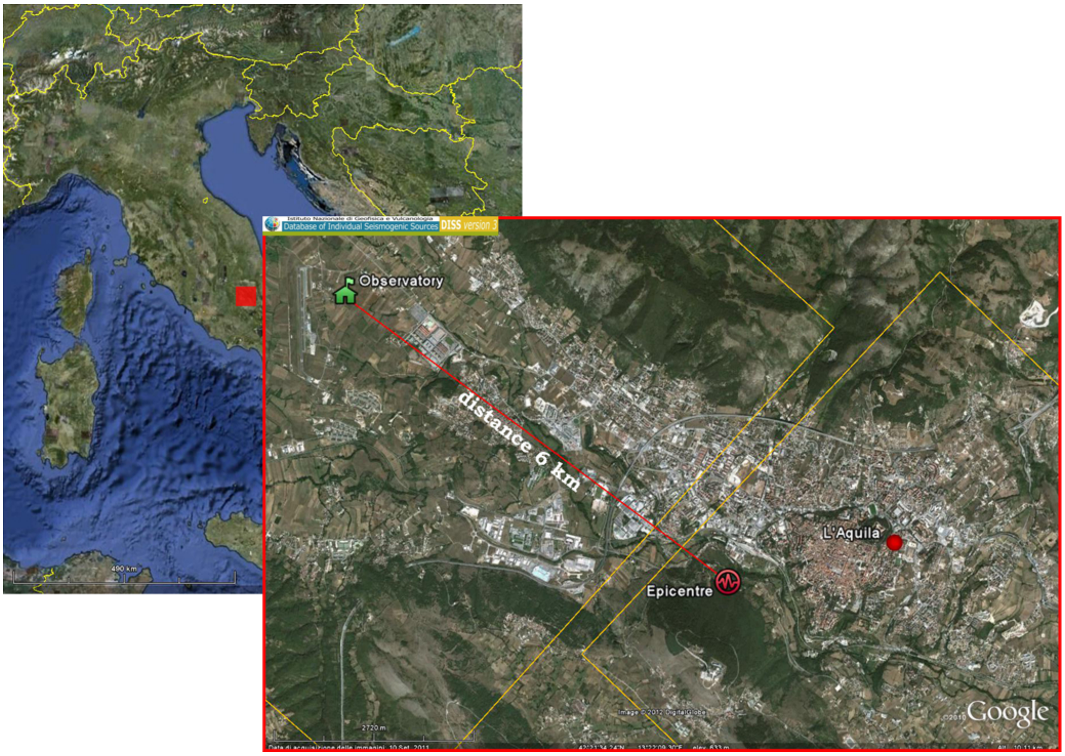

Fig. 1. Location of L'Aquila geomagnetic observatory with respect to the city of L'Aquila and the main shock epicentre. The instruments are located approximately 6 km from the estimated epicentre. The figure also shows some known fault structures near L'Aquila (yellow lines indicate fault plane projections), as extracted from the INGV DISS ver. 3 database (http://diss.rm.ingv.it/dissNet/) (Basili et al., 2008; DISS Working Group, 2010).

(Banks, 2007):

$$\delta = \sqrt{\frac{2}{\mu\omega\sigma}} \tag{1}$$

where δ is also called the *skin depth* and μ is the magnetic permeability, here considered to have its value in a vacuum of $4\pi \times 10^{-7}\,\mathrm{Hm}^{-1}$.

GDS theory asserts that, in the frequency domain and under certain conditions, the relationship between the horizontal $\mathbf{H} = (X,Y)$ and vertical (Z) components of the measured magnetic field contains the response of the radial resistivity structure of the crustal layers. This relationship is expressed as follows (e.g. Egbert and Booker, 1986)

$$Z(\omega) = \mathbf{T}^{\mathbf{T}}(\omega)\cdot\mathbf{H}(\omega) = A(\omega)\cdot X(\omega) + B(\omega)\cdot Y(\omega) \tag{2}$$

where the complex quantity $\mathbf{T}(\omega) = (A(\omega), B(\omega))$ is called a *transfer function*; usually its elements A and B are also called transfer functions; $\mathbf{T}^T$ is the transpose of $\mathbf{T}$.

The use of Eq. (2) presupposes a downward e.m. plane wave from an infinite horizontal external source penetrating a semi-infinite homogeneous medium, an assumption which is usually applied in GDS and approximately satisfied at middle latitudes.

Since the ionospheric and magnetospheric sources may have a strong variability both in intensity and in distribution, it is important to compute the transfer functions for a wide range of frequencies in order to reduce the influence of such variability on the results. These are considered meaningful when they are stable, i.e. invariant, with respect to the different epochs of the data used in their computation (Niblett and Honkura, 1980). For this reason, in the following analyses we will apply some appropriate smoothing to the transfer functions.

Before going ahead, a word of caution is necessary about the significance of the transfer function concept. This is a physically very approximate algorithm with intrinsic logical flaws, as it is clearly shown by Gregori et al. (1982). Any parameters that can be inferred from the transfer functions have to be considered as approximate, although reliable, physical information. Therefore, the analysis carried out in this paper can be certainly accepted as a sound analysis of observations, and it makes sense even independently of this drawback. In addition, the skin depth concept does not suffer from this specific drawback and can be reliably used when comparing it with the hypocentre depth distribution as in Sect. 5.

3 Data and method of selection

The analysis was performed on the three-component magnetic data recorded at 1 Hz sampling in the time span 2006–2010 at the INGV L'Aquila magnetic observatory, thus very close to the epicentre of the main shock and for a significant time interval including the seismic sequence under study. We were not able to extend the analysis of data before 2006 because of some clear contamination in the spectral frequency band of concern during most of 2005 with an unknown (likely artificial) origin in the magnetometers recordings (for this reason, at the beginning of 2006 the instrument was moved to a quieter site). However, we are satisfied with the analysed time span because it includes both the 2008–2009 seismic sequence and the remainder (more than 80 % of the total time span) that is without significant seismicity, if compared with the analysed sequence. Figure 1 shows the locations of L'Aquila city and the observatory with respect to the main shock epicentre: these two latter are almost 6 km distant. The long duration of observations permitted us to analyse a large number of data, to explore a wide range of frequencies and so to try to reduce the effects of time variability in the various sources. Nevertheless, for transfer function determination a strong selection rule was applied to the magnetic data with the aim of extracting, for every analysis that solves Eq. (2) in terms of $\mathbf{T}(\omega)$, a significant number of segments, each characterised by (i) a length of 8192 s; (ii) absence of spikes and gaps due to missing values (whose effect would have been a high frequency spectral contamination); (iii) a sufficient data sampling in the daytime from 06:00 a.m. to almost 19:00 p.m. (in order to increase the number of sources that "illuminated" the crustal structures). Although data from some days have been discarded because they did not satisfy these requirements, the selection stage resulted in a large number of days with three complete segments each.

3.1 The "Conventional Analysis" of transfer function determination

In order to average out the possible influences on data due to non-uniform strong sources on a specific day which could have biased the results, we collected the data enclosed into a ten-day sliding window (composed of a total of 30 segments of 8192 s each) moving one day ahead each time: in this way, even if the obtained transfer functions refer to the most recent day of the window, they actually contain a memory of data measured also during the previous nine days. In addition to the issue about having a stable and robust transfer function inversion, this choice of the length of the sliding window was made because we are not interested in sporadic and abrupt magnetic changes with short durations (say, of few days), but are looking for persistent features which could be related to the preparatory phases of earthquakes.

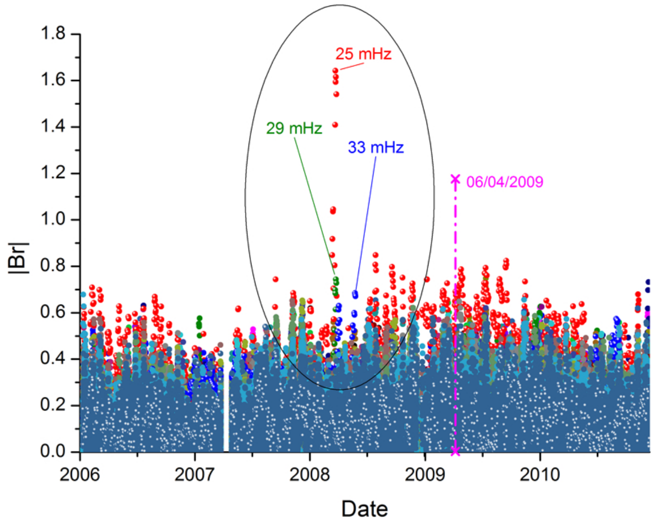

Fig. 2. The behaviour of the modulus of the real transfer functions B_r for all harmonics with periods 0.5–5 min (frequency range 3–33 mHz) in the wider time interval from 1 January 2006 to 31 December 2010. The vertical line indicates the main shock onset. Around one year before the main shock, some harmonics (frequencies of 25, 29 and 33 mHz: red, green and blue spheres, respectively) emerge from the background of 0.4 units.

The spectral transfer function analysis explored the time behaviour of the complex functions $A(\omega) = (A_r, A_i)$ and $B(\omega) = (B_r, B_i)$ of Eq. (2) for periods ranging from around 0.5 min to a maximum of 136.5 min for a total of 300 harmonics.

For a better comparison of their overall behaviour, some of the absolute values of B_r (the real part of B) related to the shorter periods (5–0.5 min) are presented in Fig. 2. As no significant differences between $|B_r|$ and $|A_r|$ appear, only $|B_r|$ is shown here. The imaginary parts of the transfer functions will not be considered because they do not have a clear and solid interpretation in the literature (Banks, 2007), though this does not mean that they might not contain some information about the system under consideration. The meaning of the real parts of the transfer functions is much clearer: for instance, $|A_r|$ and $|B_r|$ are used to define magnetovariational tools, such as the *Parkinson arrows*, which point towards concentrations of currents (e.g. Banks, 1973). The vertical solid line in Fig. 2 corresponds to the day of the main shock onset (6 April 2009). What can be seen is that most transfer function variability seems to be random below a background level of 0.4 units without any apparent connection with the main shock occurrence. However, a notable exception appears in the first part of 2008 (i.e. almost one year before the main shock) at frequencies of 25–33 mHz (periods of 30–40 s): thus, those harmonics seemed to us to be worthy of further study. By applying Eq. (1) for those frequencies and considering a plausible average crustal conductivity of $\sigma = 0.02\ \mathrm{S\,m^{-1}}$ (an intermediate value of the conductivity beneath Central Italy among those given by Armadillo et al., 2001), we have found that the range for skin depth δ corresponding to them is a little deeper than the main shock hypocentre. This outcome has been the starting point for a more detailed study in order to verify whether (i) the adopted conventional method had succeeded in removing some unwanted sources which may have biased the analysis; and in case of success, (ii) whether the suggestive hypothesis about their possible seismogenic origin may have a foundation. The need of better discriminating the behaviour of the very few harmonics between all the others prompted us to look for a technique which could serve the purpose.

3.2 The "Unconventional Analysis": the transfer function entropy

In consideration of the large number of harmonics we studied (and thus of the depth of the crustal layers), we resorted to a new unconventional approach that takes into account all the harmonics in a *holistic* analysis of the whole seismogenic volume. This is an example of the application of the point of view of *geosystemics* (De Santis, 2009). In particular, this analysis attempts to treat the volume of rock beneath the measurement site as a whole and to disclose some anomalous behaviour with time of some appropriate macroscopic quantity which can be defined by means of the computed transfer functions. In other words, if we imagine that this volume of rocks is a complex system made of several parts at different depths, then it is not difficult to understand that each harmonic that probes the crust may contribute to the definition of a suitable quantity in order to investigate the order/disorder characteristics of the same system. This macroscopic quantity is the informational (or Shannon) entropy (Shannon, 1948). This approach is also supported by recent results which find a strong nonlinear time behaviour in the seismic sequence of L'Aquila (e.g. De Santis et al., 2010, 2011).

Applying Shannon's theory (Shannon, 1948) to the real parts of the magnetic transfer functions (A_r or B_r) we define the normalised "transfer function entropy" by

$$E(t) = -\sum_{i=1}^{n} \frac{p_i(\omega,t)\cdot \log p_i(\omega,t)}{\log n} \qquad (3)$$

where $p_i(\omega,t)$ is the probability of having a certain transfer function contribution at a given ω, defined as $p_i(\omega,t) = [K_i(\omega,t)]^2 / \sum_{i=1}^{n} [K_i(\omega,t)]^2$ with K_i equal to either $A_r(\omega,t)$ or $B_r(\omega,t)$ at time t, and n the total number of harmonics. In the following we will consider the average between the entropies coming from A_r and B_r , i.e. $\langle E(t)\rangle = \left(E_{A_r} + E_{B_r}\right)/2$, even though the application of Eq. (3) to only one of them provides similar results.

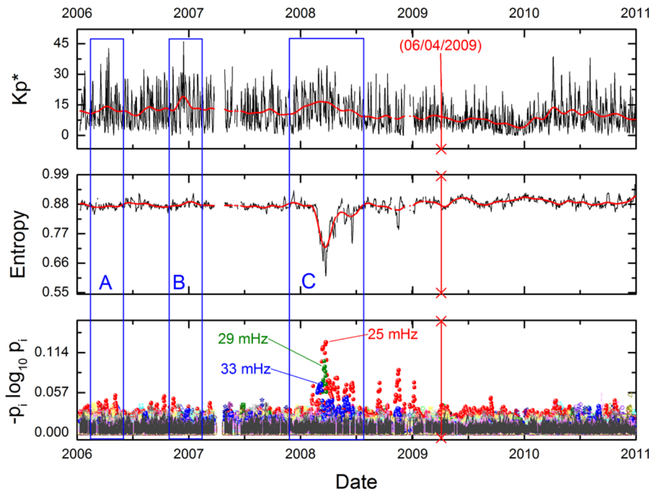

Fig. 3. Time behaviour of daily K_p geomagnetic index (top panel; $K_p* = 10 \times K_p$) along with that of the total entropy averaged between A_r and B_r (middle) and of the contributions of individual harmonics to the total entropy (bottom). The red lines in the top and middle panels are 20-day moving averages. The main features, visible in the middle and bottom plots, are the increases in the values of a few harmonics (with frequencies between 25 and 33 mHz) with distinctly larger contributions to entropy than the general background of 0.03 units. These increases do not seem to be caused by external magnetic activity. Intervals indicated with A, B and C are useful for comparing increases of K_p* with different behaviours of the entropy.

4 Results

Figure 3 shows the results obtained from the entropy analysis. To exclude the possible influence of external magnetic activity, we have also shown in the top panel the time behaviour of K_p*, which is ten times the daily mean of the K_p geomagnetic index. The K_p index is a three-hour global measure of external geomagnetic activity; it takes values from 0 to 9, and the higher its value, the more perturbed the magnetic field is from external sources (Menvielle and Berthelier, 1991). The middle and bottom panels show the average total entropy $\langle E(t)\rangle$ and the contribution of the individual harmonics to $\langle E(t)\rangle$, respectively. The red lines in the top and middle panels are 20-day moving averages for a better visualization of the smoothed behaviour in time. The main feature that stands out in the entropy (middle and bottom panels) plots is the remarkable decrease of the total entropy (middle panel) in the first part of 2008 and the corresponding emergence of a few specific harmonics (with frequencies between 25 and 33 mHz), which have distinctly larger values of entropy contribution than the background level of 0.03 units (bottom panel). These anomalous frequencies are the same that appeared in Fig. 2. We call them "anomalous" because when compared with K_p* we cannot find any indication that these increases are caused by the external magnetic activity. Indeed, if we compare their behaviours, say in the interval labelled C of Fig. 3, we can notice that K_p* has higher values for a wide time interval in approximate coincidence with the increases in the single contributions (and the consequent decrease of entropy). On the contrary, the same behaviour is not repeated at earlier times (see intervals A and B), when even higher values of K_p* do not cause any significant reduction of the total entropy due to the remarkable increase of the contributions of some few harmonics, but rather cause all the other harmonics to contribute to a more generalized increase of the total entropy. From the point of view of information theory, the emergence of some particular harmonics as opposed to the others is interpreted as the emergence of these less probable components, which thus contain more "surprise" (or "information") than all the others that share almost the same probability: the capability to highlight the collective ordered/disordered behaviour of a system, no matter how complex, in relation to that of its single components, is the strength of this method.

In order to give a more robust proof than the simple visual inspection, the correlation analysis of Fig. 4 shows the entropy plotted against K_p*, where the dashed line indicates the lowest entropy values reached mostly during the first part of 2008. No linearly correlated behaviour appears, confirming that the decrease in entropy cannot simply be ascribed to some external contamination.

Why do those particular harmonics show such a peculiar increase in contrast with all the others? How could that be explained? Is it possible to link those features to some tectonic causes, such as, for instance, the fluid migration during crack opening (e.g. Lucente et al., 2010)?

From Eq. (1) and considering an average conductivity of $0.02\,\mathrm{S\,m^{-1}}$, the previously mentioned anomalous harmonics correspond to depths of 20 ± 6 km. The given error of 6 km not only takes account of the frequency range width of the anomalous harmonics, but also of some uncertainty in the chosen average conductivity. This depth of around 20 km, even considering the possible associated error of 6 km, is beneath the typical hypocentral depth of about 10 km as deduced from the whole set of events (i.e. foreshocks, main shock and aftershocks) of the L'Aquila seismic sequence (Chiarabba et al., 2009). Taking into account this physical meaning of the transfer function harmonics in relation with depth, we might argue that the series of entropic bursts occurring at around 12, 10 and 6 months before the main shock could be interpreted as a possible precursory pattern. Actually, the internal (lithospheric) origin of these signals is unequivocal since their contribution to entropy is in the opposite direction to that of all the other harmonics which decrease as the external ionospheric and magnetospheric disturbances increase (Fig. 3). The only other phenomena known to us that occur at depths lower than seismogenic zones are non-volcanic tremors (e.g. Nadeau and Dolenc, 2004) and fluid-diffusion (F-D) mechanisms (e.g. Bonafede et al., 2007; Stefánsson et al., 2011). The former phenomena have been found for the first time outside subduction zones at 20–40 km beneath the San Andreas fault, and therefore deeper by around 10–30 km than the typical earthquake hypocentres in that region. The corresponding frequency content is 1–10 Hz and each event can last several minutes (but less than 20 min), releasing an energy equivalent to earthquakes with $M < 1.5$.

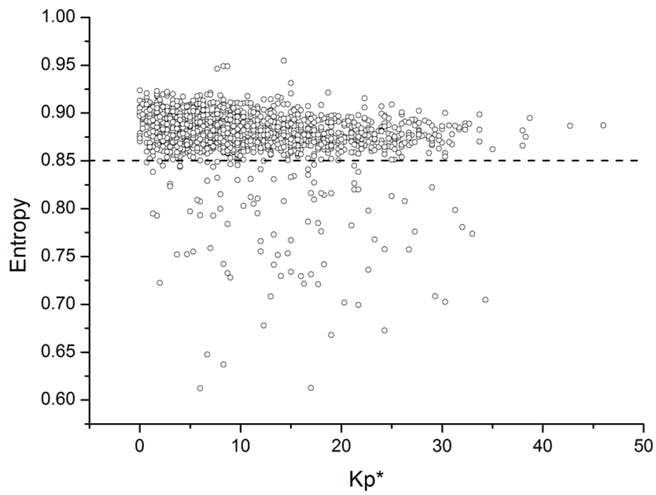

Fig. 4. Entropy vs. K_p*. Below the dashed line there are the lowest entropy values (entropy lower than 0.85) mostly reached during the first part of 2008. There is no clear correlation between the two quantities, especially during the latter period.

The F-D processes are assumed to be possible mechanisms for producing earthquake instability at depth. Stefánsson et al. (2011) consider this beneath part of Iceland, where magma released from the deep crust may be the most important cause of the intermediate and large earthquakes originating in that area. It is expected that a similar mechanism could be present in other parts of the world, including Central Italy (Chiodini et al., 2004).

The characteristics of both the above processes are not in conflict with the magnetic bursts that we find beneath the L'Aquila seismogenic zone, and the two phenomena could be parts of the same physical process of earthquake preparation.

What we can affirm is that on the one hand, our results exclude the possibility that some significant change of electrical conductivity occurred in close proximity (one week before) to the main shock, i.e. in the same epoch when Lucente et al. (2010) hypothesised the occurrence of some diffusion of fluids under the hypocentral region; on the other hand, the same results do not exclude the possibility that these processes actually happened. If they occurred, they were too rapid and not detectable with our techniques. In any case, our analysis points towards a former epoch, i.e. around 1 yr, 10 and 6 months before the main shock and this merits further investigation.

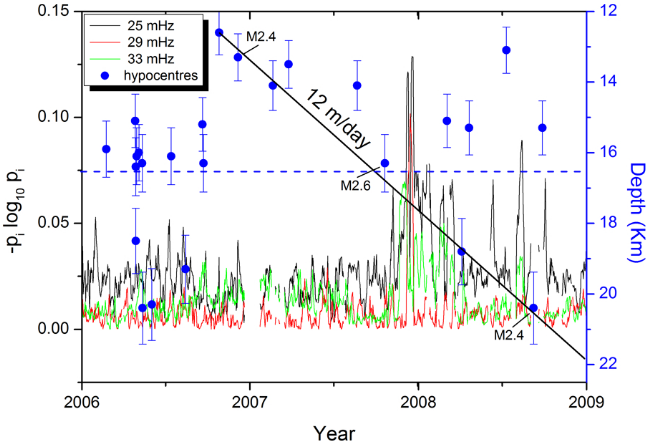

Fig. 5. Temporal behaviour of individual contributions with 25–33 mHz frequencies to total entropy (values indicated on the left y-axis) and deeper ($h > 12$ km) earthquakes (depths indicated on the right y-axis) in the L'Aquila seismic sequence in the three-year interval 6 April 2006–5 April 2009, in a radius of 20 km from the main shock epicentre and with magnitude $M \geq 1.5$. The horizontal dashed line represents a hypothetical depth where the brittle-ductile discontinuity might occur. The oblique solid line approximates the deepest earthquakes from the end of 2006 to the main shock occurrence. We can notice a slow downward hypocentre migration with apparent velocity of 12 m day^{-1}. The figure also shows the magnitude of the largest deep earthquakes with $M \geq 2.4$ (all the others shown in this figure have smaller magnitudes): they all lie close to the solid oblique line.

5 Comparison with seismic data analysis

To investigate the above aspects much better, we first made a careful inspection around the times of the magnetic bursts of the seismic record of the available AQU station, which is the closest to the main shock epicentre. However, due to its location near the centre of L'Aquila, the strong anthropic noise of the area did not allow us to detect the possible presence of non-volcanic tremors. We therefore went one step further in searching for deeper earthquakes of the seismic sequence by investigating their behaviour in time and space. To do this, we have extracted from the ISIDE seismic database (http://iside.rm.ingv.it/iside/standard/index.jsp) all earthquakes in the preceding three years (i.e. from 6 April 2006 to 5 April 2009) and within 20 km of the main shock epicentre, all with hypocentral depth greater than 12 km (i.e. $h > 12$ km; deepest hypocentre at 21 km in that period). We stopped the analysis on 5 April 2009 because we are more interested in what happened in the time preceding the main shock and also to avoid analysing the huge number of events that occurred after it. The reason for choosing this range of depths was threefold: first, it is just below the main shock epicentre; second, it includes the seismogenic layer of the area together with the brittle-ductile (B-D) boundary of the crust, estimated in 10–18 km in this area (Chiarabba et al., 2005); third, it roughly corresponds to the depths indicated by our transfer function entropy analysis. Figure 5 shows the corresponding temporal behaviour of the single contributions to the total entropy from 25–33 mHz frequencies along with the distribution in depth of the deeper events. We associated an error of 5 % with each depth estimate, and this corresponds (at these depths) to uncertainties between ±0.6 and ±1.1 km. The oblique solid line approximates most of the deeper earthquakes of the seismic sequence after the end of 2006. The figure shows also the magnitude of earthquakes with $M \geq 2.4$: they all lie close to the solid oblique line. This line would correspond to a downward earthquake migration with apparent velocity of 12 m day^{-1}, a value which is comparable with those occurring in other situations; for instance, it is similar to the velocities found in reservoir induced seismicity where pore pressure diffusion causes hypocentre vertical migration rates between 15 and 58 m day^{-1} (El Hariri et al., 2010). We also notice that the first magnetic entropic burst occurs just after the largest deep event ($M_L = 2.6$), while the other two occur between the last two deepest earthquakes preceding the main shock, just during the downward migration of hypocentres. Although we cannot exclude the possibility that our results are just a coincidence, a simple "gravitative-diffusive" model could also be an explanation: the larger rupture (when compared with those produced by smaller magnitude events) caused by the $M_L = 2.6$ event might have allowed over-pressured fluids to diffuse upwards. If this is correct, the horizontal dashed line in Fig. 5 would give the possible position of the B-D boundary across which the fluids can start to migrate: according to this interpretation the B-D boundary would represent a sort of barrier to fluids, that once it is broken may allow them to diffuse in the upper crustal medium. The tectonic process that started the dynamical evolution of the seismic sequence culminating with the strong main shock of 6 April 2009 could have originated at depths (around 20 km or deeper) lower than the main shock hypocentre, where the Adriatic plate subducts under the Central Apennines: an initial instability at those depths, whose specific origin is still unknown, could have been propagated upwards to the shallower brittle crust, finally causing the foreshocks and the subsequent main rupture. Since the crust is believed to be almost entirely filled with fluids and fluid flows are generally accompanied by (or due to) crack opening and pore-pressure diffusion (e.g. Nur and Booker, 1972; Scholz, 2002), the role of fluids in our model is very important. The upward instability propagation could have been accentuated by upward fluid migration, in a fashion similar to the F-D model (Stefánsson et al., 2011). This model would be supported by the conductivity structure proposed by Armadillo et al. (2001) using independent magnetic data, of a consistent conductive volume beneath L'Aquila at depths between 20 and 40 km. This volume could be interpreted as the possible reservoir releasing the fluids upward, in some way analogous to what was found for Iceland, supporting the F-D model there (Stefánsson et al., 2011). Moreover, this is also partly confirmed by evidence for the presence of high-pressure fluids at hypocentral depths (Terakawa et al., 2011). However, we are aware that in order to confirm or reject our model, further studies will be needed, since the phenomena we have observed have emerged as result of this unconventional analysis, and so they certainly merit further investigation in the near future, using other seismic sequences and corresponding magnetic data from a nearby geomagnetic station or observatory, as a reference station.

6 Conclusions

During the interval 2006–2010, a "conventional" transfer function analysis of L'Aquila magnetic data detected, in a single harmonic, some unconvincing anomalies which were confined to a limited period, around one year before the main shock. In order to evidence the significance of that "exception" in relation to the other components, we resorted to the subsequent analysis of the same magnetic data in terms of the Shannon entropy, the so-called transfer function entropy. This new approach allowed us to identify some distinct temporal burst regimes in which there is a significant abrupt increase of some harmonic contributions to entropy in contrast to a general decrease of all the others. These bursts start about one year before the main shock after the largest deep earthquake, with subsequent occurrences 10 and 6 months earlier than the main shock. In particular, the periods between 30 and 40 s (frequency of 25–33 mHz, corresponding to skin depths of around 20 km) show values of entropic contributions significantly higher than the background value (Fig. 3). We interpret this result as due to the complex phenomenology associated with the evolution of the L'Aquila seismic sequence occurring in the deep crustal layers. These entropic anomalies could probably be related to migration of fluids and/or changes in the micro- and meso-rupture process that affected much of the lithosphere beneath the region of L'Aquila, with particular concentration at 20 ± 6 km depth and therefore just below the average hypocentral depth (approximately 10 km) of the L'Aquila earthquake sequence. This upward diffusion of fluids is not clearly visible with a conventional transfer function analysis, because it probably does not affect too much the vertical profile of conductivity but only how it is organised, i.e. its volumetric distribution. For this reason, the entropic approach can better reveal the rapid increases of entropy at some harmonics. This fact is further corroborated by a temporal comparison of the magnetic entropic bursts with the sequence of deeper earthquakes preceding the main shock of the L'Aquila seismic sequence (Fig. 5).

Concluding, we can affirm that the transfer function entropy is very effective in getting rid in some way of the physical approximations introduced by the "conventional" transfer function analysis. This depends on the robustness of the entropy concept through its definition, to which all harmonics (in this paper 300) contribute. Differently stated, the problem is tackled in terms of an underestimating holistic viewpoint, by which we search for some overall general regularities, independent of any kind of more or less intuitive physical modelling, such as that implied by the "conventional" and approximate transfer functions. We then compare the resulting intuitive scheme with the seismic results and thus arrive at a more complete and general interpretation. Summarising, the real meaning of the paper is a proof that this kind of approach, that we called *geosystemics*, may be heuristically effective, and it is worthwhile pursuing it by means of improved data analysis and further investigations on other records and concerning other seismic areas.

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