Umbria-Marche sequence (central Italy): a study on its aftershock sequence

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ABSTRACT   We studied the aftershock sequence following the mainshock (M_s 6.4) of September 26, 1997 with its epicenter at Colfiorito (Italy). This work studies a methodological aspect finalized to observe possible anomalies in the temporal decay before a large aftershock with magnitude $M>5.5$. Such anomalies are thought to be variations from the mean temporal trend. In fact, modeling the dacay as a non-stationary Poissonian process, the number of aftershocks in a small time interval Δt is the mean value $n(t) \cdot \Delta t$, with standard deviation $\sigma = \sqrt{n(t) \cdot \Delta t}$. In this work, besides the analyzed parameters of other sequences, we have extended the spectrum of parameters by using fractal properties of the temporal sequence. Earthquakes belong to a class of phenomena known as multifractals. In general, it is important to define the fractal dimension D, but sometimes is not useful if we are describing a natural phenomenon; so it is necessary to define the box-counting dimension D_0, and the correlation dimension D_2. Usually $D_0 \geq D_2$. The data concerning the temporal series, checked according to completeness criteria, come from the NEIC-USGS data bank (http://neic.usgs.gov/neis/epic/).

1. Introduction

On September 26, 1997 an earthquake of M_w 6.0 (computed by Regional Centroid Moment Tensor www.ingv.it) occurred at Colfiorito (central Italy) at 9:40 GMT. The aftershock sequence (Umbria-Marche sequence) was closely monitored by the Istituto Nazionale di Geofisica (now Istituto Nazionale di Geofisica e Vulcanologia) through a portable network of ten stations.

Aftershocks following any major earthquake tend to cluster in space and time (Console and Di Giovambattista, 1987). Their temporal distribution often follows a regular trend, as first observed by Omori in 1894. Most of the large earthquakes that occurred in the past years (Japan 1995, Turkey, 1999, Taiwan 1999, India 2001, Indonesia 2004, etc.) had significant damage to structures and injured or killed many people.

The purpose of this paper is to identify anomalies in the temporal decay occurring in the seismic sequence before a large aftershock.

The temporal and spatial evolution of seismicity, following the mainshock in the area, has been described by Amato *et al*. (1998). A detailed study of the mechanism of the aftershock sequence was made by Chiaraluce *et al*. (1999). The tectonics of the region is affected by the Quaternary activity of normal faults oriented NW-SW. There is also evidence of a NE-SW transversal fault system, both extensional and strike slip (Deschamps *et al*., 1984; Ekstrom *et al*.,

1998). In this article, various features of the Colfiorito seismic sequence are taken into consideration. The Colfiorito earthquake had serious consequences, twelve people died and several buildings were seriously damaged. The sequence was developed along an approximately NNW-SSE line between Bagno di Romagna and L'Aquila. A detailed analysis of the temporal series of the sequence supports the hypothesis that it was developed by the activation of a fault segment (Nocera-Verchiano) on September 26. A large number of events in the area have a magnitude greater than 4.0. The sequence is characterized by an evident migration of seismicity (Cocco *et al.*, 2006). The temporal and spatial evolution of seismicity following the mainshock in the area, initially, shows a cluster of activity prevalently between Nocera Umbra and Verchiano. An occurrence of another event on October 14, 1997 shows that the activity is close to Sellano (south). The aim of this study is to explain any anomaly in the temporal distribution of seismic activity, particularly by focusing on the seismic anomalies in the temporal decay of the sequence. These anomalies are addressed as seismic activity variations, in a short time interval, before the occurrence of a large aftershock. The possible explanation of this phenomenon could be found in terms of fracture of an asperity related to the large aftershock. The fracture of neighbouring asperities initiated after the mainshock is presumably due to the redistribution of the strain energy, which is released by the mainshock, resulting in enhancement of the stress concentration around the nearest neighbouring intact asperities (Lei, 2003).

2. Aftershock temporal series: some theories and models

The probability of occurrence of future large earthquakes are very interesting from the point of view of earthquake physics, and are crucial for attempts to forecast the hazards due to large, damaging earthquakes. A theoretical model that successfully describes the earthquake recurrence is unknown. So it is necessary to adapt probability distribution based on earthquake history. Usually, the frequency of occurrence of aftershocks decays rapidly according to Omori's law (Utsu, 1961):

$$n(t) = k / (c + t)^{-p} \qquad\qquad (1)$$

where $n(t)$ is the frequency of aftershocks at time t after the mainshock; k, c and p are constants that depend on the characteristics of the area (Omori, 1894; Matsu'ura, 1986). The p-value varies between 1.0-1.4. Earthquakes located within a characteristic distance from the main event can be considered as aftershocks. The spatial distribution of the aftershocks is often related to the fault area or its length. Fault distance is usually considered equal to one or two lengths of the fault segment related to the mainshock. The empirical relations between fault area A and magnitude M_s [Eq. (2)], fault length L and magnitude M_s [Eq. (3)], and size and frequency of occurrence of earthquakes [Eq. (4)] are
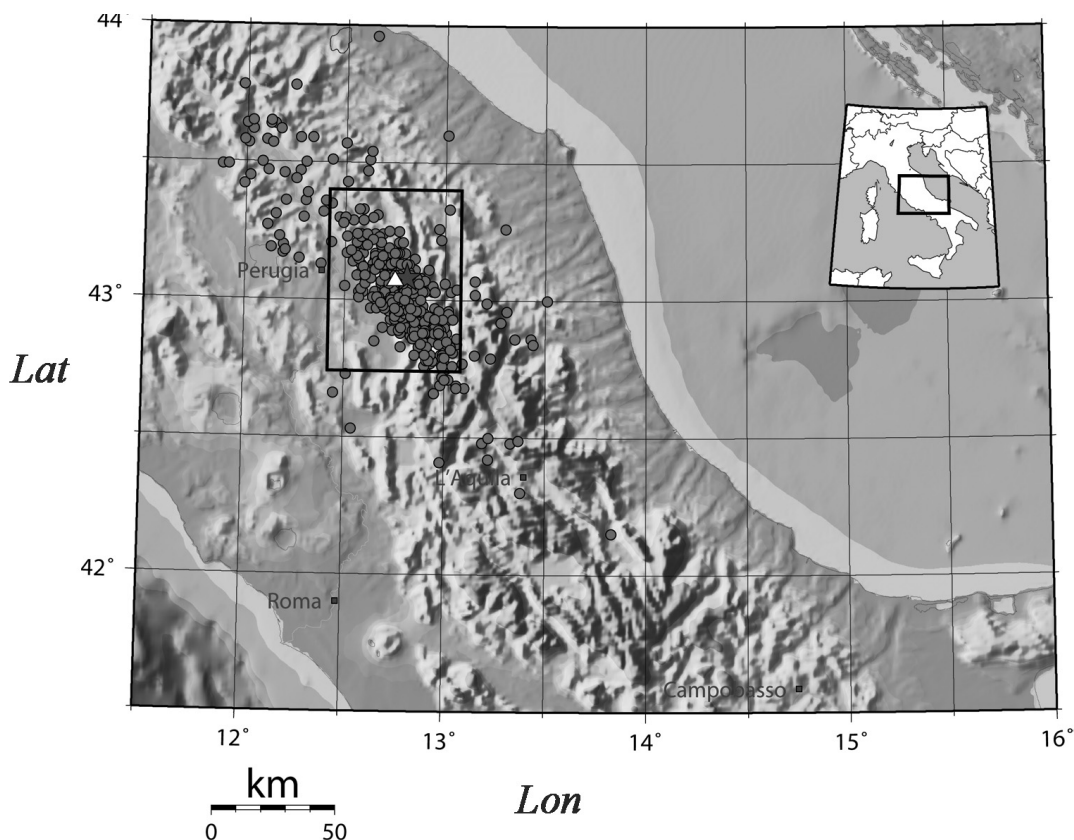
$$\log A = 1.02 M_s + 6.0 \qquad\qquad (2)$$

Fig. 1 - Umbria-Marche seismic sequence. The map is centered on the coordinates (triangle) of the "barycenter": 43.07° N, 12.75° E. The epicentral distribution of events with magnitude $M \geq M_c$ in a period d=121 days since the occurrence of the mainshock (star) is shown. The latitude, longitude and depth of events are respectively ranged from 42.74° N to 43.40° N, from 12.42° E to 13.08° E. and from 0 km to 70 km.

$$\log L = 0.5 M_s - 1.8 \quad \text{(Utsu, 1969)} \tag{3}$$

$$\log N = a - bM \quad \text{(Gutenberg and Richter, 1954).} \tag{4}$$

Here N is the number of earthquakes with a fixed magnitude M, a and b are constants characteristic for the given area. The relation is valid for earthquakes occurring regionally and globally. The constant b varies from region to region, but it is generally in a range of $0.8 < b < 1.2$ (Frohlich and Davis, 1993).

3. The data set

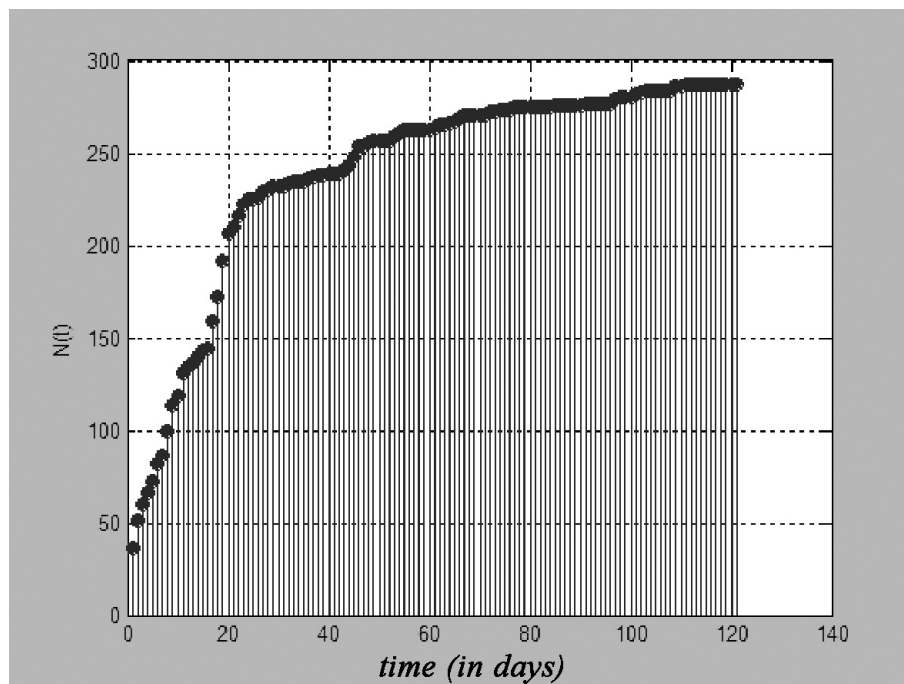Our data set consists of 482 events with magnitude greater than 1.0 recorded from September

Fig. 2 - Cumulative number of shocks per day during the duration of the Umbria-Marche seismic sequence on September 26, 1997 (121 days).

26, 1997 to September 26, 1998. The data, concerning the temporal series analyzed here, come from the NEIC-USGS data bank (http://neic.usgs.gov/neis/epic). The United States Geological Survey reported the mainshock (*M*=6.4) that occurred at 43.08° N, 12.81° E. Its time of origin was 9:40 (UTC) or 11:40 local time. For our analysis, we calculated a priori the dimensions of the area using the empirical relation log L=0,5 M_s-1.8 (Utsu, 1969). We acquired data in a square with sides at a distance 3L from the mainshock epicenter, where L represents the fault length of the mainshock computed by Eq. (3), and from the first 10 days starting from the mainshock. Using these data, we calculated the "barycenter" of the aftershock sequence. The "barycenter" coordinates are calculated as an arithmetical mean of the latitude and longitude values of the shocks (D'Amico *et al*., 2004). This was introduced to better frame the aftershock epicenters, and so, to eliminate the possibility that a lot of them might remain out of the sector considered. The final, rectangular sector considered for aftershocks is centred on the sequence "barycenter" and has its sides at a distance from this point, in terms of latitude and longitude, equals to 1.5L. Following Caccamo *et al*. (2005, 2007), we computed the completeness threshold M_c and the duration for the seismic sequence. The completeness threshold, M_c, of a data set is made using the Gutenberg-Richter (1954) relation. In this case, M_c is 3.1 computed with the data related to the epicentral distribution for events with magnitude $M \geq 1$ reported by the web site of the NEIC-USGS data-bank in the first 10 days after the mainshock and, in a square centered on the barycenter (lat. 43.07° N, lon. 12.75° E). Fig. 1 shows the data used to obtain the temporal trends related to the parameters shown in Figs. 2, 3 and 4. The map is centered on the coordinates (triangle)
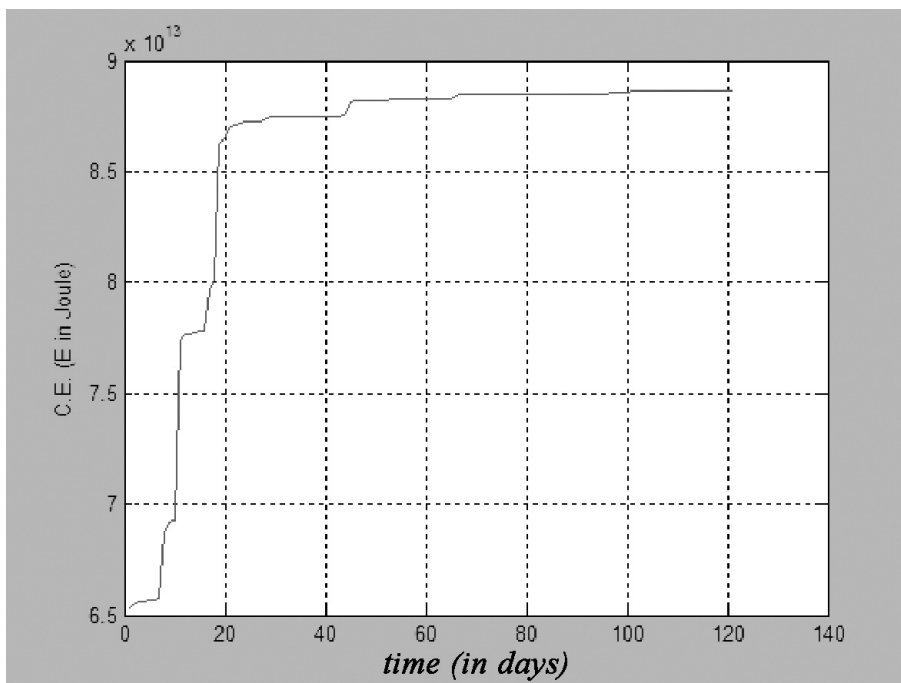
Fig. 3 - Cumulative energy of the aftershocks sequence. The energy involved in the decay process was about 9×10^{13} J.

of the "barycenter": 43.07° N, 12.75° E. The epicentral distribution of events with magnitude $M \geq M_c$ in a period $d=121$ days from the occurrence of the mainshock (star) is shown. The latitude, longitude and depth of events range from 42.74° N to 43.40° N, from 12.42° E to 13.08° E and from 0 km to 70 km, respectively.

Fig. 2 shows a cumulative number of aftershocks per day. It is possible to see how the sequence is stopped after ten days without shocks.

Fig. 3 shows the cumulative energy of the aftershock sequence expressed as

$$E_n = \sum_{i=1}^{n} \sum_{j=1}^{m_i} e_{ij} \qquad n=1,...,d \tag{5}$$

where e_{ij} represents the energy of aftershocks in the i-th day, m_i the number of aftershocks in the i-th day and n is the temporal length, in days, of the sequence.

The empirical magnitude-energy relation used here is

$$\log_{10} E = 1.5M + 4.2 \tag{6}$$

with energy E in Joule (Gutenberg and Richter, 1942, 1956; Udias, 1999). The energy involved in the decay process was about 9×10^{13} J.

In Fig. 4 *p*-values are shown, where the central graph shows the *p*-value variations and the others represent the 95% confidence interval. The *p*-value may be related to the temporal trend of shocks. In fact, with an increase of the *p*-value, corresponds an increase of the fracturing level of the rocks which indicates a relative reduction in the stress pattern. Its value also depends on the characteristics of the area (Omori, 1894; Bottari and Caccamo, 1981; Matsu'ura, 1986; Kisslinger, 1993).

The *p*-value is computed by a non-linear least squares method (Dennis, 1977; More, 1977; Coleman and Li, 1994, 1996) and its computation is made during the entire temporal duration. Its value is about 0.9 at the end of the sequence. From this graph it is possible to note two minima on October 6 and 15 of 1997, respectively. The *p*-value decrease, preceeding October 15, 1997, could indicate a resumption of activity in a quiescent area, adjoining the already active segment. This is hypothesised to indicate the possibility of a strong event.

4. Time clustering of seismicity

The fractal nature of earthquakes is widely observed. The last 20-30 years have seen a growing interest in evidence of fractal properties both in the spatial and the temporal distribution (Smalley *et al.*, 1987; De Rubeis *et al.*, 1993, 1997). A Laboratory experiment (Hirata *et al.*, 1987) shows fractal distribution of rock fractures. These observations imply that a description of scaling properties of earthquakes can be carried out.

Complex phenomena often exhibit power law (fractal) scaling (Mandelbrot, 1967, 1992; Turcotte, 1997). Aki (1981) showed that the Gutenberg-Richter relation and the formula $N=CA^{-y}$ are equivalent. Thus, the Gutenberg-Richter relation implies universal fractal behaviour of earthquakes. In the relation, N represents the number of earthquakes per unit time with rupture area greater than A occurring in a specified area; C and y are constants. The fractal dimension is $D=2y$ (Turcotte *et al.*, 2000). Earthquakes belong to a class of phenomena known as multifractals. The concept of topologic dimension for any object is known. For example, a segment, a plane figure or a solid figure have, respectively, dimensions equal to 1, 2 or 3. A fractal can instead be defined as an object of fractionary dimension. To define its dimension it is important to refer to the self-similarity concept: an object is selfsimilar when it can be divided into similar parts of the whole. Fractal dimension, D, is defined, using the ratio of homothety, as:

$$D = \frac{\log N}{\log\left(\frac{1}{k}\right)} \tag{7}$$

where k is the ratio of homothety and N is the number of the selfsimilar copy obtained.

Given a volume, we divide the space into cubes of side ε, and we calculate the number of occupied cubes, $N(\varepsilon)$. The fractal dimension based on box-counting, D_0, is:

$$D_0 = \lim_{\varepsilon \to 0} \frac{\log N(\varepsilon)}{\log\left(\frac{1}{\varepsilon}\right)} \tag{8}$$
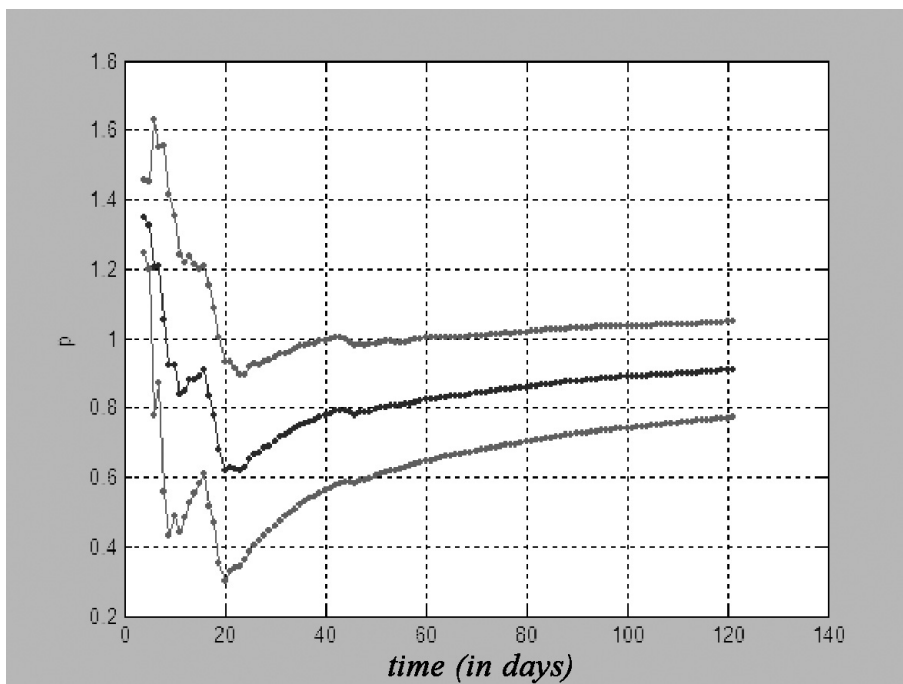
Fig. 4 - Variations of *p*-value during the decay process for the Umbria-Marche seismic sequence on September 26, 1997. This value is related to the temporal trend of shocks. The central graph represents the parameter variations, the others represent the confidence interval at 95% level.

If we want to consider "how many of the cubes are occupied", we must introduce another important fractal dimension, the correlation dimension D_2, defined as:

$$D_2 = \lim_{\varepsilon \to 0} \frac{\log C(\varepsilon)}{\log(\varepsilon)} \tag{9}$$

where $C(\varepsilon)$ represents the "correlation sum" function defined as:

$$C(\varepsilon, q) = \sum_i p_i q_i^{q-1} = \frac{1}{N} \sum_i \left\{ \frac{1}{N-1} \sum_{j \neq l} \theta\left(\varepsilon - \left|\overline{x(i)} - \overline{x(j)}\right|\right) \right\}^{q-1} \tag{10}$$

considering the Heaviside step function:

$$\theta(x) = \begin{cases} 1 & x \geq 0 \\ 0 & x < 0 \end{cases}. \tag{11}$$

N represents the number of points, $x(i)$ and $x(j)$ the position of various points along the axes, the absolute value represents the Euclidean distance of a pair of points that characterize the distribution and, finally, ε is the length of intervals which fill the entire ensemble.

$P(i)$ represents the probability of a point belonging to the i-th box with size r. The real number q is a weight that permits the investigation of more or less densely clustered parts of the set. In particular if $q=2$, p represents the probability that 2 points are in the same box (Schuster, 1988).

It is possible to obtain the box-counting D_0 and the correlation dimension D_2 for $q=0$ and $q=2$, respectively. In general $D_0 \geq D_2$, when $D_0 = D_2$ the phenomenon is a monofractal (Goltz, 1998). Moreover, these dimensions are less than 1 where events are clusterized in time (D'Amico *et al.,* 2005).

The temporal distribution of seismicity has been investigated by several authors using a diverse range of techniques. Time variations of the fractal dimensions of seismic sequences have been investigated, highlighting a connection with the b-value of the Gutenberg-Richter relation (Hirata, 1989; Main, 1991a, 1991b; Lomnitz-Adler, 1992; Oncel *et al.*, 1996).

In the present work, the study on fractal dimensions D_0 and D_2 highlights the time clustering of seismicity related to the epicentral area of September 26, 1997 Colfiorito earthquake. This aftershock sequence has been empirically divided into four periods of interest each of which contains the events of the preceeding time intervals, expressed in days:

period 1   1....5;
period 2   1...10;
period 3   1...15;
period 4   1...20.

Fig. 5 shows fractal dimensions D_0 and D_2 calculated for each period separately. We obtained $D_0 \approx 1.0301 \pm 0.1930$ and $D_2 \approx 0.5372 \pm 0.0198$ for the first period, $D_0 \approx 0.8934 \pm 0.0709$ and $D_2 \approx 0.6494 \pm 0.0071$ for the second period, $D_0 \approx 0.8815 \pm 0.0413$ and $D_2 \approx 0.6480 \pm 0.0178$ for the third period, $D_0 \approx 0.8786 \pm 0.0301$ and $D_2 \approx 0.6986 \pm 0.0058$ for the fourth period. All the calculations have a high correlation coefficient R close to 1. Fig. 6 shows the trend of D_0 and D_2 for the four periods considered. We observe that D_2 values appear almost constant from period 2 to period 3, while there is an increase in the first and last periods. This seems to be consistent with the occurrence of the strong aftershocks after the increase in D_2 values. This phenomenon should be further investigated.

5. The Δ/σ method

Following the approach described by Caccamo *et al*. (2005, 2007) and Parrillo *et al*. (2005), we detected the temporal decay of this aftershock sequence finding some anomalies. In mathematical terms, the observed temporal series of the aftershocks, per day, can be considered as a sum of a deterministic contribution and a stochastic contribution. These contributions are respectively the temporary decay of aftershocks with a power law, i.e. the modified Omori formula (Utsu, 1961), and the random fluctuations around a mean value represented by the above mentioned power law.

Since the decay phenomenon can be modelled as a non-stationary Poissonian process, where the intensity function is equal to $n(t)=k \cdot (t+c)^{-p}$ (Page, 1968; Matsu'ura, 1986), the average number of aftershocks in a time interval Δt is $n(t) \cdot \Delta t$, with a standard deviation of $\sigma = \sqrt{n(t) \cdot \Delta t}$. With
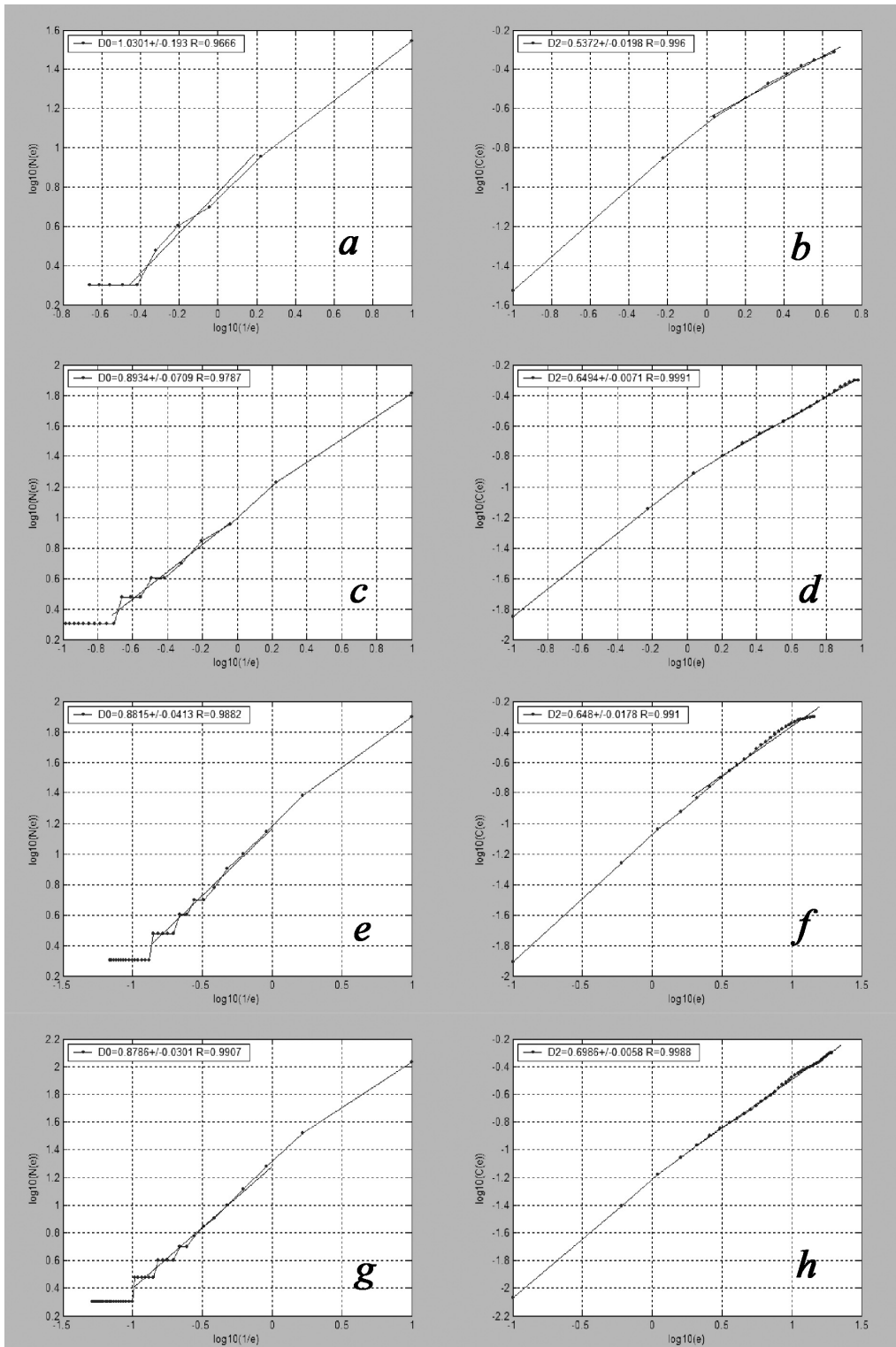
Fig. 5 - Calculations of the fractal dimensions D_0 and D_2 respectively for: Period 1 (a and b); Period 2 (c and d); Period 3 (e and f); Period 4 (g and h). These dimensions are calculated from the slope of the best straight linear the log-log plot.
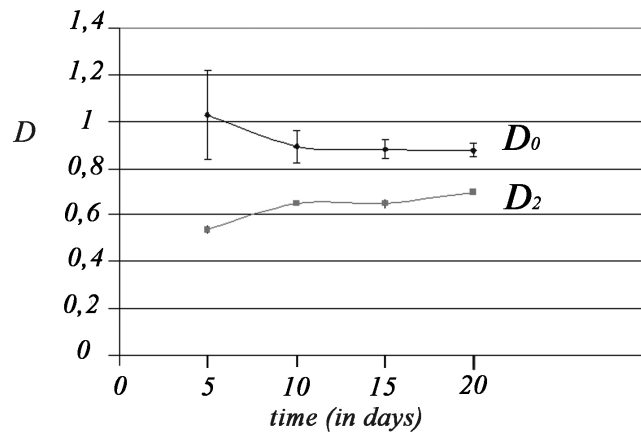
Fig. 6 - Fractal dimensions D_0 and D_2 trend with their uncertainties in the four periods considered.

Δt =1 day, the mean and standard deviations become $n(t)$ and $\sqrt{n}$, respectively. We may expect that the random fluctuations around the mean value $n(t)$ will be within a 2.5σ range (~99%): in particular, data with $\Delta(t)|n_{obs}\text{-}n_{calc}|>2.5\sigma$ have a probability of occurrence less than about 1% (Caccamo *et al.*, 2005, 2007; Bussetti, 1983). We find the anomalies (the ratios of the differences between the observed temporal trend and the theoretical trend) from the best fit, and the standard deviations of the theoretical values. We considered anomalies with Δ/σ values greater than 2.5.

The complete series of the real data is

$$\{n_{obs}(t_j)\} \qquad \text{with} \quad j=1,\ldots, d. \tag{12}$$

Hypothesizing that an anomaly occurs several days before a large aftershock, depending on its magnitude, we introduced a constant shift s=6. This is a starting assumption to avoid including the possible anomalies in the extrapolation (in this temporal range). The predicted theoretical value of the series, obtained by extrapolation, is

$$\{n_{calc}(t_k)\} \qquad \text{with} \quad k=h+s,\ldots, d \tag{13}$$

and h=2·v, where v=2 is the number of the unknown variables, k and p, of the function we use here:

$$n(t)=k\cdot t^p+k_1. \tag{14}$$

The generic element of series (13), $n_{calc}(t_g)$, where g=$h+s$,..., d, is obtained by using the sub-set $\{n_{obs}(t_e)\}$ with e=1,…, g-s (excluding the last six data from the elaboration), where g is the number of time intervals according the sampling time (in this paper the sampling time is 1 day). The value "ten" is the sum of the shift (s=6) and the minimum number of points used for the extrapolation (h=4).
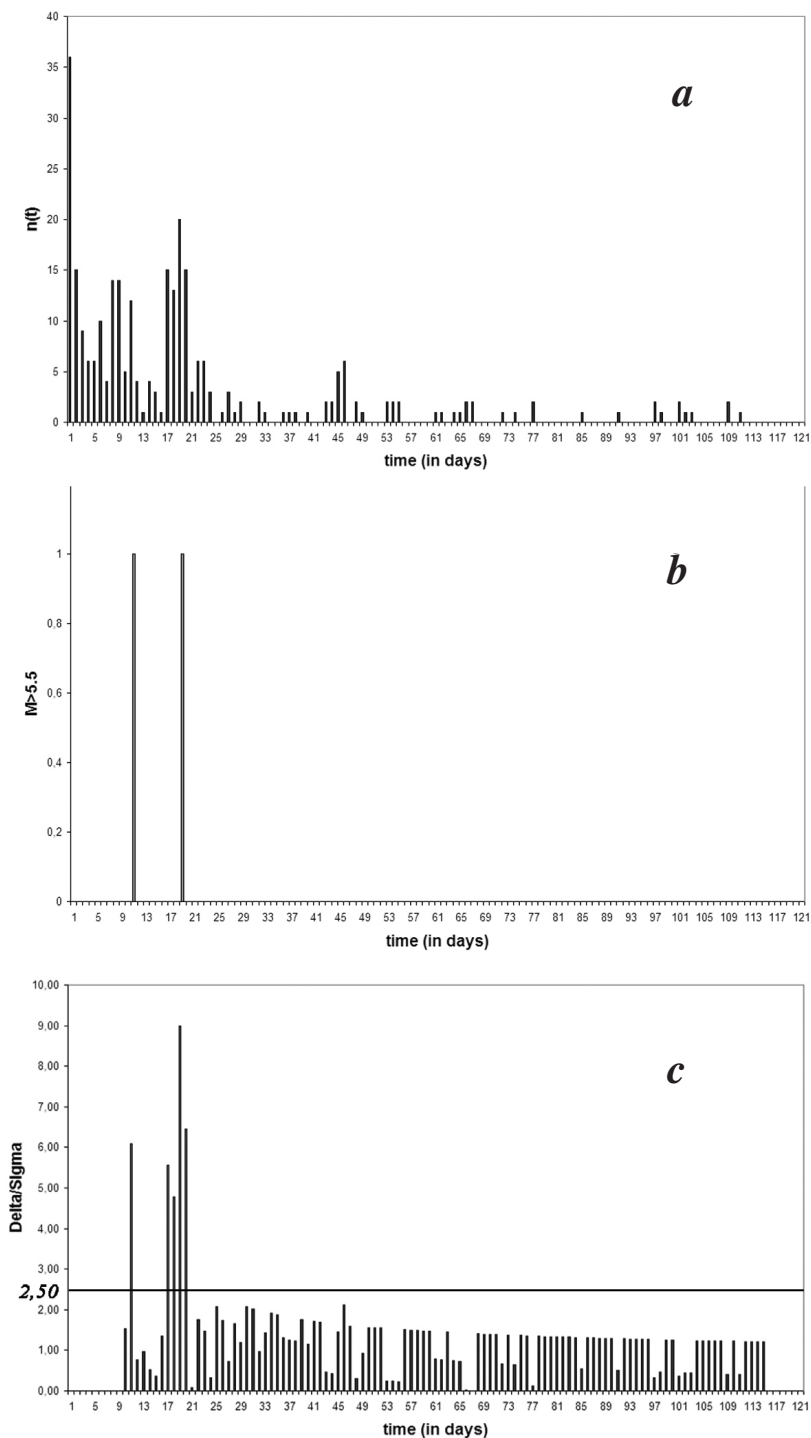
Fig. 7 - a) Temporal trend $n(t)$, number of aftershocks per day, for the events with a magnitude $M \geq M_c$, where M_c=3.1, of the Umbria-Marche sequence; b) temporal trend of the aftershocks with magnitude M>5.5; c) Δ/σ values versus time t, are shown: there are 5 Δ/σ values, in this case, exceeding this threshold: Δ/σ=6.08 on the 11th day, Δ/σ=5.56 on 17th day, Δ/σ=4.79 on the 18th day, Δ/σ=8.99 on the 19th day, Δ/σ=6.45 on the 20th day, with shocks with magnitude M=5.8 and M=5.7 on the 11th and 19th day, respectively.

The differences between the calculated data and the observed data, $\Delta(t_k)$, are obtained by

$$\Delta(t_k)= \mid n_{obs}(t_k)\text{-}n_{calc}(t_k) \mid . \tag{15}$$

When $\Delta(t_k) > 2.5 \cdot \sqrt{n_{obs}(j) \cdot \Delta t}$, we considered this as an "anomaly". So, how can the above be explained, the first point of the extrapolation is relative to t_k=10 days.

We considered an additional constant term K_1 to take into account the background seismicity. This parameter is typical of the seismicity of an area. To evaluate it, we should consider several interseismical periods of the area, to estimate the background seismicity. We assumed, as a starting assumption that K_1=1, since the aftershock decay is a discrete phenomenon and so the function we consider here stretches asymptotically to 1 (Utsu *et al.*, 1995; Caccamo *et al.*, 2005, 2007).

The temporal trend $n(t)$, number of aftershocks per day, is shown in Fig. 7a.

In Fig. 7b shocks with magnitude M>5.5 are shown. In Fig. 7c, the Δ/σ values versus time t (in days) are shown, excluding the initial 10 days.

The three plots of Fig. 7 have the same temporal scale.

There are 5 Δ/σ values exceeding the threshold mentioned above: Δ/σ=6.08 on the 11th day, Δ/σ=5.56 on 17th day, Δ/σ=4.79 on the 18th day, Δ/σ=8.99 on the 19th day, Δ/σ=6.45 on the 20th day, with shocks with magnitude M=5.8 and M=5.7 on the 11th and 19th day, respectively. From Fig. 7, it can be seen that the first anomaly is contemporary to the October 6 (M=5.8) aftershock, while the other ones occurred before the October 14 large aftershock, except the last Δ/σ value, that we consider as a false alarm. We identify a correlation between the Δ/σ anomalies and the occurrence of the large aftershocks.

6. Discussions and conclusions

The aim of this work is the identification of anomalies in the temporal trend occurring in the seismic sequences before a large aftershock. These anomalies are thought to be variations in seismic activity, in a short time interval, before the occurrence of a large aftershock. It is observed that, before the occurrence of the large aftershock, it is possible to indicate these kinds of anomalies in the temporal decay, considered as deviations from the mean value $n(t)$ greater than a 2.5 quantity, that, likely, are not random-like fluctuations but a possible precursor.

The most sensitive parameter for the individuation of seismic anomalies of this aftershock series is the Δ/σ value relative to the temporal trend $n(t)$. The other parameters we use for completeness of information, do not furnish, for this purpose, significant results.

Within the previous approximations, it seems that these fluctuations are consistent with the occurrence of the large aftershocks (Fig. 7b) and the trend of D_0 and D_2 values (Fig. 6) too.

REFERENCES

Aki K.; 1981: *A probabilistic synthesis of precursory phenomena.* In: Simpson D.W. and Richards P.G. (eds.). *Earthquake prediction: an international review,* American Geophysical Union, Washington D.C., pp. 566-574.

Amato A., Azzara R., Chiarabba C., Cimini G.B., Cocco M., Di Bona M., Margheriti L., Mazza S., Mele F., Selvaggi G., Basili A., Boschi E., Courboulex F., Deschamps A., Gaffet S., Bittarelli G., Chiaraluce L., Piccinini D. and Ripepe M.; 1998: *The 1997 Umbria-Marche, Italy, earthquake sequence: a first look at the main shock and aftershocks.* Geophys. Res. Lett., **25**, 2861-2864.

Bottari A. and Caccamo D.; 1981: *A study of the temporal sequence of aftershocks of the 1908 Messina earthquake.* Annali di Geofisica, **34**, 149-160.

Bussetti G.; 1983. *Esercitazioni pratiche di fisica. IV edizione*. Libreria Editrice Universitaria Levrotto & Bella, Torino, 670 pp.

Caccamo D., Barbieri F.M., D'Amico S., Laganà C. and Parrillo F.; 2005: *The temporal series of the New Guinea 29 April 1996 aftershock sequence*. Phys. of the Earth and Plan. Int., **153** (4), 175-180.

Caccamo D., Barbieri F.M., Laganà C., D'Amico S. and Parrillo F.; 2007: *A study about the aftershocks sequence of 27 December 2003 in Loyalty Islands.* Boll. Geof. Teor. Appl., **48** (1), 53-63.

Chiaraluce L., Chiarabba C., Cocco M., Di Bona M., Margheriti L., Selvaggi G., Deschamps and Courboulex; 1999: *A detailed image of the seismic sequence strain relased by the aftershocks of the Umbria-Marche 1997 sequence.* Geophys. Res. Abs., **1**, 96.

Cocco M., Chiaraluce L. and Catalli F.; 2006: *Seismicity migration and time-dependet stress transfer: supporting and conflicting evidence from the Umbria-Marche (Central Italy) seismic sequence.* In: ISM Report on Research and Education. The 4th International Workshop on Statistical Seosmology, Proceedings, Japan, **23**, pp. 44-49.

Coleman T.F., and Li Y.; 1994: *On the convergence of reflective Newton methods for large-scale nonlinear minimization subject to bounds*. Mathematical Programming **67**(2), 189-224.

Coleman T.F. and Li Y.; 1996: *An interior, trust region approach for nonlinear minimization subject to bounds.* SIAM Journal on Optimization **6**, 418-445.

Console R. and Di Giovambattista R.; 1987: *Local earthquake relative location by digital records.* P.E.P.I., **47**, 43-49.

D'Amico S., Caccamo D., Barbieri F., Laganà C.; 2004: *Some methodological aspects related to the prediction of large aftershocks*. In: Book of abstracts E.S.C. 2004, XXIX General Assembly, Potsdam, pp. 132-133.

D'Amico S., Caccamo D., Barbieri F.M., Laganà C. and Parrillo F.; 2005: *Use of fractal geometry for a study of seismic temporal series.* In: European Geosciences Union, General Assembly, Geophysical Research Abstract. Vol. 7, p. 103.

Deschamps A., Iannaccone G. and Scarpa R.; 1984: *The Umbrian earthquake (Italy) of 19 Septmber 197*9. Ann. Geophysicae, **2**, 29-36.

Dennis, J.E. Jr.; 1977: *Nonlinear Least Squares.* In: Jacobs D. (ed), State of the Art in Numerical Analysis, Academic Press, New York, pp 269-312.

De Rubeis V., Dimitriu D. Papadimitriou E. and Tosi P.; 1993: *Recurrent patterns in the spatial behaviour of italian seismicity revealed by the fractal approach.* Geoph. Res. Lett., **20**(18), 1911-1914.

De Rubeis V., Tosi P. and Vinciguerra S.; 1997: *Time clustring of seismicity in the Etna region between 1874 and 1913.* Geoph. Res. Lett., **21**(18), 2331-2334.

Ekstrom G., Morelli A., Boschi E. and Dziewonski A.; 1998: *Moment tensor analysis of the Central Italy earthquake sequence of September-October 1997.* Geoph. Res. Lett., **25**, 1971-1974.

Frohlich C. and Davis S.D.; 1993: *Teleseismic b values; or much ado about 1.0.* Geoph. Res. Lett., **98**, 631-644.

Goltz C.; 1998: *Fractal and chaotic properties of earthquakes.* Springer-Verag Telos, 178 pp.

Gutenberg B. and Richter C.F.; 1942: *Earthquake magnitude, intensity, energy and acceleration.* Bull. Seism. Soc. Am., **32**, 162-191.

Gutenberg B. and Richter C.F.; 1954: *Seismicity of the Earth.* Princeton University Press, Princeton, New Jersey (USA), 310 pp.

Gutenberg B. and Richter C.F.; 1956: *Magnitude and energy of earthquakes.* Ann. Geoph., **9**, 148-149.

Hirata T.; 1989: *A correlation between the b-value and the fractal dimension of earthquakes*. J. Geoph. Res., **94**, 7507-7514.

Hirata T., Satoh T. and Ito K.; 1987: *Fractal structure of spatial distribution of micrifracturing in rock*. J. of Geoph. Res. Abstr. Soc, **90**, 369-374.

Kisslinger C.; 1993: *The stretched Exponential Function as an alternative model for aftershock decay rate*. J. of Geoph. Res., **98**, 1913-1921.

Lei X.; 2003: *How do asperities fracture? An experimental study of unbroken asperities.* Earth and Planetary Science Letters. **213**, 347-359.

Lomnitz-Adler J.; 1992: *Interplay of fault dynamics and fractal dimension in determining Gutenberg & Richter's b-value.* Geoph. J. Int., **107**, 531-541.

Main I.G.; 1991a: *Damage mechanics with long-range interactions: correlation between the seismic b-value and the fractal two-point correlation dimension*. Geoph. J. Int., **107**, 31-541.

Main I.G.; 1991b: *A modified Griffith criterion for the evolution of damage with a fractal distribution of crack lengths: application to seismic events rates and b-values*. Geophys. J. Int., **111**, 353-362.

Mandelbrot B.B.; 1967: *How long is the coast of Britain? Statistical self-similarity and fractional dimension.* Science, **156**, 636-638.

Mandelbrot B.B.; 1992: *The Fractal geometry of nature*. Freeman, San Francisco, 468 pp.

Matsu'ura R.S.; 1986: *Precursory quiescence and recovery of aftershock activities before some large aftershocks.* Bulletin of the Earthquake Research Institute University of Tokyo, **61**, 1-65.

More J.J.; 1977: *The Levenberg-Marquardt algorithm: implementation and theory.* In: Watson G.A. (ed), Numerical Analysis, Lecture Notes in Mathematics, No 630, Springer Verlag, Berlin, pp. 105-116.

Omori F.; 1894: *On the aftershocks of earthquakes*. J. Coll. Sci. Imp. Univ. Tokyo, **7**, 111-216.

Oncel A.O., Main I., Alptekin O., and Cowie P.; 1996: *Temporal variations in the clustering property of seismicity in the North Anatolia fault zone between 31°E and 41°E*. Pageoph, **147**, 147-159.

Page R.; 1968: *Aftershocks and microaftershocks of the great Alaska earthquake of 1964*. Bull. Seismol. Soc. Am., **58**, 1131-1168.

Parrillo F., D'Amico S., Caccamo D., Barbieri F.M. and Laganà C.; 2005: *The "Delta/Sigma method" to evaluate the occurrence of large aftershocks*. Geoitalia 2005, Quinto Forum Italiano di Scienze della Terra, Spoleto, 1.

Schuster H.G.; 1988: *Deterministic Chaos*. VHC Verlag Gesellschaft, Weinheim (D), 312 pp.

Smalley R.F., Chatelain J.L., Turcotte D.L. and Prevot R.; 1987: *A fractal approach to the clustering of earthquakes: application to the seismicity of the New Hebrides*. Bull. Seism. Soc. Am, **77**, 1368-1381.

Turcotte D.L.; 1997: *Fractals and chaos in Geology and Geophysics*. *2nd ed.*. Cambridge University Press, Cambridge, 412 pp.

Turcotte D.L., Newman W.I. and Gabrielov A.; 2000: *A statistical physics approach to earthquakes.* In: GeoComplexity and the Physics of Earthquakes. Geophysical monograph 120, American Geophysical Union, Washington D.C.

Udìas A.; 1999: *Principles of seismology*. Cambridge University Press, 512 pp.

Utsu T.; 1961: *A statistical study of the occurrence of aftershocks.* Geophys. Magazine, **30**, 521-605.

Utsu T.; 1969: *Aftershocks and earthquake statistics (I) source parameters which characterize an aftershock sequence and their interrelations.* J. Fac. Sci. Hokkaido Univ., **3**, 129-195.

Utsu, T., Ogata Y. and Matsu'ura R.S.; 1995: *The centenary of the Omori formula for a decay law of aftershocks activity*, J. Phys, Earth, **43**, 1–33.

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