Seismicity and mean magnitude variations in correlation with the strongest earthquakes of the 1997 Umbria-Marche sequence (Central Italy)

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Abstract. Nearby faults can interact, affecting the timing of a future earthquake. A large earthquake can alter the static stress surrounding faults, possibly activating an aftershock sequence. Here we test the hypothesis of earthquake triggering by examining the earthquakes that affected the Umbria-Marche region (Central Italy) during September 1997 – April 1998. The analysis was performed combining the information contained in the parameters $a$ and $b$ of the Gutenberg-Richter relationship, before and after the strongest mainshocks, using a catalog of 6,211 earthquakes ($M \geq 1.5$). By analyzing the changes in space of the $a$ parameter, we found that Colfiorito mainshocks (5.7 and 6.0 $M_W$) on September 26, 1997 increased the seismicity rate within 40-50 km of the epicenter. We interpret this behavior as the effect of stress redistribution. After analyzing the $b$-value changes in space and time, we found the strongest increase in mean magnitude in the Sellano area, between the occurrence of the Colfiorito shocks and the two strong events (5.2 and 5.6 $M_W$, respectively), which hit this region on October 12 and October 14, 1997. Based on these observations, we postulate that variations in $b$-value just before such events indicate a probable causal connection between the largest earthquakes in the sequence.

1. Introduction

It has been observed that the occurrence of an earthquake changes the stress field of neighboring zones as far as 100 km from the previous mainshock (as observed in western Turkey) [Eyidogan and Jackson, 1985]. Some zones were defined as “stressed up”, in that
increased static stress can increase seismicity rate and trigger moderate to large earthquakes, and some other zones characterized by lack of seismicity were defined in “stress shadow” [e.g. Reasenberg and Simpson, 1992; Harris and Simpson, 1996; Hodgkinson et al., 1996; Stein et al., 1997; Harris, 1998; Toda et al., 1998; Stein, 1999; King and Cocco, 2001]. King et al. [1994] have calculated that a stress increase of less than one-half bar appears sufficient to trigger earthquakes and a similar stress decrease will suppress them. Kagan [1994] posits that incremental shear stress is more likely to induce an earthquake if the stress is coherent with the moment tensor of the future event.

In different regions a clear correlation between locations of large earthquakes and the pattern of previous Coulomb stress change has been found in both transcurrent and normal faulting systems. This pattern suggests a strong elastic interaction among the different faults, favoring their rupture. Some important examples are: the 1981 Gulf of Corinth events (Greece) [Jackson et al., 1982], the 1954 Rainbow Mountain – Fairview Peak – Dixie Valley sequence (Nevada) [Hodgkinson et al., 1996], the 1999 Izmit and Duzce earthquakes (southwest Turkey) [Hubert-Ferrari et al., 2000], as well as the 1992 Landers–Big Bear and 1999 Hector Mine events (California) [e.g. Harris and Simpson, 2002; Pollitz and Sacks, 2002].

Furthermore, Toda et al. [1998] have calculated the change in probability for future earthquakes caused by the stress changes associated with the 1995 Kobe (Japan) shock, analyzing the stress pattern on major faults within 100 km of Kobe. Stein et al. [1997] calculated how much stress change associated with strong events from 1939-1992 along the Northern Anatolian fault altered the earthquake probabilities on neighboring faults.

Important information about the stress state of a volume in which earthquakes occur is also contained within variations in the frequency-magnitude distribution (FMD) of earthquakes:

$$\log_{10} N = a - bM$$  \hspace{1cm} (1)
Where $N$ is the cumulative number of earthquakes with magnitude larger or equal to $M$, and $a$ and $b$ are constants related to activity and earthquake size distribution, respectively.

The $b$ parameter, a measure of the ratio of small to large earthquakes, has been shown to vary temporally and spatially on scales of hundreds of meters to tens of kilometers [Ogata and Katsura, 1993; Wiemer and Benoit, 1996; Wiemer and McNutt, 1997; Wiemer and Wyss, 1997] although the average $b$-value in large volumes tends to show constant values near 1 [Kagan and Jackson, 1991; Frohlich and Davis, 1993].

It has been noted, in ‘retrospective analysis’, that asperity zones are characterized by $b$-values as low as 0.5 with respect to the background level. This perturbation in favor of larger events has been interpreted to indicate higher than average stress, a property thought typical of asperities [Wiemer and Wyss, 1997; Sylvander, 1999; Wyss et al., 2000]. In fact these zones are defined as areas where shear stress is higher than in neighboring zones. This interpretation was also confirmed by works on rock mechanics (such as Scholtz [1968]) to infer that regions of low $b$-values are regions of high stress.

After studying how the June 1992 Landers magnitude 7.3 earthquake (Southern California) influenced seismicity in the neighboring area, Wyss and Wiemer [2000] found good correspondence between the patterns of seismicity rate change and Coulomb stress change. Furthermore, by comparing the $b$-values before and after the Landers earthquake, the authors found a decreased $b$-value associated with an increased seismicity rate that leading to a higher estimate of local probability in the zone of the Hector Mine event (October 1999).

By applying a high-resolution spatial mapping of the frequency-magnitude relationship parameter ($a$ and $b$ values), we may be able to map areas of higher hazard for a future mainshock (i.e., short local recurrence times and consequently lower $b$-values) as well as areas not likely to experience a large rupture. This method implies the hypothesis that equation (1) is valid from small up to large magnitudes.
Using this idea, we analyzed the 1997 Colfiorito sequence, which occurred in Central Italy along the Apennines chain and lasted for several months. It was characterized by six earthquakes of magnitude ($M_W$) larger than 5 in a period of 20 days (Table 1). Here we analyze how the seismicity rate, $b$-value and then the expected rate of large earthquakes changed after two shocks of 5.7 and 6.0 $M_W$ occurred near Colfiorito on September 26. We consider the September 1997 events as a pair (physically related to each other), because the second shock occurred within nine hours, on the same day, and at a small distance, 3 km WNW of the first.

Our goal is to see if the $a$ and $b$ parameters indicate any interdependence among the Colfiorito events pair and the later three large shocks (# 11, 12 and 18 in Table 1). They occurred on October 12 and 14 (5.2 and 5.6 $M_W$, respectively) with a delay of 16 and 18 days, in the Sellano area, 12 km SSE from Colfiorito, and on April 3, 1998 (5.1 $M_W$) near Gualdo Tadino, around 10 km NNW from Colfiorito (Figure 1). Cocco et al. [2000] already found good agreement between the location of the Sellano event on October 14 and the pattern of Coulomb stress change after the Colfiorito pair.

2. The seismic sequence

In 1997 the Umbria-Marche region of Central Italy was hit by a long seismic sequence at started on September 3 with a 4.5 $M_W$ foreshock and was followed by several small events in the following two weeks. The three largest shocks occurred on September 26 close to Colfiorito (5.7 and 6.0 $M_W$) and on October 14 in the Sellano area (5.6 $M_W$) (Figure 1 and Table 1). Their focal mechanism and spatio-temporal distribution indicate that different, parallel normal faults oriented NW-SE with a constant 35°- 45° SW dip ruptured successively in time. Also the aftershocks distribution, covering a 12 x 40 km² area along the Apennines chain’s main direction, is coherent with the ruptures’ orientations [Amato et al., 1998;
Ekström et al., 1998; Deschamps et al. 2000; Morelli et al., 2000]. The earthquakes were mostly located in the crust’s upper 9 km and largely presented a normal faulting mechanism with a NE-SW tension axis (Figure 2) in agreement with the seismotectonic model of the Central Apennines [Calamita and Pizzi, 1994].

Two other earthquakes of comparable magnitude have occurred in this region in the last two decades. The first event (Val Nerina, September 19 1979, 5.7 $M_w$) was located about 30 km SE of Colfiorito; the second (Gubbio, April 29 1984, 5.3 $M_w$) was located about 40 km to the NW of it. Several earthquakes within this area are reported in the historical catalog [Boschi et al., 1997], some with comparable magnitudes in the Gualdo Tadino area (1747, 1751) and others with larger magnitudes to the south of Val Nerina (1703, 1730) (Figure 1). Based on this information, the recent seismic sequence can be considered typical for this area with events of moderate magnitude and fault length not exceeding 15 km. Furthermore, the sequence occurred in a zone where major earthquakes have not been reported since 1279 (I=IX) as shown in Figure 1 [Boschi et al., 1997].

The following features characterize the Colfiorito sequence (Table 1):

- An unusually high number of strong events: 12 earthquakes with $M_L$≥4.5 in the whole sequence, 6 with $M_L$≥5.0;

- An evident SE migration of seismicity toward the Sellano area, since the beginning of October 1997, where the shock of October 12 (5.1 $M_L$) occurred (labeled with # 11 in Table 1);

- Low dip angle of the principal faults;

- A cluster of earthquakes at the northwestern border of the seismogenic zone following the 4.9 $M_L$ earthquake of April 3, 1998, after a period of decreased seismicity rate.
Here we analyze the changes in the $z$-value, $b$-value and rate density of this sequence to find further evidence of correlation with a seismicity-triggering mechanism already suggested by previous studies [e.g. Cocco et al., 2000; Deschamps et al., 2000].

3. Data and methods

For this study we have used data collected by the Italian National Seismic Network (RSNC) of INGV (Istituto Nazionale di Geofisica e Vulcanologia) from July 1987, when a new seismic acquisition system was adopted, to December 2001. The seismological network consists of more than 90 stations equipped with vertical short-period seismometers. We have considered the shallow earthquakes contained in the area indicated by the black rectangle in Figure 1 and delimited by the co-ordinates 42.3° N and 43.6° N in latitude and by the co-ordinates 12.3° E and 13.6° E in longitude, where the 1997 Colfiorito sequence took place. For the whole period analyzed (1987-2001), we selected 6211 events with depths smaller than 70 km and $M \geq 1.5$: the minimum magnitude of completeness estimated for this catalog. This data set is characterized by a $b$-value of 1.09 ± 0.01, obtained through the maximum likelihood method [Aki, 1965]. The estimate of the standard deviation was obtained following Shi and Bolt [1982] (Figure 3). The epicentral accuracy is retained to be typically better than ≈3 km. We assume that the epicentral errors are not a significant function of magnitude. Due to the relatively large distances between the seismic stations, the hypocentral errors are definitely larger than those of the epicentral co-ordinates. The depth distribution of hypocenters in the study region reveals two distinctive peaks at 5 and 10 km (depths fixed by the analysts), containing about 50% of all the events. This depth distribution, common in areas monitored by sparse networks, prohibits any cross-sectional analysis of seismicity. We have analyzed local magnitude, $M_L$, because it is more frequently reported and is the only magnitude calculated for strong earthquakes. The accuracy of the procedure used to determine
$M_L$ has been considered by Gasperini [2001]. Analyzing true Wood-Anderson instruments data (from 1982 to 1989) and simulating Wood-Anderson amplitudes computed from Very Broad Band recordings (from 1990 to 1995), he found that the amplitude attenuation law for Italy does not significantly differ from that proposed by Richter [1935] for Southern California. Moreover, Gasperini [2001] found that the local magnitude could be confidently computed from vertical components using the standard procedure, simply adding 0.1 units to the resulting value. Our choice of using $M_L$ rather than $M_D$ (estimated from the signal duration) is supported by the relation found by Gasperini [2001] between $M_D$ and the standard Wood-Anderson $M_L$. The relative plot exhibits a slope significantly different from 1. The fact that procedures currently in use at INGV underestimate $M_D$ at the highest values and overestimate it at the lowest implies a $b$-value systematically bigger than 1.0: the typical value for Italy in the literature.

We analyzed the homogeneity of the INGV catalog to eliminate the possibility that variations observed in the background seismicity are unreal, that can be caused for example by changes in the data acquisition system or in their processing [Zúñiga and Wiemer, 1999]. To point out these variations with respect to background seismicity, we have declustered the catalog using the Reasenberg algorithm [Reasenberg, 1985]. Then we applied a homogeneity analysis using the magnitude signature technique and GENAS algorithm [Habermann, 1983, 1987]. This approach evaluates stationarity of magnitude reported in different magnitude bins with time. The basic hypothesis assumed herein this analysis is that the background seismicity rate and the $b$-value are constant on a large scale.

We found a magnitude shift of 0.4 units affecting $M_L$ in 1996, probably due to changing the type of amplifier and filter in the seismic stations. The apparent decrease in seismicity rate after this year is shown in Figure 4a. Based on results obtained by Gasperini [2001], we should have increased all the $M_L$ in the raw catalog after 1996. But we prefer to correct the
period 1987-1996, with a constant value for the $M_L$ of −0.4, to obtain a homogeneous catalog in case new data are added in the future. The effect of this correction is visible in Figure 4b. Note, however, that even if the applied correction results in a homogeneous catalog, results homogeneous after the applied correction, the problem of determining the absolute value of local magnitude remains open.

Lastly, some missing values of $M_L$ for events in which $M_D$ was available have been calculated using an empirical linear fit relation between $M_D$ and the corrected $M_L$.

Afterwards, for the successive analysis to calculate $b$-value and local occurrence rate density before and after the Colfiorito pair (# 2, # 3 as labeled in Table 1) and the Sellano earthquake of October 12 (# 11 as labeled in Table 1), we focused on a smaller area of 35 km x 55 km to include just the sequence containing 3,024 events with $M \geq 1.5$ (Figures 8, 9, 11 and 12).

To analyze our dataset we have used the ZMAP software package [Wiemer and Wyss, 1994], that allows computation and mapping of $a$-value, $b$-value, and other seismological parameters as a continuous function of space, time and magnitude by creating dense spatial grids of such parameters. Furthermore, this package enables parameter comparison for different zones or periods of time. ZMAP calculates the statistical significance of these results by using several statistical tests.

As mentioned, we used this technique to focus on variations in seismicity rate, $b$-value and local occurrence rate density ($\lambda_L$) after the two Colfiorito mainshocks to see if these parameters give some useful information. Do they correlated with the preparation of the Sellano large earthquakes and the following April 3, 5.1 $M_W$, event? We represented the change in $a$ parameter by the $z$-value, which is related to the statistical significance of the rate change at a given time, $t_c$, by the standard deviate $Z$ test [Habermann, 1983; Wyss and Burford, 1985] with
Here $\_\_ \$, $\sigma$ and $n$ are respectively the mean seismicity rate, the standard deviation and the number of samples in the two periods (before and after $t_c$ indicated with subscript index 1 and 2, respectively). The value of $z$ measures the significance of the difference between the two mean seismicity rates. The larger the difference between the two means, $\_\_ \$, and the larger the numbers of samples, $n$, before and after $t_c$, the more confidence one can place in the significance of that difference, whereas large variances of the means, $\sigma$, decrease the confidence. The $z$-value is generally estimated from declustered catalogs.

Starting from the frequency-magnitude distribution (eq. 1), one estimates the local occurrence rate density $\lambda_L(M)$ of an earthquake with magnitude equal to $M$ and larger in a given circle of area $A$ by

$$\lambda_L(M) = 10^{(a-bM)} / (T A) \quad (3)$$

where $T$ is the period over which $N$ earthquakes occurred within the circle.

Obviously, $\lambda_L(M)$ changes when $a$ and $b$ change. In this paper, we map $\log \lambda_L$ (for a given magnitude), i.e. the logarithm of the ratio of $\lambda_L$ before and after a particular event as a function of space, combining the information contained in the $a$ and $b$ values and testing the hypothesis that its increase may be correlated with the occurrence of mainshocks. Hence a $\log \lambda_L$ of 2 means a 100 times increase in $\lambda_L$.

After a preliminary declustering of the catalog, we generated the maps of $Z$ and $\log \lambda_L (M \geq 5)$ before and after the Colfiorito pair using a grid with node spacing of 1 km. We sampling a minimum number of 50 earthquakes, in cylindrical volumes, within a fixed radius of 5 km around each node. To map the $b$-value during different time periods, we used a grid with node spacing of 1 km and sampled a fixed number of 100 earthquakes, occurring within a radius ranging between 2 and 10 km. The parameters used for the analysis shown in this paper (grid
size, sampling radius, minimum number of events, etc.) represent a choice obtained after numerous tests.

4. Results and Discussion

By analyzing the seismicity rate variations throughout the study region, we have compared the period before the Colfiorito pair, from July 1987 to September 25, 1997, with the following period including the Colfiorito events and lasting until December 31, 2001. The declustered catalog contains, in this area and for the entire period considered, 1823 events with $M_c \geq 1.5$. We obtained the map of $z$-values, reported in Figure 5. It shows how seismicity increased (indicated by red) after those events almost everywhere in the area with larger increases in the zones of the larger shocks and where the aftershocks concentrated. Similar results have been obtained with the un-declustered catalog that has not been declustered. The resolution maps for the background and the foreground periods (Figure 6, a and b) outline the areas where a statistical analysis is possible. The central part of these areas is characterized by a resolution better than 10 km. This value indicates the smallest sizes that the method can allow to investigate. After analyzing the INGV catalog preceding the Colfiorito events, Console et al. [2000] found a marked clear area of quiescence nearly overlapping the epicenter of the Colfiorito mainshocks. The quiescence period, with no events of $M_D > 3.2$, started in June 1994 and lasted till the sequence began. The greatest decreases in seismicity rate were located in the Sellano area, the zone that failed successively, indicating that the affected volume was primed for rupturing.

The $b$-value maps were created using the maximum likelihood method proposed by Utsu [1965; 1967] on the original undeclustered catalog (not declustered).
To study the variation in the rate of small to large magnitude earthquakes, we retain the original un-declustered catalog, because the declustering process affects the results by removing mostly small earthquakes.

By mapping the $b$-value for the whole region and period (Figure 7a), one can see how this parameter changes significantly in space, with minimum values in the area of larger shocks (inside within a maximum radius of 15 km). In Figure 7b the error associated in calculating the $b$-value is mapped for each grid node.

Furthermore, by dividing the entire period into different temporal windows delimited by the strongest shocks (# 2-3, # 11, # 18 as labeled in Table 1), we have found that the $b$-value also changes in time (Figure 8). In particular it is very low we observe a very low $b$-value ($\sim 0.5$) in the Sellano area after the Colfiorito mainshocks (Figure 8b) compared with the periods before Colfiorito (Figure 8a). So the Colfiorito mainshocks in September 1997 have affected the ratio of small to large events in favor of big earthquakes within about 10 km of the epicenter.

We tested the significance of this difference, noted in two time periods, using Utsu’s method [Utsu, 1992] sampling the volumes A and B (11.29 and 6.20 km in radius, respectively) (Figures 8a and 8b, respectively) at the same point of coordinates (42.9°N - 13.0°E) but for a different time period of time. The probability that these two samples come from the same population is very low ($P = 1.2 \times 10^{-4}$) (see Figure 9).

As noted remarked by Wiemer and Katsumata [1999] and by Wiemer et al. [2002], the increase in the magnitude of completeness, $M_c$, in an aftershock sequence soon after a mainshock, is an important aspect for analyzing in the analysis of seismicity. In particular, one would expect that this increase would cause an apparent reduction of the $b$-value. To consider this aspect in the present analysis, we mapped $M_c$ in the four periods of time selected for Figure 7a-d (Figure 9). Actually, by comparing from a comparison between
the corresponding maps of Figures 7 and 9, one might infer a negative correlation between $M_c$ and $b$. However, we may note also that this circumstance doesn’t change the reliability of the Utsu test, which is based only on the magnitudes above the actual completeness threshold.

In Figure 11a, we compared the period that preceded the Colfiorito mainshocks with the period that followed the Sellano event. For events of $M_L \geq 5.0$, we found a strongly increased rate of occurrence south of the Colfiorito to the Sellano area, for events of $M_L \geq 5.0$. In the volume of the Sellano mainshocks the increase was nearly 200-times. The significance of this change in $\lambda_L$ is above the 95% confidence limit according to Utsu [1992] (see Figure 11b).

We propose that the concentrated aftershocks in the southern part of the Colfiorito area transferred stress into the Sellano region, already in a state of precursory quiescence and in preparation for failure [Console et al., 2000]. This zone broke later with bigger magnitude as if that area was more resistant and needed a larger stress increase to be able to fail. Only after the stress built up by the aftershocks exceeded the strength of the Sellano patch could the seismicity propagate to the south-east (Figure 8c). In this respect, the usual aftershock model as described by the modified Omori law [Utsu, 1961] does not apply to this sequence.

The fact that the $b$-value decreased (Figure 8b) before the Sellano $M_W 5.2$ event indicates anomalously large average magnitudes for the Colfiorito aftershocks were anomalously large there. These observations suggest that the Colfiorito aftershocks may have contributed to triggering the Sellano events. For example, the October 4 event (# 7 in Table 1 and Figure 2) could be interpreted as a Colfiorito aftershock or a foreshock to the Sellano events. The idea that aftershocks, rather than the mainshock, can trigger later big events has also been proposed by Felzer [2001] for the 1999 $M_W 7.1$ Hector Mine earthquake.
The area around Sellano where we found the low $b$-value (Figure 8b) and the significantly increased $\lambda_L$ after the Colfiorito mainshocks (Figure 11a) roughly matches fairly consistent with that where Cocco et al. [2000] calculated a Coulomb stress increase of about 0.5 bars, caused by the Colfiorito’s first foreshock and two mainshocks. We suggest may suppose that faults favorably oriented to fail in the Sellano area where stress was transferred after the Colfiorito mainshocks enabled triggering like those reported in southwest Turkey [Eyidogan and Jackson, 1985], Dixie Valley (Nevada) [Hodgkinson et al., 1996] and Greece [Hubert et al., 1996]. This idea is supported by several recent papers showing the possibility of earthquakes triggered by a static Coulomb stress change on the order of a few tenths of a bar. Such an effect has been observed for short terms, typically of some days to several months, in as-for aftershock sequences (e.g. Harderbeck et al. [1998], Toda et al. [1998], Seeber and Armbruster [2000]). Other papers deal with a longer time delay, on the order of tens of years, between pairs of large events (e.g. Stein et al. [1997], Nalbant et al. [1998], Pollitz and Sacks [2002], Papadimitriou [2002]).

Fault frictional properties and rheology can explain the short time delay observed in most aftershock sequences [Boatwright and Cocco, 1996; Gomberg et al., 1998]. Other processes such as fluid flow, viscoelastic relaxation [Hudnut et al., 1989; Freed and Lin, 1998] and creep must be considered to justify the much longer time needed for fault interaction and stress transfer in phenomena like seismicity migration. The Colfiorito mainshocks have been followed by the progressive activation of the Sellano area over a couple of weeks. This time delay is too short to involve a process of viscoelastic stress relaxation, as invoked by Pollitz and Sacks [2002] for the 1999 Hector Mine earthquake after the 1992 Landers shock. In fact, due to the viscosity of the lower crust and of the mantle, a process of post-seismic viscoelastic stress migration is generally characterized by a time delay on the order of tens of years. Jaumé and Sykes [1992] and Simpson and Reasenberg
[1994] have argued that post-seismic fluid flow could raise the effective coefficient of friction, causing long-term increases in the static stress changes. The presence of fluid during the Colfiorito sequence has been previously suggested based on seismological [Ripepe et al., 2000], geochemical [Quattrocchi, 1999] and fault zone data [Cello et al., 2001a/b]. One might suppose that the first ruptures favored fluid migration along the strike of the main structures. A fluid presence also supports the idea of reactivating pre-existing thrust planes to justify the low-angle dip for larger shocks in the sequence [Cello et al., 2000]. Furthermore, for other long extensional seismic sequences characterized by several events of similar magnitude, it has been recognized an important role of fluid diffusion processes in non-events triggering events has been recognized[Nur and Booker, 1972; Noir et al., 1997].

After considering the aftershocks distribution [Amato et al., 1998] and analyzing the rupture directivity on broadband waveforms [Pino et al., 1999 and Pino and Mazza, 2000], we have inferred that the rupture associated with the Colfiorito mainshock (UT time 00:33) propagated southwards, while that due to the second large event (UT time 09:40) went northwards. Given these different directions of rupture propagation, we looked for any correlation with the large April event in April. In Figure 12(a, b) where we compared the periods from July, 1987-September 25, 1997 to September 26, 1997-April 3, 1998 into mapping the variation of local occurrence rate density, we found did not find a significantly clear result to support for a possible influence from the Colfiorito pair. Figures 8d and 10d show that from April, 1998 to the end of 2001 the whole study area has been characterized by a high b-value and a normal magnitude of completeness.

5. Conclusions

By studying the seismicity off the last 13 years in a region of Central Italy, we have shown how the a and b parameters of the frequency-magnitude distribution change significantly in
space and time, at short scale, especially in relation to an important seismic sequence. The $b$-values were anomalously low, indicating an anomalously large average magnitude in the regions later hit by the Sellano and Gualdo Tadino earthquakes (Figures 8b and 8c, respectively). In particular, we show that just after the larger mainshocks of the sequence, at a distance of about 12 km where two large events occurred about two weeks later, the ratio of small to large earthquakes decreased. This decrease caused an anomalous low $b$-value (Figure 8b). The combined effects of the changed $a$-value and $b$-value changes resulted in high values for the theoretical rate of $M_L \geq 5.0$ events near the sites of the future Sellano and Gualdo-Tadino earthquakes (Figure 11a). Anomalously large and frequent aftershocks of the Colfiorito events may have helped trigger the Sellano and Gualdo Tadino events. Based on these observations and a previous study about concerning Coulomb stress variations in the same area, we suggest a possible correlation between the September Colfiorito pair and the events of October near Sellano. Explicit stress history calculations for this area by Cocco et al. [2000] show that areas of anomalously high seismicity and average magnitude were also regions of highly resolved stress changes.

Evidence of fluid migration in the study area can justify a delay of about two weeks for the Sellano events, as well as seismic progression, the migration of seismicity along the main geological structures.

Statistical analysis of the $b$ parameter and local rate density change before and after a target event, even without a clear explanation of the physical processes connected with the earthquake preparation, gives useful information about time-dependent hazards.

We are confident that mapping changes in $z$-value, $b$-value and $\lambda_{l,5}$ as a function of space and time, will help identify locations where large earthquakes are more probable. The parameters may indicate resolved stress and may contain information not revealed in traditional stress history calculations. For example, our values may be sensitive to stress
changes from earthquakes too small to compute explicit stress effects. The results of this study favor the use of $b$ values in a local spatial context as well as temporal context.

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Figure Captions

Figure 1. Epicenter map of shallow seismicity in black ($M \geq 1.5$) for the Central Apennines region (July 1987-December 2001). Historical seismicity is indicated with red squares (the numbers indicate the years of occurrence). The rectangle traced delimits the area under study. The principal mainshocks of the 1997-1998 sequence occurred in the Colfiorito, Sellano and Gualdo Tadino areas are indicated with numbers (see for details Table 1). The two strongest events in the last two decades before 1987 are shown by triangles.

Figure 2. The Harvard centroid moment tensor (CMT) fault plane solutions for the 20 strongest shocks (labeled by their date) of the sequence. The list of events is reported in Table 1. The focal mechanisms of the four time periods considered in our analysis are indicated with different colors. The letters from A to D in the legend refer to the groups listed in Table 1. The radius of each focal sphere is proportional to the moment magnitude. (Redrawn after Morelli et al., 2000).

Figure 3. Frequency-magnitude distribution (FMD) for the area indicated by the black rectangle in Figure 1.

Figure 4. (a ) Cumulative frequency-magnitude distribution per year (upper panel) and non-cumulative frequency-magnitude distribution (down panel) for the period 1987.5-1996 (‘0’) and 1996-2001.99 (‘x’). The decrease in rate detected after 1996 is clearly visible. (b) Same as (a) after the catalog correction of the magnitude shift –0.4 in the period 1987.5-1996.

Figure 5. Map of seismicity rate variations after Colfiorito mainshocks. The period from July 1987 to September 25, 1997 is compared with that from September 26, 1997 to December 31, 2001. For this analysis we have correlated earthquakes from the two periods within a radius of 5 km around each node of a grid with 0.01° spacing. The negative values of $z$, shown in red, indicate a significantly increased seismic activity in the second period. The numbers labeled near yellow stars are referred to the mainshocks reported in Table 1. The principal towns of the area, mentioned in the text, are also indicated. The rectangle shows the area of the specific sequence analyzed in this study.

Figure 6. (a) Resolution map for the background period (July 1987 – August 1997). (b) Same as (a) but considering the foreground period (September 1997 – December 2001).
The colors indicate the minimum radius of a circle including 50 events at each point coordinates.

**Figure 7.** (a) Map of \( b \)-values (i.e. mean magnitude of earthquakes) for the area under study defined in Figure 1. The seismicity for the whole period (1987-2001, 6211 events with \( M \geq 1.5 \)) has been analyzed, using a grid with 0.01° spacing for the 150 nearest earthquakes in radii between 1.6 and 15 km. (a) Map of the error associated with the \( b \)-values.

**Figure 8.** \( b \)-value maps for four different periods. For the analysis we used a grid with 0.01° spacing sampling the 100 nearest earthquakes, within a maximum radius of 10 km. (a) Period from July 1987 to September 25, 1997 (459 events with \( M \geq 1.5 \)). (b) Period from September 26, 1997 (including the September Colfiorito pair indicated with red stars) to October 12, 1997, a few hours before the strongest shocks in the Sellano area (green stars) (1366 events with \( M \geq 1.5 \)). (c) Period from October 12, 1997 (including the Sellano strongest events indicated with green stars) to April 3, 1998 before the shock of April 3 (blue star). This subset includes 1,662 events with \( M \geq 1.5 \). (d) Period from April 3, 1998 (including the Gualdo Tadino shock indicated with blue star) to December 31, 2001 (832 events with \( M \geq 1.5 \)). The number labeled near the stars are refered to Table 1. In Figs. 5a and 5b the letters A and B, respectively, indicate circles with a center in the same point (long. 13.0° E, lat. 42.9° N) where we calculated the Utsu’s test reported in Figure 9.

**Figure 9.** Frequency-magnitude distribution of earthquakes in two cylindrical volumes defined by the letters A and B in Figure 5a and 5b, containing 100 and 60 events, respectively. The \( b \)-values of the two sets are also indicated. The \( P \) probability according to Utsu’s test is also reported.

**Figure 10.** Maps of the spatial distribution of \( M_c \), computed by measuring the deviation from an assumed power law, for the four different periods. Maps were obtained by sampling events within a 10 km radius from each node. (a), (b), (c), and (d) refer to the same periods as in Figure 8 (a-d), respectively.

**Figure 11.** (a) Change in local occurrence rate density (\( M_L \geq 5.0 \)), obtained by comparing the period from September 26, 1997 (after the Colfiorito pair) to October 12, 1997, just before the \( M_w5.2 \) earthquake (1,366 events with \( M \geq 1.5 \)), with that from July, 1987 to September 25,
1997. We analyzed 459 events with $M \geq 1.5$. For the comparison we used a grid with 0.01° spacing and analyzed a minimum number of 50 events in a fixed radius of 5 km around each node. The Colfiorito (5.7 and 6.0 $M_w$) and Sellano (5.2 and 5.6 $M_w$) are indicated by colored stars and labels. (b) Significance of the local occurrence rate density change, as mapped in (a).

**Figure 12**. (a) The same as Figure 11a but for a longer period of time, after the Colfiorito mainshocks. We have compared the period from September 26, 1997 to April 3, 1998, just before the $M_w 5.1$ earthquake (3,024 events with $M \geq 1.5$), with that from July 1987 to September 25, 1997. This subset includes 459 events with $M \geq 1.5$. For the comparison we used a grid with 0.01° spacing and analyzed a minimum number of 50 events in a fixed radius of 5 km around each node. The Colfiorito (5.7 and 6.0 $M_w$) and Gualdo Tadino main shocks (5.1 $M_w$) are indicated by colored stars and labels. (b) Significance of the local occurrence rate density change, as mapped in (a).
Table Captions

Table 1. Main parameters of the 20 major shocks of the sequence. The letters reported refer to the events included in the four periods considered. (Redrawn after Morelli et al., 2000).