Space-time variations of the Umbria-Marche region instrumental seismicity

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Abstract
In the Umbria-Marche region, space and time variations concerning the *b* value were studied by instrumental seismicity from January 1987 to May 1999, according to the Bender method. Data were divided into two partially independent data sets. The first set, (January 1987 - December 1996), does not include the Colfiorito seismic sequence that occurred in the autumn of 1997. The second data set includes all events from January 1987 to May 1999. Using square cell dimensions of 80, 40 and 20 km, the examined area was divided respectively into three grids. The *b* value was estimated for each cell using the first data set, thus allowing us to reveal *b* value space variations and determine the resolution. To evaluate the stability of our result we estimated the *b* value on the basis of historical seismicity within the region. Several synthetic tests were also performed to estimate the stability of the Bender method and to verify its consistency with respect to other methods commonly used. Finally we estimated the *b* values using the second data set to prove the time variations. Results from the area examined show that the lowest possible spatial resolution of the *b* value is about 40 km and that there is a correlation between the *b* value pattern and the main active tectonic structures of the area. The most important time variations occur within the Colfiorito area, in which the *b* value drops significantly within the second data set. Results suggest two different ways of strain release: the first one produces continuous seismicity that spreads all over the examined area, while the second, which concerns stronger earthquakes, is localized.

Key words *b value – seismicity – Apennines*

1. Introduction

The study of seismicity is an important factor in the evaluation of seismic hazard and to better understand the different geodynamic aspects of a region (Papazachos, 1999, and references therein). Seismicity, by means of the *b* parameter of the Gutenberg-Richter relation, also assists with the determination of return time for seismic events of a given magnitude.

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The seismotectonic setting of the Umbria-Marche region in Central Italy (fig. 1) is characterized by three deformation patterns along the Apennines arc (Lavecchia *et al.*, 1994; Boncio *et al.*, 1998). Along the Adriatic coast-line, a contractional deformation of the outer domain combines Pleistocene layers into the fold and thrust system, whereas two styles of extensional deformation prevail in the two inner domains of the Tuscan ridges and the Umbria pre-Apennines. Both *T*-axes in the inner domains and *P*-axes in the outer domain are oriented SW-NE or WSW-ENE and this coaxiality reflects the situation depicted by the geological strain field. By way of geophysical surveys (Ponziani *et al.*, 1990), together with structural geological inspections performed within the

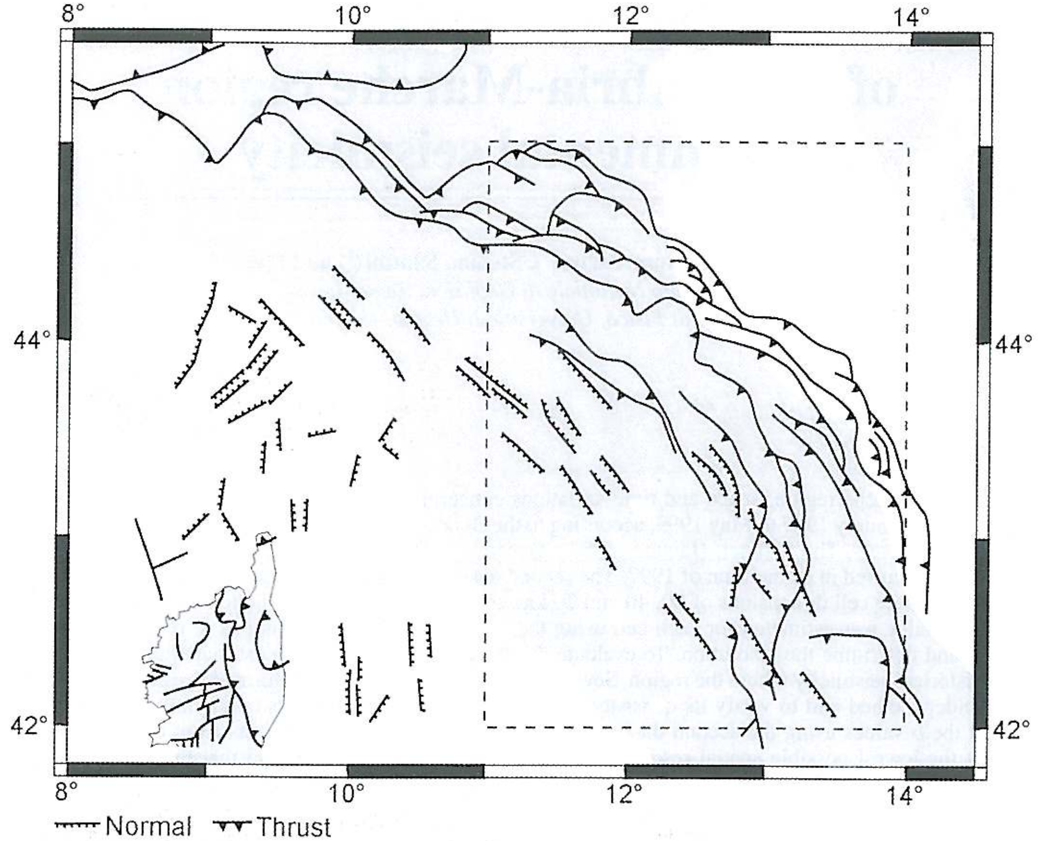


Fig. 1. Main structural features of the Central and Northern Apennines (barbed lines: normal faults, lines with triangles: thrusts). The dashed box evidences the studied region (modified from Ambrosetti *et al.*, 1987).

Central-Northern Apennines (Menichetti, 1991; Pialli and Alvarez, 1997; Pialli *et al.*, 1998; Boncio and Lavecchia, 2000), it is possible to determine a complex tectonic style, with adjacent zones having different sismotectonic characteristics (Malinverno and Ryan, 1986; Frepoli and Amato, 1997). This phenomenon is reflected in stress field heterogeneities and, as a consequence, in variations of the b-value (Gutenberg and Richter, 1944).

In literature, some authors (Urbancic *et al.*, 1992; Wiemer and Wyss, 1997; Wyss *et al.*, 1997) have emphasized that b value anomalies higher than the usual ($b \approx 1$) are connected to variations in stress, or in the fragmentation condition – both present in the examined area. As a rule, zones with homogeneous seismogenic settings should be chosen to estimate the b parameter to enable assessment of the seismic hazard level. However, determination of these zones is not easy if both the geologic and seismotectonic settings of the studied area change within a range of a few kilometres due to surface heterogeneities. In this case, division into zones cannot be made on the basis of seismic distributions and seismotectonic knowledge alone. If such complex areas are divided into zones, or cells, it is reasonable to suppose that the b value may vary slightly if some spatial correlation between adjacent cells is made (Papazachos, 1999).

The studied area complies with the above mentioned case due to its complex structural setting. The aim of our study is to determine the minimum zone, or cell dimensions required to obtain a robust estimation of the b parameter of the Gutenberg-Richter relation, independently from the structural and seismotectonic aspects. Furthermore, the result could allow us to consider and to evaluate the variations in the b value between adjacent cells and in two different time intervals.

2. The method

It is possible to determine the connection between seismic event frequency and magnitude, by way of the empirical Gutenberg-Richter relation

$$\ln N(m) = \alpha - \beta m \tag{2.1}$$

where $N(m)$ is the (non-cumulative) number of seismic events having a determined magnitude equal to m within an observed time period, and where α and β are linked to the well-known a and b parameters by the relations

$$\alpha = (\ln 10)\, a$$

$$\beta = (\ln 10)\, b.$$

It is necessary to reduce the seismic events observation field by selecting the minimum value of magnitude, say m_0, above which the eq. (2.1) is experimentally verified.

Various methods to determine the β value and its confidence limits are found in literature; most of them generally set the upper limit as an infinite magnitude (Aki, 1965; Cornell, 1968; Weichert, 1980). However, events have a maximum dimension, so the Gutenberg-Richter relation is also limited on the upper side (Cosentino *et al.*, 1977; Bender, 1983). The lower limit, as previously stated, is experimentally determined; the magnitude of the upper limit (m_{max}) is set considering the historical seismic events catalogues (see Section 3).

Supposing that seismic events are independent in magnitude and regarding them as random variables, the probability distribution in the magnitude range $m_0 \div m_{max}$ (for magnitude M values smaller than a fixed magnitude m)

$$F_M(m) = P(M < m \,|\, m_0 < m < m_{max})$$

may be written as (Der Kiureghian and Ang, 1977)

$$F_M(m) = \frac{N(m_0) - N(m)}{N(m_0) - N(m_{max})}$$

hence

$$F_M(m) = \frac{1 - \exp[-\beta(m - m_0)]}{1 - \exp[-\beta(m_{max} - m_0)]}. \tag{2.2}$$

However, these statistical methods consider the Gutenberg-Richter empirical relation as a continuous function of the magnitude for grouped data. In order to estimate the b value, it is necessary to divide the magnitude range, $m_0 \div m_{max}$, into n equal intervals of width Δm. The probability P_i that an event occurs in the i-th interval is a factor $\exp(-\beta \Delta m)$ less than the probability P_{i-1} that an event will occur in the interval $(i-1)$-th (Bender, 1983), *i.e.*

$$P_i = P_{i-1} \exp(-\beta \Delta m).$$

Considering expression (2.2) we can rewrite the probability as follows:

$$P_i = \frac{1 - \exp(-\beta \Delta m)}{1 - \exp(-n \beta \Delta m)} \exp[-(i-1)\beta \Delta m]. \tag{2.3}$$

Since events are independent, the probability P_i for N observed data to be distributed in n magnitude intervals is a multi-nominial distribution

$$f(k_1, k_2, \ldots, k_n) = \frac{N!}{\Pi_{i=1}^{n} k_i!} P_i^{k_i}$$

where P_i is the probability that an event occurs in the i-th class, and k_i the number of events of the same class. Considering the expression (2.3) we can rewrite the probability distribution as follows:

$$f(k_1, k_2, \ldots, k_n) =$$

$$= \frac{N!}{\Pi_{i=1}^{n} k_i !}\left[\frac{1-\exp(-\beta\Delta m)}{1-\exp(-n\beta\Delta m)}\right]^N \cdot$$

$$\exp\left[-\beta\Delta m\sum_{i=1}^{n}(i-1)k_i\right].$$

The most probable estimate for β is the value for which $\partial \log f/\partial\beta$ is null

$$\frac{\exp(-\beta\Delta m)}{1-\exp(-\beta\Delta m)} - \frac{n\exp(-n\beta\Delta m)}{1-\exp(-n\beta\Delta m)} -$$

$$-\sum_{i=1}^{n}\frac{(i-1)k_i}{N} = 0.$$

The number of events, $q(m_i)$, within a magnitude interval $m_i - \Delta m/2 \leq m \leq m_i + \Delta m/2$ may be written

$$q(m_i) = N\int_{m_i-\Delta m/2}^{m_i+\Delta m/2} f(m)dm = \exp(\alpha - \beta m_i),$$

where N is the total number of events in the range $m_0 \leq m \leq m_{max}$, and hence

$$\exp(\alpha) =$$

$$= \frac{N\exp[\beta(m_0 + \Delta m/2)]}{1-\exp[-\beta(m_{max} - m_0)]}[1-\exp(-\beta\Delta m)].$$

The error on b has been widely treated in Bender (1983); for our purposes it is enough to consider $\sigma_b \approx 0.1b$ when $N \geq 70$. As far as the error δa on a is concerned, for the sake of simplicity we assume δa to be the maximum deviation in the Bender method, and δa to be the standard deviation for least-squares and maximum-likelihood methods.

For estimation and stability of the b value, we generated five synthetic tests (fig. 2a-e). In the hypothesis that events might be independent (or rather they might follow a poissonian distribution), we compared results of Bender's Method (hereinafter BM) with the results of other frequently used statistical methods, Least Squares (LS) and Maximum Likelihood (ML).

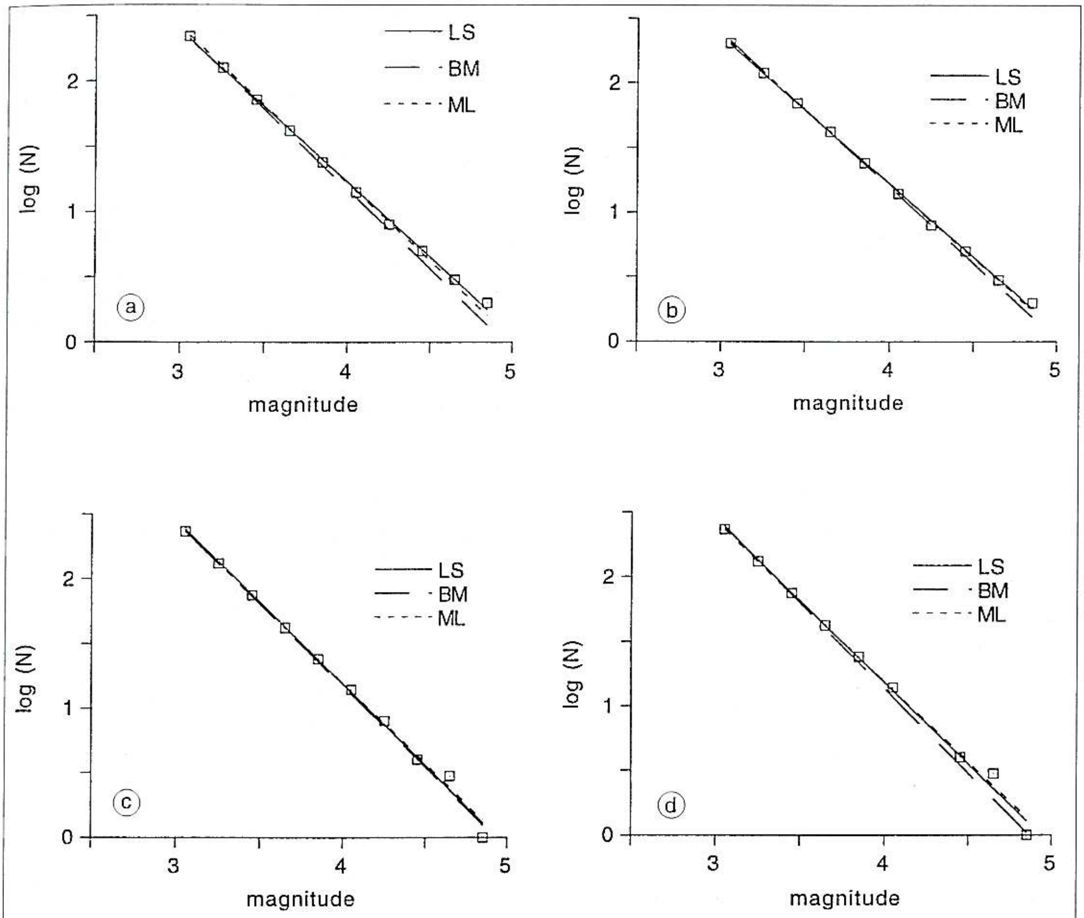
In performing such tests we attempted to simulate situations that could be considered as realistic. In all cases we produced a distribution of events following the Gutenberg-Richter rule, with pre-fixed values of $b = 1.2$ and $a = 6.0$, in a magnitude range between $m = 3$ and $m = 5$. Each result was then divided into magnitude classes with $\Delta m = 0.2$. We then added «noise» to the resulting distribution, either changing at random the number of events in some magnitude classes, or omitting classes entirely in some instances.

In the first test (case A), only approximation errors are present in the distribution. If we use the least squares method, the b value ($b = 1.15$) is very different from its pre-fixed value. It is possible to obtain a better estimate ($b = 1.19$) using the maximum likelihood method, whereas the Bender method gives an estimate of $b = 1.22$ (fig. 2a).

In case B (fig. 2b), lower magnitude classes were undervalued, that is the event number is less than in case A, leaving other classes unchanged. Using this approach, the b value decreases and the greatest variation of 0.04 occurs with the maximum likelihood method.

In case C (fig. 2c), lower magnitude classes were overestimated and upper magnitude classes were underestimated. For this test, the b value increases and the most notable variation ($\Delta b = 0.11$) is due to the least squares method. To allow us to simulate a situation in which the events distribution is referred to an excessively short time period, we set the events number in an upper magnitude class to zero, reflecting an incomplete catalogue.

Following on from test C, distribution results in cases D (fig. 2d) and E (fig. 2e) were obtained by using zero settings in the magnitude classes $m = 4.25$ and $m = 4.65$ respectively. The b value does not change in either test when using the maximum likelihood method, whereas there is a variation of 0.03 in only one case when using the least squares method. However, the b value alters considerably when following the Bender method, with a variation of $\Delta b = 0.06$ in the case D and $\Delta b = 0.02$ in the case E. Results for all three methods have been summarized in table I.



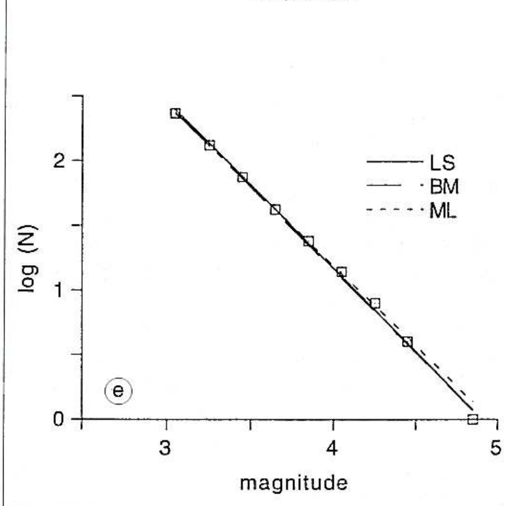


Fig. 2a-e. Frequency-magnitude distribution of N events with a pre-fixed $b = 1.2$ value, computed using the Least Squares (LS), Bender Method (BM), and Maximum Likelihood (ML) technique. a) case A, initial calculus; b) case B, underestimating the lower magnitude classes; c) case C, overestimating the lower magnitude classes and underestimating the upper ones; d) case D, the same as case C without the 4.2 magnitude class; e) case E, the same as case C without the 4.6 magnitude class. Results in table I. See text for further details.

Table I. Values of b and a and associated errors for the 5 tests shown in fig. 2a-e. δ is the standard deviation for LS and ML, or the maximum error for BM.

	$b \pm \sigma_b$			$a \pm \delta_a$		
Method	LS	BM	ML	LS	BM	ML
Case A	1.15 ± 0.01	1.22 ± 0.12	1.19 ± 0.06	5.82 ± 0.06	6.05 ± 0.40	5.98 ± 0.04
Case B	1.13 ± 0.02	1.19 ± 0.12	1.15 ± 0.06	5.75 ± 0.07	5.96 ± 0.40	5.82 ± 0.04
Case C	1.26 ± 0.04	1.26 ± 0.13	1.24 ± 0.06	6.23 ± 0.15	5.21 ± 0.42	6.14 ± 0.04
Case D	1.26 ± 0.05	1.32 ± 0.13	1.24 ± 0.06	6.24 ± 0.21	6.42 ± 0.44	6.14 ± 0.04
Case E	1.29 ± 0.06	1.28 ± 0.13	1.24 ± 0.06	6.33 ± 0.24	6.28 ± 0.42	6.14 ± 0.04

Results of these synthetic tests show that the maximum likelihood method allows a more stable estimate of the b value. Nevertheless, we preferred to use the Bender method in view of the fact that it considers the magnitude range to be limited in the upper part. This method also takes into account any slight variations in each class frequency, together with uncertainties in evaluating magnitude. With data extracted from fig. 2a-e, in fig. 3 we summarise distribution results from test cases A, C and E, using the Bender method only.

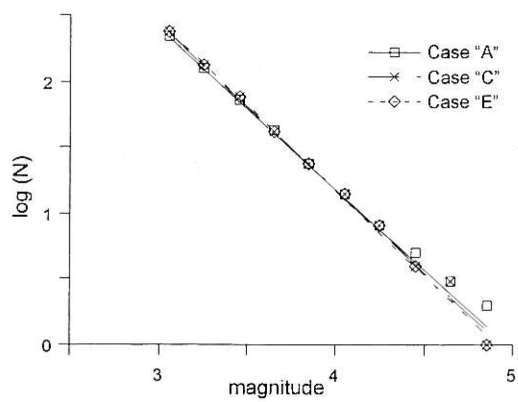


Fig. 3. Comparison of fit results computed by the BM for different synthetic tests: A, C and E cases only. Fit results in table I.

3. Data

In this paper we have analysed instrumental and historical seismicity in the Umbria-Marche region as bounded in longitude by 11°E and 14°E and in latitude by 42°N and 45°N.

We considered earthquakes shallower than 20 km, occurring between January 1987 and May 1999, as recorded by the Italian National Seismic Network (RSNC).

The data have been divided into two partially independent sets; data set I concerns 3895 events occurring between January 1987 and December 1996, data set II concerns 7885 events occurring between January 1987 and May 1999. Data set I does not include the Colfiorito seismic sequence that occurred in the autumn of 1997. Owing to the time period overlap, we attempt to evidence how the b estimate may change with the occurrence of a noteworthy seismic sequence. Figure 4 shows the studied area, together with the epicentres of seismic events occurring prior to the end of 1996 (data set I).

Since 1987, the RSNC has used the same type of instrumentation and, in the studied area, the network coverage can be considered approximately uniform. These hypotheses allow us to compare results referring to different time intervals.

The examined data were grouped into 0.2 magnitude intervals. They presented a 2.4 threshold magnitude, as shown in fig. 5 for the data set I, and a b value of little more than one, as previously shown by other authors in the same area (Console *et al.*, 1990; Mantovani *et al.*, 1990). For «magnitude» we mean the duration magnitude or, where present, the local magnitude.

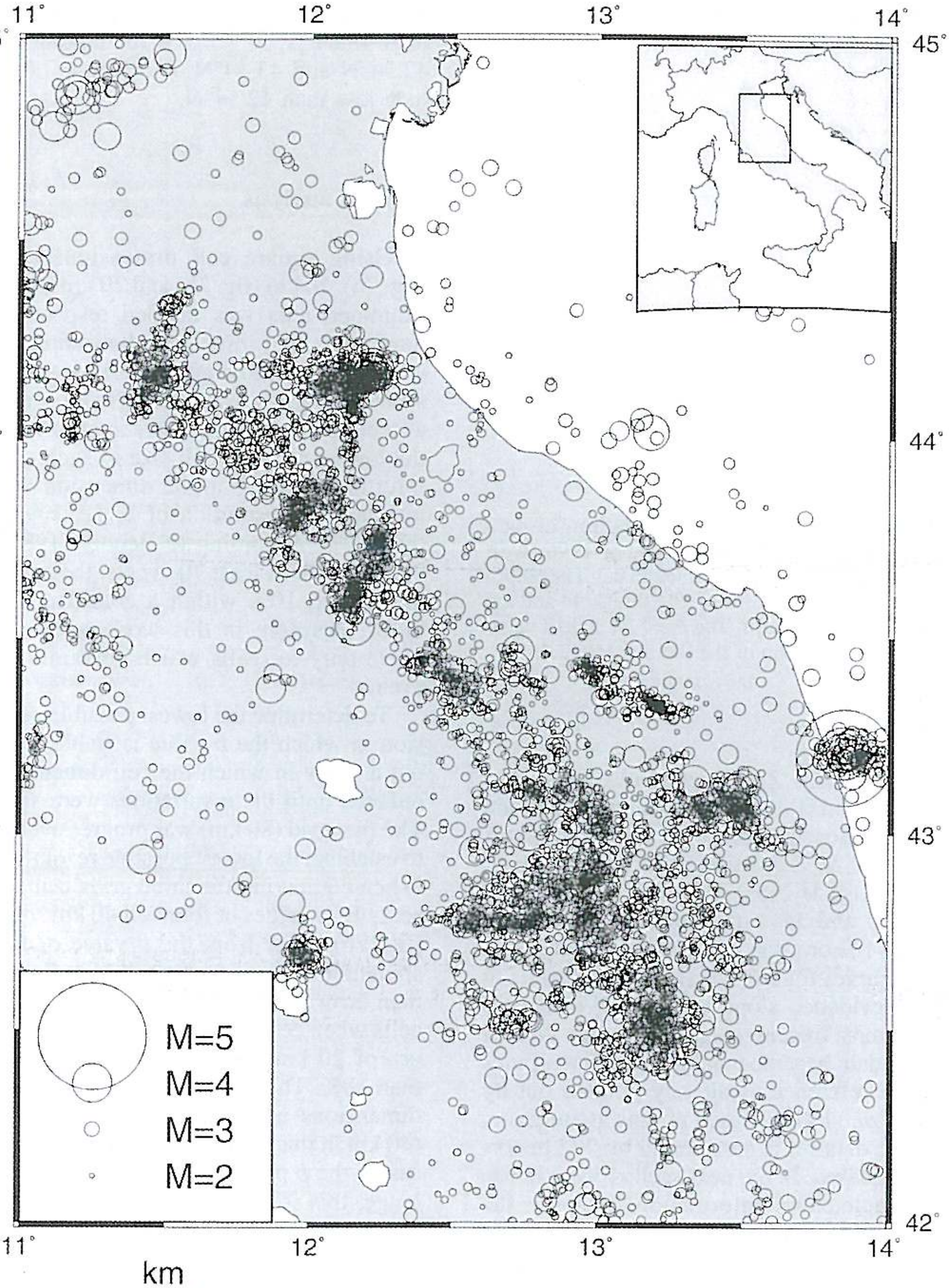


Fig. 4. Epicenters of the examined area from January 1987 to December 1996 with magnitude $M \geq 1.8$ (meridians 11° and 14°; parallels 42° and 45°.

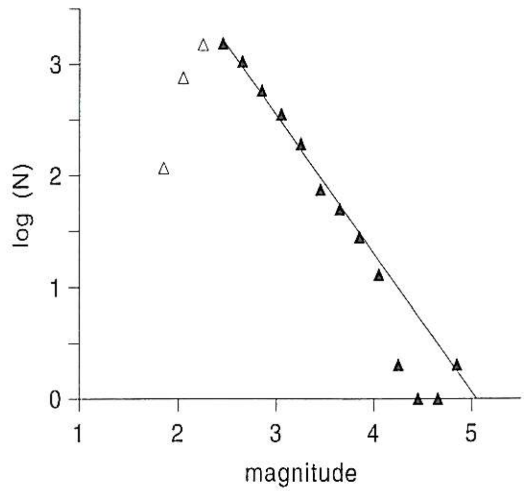


Fig. 5. Frequency-magnitude distribution of data set I, from January 1987 to December 1996, with respect to magnitude intervals of width 0.2. The empty triangles correspond to the $M < 2.4$ threshold and are not included within the fit. The $b = 1.24 \pm 0.12$ value results from application of the BM for $M \geq 2.4$.

We considered 2187 historical earthquakes reported since 1500, with equivalent magnitude of $M_e \geq 4.0$ (Boschi *et al.*, 1999). This data set is complete for $M_e \geq 4.0$ since 1925, for $M_e \geq 4.5$ since 1870, for $M_e \geq 5.0$ since 1780, for $M_e \geq 6.0$ since 1635 and $M_e \geq 6.5$ since at least 1500. Figure 6a-f shows the cumulative number *versus* time curves for some magnitude intervals in order to evidence slope changes. We assume that the most recent change in slope occurs when the data became complete for magnitudes above the reference as already pointed out by Boschi *et al.* (1997). Based on this assumption, the usable data set is constituted by 272 historical earthquakes. In the next section we will use such completeness information to apply the «mean method», as in Papazachos (1990), to estimate the reliability of instrumental data results.

In order to evaluate the b value by way of the Bender method, we divided the analysed area into three sectors and considered the maximum magnitude of each sector by Boschi *et al.* (1999). This resulted in a $m_{max} = 6.2$ for latitude greater than 43.44°N, $m_{max} = 6.8$ for latitude between 42.54°N and 43.44°N, and $m_{max} = 7.0$ for latitude less than 42.54°N.

4. Data analysis

Using square cell dimensions of 80 km (fig. 7a), 40 km (fig. 7b) and 20 km (fig. 7c), the examined area was divided respectively into three grids. All three grid dimensions selected were sufficient to consider events generated by several faults in every cell. Given a maximum acceptable error of 0.1 b, we aimed to determine the lowest possible cell size in such a way that a further decrease in the dimension would not increase the resolution of b, *i.e.* b varies less than 10%. According to Bender (1983), if the cell contains at least 70 events, the error on b is lower than 10% within a confidence level of 90%. Therefore in this work calculations are made only for cells which contain at least 70 events.

To determine the lowest possible cell dimension in which the b value is stable, we carried out a study in which the cell dimensions were reduced until the b variations were significant. The first grid (80 km) was progressively divided to establish the lowest possible resolution value. When comparing the three grids using the data set I, the b values of four cell 40 km square cells (fig. 7b) differ from the b value of the corresponding 80 km square cell (fig. 7a) by more than error. Whereas, b values of 40 km square cells when compared with corresponding b values of 20 km square cells (fig. 7c) differ less than 10%. These results indicate that if the cell dimensions are reduced below a fixed value (40 km in this case), the information required to assess the b parameter is not enhanced. On average, this is due to the number of events decreasing in proportion to the cell size, consequently giving an increased error value to the b estimate.

In order to obtain slow variations of b value, we applied a spatial smoothing to the grids with cells of 20 km and 40 km. The smoothing entails weighting data of adjacent cells by 1 for

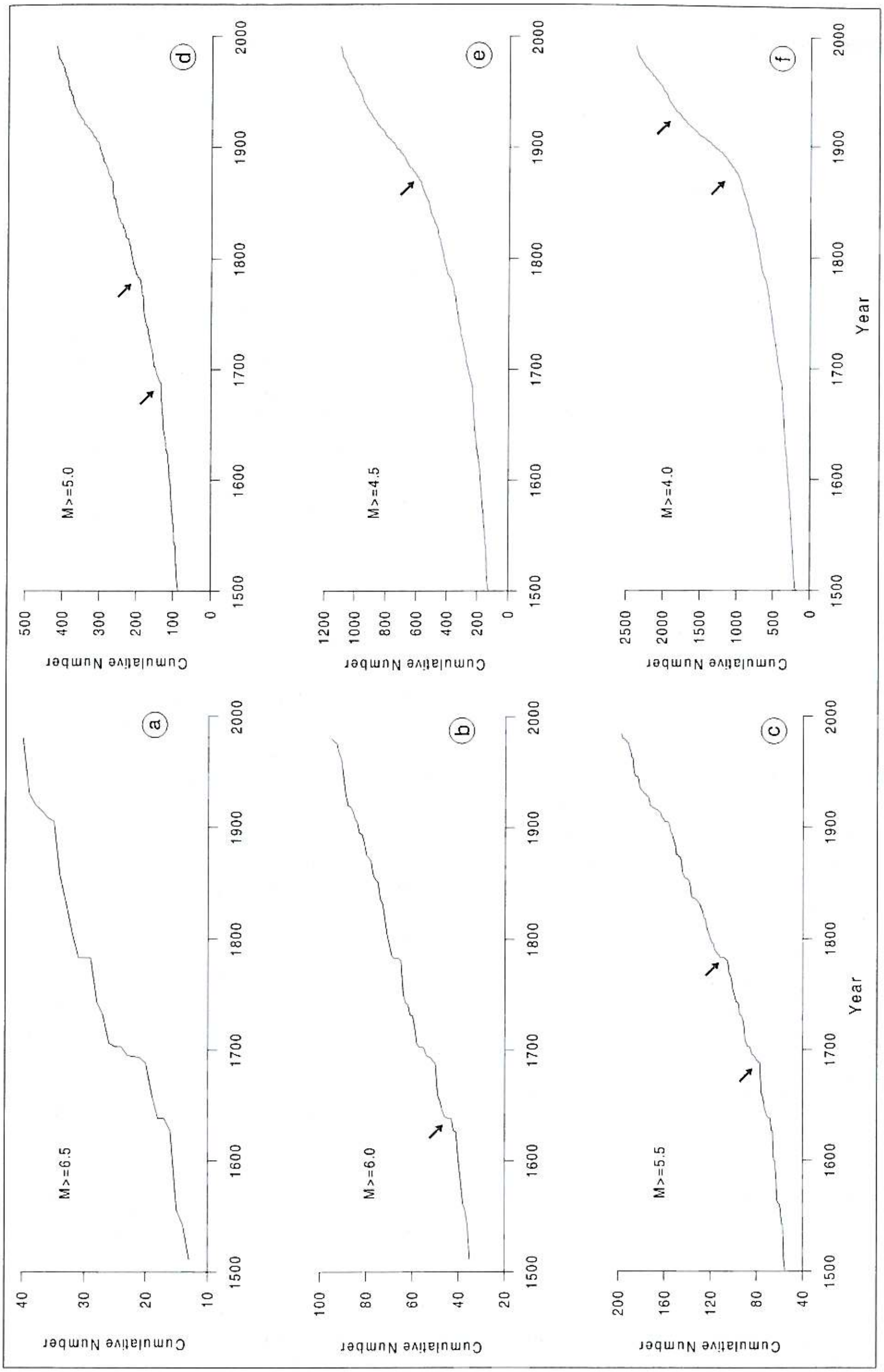


Fig. 6a-f. Cumulative number of historical earthquakes for magnitude thresholds from (a) 6.5 down to (f) 4.0 in increments of 0.5. Arrows indicate major slope changes, the most recent ones represent the completeness year for the specific class, respectively: a) 1500; b) 1635; c) 1780; d) 1780; e) 1870; f) 1925.

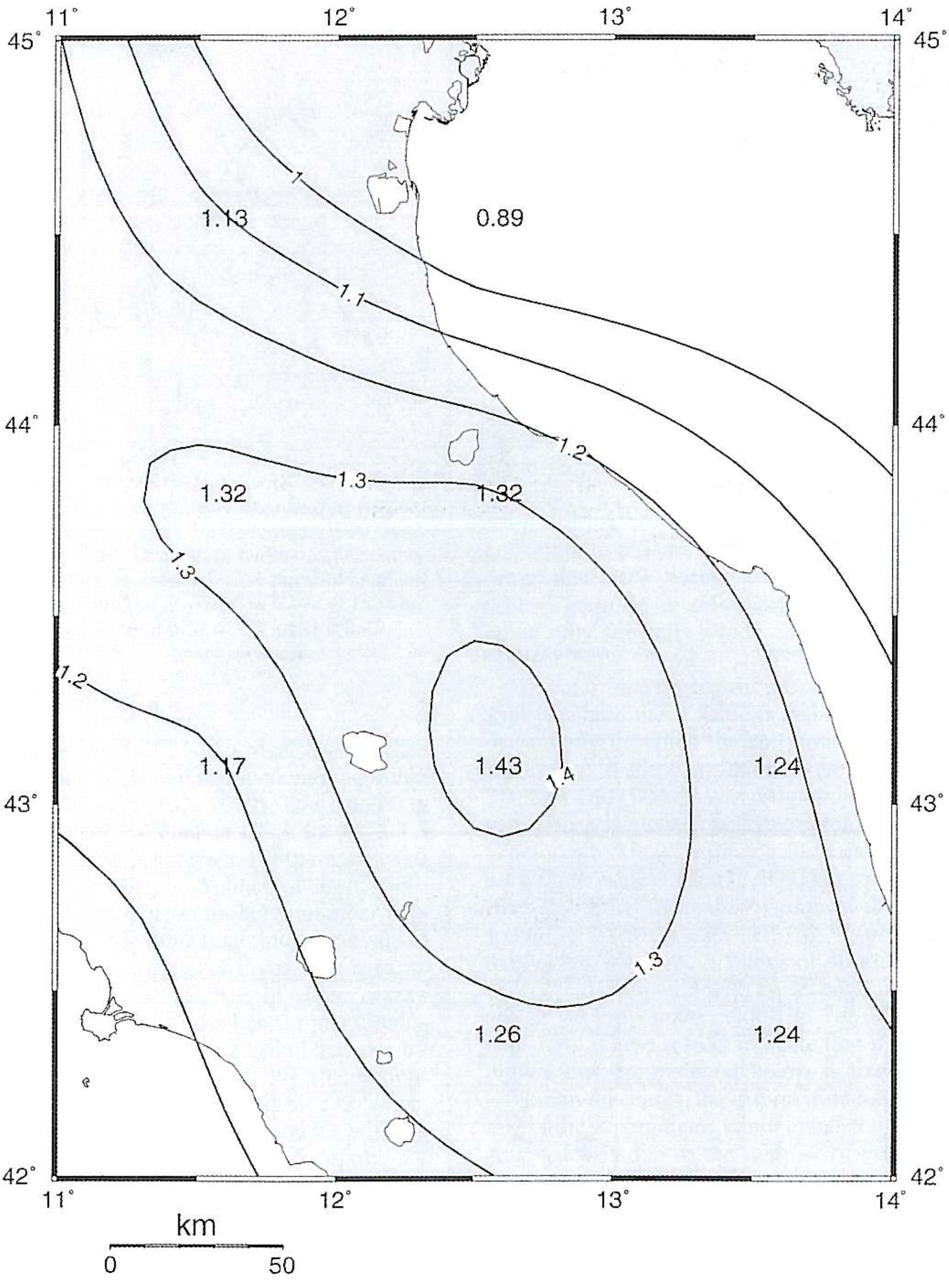


Fig. 7a. Contour lines of *b* computed by GMT (Wessel and Smith, 1991) where the event number inside every cell is ≥ 70. Grid cell dimension: 80 km.

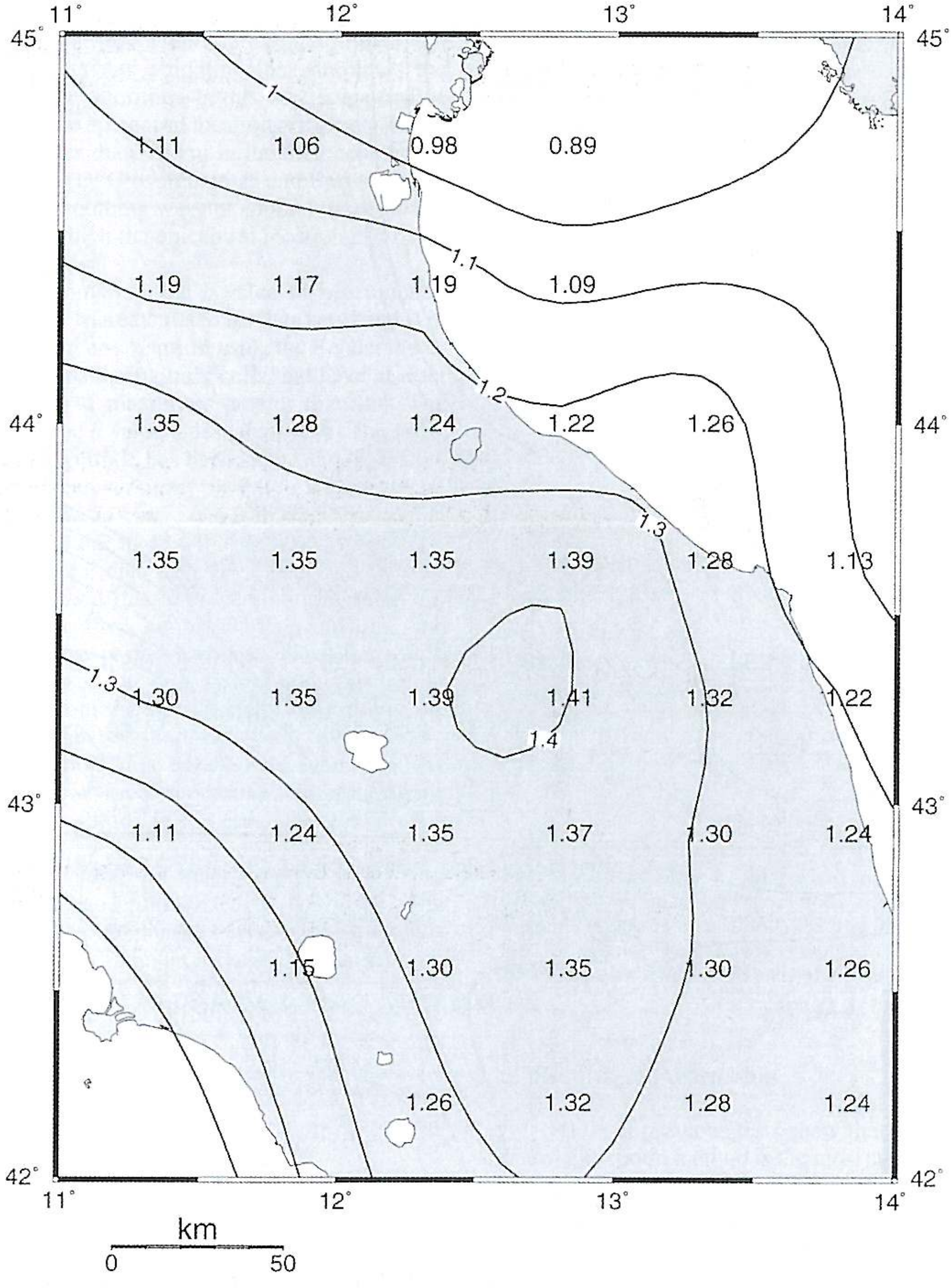


Fig. 7b. Contour lines of *b* computed by GMT (Wessel and Smith, 1991) where the event number inside every cell is ≥ 70. Grid cell dimension: 40 km.

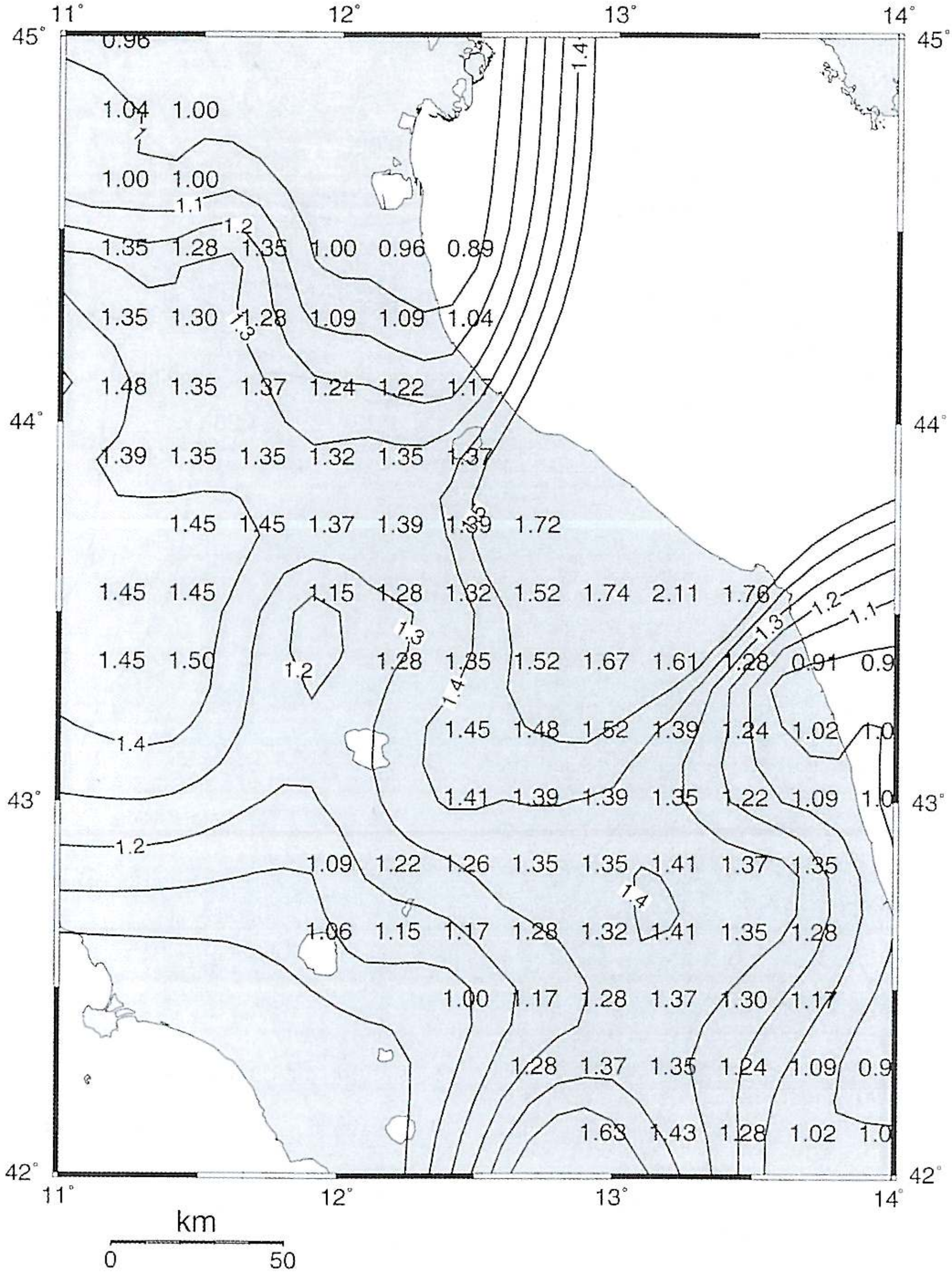


Fig. 7c. Contour lines of b computed by GMT (Wessel and Smith, 1991) where the event number inside every cell is ≥ 70. Grid cell dimension: 20 km.

contiguous and by $\sqrt{2}$ for diagonal cells. In a square grid, this is an approximation of a distance-dependent weighting that simplifies the computing algorithm. In this way, it is possible to allow for epicentral location error, considered to be lower than 10 km in the area covered by our study (Di Giovambattista and Barba, 1997). Spatial smoothing was not applied to the 80 km grid for which the epicentral location error was ignored.

To summarize, the b value of instrumental seismicity was calculated for data sets I and II in three different-size grids, using the Bender method and considering only cells that have at least 70 events of magnitude greater than 2.4. The error on the b value is less than 10%. The maximum magnitude has been suggested by historical data and set to $m_{max} = 7.0$ in latitude range 42.00° ÷ 42.54°, $m_{max} = 6.8$ in 42.54° ÷ 43.44° and $m_{max} = 6.2$ in 43.44° ÷ 45.00°.

Historical data have been analyzed by way of a slightly different process in order to obtain a representative sample of data covering the longest time period and the widest magnitude range. Based on this assumption, the usable data set from the area studied is constituted by 272 historical earthquakes.

The whole time interval for which data is available was separated into reliable data subintervals, as shown in the previous section. The number of events in each subinterval was computed for classes of magnitude $\Delta m = 0.5$ and $\Delta m = 0.2$, and normalized to 100 years. The start year, corresponding completeness magnitude and the number of events occurring in each subinterval for $\Delta m = 0.5$ (YYYY, MAG, N) are; 1500, 7.0, 1; 1500, 6.5, 3; 1635, 6.0, 5; 1780, 5.5, 20; 1775, 5.0, 39; 1870, 4.5, 95; 1925, 4.0, 109. The ending year for all subintervals is 1990. The Bender method was then applied to the normalized event numbers resulting into a $b = 0.89$ for $\Delta m = 0.5$ and a $b = 0.98$ for $\Delta m = 0.2$, both with a standard deviation of slightly less than 10%. Figure 8a,b shows the normalized event numbers along with the fit. We observe that the number of usable historical events, 272, is not sufficient to mesh the analyzed area, as performed with the instrumental data.

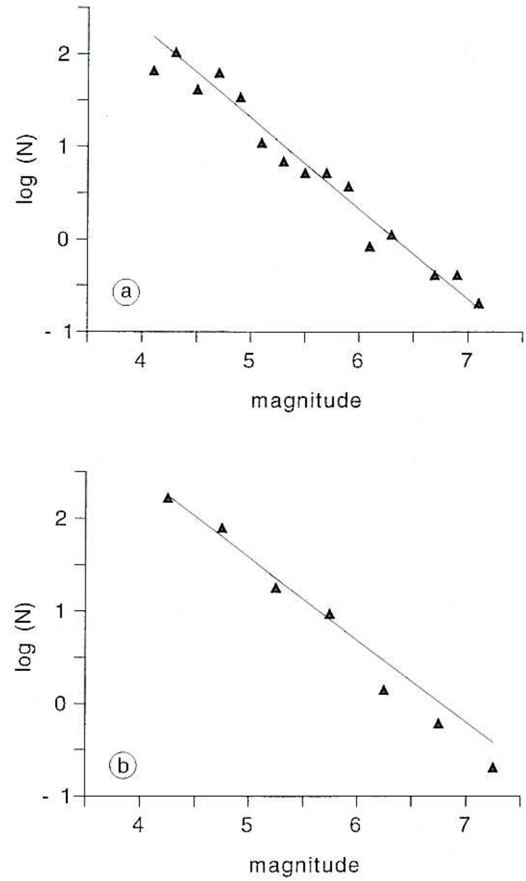


Fig. 8a,b. Number of historical earthquakes normalised within a 100 year period, with respect to magnitude classes of width (a) 0.2 and (b) 0.5. See text for details. The b values computed through the BM are (a) $b = 0.98 \pm 0.10$ and (b) $b = 0.89 \pm 0.09$.

5. Results and discussion

Results of previous tests show that the maximum likelihood method is the most stable technique to assess the b variation, particularly when a class is lacking. Furthermore, with this method the b variation from a pre-fixed value $b = 1.2$, is less than to the two other methods discussed, randomly changing the frequency of some classes. In this paper, we preferred to use the Bender method as the Gutenberg-Richter relation is upper limited and there are fluctuations in every class, either in event number, or in magnitude.

When the square cell dimensions become smaller than a fixed value (40 km in this case), the information required to assess the *b* parameter is not enhanced. This is because, on average, the number of events occurring within a smaller cell decreases, and consequently the *b* estimate is subject to greater error. Also, in using a 40 km grid, more than one fault can lie in each cell and also the epicentral error can be neglected.

Figure 7b shows the *b* values computed into the 40 km grid for data set I (January 1987 - December 1996). To evidence how *b* may change when a noteworthy seismic sequence occurs, we compared the values with those computed using the same procedure and grid size for data set II (January 1987 - May 1999), in which the Colfiorito seismic sequence is present. With the addition of the later events the data number is almost doubled. These data have the same 2.4 threshold magnitude as those of data set I. The difference in *b* values between data set I and data set II are within acceptable error margins for most of cells. However, differences are most notable between cells that lie within the Colfiorito zone (43.0°; 12.9°) and neighbouring cells. These *b* values, as shown in table II, range from *b* values 1.30 ÷ 1.41 for data set I to *b* values 1.13 ÷ 1.17 for data set II.

Table II. *b* values concerning the Colfiorito area for both data sets I and II.

Latitude (N)	Longitude (E)	Data set I (1987-1996)	Data set II (1987-1999)
42.54	12.25	1.30	1.13
42.54	12.75	1.35	1.15
42.54	13.25	1.30	1.15
42.90	12.25	1.35	1.13
42.90	12.75	1.37	1.15
42.90	13.25	1.30	1.15
43.26	12.25	1.39	1.17
43.26	12.75	1.41	1.15
43.26	13.25	1.32	1.15

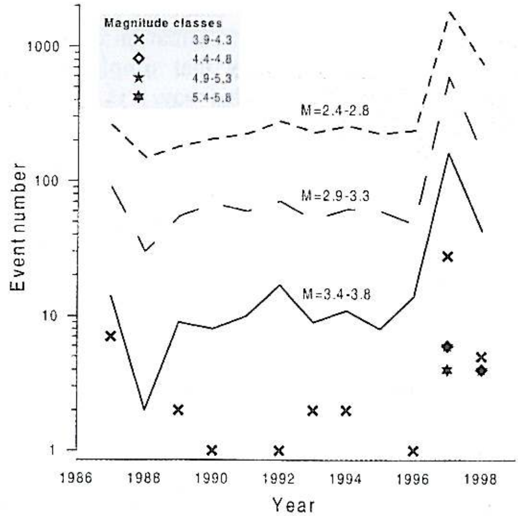


Fig. 9. Rate of instrumental earthquakes per year, from 1987 to 1998, divided into magnitude classes; small-dashed line, large-dashed line and solid-line represent magnitude classes of 2.5-2.9, 3.0-3.4 and 3.5-3.9 respectively; the symbols represent higher magnitudes (see the legend).

The *b* variations can be strongly connected to stress variations or fragmentation conditions (Wyss *et al.*, 1997). After one or more large earthquakes, which may change the stress or fragmentation condition, the *b*-value must be the same if different intervals of magnitude are examined. To verify that the decreasing *b* value for data set II is really due to the Colfiorito sequence, we have estimated the *b* value by taking into consideration the magnitude range $m = 2.4$ to $m = 4.0$. The computed *b* value for the Colfiorito zone is 1.15 ÷ 1.22, therefore the *b* value decreases with respect to data set I. This is also true if the magnitude range is limited to $m = 4.0$. Results prove that the *b* variation depends on the Colfiorito sequence due to the fact that the slope of the Gutenberg-Richter relation is the same when based on the whole of data set II, as for when limited to $m = 4.0$. These results indicate that in the presence of several segmented active faults, earthquakes occur in a larger magnitude range and at a higher rate than back-

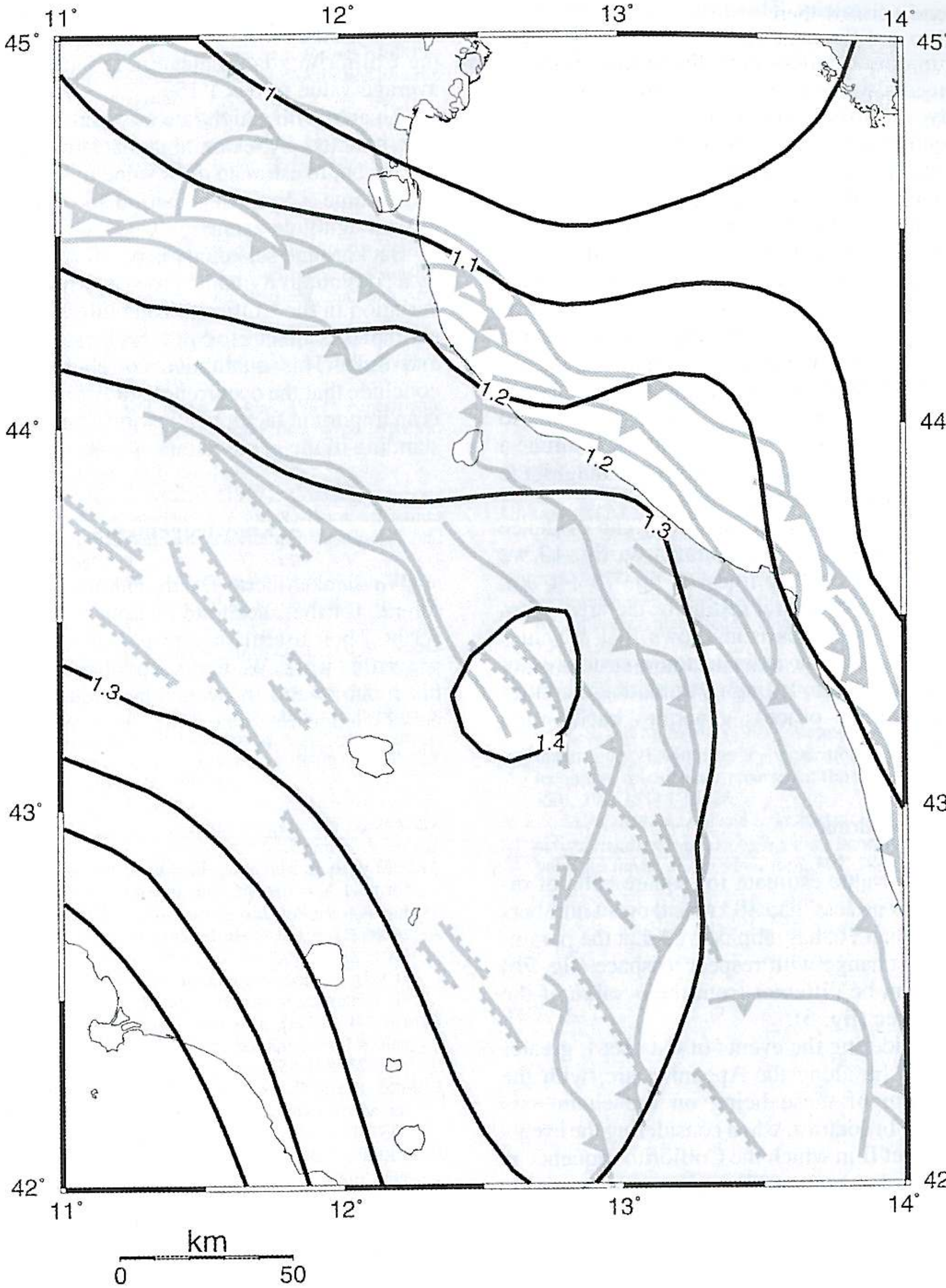


Fig. 10. Contour lines of *b* (black) for cell grid of width 40 km (same as fig. 7b) and main structural features in the area (light grey), as in fig. 1.

ground seismicity. This phenomenon seems to lead to a different slope of Gutenberg-Richter and to dominate the previous seismicity, as evidenced also by Scholz and Gupta (2000). Our study is further confirmation that b variations depend strongly on the examined time period, because the number of event varies across time for several classes. Figure 9 shows the annual rate fluctuations for different ranges of magnitude. If the time period for data extraction is too short, events of large magnitude can be absent and the instrumental seismicity can be insufficient to estimate the b value (fig. 9). Therefore, areas with a high rate of seismicity can be determinated by examining background seismicity within a short time period, say 10 years, but to estimate the b value it is necessary to examine a longer time period, to take large-magnitude earthquakes into account.

In order to verify if seismicity can evidence different seismogenetic settings, in fig. 10 we overlaid the contour lines of fig. 7b (40 km cells) on the structural setting of the area (from fig. 1). This comparison shows that b values closely follow the main tectonic structures of the Apennine arc, letting us think that a continuous dynamic process generates background seismicity.

6. Conclusions

The b value estimate for square cells of dimensions no less than 40 km and event numbers greater than 70, has emphasized that the parameter can change with respect to space (fig. 7b) and it can be different from the b value of the whole area (fig. 5).

Considering the events of data set I, greater b values lie along the Apennine arc, with the maximum of these being on the chain axis (fig. 10). In contrast, when considering the events of data set II in which the Colfiorito sequence is present (Autumn 1997), cells with longitude between 12.0° and 13.5° and with latitude near 43° have greatly decreased b values (table II). A decrease is also notable if we consider a limited magnitude range within data set II. These different b values (between data set I and data set II) infer that seismicity within the first set should be considered background seismicity. On the contrary, data set II b values within and near to the Colfiorito zone indicate a reduction in the average value to $b = 1.15$.

An area with a high rate of seismicity can be determinated by examining background seismicity, but to estimate the b value it is necessary to examine a long time period to account for large-magnitude events.

Background seismicity is possibly generated by a continuous dynamic process, whereas the b variation in the Colfiorito zone infers that large earthquakes are due to a process located in space and time. This qualitative correlation lets us conclude that the occurrence study of seismicity is an important factor in allowing a better understanding of the geodynamic aspects of a region.

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