ANNALS OF GEOPHYSICS, VOL. 46, N. 6, December 2003

A microseismic study in a low seismicity area of Italy: the Città di Castello 2000-2001 experiment

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Abstract
Recent seismological studies contribute to better understand the first order characteristics of earthquake occurrence in Italy, identifying the potential sites for moderate to large size earthquakes. *Ad hoc* passive seismic experiments performed in these areas provide information to focus on the location and geometry of the active faults more closely. This information is relevant for assessing seismic hazard and for accurately constraining possible ground shaking scenarios. The area around the Città di Castello Basin, in the Northern Apennines (Central Italy), is characterized by the absence of instrumental seismicity ($M > 2.5$), it is adjacent to faults ruptured by recent and historical earthquakes. To better understand the tectonics of the area, we installed a dense network of seismic stations equipped with broadband and short period seismometers collecting data continuously for 8 months (October 2000-May 2001). The processing of ~ 900 Gbyte of data revealed a consistent background seismicity consisting of very low magnitude earthquakes ($M_L < 3.2$). Preliminary locations of about 2200 local earthquakes show that the area can be divided into two regions with different seismic behaviour: an area to the NW, in between Sansepolcro and Città di Castello, where seismicity is not present. An area toward the SE, in between Città di Castello, Umbertide and Gubbio, where we detected a high microseismicity activity. These findings suggest a probable different mechanical behaviour of the two regions. In the latter area, the seismicity is confined between 0 and 8 km of depth revealing a rather well defined east-dipping, low angle fault 35 km wide that cuts through the entire upper crust down to 12-15 km depth. Beside an apparent structural complexity, fault plane solutions of background seismicity reveal a homogeneous pattern of deformation with a clear NE-SW extension.

Key words *microseismicity – low-angle normal fault – seismic gap – seismic hazard*

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1. Introduction

The Northern Apennines (Italy) undergo active crustal extension of about 0.5×10^{-7} yr^{-1} (Hunstad *et al.*, 2003) accommodated by moderate to large normal faulting earthquakes along a NW-elongated seismic belt (fig. 1).

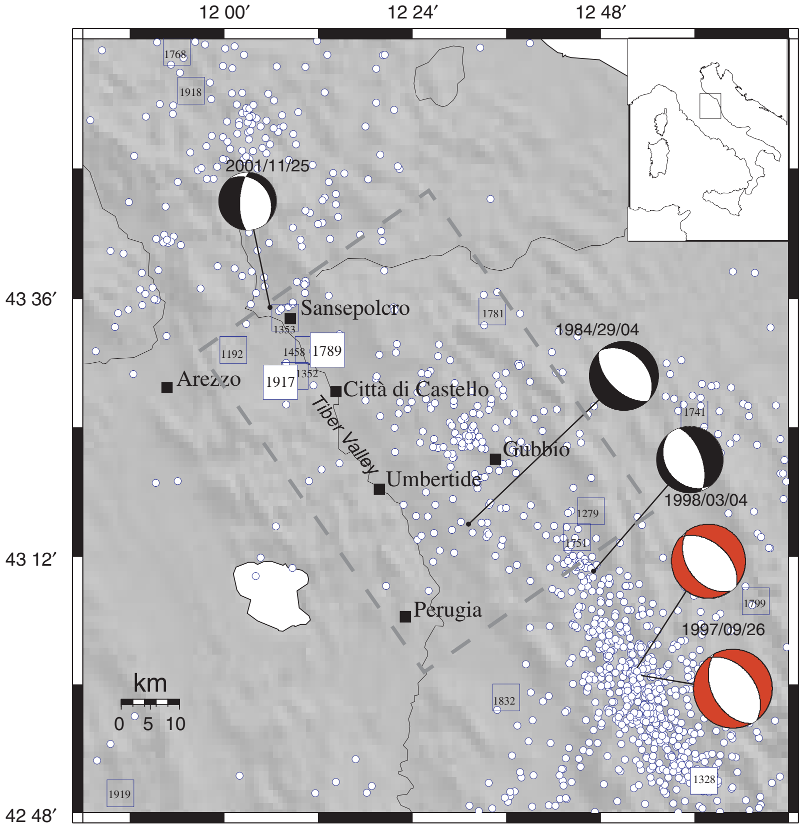

Fig. 1. Historical earthquakes (empty squares, from CFTI-CPTI) and instrumental seismicity ($M_L > 2.5$) of the area (from earthquake bulletin 1990-1999): in black the focal mechanisms of Gubbio (1984), Gualdo Tadino (1998) and Sansepolcro (2001); in red the mainshock of Colfiorito seismic sequence (1997). The dashed box enclose the studied area.

In the past 30 years, three seismic sequences following M_w 5.0-6.0 mainshocks, struck a 70 km long section of the Northern Apennines from Norcia (1979, $M_S = 5.8$) to Gubbio (1984, $M_S = 5.3$). For all these sequences, seismological data show that earthquakes occur on NW-trending normal faults the lengths of which are usually less than 10 km and dipping towards the southwest (Deschamps *et al.*, 1984; Haessler *et al.*, 1988; Westaway *et al.*, 1989). During the most recent sequence that struck the region in 1997 and 1998, seven mainshocks ($5.0 < M_W < 6.0$, Ekström *et al.*, 1998) ruptured adjacent SW-dipping fault segments composing a 50 km long portion of the seismic belt (Chiaraluce *et al.*, 2003). At the northwestern end of this «seismically active» belt lies a «silent» region located near the town of Città di Castello (hereafter CDC). The southern portion of the study area had no significant earthquakes for the past 1.0 kyr (see the historical catalogue of Boschi *et al.*, 1997) and an impressive lack of instrumentally recorded seismicity $M_L > 2.5$ earthquakes, which represent the lower threshold magnitude for the located earthquakes detected by the Italian network (RSNC).

To investigate the CDC seismically silent region, we acquired digital, mostly broadband, data with a seismic array consisting of 30 three-component stations. The experiment lasted for 8 months.

The experiment aims at defining possible fault location, geometry and kinematics by using the recorded microearthquakes. In this study, we present the recorded data with preliminary hypocentral determinations and focal mechanisms of the recorded seismicity.

2. The CDC experiments

Between October 2000 and May 2001, a dense temporary local network of 30 digital three-component stations was installed in an area extending $70 \times 40 = 2800$ km^2, from Sansepolcro (in the north) to 20 km south of Gubbio (fig. 2). We installed Reftek, Nanometrics and Lennartz digital stations equipped with three-component sensors (Lennartz LE3d-5s, Guralp CMG40T, Lennartz LE3d-1s, Mark L22).

The geometry of the network varied during the experiment in order to fulfil the aims of the project.

2.1. *The test*

In the first week of October 2000, a data acquisition test was performed to provide a first look at the ongoing local seismicity. This initial part of the

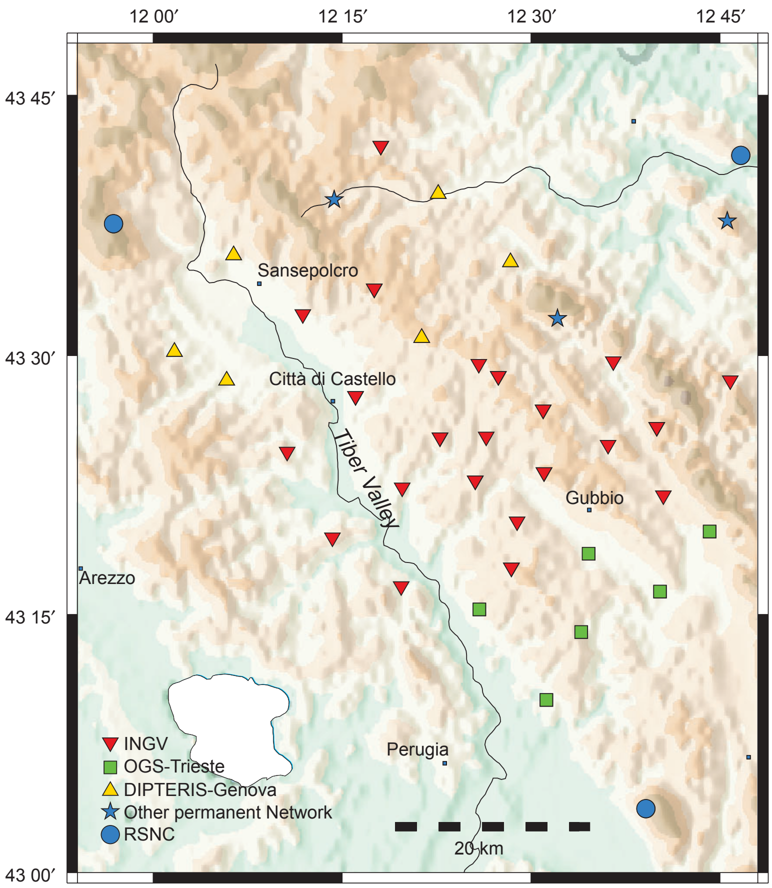

Fig. 2. Map of the temporary seismic network operating in the area.

experiment was also helpful to optimize the network geometry. The test network consists of 10 stations positioned around the nodes of a regular grid spaced at about 10 km. The stations continuously telemetered the data to a seismic acquisition center located in a mobile van. After optimizing the trigger and the coincidence parameters of the acquisition system, it was possible to monitor the seismicity of $M_L \leq 1$ in real-time. A total of 181 events (143 both local and regional, and 38 teleseisms) were recorded during the month-long experiment.

2.2. *The main experiment*

As the test confirmed the presence of a significant level of background seismicity, we distributed the remaining 20 stations around the 10 stations of the test network (see fig. 2). From the beginning of November 2000, the network was composed of «stand alone stations» only recording continuously. The radio transmitters were replaced by systems recording on hard disks (Lennartz Digital Uher). At a sampling rate of 125 sps per channel, the memory capacity of the stations lasted for more than 1 month.

The instruments were powered by high capacity batteries (200 Ah) and solar panels. Stations were located along secondary country roads, mainly in woodland, far away from any anthropic noise. Sites characterized by solid rock outcrop, good insulation for solar panels and visible sky for GPS were favoured. The seismometers were buried, when possible, and covered with a heavy cap to avoid wind and rain disturbances. The stations were visited once a month for data collection and servicing.

After processing the data of the first 4 months, a cluster of shallow (from 2 to 4 km depth) hypocenters was identified roughly 10 km to the north of Gubbio. Thus, we decided to move some of the stations from the peripheries of the network to the Gubbio area to improve the network geometry around the hypocentral cluster.

2.3. *Event detection*

In order to extract seismic events from the continuous raw data, we developed a detection algorithm, denominated DETECTOR hereafter. DETECTOR is based on the same principles as the automatic detection systems used by centralized networks. We applied an *sta/lta* (short time average/long time average) trigger algorithm to the continuous data. The values for *sta*, *lta*, ratio threshold and other trigger parameters were optimized by performing numerous test runs: *sta* = 2 s, *lta* = 40 s, *threshold* = 3, *duration* = 2 s, *coincidence interval* = 40 s, *coincidence level* = 5. This resulted in an earthquake detection level less than $M_L \approx 1$ and minimum number of false events.

The detection procedure was composed as follows: firstly we selected intervals with *sta/lta* (*i.e.* short *versus* long time averages) greater than a chosen threshold on each station data. Then a trigger coincidence of five stations was chosen to declare an event. Event time windows were extracted from the continuous data of all 30 stations based upon the selected triggers. We also find the best combination for the time length of *sta/lta* required to be greater of the threshold, the time interval during which triggers have to be considered coincident and finally the number of coincident triggers necessary to define a possible seismic event.

3. Data analysis

Despite the low magnitude of the events, a high signal/noise ratio allowed us to read *P*-wave and *S*-wave onset on digital waveforms with the accuracy of few samples for the *P* onset and less than 0.1 s for *S*-wave readings. The phase readings has been weighted according to reading accuracy and the following «01234»-HYPO71 (Lee and Lahr, 1975) scheme was adopted: 0 ($\leq$ 0.03 s), 1 (0.03-0.06 s), 2 (0.06-0.1 s), 3 ($\geq$ 0.1 s), and 4 (ignore).

A total of 2200 earthquakes were located by using HYPOELLIPSE (Lahr, 1989) and a gradient 1D velocity model obtained by inverting 1994 earthquakes for a total of 14245 *P*-phases and 14105 *S*-phases by using VELEST (Kradolfer, 1989; Kissling *et al.*, 1994) code.

The starting velocity model was derived from studies performed on adjacent region (Deschamps *et al.*, 1984; Chiaraluce *et al.*, 2003). The best 1D velocity model was reached after 4 iterations, with a final unweighted RMS of 0.04. The *P*-wave velocity values found are consistent with those derived by logs and laboratory experiment (Bally *et al.*, 1986).

Inversion results (fig. 3) show the presence of a well constrained, low velocity layer (5.7 km/s) at a depth of 20 km. Although the distribution of seismicity is mainly located in the first 15 km in depth, the crustal volume affected by the low velocity layer is sampled by a good number of seismic rays (> 500 rays and about 20 hypocenters). The deepest earthquakes recorded were localized around 60 km at depth. The V_p / V_s ratio computed by VELEST is also well constrained around the value of 1.84. Similar values were found in adjacent regions (Haessler *et al.*, 1988; Ripepe *et al.*, 2000). Finally, a total of 2215 events were located by using HYPOELLIPSE (Lahr, 1989).

To determine M_L, synthetic Wood-Anderson seismograms were calculated by removing the instruments response of each recording and convolving the displacement ground motion with the standard Wood-Anderson torsion seismograph response (Anderson and Wood 1925; Richter, 1935). Observed NS and EW amplitude values were averaged to constitute a single measurement (*i.e. sqrt* $(ns^2 + ew^2)$).

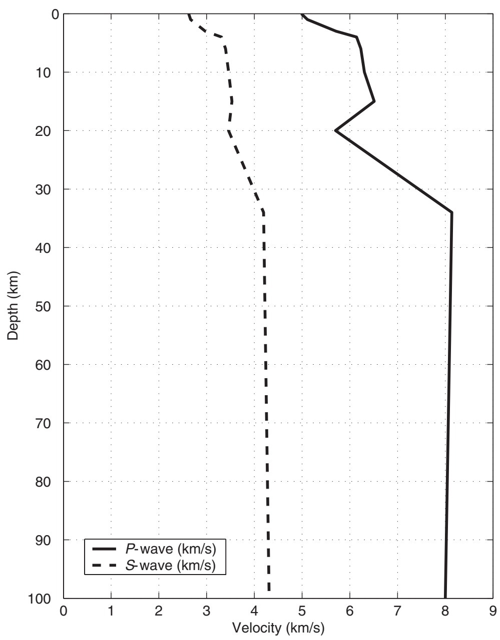

Fig. 3. Best 1D velocity model for the study area.

Following the method originally used by Richter (1935,1958), M_L is given by

$$M_L = \text{Log}A - \text{Log}A_0 - S \qquad (3.1)$$

where A is the maximum trace amplitude in millimetres measured from a Wood-Anderson seismogram, A_0 is a distance dependent attenuation curve, and S is a station dependent M_L adjustment.

We assume a parametric form of Log A_0 proposed by Alsaker *et al*. (1991) and derived by Bakun and Joyner (1984) as

$$\begin{aligned}\text{Log}A_0(R) = n\text{Log}\left(R / R_{\text{ref}}\right) + \\ -k\,(R - R_{\text{ref}}) - V\,(R_{\text{ref}})\end{aligned} \qquad (3.2)$$

where R is the hypocentral distance (in km), n and k are parameters related to geometrical spreading and attenuation of *S*-waves. R_{ref} is the reference distance and $V(R_{\text{ref}})$ is the reference magnitude value corresponding to R_{ref}.

For instance, $R_{\text{ref}} = 100$ km corresponds to $V(R_{\text{ref}}) = 3$ whereas $R_{\text{ref}} = 17$ km corresponds to $V(R_{\text{ref}}) \sim 2$. To establish the new magnitude scale, we combined eqs. (3.1) and (3.2) to give an over determined sparse system of equations. Under the constraints that station corrections sum to zero, the inversion of the linear system was performed by a least-squares procedure based on LSQR algorithm (Paige and Saunders, 1982). Bootstrap resampling techniques (*e.g.*, Efrom, 1979) were used to estimate the inversion stability; station corrections, attenuation parameters and magnitude values are computed by averaging the results of 5000 repeated inversions on randomly resampled data sets. Bootstrap results are also used to estimate the standard deviation of the inverted parameters.

Inversion was performed normalizing the local magnitude scale to motions at Ref = 17 km ($V_{\text{ref}} \sim 2.0$). We obtain $n = (1.420 \pm 0.03)$, $k =$ $= (0.009 \pm 0.0009)$ and station corrections ranging between 0.46 and -0.30.

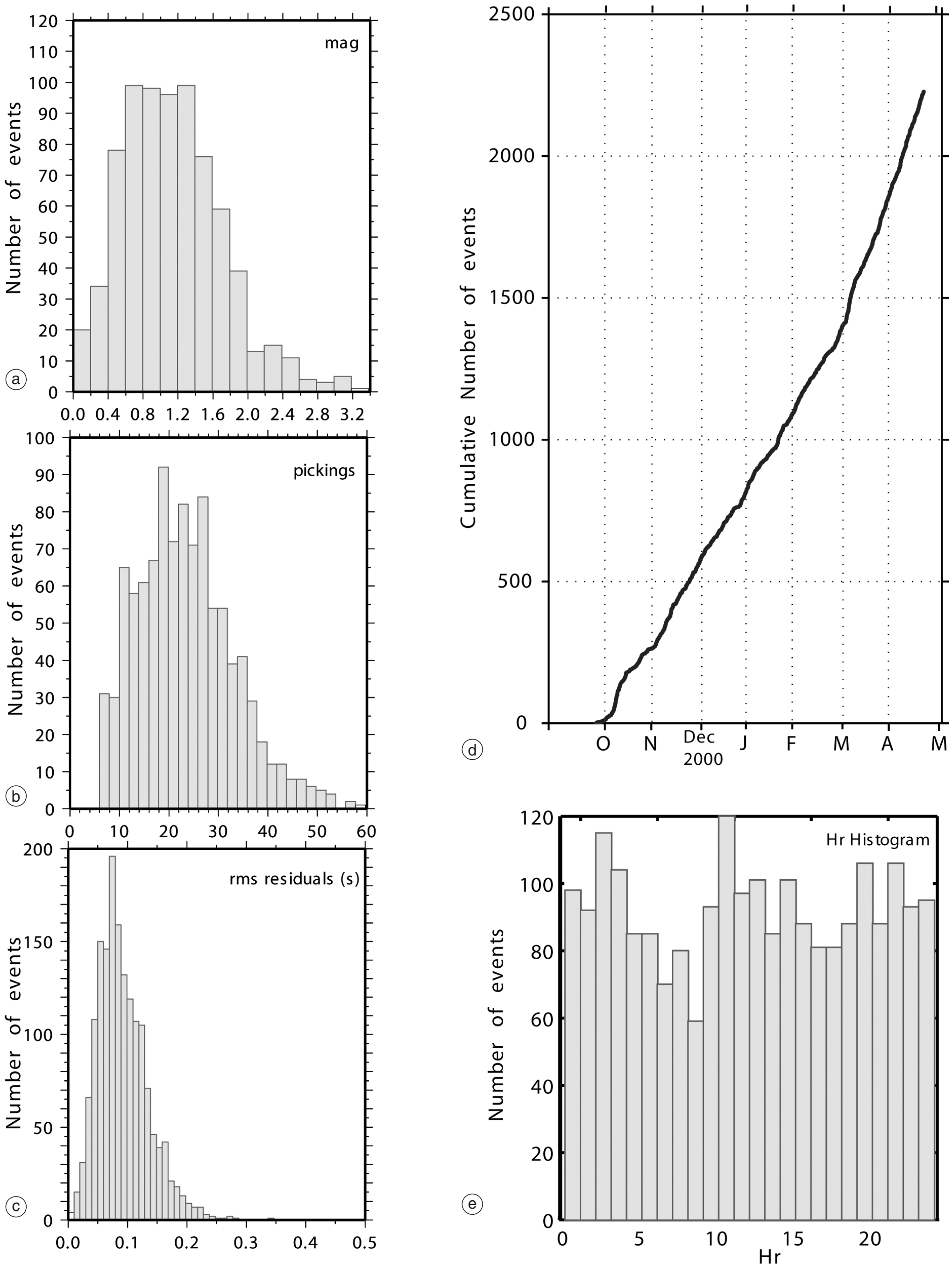

Fig. 4a-d. Statistical plots for the localized events.

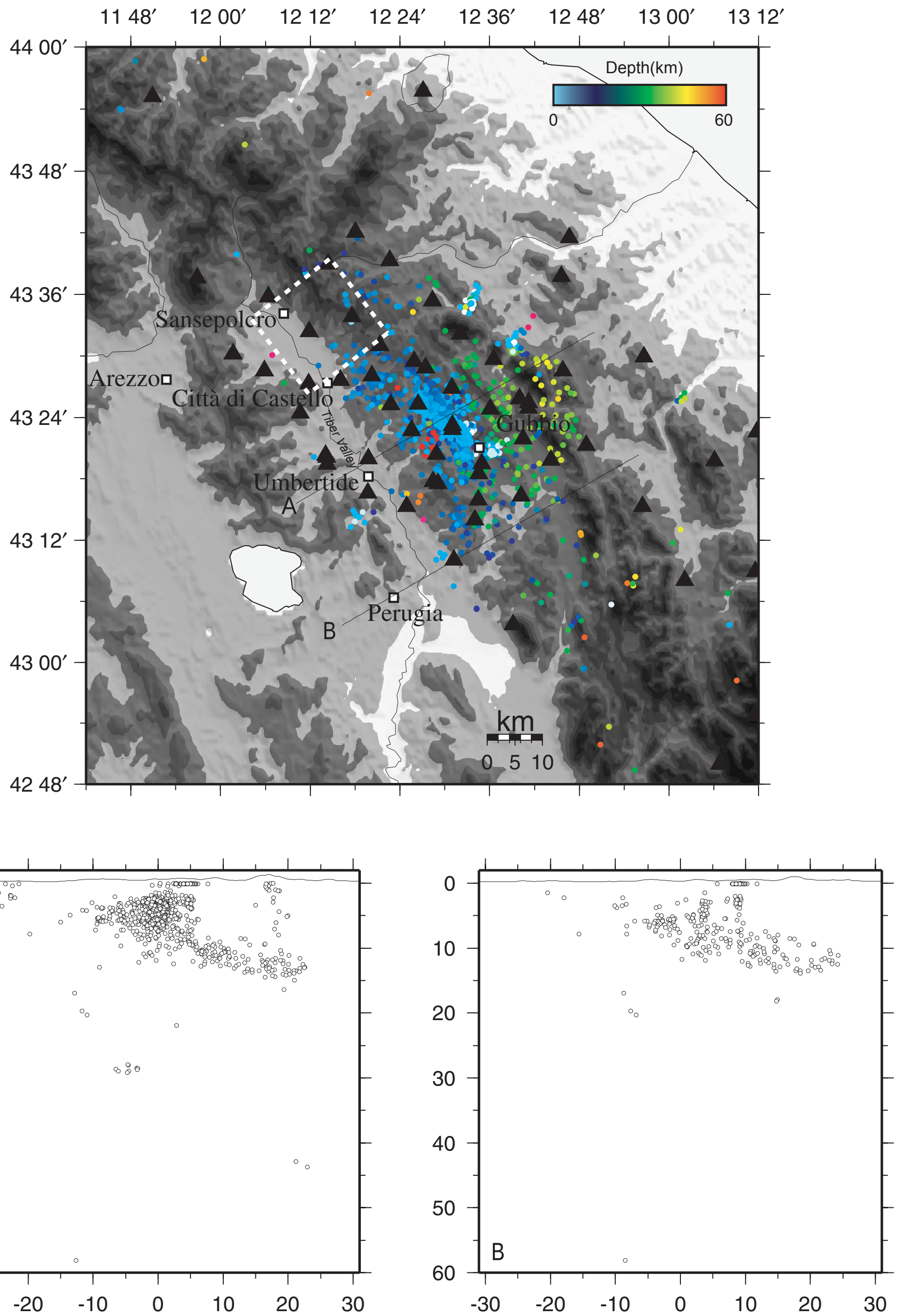

Fig. 5. Map and two vertical sections of located earthquakes. The dashed box enclose the hypothesized seismic gap discussed in the text. White dots represent quarry blasts localized at very shallow depth.

Finally some statistical analyses were performed on the temporal distribution of seismicity. The main results can be summarized as follows:

i) The distribution of the local magnitude for the recorded events is complete down to $M = 1.2$, while ranging between 0 and 3 (fig. 4a).

ii) The number of phases per events spans from a minimum of 8 phases (about 30 events) to 60 phases per event (including *P*- and *S*-wave readings from the INGV-National Seismic Network), indicating the good design of the acquisition experiment, the choice of low-noise sites and minimum failure of the instrumentation (fig. 4b).

iii) The RMS mean value around 0.08 s indicates an overall high quality of phase data and of the velocity model adopted (fig. 4c). This is also shown by the final locations depicted on the plan view map and along the vertical sections. The number of events *versus* hour of the day is almost constant. The cumulative number of events (fig. 4d) shows a constant increase between October 2000 and February 2001 while between March and April 2001 this trend shows an evident tilt-up (fig. 4d,e), corresponding to the period when the network was densified around the main clusters of seismicity. We also used blasts produced by some quarries to test the goodness of earthquake location (A in fig. 5).

For the best located events, we computed focal mechanisms by using those events which has at least 12 *P*-wave polarities (fig. 6). Here we present the solutions for 150 events computed by using FPFIT code (Reasemberg and Oppenheimer, 1985) which have quality factors equal to A.

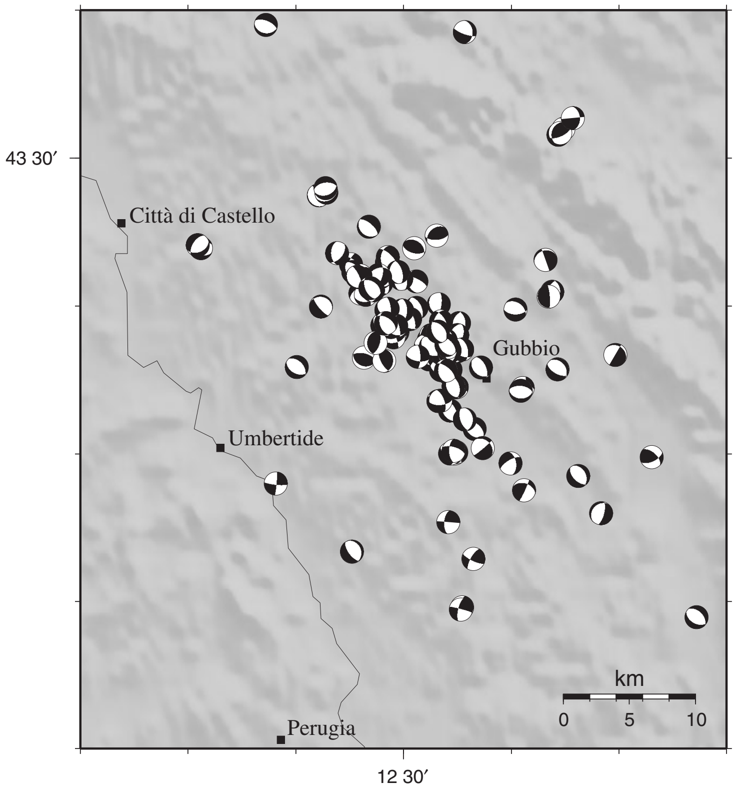

Fig. 6. 150 focal solutions for the best located earthquakes.

4. Data interpretation

Figure 5 shows the epicentral map of the located earthquakes with horizontal and vertical errors less than 1 km and an azimuthal gap less than 240° (80% of the seismicity has a gap less than 180°). Earthquakes concentrate on a NW-trending structure elongated for 20-30 km to the north of Gubbio and toward the south. Background seismicity in the area close to CDC town is scarse, leaving the hypothesized seismic gap almost silent (A in fig. 5). This behaviour is neither referable to the network geometry nor to the detection method which was applied to the data using the same trigger parameters. This is granted by the presence of some microearthquakes in the north-western part of the network (see A in fig. 5) which were localized also using data from the RSNC network.

From the vertical sections across the fault system (fig. 5), we observe an east-dipping cloud of hypocenters that, from around 2 km depth beneath the western side of the Tiber Valley, reach about 12-15 km of depth to the east, cutting the upper crust beneath the Apenninic belt. We interpret this seismicity to occur on a gently east-dipping fault, possibly related to the Alto Tiberina Fault previously recognized comparing available seismological data with seismic reflection profiles and geological investigation (Barchi *et al.*, 1998; Boncio *et al.*, 2002; Collettini *et al.*, 2002). In the central part of the study area, the cluster of hypocenters located to the north of Gubbio is confined at depth by the east-dipping fault intersecting the west-dipping seismicity at a depth of about 6 km. A few earthquakes occur at mid-crustal depth and down to 50 km in the uppermost mantle. The occurrence of this sub-crustal seismicity could be related to the flexured subducting Adriatic lithosphere as previously observed by Selvaggi and Amato (1992).

The focal solutions for a very high number of events (> 80%) show a general NE-trending direction of extension, with NW-trending both SW and NE dipping normal fault planes, consistent with the regional stress observed by large earthquakes and borehole breakouts (Mariucci *et al.*, 1999; Chiaraluce *et al.*, 2003). Some inverse solutions are found for the deepest earthquakes.

5. Conclusions

The microseismicity distribution, revealed by the high detection capability of the local array, let us define two different zones: one in the SE, where low magnitude background seismicity is present and well organized, and the other to the NW characterized by the lack of microseismicity. Consistently with larger magnitude events, background seismicity occurs on small NW-trending normal fault segments governed by the regional NE trending extension.

In the southern and central portions (from Gubbio to CDC), the microseismicity defines a seismic volume elongated for 50 km and from 0 to 15 km in depth, confined on the hanging-wall of the low angle east-dipping normal fault (Alto Tiberina Fault). This latter fault seems to play a dominant role in controlling the deformation style of the area, cutting the upper crust almost completely.

In the northern part, background seismicity is nearly absent defining a ≈ 20 km long area northward limited by an area (Sansepolcro) where three small seismic sequences ($M_L < 4.5$) occurred in the past 15 years (Braun *et al.*, 2003). These sequences are referable to the northern segment of the ATF fault, or to the antithetic SW dipping normal fault. In the silent area, no earthquakes occur on the east-dipping low angle normal fault hypothesized by active seismic data. The presence of historical earthquakes (1458, MCS = IX; 1789, MCS = VIII-IX; 1917, MCS = IX) suggests that this portion of the fault could be presently locked, supporting the hypothesis of a seismic gap.

Acknowledgements

This experiment was partially founded by Gruppo Nazionale per la Difesa dai Terremoti within the project: «Development and Comparison among Methodologies for the Evaluation of Seismic Hazard in Seismogenic Areas: Application to the Central and Southern Apennines», coordinated by M. Cocco.

We thank all the people who contributed to the field work and in particular the technical staff of the INGV Mobile Network Laboratory.

Thanks to M. Ibs-von Seht for the discussion useful to the improvement of the manuscript, and to Salvatore Barba and Massimiliano Rinaldo Barchi for the useful comments in reviewing the manuscript.

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(received June 16, 2003;
accepted October 20, 2003)