From earthquake intensities to earthquake sources: extending the contribution of historical seismology to seismotectonic studies

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Abstract
The epicentral locations and magnitudes of the events reported in the *Catalogue of Strong Italian Earthquakes* are obtained from intensity data through a standardized and established algorithm. However, we contend that the dense and homogeneously collected data sets presented in this catalogue can also be used to assess the location, physical dimensions and orientation of the earthquake source on purely historical grounds. The method we describe is of special value for older earthquakes and for all events that fall in areas where the understanding of faulting and tectonics is limited. At the end of the calculations the seismic source is represented as an oriented «rectangle», the length and width of which are obtained from moment magnitude through empirical relationships. This rectangle is meant to represent the actual surface projection of the seismogenic fault or, at least, the projection of the portion of the Earth crust where a given seismic source is likely to be located. Sources derived through this procedure can then be juxtaposed to sources derived from instrumental and geological data for constructing fault segmentation and earthquake recurrence models and for highlighting linear gaps in the global seismic release. To test the method we applied it systematically to all $M > 5.5$ earthquakes that occurred in the Central and Southern Apennines in the past four centuries. The results are encouraging and compare well with existing instrumental, direct geological and geodynamic evidence. The method is quite stable for different choices of the algorithm parameters and provides elongation directions which in most cases can be shown to be statistically significant. The resulting pattern of source locations and orientations is homogeneous, showing a consistent Apennines-parallel trend that agrees well with the NE-SW tectonic extension style of the central and southern portions of the Italian peninsula.

Key words *historical earthquakes – source orientation – seismotectonics – seismic intensity*

1. Introduction

Assigning the most severe historical earthquakes to specific individual seismogenic sources is one of the most important aspects of the modern practice of seismic hazard assessment. This effort forms the basis for supplying more faithful predictions of ground shaking than those obtained from conventional seismotectonic analyses, both in the near-field and for the more populated areas away from the main tectonic belts. Defining a set of homogeneous seismogenic sources, however, is not an easy assignment. As Reiter (1991) puts it «*Defining and understanding seismotectonic sources is often the major part of a seismic hazard analysis and requires knowledge of the regional and local geology, seismicity and tectonics*». Perhaps due to the difficulties in identifying indi-

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vidual large sources, either by recourse to geological data alone or through a combination of historical and geological evidence, most national seismic hazard plans still rely almost exclusively on catalogues of historical seismicity with very little support from field and instrumental data (*e.g.*, Muir-Wood, 1993). This condition is further stressed by the circumstance that in many countries the amount of knowledge available on historical seismicity is often considered sufficient to supply a satisfactory representation of the earthquake potential. Historical seismicity is commonly spread over wide tectonic regions simply to derive rates of occurrence for different classes of ground acceleration, and any direct reference to a specific fault or to the physics of the earthquake phenomenon is invariably lost.

We certainly agree that historical catalogues supply a reasonable first-cut representation of regional seismicity, but we want to demonstrate that the information contained in homogeneously collected intensity observations such as the *Catalogue of Strong Italian Earthquakes* can be used to quantify the essential parameters of the seismic source. In other words, intensity observations may provide a wider and more valuable contribution to the assessment of seismic hazard than is commonly assumed. We propose a strategy through which seismogenic sources documented solely by historical information can be described by the same set of physical parameters normally used to describe sources for which modern instrumental observations are available, though with a lesser degree of confidence.

Our approach is similar to that taken by Johnston (1996) and Bakun and Wentworth (1997) for quantifying the seismic moment of some large pre-instrumental earthquakes of stable continental regions and California, respectively, but somewhat more ambitious. Our goals is to process intensity data: 1) to estimate the location of the seismogenic source, expressed as the center of the distribution of damage, and of its size, to be derived from the overall extent of the damage pattern (these first two steps are described in detail in the contribution «Deriving numerical estimates from descriptive information: the computation of earthquake parameters», by Gasperini and Ferrari (2000, this volume); 2) to make inferences about the physical dimensions of the seimogenic source (length and width) using empirical relationships (Wells and Coppersmith, 1994), and 3) to calculate the orientation of the seismogenic source using an original algorithm described later on in this paper. The estimated parameters are then calibrated using earthquakes for which both intensity and modern instrumental data are available. Each earthquake source is then conceptually and graphically delineated by an oriented «rectangle» representative of the fault at depth. This rectangle is meant to represent either the actual surface projection of the seismogenic fault or, at least, the projection of the portion of the Earth crust where a given seismic source is more likely to be located.

Our ultimate goal is to complement and strengthen with historical information the usually limited instrumental or surface faulting evidence forming the current earthquake distribution and recurrence models. This is done under the assumption that each source tends to produce *characteristic earthquakes* and that each earthquake is representative of the maximum source potential. To test our method we applied it systematically to a subset of the *Catalogue of Strong Italian Earthquakes* including all the $M > 5.5$ earthquakes that occurred in the Central and Southern Apennines during the past four centuries. After a first catalogue of historical sources has been obtained, additional and often unknown seismogenic faults can be inferred from the analysis of the spatial relations between adjacent sources, or between them and other non-damaging historical or instrumental earthquakes, or by recourse to more focused geological observations. The process allows the extent of overlap between adjacent sources, that is, the regularity of the seismic release in space, to be tested (if the sources tend to a *characteristic* behavior they should not overlap nor leave substantial gaps). Proving or disproving such regularity has obvious and important implications for the assessment of regional seismic hazard.

This contribution represents a condensed version of a paper published by Gasperini *et al.* (1999) in the *Bulletin of the Seismological Society of America*. Since its first elaboration the algorithm for deriving three-dimensional earthquake sources from intensity data has been widely used on Italian, Greek, Spanish, French and Northern European data in the framework of the EC project termed «Faust» (*e.g.*, Leschiutta *et al.*, 2000). In particular, the algorithm was used to derive over 150 «purely historical» sources in the framework of the «Database of potential sources for earthquakes larger than 5.5 in Italy» (Valensise and Pantosti, 2000).

The program «Boxer», that performs all the computations described in this paper, is available in executable form (for both Macintosh and Windows platforms) from: http://ibogfs.df.unibo.it/user2/paolo/www/boxer/boxer.html.

2. Previous efforts

Estimating fault parameters from Mercalli intensity data has been the object of numerous investigations during the past twenty years and indeed represents one of the most challenging developments of modern seismic hazard assessment. Probably the first attempt to derive quantitative information on the earthquake source from macroseismic data can be traced back to Shebalin (1973), who proposed to estimate the dimension and orientation of a seismogenic fault from the ellipticity of the highest degree isoseismals. This study, however, analyzed the shape of hand-drawn isoseismals of a limited number of selected earthquakes, and therefore its conclusions are still essentially qualitative.

Starting with the contribution of Ohta and Satoh (1980), several attempts have been made at modeling macroseismic intensities generated by sources of known geometry with various techniques. Among them are the Kinematic Function KF (Chiaruttini and Siro, 1981), the generation of synthetic seismograms by normal mode summation (Panza and Cuscito, 1982; Suhadolc *et al.*, 1988; Pierri *et al.*, 1993) or by ray-tracing (Zahradník, 1989). Some of these investigators have subsequently tried to infer the focal parameters of historical earthquakes from intensity data (*e.g.*, Chiaruttini and Siro, 1991; Sirovich, 1996), but so far the results of these attempts have not been extensively tested against instrumental data, partly due to the fact that good quality homogeneously collected intensity data were not widely available in revised catalogues until 1995.

Notwithstanding possible rapid developments of these techniques in the near future, we believe that at present most of them are not reliable enough for widespread application. In the absence of a physical model explaining the spatial pattern of intensity data (in particular, of a function for converting ground displacement, velocity and acceleration into felt intensity), macroseismic data alone have not been able to constrain efficiently parameters of the seismogenic source which do not have a straightforward correspondence with the observed intensity, such as the fault dip and the sense of slip (fault rake). On the contrary, the strike of the seismic source (that is, the azimuth of the seismogenic fault) is somehow related to the distribution of the earthquake effects. In the past, the fault azimuth was commonly inferred by means of a visual inspection of hand-drawn isoseismals (*e.g.*, in Shebalin, 1973 and subsequent Shebalin-type approaches). This procedure obviously introduces a great deal of arbitrariness, since the person in charge of drawing the isoseismals may somehow convey in the artwork his or her own preconceptions about the location and geometry of the seismogenic fault, often forcing the data to say more than they really show. More recently other workers have produced «objective» isoseismals through automatic computer contouring (*e.g.*, De Rubeis *et al.*, 1992), but also in this case all possible inferences can only be visual, and therefore subjective and almost impossible to test statistically.

A visual analysis of isoseismal lines may indeed be helpful for identifying survey blunders or anomalous intensity points resulting from site effects. However, the statistical analysis of individual observed intensity values must be preferred when the goal of the analysis is to derive global quantitative estimates of the main source parameters. This can now be done using the homogeneously collected and interpreted data supplied by the *Catalogue of Strong Italian Earthquakes* in its various versions.

3. Modeling approach

As a test of the algorithm we focused on the Central and Southern Apennines (a portion of the Italian territory between 40.0° and 43.2° latitude north, 12.8° and 17.0° longitude east) for this is the most seismically active region of the whole Italian peninsula and is characterized by a relatively well defined uniform tectonic pattern. For the region of our interest the second release of the *Catalogue of Strong Italian Earthquakes* (1997) lists 41 large earthquakes, for a total of over 7500 intensity data points (see table I). We decided to include in our analysis the felt reports gathered during a preliminary survey of the effects of the 26 September 1997, Colfiorito, Central Italy earthquakes (WGMSCE, 1997), as it allows for an interesting *a posteriori* test of our approach.

Our modeling strategy involves five steps (fig. 1):

Step 1 – Locating the source – We first compute the epicenter of each of the 42 earthquakes from macroseismic data alone. The epicenter, that is found through the averaging technique described by Gasperini and Ferrari (2000, this volume), is then used as the origin of the reference system for locating the extended source and for analyzing the azimuthal distribution of felt intensities to determine the source strike.

Step 2 – Assessing the earthquake seismic moment – The distribution of felt intensities of each earthquake is then used to infer the earthquake seismic moment M_0 and the corresponding moment magnitude $\boldsymbol{M}$ using the algorithm also described by Gasperini and Ferrari (2000, this volume).

Step 3 – Assessing the source dimensions (length and width) – The seismic moment of each individual earthquake is then used to infer the physical dimensions of the relevant source under the hypotheses set forth above (each source tends to produce characteristic earthquakes and each earthquake is representative of the maximum source potential). We used Wells and Coppersmith's (1994) empirical relationships to calculate the full rupture length and width of the seismogenic source. Although most of the best studied strong Italian earthquakes exhibit pure normal faulting, reverse and strike-slip faulting earthquakes are expected to take place particularly along the Adriatic margin. For this reason we used the relationships that were derived by these investigators as an average of «All» possible faulting styles

$$\mathrm{Log}_{10}(\mathrm{RLD}) = 0.59(\pm 0.02) \cdot \boldsymbol{M} - 2.44(\pm 0.11)$$
$$\mathrm{Log}_{10}(\mathrm{RW}) = 0.32(\pm 0.02) \cdot \boldsymbol{M} - 1.01(\pm 0.10)$$

where RLD and RW are the subsurface rupture length and the down-dip rupture width, respectively, and $\boldsymbol{M}$ is the moment magnitude.

Step 4 – Assessing the source orientation (azimuth) – Once the source has been located and its physical dimensions evaluated, this step involves assessing its true orientation. This is accomplished by a new algorithm described in detail in the following section.

Step 5 – Representing the source – The seismic source is finally drawn as a rectangle centered on the macroseismic epicenter. The rectangle represents either the actual surface projection of the causative fault or, at least, the surface projection of the portion of the Earth crust within which the fault is more likely to be located. Since Italian faults tend to be predominantly dip-slip, as a first approximation the width of the rectangle delineating each source is plotted as if it represented the projection of a fault dipping 45° in an unspecified direction perpendicular to the fault strike (see fig. 2a).

4. The assessment of the orientation of the seismogenic source

Our reasoning starts from the common observations that the direction of maximum elongation of the highest degree isoseismals is controlled by the geometry of the seismogenic structure and that the highest degree isoseismals tend to approximate the projection of the fault upon the Earth surface (fig. 2a). We assume that these observations are the perceptible expression of a physical link between the source at depth and

Table I. The table lists 41 $M > 5.5$ earthquakes of the Central and Southern Apennines taken from the second release of the *Catalogue of Strong Italian Earthquakes* (1997), plus the 1997 Colfiorito earthquake. M_e is the «equivalent magnitude» computed analyzing the distribution of Mercalli intensity data (Gasperini and Ferrari, 2000, this volume), that we assume to represent the moment magnitude M. NTot and NAz are total number of data available for the given earthquake and number of data used for computing the source azimuth, respectively. The reported azimuths are those obtained using our preferred choice for distance weighting and lower bound for the intensity threshold (see text) and correspond to the orientations of the solid white rectangles in fig. 5. The table also lists the standard deviation of the computed azimuths (under the assumption of Von Mises-type distribution) and the significance levels of the Rayleigh and Kuiper tests («uniform» means that the text returns a significance level higher than 0.10 and therefore the hypothesis H_0 of uniformity of the data distribution cannot be rejected: see Appendix 3 in Gasperini *et al.*, 1999, for details).

Date	Lat.	Long.	M_e	Locality	NTot	NAz	Azimuth	Rayleigh	Kuiper
07/30/1627a	41.74	15.34	6.7	Gargano	65	22	111 ± 37	< 0.10	< 0.01
07/30/1627b	41.69	15.38	5.8	San Severo	1	–	–	–	–
08/07/1627c	41.76	15.33	5.9	Gargano	5	3	111 ± 44	Uniform	Uniform
09/06/1627d	41.60	15.36	5.8	Gargano	2	–	–	–	–
10/15/1639	42.65	13.25	6.0	Monti della Laga	15	10	062 ± 20	< 0.01	< 0.05
05/31/1646	41.87	15.94	6.1	Gargano	18	5	062 ± 156	Uniform	Uniform
07/23/1654	41.63	13.68	6.1	Sorano-Marsica	44	16	110 ± 47	Uniform	< 0.01
06/05/1688	41.28	14.56	6.6	Sannio	26	17	118 ± 14	< 0.01	< 0.01
09/08/1694	40.88	15.34	6.8	Irpinia	294	31	121 ± 12	< 0.01	< 0.01
03/14/1702	41.12	14.99	6.3	Beneventano	37	4	107 ± 189	Uniform	Uniform
01/14/1703a	42.68	13.12	6.7	Norcia	196	33	165 ± 30	< 0.05	< 0.01
01/16/1703b	42.62	13.10	6.0	Roio Piano	22	10	117 ± 146	Uniform	< 0.05
02/02/1703c	42.46	13.21	6.6	Aquilano	70	7	114 ± 28	< 0.05	< 0.05
11/03/1706	42.08	14.08	6.6	Maiella	99	7	136 ± 18	< 0.01	< 0.01
05/12/1730	42.74	13.12	6.3	Umbrian Apennines	22	10	004 ± 59	Uniform	< 0.01
03/20/1731	41.27	15.76	6.5	Foggiano	50	5	120 ± 35	< 0.10	< 0.10
11/29/1732	41.08	15.06	6.5	Valle Ufita	168	4	092 ± 96	Uniform	Uniform
10/06/1762	42.31	13.59	5.6	Aquilano	6	–	–	–	–
07/31/1786	42.32	13.37	5.6	San Demetrio	7	–	–	–	–
03/18/1796	40.75	13.91	5.7	Casamicciola Terme	1	–	–	–	–
07/26/1805	41.50	14.47	6.5	Molise	223	8	124 ± 27	< 0.05	< 0.05
02/01/1826	40.52	15.73	5.8	Basilicata	18	4	162 ± 44	Uniform	< 0.10
11/20/1836	40.14	15.78	6.3	Southern Basilicata	17	6	151 ± 34	< 0.10	< 0.10
08/14/1851a	40.96	15.67	6.3	Basilicata	102	6	161 ± 33	< 0.10	< 0.05
08/14/1851b	40.99	15.65	5.6	Melfi	10	–	–	–	–
04/09/1853	40.82	15.22	5.9	Irpinia	47	6	005 ± 40	Uniform	< 0.05
12/16/1857	40.35	15.84	6.9	Basilicata	337	18	127 ± 11	< 0.01	< 0.01
03/12/1873	43.09	13.24	6.0	Polverina	196	3	080 ± 09	< 0.01	Uniform
09/10/1881	42.23	14.28	5.6	Lanciano	29	4	037 ± 17	< 0.01	< 0.05
07/28/1883	40.74	13.89	5.7	Casamicciola Terme	27	8	092 ± 77	Uniform	< 0.05
02/24/1904	42.10	13.32	5.6	Marsica	22	7	086 ± 48	Uniform	Uniform
06/07/1910	40.90	15.42	5.8	Irpinia	376	4	126 ± 16	< 0.01	< 0.10
01/13/1915	41.99	13.65	6.9	Avezzano	860	25	122 ± 16	< 0.01	< 0.01
07/23/1930	41.05	15.36	6.7	Irpinia	511	16	108 ± 11	< 0.01	< 0.01
09/26/1933	42.05	14.19	6.0	Maiella	326	3	025 ± 05	< 0.01	< 0.05
10/03/1943	42.91	13.65	5.9	Offida	131	16	149 ± 22	< 0.01	< 0.01
08/18/1948	41.58	15.75	5.7	Zapponeta	59	9	010 ± 31	< 0.05	< 0.01
08/21/1962	41.14	14.97	6.2	Irpinia	214	11	160 ± 28	< 0.05	< 0.01
09/19/1979	42.71	13.07	5.8	Valnerina	691	30	156 ± 24	< 0.01	< 0.01
11/23/1980	40.84	15.28	6.9	Irpinia	1319	15	126 ± 25	< 0.01	< 0.01
05/07/1984	41.67	14.06	5.9	Val Comino	913	3	152 ± 34	< 0.10	Uniform
09/26/1997	*43.02*	*12.87*	*5.8*	*Colfiorito*	*182*	*19*	*145 ± 10*	*< 0.01*	*< 0.01*

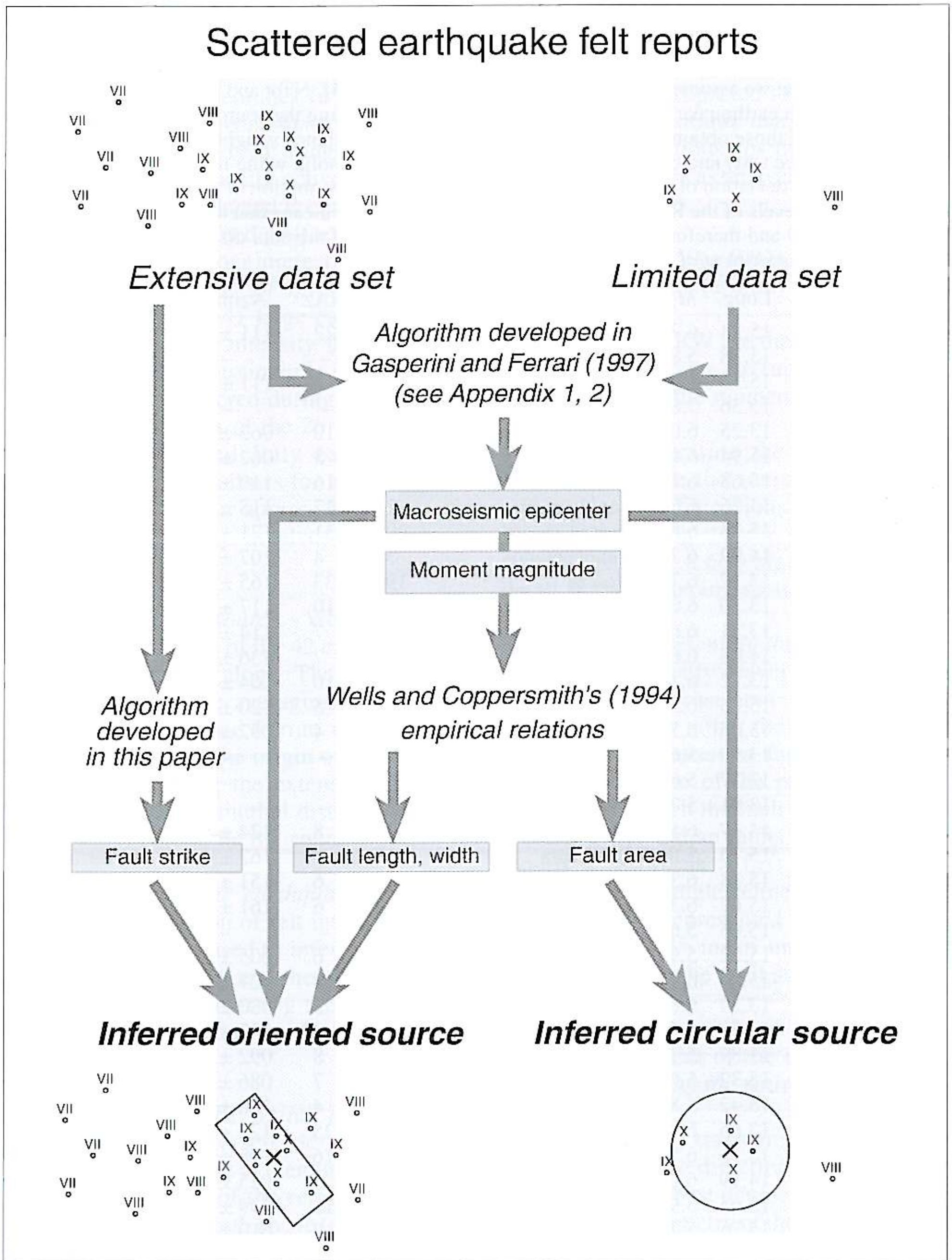

Fig. 1. Block diagram showing the various steps of our analysis.

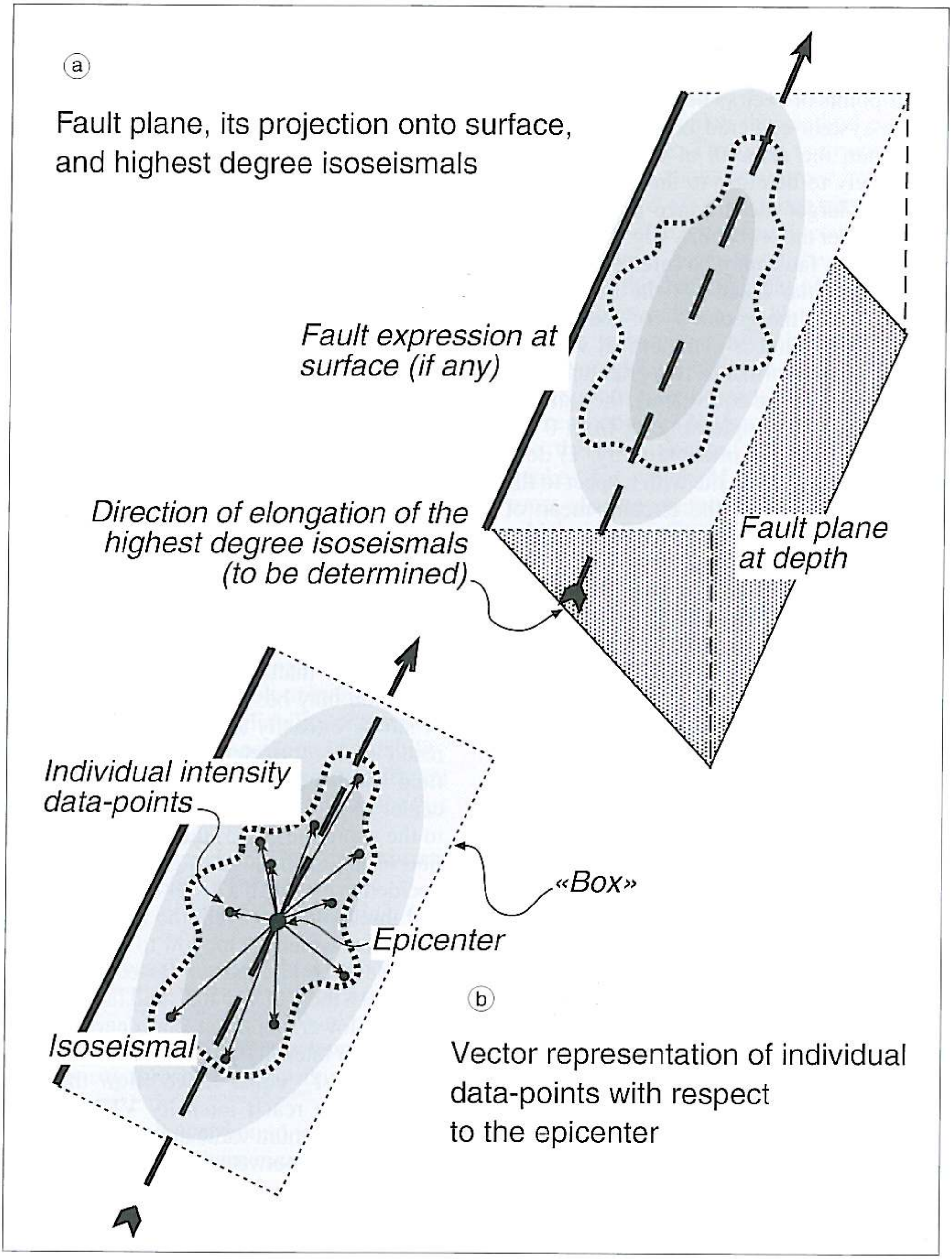

Fig. 2a,b. Geometry of the problem. a) Fault plane, its surface projection, and isoseismals. b) Vector representation of Mercalli intensity data points with respect to the macroseismic epicenter (*i.e.* the mid point of the data distribution).

the pattern of ground shaking at the surface. Under this assumption, if all the sites where the largest intensities were observed are considered as the end-points of vectors belonging to a polar coordinate system centered on the macroseismic epicenter, the azimuth of each individual vector is likely to be close to the true strike of the fault; the larger the distance from the epicenter, the higher the probability (fig. 2b). Hence, the strike of the fault may be inferred by computing the «circular mean» of the azimuth of these sites. Since the geometry of the surface projection of the fault is symmetrical with respect to the epicenter (that is, two orientations at 180° to one another are equivalent), the azimuth is represented by an angle ranging from 0° to 180°. We calculate this orientation by: 1) doubling the azimuth of each site with respect to the epicenter; 2) calculating the circular mean of these angles, and 3) halving the resulting circular mean.

Similar to any other analysis of angular data, the reliability of the circular mean rests upon the uniformity in the distribution of the data themselves. Since the angular location and the dispersion are not independent variables, a uniform distribution has no significant central value and any further statistical analysis is therefore generally meaningless. A number of statistical tests are available to analyze the uniformity of a circular distribution. Among them the Kuiper test proved preferable for small data sets, while the Rayleigh test proved most powerful where the distribution of the parent population is Von Mises-type (Rock, 1988). The mathematical details of the procedure for calculating the circular mean, the associated standard deviation and the significance levels of the distribution uniformity tests are described in Appendix 3 of Gasperini *et al.* (1999).

Before running the algorithm with real data we must select an appropriate lower threshold for the macroseismic intensity of the data points to be included in the averaging process. Ideally, the data set of each earthquake source should include only the localities where the observed intensity is largest. In real applications, however, such maximum effects often occur at a limited number of scattered sites as a result of local amplifications induced by the near-surface geology, of particular characteristics of the local buildings, of focusing of the seismic energy or constructive interference of wave-trains from different portions of the source. Under such circumstances the highest intensities of a given earthquake may represent outliers in the data distribution, in which case the source is more correctly represented by the pattern of sites that experienced an intensity one or even two degrees lower than the maximum observed.

A rational criterion to choose the intensity threshold is to select a value such that the average epicentral distance of sites having a larger intensity is comparable to the fault size. As we have seen in Step 3, we can use Wells and Coppersmith's (1994) empirical relationships to calculate an approximate source length as a function of $\boldsymbol{M}$. We can then pick the intensity threshold that gives the best equivalence between half of the fault length and the average distance of the data points from the epicenter. Nevertheless, our experience shows that the plain application of these criteria may lead us to retain data points having an intensity more than two or three degrees lower than the maximum intensity. This condition may be: 1) the effect of the presence of strong intensity amplification effects; 2) the result of incompleteness of the macroseismic field and hence of mislocation of the true epicenter (*e.g.*, when this occurs offshore or close to the shoreline), or 3) the result of overestimation of the earthquake magnitude. We therefore decided to establish a lower bound for the intensity threshold to prevent the inclusion of intensities data which are too low to be representative of the source orientation. Based on our experience with the data set analyzed in this paper, we set this lower bound at one degree below the maximum intensity plus uncertainty (for example, for an I_{max} equal X we allow the intensity threshold to reach intensity VIII-IX). This is also the maximum value normally attained from the difference between the epicentral intensity I_0 and the maximum intensity I_{max}.

A further important issue is the choice of an appropriate distance weighting scheme. Under the assumptions mentioned above and for any given intensity, the farther a certain site, the higher the probability that the azimuth of that site approximates the strike of the fault. We assume that this probability is proportional to some function of the distance, normalized by the average epicentral distance of all data points having the same intensity. This function should be somehow related to the attenuation of the intensity with distance. A simple relation, which proved to fit well the attenuation of earthquake intensity for the Italian territory was proposed by Berardi *et al.* (1993). This relation, termed CRAM (Cubic Root Attenuation Model), is given by the following expression:

$$\Delta I = \mathrm{a} + \mathrm{b} D^{1/3}$$

where $\Delta I = I_0 - I$ is the difference between the epicentral intensity and the intensity observed at a given site, and D is this site's distance from the epicenter. A least squares fit over all the sites having an assigned intensity in our database returned a = −0.46 and b = 0.93 (the corresponding coefficient of variation is $R^2 = 0.52$). We can then invert the CRAM relation to estimate the average normalizing distance for each intensity and use the cubic root of the normalized distance as a weight assigned to each of the data used to estimate the fault azimuth.

To evaluate the overall reliability of Step 4 of our modeling approach we performed a stability analysis by comparing the results of different weighting schemes. Reasonable choices for this test include:

a) No distance weighting (all data are assigned the same weight).

b) Cubic root of distance weighting (see above).

c) Distance weighting (weight is proportional to the normalized distance of the point from the epicenter).

Similarly, we tested different lower bounds for the intensity thresholds according to the following schemes:

d) Zero degree lower bound (which implies that only data points where *maximum intensity* is observed are used).

e) One degree lower bound (see discussion above).

f) No lower bound (all available data could be used).

Notice that in both cases we are essentially comparing the results of our preferred or «central» schemes, indicated by (b) and (e) and already described in the text, with those obtained using two extreme scenarios.

Finally, we need to define the minimum size earthquake for which the method can be used with confidence. This step is crucial since the analysis of earthquake sources comparable in size with the average distance between the sites used to estimate the azimuth could yield meaningless results because of the low «resolving power» of the data distribution itself. Based on the average spacing of historical settlements in Italy, we assume this minimum fault length to be somewhere between 5 and 10 km, which corresponds to a moment magnitude of 5.3 and 5.8 respectively (Wells and Coppersmith, 1994). We therefore decided to analyze only earthquakes for which $M > 5.5$. Good candidates must also be characterized by at least 5 data, which is the minimum figure for which the statistical tests hold rigorously. This condition applies to 27 out of 42 earthquakes of magnitude 5.5. and above reported by the *Catalogue of Strong Italian Earthquakes* for the region of interest (see column NAz in table I). To maximize the use of available data, however, we tentatively extended the application of the algorithm to 9 additional events for which at least 3 data points are made available by the selection criteria.

5. Modeling results

Figure 3a-d shows the full modeling procedure applied to the 23 July 1930, Irpinia (Southern Italy) earthquake (see also fig. 1 and discussion on modeling strategy in previous section). For this earthquake there exists an instrumental estimate of $M_S = 6.6$ (Margottini *et al.*, 1993). The macroseismic data set includes 511 localities with MCS intensities in the range II to X. The epicenter (a) is computed by averaging the coordinates of the 3 sites where the maximum intensity (X) was observed (see Gasperini and Ferrari, 2000, this volume). The moment magnitude ($M = 6.7$: (b), that was computed with the mixed epicentral intensity-isoseismal radii method (see Gasperini and Ferrari, 2000, this volume), is slightly larger than the instrumental

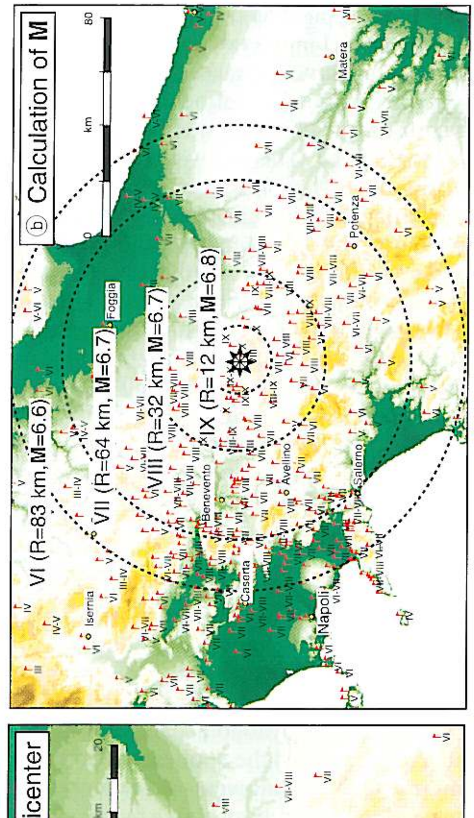

Fig. 3a-d. Application of the procedure to a sample earthquake (23 July 1930, Irpinia earthquake: see text). a) Determination of the macroseismic epicenter; b) of the macroseismic moment magnitude; c) of the source azimuth, and d) final representation of the inferred source.

estimate. The source azimuth (N108°W: (c)) was determined using 16 intensity data in the range VIII-IX to X. The length and width of the inferred source (32.6 and 13.6 km, respectively: (d)) are computed from Wells and Coppersmith's (1994) relationships; to account for the presumable dipping geometry of the fault, however, we plotted the source width as the surface projection of a 45°-dipping plane (multiply the width by the cosine of 45°).

A similar procedure was followed for all the 42 M > 5.5 Central and Southern Apennines earthquakes that we selected to test our approach. Figures 4 and 5 show the location, extent and orientation of the inferred sources. These earthquakes occurred between 1600 A.D. and present, and are shown by rectangles constructed using exclusively historical information following the five steps of our modeling scheme. We recall that such rectangles comprise a *synthetic representation* of the source that is coherent with standard schematizations based on instrumental or field evidence. Figure 3a-d shows the results obtained using different distance weighting schemes, while fig. 5 shows the effect of different choices of different lower bounds for the intensity threshold. Under the relatively strict requisites of our preferred schemes ((b) and (e)), for 6 out of 42 earthquake sources we could calculate only the location and size but not the azimuth (all of them have M < 6.0), and for this reason they are shown with circles in which the diameter is the estimated fault length (except for three solutions obtained with the more tolerant scheme (f), for which a rectangle is shown).

For many sources the estimated azimuth does not differ much for different distance weighting schemes (see fig. 4). The most evident discrepancy concerns the large 1627a, Gargano earthquake, which appears to vary from a trend almost parallel to the Apennines (from about N60°W for the (a) and (b) schemes to about N30°E for the (c) scheme). Less pronounced differences (within 10°) can be observed for some of the largest earthquakes such as the 1980, Irpinia; the 1915, Avezzano; the 1732, Valle Ufita, and the 1703a, Norcia. In general, the algorithm seems quite stable for different weighting schemes; in particular, in almost all cases the (b) scheme returns a result that is intermediate with respect to the other two, and for this reason we decided to regard it as our best choice.

Choosing different lower bounds for the intensity thresholds (see fig. 5) also does not appear to return drastically different results. The solutions obtained using the (e) and (f) schemes are almost coincident for most of the earthquakes. The only significant difference concerns the 1703a, Norcia earthquake and three relatively small (M < 6.0.) earthquakes (1762, 1786 and 1851b), for which the azimuth can only be computed using the (f) scheme. On the contrary, larger deviations exist between the (d) scheme and the other two; the largest of them again concerns the 1627a, Gargano earthquake, the source of which varies in orientation by nearly 90° from one scheme to another. For two other large earthquakes (1688, Sannio and 1980, Irpinia) the deviation ranges between 10° and 20°. It should also be noted that in 21 cases (*versus* 6 for scheme (e) and 3 for scheme (f)) the azimuth cannot be computed with the more demanding scheme (d) due to an insufficient number of data points (less than 3). In contrast, the algorithm is rather stable with respect to the other two schemes. This suggests that the choice of using only data having the same intensity as the epicentral intensity (scheme (d)) is too restrictive and represents an unjustified limitation of the applicability of the algorithm. We therefore decided to assume the (e) scheme, which is also more plausible from the point of view of the physics of the problem, as the most reasonable and reliable choice.

Our preferred solutions are shown by white rectangles enclosed by a solid line in fig. 5 and are listed in table I along with our estimated macroseismic epicenter and moment magnitude M. For each earthquake table I also reports the significance level (s.l.) obtained from the Rayleigh and Kuiper distribution uniformity tests (see Appendix 3 in Gasperini *et al.*, 1999). In most cases the Kuiper test allows the H_0 hypothesis (data distribution is uniform) to be rejected at least at s.l. < 0.05, and therefore the source orientation can be estimated with confidence. For 11 earthquakes the significance level is larger than 0.05 and the H_0 hypothesis cannot be

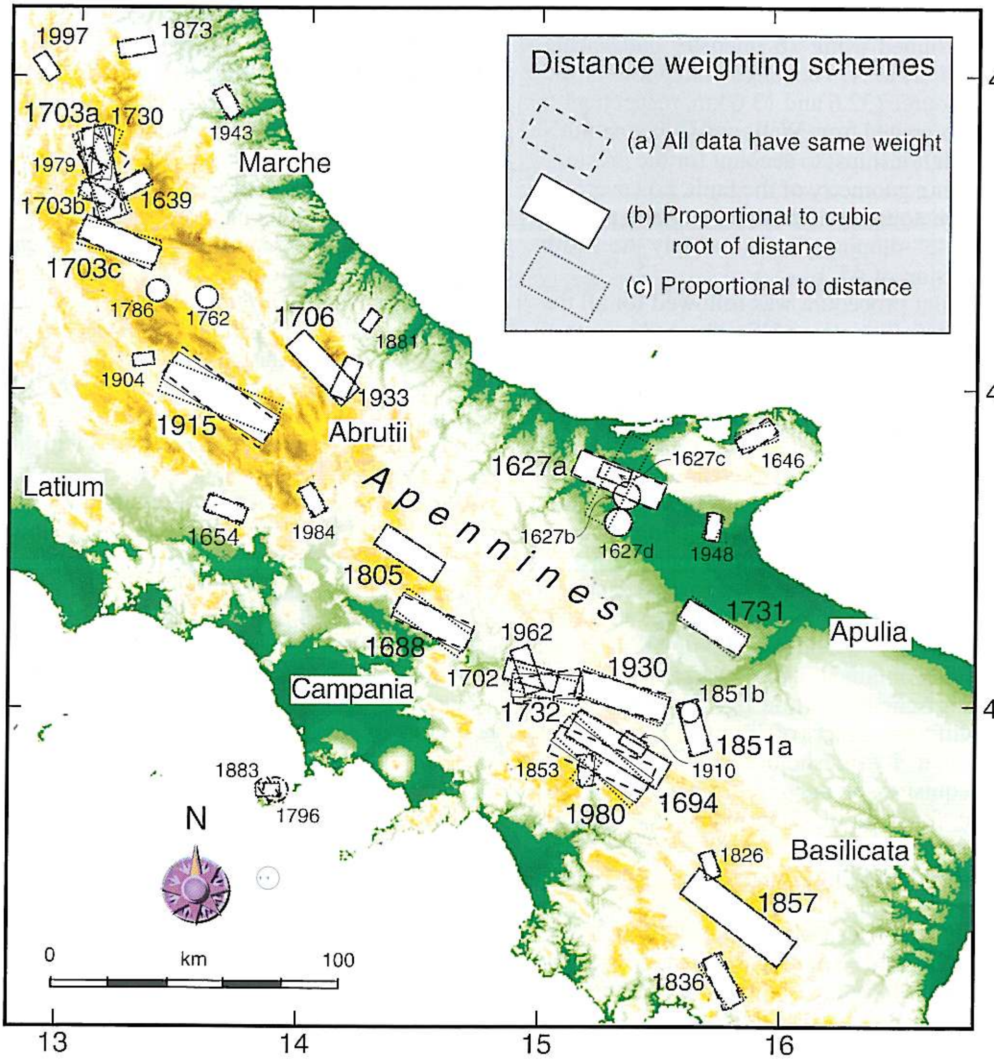

Fig. 4. $M > 5.5$ earthquakes in the Central and Southern Apennines from the year 1600 to 1997, with rectangles representing the surface projection of the inferred seismogenic sources. The source azimuths are computed as described in the text. The larger side of the rectangle represents the fault length computed as function of the moment magnitude $\boldsymbol{M}$ using Wells and Coppersmith's (1994) relationships; the smaller side represents the surface projection of the fault width assuming a dip angle of 45°. The figure shows the solutions obtained for different distance weighting schemes (see text and legend in figure). The rectangles drawn with solid lines represent our best guess and were obtained using the (b) scheme (cubic root weighting). Notice that (a) and (c) scheme solutions may not appear if identical to the corresponding (b) solution. A circle having the diameter equal to the fault length replaces the rectangle for all sources for which the azimuth could not be computed due to insufficient number of data points (less than 3).

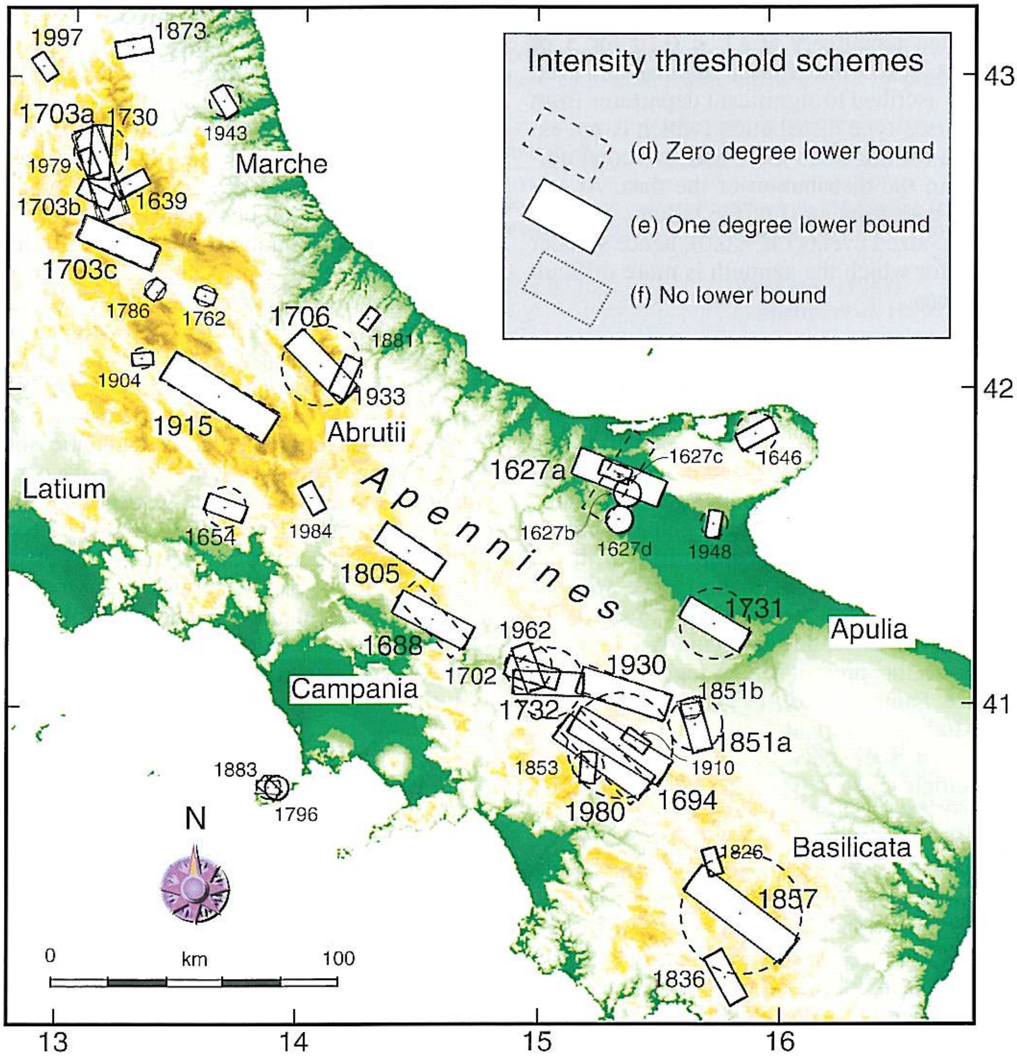

Fig. 5. Same as fig. 4 except for the rectangles (seismogenic sources), which are obtained using the (b) weighting scheme (cubic root weighting: see text and fig. 4) and varying the lower bound for the intensity threshold according to three different schemes (see text and legend in figure). The rectangles drawn with solid lines represent our best guess and were obtained using the (e) scheme (one degree lower bound). Notice that (d) and (f) scheme solutions may not appear if identical to the corresponding (e) solution.

confidently rejected. We could tentatively reject the H_0 hypothesis for 4 of these 11 earthquakes where s.l. < 0.10, whereas for the remaining 7 events the results must be considered with caution. The reason why we do not simply discard these results is because for most of these events the test statistics are not rigorous as the source azimuth was computed using less than five intensity data. In turn, the Rayleigh test does not allow the uniformity hypothesis to be rejected for about half of the computed azimuths (it can be rejected tentatively at s.l. < 0.10 for 5 of them). At least some of these failures, however, could be ascribed to significant departures from a Von Mises-type distribution (which is not established for our data) more than to actual uniformity in the distribution of the data. At any rate, for both tests most of the failures concern moderate-sized events (**M** < 6.0), hence smaller sources for which the azimuth is more difficult to estimate.

6. Comparing the inferred sources with instrumental and geologic evidence

As most of the 42 analyzed earthquakes occurred in the pre-instrumental era, very few extended fault models and focal plane solutions are available for a direct comparison with our intensity-derived sources. In this respect we wish to recall that, due to the peculiar characteristics of Italian tectonics, and particularly the youthfulness of the present stress regime (see, for example, Pantosti *et al.*, 1993), very few of the major Italian historical earthquakes have been positively associated with a well-identified active tectonic feature. The CMT database (published in Dziewonski *et al.*, 1981 and subsequent quarterly papers on *Physics of the Earth and Planetary Interiors*) supplies data for the four most recent earthquakes (1979, Valnerina; 1980, Irpinia; 1984, Val Comino; 1997, Colfiorito). For the 1962, Irpinia earthquake, a reasonably reliable focal mechanism computed using *P*-wave polarities is given by Westaway (1987). Direct surface faulting evidence was documented for the 1915, Avezzano and for the 1980, Irpinia earthquakes, both of which were modeled also by inversion of coseismic elevation changes (see discussion below).

Table II shows a summary of the comparison between these instrumentally or geologically derived azimuths and the results of our intensity-based computations. In general the agreement is quite satisfactory. In particular for the 1980, Irpinia earthquake our estimate is very close (within 10°) to the orientation of both of the CMT nodal planes and of the geologically inferred fault. For the 1979, Valnerina earthquake our solution is almost coincident with one of the two nodal planes, but unfortunately no geological or seismological evidence is available to date to decide which is the actual rupture plane. For the 1915, Avezzano earthquake the maximum difference is 13°. For the remaining two earthquakes (1962, Irpinia and 1984, Val Comino) our result lies almost in the middle of the two instrumental solutions, with differences in the order of 20°-30°. Also for these two events no conclusive evidence exists to date as to which of the two nodal planes is the actual rupture plane.

For three especially well documented earthquakes we decided to extend the comparison to the full definition of the seismogenic source. Figures 6a-c summarize the results of a comparison of evidence available respectively for the 1915 Avezzano, 1980 Irpinia, and 1997 Colfiorito earthquakes *versus* the estimates derived in this paper. The following discussion focuses on the most evident discrepancies that emerge from this comparison. For the source parameters that are fit satisfactorily the reader may refer directly to the information shown in figs. 6a-c.

1915 Avezzano – The 1915, Avezzano (Central Italy), is the second deadliest earthquake of Italian history (fig. 6a). The concentration of population in the depression of the former Fucino Lake, which had been reclaimed in the 1860's and soon after re-utilized for extensive agricultural development and for new settlements, and the widespread amplifications of the ground motion induced by the particular configuration of the area, conspired in turning this earthquake into an immense catastrophe. Perhaps for this reason in current catalogues this earthquake is characterized by a large number of localities which were assigned intensity XI (see fig. 6a). This circumstance has driven the inferred **M** up to 6.9, which implies a nearly 40 km-long causative fault. The geodetic model proposed by Ward and Valensise (1989) implies a **M** 6.6, but this is a minimum figure as it is based on the portion of the fault that could be resolved by observations of coseismic strain. In view of this limitation and given the extent of the observed surface ruptures, we may conclude that the true **M** of the 1915 earthquake was between 6.7 and 6.8.

Table II. Source azimuths computed in this paper are compared with the corresponding seismological, geological or geodetic estimates for six of the largest earthquakes that occurred in the study region during this century. Azimuths are derived from strikes of focal mechanism nodal planes by reducing them to the 0°-180° range. Note that available published literature suggests that all of these earthquakes were characterized by predominantly normal faulting.

Date	Locality	Azimuth(s)	Reference
01/15/1915	Avezzano	~ 130°	Serva *et al.* (1986)
		135°	Ward and Valensise (1989)
		122°	Gasperini *et al.* (1999)
08/21/1962	Irpinia	130° or 6°	Westaway (1987)
		160°	Gasperini *et al.* (1999)
09/19/1979	Valnerina	3° or 161°	CMT
		156°	Gasperini *et al.* (1999)
11/23/1980	Irpinia	135° or 123°	CMT
		125°÷135°	Pantosti and Valensise (1990)
		126°	Gasperini *et al.* (1999)
05/07/1984	Val Comino	174° or 132°	CMT
		152°	Gasperini *et al.* (1999)
09/26/1997	Colfiorito (2 shocks)	143° and 154°	Ekström *et al.* (1998)
		145°	Gasperini *et al.* (1999)

Part of the misfit in the orientation of the fault could be accounted for by the northwestward propagation of the coseismic rupture (Berardi *et al.*, 1995) and by the lack of settlements to the north and south of the epicenter.

1980 Irpinia – Our intensity-based source for the 1980, Irpinia earthquake (fig. 6b) is quite surprising for it fits the real seismic source nearly to perfection except for its location, which is shifted to the northwest by about 8 km. Indeed the macroseismic solution could not capture the intrinsic complexity of the earthquake rupture, that was characterized by at least three discrete subevents occurring within a 40 s time span, but it somehow responded to the northwestward propagation of the rupture (*e.g.*, Bernard and Zollo, 1989), which caused an asymmetry in the distribution of the highest reported intensities with respect to the location of the source.

1997 Colfiorito – The Colfiorito earthquakes (fig. 6c) make an especially interesting case as they occurred immediately after the modeling procedure and its parameters had been firmly established based on the experience gained from the rest of our data set. The analysis uses the results of a preliminary survey of the earthquake (WGMSCE, 1997) completed on 2 October, that is, a week after the mainshocks, because the damage pattern was soon after worsened by a series of strong aftershocks ($M > 5$), which effectively extended the region that ruptured during the sequence.

The main limitation of our macroseismic solution is represented by its inadequacy to account for multiple ruptures occurring closely spaced in time. Unlike the case of 1980, when the moment release was dominated by the first mainshock subevent, the two mainshocks of the Colfiorito sequence were comparable in size and are presumed to have ruptured in opposite directions, generating a pattern of cumulative damage that does not fully reflect the actual energy release. We believe that, had the two shocks occurred separated in time by a few years, our approach would have retrieved the correct extent of each individual source.

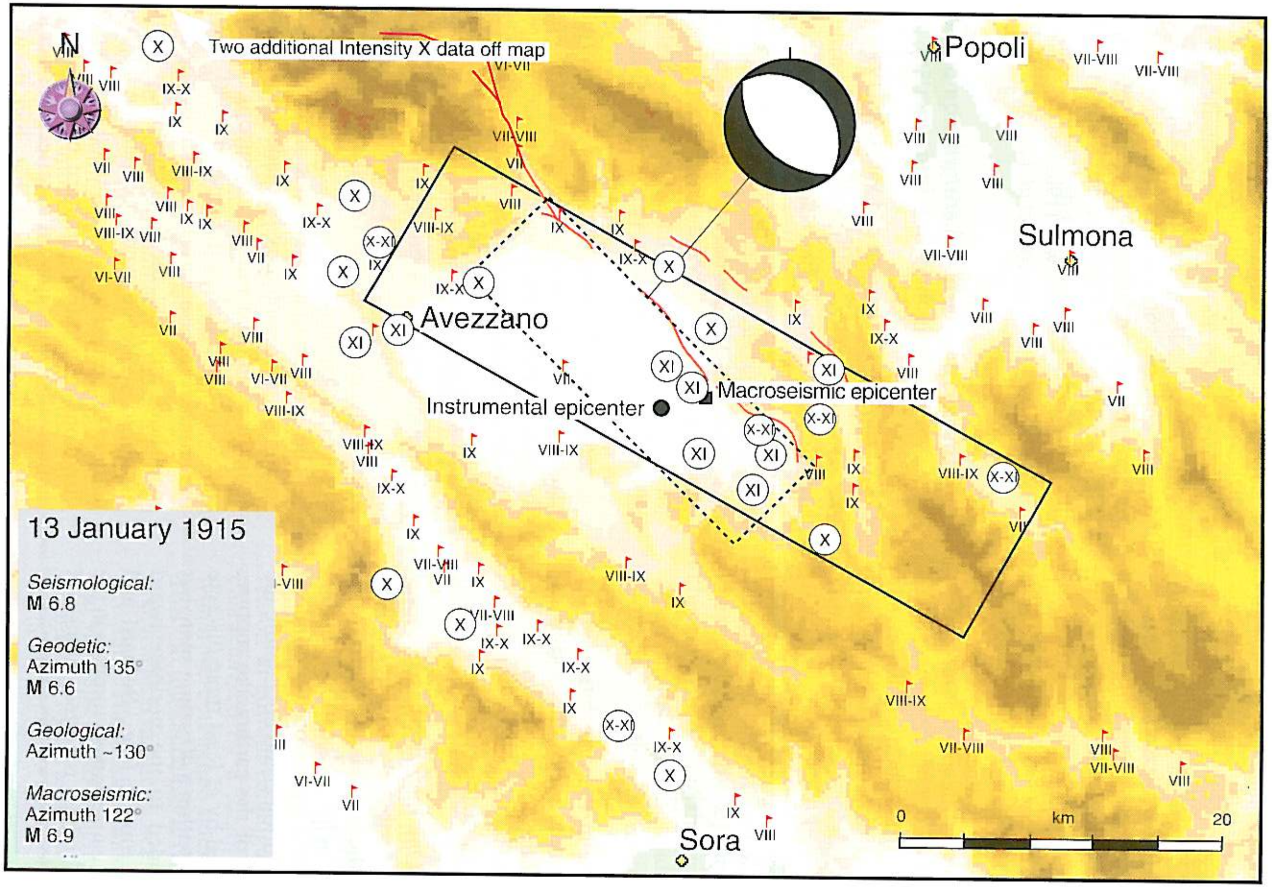

Fig. 6a. Visual comparison of source parameters obtained in this paper with published estimates from seismological, geological and geodetic data for Avezzano earthquake of 13th January 1915 (see also table II). The solid box represents our best guess of the source dimension, size and location obtained from macroseismic data alone (same as in figs. 4 and 5). The dashed box delineates a solution available in current literature (from geological and geodetic data: Serva *et al.*, 1986; Ward and Valensise, 1989). Intensity data from the *Catalogue of Strong Italian Earthquakes*, second release (1997). Discontinuous red lines indicate fault scarps from Serva *et al.* (1986). The instrumental epicenter of the 1915 earthquake was calculated by Basili and Valensise (1986). The symbol ***M*** indicates moment magnitude. Note the strong asymmetry of the damage pattern with respect to the surface expression/projection of the fault, which is strong evidence for a significant fault dip.

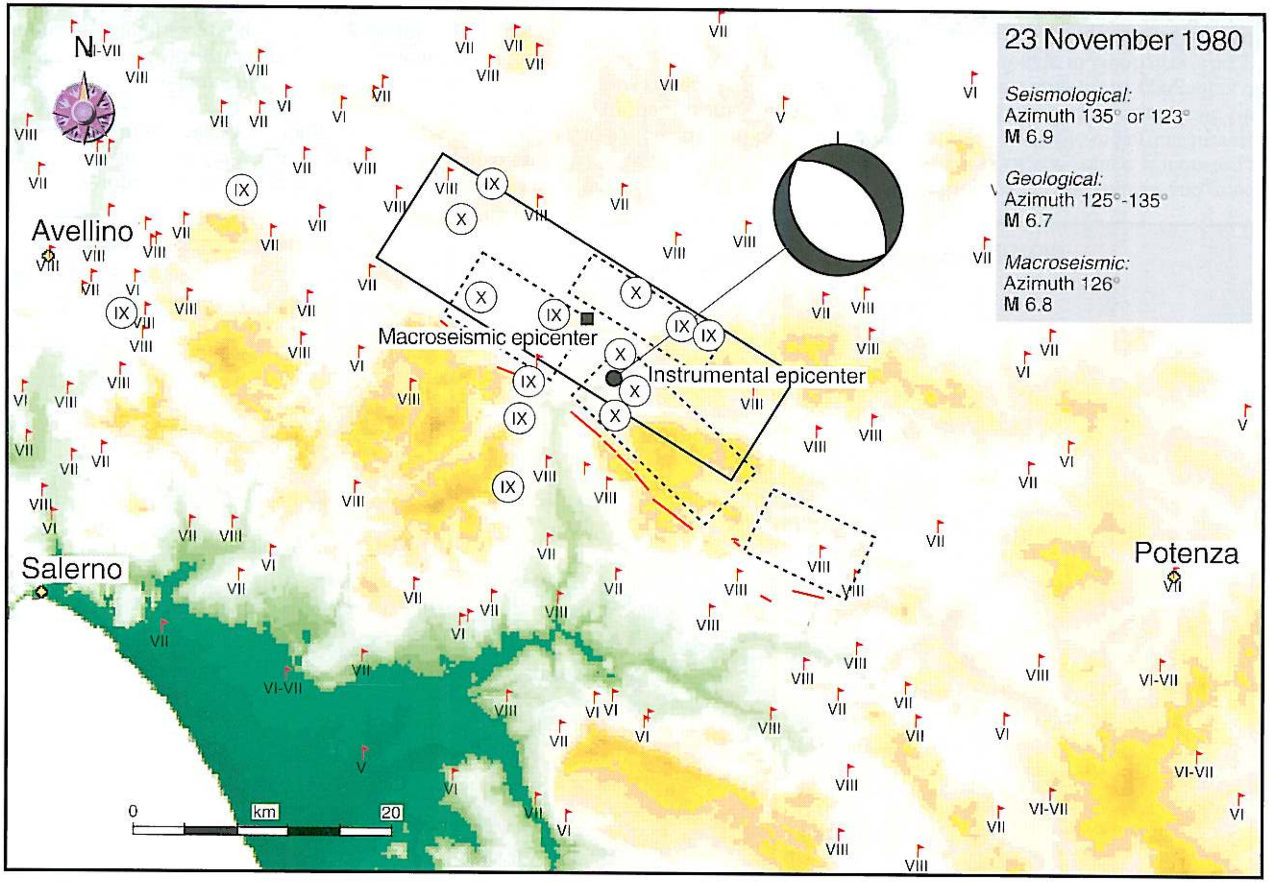

Fig. 6b. Visual comparison of source parameters obtained in this paper with published estimates from seismological, geological and geodetic data for Irpinia earthquake of 23th November 1980 (see also table II). The solid box represents our best guess of the source dimension, size and location obtained from macroseismic data alone (same as in figs. 4 and 5). The dashed box delineates a solution available in current literature (from seismological, geological and geodetic data: Westaway and Jackson, 1984; Bernard and Zollo, 1989; Pantosti and Valensise, 1990). Intensity data from the *Catalogue of Strong Italian Earthquakes*, second release (1997). Discontinuous red lines indicate fault scarps from Pantosti and Valensise (1990). The instrumental epicenter of the 1980 earthquake was calculated by Giardini *et al.* (1996). The symbol ***M*** indicates moment magnitude. Note the strong asymmetry of the damage pattern with respect to the surface expression/projection of the fault, which is strong evidence for a significant fault dip.

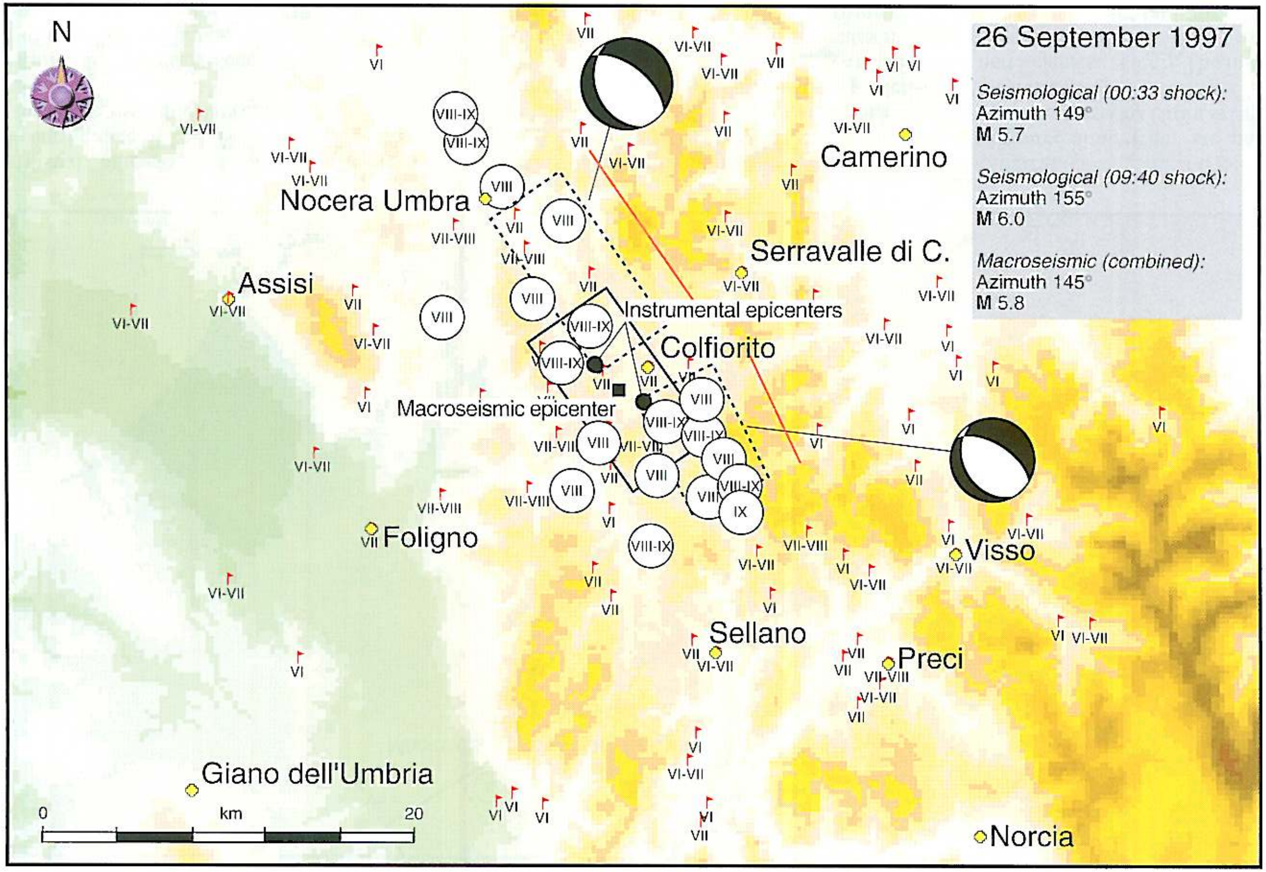

Fig. 6c. Visual comparison of source parameters obtained in this paper with published estimates from seismological, geological and geodetic data for Colfiorito earthquake of 26th September 1997 (see also table II). The solid box represents our best guess of the source dimension, size and location obtained from macroseismic data alone (same as in figs. 4 and 5). The dashed box delineates a solution available in current literature (from seismological data: Amato *et al.*, 1998; Ekström *et al.*, 1998). Intensity data from WGMSCE (1997). Continuous red lines indicate the surface projection of the causative faults of the 1997 earthquakes (no unambiguous surface faulting was reported for these events). The epicenters of the two mainshock of the 1997 Colfiorito earthquake sequence were taken from the *Monthly Bulletin* of the Istituto Nazionale di Geofisica. The symbol ***M*** indicates moment magnitude. Note the strong asymmetry of the damage pattern with respect to the surface expression/projection of the fault, which is strong evidence for a significant fault dip.

In summary, our results agree fairly well with instrumental or surface faulting evidence available for a limited number of large earthquakes, with typical inaccuracies in the order of 5-10 km for the source location, 0.2-0.3 moment magnitude units for the source size, and 10°-15° for the source orientation. A limitation of the approach is represented by its inability to filter out possible distortions of the macroseismic field associated with source directivity or extreme source complexity. This characteristic, however, could eventually be turned to our advantage for exploring the dynamic properties of the source if independent information on the rupture timing and propagation direction becomes available. Nevertheless, the examination of the information available for modern earthquakes shows that in several cases the inaccuracy of our intensity-based estimates is comparable with the uncertainties in instrumental determinations.

7. Tectonic constraints and implications

Although this work was not expressly intended to contribute to the understanding of recent Italian geodynamics, we feel that a discussion of our modeling results in the framework of the general seismotectonic context of this region may help assessing the ability of the algorithm to evaluate the true location, extent and orientation of major seismogenic sources. At the same time, some of our results may support on a more quantitative basis some of the current ideas concerning the Apennines seismicity.

A general conclusion from a simple visual inspection of fig. 5 is that the main earthquake sources of this region align along the crest of the Apennines within a < 50 km-wide corridor, suggesting the existence of a relatively simple yet extremely continuous seismogenic belt. This intriguing circumstance was first pointed out exactly 150 years ago by Perrey (1848) based on a qualitative examination of intensity data, and has later become the basis for the development of modern earthquake recurrence models for the region (*e.g.*, Valensise *et al.*, 1993). Notable exceptions are represented by the 1627, 1646, 1731, 1881, 1943, 1948 earthquakes, which following Frepoli and Amato (1997, 2000) could be interpreted as the manifestation of the existence of an active compressive belt rather well separated from the main active extensional belt straddling the crest of the Apennines.

A subsequent observation is that there appears to be limited overlap between adjacent sources. This condition supports the earlier assumption that our data set of 42 large historical earthquakes is representative of as many individual sources belonging to a segmented belt. In conjunction with additional tectonic and instrumental evidence, this circumstance may form the basis for a systematic search for potential gaps in historical seismic release throughout the investigated region.

The combination of a mildly heterogeneous tectonic regime, some scatter in the input data and some instability in the processing algorithm could indeed be reflected in a tendency for the investigated sources to exhibit a rather scattered orientation. Quite surprisingly, no such tendency appears from the results shown in fig. 5; on the contrary, most of the sources seem to align in a rather orderly fashion along the trend of the Apennines. In particular, while the main sources of the Southern Apennines all trend between N40°W and N60°W, the sources inferred for the two largest shocks of the 1703 sequence (1703a and 1703b in fig. 5) seem to testify a known transition from the N70°W-trending Northern Abrutii tectonic structures to the decidedly more north-south trend of the Umbria-Marche Apennines (see, for example, Cello *et al.*, 1997). The only significant departures from the general trend concern smaller-size, less constrained earthquakes such as the 1639, 1646, 1853, 1873, 1881, 1904, 1933, 1948. Following the interpretation proposed by Valensise *et al.* (1993), at least some of these earthquakes (particularly those closer to the axis of the Apennines) might reflect the activity of known *transverse* tectonic lineaments (that is, perpendicular to the main trend of the Apennines) which are known to predate the onset of the present stress regime.

Overall, our modeling results are compatible with the general notion that the central and southern portions of peninsular Italy are actively extending in a direction perpendicular to the local strike of the Apennines. This circumstance has been qualitatively known for some time based on conventional geological evidence (*e.g.*, Scandone (1983), but it has been recently demonstrated to hold also for present-day tectonics by Valensise *et al.* (1993) and Amato and Montone (1997), respectively based on the analysis of the largest historical earthquakes comprising the Central and Southern Apennines segmented seismogenic belt and on a careful examination of direct indicators of the modern stress field (earthquake focal mechanisms and borehole breakout data). The uniformity of the trend delineated by our intensity-based sources and its consistency with the information supplied by several independent lines of evidence represent an implicit validation of the approach itself.

We recall that our approach relies on the working hypothesis that each analyzed historical earthquake represents the maximum-size event that can be generated by its respective source. If proven, this hypothesis allows each earthquake to be regarded as an individual characteristic source that can be used to construct a fault segmentation model, or to integrate an existing one. The overall arrangement of the inferred source zones suggests only minimal overlap between adjacent sources and highlights «missing» source areas that may be incorporated in current fault segmentation models and should become the locus of more focused investigations in the future.

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