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The April 1996 Irpinia seismic sequence: Evidence for fault interaction

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Abstract

The analysis of the Irpinia earthquake of 3 April 1996 ($M_L = 4.9$), based on strong motion and short period local data, shows that it was a normal faulting event located within the epicentral area of the M_S 6.9, 1980, earthquake. It was located at 40.67° N and 15.42° E at a depth of 8 km. The local magnitude (4.9) has been computed from the VBB stations of the MedNet network. The moment magnitude is $M_w = 5.1$ and the seismic moment estimated from the ground acceleration spectra is 5.0 10^{23} dyne cm. Spectral analysis of the strong motion recordings yields a Brune stress drop of 111 bars and a corner frequency of 1 Hz. The source radius associated to these values of seismic moment and stress drop is 1.3 km. The focal mechanism has two nodal planes having strike 297°, dip 74°, rake 290° and strike 64°, dip 25° and rake 220°, respectively. A fault plane solution with strike 295° ± 5°, dip 70° ± 5°, and rake 280° ± 10° is consistent with the S-wave polarization computed from the strong motion data recorded at Rionero in Vulture. We discuss the geometry and the dimensions of the fault which ruptured during the 1996 mainshock, its location and the aftershock distribution with respect to the rupture history of the 1980 Irpinia earthquake. The distribution of seismicity and the fault geometry of the 1996 earthquake suggest that the region between the two faults that ruptured during the first subevents of the 1980 event cannot be considered as a strong barrier (high strength zone), as it might be thought looking at the source model and at the sequence of historical earthquakes revealed by paleoseismological investigations.

Introduction

In areas where large magnitude earthquakes ($M > 6.0$) have long repeat times, the study of moderate magnitude seismicity ($4.5 < M < 6.0$) plays an important role for the understanding of the seismogenic behavior of active faults. This is true for the Southern Apennines, where the single faults have recurrence times of thousands of years (Pantosti et al., 1993), and large magnitude earthquakes occur along the seismogenic belt few times within a century (Boschi et al., 1995).

The Southern Apennines seismogenic belt consists of different fault segments along which normal faulting earthquakes occur (Valensise et al 1993; Nostro et al., 1997). The best known is the Irpinia fault that ruptured during the 1980 earthquake. This earthquake ruptured several segments, whose geometry has been determined in detail by investigating different geophysical data (Deschamps and King, 1983 and 1984; Westaway and Jackson, 1987; Bernard and Zollo, 1989; Pantosti and Valensise, 1990; Pingue and De Natale 1993; Cocco and Pacor, 1993). Figure 1 shows a map of the Irpinia fault system with the segments that ruptured during the 1980 earthquake, its focal mechanism and the distribution of aftershocks (shown in the map and in the cross-section A–A' in the top right-hand corner) relocated by Amato and Selvaggi (1993). The segment dimensions and seismic moments are summarized in Table 1. According to the literature published, we refer to the three subevents of the 1980 event as 0s, 20s and 40s, respectively. The total seismic moment of this earthquake ranges between 1.3 and 3.0 10^{26} dyne cm (Giardini, 1993).

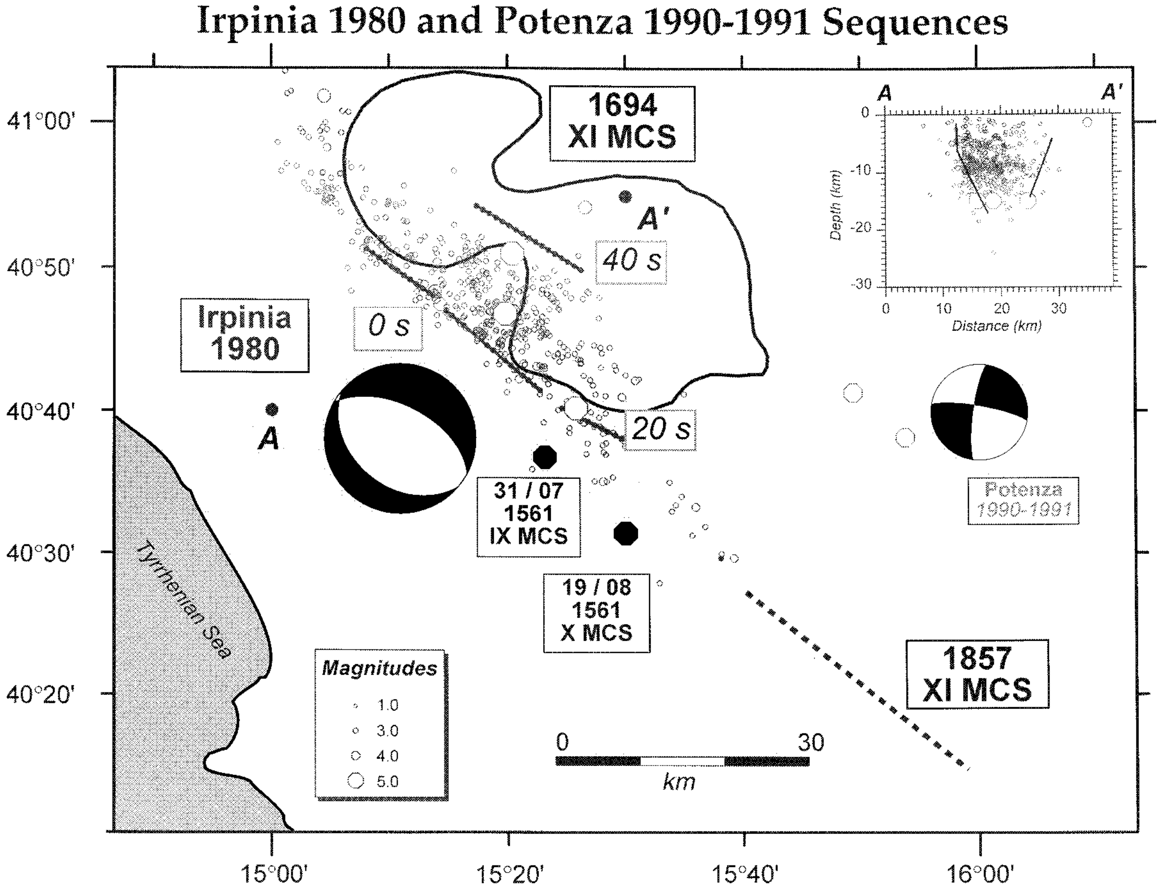

Figure 1. Map of the Irpinia and Lucania fault zones. The three fault segments that ruptured during the 1980 earthquake are named 0s, 20s and 40s, respectively. Aftershock distribution is shown on the map and on the cross-section A–A' on the top right-hand corner. The XI degree MCS isoseismal curves of the 1694 historical earthquake has been drawn. The dotted line indicates the fault segment which is believed to have ruptured during the 1857 earthquake. The locations of the two 1561 moderate magnitude earthquakes (Boschi et al., 1995) are shown on the map. The 1990 and 1991 Potenza mainshocks are plotted on the map (open circles). The figure shows the focal mechanisms of the 1980 Irpinia ($M_S = 6.9$) and of the 1990 ($M_S = 5.4$) and 1991 ($M_S = 4.8$) Potenza earthquakes.

Figure 1 also shows the historical earthquakes that occurred in the study area: the largest magnitude events are the 1694 and the 1857 earthquakes (M 6.9 and 7.0, respectively), while the two 1561 events (both located in Vallo di Diano) have moderate magnitudes (5.5 and 6.4 respectively; Boschi et al., 1995). We have plotted in Figure 1 the X degree MCS isoseismal curve of the 1694 seismic event. There are two important considerations associated with the historical earthquakes in this area. First, the area struck by the 1561 earthquakes has been highly loaded by the stress redistribution after the 1980 event (Nostro et al., 1997). The 1980 Irpinia earthquake increased the static stress of nearly 1 bar in the area where the 1561 moderate earthquakes occurred (Nostro et al., 1997). According to these results the Vallo di Diano area is a candidate site for a $M \sim 6$ earthquake in the next future. Second, according to paleoseismological investigations, the 1694 earthquake did not rupture the 0s and 20s fault segments of the 1980 event (Pantosti et al., 1993). The shape and the position of the 1694 isoseismal curve, point out that this event struck the same area of the 1980 earthquake and suggest that different active faults should be present in the area. These observations led Nostro et al. (1997) to suggest that the 1694 earthquake might have ruptured a fault segment, to which the 40s fault might belong, antithetic to the 0s and 20s faults. According to this interpretation and to the static stress changes caused by the rupture of the first two subevents, Nostro et al. (1997) suggested that the 40s subevent could be considered as a reactivation of a pre-existing fault, in agreement with the

Table 1. Fault geometry and seismic moments of the 1980 Irpinia earthquake

Subevents	Authors	$M_o \times 10^{26}$ *dyne cm*	Mw	fault length L (km)	fault width W (km)
(1) 0s	B&Z	0.90	6.6	25	15
	P&V	1.24	6.6	28	15
(2) 20s	B&Z	0.40	6.4	20	15
	P&V	0.31	6.3	10	15
(1 + 2)	B&Z	1.3	6.7		
0s + 20s	P&V	1.55	6.8		
	RBCDM	2.3	6.9		
(3) 40s	B&Z	0.3	6.3	15	15
	P&V	0.24	6.2	15	15
	RBCDM	0.40	6.4		
Total	B&Z	1.6	6.8		
0s + 20s + 40s	P&V	1.8	6.8		
	RBCDM	2.7	6.9		

B&Z = Bernard and Zollo (1989)
P&V = Pantosti and Valensise (1990)
RBCDM = Rovelli et al. (1988).

findings of Amato et al. (1992) and Chiarabba and Amato (1994).

The last moderate magnitude seismic sequence in the study area was located near Potenza during 1990 and 1991 (see Figure 1). The hypocenters of the Potenza seismic sequences are deeper than that of the 1980 Irpinia earthquakes (Azzara et al., 1993) and their mainshocks have both a strike slip focal mechanism (Ekstrom, 1994). These fault plane solutions are consistent with the direction of regional extension for this area as resulting from borehole breakout data (Amato et al., 1995).

Paleoseismological studies performed by Pantosti et al. (1993) have shown that the 0s and the 20s segments have almost the same temporal sequence of past earthquakes. This implies that these two fault segments have a quite similar seismic cycle. Despite the errors in dating paleo-earthquakes, it might be speculated that the 0s and the 20s segments ruptured together during large magnitude earthquakes. In this case, the area comprised between the two segments can be considered as a permanent dynamic feature during the seismic cycle of this active fault.

In this paper, we present and discuss the 3 April 1996 seismic sequence, that occurred within the epicentral area of the 1980 earthquake. The goal is to understand the spatial and temporal pattern of seismicity along the best known fault segment of the Southern Apennine seismogenic belt, taking into account the fault geometry and the state of stress of the area.

The 1996 mainshock

Earthquake location. The main shock occurred on 3 April 1996, at 13:04:34.99 GMT; the local magnitude (M_L 4.9) has been computed from broad-band recordings at MedNet stations (S. Mazza and N. A. Pino personal communication). It was recorded by the stations of the Italian permanent seismic network of the Istituto Nazionale di Geofisica (I.N.G.), and of the permanent seismic network of the Osservatorio Vesuviano (O.V.), both equipped with vertical component short period seismometers, and by several digital three-component seismic stations (Guralp CMG40T or Mark 1s) deployed near Potenza by the I.N.G., the University of Grenoble and the University of Liege during a cooperative monitoring experiment. It was also recorded by a broad-band seismic station (Guralp CMG3) deployed on the Vesuvius volcano by the I.N.G. during a cooperative experiment with the University of Nice, the University of Naples and the O.V. for tomographic studies of the Vesuvius area. Most of the aftershocks were also recorded at the digital stations (equipped with LE/3D 5s) deployed by the I.N.G. during a teleseismic transect experiment (Amato et al., 1998). Using all the available data, the mainshock hypocenter has been located at 40.67 N and 15.42 E at a depth of 8 km (see Figure 2); the error is 0.5 km for the horizontal coordinates, and 1.0 km for depth.

The crustal structure in the Irpinia area is well known, thanks to tomographic studies performed us-

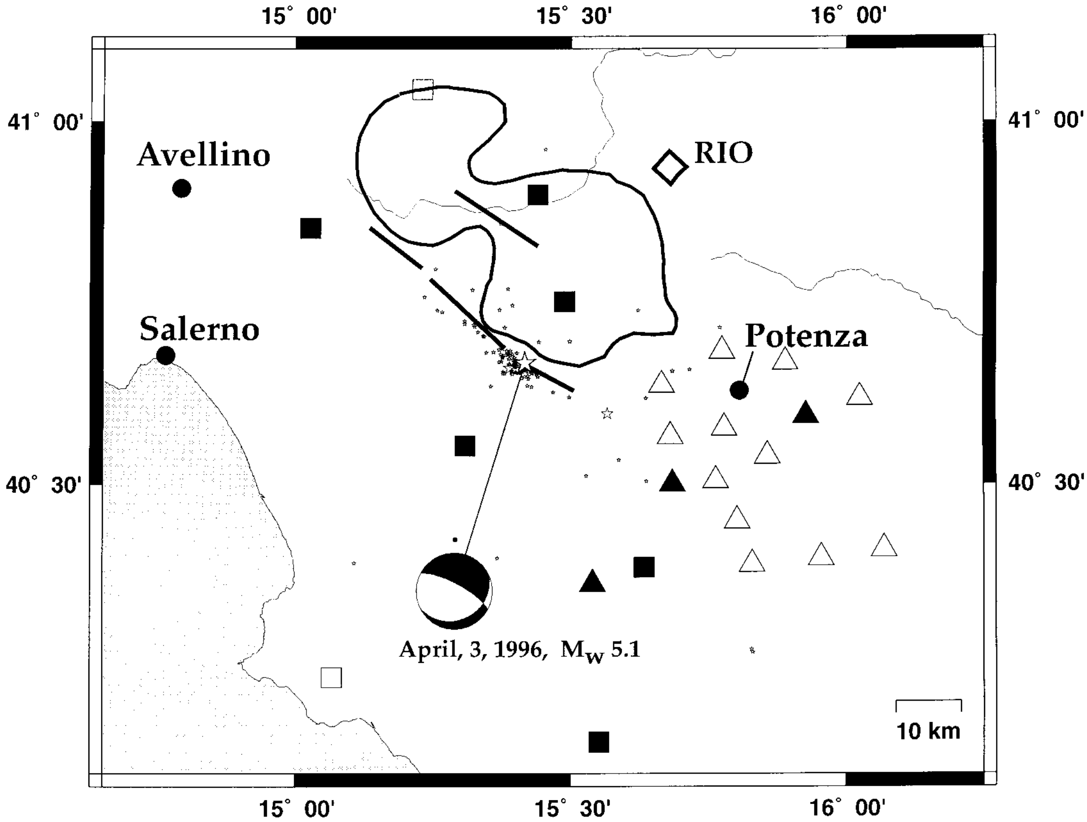

Figure 2. Distribution of seismic stations closest to the epicenter of the 3 April 1996 Irpinia earthquake (indicated by a star). The solid and the open squares indicate the permanent seismic stations of the ING National network and of the OV Regional network, respectively. The solid triangles show the stations of the temporary teleseismic transect experiment deployed by ING (Geomodap EC project); while the open triangles indicate the stations of a temporary local network deployed near Potenza by the ING, the University of Grenoble and the University of Liege. The open diamond indicates the position of the accelerograph located in Rionero in Vulture (RIO). The focal mechanism, the epicenter and the aftershock distribution of the 1996 earthquake are shown. We have also drawn the surface projections of the three segments that ruptured during the 1980 earthquake.

ing the recordings of the 1980 aftershock sequence (Amato and Selvaggi, 1993; Amato et al., 1992). In this study we use a 1-D velocity model obtained by Amato and Selvaggi (1993) and the code hypoinverse (Klein, 1989) to locate mainshock and aftershocks. Figure 2 shows the distribution of seismic stations used to locate the mainshock, its location and the fault segments that ruptured during the 1980 earthquake. It emerges that the 1996 earthquake occurred within the epicentral area of the 1980 event and very close to the 20s segment.

Source Parameters. The 1996 Irpinia M_L 4.9 mainshock triggered several strong motion accelerometers belonging to the ENEL strong motion network. Most of them are mechanical SMA1 Kinemetrics accelerometers; only the accelerograph located at Rionero in Vulture (RIO, see Figure 2) is a digital SSA1 Kinemetrics (16 bit) instrument and it was triggered by P-waves. Figure 3 shows the time history of ground acceleration recorded at RIO 36 km far away from the epicenter. Peak ground acceleration is 12.5 gals recorded on the EW component; for this reason, most of the mechanical accelerographs (that have a triggering threshold of 10 gals on the vertical component) in a similar distance range were not triggered by the main shock.

In this study, we use the strong motion data recorded at RIO to compute the seismic moment and the corner frequency using both a frequency and a time domain analysis. We have integrated the ground acceleration to get ground displacement. Figure 4 shows the horizontal ground displacement recorded at RIO. The duration of the displacement pulse is about 1s. We

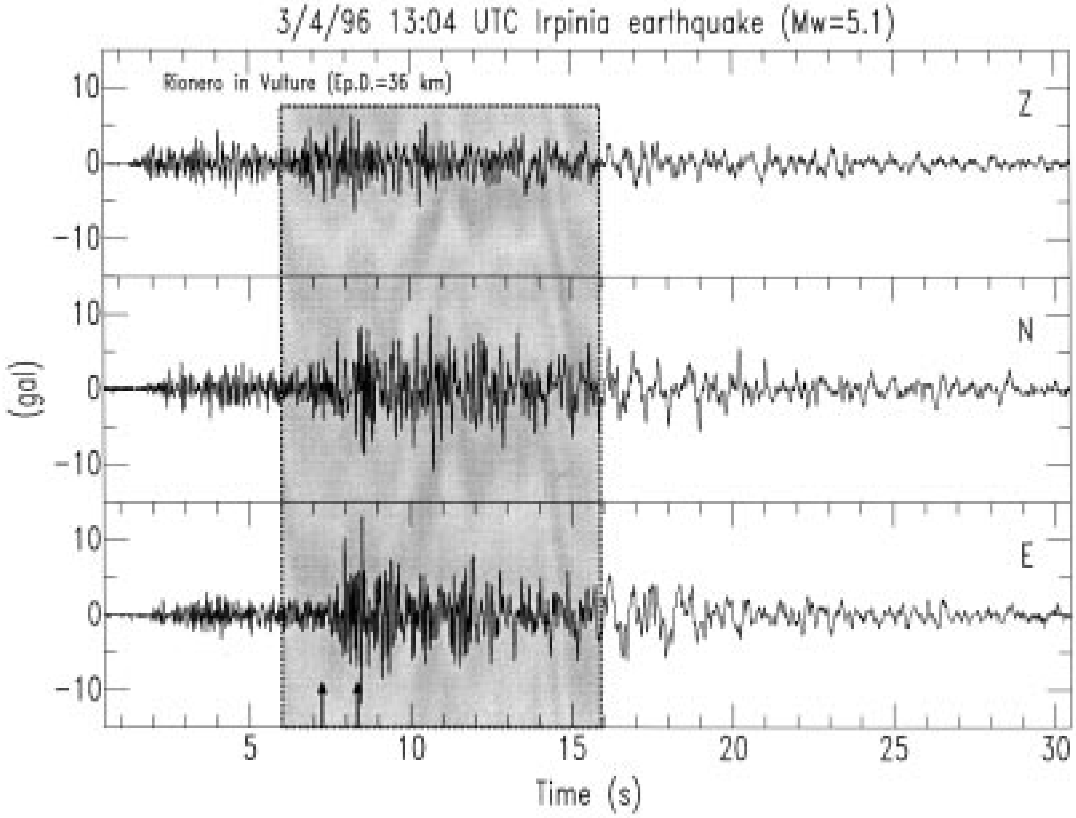

Figure 3. Acceleration time history recorded during the April 3, 1996, Irpinia mainshock at Rionero in Vulture (RIO) located 36 km from the epicenter (see Figure 2). The shaded area shows the time window used to compute the acceleration spectrum. The two arrows indicate the S-wave window used to compute the seismic moment (see Figure 4).

have computed the seismic moment from the area of the displacement pulse (indicated by the shaded area in Figure 4) following Brune (1970) using the equation (Frankel, 1981) $M_o = R\Omega_o/C$, where R is the hypocentral distance and C is a constant given by

$$C = \frac{F_s R_{\vartheta\varphi} P}{4\pi\rho\beta^3}, \tag{1}$$

which includes the effects of free surface correction ($F_s = 2$), the average *rms* radiation pattern ($R_{\vartheta\varphi} = 0.63$) and the energy partition between the horizontal components ($P = 1/\sqrt{2}$); ρ and β are the density and the S-wave velocity in the crust. According to Boore and Boatwright (1984) the *rms* radiation pattern coefficient, averaged over the whole focal sphere (0.63), does not considerably differ from the effective radiation coefficients for a dip-slip fault averaged over the appropriate distance range ($0.67 \div 0.70$). The resulting value is $M_o = 2.3\ 10^{23}$ dyne cm. We also computed the seismic moment in the frequency domain by fitting the spectra of ground acceleration $A(f)$ recorded at RIO, assuming an omega-squared model $S(f, M_o, f_c)$ for the source radiation (see Rovelli et al., 1988 and 1991, and Cocco and Rovelli, 1989, for further details)

$$A(f) = \frac{1}{R} S(f, M_o, f_c) e^{-\pi k f} e^{-\frac{\pi R}{\beta Q_o}}, \tag{2}$$

where $S(f, M_o, f_c)$ is defined as follows:

$$S(f, M_o, f_c) = C M_o \frac{(2\pi f)^2}{1 + \left(\frac{f}{f_c}\right)^2}, \tag{3}$$

where C is defined by Equation (1) and f_c is the corner frequency. Equation (2) also includes the attenuation contribution. The values of the attenuation parameters k and Q_o for RIO have been taken from previous studies (Rovelli et al., 1988; Cocco and Rovelli, 1989); they result 0.06 sec and 100, respectively. The station estimates for the k parameter are very close to the average value obtained for all the Apenninic earthquakes by Rovelli et al. (1988). The corner frequency has been estimated using the approach proposed by Andrews (1986) and currently applied by Rovelli et al. (1988

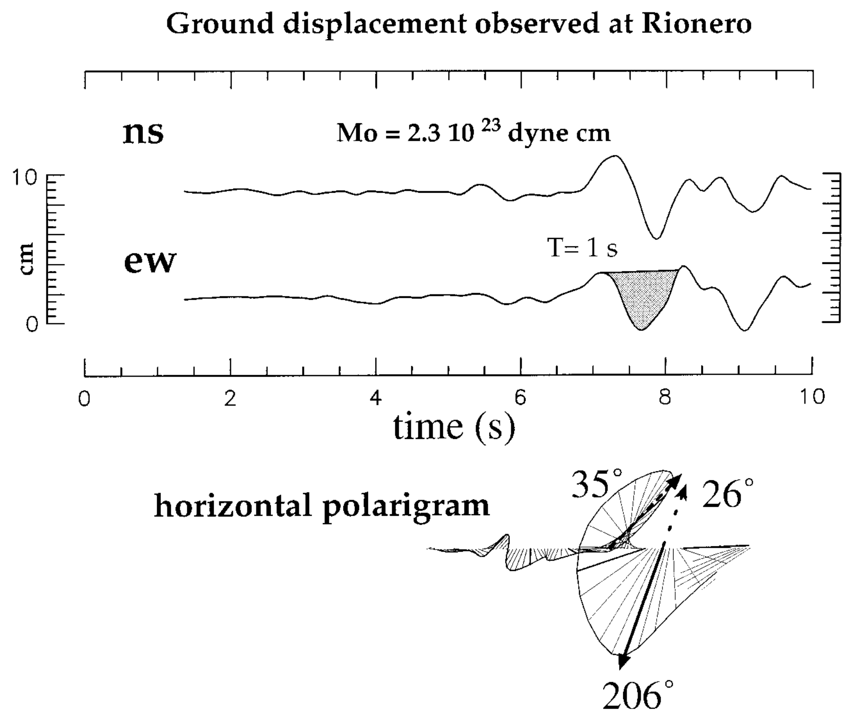

Figure 4. Horizontal components of the displacement time history obtained integrating the ground acceleration recorded at Rionero in Vulture (RIO). The shaded area indicates the displacement pulse where the seismic moment has been computed. The duration of the displacement pulse T is 1 s and the resulting seismic moment is 2.3 10^{23} dyne cm. The horizontal polarigram of ground displacement is shown. The computed S-wave polarization is 26° ± 10° (see the arrows).

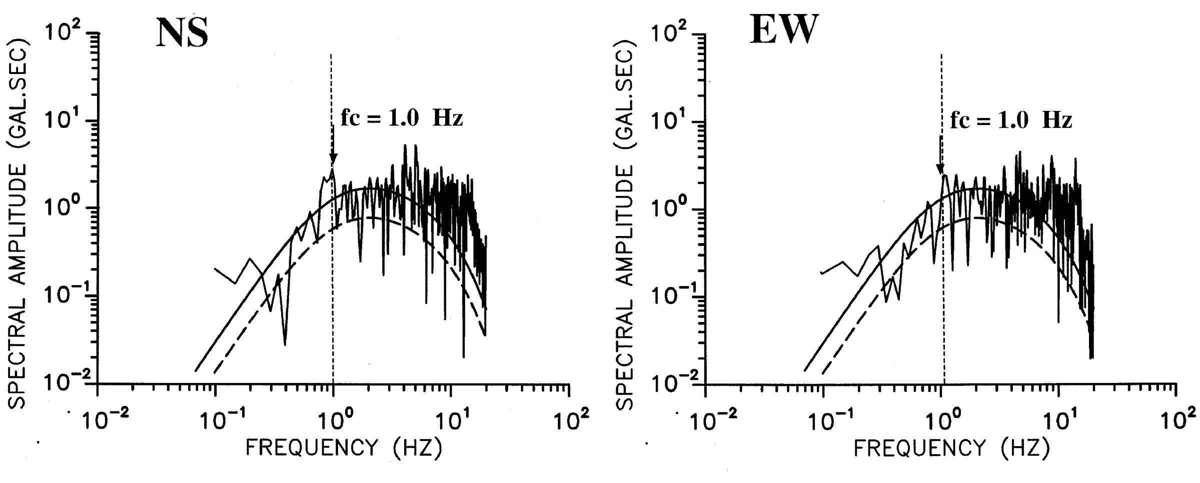

Figure 5. Fourier spectra of the horizontal components of ground acceleration recorded at RIO. The solid line indicates the theoretical curve computed for an omega squared Brune spectral model using a corner frequency of 1 Hz and a seismic moment of 5.0 10^{23} dyne cm, estimated in the frequency domain. The dashed line shows the theoretical curve using a seismic moment of 2.3 10^{23} dyne cm, estimated in time domain. The attenuation parameters used in the theoretical spectral model are given in the text.

and 1991) and by Cocco and Rovelli (1989); the resulting value is 0.95 Hz in good agreement with the time domain estimate. The seismic moment estimated by fitting the acceleration spectra is $M_o = 5.0\ 10^{23}$ dyne cm. Figure 5 shows the acceleration spectra of horizontal components fitted by the theoretical model defined by (2) and (3) using the values for k and Q_o given above, and the frequency domain estimates for f_c and M_o (solid curves); for comparison, the fit to the recorded ground acceleration spectra resulting from the time domain estimates of f_c and M_o is shown in Figure 5 (dashed curves). A value of seismic moment of 5.0 10^{23} dyne cm provides a better fit of acceleration spectra. In particular, the fit is quite good up to 4 Hz, while at frequencies higher than 5 Hz there are evident amplifications due to the local geological conditions of the recording site.

The difference between time and frequency domain estimates of seismic moment found in this study (roughly a factor of 2) is consistent with the results obtained by Rovelli et al. (1991), who discuss the problems related to the different estimates of source parameters from strong motion accelerograms. Those authors also found a similarity between the corner frequencies estimated in time and frequency domain for earthquakes having magnitudes close to the size of the Irpinia 1996 event analyzed in this study.

The moment magnitude (Hanks and Kanamori, 1982) associated to the frequency domain estimate of seismic moment is M_w 5.1. The local magnitude computed from synthetic Wood-Anderson seismograms obtained from the strong motion accelerogram (Kanamori and Jennings, 1978; Bonamassa and Rovelli, 1986; Di Bona et al., 1995) is 5.4.

Using the frequency domain estimates of corner frequency (1.0 Hz) and seismic moment (5.0 10^{23} dyne cm), which better fit the acceleration spectra (see Figure 5), we have computed the Brune stress drop according to the following relation:

$$\Delta\sigma = \frac{M_o f_c^3}{(4.9\,10^6 \beta)^3}.$$

The resulting Brune stress drop is 111 bars, which is in agreement with the scaling law between seismic moment and corner frequency found by Rovelli et al. (1988) for the southern Apennines earthquakes (see Figure 9 in that paper). The source radius can be computed by means of the following equation:

$$r = 2.34 \frac{\beta}{2\pi \cdot f_c},$$

which yields a source radius of 1.3 km.

Figure 2 shows that the 1996 Irpinia mainshock was located between the two fault fragments that ruptured during the 0s and 20s subevents of the 1980 earthquake. The size of the surface faulting gap observed during the 1980 event is roughly 3 km (see Figure 1, and Pantosti and Valensise, 1990). It is important to point out that the source dimension (the diameter of a circular fault is 2.6 km) of the 1996 earthquake is close to the spatial separation of the first two subevents of the 1980 mainshock. In other words, it is interesting to discuss the dimension and the geometry of the fault and slip direction during the 1996 event with respect to the rupture history of the 1980 earthquake.

Focal Mechanism. We have determined the fault plane solution of the mainshock from the first motion P-wave polarities observed at the stations of the permanent networks of I.N.G. and O.V. and at those of the temporary network near Potenza using the code FPFIT (Reasemberg and Oppenheimer, 1985). The resulting focal mechanism is shown in Figure 6 (and included in Figure 2). The two nodal planes resulting from this analysis are: strike 297°, dip 74° and rake 290° and strike 64°, dip 25° and rake 220°.

We have analyzed the S-wave polarization resulting from the accelerograms recorded at Rionero in Vulture (RIO) in order to verify the focal mechanism computed from P-wave polarities. We have calculated the S-wave polarization on both velocity and displacement recordings. Figure 4 shows the horizontal polarigrams resulting from the ground displacement waveforms. The observed S-wave polarization at RIO is $26° \pm 10°$. A fault plane solution having strike 297°, dip 70° and rake 280°, yields a theoretical S-wave polarization of 32°; while a pure normal faulting event (rake is 270°) yields a theoretical S-wave polarization of 23°. Thus, we conclude that the S-wave polarization observed at RIO is consistent with the focal mechanism resulting from first motion polarities.

The seismic sequence

The 1996 earthquake is the largest seismic event since 1980 located within the epicentral area of the M_S 6.9 Irpinia earthquake. The magnitude of the 1996 mainshock is similar to the magnitudes of the largest aftershocks of the whole 1980 seismic sequence (Deschamps and King, 1984). The 1996 Irpinia earthquake was preceded by few microseismic events with

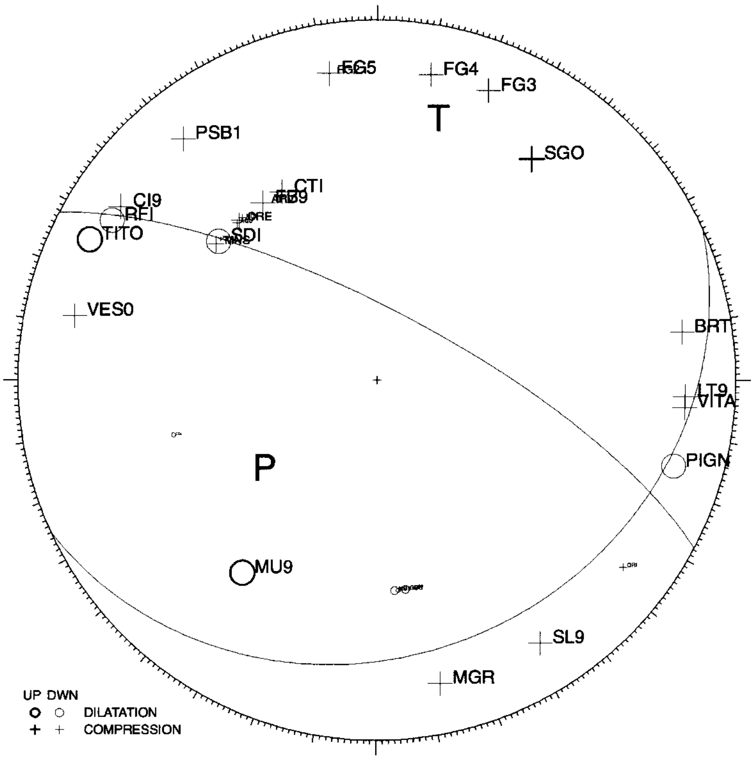

Figure 6. Fault plane solution computed using P-wave polarity data. The two nodal planes are: $\phi = 297°$, $\delta = 74°$, $\lambda = 290°$ and $\phi = 64°$, $\delta = 25°$ $\lambda = 220°$, where ϕ, δ and λ are strike, dip and rake, respectively. The dimension of the symbols scales with the weight assigned to each reading.

magnitudes less than 2.2. The largest aftershock of 3 April 1996 at 14:15 GMT, had a magnitude of 3.5. Figure 7 shows the temporal distribution of seismicity during the 1996 sequence and the maximum magnitude per day. The temporal pattern of seismicity is characterized by few periods of more intense seismic activity. In particular, on July 17 there was a sharp increase of activity with several $M \sim 3$ aftershocks. The whole sequence lasted until the end of August 1996.

The aftershocks were located using the data recorded at the permanent stations of the ING and OV networks and, only for the earthquakes that occurred in the period April–June, also using the data recorded at the digital stations deployed during the temporary experiments (near Potenza, along the Apenninic transect and on the Vesuvio Volcano). Mainshock and aftershocks were located using a 1-D velocity model proposed by Amato and Selvaggi (1993). The aftershocks occurred during April–June, 1996, have formal location errors that range between 0.5 and 1.0 km for the horizontal coordinates, and 1.0 km for the depth; for the remaining period the formal location errors are 1.5 km for the horizontal coordinates and 2.0 km for the depth.

In order to verify the dependence of hypocentral determination on velocity model, we performed a numerical test varying the P wave velocity by 10% in each layer and Vp/Vs ratio between 1.7 and 1.9. We relocated the earthquakes that occurred in the period April–June 1996 using all the resulting velocity models. We found that 75% of the events has a horizontal displacement from our preferred 1-D location less than 1 km (that is, of the order of the formal horizontal error). The effect of the crustal model on the hypocentral depths is larger: only 65% of the events shows variation in depth less than 2 km. This sensitivity test allows us to conclude that the earthquake locations are good enough to study the spatial pattern of seismicity during the 1996 seismic sequence, due to the good azimuthal coverage guaranteed by the permanent seismometric stations of the ING and OV networks in the study area (see Figure 2).

The spatial distribution of seismicity during the 1996 sequence is plotted in Figure 8: the open circles

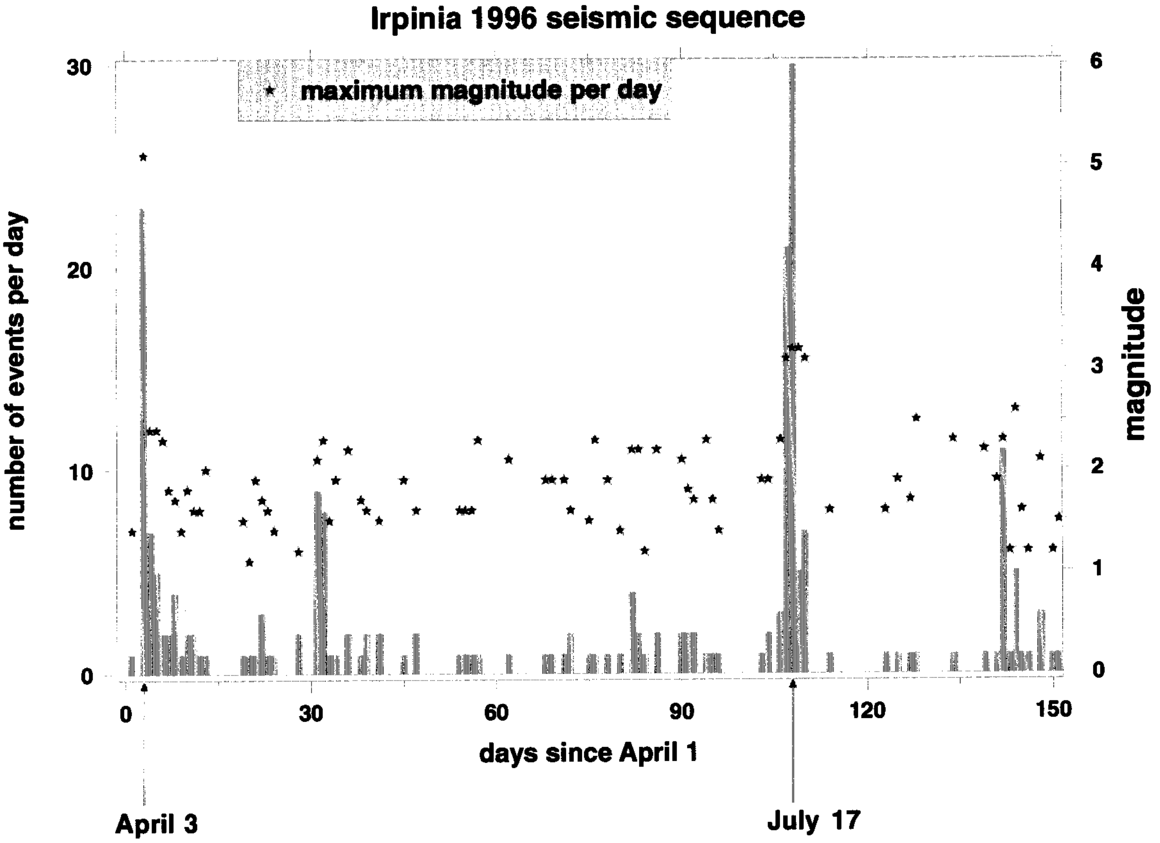

Figure 7. Number of events as a function of time. Stars indicate the maximum magnitude per day.

indicate the aftershocks that occurred between May and June, while the gray circles depict the aftershock distribution between July and September. Figure 8 shows that the 1996 seismic sequence was located at the southern end of the 0s segment of the 1980 earthquake and it extends on the 20s fault segment, in the area where Pantosti and Valensise (1990) observed a gap in surface faulting. The two cross sections show the distribution of hypocenters along and perpendicularly to the Apenninic chain: most of aftershocks cluster at depths shallower than 12 km. Aftershock distribution does not allow to constrain the geometry of the fault plane at depth. However, the fault plane solution and the distribution of seismicity on a SW–NE cross section better agree with the geometry of the 20s subevent proposed by Pantosti and Valensise (1990; see also Nostro et al., 1997, and references therein).

Discussion and conclusive remarks

The 1996 mainshock was located in the area which separates the two first subevents of the 1980 Irpinia earthquake. The focal mechanism shows a normal faulting on a plane striking 297° and dipping 74° NE. This faulting mechanism is consistent with the Coulomb stress changes caused by the 1980 earthquake as pointed out by Nostro et al. (1997). The mainshock fault plane solution and the aftershock distribution of the 1996 sequence provide further evidence to reject the hypothesis that the 20s subevent ruptured a shallow angle fault (see Nostro et al., 1997 and reference therein). The fault dimension of the 1996 mainshock ($M_o = 5.0\ 10^{23}$ dyne cm) is similar to the size of the surface faulting gap observed by Pantosti and Valensise (1990) between the two first subevents of the 1980 earthquake. Figure 9 allows us to compare the distribution of hypocenters during the 1996 sequence with the surface displacement and the aftershock distribution during the 1980 earthquake. The aftershock distribution during the 1980 sequence

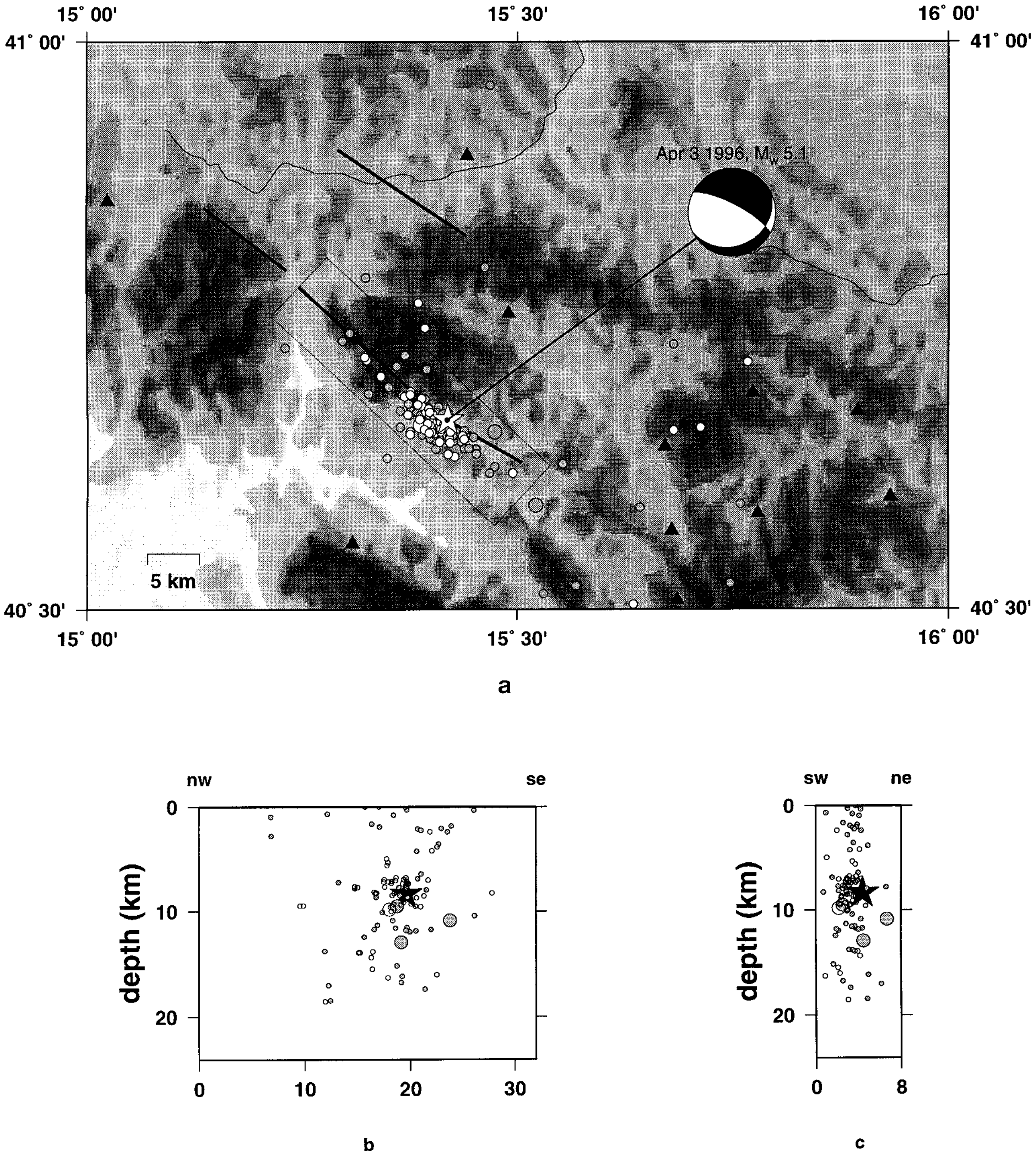

Figure 8. Distribution of seismicity during the 1996 sequence. The topography is shown on the map. The open circles indicate the aftershocks that occurred between April and June 1996, while the gray circles show the aftershock distribution between July and September. The star indicates the mainshock location. The four fault fragments which ruptured during the 1980 event have been drawn. Triangles show some of the seismic stations which have been used to locate earthquakes. The fault plane solution of the 1996 mainshock is shown. The box in the map indicates the zone used to select the data for the two cross-sections shown in the bottom panels ((b) is an Apenninic cross section, and (c) is taken perpendicular to the chain).

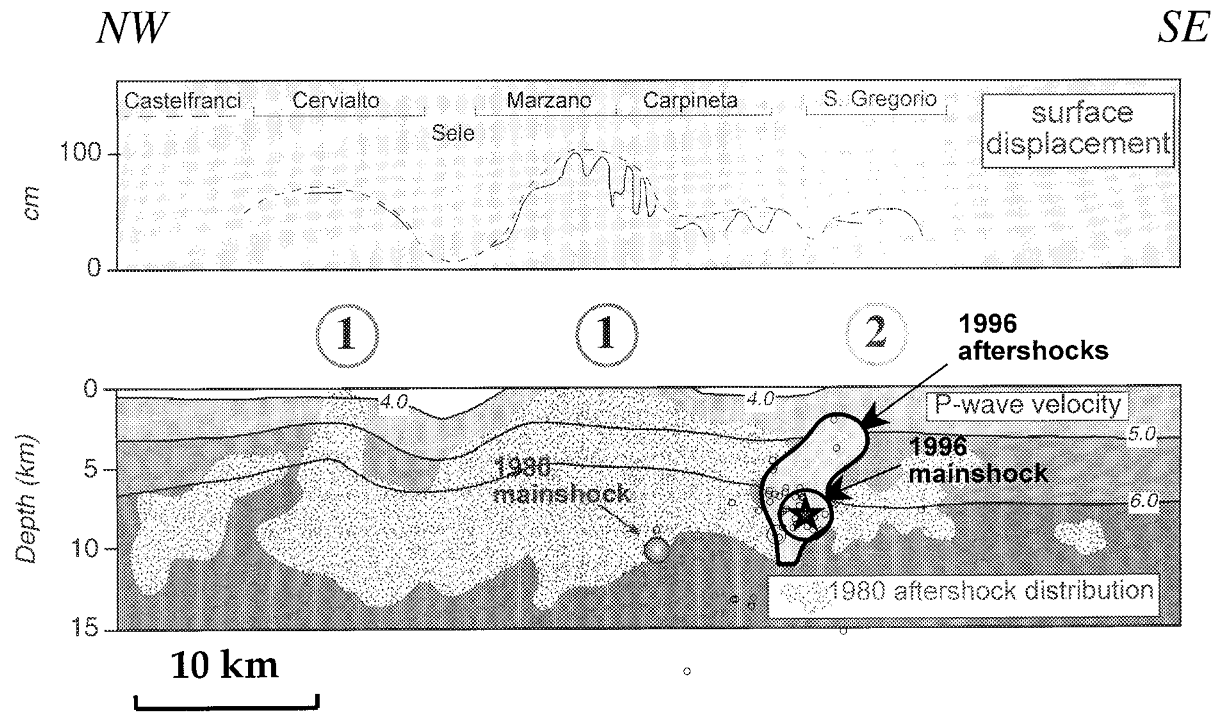

Figure 9. Comparison between the aftershock distributions (shaded areas) and the mainshock locations of the 1980 and 1996 seismic sequences along a NW–SE section (P-wave velocities are shown in the lower panel, from Selvaggi and Amato 1993). Surface displacement (upper panel) is taken from Pantosti and Valensise (1990). All the aftershocks of the 1996 sequence occurred within the solid curve draw in the lower section. The earthquake selection criterion is the same as in Figure 8. The solid circle indicates the fault area that ruptured during the 1996 mainshock. The 1996 mainshock is located at the southern edge of the 1980 first subevent; the 1996 sequence has been located in a area where few aftershocks occurred during the 1980 sequence and that did not sustain surface breakage.

is shown by the shaded area, while the circle indicates the area which ruptured during the 1996 mainshock. This figure points out that the 1996 seismic sequence occurred in an area where few aftershocks of the 1980 earthquake were located and which coincides with the gap in surface faulting.

Based on the source model proposed by Bernard and Zollo (1989) and by Pantosti and Valensise (1990), the area comprised between the two first subevents can be thought as a barrier which spatially separates the two segments of the Irpinia fault (see Figure 1 and 8). Paleoseismological studies have shown that the 0s and the 20s subevents of the 1980 earthquake experienced almost the same earthquakes during the past 10.000 years (Pantosti et al., 1993). Even if the errors in dating the paleo-earthquakes are of the order of 100 years, it might be possible to speculate that this barrier is a permanent dynamic feature (high strength zone) of the Irpinia fault. That is, the Irpinia rupture zone consists of several well known faults which rupture together during repeated earthquakes. The occurrence of the 1996 $M_w = 5.1$ earthquake and its sequence of aftershocks makes this hypothesis difficult to be sustained. In fact, we should not expect any moderate magnitude earthquake as well as microearthquakes in an area with high strength (as a barrier; Scholz, 1990).

The slip heterogeneity (found by Cocco and Pacor, 1993) and the delayed subevent triggering observed during the 1980 earthquake (Westaway and Jackson, 1987) can be interpreted in terms of lateral variation of frictional properties (Boatwright and Cocco, 1996). Accordingly, the area located south of the first subevent might be considered a weak zone (following the definition given by Boatwright and Cocco, 1996), which arrested the rupture propagation during the first rupture episode (0s subevent), delayed the rupture nucleation of the 20s subevent and where both aftershocks and interevent seismicity can occur because it is a velocity weakening zone (see Boatwright and Cocco, 1996). It is expected that a velocity strengthening zone, even when its frictional behavior is nearly neutral, can have aftershocks but not interseismic earthquakes. This, because such velocity strengthening zones can be forced to dynamic instability by the dynamic stress redistribution associated with large earthquakes, but its response to a nearly constant tectonic load is always stable.

Belardinelli et al. (1998) investigated the dynamic stress changes caused by the rupture of the first subevent of the 1980 Irpinia earthquake. They found that after 8.75s the dynamic stress on the 20s fault plane reached the peak value, but the subevent nucleated nearly 10 s later, when the dynamic stress had already reached the static configuration. These authors interpreted their results suggesting that the area located south of the main (first) subevent of the Irpinia earthquake is a weak zone; for this reason, the nucleation of the 20s subevent needed this long time interval to accelerate to a dynamic instability. The occurrence of the 1996 sequence between the 0s and 20s faults is consistent with the hypothesis of the weak zone and with the Belardinelli et al. (1997) findings. Therefore, this sequence can be considered as interevent seismicity. We emphasize that lateral variation of fault friction affects the phenomenology of earthquake faulting.

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