



Seismic hazard assessment at Mt. Vesuvius: Maximum expected magnitude

E. Cubellis^a, G. Luongo^b, A. Marturano^{a,*}

^a *Istituto Nazionale di Geofisica e Vulcanologia-Osservatorio Vesuviano, Napoli, Italy*

^b *Dipartimento di Scienze della Terra, Università Federico II, Napoli, Italy*

Received 3 January 2006; received in revised form 17 January 2007; accepted 1 March 2007

Available online 21 March 2007

Abstract

This paper deals with the problem of seismicity at Mt. Vesuvius with a view to providing an estimation of the maximum expected earthquake. Integrated analysis of both historical and current seismicity as well as the geological conditions of Vesuvius and the surrounding areas show that seismogenetic structures may fall within the crater axis and at the boundaries of the volcanic complex. While activation of the whole seismogenetic volume detected by seismicity in the past 30 years would indicate a total seismic moment of $M_0 = 7.1 \times 10^{15}$ Nm for a magnitude $M = 4.5$, knowledge of the area's geological structure suggests faulting surfaces of about 32 km² with an associated magnitude of $M = 5.4$. The areas of maximum expected damage differ according to the orientation of the hypothesized structure. Analysis of geological and geophysical data and the damage associated to the AD 62 earthquake shows that the prevailing directions in the faulting planes are NE–SW in the eastern sector of the volcanic complex, and roughly WNW–ESE in the southern part of the volcano along the coast.

Comparison of instrumental seismicity and historical data reveals two significantly different energy levels: a lower earthquake level with $M_{\max} = 4.5$, corresponding to current seismicity and that which accompanied volcanic activity in the eruptive period from 1631–1944; an upper level with $M_{\max} = 5.4$, represented by the AD 62 earthquake. The two levels correspond to two stress states and different seismogenetic structures.

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Keywords: Mt. Vesuvius; seismic hazard; historical seismicity

1. Introduction

To evaluate the maximum earthquake expected at Vesuvius, data are required on the location, source mechanism and damage levels of historical earthquakes, in addition to understanding how the volcano works. From the historical data there is no clear evidence for events with magnitudes greater than 4.0–4.5 except the earthquake which occurred in 62 AD, 17 years before

the 79 AD plinian eruption which completely buried the towns of Pompeii, Herculaneum, Oplontis and Stabiae. It is, however, not yet clear whether this event was located in the magma feeding shallow structure or rather along an active fault at the boundary of the volcano (Sigurdsson et al., 1985; AA.VV., 1995; Marturano and Rinaldis, 1998).

At Vesuvius the risk associated to a large eruption is very high since 600,000 people live on the volcano and a further 2,000,000 are located around it, but complete assessment means including the potential damage due to earthquakes accompanying the largest eruptions.

* Corresponding author.

E-mail address: marturano@ov.ingv.it (A. Marturano).

Moreover low-moderate energy earthquakes are also observed in volcanically active areas during quiescent periods. Generally such events are shallow and produce high intensities in the epicentral area. Today at Vesuvius the high housing density and economic value exposed make the area of considerable importance for mitigating seismic risk (Cherubini et al., 2001).

In volcanoes the stress field generating seismicity is made even more complex than in tectonic areas because, in addition to the regional stress field, there is that generated by the migration of magmatic masses towards the surface, as well as the rheological properties of the rocks conditioned by the high geothermal gradient.

In the present paper the data set of historical seismicity is built up using reliable sources from classical and medieval times and instrumental seismicity which has been available since the 1944 eruption. During this last time span the most reliable data have been available since 1970 when a modern seismic network was installed for monitoring the seismicity of Vesuvius. Such data will be correlated to the structural framework of the volcano basement to define the suspected seismogenetic sources and evaluate the maximum earthquake expected at Mt. Vesuvius.

2. Volcanic history and seismicity

Eruptive activity in the Vesuvius area dates back about 400 ka although the volcano edifice formed only in the last 25 ka. The recent crater was built inside an older stratovolcano, Monte Somma, and since 1944 has been quiescent (Santacroce, 1987; Brocchini et al., 2001; Santacroce and Sbrana, 2003). The earliest eruptive history of Mt. Vesuvius has been reconstructed on the basis of stratigraphic and radiometric studies of volcanic products and palaeosoils, whereas its later history, which begins with the eruption of 79 AD, has been reconstructed from the direct descriptions of eruptive events as well. The volcano produced at least five large-scale plinian eruptions ($18,300 \pm 180$ years BP, $16,130 \pm 110$ years BP, 8010 ± 35 years BP, 3760 ± 70 years BP, and 79 AD) and smaller-scale subplinian eruptions (Santacroce and Sbrana, 2003). After the Avellino eruption (3760 ± 70 years BP) the volcano continued with subplinian and strombolian eruptions for a few centuries (Rolandi et al., 1998; Santacroce and Sbrana, 2003), and then became quiescent for several hundred years until the eruption in AD 79 (Sigurdsson et al., 1985; AA.VV., 1995; Marturano and Rinaldis, 1998; Luongo et al., 2003; Marturano, 2006).

The activity of Mt. Vesuvius during the early centuries of the first millennium and Middle Ages is not very well known. After the subplinian eruption of 472

Vesuvius was active until the twelfth century (Arnò et al., 1987; Figliuolo and Marturano, 1995; Mastroloranzo et al., 2002; Rolandi et al., 2004). There then followed a long period during which no significant eruptive events took place until 1631 when the volcano resumed its activity with a subplinian eruption (Rolandi et al., 1993; Rosi et al., 1993). From 1631 to 1944 the volcano produced 18 eruptive periods of small and medium-sized eruptions from both terminal and side vents (Santacroce and Sbrana, 2003) (Fig. 1).

The oldest seismic events recorded by Suetonius (Tiberius 74,2) in the Vesuvian area occurred in AD 37 when, a few days before the death of the emperor Tiberius, an earthquake brought down the lighthouse on Capri. On 5 February in AD 62 a strong earthquake hit Campania, causing considerable damage to Pompeii and Herculaneum and relatively minor damage to Naples and Nuceria (Tacitus, Ann. XV.22.1; Seneca, NQ VI. 1.1–3), with maximum intensity of IX MCS degree and a magnitude of about $M=5$ (Fig. 2). In light of recent archaeological and epigraphic evidence, this earthquake and subsequent low-moderate energy seismic swarms have been considered as precursors of the AD 79 eruption (AA.VV., 1995; Marturano and Rinaldis, 1998; Cubellis and Marturano, 2002a; Luongo et al., 2003; Cubellis and Marturano, 2006).

In AD 64 a seismic event occurred during a representation in the theatre in Naples when Nero was present

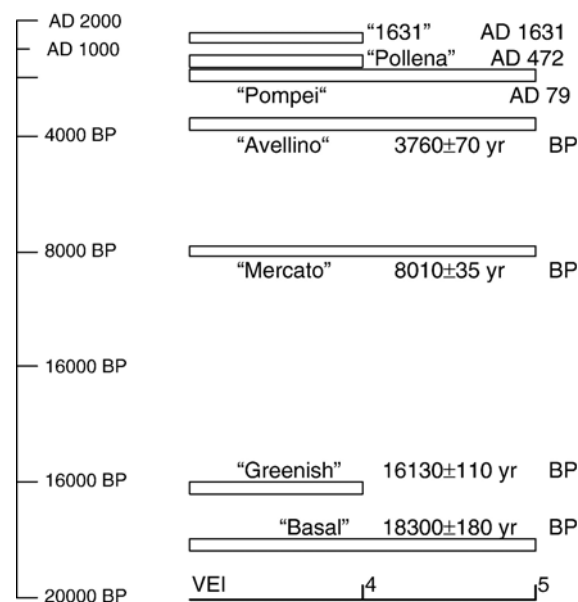


Fig. 1. Chronogram of Vesuvian activity in the last 18 ka (modified by Andronico et al., 1998). Plinian and subplinian eruptions (VEI 4–5) are reported.

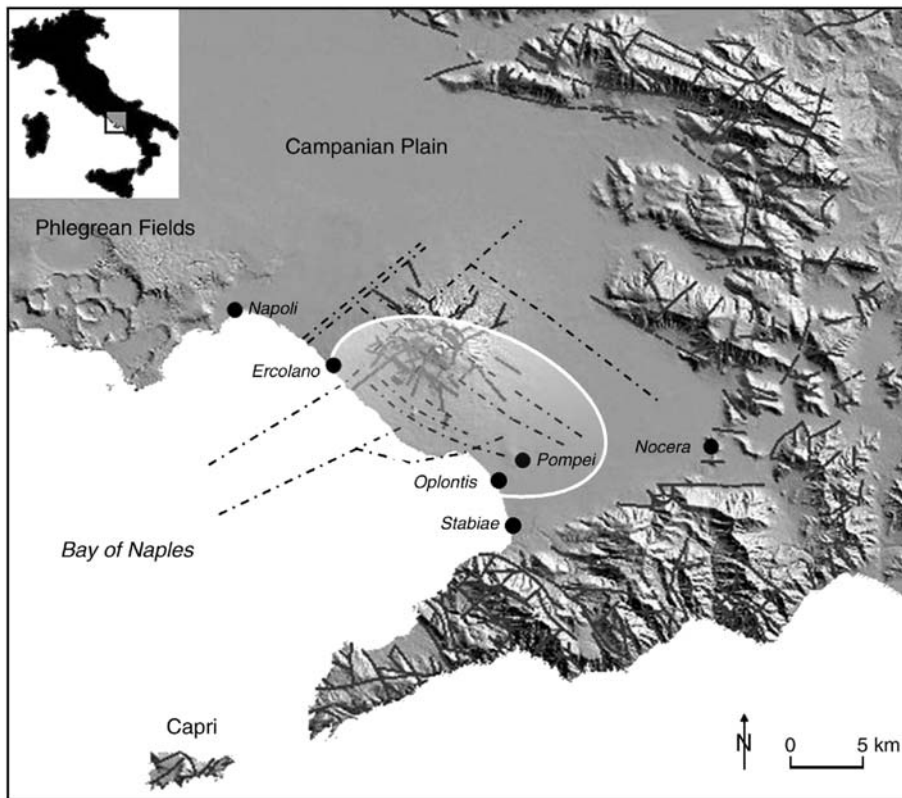


Fig. 2. Mt. Vesuvius 62 AD earthquake. The ellipse defines the area of maximum damage. The main tectonic structures of the Mt. Vesuvius and the Campanian Plain are reported. Continuous line: Topographic lineaments and faults by DTM analysis. Dashed lines and points: faults and fractures of the shallow basement from geophysical surveys (from: Maino et al., 1964; Bruno et al., 1998; Ventura and Vilardo, 1999; Cubellis et al., 2001).

(Tacitus, Ann. XV. 34.1; Suetonius, Nero, 20.2). According to Tacitus the theatre was damaged and Nero escaped the danger.

In the last years of the Roman Empire, the sources describe a large eruption at Vesuvius in the 5th century well known to volcanologists, but for this period there are no data on seismicity. In Medieval times the available sources record no significant seismic activity in the Vesuvian area despite the occurrence of some large eruptions (Santacroce, 1987; Figliuolo and Marturano, 1995, 1997; Principe et al., 2004).

The largest eruption in modern times occurred in 1631: there is reliable documentary evidence for seismic activity during the night between 15 and 16 December from primary sources in State and Ecclesiastical archives. At 10 p.m. on 15 December a shock at Vesuvius was felt in Naples; its intensity distribution suggests a magnitude of 4.0. Several light shocks occurred until midnight and then a particularly strong tremor followed; other earthquakes were felt afterwards until 6:30 a.m. when the eruption began (Santacroce, 1987; Marturano and Scaramella, 1995, 1997; Principe et al., 2004). After this eruption up to

recent times the earthquakes were generally of low-moderate energy and related to eruptive activity. The most dangerous occurred on 15 June 1794 during the lateral eruption which destroyed the town of Torre del Greco (Della Torre, 1794a,b; Breislak and Winspeare, 1794). The shocks caused damage to buildings in the Vesuvian area and shattered window panes in Naples.

Since 1944, seismicity at Vesuvius has been marked by events with a frequency of a few hundred per year concentrated in the summit caldera at a depth of less than 6 km below sea level (Vilardo et al., 1996; Bianco et al., 1999; Capuano et al., 1999; Zollo et al., 2002; De Natale et al., 2004, 2006). The most significant event recorded before the development of the surveillance seismic network at Vesuvius in 1970 is that of 11 May 1964, in the crater floor (Imbò et al., 1964). The felt area covered the whole Vesuvian region with maximum effects at the upper part of the volcano (V MCS degree) which gradually decreased towards its apron (II–III degree).

Several periods of greater activity were recorded in 1989, 1990, and more recently, in 1995 and 1996. At that time the strongest event occurred on 25 April 1996

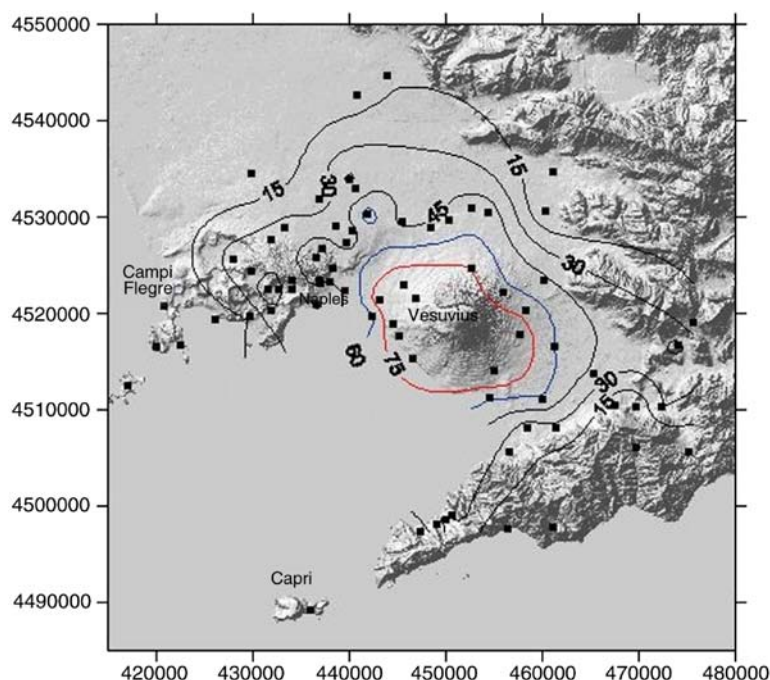


Fig. 3. Felt index map of the 9 October 1999 earthquake (from Cubellis and Marturano, 2002a modified). The felt index is the percentage response to the question *Did you feel the earthquake?* The area of maximum value (75%) covers the whole Vesuvian area. The felt index weakens strongly in NE and SE directions. Towards Naples a slighter decrease in the felt index is observed.

($M_L=3.4$, $h=2$ km) and was clearly felt all over the Vesuvian region, including Naples and some zones in the Phlegraean Fields and on the island of Capri. In the early months of 1999 seismicity increased slightly. During August 1999 a seismic sequence began and culminated in the most energetic event on 9 October; the instrumental data were the following: Epicentre location (crater area); $M_L=3.6$; $h=3$ –4 km (Zollo et al., 2002; Del Pezzo et al., 2004; De Natale et al., 2004). The M_L currently used at the Vesuvius Observatory is a duration-based scale calibrated using a Wood–Anderson equivalent station in Rome as reference (Del Pezzo and Petrosino, 2001).

To sum up, the Vesuvian area experienced moderate energy events from 1631 to 1944. During this span of time the magnitude of the most severe event, that of 1794, is the same as that of the strongest event prior to the 1631 eruption and close to that of 1999. The strongest event in Classical times was that of AD 62; for the Middle Ages, the least known period, there is no reliable information on seismicity although a number of reported tremors generally affected the city of Naples.

3. Intensity–magnitude and attenuation law

The earthquake of 9 October 1999 was felt over a wide area within about 25 km from the crater and caused fear

and anxiety among the people. A macroseismic study of this event was carried out from an analysis of questionnaires sent out to all secondary schools in each municipality of the Vesuvian and Neapolitan area and surrounding towns (Cubellis and Marturano, 2002a). Maps of earthquake intensity and a felt index were drawn up (Fig. 3). Marked attenuation areas were found in the epicentral area in the zones E and NE of the volcano and in the Phlegraean Fields produced by the source mechanism, scattering and low rigidity of the shallow rocks. Several authors (Zollo et al., 2002; Del Pezzo et al., 2004; De Natale et al., 2004; Guidarelli et al., 2006) agree with a normal fault solution for the 9 October 1999 earthquake but their solutions differ in the strike of the fault plane (E–W or NW–SE) and the strike slip and non-DC components. The contours of felt index areas illustrate that maximum intensities correspond to the volcanic edifice from which they decrease radially. On analyzing the whole data set Cubellis and Marturano (2002a) estimated epicentral intensity (I_0 : 6–6.4), attenuation coefficient (Blake's formula), absorption factor, equivalent macroseismic magnitude ($M=3.6$ –4.1), and the quality factor ($Q=60$ –90), and measured these against the values obtained by instrumental records.

The data on the 9 October 1999 earthquake can be used to assess the energy of historical earthquakes in

order to determine the level of seismicity and relate it to the volcano's seismogenetic structures and/or to explosive activity. Fig. 4 shows theoretical intensity curves for magnitudes between 3 and 5, and demonstrates that $M \leq 3$ earthquakes are felt up to the foot of Mt. Vesuvius, while those with $M=3.5$ and 4.0 are felt throughout the Bay of Naples and as far as the Phlegraean Fields. In addition, earthquakes with $M=4.5$ can cause slight damage in central Naples, at a distance of 12 km from Mt. Vesuvius.

Using the intensity versus distance relation for the 3–5 Magnitude range, according to Karnik's modified magnitude-epicentral intensity relation Cubellis and Marturano (2002a) estimated the magnitude of earthquakes preceding the eruptions of AD 79 and 1631. The earthquakes from 1631 to 1944 do not appear to have exceeded the value of $M=4.5$. In particular, the seismicity which occurred before and accompanied the eruption of 1631 probably lasted only several hours or a few days during which numerous events of moderate-low energy were released.

The greatest pre-eruptive event, however, was rated at $M=4.0$, a value that corresponds to those recorded in recent times. Finally, Fig. 4 shows the intensity values expected for a magnitude of 5 and their decrease with distance from the crater; these values agree with those experienced during the AD 62 earthquake, the largest recorded in the Vesuvian area.

However the level of intensity in volcanic and tectonic areas differs for the same magnitude earthquake. Indeed, comparison between attenuation laws in the Apennines and Vesuvius and Campi Flegrei areas evidences epicentral intensity differences for equal magnitude events. Fig. 5 shows that the earthquakes at

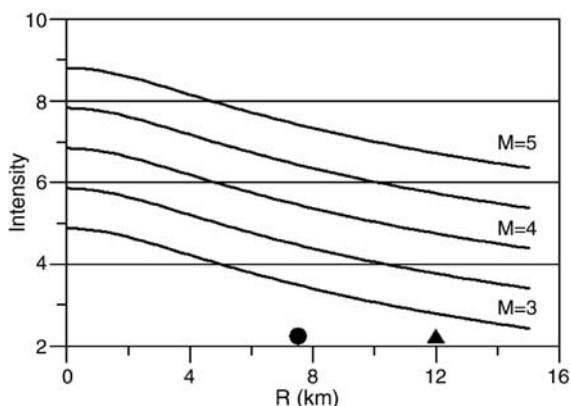


Fig. 4. Expected earthquake intensity for magnitude 3–5 as a function of epicentral distance (from Cubellis and Marturano, 2002a modified). The circle represents the mean distance of the towns at the foot of Vesuvius from the crater axis; the triangle the distance of Naples.

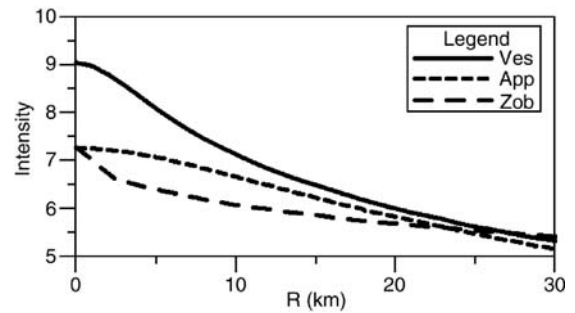


Fig. 5. Intensity for $M=5.1$ earthquake as a function of epicentral distance. Ves: Vesuvius; App: Apennines; Zob: Zobin (2001), focal depth $h=3$ km. As may be observed the Zob relation like that for tectonic areas (App) does not fit the data in the epicentral region at Vesuvius.

Vesuvius are as much as one to two epicentral intensity degrees greater than equivalent magnitude events occurring in the Apennines. In the figure we report the expected intensities for equal magnitude events ($M=5.1$) occurring on Mt. Vesuvius and the nearby Apennine Chain respectively (Grandori et al., 1988; Cubellis and Marturano, 2002a) as well as the expected intensities using the relation proposed by Zobin (2001) for volcanic earthquakes.

4. Source dimension scaling

The faulting parameters of most earthquakes appear to be systematically related, thereby providing scaling relations: seismic moment (M_0) is proportional to the cube of fault length (L^3) for a wide range of earthquakes (Aki, 1967; Kanamori and Anderson, 1975; Kasahara, 1981). An upper limit for “small” earthquakes following relation $M_0 \approx L^3$ is set at $M_0 = 7.5 \cdot 10^{25}$ Nm ($M_w = 6.6$), after which $M_0 \approx L^2$ (Shimazaki, 1986; Lay and Wallace, 1995). Abercrombie (1995) and Prejean and Ellsworth (2001) used spectral analysis from borehole recordings for Cajon Pass and Long Valley caldera (California) earthquakes, respectively, showing that constant stress drop scaling continues to $M_0 = 10^9$ N m ($M_w = 0$), appearing generally from 0.1 to 30 MPa. Recently, Stork and Ito (2004) investigated $0.6 < M_w < 3.0$ earthquakes in western Nagano, Japan, obtaining a scaling relationship of static stress drop in the range 0.01 – 10 MPa with $10^{10} < M_0$ (Nm) $< 10^{13}$.

Seismic moments, source radii and stress drops of the largest events at Mt. Vesuvius during the period 1989–1999 showed an approximate constant stress drop scaling moment in the range 1.5–9.2 MPa (Zollo et al., 2002). Similar values were obtained by De Natale et al. (2004) and Galluzzo et al. (2004). The latter,

nevertheless, observed that the scaling law clearly deviates from a constant stress drop relation in the whole range of magnitude $0.4 < M_L < 3.6$, analyzing the earthquakes recorded in the period 1997–2000.

The inferred mean radii of seismic sources at Mt. Vesuvius are plotted against the mean seismic moments in comparison with the results of Long Valley and Nagano (Prejean and Ellsworth, 2001; Stork and Ito, 2004) (Fig. 6). This figure also reports the corresponding values for the 9.10.1999 and AD 62 earthquakes obtained from estimated macroseismic magnitude ($M=3.6$ and $M=5.1$ respectively). For these two earthquakes the seismic moment (M_0) and circular crack radius (r) were calculated utilizing the relations $\log M_0 = 1.5 M + 16.1$ (Lay and Wallace, 1995) and $\Delta\sigma = 7/16 M_0/r^3$ (Elsheiby, 1957) respectively, assuming the mean value of stress drop $\Delta\sigma = 3$ MPa. The values obtained are $M_0 = 5.6E+16$ Nm and $3.2E+14$ Nm, $r = 2.0E+3$ m ($A = 12.8$ km²) and $3.5E+2$ m ($A = 0.4$ km²) for the AD 62 and 9.10.99 earthquakes respectively. The M_0 estimated by macroseismic magnitude for the 9.10.99 earthquake has the same value as the instrumental ones (Zollo et al., 2002; De Natale et al., 2004; Del Pezzo et al., 2004).

According to historical data the maximum magnitude observed at Mt. Vesuvius is 5.1 for the 62 AD earthquake. For this event the estimated error for the intensity is 0.5 degrees, which corresponds to a value of 0.3 for

the magnitude (Cubellis and Marturano, 2002a). Therefore if we use the value of $M=5.4$ for the maximum value observed in historical times, we obtain a rupture area of 25.5 km² and a radius $r=2850$ m according to the empirical relation of Lay and Wallace (1995) reported above.

To obtain the maximum magnitude expected at Mt. Vesuvius we set geological and geophysical constraints for modelling reasonable tectonic structures as well as the seismic source below the volcano. The seismicity of the last three decades at Mt. Vesuvius, in the magnitude range 1.8–3.6, shows that the Gutenberg–Richter distribution of these data provides an $M_{\max}$ of 4.1 (Galluzzo et al., 2004) corresponding to a source radius of $r = 6.4 \times 10^2$ m and $A = 1.3$ km² using the above relations. The gap between the source dimension of the AD 62 earthquake and those of the maximum magnitude obtained for the present seismicity shows that the seismogenetic sources have to be considerably different. The poor sequence of historical seismicity prevents robust statistical analysis for the evaluation of the expected maximum earthquake. Hence it is necessary to analyze in greater detail the structure of the volcano and the basement to obtain a coherent framework for a seismic source larger than the present one.

5. Geological and geophysical constraints

The Somma Vesuvius volcano rises in the Campanian Plain graben, a region affected by tensile quaternary tectonics (Fig. 2). NW–SE and NE–SW normal faults dissect the Apennines units and submerged the carbonate basement forming the graben filled up with marine and alluvial sediments. The basement rocks crop out on the rims of the depressed area and are present in drill holes in the northern-western sector of the graben at about 1.3 km deep and in the Vesuvian area, at about 1.9 km deep. In drill holes deeper than 3 km in the northwestern coastline of the graben in the vicinity of the mouth of the Volturno river, and in the Phlegraean area, the carbonate basement was not found, testifying to its considerable lowering and/or disruption. According to gravity and seismic data the carbonate basement deepens towards the Bay of Naples and the centre of the Campanian Plain where it reaches a maximum depth approximately of 3000 m on land. The tectonics around and across the volcano show the same regional trend (Fig. 2). The shallow structure of Mt Vesuvius has been deduced both by magnetic, gravimetric and seismic surveys and the borehole drilled down to a depth of 2200 m b.s.l. (Cassano and La Torre, 1987; Cubellis et al., 1995; Berrino et al., 1998; Bruno et al.,

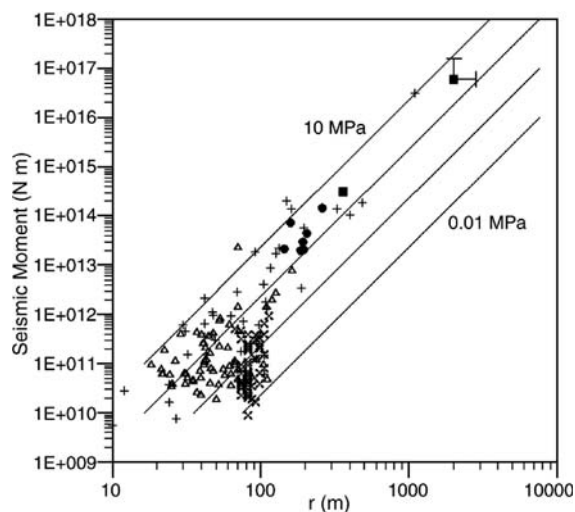


Fig. 6. Seismic moment versus source radius. The lines are those of constant stress drop (0.01–10 MPa). Results from Zollo et al. (2002) (closed circles), Galluzzo et al. (2004) (xs), Prejean and Ellsworth (2001) (crosses) and Stork and Ito (2004) (triangles) are also shown. The squares indicate the 9.10.99 and AD 62 Vesuvius earthquakes by macroseismic analyses; the error bars are ($r=+850$ m; $M_0=1$ E+17 Nm) for the AD 62 earthquake.

1998; Bruno and Rapolla, 1999; Brocchini et al., 2001; Cubellis et al., 2001).

Seismic tomography (Zollo et al., 1996, 1998) and surface wave dispersion measurements (Nunziata et al., 2006) suggest the presence beneath Vesuvius of a melting zone with the top at a depth of about 8 km, but small and shallower magma chambers, hypothesized by petrological and geochemical data, are undetected. A high P-wave velocity body corresponding to the crater axis zone was defined above the carbonate basement at a depth of approximately 2 km. Moreover the decay of intensity of local earthquakes in the Campanian Plain was interpreted with the brittle–ductile transition at a depth of 8 km around the volcano (Cubellis and Marturano, 2002b).

According to the model proposed by Solaro (2004) and Borgia et al. (2005), on the rheological stratification of the crust beneath Mt. Vesuvius, the brittle–ductile transition occurs at a depth of about 4 km that corresponds to the cutoff depth for seismicity. A thermal model for the Vesuvius magmatic source (Civetta et al., 2004) indicates that the 350 °C isotherm — which represents the transition between brittle and ductile layers (Scholtz, 2002) — reaches depths of about 3000–4000 m.

The hypocentres of earthquakes at Mt. Vesuvius cluster around the crater axis at a distance up to 1 km, down to 5–6 km b.s.l., but 80% of the hypocentres are located between depths of 0 and 3 km at a horizontal distance of 0.5 km from the crater axis (Lomax et al., 2001; Scarpa et al., 2002; Del Pezzo et al., 2004, 2006). In addition, the focal mechanism solutions generally show the occurrence of normal faults with a highly non-regular fault plane pattern. Variable space–time slip directions with both right and left lateral solutions occur.

The structural data, as well as historical and present-day seismicity, suggest two potential seismogenetic structures in and around the volcano: one associated to the most superficial part of the system feeding the volcano; the other as faults that cross the basement of the volcano.

Therefore the seismogenetic volume of the recent seismicity at Mt. Vesuvius may be represented by a cylinder with a radius of 0.5 to 1 km and a height of 4 to 6 km.

The energy release from the unit volume of seismic source has a rather uniform value of about 10^2 J m^{-3} according to the empirical law $\log E = \log V + 3$ (Tsuboi, 1956). This value is similar to 10% of the potential energy in rocks at ultimate strain (10^{-4}), taking into account the seismic efficiency. Consequently, the volume of strain release specifies the magnitude of the resultant earthquake. Using the above empirical law,

which correlates seismogenetic volume to energy and hence to magnitude, $M=5$ corresponds to the greatest volume, while there is a smaller magnitude, $M=4.5$, at the lowest volume.

Within the volcano's current state of activity and allowing for the data of historical seismicity accompanying eruptive crises, it may be expected that the seismogenetic volume of rock capable of releasing seismic energy corresponds to the lowest value obtained from the above analyses. This hypothesis is also strengthened by the fact that the magnitude expected in this case ($M=4.5$) is very close to that obtained from data relating to more recent activity which supply an $M_{\max}$ of 4.1 (Galluzzo et al., 2004). In this case the seismogenetic volume corresponds to about 3 km^3 , while to reach $M=5.4$ of the AD 62 earthquake the seismogenetic volume should be about 80 km^3 . For an earthquake of such magnitude to occur, a much larger-size structure has to be activated compared with those associated to recent earthquakes and those that accompanied the eruptive crises from 1631 onwards ($M=3.6$ – 4.5).

To obtain the dimensions of a seismogenetic source associated to a fault rather than a rock volume we use Lay and Wallace's formula (1995) which correlates magnitude to seismic moment. Thus, once the magnitude is known from the seismogenetic volume and the seismic moment obtained, we calculate the seismogenetic surface $A = M_0 / U\mu$. The fault plane source for the 5.4 magnitude earthquake which best approximates the result of the seismogenetic volume of about 80 km^3 has the following parameters: $U = 10 \text{ cm}$; $\mu = 5 \times 10^{10} \text{ Nm}^{-2}$; $A = 32 \text{ km}^2$.

6. Discussion and conclusions

In the last three decades the seismic and ground deformation networks for monitoring the Neapolitan volcanic area and the vicinity of the volcanoes have shown no significant dynamics. Indeed, high precision levellings, tiltmeters, tide gauges, along the coast and GPS measurements (Pingue et al., 2004, 2005) indicate that at Vesuvius the ground is subsiding with a velocity less than 0.5 cm/year in the upper part of the volcano, while in the plain around the volcanoes up to the Apennine Chain the ground is sinking with a velocity similar to that of long-term tectonic deformations (1–2 mm/year) (Ambrosetti et al., 1987). In addition the seismicity is located only in the volcanic structures. No tectonic fault surveyed in the Campanian Plain by geological and geophysical exploration (gravimetric, magnetic, seismic tomography) has released seismic energy in historical times.

In the investigated area the crust is thinned by the uprising of the mantle; it shows the Moho at a depth of 25 km and a ductile layer with its top at 8 km around Vesuvius and less than this value under the volcano.

The tensile tectonics, the low regional dynamics and the rheological stratification of the crust characterized by a shallow brittle–ductile transition preventing a large strain in the upper crust strongly limit the seismogenetic volumes and facilitate creep processes in the crust. In such conditions the level of seismicity is low, with events of moderate magnitude and low frequency.

Instrumental seismicity data (1970–2005) show that the maximum earthquake recorded at Vesuvius has a magnitude $M_L=3.6$ (9.10.1999). For this earthquake, macroseismic analyses supply a slightly higher value ($M=3.6–4.1$). Instrumental data analysis using the G–R relation shows that the maximum expected earthquake would have a magnitude of 4.1 (Galluzzo et al., 2004).

Very similar magnitudes ($M=4.0–4.5$) were obtained for the seismic events that preceded and accompanied the most intense eruptions of the last eruptive period from the subplinian eruption of 1631 to the 1944 strombolian eruption, after which the volcano entered a state of quiescence.

Only the AD 62 earthquake shows a clearly higher magnitude ($M=5.4$) than that of all known events. Although for this earthquake there has been no reconstruction of the macroseismic field such as to obtain a satisfactory interpretation of source location and size, knowledge of the damage area allows us to estimate the magnitude and depth of the event with good reliability. In general, the seismicity falls within two distinct energy levels: one with events of magnitudes lower than 4.5 and the other with magnitudes over 5.0, consisting only in the AD 62 earthquake ($M=5.4$). Such distinct energy levels are probably activated by different stress fields. The seismogenetic volume of the quiescent phase and that of persistent activity is very modest ($\sim 3 \text{ km}^3$), similar to that defined by the distribution of the hypocentres of the last 30 years. By contrast, the AD 62 event has a source very close to that of tectonic events in the nearby Apennine Chain. In this case, it cannot be ruled out that the AD 79 eruption was prepared by intensification of the deformation processes in the area that gave rise to the AD 62 earthquake, which appears a singularity in the recent history of Vesuvius.

If the seismic data are supplemented by the knowledge of the rheological stratification of the crust, it may reasonably be affirmed that the reduced thickness of the brittle layer below Vesuvius does not allow the development of seismic sources of the same dimensions as those observed in the Apennines. Taking into account the

thickness of the brittle layer below Mt. Vesuvius (about 4 km) and the estimated dimension of the surface of the 62 AD earthquake source (32 km^2), the vertical dimension of the fault generating this event would represent the limit value for the earthquakes in the Vesuvius area. As this seismic source does not show such characteristics that it can be considered an evolution of current low-moderate energy seismicity, an earthquake like that of AD 62 could only be generated with a significant change in stress field, while in the short term the strongest earthquake expected should fall within the 4.0–4.5 range of magnitude; with the 9.10.99 earthquake we came close to this limit.

To define the source of the AD 62 earthquake accurately it will be necessary to integrate the regional stress field with the stresses produced by volcano loading and magma pushing. It is well known that Mt. Vesuvius is located in the regional stress field with a tensile tectonics that produces the spreading of the Tyrrhenian Sea (Patacca et al., 1990; Luongo et al., 1991; Vai and Martini, 2001). The complexity of the geological structure of the Bay of Naples and the fuzzy sequence of the historical seismic data make it difficult to assess the accurate location, dimension and strike of the seismogenetic source.

However, different seismogenetic sources can be reliably identified for the two energy levels of earthquakes by tectonics (Fig. 2) and physical properties of rocks. The higher level may be associated with the carbonate basement which is characterized by tensional Quaternary tectonics with the main fault systems trending NW–SE and NE–SW. The most foreseeable faults as active deformation sources are located at the edge of the plumbing system of Mt. Vesuvius because the thickness of the brittle layer increases from the central to peripheral volcano structure. Hence it can be ruled out that the seismogenetic source of the strongest earthquakes at Mt. Vesuvius is a fault which crosses the volcano edifice through the magma feeding conduit.

Instead, the seismicity of the lower energy level appears associated to faults related to volcanic cover a few kilometres thick. In this last case numerous structural lineaments constitute possible seismic sources.

Acknowledgments

We gratefully acknowledge Margaret Mangan and the anonymous referees for their helpful comments and suggestions in improving the quality of the paper. We thank Edoardo Del Pezzo, for his critical reading and useful discussions, and Hermann Engelen for his editorial advice and cooperation.

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