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Upwards migration of seismic focii: A forerunner of the 1989 eruption of Mt Etna (Italy)

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Abstract. Data from a portable array of three-component digital stations, run at Mt Etna from 1988 to early 1990, highlight the seismic behaviour of the volcano before the 1989 eruption, one of the most significant in terms of energy of the last two decades. After a two-year period of weak and discontinuous seismicity, the depth of the seismically active volumes was observed to become shallower a few months before the volcanic event. The overall migration of the events, inferred by hypocentral locations and decreases of S-P time differences at two stations, agrees with other geophysical forerunners and allows further insights into the changes in the stress field leading to the eruption.

Key words: Mt. Etna – seismic network – eruption – migration – hypocenter – S-wave – forerunner.

Introduction

Increases in magnitude and frequency of occurrence of earthquakes preceding an eruption is commonly attributed to major changes in the internal stress field originating from magma movements. Also, at Mt Etna (Fig. 1) a change in the seismic energy release was observed a few months before the outbreak of the major eruption of September–October 1989 (Fig. 2).

No noteworthy seismicity had been observed in the months following the end of the previous (1986–1987) eruption, which also originated at the subterminal 'S-E Crater', with the exception of a few shallow sequences located on the lowermost flanks of the volcano (Ferrucci et al. 1991). This type of seismic activity, common to the peripheral structures of Mt Etna (Lo Giudice and Rasà 1992), displays extremely shallow depths (typically even less than 1 km) which require proper narrow seismic arrays to constrain the focii instead of networks operating at the scale of the volcano.

The seismic sequences which occurred during 1988 and 1989 do not display space-time patterns clearly linked to the resumption of the volcanic activity and, therefore, cannot be considered effective eruption precursors. Nevertheless, they mirror the overall tectonic behaviour of the Etnean substratum (Lo Giudice et al. 1982; Lo Giudice and Rasà 1986) and may occur in response to abrupt modifications of the pattern of a central stress-field (e.g. at the end of the 1984 eruption, Gresta et al. 1987; Hirn et al. 1991; Nercessian et al. 1991) and/or to the eastwards gravity-slide tendency of the eastern part of the volcano (Murray and Guest 1982; McGuire and Pullen 1989; McGuire et al. 1990; 1991).

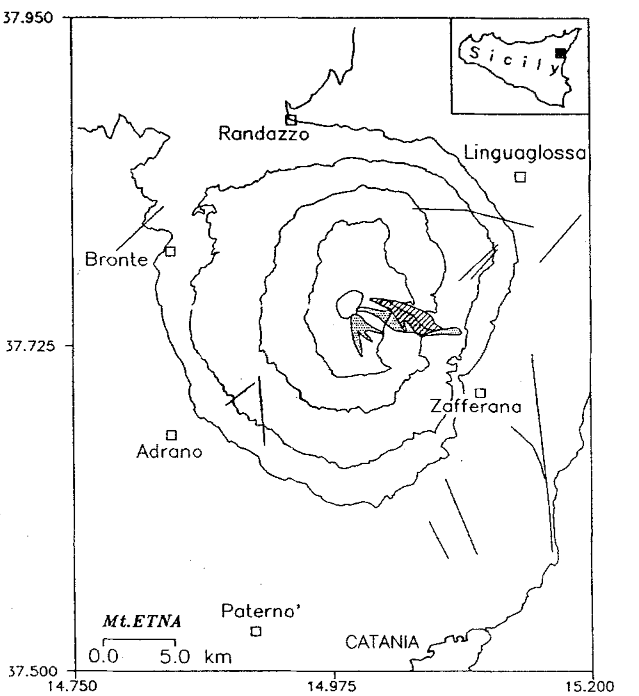

Fig. 1. Tectonic sketch of Mt Etna volcano with the main structural trends (*lines*) and the lava flows of the 1986–1987 (*hatched*) and 1989 (*dotted*) eruptions

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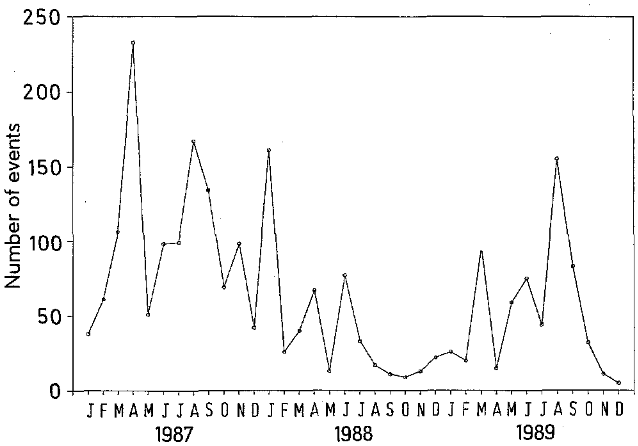

Fig. 2. Seismic monthly frequency at Mt Etna from 1987 to 1989 recorded at station SLN

During 1988 and until early 1989 seismicity was characterized by a few isolated events and minor swarms, the most important of which occurred on 10 April (seven events: $5 \leq z_{km} \leq 10$; $M_L \leq 2.3$) and 19 June 1988 (ten events; $18 \leq z_{km} \leq 23$; $M_L \leq 2.9$).

The resumption of major seismic activity was observed beginning 7 July 1989, when a 10-km-deep $M_L = 3.0$ event was located beneath the southeastern Valle del Bove. Several swarms of variable intensity followed after 27 July.

The 1989 eruption started on 11 September, accompanied and preceded by vigorous tremor bursts but by weak-to-nil earthquake activity. Two major swarm sequences, respectively composed of some tens of deep (23–24 September) and some hundreds of shallow (30 September–3 October) events, were later observed before the end of the eruption. The 23–24 September 1989 sequence was composed of regional and local (Etnean) earthquakes and it is not possible to accurately locate the Etnean events because they occurred near the southwestern edge of the array. The S-P times at the nearest stations to the epicentral area seem to indicate a depth ≥ 15 km.

Seismicity, other than skin-deep, reflects the changes with time of the stress field acting on the volcano and leading to an eruptive events. Three-component data, gathered at a portable array of digital stations operated at Mt Etna beginning in 1988 (Fig. 3), provide evidence for upwards migration of the increased seismicity which preceded the 1989 eruption.

The data

As a function of both level and type of the ongoing seismo-volcanic activity, the geometry of the portable array underwent several changes during the operation period (April 1988–March 1990). We left unchanged the position of two stations (NIK – event-trigger station and SLN – transmitted via cable and continuously recorded at Osservatorio Vesuviano in Napoli; Fig. 3) throughout the experiment to maintain a minimal reference geometry with time.

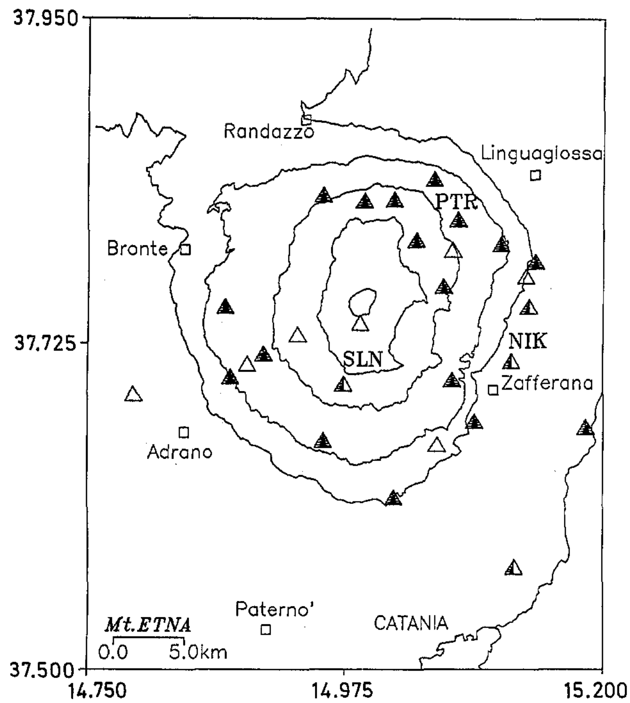

Fig. 3. Temporary digital array operating during 1988 (*filled triangles*) and 1989 (*open triangles*). Stations operating for the whole period are indicated with *half-filled triangles*

The number of three-component digital stations contemporaneously operated on the volcano ranged from 7 to 22, with the only exception during July–December 1988 when there were only four stations. In this period, triggering and recording using the standard short-time/long-time signal averages ratio technique, was difficult because of continuous and strong tremor activity, which raised the long-time average to very high levels. However, this period was characterized by a substantial lack of significant local seismicity as marked by the continuous recording of SLN station.

Slightly more than 50 events were detected at the array during the whole year 1988. Early 1989 was characterized by a substantial lack of seismic activity (Ferrucci 1990). Resumption of seismicity in July 1989 (Fig. 2) resulted in the detection of more than 250 events at the array of Fig. 3. The strongest events recorded reached $M_L = 3.3$.

Since field equipment was self-triggered on event detection, lower Magnitude events were generally not recorded at enough stations to allow reliable location. The estimated Magnitude threshold for triggering the whole network is slightly larger than $M_L = 1.5$.

The fully usable data set (36 events, 25 of which belong to the 1989 pre-eruptive period) contains the events for which at least eight phase arrivals, recorded at least five stations, are available. Such a subset is representative of the larger energy seismicity observed during the operation period.

The events (epicenters shown in Fig. 4) were located using the standard location routine HYPO71 (Lee and Lahr 1975) with a six-layer veloctiy model taken from Hirn et al. (1991). The model, character-

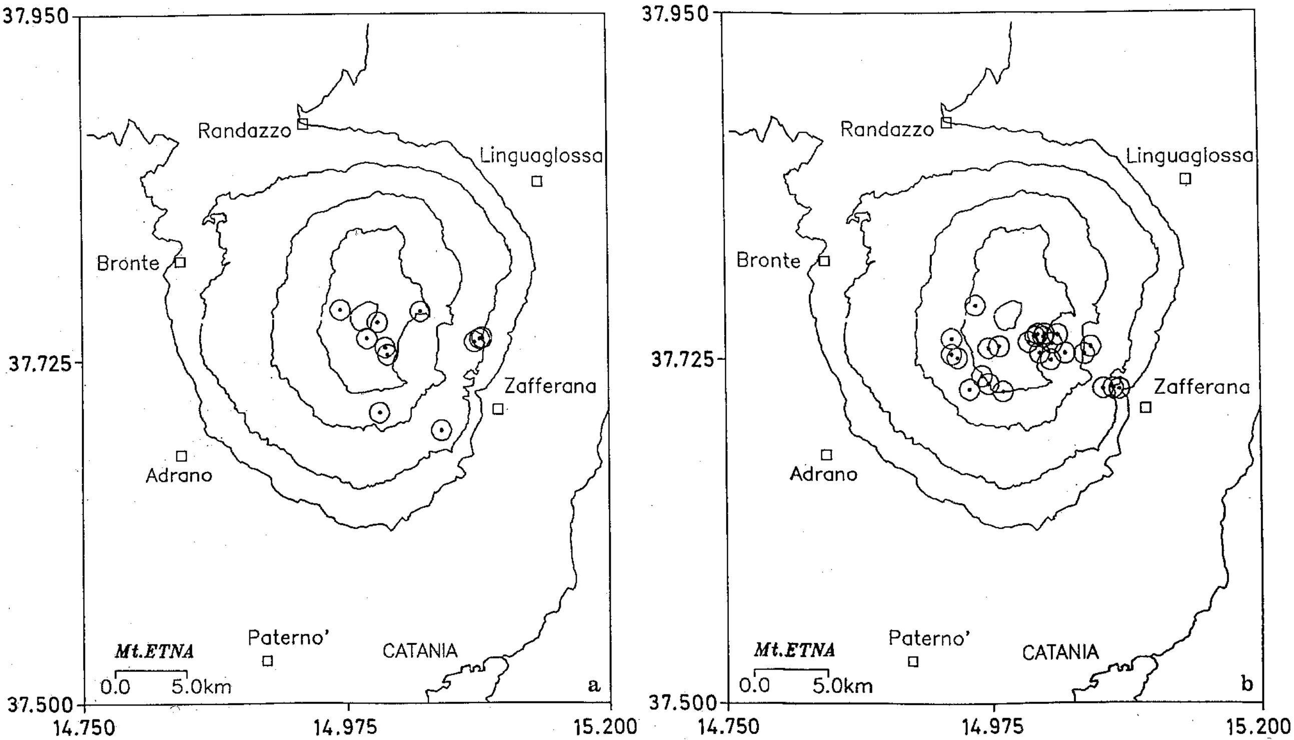

Fig. 4a, b. Location map: **a** Epicentral distribution of the events occuring before summer 1989. Skin-deep peripheral events (see text) are not reported. **b** As in (a), from early summer 1989 to the onset of the 1989 eruption

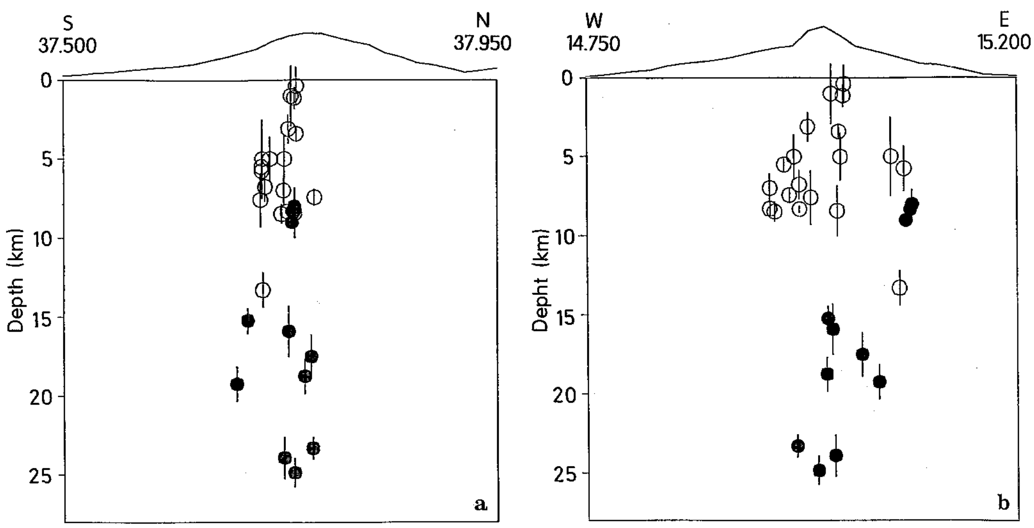

Fig. 5a, b. Hypocentral distribution along cross-section NS (**a**) and EW (**b**). *Full circles* indicate 1988 hypocenters; *empty circles* the 1989 ones; *error bars* the statistical standard arrors (ERZ) from HYPO71 Hypocenter Output. Note that W-E spreading of the hypocentral volume exceeds the plane location errors (<1.0 km)

ized by a generally smooth increase of P–velocities with depth, contains two major discontinuities at 3.0 km (from 3.1 to 4.0 km/s) and 8.0 km b.s.l. (from 4.8 to 5.8 km/s) respectively.

A Vp/Vs ratio (1.74) slightly higher than that of the reference model was found to best fit the data. However, slightly different Vp/Vs values were needed for best locating the events epicentered on the western (>1.74) or on the eastern flanks of the volcano (<1.74). Although negligible, such a difference may agree with the presence of a significantly large P and S velocity high on the eastern side, beneath the Valle del Bove (0–8 km b.s.l.; Hirn et al. 1991). The largest errors in depth and epicenter do not exceed 1.5 and 1.0 km respectively, while the RMS (root mean square) residuals are typically less than 0.15 s.

As shown in Fig. 5, the events preceding the 1989 eruption by a few months display shallower depths than those occurring in the two years following the 1986–1987 eruption.

Three-component stations, essential for unambiguous identification of the S-wave onset (see Fig. 6), allow us to provide evidence that this observation is independent of changes in the recording array geometry and/or from the limited number of best-constrained events shown in Fig. 3.

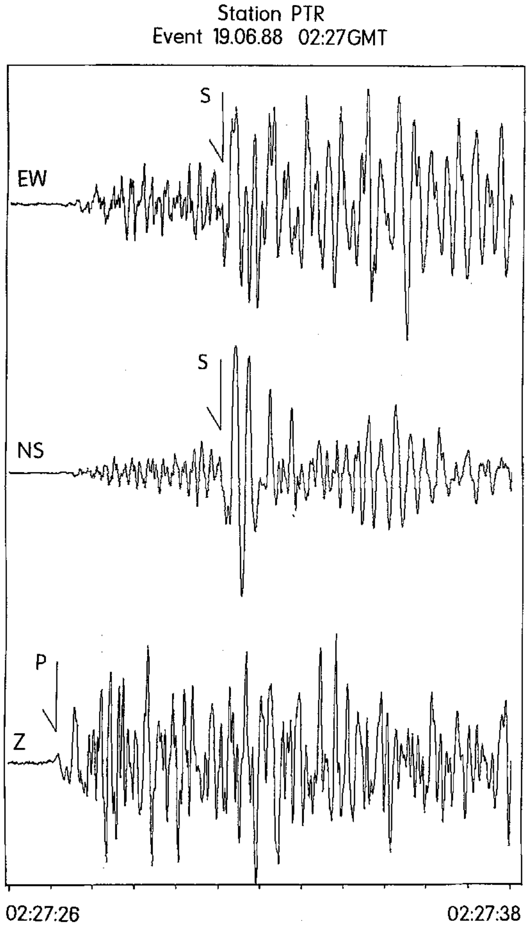

Fig. 6. Example of an earthquake recorded at station PTR during 1988. Note how the S-phase is evident only on the horizontal components while it is quite indecipherable on the vertical one

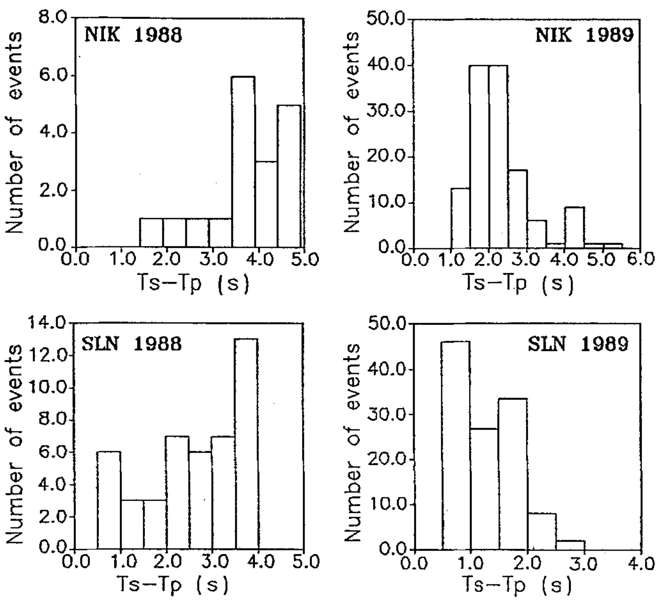

Fig. 7. Ts-Tp differences for all the events recorded at three-component stations NIK (*top*) and SLN (*bottom*), respectively before summer 1989 (*left*) and from early summer 1989 to the onset of the eruption (*right*)

Figure 7 shows the observed S-P time differences for *all the events recorded* at stations SLN and NIK, respectively located 5 km SSW and 9 km ESE of the SE active crater, and kept in continuous operation during the two-year observation period. Station SLN (Fig. 3) was continuously recorded and provided a larger data set than event-triggered station NIK.

On the whole, Ts-Tp differences are significantly smaller for the events recorded in the three months preceding the 1989 eruption than for the former observation period. This indicates an overall shortening of the foci-to-stations distances with time, in agreement with the calculated focal depths of the higher quality data subset.

Discussion and conclusions

Although migration of seismic events with time is considered to be a likely precursor to volcanic eruptions, the complexity of the magma transport towards the surface and inadequacies of many seismic arrays make this event difficult to constrain (e.g. Hirn and Michel 1979).

We can state that, in our case, the quality of the data are sufficient to place satisfactory constraints on hypocenter depth which allow us to observe the upward displacement with time of the seismically active volumes.

Such a forerunner is in agreement with other forerunners of the 1989 eruption of Mt Etna.

For instance, sharp inflation of the volcanic edifice was observed beginning in the spring of 1989 at two tilt stations, respectively located 7 km SSE and 10 km NW of the summit craters' area (Bonaccorso et al. 1990; Briole et al. 1990a, 1990b). This preceded renewal of stronger and shallower seismic activity by a few months.

On the other hand, in the biennial period following the 1986–1987 eruption, the gravity field displayed an increase as large as ca. 100 microGal over a horizontal distance of 10 km (Budetta et al. 1989). The large amount of this anomaly, centered a few kilometers SSE of the central craters, indicates a probable participation of magma uplift causing changes in the stress and strain fields acting on the volcano. Since such a major gravity variation occurred during a period of weak-to-nil seismicity, it can be deduced that magma ascent occurred almost 'silently', with the only exception being the minor swarms of April and June 1988.

The clustering of deeper seismicity (typically slightly less than 10 km and slightly more than 20 km b.s.l.) could thus point to the presence of major structural boundaries, which acted as short-lasting barriers to rising mantle magma. Some attempts have been made to model seismicity, magma recharge and transport at Mt Etna (e.g. Cristofolini et al. 1987; Hughes et al. 1990) but petrological constraints on the erupted products (Armienti et al. 1989, 1990) allow us to exclude the possibility that the clustering of seismicity at such depths is indicative of the existence of intermediate or deep magma chambers.

Intermediate-depth seismicity (z≥15 km) could be relate to readjustment of the internal (local) stress-field in response to change in the external (regional) one, or to changes of the internal state of stress in response to displacement of magma masses towards shallower depths. Coexistence of both tensile and compressive focal solutions for such a depth class of events (Scarpa et al. 1983) suggest that both hypotheses may be equally valid.

Conversely the deeper seismic zone (about 20 km), unfortunately laying beyond the geometrical limits of a passive seismic tomography investigation (Hirn et al. 1991), was activated twice during the observation period (19 June 1988 and 23–24 September 1989).

In both cases the seismic sequences displayed a swarm character. In the first case, resumption of major volcanic and tremor activity (without overflows) followed a few days later. In the second, the swarm accompanied and/or preceded by a few hours a sharp increase of the already high eruptive activity levels (Bertagnini et al. 1990). Inference on the direct relationship between this seismicity and short-term changes in type and intensity of the volcanic activity is certainly not allowed by only these two examples. Furthermore, these are not directly comparable to similar sequences of the past, where hypocenters were located without the constraint of S-wave data from three-component stations (e.g. Scarpa et al. 1983; Gresta and Patanè 1987).

However, such examples suggest that the occurrence of deep swarm events beneath the volcano may be relevant for constraining the recharge mechanisms and, therefore, for forecasting future volcanic events at Mt Etna.

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