Retrieval of large volcanomagnetic effects observed during the 1981 eruption of Mt. Etna

Ciro Del Negro [1], **Fabrizio Ferrucci** [2] **and Rosalba Napoli** [1]
[1] *Istituto Internazionale di Vulcanologia, CNR, Catania, Italy*
[2] *Dipartimento di Scienze della Terra, Università della Calabria, Arcavacata di Rende, Cosenza, Italy*

Abstract
A large temporal anomaly was retrieved in the total geomagnetic field series recorded in 1981 on Mt. Etna at two continuously recording magnetometers, and associated with the March 17-23 eruption of the volcano. Variations were of such large scale that a 10 nT anomaly was observed at a distance of some 7 km from the eruptive events, calling for a significant extension and depth of the magnetic anomaly source. We discuss here some models which may account for such magnetic changes in relation to the eruption mechanism inferred by other data. The anomaly is thought to be accounted for by the joint effect of piezomagnetism of the country rocks and thermal demagnetisation engendered by a large intrusive dyke.

Key words *Mt. Etna – thermomagnetism – electrofiltration – piezomagnetism*

1. Introduction

Since the early 20th century, observed anomalous variations of the geomagnetic field were thought to accompany volcanic activity (Wilson, 1922). The first reliable measurements on active volcanoes were carried out on Oshima volcano central cone, Mihara-yama during the 1950s, aimed at observing changes in inclination and declination of the magnetic field. Immediately after the 1951 eruption, Rikitake (1951) detected changes in inclination as large as 30 min of arc over 2 months, in the area surrounding the central cone; Yokoyama (1957) observed a gradual increase in declination (up to 7 min of arc) during the moderate activity of 1953-1954.

With the development of the proton precession magnetometer (Overhauser, 1953; Sigurgeirsson, 1970), regular and accurate measurements of the geomagnetic field intensity became feasible. Quantitative studies of volcanomagnetic effects developed, leading to many successful observations of the correlation between volcanic activity and magnetic changes (*e.g.*, Johnston and Stacey, 1969; Davis *et al.*, 1979; Pozzi *et al.*, 1979; Zlotnicki, 1986; Johnston *et al.*, 1981; Davis *et al.*, 1984; Yukutake *et al.*, 1990; Sasai *et al.*, 1990).

Overall, large and slow magnetic changes of the order of several tens of nanoteslas observed on volcanic edifices are generally attributed to the thermal effect (*e.g.*, Tanaka, 1993), the temperature dependence of the magnetisation of rocks being well established (*e.g.*, Nagata, 1943). Conversely, small (say, less than 15 nT) and swift or sudden changes in the magnetic field intensity at active volcanoes can be ac-

Mailing address: Dr. Ciro Del Negro, Istituto Internazionale di Vulcanologia, CNR, P.zza Roma 2, 95123 Catania, Italy; e-mail: delnegroc@iiv.ct.cnr.it

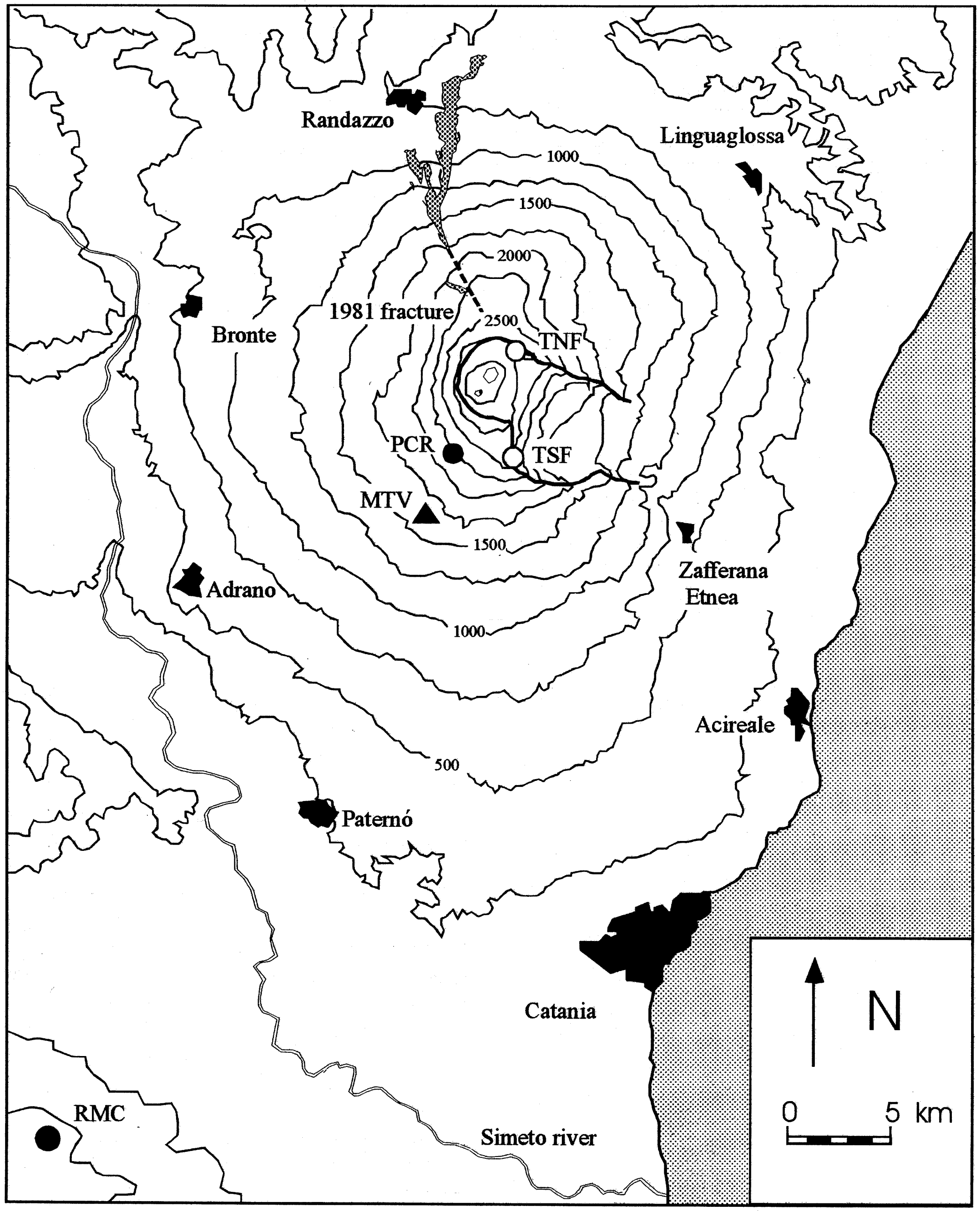

Fig. 1. Location map of continuously recording stations run on uppermost Mt. Etna in March 1981. Filled circle: proton precession magnetometer. Open circle: bore-hole tilt gauge. Filled triangle: seismic station. The dashed line indicates the 1981 eruptive fissure. The length of the flow was circa 10 km. The distance between the uppermost events and the nearest magnetic station (PCR) was about 7 km.

counted for by two types of mechanisms: electrofiltration (*e.g.*, Zlotnicki and Le Mouël, 1990) and piezomagnetism (*e.g.*, Stacey and Johnston, 1972). The electrokinetic effect is linked to the observation that, in the presence of heterogeneous conductive regions, the water flow through porous rocks produces electric current, which in turn yields the magnetic field (*e.g.*, Fitterman, 1979a). Piezomagnetism, conversely, relies upon the observation that the magnetisation of titanomagnetite-bearing rocks varies under the application of mechanical stresses (*e.g.*, Nagata, 1970).

In general, volcanomagnetic anomalous fields can be considered the integrated effect of any of these sources distributed within the volcanic edifice. At a large basaltic volcano like Mt. Etna, that contains large amounts of magnetic minerals, large amplitudes of volcanomagnetic anomalies are predictable since magnetic minerals change their magnetisation when they are either subjected to heating (or cooling), or placed in a rapidly changing deviatoric stress field. We can conclude that, in principle, Mt. Etna is a good candidate site for testing the suitability of magnetic surveillance aimed at volcanic activity prediction. Since the historical time series provided by the present permanent magnetic network is still short (Del Negro *et al.*, 1994a), we searched for such a suitability with a four-year series of magnetic data recorded between 1979-1983 (Budetta and Pinna, 1979), which includes the major eruption of March 1981 (fig. 1).

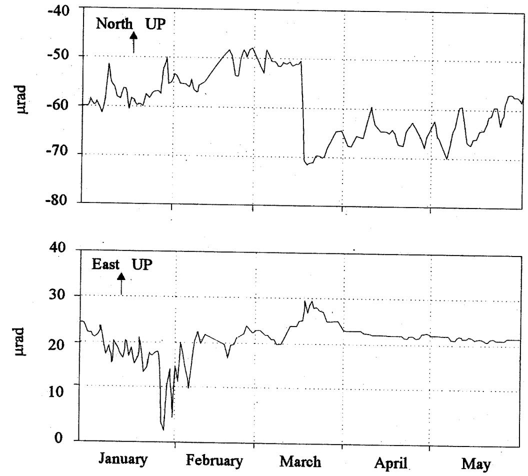

Fig. 2. Variations of the radial and tangential components of the bore-hole tiltmeter (TSF) located on the southern flank of Mt. Etna during 1981 (modified after Villari, 1983).

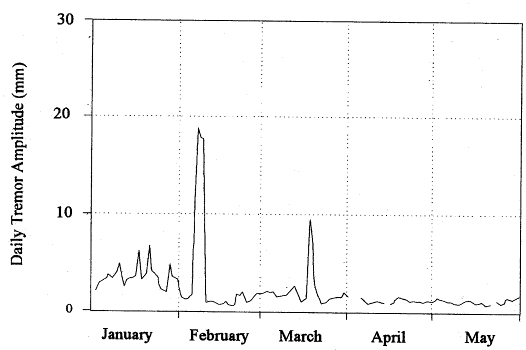

Fig. 3. Daily series of the tremor amplitude recorded at the Mt. Vetore (MVT) station during 1981 (modified after Tanguy and Patanè, 1984).

2. The 17-23 March 1981 eruption of Etna

Since early 1981 the volcanic activity of Mt. Etna was characterised by fairly strong ash emissions from the Bocca Nuova summit crater. In early February an intense explosive activity was observed at the NE crater, followed by a short-lasting lava effusion from it between February 5 and 8. This eruption was preceded by a sharp offset (fig. 2) of the East component (tangential) of the two-component tilt station (TSF) located on the southern flank of Mt. Etna (Villari, 1983), and accompanied by a rapid increase in the volcanic tremor amplitude (fig. 3) from February 5 to 7 (Tanguy and Patanè, 1984). No earthquake was recorded during this period (Gresta and Patanè, 1983).

After the end of activity at the NE crater, ash emission from the Bocca Nuova and the SE crater, and high volcanic tremor level were observed until the first days of March 1981 when, preceded by two days of intense seismic activity, an eruptive fissure system with approximately NW-SE trend opened on March 17 on the north-western flank of the volcano 2550 m a.s.l.

The fissure developed northwards quickly reaching 1400 m a.s.l., for a length of about 6 km overall. The eruption was accompanied by strong lava fountaining, and by an impressive lava flow that moved to lower altitudes at an average velocity of 1 km/h. The next day, the eruptive fracture extended by another 2 km, reaching 1120 m a.s.l.; however, most of the lava volume had already been emitted, reaching the Alcantara river 650 m a.s.l. and seriously threatening the town of Randazzo (fig. 1).

Eruptive activity continued sporadically until 23 March: at its end, the fissure extended for some 7.5 km between 2500 and 1100 m a.s.l. (Romano, 1981). Overall, a 5 m thick lava flow had covered a surface of about 6 km^2, leading to an estimated total volume of erupted lava between 18×10^6 m^3 (Murray, 1982) and 35×10^6 m^3 (Romano, 1981). If we consider that some 70% of lavas were erupted during the first 24 h, we can conclude that the average effusion rate in this period may have approached 300 m^3/s: this value is approximately one order of magnitude larger than, for instance, the rates of the 1989 and 1991-1993 flows (Barberi *et al.*, 1990, 1993), and can be included in the top values observed or measured in historical eruptions.

The picture of seismic activity that characterised early 1981 on Etna is loose, because of the lack of a sufficient number of stations for properly constraining the foci in space: then, the indicated focal depths (< 5 km in general, or < 1 km for weakest events recorded at one or two stations; Gresta and Patanè, 1987) are unreliable.

Conversely, the rate of seismic energy release is reliable. It indicates that more than 50 events per hour occurred next to the eruptive area immediately before the onset of the eruption, while the earthquake occurrence frequency sharply decreased after the eruptive fractures opened (Gresta and Patanè, 1987). Large tremor amplitudes were observed only during the first two days, in agreement with the observed paroxysmal onset of the eruption (fig. 3). The drop in seismic activity was qualitatively associated (Tanguy and Patanè, 1984) with «changes in the (local) stress field caused by the opening of new fractures or the closing/opening of pre-existing ones».

The seismic swarm was accompanied by a sudden sharp deformation episode observed at the tilt gauge TSF, installed 2500 m a.s.l. on the upper-southern flank of the volcano, about 8 km from the nearest end of the eruptive fissure (fig. 2). The pattern of the tilt shows a clear offset prior to the March eruption (on the East component, tangential) and an abrupt deflation of the radial component once the eruption began (North component, radial). This second phase of tilt correlates well with the tilt offset observed at the second tilt gauge installed at TNF, on the north-eastern flank (fig. 1). From this pair of tilt data Villari (1983) hypothesised that a dyke intrusion might have taken place in the north-eastern flank, along the NE rift of the volcano.

Gravity changes measured immediately after the eruption (Sanderson, 1982), in March-April 1981 indicated a gravity decrease in the central and upper southern flanks of the volcano, which was interpreted as the effect of magma drainage from a radial dyke 14 km long and $1 \div 5$ m wide, running at an elevation of 0-1000 m a.s.l. in a direction close to that of the eruptive fissure. During summer 1981, precise levelling and gravity surveys were carried out simultaneously. Upward movements of more than 17 cm were observed near the new fissure. Gravity data showed changes of up to $+63$ μGal, and displayed a good correlation with the elevation changes. In conclusion, since the ground uplift was accompanied by gravity increase (fig. 4a,b), both sets of measurements appeared to carry the signature of a new magma intrusion.

Furthermore, a zone of slight (approximately 2 cm) subsidence, then of possible deflation, was observed between 3-8 km to the east of the fissure. However, the volume of depression was insufficient to account for the volume of magma erupted. This led to uncoupling the data sets and drawing a two-source

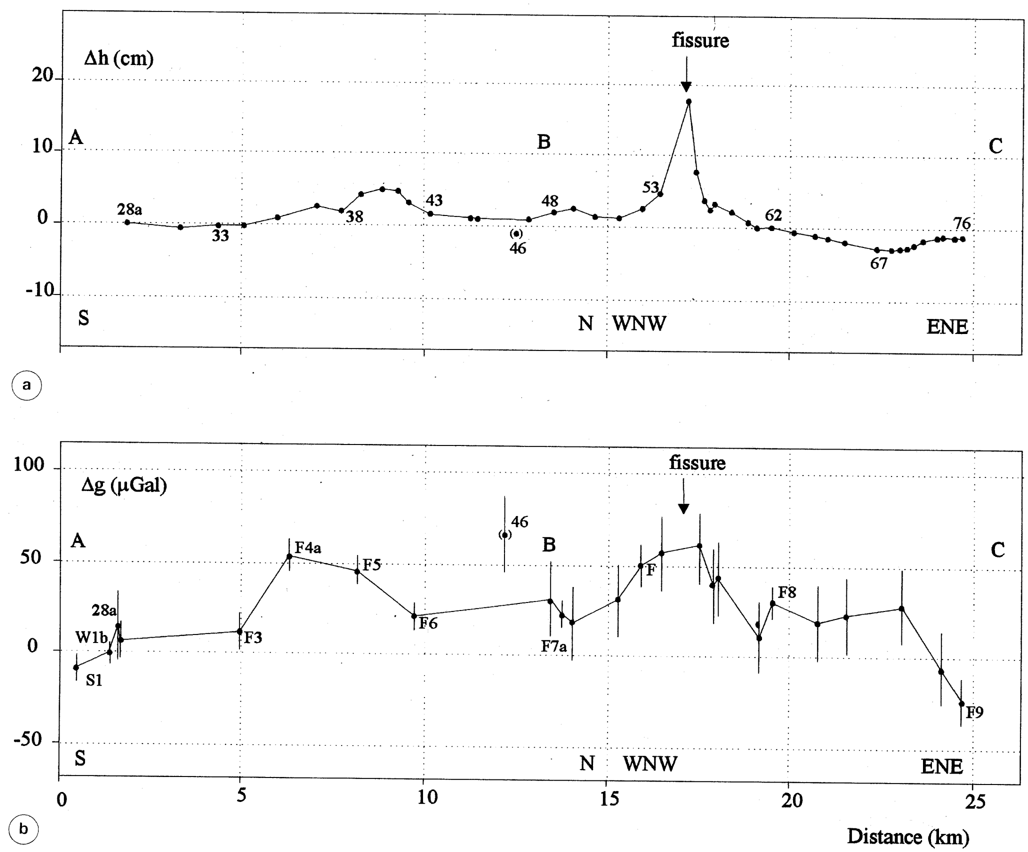

Fig. 4a,b. Plot of (a) elevation change in centimetres and (b) gravity change in μGal as a function of distance in kilometres along a profile across the eruptive fracture, whose position is marked by an arrow (modified after Sanderson *et al.*, 1983).

model, in which (fig. 4a) the ground deformations near the eruptive fissure were attributed to a dyke intruded 0.1-0.5 km below the surface, with strike along-fissure and dip between 75-90°, while (fig. 4b) the gravity changes were attributed to a 1.5-2.0 km deep intrusive dyke with the same strike (Sanderson *et al.*, 1983).

3. Volcanomagnetic signals

The array – Between 1979 and 1983 two continuously recording magnetic stations were run in eastern Sicily (fig. 1), aimed at testing the differential technique on Mt. Etna: one station was installed at Piccolo Rifugio (PCR) on the southern flank of the volcano, 2500 m a.s.l., while the second was located at Ramacca (RMC), about 40 km to the S-W of the summit craters (Budetta and Condarelli, 1983). The latter was aimed at providing the reference field external to the volcanic area.

Both stations were equipped with a proton precession magnetometer with 1 nT resolution, and measured the intensity of the Earth's magnetic field at a 5-min repetition rate. The data were radio-transmitted to the Istituto Internazionale di Vulcanologia in Catania, then recorded in punched-tape form. Even though this recording technique is now largely obsolete, and punched-tape supports have undergone material damage in the meantime, our attempts at retrieving the data set were successful and we recovered a complete series spanning five months across the March 1981 eruption.

Data Processing – Most of the published works on volcanomagnetic signals focus on simple simultaneous differences between the magnetic field amplitudes recorded at several points on a volcano. Generally, the difference is calculated with respect to a reference station located in a «magnetically quiet» site, that is, a site undisturbed by volcanomagnetic signals and with low local static gradient. Differences are averaged on hourly averages and daily mean values. Daily mean values of the differences allow typically reducing abnormal magnetic transients to approximately 1-2 nT. The amplitude of volcanomagnetic signals prior to the eruptions can locally exceed the reduced value of the abnormal transient variations by up to one order of magnitude: then this technique can be used with some degree of confidence.

Figure 5 plots the night-time and whole-day means of simple differences between PCR and RMC for the months of January and February, 1981. Data from 00h00 to 03h55 (LT) are used for the night-time mean (the magnetically quiet night hours): while those from midnight to 23h55 for the whole-day average.

We find a major oscillatory change in the whole-day mean. Day-to-day fluctuations in the whole-day mean difference values are partly due to local induction, and partly to the aliasing effect when full-time data are not available. One way to eliminate such noise is that of using the Wiener filtering method (*e.g.*, Del Negro, 1996).

It is worth noting that even the night-time mean data were not quiet enough to reduce the effects of local disturbances of the external field to less than 2-4 nT. In such case, mag-

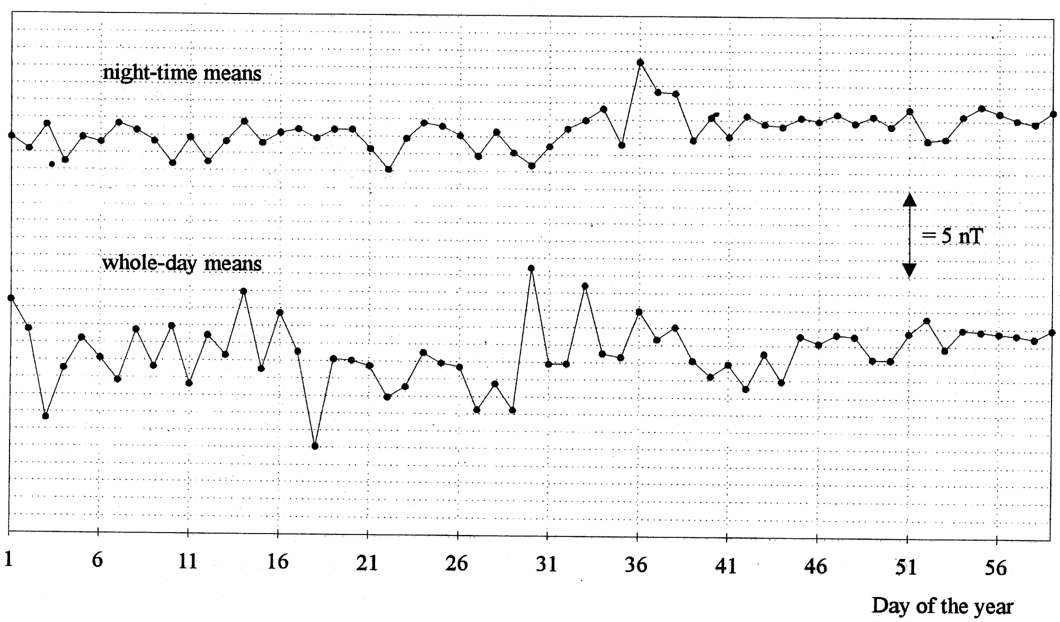

Fig. 5. Daily means of simple differences of the total intensity between PCR and RMC during January and February 1981. Top: night-time mean; bottom: whole-day mean.

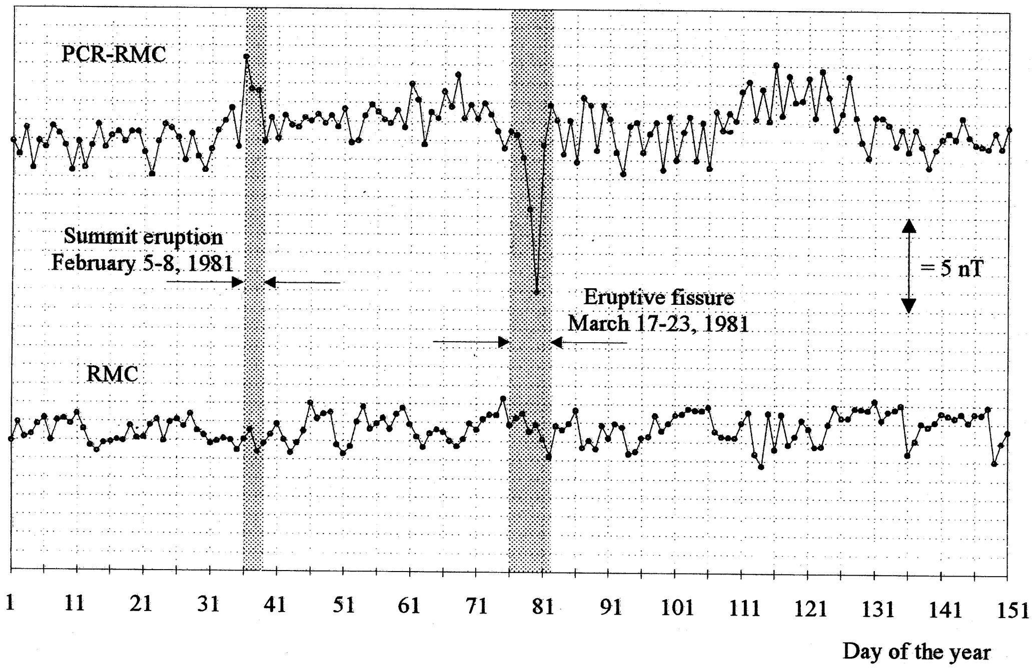

Fig. 6. Mean night-time values of the magnetic field at RMC and of differences PCR-RMC, between January and May 1981. Eruption is indicated by the shaded area.

netic changes associated to volcanic activity may not be identified from the night-time mean data only when variations are small. On another hand, on a basaltic volcano like Etna, magnetic variations arising from volcanic activity are expected to be large enough to be extracted from the data set. We discuss below the variation of the magnetic field only through the night-time mean data in the following.

Figure 6 reports the night-time mean value series (from January to May 1981) of the Earth's magnetic field intensity at RMC (reference station), and the differences between field intensities PCR- RMC. At RMC, no appreciable change exceeding 2-3 nT was detected during the whole period. Conversely, a large change in the PCR-RMC differences was observed when the eruptive fissures opened.

In the absence of volcanic activity, the differences had a standard deviation $\sigma = 1.7$ nT, increasing to $\sigma = 3.7$ nT during activity periods. These changes were reversible, the differences returning to pre-eruption values in a very short time.

The sharpest magnetic transient occurred shortly after the onset of the eruption between 19 and 21 March. It lasted three days, reaching the largest amplitude slightly more than 10 nT. This is one of the top-values among those reported in magnetic literature (table I): its real magnitude on-site might also be larger, once one considers that it was observed at a station offset 7 km from the site of the eruptive events.

Spectral analysis – Another method for the detection of the volcanomagnetic transients is that of analysing the spectral amplitude changes in signals recorded at permanent magnetic stations. An amplitude increase in power spectrum, caused by a significant change in the local magnetic field, is expected to occur in the

Table I. Transient magnetic anomalies observed to date at active volcanoes. Except for thermomagnetic signals, that may reach several tens of nanoteslas (Yukutake *et al.*, 1990; Hamano *et al.*, 1990; Tanaka, 1993), other observed volcanomagnetic signals are always less than 15 nT (Davis *et al.*, 1979, 1984; Pozzi *et al.*, 1979; Johnston *et al.*, 1981; Zlotnicki and Le Mouël, 1988).

La Soufrière, Guadeloupe, French West Indies	$\Delta B_{max} \approx 20$ nT	Mid-length transient associated with activity sources at crustal depth (Pozzi *et al.*, 1979)
Mt. St. Helens, Washington State, U.S.A.	$\Delta B_{max} \approx 10$ nT	Possible transient associated with one eruptive event (Davis *et al.*, 1984)
Piton de la Fournaise, Réunion Island	$\Delta B_{max} \approx 20$ nT	Several transients associated with eruptive episodes (Zlotnicki and Le Mouël, 1988)
	$\Delta B_{max} \approx 40$ nT	Magnetic field variations associated with the whole eruptive period (Zlotnicki *et al.*, 1993)
Izu-Oshima, Japan	$\Delta B_{max} \approx 20$ nT	Decrease of magnetic field prior to the main eruption (Yukutake *et al.*, 1990)
	$\Delta B_{max} \approx 10$ nT	Transient associated with the eruptive event (Hamano *et al.*, 1990)
Aso, Japan	$\Delta B_{max} \approx 30$ nT	Magnetization changes generated from temperature variations (Tanaka, 1993)

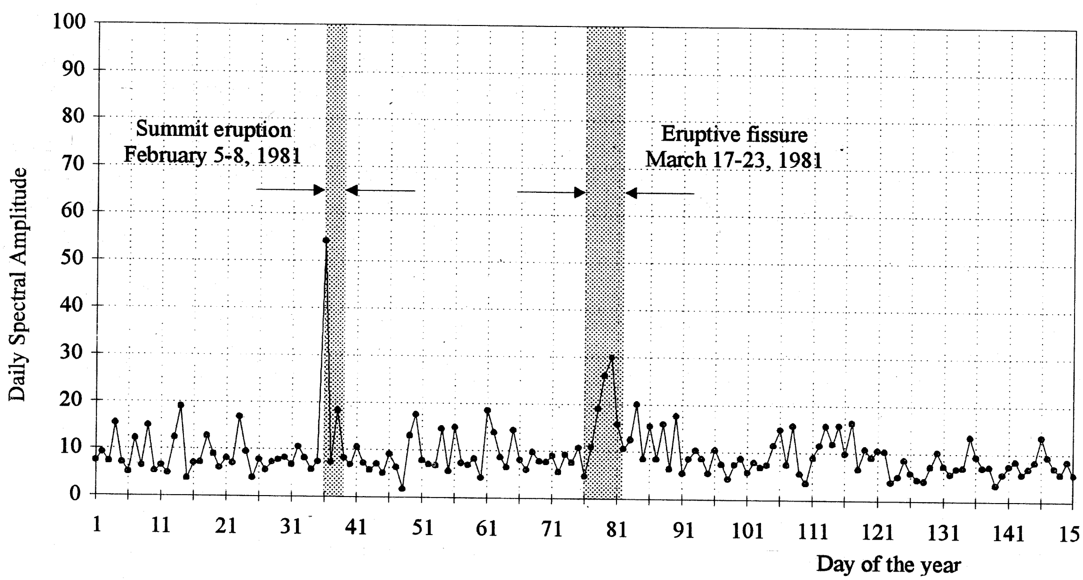

Fig. 7. Variations of daily spectral amplitudes of the magnetic signal at Piccolo Rifugio (PCR) during January and May, 1981.

presence of volcanic activity. This method looks suitable for time series in which the external noise reaches high power values. In such case, the daily averages are affected by a large statistical error, and the simple difference method for the detection of the volcanomagnetic transients becomes quickly ineffective. In our case, we removed mean value offsets from the observed data and calculated the daily spectra. The latter always show a main peak of variable amplitude at low frequencies, and near-nil amplitude at high frequencies.

Figure 7 shows the daily maximum peak of the spectral amplitude changes with time observed at PCR station, plotted between January and May. These show a complex trend characterised by a relatively low average, where it is possible to notice a distinct peak during the first days of February (6-7) and a significant increase in the mean level taking place with the eruption of March. The isolated peak observed in February is associated in time with the intense explosive activity from the NE crater. It is worth noting that the amplitude spectral trend of the magnetic signal observed at PCR station is well correlated with the trend of the daily tremor amplitude at Mt. Vetore (MVT) seismic station during the whole period (fig. 3).

Even though the present test is carried out on one station only, this result highlights the potentiality of spectral methods in discriminating volcanomagnetic transients.

4. Modelling and interpretation

We have shown that a significant change in the local magnetic field was correlated with the opening of eruptive fissures on Mt. Etna between March 17-23, 1981. This change might have been generated by different mechanisms, *i.e.*: i) thermomagnetism; ii) electrofiltration; and iii) piezomagnetism.

4.1. *Thermo-magnetic processes*

The thermomagnetic effect may play an undoubted role on active volcanoes, where large amounts of thermal energy concentrate and dissipate in short time lapses through fumarole activity and emission of lava and ash. Rocks subjected to temperature change are also subjected to demagnetisation and/or remagnetisation. When the temperature exceeds the Curie point, the rocks lose their magnetisation and consequently modify the static crustal magnetic field (Rikitake, 1952; Rikitake and Yokoyama, 1955). Conversely, emplaced lava flows or pyroclastic deposits that cool below their Curie temperature, acquire a thermoremanent magnetisation that is related to the intensity and direction of the Earth's magnetic field (*e.g.*, Stacey and Banerjee, 1974).

Thermomagnetic effects strongly depend on the content and type of rock-forming ferromagnetic minerals (*e.g.*, Nagata, 1943; Rikitake and Yokoyama, 1955). Titanium-rich magnetites (Carter, 1976; Pozzi, 1977), characterised by Curie temperatures ranging from 200 to 300°C, and acquiring a great deal of NRM in the interval between 100-200°C (Nagata, 1961), were found in Etna's lavas.

In such minerals, small variations in temperature in this interval can produce notable changes of magnetic field. Emeleus (1977) calculated that thermal oscillations of only 1°C in a limited volume of basaltic rock such as a plug, relatively cold and close to the observation point, can produce magnetic field variations of some 25 nT.

If the Etna volcano partially loses its magnetisation by heating the materials beneath the almost permanently active central craters, the total intensity is expected to decrease to the south of them and to increase to the north. During the 1981 eruption, although there was only one observation point, a negative zone of field change could be invoked as extending from the fracture to the southern flank of Mt. Etna. The PCR station is located some 7 km south of the eruptive fissures. If the observed magnetic changes were entirely brought about by the thermal effect, the thickness of a nonmagnetic dyke should have been of several tens of meters: this value is not consistent with the mechanical estimate of the width of the opening crack (a few meters at most, after Villari, 1983).

Table II. Parameters of the modelled dyke.

Dyke length	8.0 km	Stress sensitivity	1.0×10^{-3} MPa^{-1}
Dyke thickness	0.5 km	Lamé's constants	3.0×10^{-4} MPa
Dislocation (crack opening)	3.0 m	Average magnetisation	9.0 A/m
Depth of burial	1.0 km	Declination	1.2°
Dip angle	90°	Inclination	53.1°
Strike	N30°W		

We computed the anomaly field produced by a non-magnetised prismatic body using Bhattacharyya's (1964) formula for a rectangular prism. The geometry of the intrusive body is taken from the models of Sanderson (1982) and Sanderson *et al.* (1983), based on vertical displacement and gravity change observations. The model parameters and the medium constants are summarised in table II. At the PCR magnetic recording site, the amplitude of the anomaly field was less than -1 nT (-0.7 nT), as shown in fig. 8.

The question is if the discrepancy between the calculated and the observed (> 10 nT) values can be ascribed to heating of the wall materials within several tens of meters from the vent. If the magma heated the dyke wall more effectively, the absolute field values must have increased in proportion to the thickness of the demagnetised region. The question arises, therefore, how can heating of such a large volume take place quickly.

When we assume a thermal conduction mechanism, the heating efficiency of the surrounding rocks is not sufficient. If we take the thermal diffusivity of basaltic rocks to be 0.01 cm^2/s (corresponding to 31.536 m^2/year; Carslaw and Jaeger, 1959) the region effectively heated by conduction would be very limited indeed, and the observed change (10 nT at a distance of 7 km from the vent) would take place in years and not in hours.

A more efficient process is that of heat transfer by fluids through pores and cracks in the rocks, which looks possible once one assumes that a volcano must contain large cracked (or fissured or faulted) volumes, and that pre-eruptive and eruptive stresses increase the overall porosity. In general, the emplacement of a dyke can lead to extensive convective thermal exchanges, and large volumes nearby can be demagnetised very quickly. Where parts of the edifice are water-saturated, both long-term and rapid volcanomagnetic variations are observed (Yukutake *et al.*, 1990; Sasai *et al.*, 1990; Hamano *et al.*, 1990). Also in this case, however, even in the presence of diffusivity rates two-three orders of magnitude larger than those observed at rest, it is impossible to fit both swiftness and amplitude of the observed anomaly.

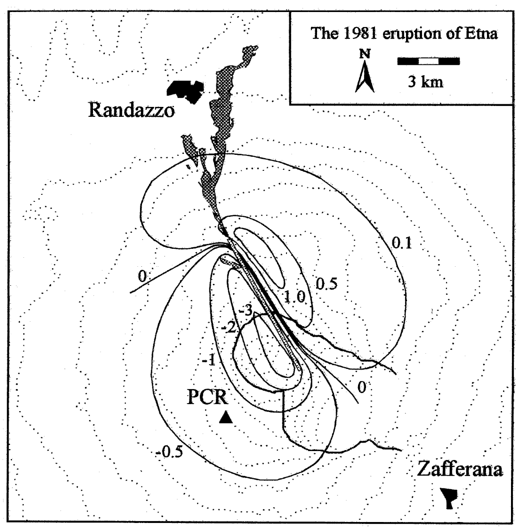

Fig. 8. Magnetic anomaly caused by a non-magnetic dyke. Units in nanoteslas.

We can conclude that the observed magnetic change cannot be interpreted by a single thermal process, and some other mechanism must be sought.

4.2. *Electrokinetic phenomena*

The observed time scale of volcanomagnetic transients associated with the 1981 eruption brings into play the electrokinetic effect. Anomalies of this type show modest maximum values but, with respect to thermomagnetic changes, they are fast.

Electrokinetic effect is related to water circulation along cracks and faults or through the interconnected pore network of rocks (Nourbehecht, 1963). Favourable conditions occur when rainfall is abundant, water tables are present, and interconnected cracks and fissures are more or less water-saturated. In addition, gas or magma flow may significantly intervene in the generation of electrokinetic phenomena. The time constants of the electrokinetic field are determined by changes of the stress field (as for piezomagnetism) and also by thermal or chemical modifications of fluids within the interconnected crack and fissure nets. Electrokinetic signals tend to be focused along channels where fluids flow, and may be associated with ground deformation (thus, with contemporary piezomagnetic effects): unlike piezomagnetic effects, however, the electrokinetic ones are always associated with variations in the electric field (streaming potential); the ratio of the voltage to the pressure drop along the flow path is called the electrokinetic coupling coefficient.

The streaming potential is linked to the existence of an electric double layer at the solid-liquid interface. It is constituted of a layer of ions firmly attached to the solid wall, and of a mobile diffusive layer of ions of opposite sign extending in the liquid phase (*e.g.*, Fitterman, 1979a; Murakami, 1989):

$$E = \frac{\varepsilon\zeta}{\eta\sigma}\nabla P \tag{4.1}$$

where ε, η and σ are, respectively, the dielectric coefficient, the viscosity and the conductivity of the fluid, and ∇P is the fluid pressure gradient (over hydrostatic pressure). The potential ζ is the potential difference between the solid-liquid interface and the bulk of the liquid.

When a pore pressure gradient is applied, a fluid flow drags the positive ions of the fluid phase. The corresponding electrical current density $\boldsymbol{j}$ is given by (Mitzutani *et al.*, 1976; Fitterman, 1978):

$$\boldsymbol{j} = -\frac{\Phi\varepsilon\zeta}{\eta}\nabla P = C\sigma\nabla P \tag{4.2}$$

where Φ is the porosity of the medium, and C the streaming coefficient ($C = -\varepsilon\zeta\Phi/\eta\sigma$).

At any time t, the electrokinetic magnetic field $\boldsymbol{B}_{VE}(Q,t)$ at the surface of a volcano is given by Biot and Savart's law:

$$\boldsymbol{B}_{VE}(Q,t) = \frac{\mu_0}{4\pi}\nabla\times\left[\iiint_{VE}\frac{\boldsymbol{j}(M,t)}{r}\,dv_M\right] \tag{4.3}$$

with Q the observation point and $\boldsymbol{j}(M,t)$ the electrokinetic current at the time t and in a point M of the volume interested by the water circulation V_E.

Several authors have calculated the electrokinetic magnetic field at the surface in the frame of seismotectonic effects (Mitzutani *et al.*, 1976; Fitterman, 1978, 1981; Murakami, 1989). In all cases, the order of the magnitude of the electrokinetic magnetic field can reach 10 nT or more for a pore pressure gradient of 1 MPa/km.

For example, Fitterman (1979b) gave analytical solutions of electric and magnetic fields for an electrokinetic source made of a rectangular vertical fault separating two media with different streaming potential coefficients, C_1 and C_2. The source intensity, $(C_1 - C_2)P$, has a constant value inside this rectangle and is zero outside. This model can be thought of as a region along the fault from which fluid is flowing ($P > 0$) or toward which fluid is flowing ($P < 0$). Fitterman shows that for pore pressure changes (in the range of 1-10 MPa) produced by tectonic stress variations and for reasonable values of C_1 and C_2 (1-100 mV/bar), the depth to the top of the source must be less than 1 km for a detectable magnetic signal (> 1 nT) to be produced.

To assess the order of magnitude of a possible electrokinetic magnetic field on Mt. Etna, we will use Fitterman's model. Let us represent the fractured zone by a rectangular fault 8 km long, 0.5 km high, with its upper limit at 1.0 km beneath ground surface, and let the source intensity $V = (C_1 - C_2)P$ be 1 V.

At station PCR, 7 km away from the fractured zone, the corresponding magnetic field (projected along the main field $\boldsymbol{B}_p$) is less than 1 nT. This result, together with the observations that the source of magnetic anomaly was most probably deep (large distance between the eruption site and the recording station), and the clayey horizon limiting the aquifers on northern Etna is very shallow (a few hundred meters only beneath the free surface), indicates that it is difficult to explain such large observed magnetic anomalies by the electrokinetic coupling.

It is worth noting further that magnetic field changes produced by the electrokinetic motion must bear an associated electric field: unfortunately, no electric data were available to be jointly discussed and modelled with the magnetic changes, and the exclusion of the electrokinetic effect from the candidate magnetic source processes relies entirely upon the indirect field and geological evidence discussed above.

4.3. *Piezomagnetic effects*

In case of magma intrusion, large changes in the local stress and strain fields can drive magnetic changes (due to the piezomagnetic effect) and displacement of magnetic materials (*e.g.*, Johnston and Stacey, 1969; Zlotnicki and Le Mouël, 1988; Del Negro *et al.*, 1994b).

The fundamental concept of stress-induced volcanomagnetic effect was first proposed by Stacey *et al.* (1965) who, having assigned a stress distribution around a magma chamber, computed a theoretical piezomagnetic anomaly. The linear relationships established by Stacey (1964) between applied stress σ and magnetisation changes $\Delta\boldsymbol{J}$, were reduced by Sasai (1980) to the following simple formula:

$$\Delta\boldsymbol{J} = \frac{3}{2}\beta\boldsymbol{T}'\boldsymbol{J} \tag{4.4}$$

where β is the stress sensitivity. $\boldsymbol{T}'$ is the deviatoric stress tensor, which is related to the stress tensor $\boldsymbol{T}$ and to the average stress σ_0 as:

$$\boldsymbol{T} = \sigma_0 + \boldsymbol{T}' \tag{4.5}$$

where $\sigma_0 = (\tau_{xx} + \tau_{yy} + \tau_{zz})/3$.

Schematically, the stress field is decomposed into a hydrostatic and a deviatoric component. The piezomagnetic effect would derive from the latter, provided that the temperature of the embedment is below the Curie point. Prior to an eruption, it is to be expected that magmatic activity may lead to an increase in the deviatoric stress, and to the consequent increase in the local magnetic field, whereas the stress release occurring during an eruption would produce a decrease in the magnetic field (Parkinson, 1983).

The piezomagnetic field ($\boldsymbol{B}_{VP}$) at the ground surface of a volcano can be estimated (Zlotnicki and Le Mouël, 1988) by:

$$\boldsymbol{B}_{VP}(Q, t) = -\frac{\mu_0}{4\pi}\nabla\left[\int\int\int_{V_P}\Delta\boldsymbol{J}(M, t)\frac{\boldsymbol{r}}{r^3}\,dv_m\right] \tag{4.6}$$

where μ_0 is the magnetic permeability in the vacuum, Q is the observation point, M is a point in the rock volume V_p submitted to a stress field, and r is the distance between M and Q.

The solution of integrals in eq. (4.6) for a piezomagnetic field caused by intrusion of a vertical rectangular dyke is given by Sasai's (1980) formula. Application of this formula requires an estimate of the value of three parameters. The intensity of the piezomagnetic effect is proportional to the product of the stress sensitivity (β), the average magnetisation (J) and the average rigidity of the Earth's crust (λ). According to Carmichael (1977), who assessed the upper crust to be characterised by approximately constant stress sensitivity, we estimated β of the order of 10^{-3} MPa^{-1}; while for the average crustal strength we here adopt a value for estimating the seismic moment of shallow earthquakes (Patanè *et al.*, 1994).

As for the average magnetisation of Etna, only a few data are available. The average magnetisation was evaluated at 9.0 A/m on the grounds of some magnetic surveys (*e.g.*, Del Negro *et al.*, 1994a) and a few laboratory tests (Roberti and Scandone, 1975; Pozzi, 1977). The estimate of average magnetisation is probably not too large for volcanoes consisting of basaltic rocks where the titanomagnetite is abundant (also 10% modal by weight; Pozzi, 1977). As for the products erupted during the 1981 eruption, their composition does not differ from that typical of Etnean lavas (Villari, 1983; Cosentino *et al.*, 1981).

Figure 9 shows the computed piezomagnetic change associated with an intrusive dyke, and relates to the entry values given in table II: very large decrease in intensity of the magnetic field above the dyke contrasts with the positive changes occurring at its northern tip and to the south-west of it.

The computed anomaly module is in agreement with the variation of approximately 10 nT recorded at station PCR, but the large anomaly peak (– 100 nT) obtained above the dyke seem in principle rather exaggerated, being it known that most of the hitherto observed piezomagnetic effects are of roughly an order of magnitude lower. Banks *et al.* (1991) found Sasai's (1980) solution yields a contradictory result for a particular configuration of the fault: «if the slip surface intersects the ground surface, the piezomagnetic field is predicted to increase without limit as the fault trace is approached». The discrepancy was noted between numerical and analytic solutions in some piezomagnetic models. Subsequently, Sasai (1991b) re-examined its solution for the piezomagnetic field to correct some of the previous results and gave formulas for uniform strike-slip and tensile faulting along a vertical rectangular fault. Recently, these contradictions

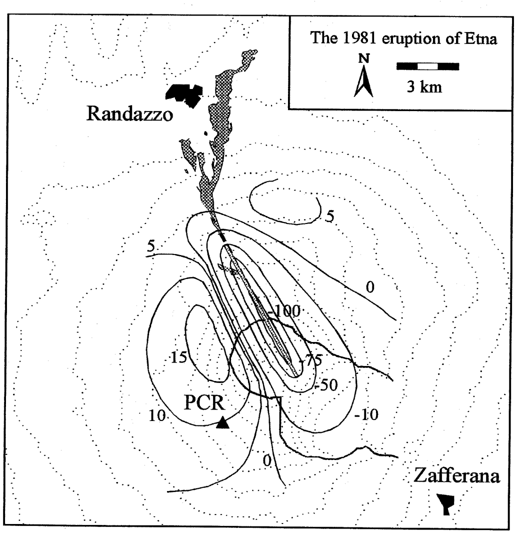

Fig. 9. Piezomagnetic anomaly field produced by an intrusive dyke computed using the Sasai's (1980) model. Units in nanoteslas.

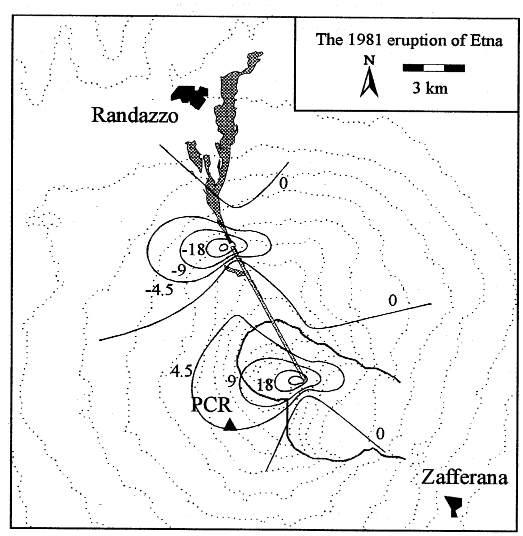

Fig. 10. Piezomagnetic anomaly field produced by an intrusive dyke computed using the Sasai's (1991b) model. Units in nanoteslas.

were also discussed by Sasai (1994a,b) himself.

Using Sasai's (1991b) solutions, it was recomputed the piezomagnetic field associated with a tensile fault (*i.e.*, an intrusive dyke). Magnetic changes are computed with the same model parameters as given in table II: with the new Sasai's model the relatively intense magnetic changes are seen only around both tips of the dyke (fig. 10). However, this new result does not justify completely (about half) the variation observed at the station measuring on the volcanic edifice. Moreover, the change observed at station PCR is negative, whereas the piezomagnetic anomaly computed here is positive.

5. Discussion and conclusions

We have demonstrated above that, having assigned the source-receiver geometry, piezomagnetism is the only way to explain most of the amplitude of the volcanomagnetic transient observed in March 1981 on Etna.

However, we encounter major difficulties in explaining the total intensity decrease at PCR station in terms of piezomagnetism due to positive stress change. According to piezomagnetic calculations based on the Mogi's (1958) model (Sasai, 1991a), the increase in hydrostatic pressure within a spherical source beneath the crater gives positive and negative magnetic anomalies on the south and north parts of the latter, respectively.

Pressurisation in the piezomagnetic model corresponds to magnetisation in the thermomagnetic model: then, depressurisation correlates with demagnetisation. In the thermomagnetic model (see fig. 8) the shape of the anomaly due to constant-geometry bodies does not depend on the susceptibility contrast, though reversal of the susceptibility contrast determines swapping of the poles of the anomalous dipole.

Since further computations cannot contribute to improving a model that can be checked at only one recording point, we can preliminarily conclude, from a qualitative standpoint, that (i) the module and the spatial extent of the volcanomagnetic transient we are dealing with are fully accounted for by a pure piezomagnetic model, while (ii) the sign is best explained by negative stresses acting on the rocks embedding the model dyke.

In other words, collapse of the dyke walls following a sudden decrease of the flow feeding the eruptive events, might have induced a decrease of the stress-induced magnetic field. This change was fast enough to prevent slow recovery of the stress in the embedment since, in the short term, the fields observed before and after the event are the same (within less than 2 nT). As a whole, the magnetometer at PCR worked as a sort of strainmeter.

This thesis is independently supported by other observations and by the course of the anomaly with time. Indeed: a) the magnetic event observed at station PCR was accompanied by sharp deflation of the radial (N-S) component of the tilt station TSF, also located on the upper southern flank of the volcano; and b) after two days of huge lava emission rate (up to 300 m^3/s) both the magnetic and the tilt events marked the transition towards the minor effusive rate of a few m^3/s that lasted until the end of the 1981 eruptive event on March 23.

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