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MAGNETIC MEASUREMENTS ON LA SOUFRIERE VOLCANO, GUADELOUPE (LESSER ANTILLES), 1976—1984: A RE-EXAMINATION OF THE VOLCANOMAGNETIC EFFECTS OBSERVED DURING THE VOLCANIC CRISIS OF 1976—1977

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ABSTRACT

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During the seismovolcanic crisis of 1976—1977 at La Soufrière on Guadeloupe, a magnetic network of 12 reference markers was set up. Measurements of the intensity of the earth's magnetic field, carried out up to once a day at each marker, showed volcanomagnetic variations of several nanoteslas (nT). The variations, at certain markers, were more or less concealed by transient magnetic variations due to anomalies in conductivity. As early as 1978, measurements were resumed and a telemetering network was coupled with the network of reference markers, to which 4 new markers were added. A detailed study of conductivity anomalies was carried out on the entire volcano. Contrasts in conductivity linked to the existence of a superficial conducting surface, on a SSW/NNE axis, located south of the volcano, caused a great lack of homogeneity in the field variations measured at the surface. Variations greater than about 10 nT appeared in the difference in intensity of the earth's magnetic field between two stations.

No long-term magnetic variation was observed between 1978 and 1984. On the network of markers, the accuracy of measurements of volcanic effects was at best 2 nT. Measurements carried out on the telemetering network during the night refined these results, since their accuracy was 1 nT. The only significant volcanic crisis between 1978 and 1984 (5—7 January 1981) seems to be observed by telemetering stations. All the measurements carried out in periods of volcanic inactivity make it possible to re-examine the crisis of 1976—1977. Though volcanomagnetic effects over short periods cannot be accurately determined, variations with a time constant of several weeks were present over the entire volcano. These variations were as high as 7—8 nT in remote stations and they can be linked to the three major phases of eruptive activity at La Soufrière during the crisis of 1976—1977.

1. INTRODUCTION

The main studies relating to the establishment of variations in the earth's magnetic field linked to volcanic activity have been conducted on the following volcanoes: Oshima in Japan (Rikitake, 1951; Yokoyama, 1957; Uyeda, 1961), Ruapehu and Ngauruhoe in New Zealand (Johnston and Stacey, 1969a, 1969b), Kilauea on Hawaii (Davis et al., 1979), Mount St. Helens in the United States (Johnston et al., 1981, Davis et al., 1984) and La Soufrière on Guadeloupe, in the French Antilles (Pozzi et al., 1979). The observed anomalies vary between 1.5 nT (Davis et al., 1979) and ca. 10 nT (Johnston and Stacey, 1969b; Pozzi et al., 1979; Johnston et al., 1981). The relation between magnetic anomaly and volcanic activity is not always clear. Sometimes, on the same volcano, no magnetic anomaly is observed during an eruptive phase, while it can be detected during another phase (Johnston and Stacey, 1969b; Pozzi et al., 1979). In the same way, magnetic anomalies are sometimes observed without any apparent connection with volcanic activity.

All these studies emphasize that the major difficulty in the search for possible volcanomagnetic effects lies not so much in the sensitivity of the magnetometers used — most often, these are proton magnetometers with a sensitivity of 0.25 nT or 0.1 nT — as in the reduction of data (Beahm, 1976; Davis et al., 1979; Pozzi et al., 1979; Johnston et al., 1981; Zlotnicki et al., 1986). Local magnetic effects (anomalies in conductivity, contrasts in induced magnetization of magnetized structures) may induce differences in the variations of the earth's magnetic field that are greater than the volcanomagnetic effect that is sought at stations only a few kilometers away (Le Mouël et al., 1984; Zlotnicki et al., 1986).

La Soufrière volcano on Guadeloupe is one of many volcanoes located along the Recent arc of the Lesser Antilles (Martin-Kaye, 1969; Briden et al., 1979). It is an andesitic volcano of explosive character where, over the past 200,000 years or so, there has been an alternation of eruptions of extrusive and explosive character. Since the last magmatic eruption, dated 1450 ± 50 A.D. (Semet and Vatin-Perignon, 1979; Vincent et al., 1979; Semet et al., 1982), several phreatic eruptions have taken place, the most substantial of which occurred in 1797, 1956, and 1976—1977 (Lacroix, 1904; Robson and Tomblin, 1966).

In this geological and historical context, after the initial phreatic eruptions of July, 1976, a magnetic network of 12 markers was set up that covered the entire volcano (see Fig. 1 and Tables 1 and 2) (IPG, 1976; Pozzi et al., 1979; Feuillard et al., 1983). At each of the markers P, measurements were made every 1—2 days. The intensity of the earth's magnetic field, at an instant t, was reduced by measurement at the reference station O which continuously measured the intensity of the earth's magnetic field (Matouba: MTB, Fig. 1) (see paragraph 2b). To each of the reduced values $\Delta B(P, \mathrm{O}, t)$, a margin of error was associated. This error was evaluated on the basis of the projection and reduction errors (Pozzi et al., 1979) that characterize abnormal transient variations of the magnetic field for periods of a few hours. The main results are the following:

(a) at several markers, the differences $\Delta B(P, \mathrm{O}, t)$ are greater than the calculated margins of error;

(b) for periods of a few days, the markers located near the volcano show only variations of low amplitudes: less than 5 nT. At some markers further away (essentially DOL, HBT, GDM) (Fig. 1), the variations reach a level of about 15 nT in a few days;

(c) the time constants of the observed variations are very variable. They oscillate between a few days (and the variations are poorly described because of the sampling of the measurements) and a few weeks (the corresponding variations then have a lower amplitude than the variations over shorter periods).

Beginning in June, 1978, differential magnetic measurements were again performed on the mountain area in the same configuration. During this period without any volcanic activity, a detailed work on reduction errors was conducted. It appeared that, even for markers located far from the volcano, variations of the transient field (periods ranging from a few minutes to 24 hours such as the daily solar variation SR) do not disappear completely in the reduction of the data (Le Mouël et al., 1984). Temporal variations $\Delta B(P, \mathrm{O}, t)$ may even be as high as 15 nT in the course of a single day at certain points in the mountain area, these variations being attributable only to the magnetic agitation of the particular day.

Adding a differential magnetic network for the continuous measurement of the intensity of the earth's magnetic field, in which reduction errors are low, to a 16-marker network (Table 2), in which reduction errors are well-known, makes it possible to study over a long period the background noise specific at each marker and to evaluate any long-term magnetic variations. The results observed during the seismovolcanic crisis of 1976—1977 can be re-interpreted in the light of those obtained between 1978 and 1984.

2. MAGNETIC NETWORKS AND VOLCANIC MONITORING

The earth's magnetic field, at a point P and an instant t, is the sum of the following (IPG, 1976, 1977; Pozzi et al., 1979; Zlotnicki, 1979):

(a) A regular field of deep origin $\vec{B}_{\mathrm{r}}$. This field is subjected to the slow century-long variation, the temporal and spatial variations of which are respectively 120 nT/year and 6 nT/km. It remains homogeneous over the area covered by a magnetic network (20 km by 20 km).

(b) A field of local anomalies $\vec{B}_{\mathrm{a}}$ generated by the various magnetized structures. We define $\vec{B}_{\mathrm{a}}$ as the constant field obtained in the absence of any volcanic activity.

(c) A transitory field $\vec{B}_{\mathrm{t}}$, the primary sources of which are located in the ionosphere or beyond (i.e. an altitude greater than 100 km). This field, which is variable in time and space, induces abnormal transient magnetic variations through the intermediary of geological structures of different conductivity. These variations are not homogeneous and are sometimes intense (more than 50 nT).

(d) A possible volcanomagnetic field $\vec{B}_v$, which is the field generated by various possible sources. We have:

$$\vec{B}(P, t) = \vec{B}_r(P, t) + \vec{B}_a(P) + \vec{B}_t(P, t) + \vec{B}_v(P, t) \quad (1)$$

In order to eliminate the variations of the main field $\vec{B}_p$ (the sum of $\vec{B}_a + \vec{B}_r$) and the transient field $\vec{B}_t$, we study the temporal variations of the difference $\Delta B(P, \mathrm{O}, t)$ between the intensities of the fields measured at the points P and O at instant t:

$$\Delta B(P, \mathrm{O}, t) = B(P, t) - B(\mathrm{O}, t) \quad (2)$$

Using the formal notation proposed by Pozzi et al. (1979) it can be shown that relation (2) can be put into the following simplified form (Zlotnicki et al., 1986):

$$\begin{aligned}\Delta B(P, \mathrm{O}, t) = {} & [\vec{B}_v(P, t) \cdot \vec{u}(P) - \vec{B}_v(\mathrm{O}, t) \cdot \vec{u}(\mathrm{O})] + [\vec{B}_t(P, t) \cdot \vec{u}(P) \\ & - \vec{B}_t(\mathrm{O}, t) \cdot \vec{u}(\mathrm{O})] + [B_p(P, t) - B_p(\mathrm{O}, t)] + \epsilon(B_v/B_p) \\ & + \epsilon'(B_t/B_p)\end{aligned} \quad (3)$$

$\vec{u}$ is the unit vector carried by the main field $\vec{B}_p$. It may be considered constant because of the slowness of the century-long variation. The first term at the right-hand side of the equation is the difference between the volcanomagnetic fields between two stations $\Delta V(P, \mathrm{O}, t)$; that is the term that should be eliminated. The second term represents the reduction error $E_r(P, \mathrm{O}, t)$ of the data due to contrasts in conductivity between two stations. The third term characterizes the difference between the main fields and remains constant in the absence of any anomaly in the century-long variation. Finally, ϵ and ϵ' are infinitesimal quantities in B_v/B_p ($\leqslant 15/39{,}000$) and B_t/B_p ($\leqslant 50/39{,}000$) (at Guadeloupe).

2b. Magnetic networks

In 1978, three telemetering stations measuring the intensity of the earth's magnetic field (MTB, SAV, PAL) were associated with the 12-marker network (Fig. 1). The reduction station (MTB) was set up less than 200 m from the site used between September 1976 and April 1977 by the Laboratoire d'Electronique et de Technologie de l'Informatique (L.E.T.I.) of the Atomic Energy Commission (Robach et al., 1976). At each station, measurement of the magnetic field is performed every minute on the minute (Geometrics magnetometers 1/4 nT) and communicated in digital form to the observatory (FSC) after a set interval of time. At the observatory an analog recorder (tapespeed 6 cm/h; sensitivity 1 nT/mm) reproduces the intensity of the magnetic field at reference station O (MTB) and the simultaneous differences between the Palmiste and Matouba stations, ΔB (PAL–MTB), and the Savane and Matouba stations, ΔB (SAV–MTB). The Matouba telemetering station was chosen as reference station because it was far from the sea (this

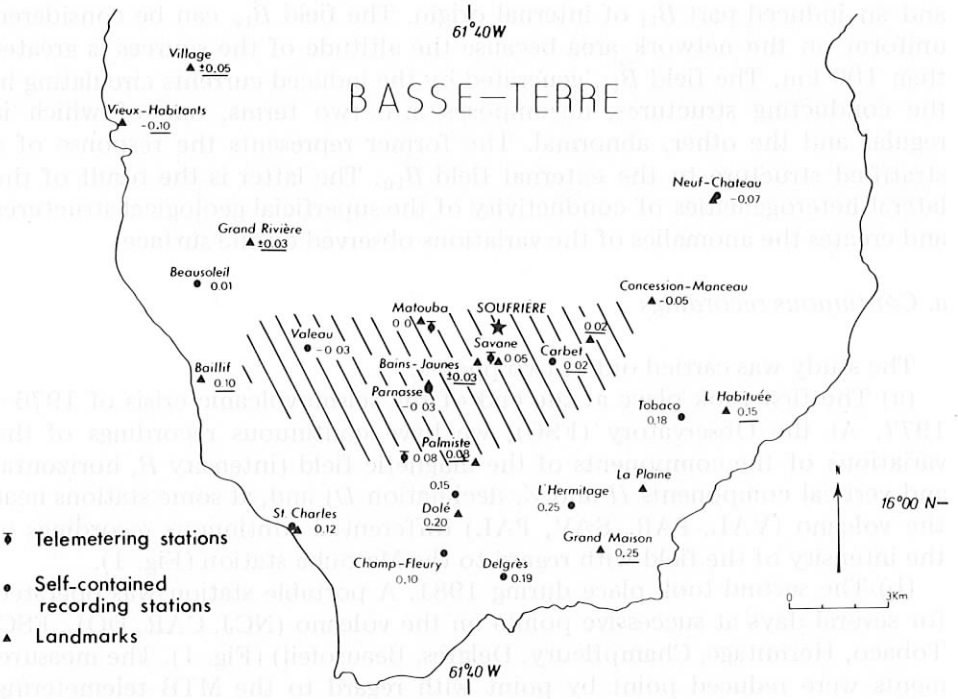


Fig. 1. Magnetic network and associated β_{po} coefficients. The hatched zone is the region close to the volcano where the reduction errors β_{po} are less than 0.10.

avoids magnetic coastal effects) and because transient magnetic variations are approximately homogeneous in this area (Pozzi et al., 1979).

Beginning in 1979, the network of markers was measured regularly. Two series of measurements of about three weeks each were performed (February—March, March—April, and May). In 1980, four markers were added (NCJ, COM, GRI, CRT) (Fig. 1 and Table 2). Since that date the network was subdivided into two parts. The first, made up of markers close to La Soufrière (MTB, SAV, BJH, PAR, PAL, FSC) was read every month or so. The second, made up of the more remote markers (NCJ, COM, HBT, DOL, GDM, BAI, GRI, VHT, CRT) was measured every two months. At each marker, two or three groups of 10 consecutive measurements 5 or 10 minutes apart were reduced in relation to the Matouba telemetering station and the value selected for $\Delta B(P, \mathrm{O}, t)$ is the average of the two or three series.

3. TRANSIENT VARIATIONS AND REDUCTION ERRORS

The reduction errors E_{r} (P, O, t) [the second term at the right-hand side of eqn. (3)] can be evaluated in the absence of volcanic activity on the basis of a study of the transient field.

The transient field $\vec{B}_{\mathrm{t}}$ is the sum of an exciting part $\vec{B}_{\mathrm{te}}$ of external origin and an induced part $\vec{B}_{ti}$ of internal origin. The field $\vec{B}_{te}$ can be considered uniform on the network area because the altitude of the sources is greater than 100 km. The field $\vec{B}_{ti}$, generated by the induced currents circulating in the conducting structures, decomposes into two terms, one of which is regular and the other, abnormal. The former represents the response of a stratified structure to the external field $\vec{B}_{te}$. The latter is the result of the lateral heterogeneities of conductivity of the superficial geological structures and creates the anomalies of the variations observed on the surface.

a. Continuous recordings

The study was carried out in two phases:

(a) The first took place at the end of the seismovolcanic crisis of 1976—1977. At the Observatory (FSC), we have continuous recordings of the variations of the components of the magnetic field (intensity B, horizontal and vertical components H and Z, declination D) and, at some stations near the volcano (VAL, PAR, SAV, PAL) differential continuous recordings of the intensity of the field with regard to the Matouba station (Fig. 1).

(b) The second took place during 1981. A portable station was operated for several days at successive points on the volcano (NCJ, CAR, DOL, FSC, Tobaco, Hermitage, Champfleury, Delgrès, Beausoleil) (Fig. 1). The measurements were reduced point by point with regard to the MTB telemetering station.

It was established that at any two points, the variations of the magnetic field are proportional to each other within the accuracy of the measurements (Fig. 2). As a result, the reduction errors $E_r\,(P, \mathrm{O}, t)$ are expressed simply in the form of a product of a spatial function a_{po} depending only on points P and O and a temporal function $R(t)$ (Pozzi et al., 1979; Zlotnicki, 1979):

$$E_r(P, \mathrm{O}, t) = \vec{B}_t(P, t) \cdot \vec{u}(P) - \vec{B}_t(\mathrm{O}, t) \cdot \vec{u}(\mathrm{O}) = a_{po} R(t) \qquad (4)$$

The characteristic function $R(t)$ of the excitation is expressed in the form of a convolution product (Le Mouël and Menvielle, 1982; Menvielle, 1984):

$$R(t) = K \times [\vec{q} \cdot \vec{B}_t(\mathrm{O}, t)] \qquad (5)$$

K is a linear operator; $\vec{q}$ is a unit vector of the horizontal plane defining the efficient direction of the transient normal field to station O. In an initial approximation, K can be considered as a scalar, and at Guadeloupe the horizontal N—S component and the intensity of the magnetic field are proportional (Fig. 2).

The temporal variations of the function $R(t)$ are therefore proportional to those of the intensity of the field, and the reduction errors can be written in the following simple form:

$$E_r(P, \mathrm{O}, t) = \beta_{po} \vec{B}_t(\mathrm{O}, t) \cdot \vec{u}(\mathrm{O}) \qquad (6)$$

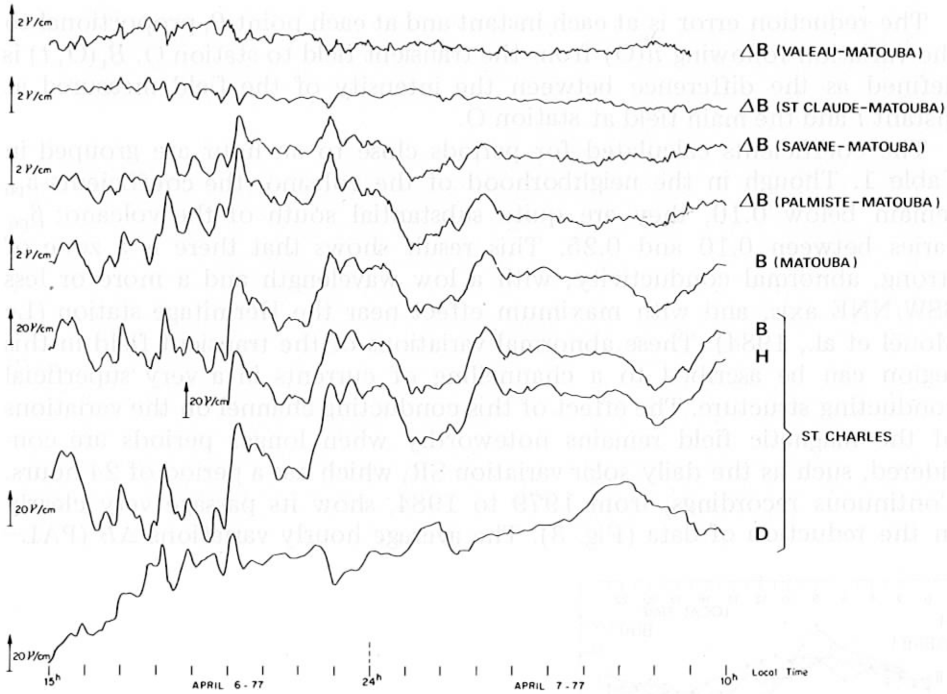


Fig. 2. Example of continuous recording of the earth's magnetic field (B: intensity; H: horizontal component; D: declination; ΔB: difference) at the end of the seismovolcanic crisis.

TABLE 1

Coefficients β_{po} calculated on the basis of continuous recordings of B for periods close to an hour. The reduction station is MTB.

Station	Year	Coefficients β_{po} (Reference: Matouba)
Savane	1977	0.05
Parnasse	1977	−0.03
Valeau	1977	−0.03
Palmiste	1977	0.08
Neuf-Chateau	1981	−0.07
Tobaco	1981	0.18
Carbet	1981	0.02
Champfleury	1981	0.10
Dolé	1981	0.15
Delgrès	1981	0.19
L'Hermitage	1981	0.25
Fort St. Charles	1981	0.12
Beausoleil	1981	0.01

The reduction error is, at each instant and at each point P, proportional to the variation following $\vec{u}(\mathrm{O})$ from the transient field to station O. $\vec{B}_t(\mathrm{O}, t)$ is defined as the difference between the intensity of the field measured at instant t and the main field at station O.

The coefficients calculated for periods close to an hour are grouped in Table 1. Though in the neighborhood of the volcano, the coefficients β_{po} remain below 0.10, they are quite substantial south of the volcano; β_{po} varies between 0.10 and 0.25. This result shows that there is a zone of strong, abnormal conductivity, with a low wavelength and a more or less SSW/NNE axis, and with maximum effect near the Hermitage station (Le Mouël et al., 1984). These abnormal variations of the transient field in this region can be ascribed to a channelling of currents in a very superficial conducting structure. The effect of this conducting channel on the variations of the magnetic field remains noteworthy when longer periods are considered, such as the daily solar variation SR, which has a period of 24 hours. Continuous recordings, from 1979 to 1984, show its passage very clearly in the reduction of data (Fig. 3). The average hourly variations ΔB (PAL—

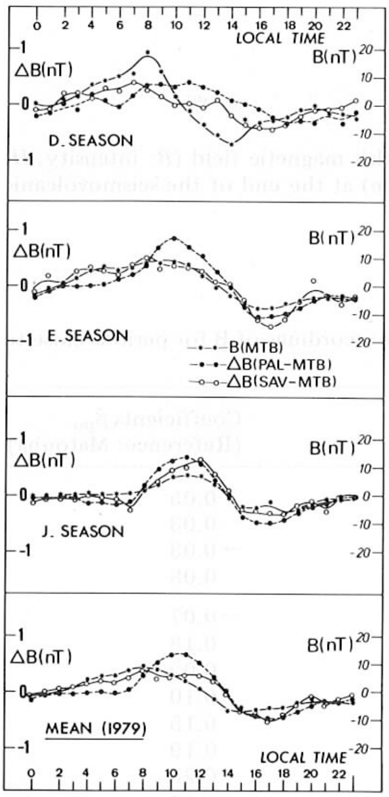


Fig. 3. Telemetering stations. Hourly mean values of the differences ΔB (PAL—MTB), ΔB (SAV—MTB) and of the intensity B (MTB) during the D, E, and J seasons and mean annual values (1979).

MTB) and ΔB (SAV—MTB) calculated for the D seasons (November, December, January, February), the E seasons (September, October, March, April), and the J seasons (May, June, July, August) do not exceed 1.5 nT in amplitude and remain more or less proportional with the average hourly variations at the Matouba station (amplitudes below 35 nT). Between 1979 and 1982, the coefficient β_{po} pertaining to Palmiste and Matouba, estimated over a year, varies between 0.03 and 0.06.

3b. Network of markers — 1979

The network of markers was measured from February to May. Seismic energy release during these three months was less than 2×10^{13} ergs and magnetic agitation, characterized by the indices aa (morning and afternoon) (Mayaud, 1973) was sometimes considerable (Fig. 4).

The markers close to the summit of the volcano showed only low variations, generally below 5 nT (MTB, SAV, BJH, PAR, CAR, PAL). At some more remote markers (DOL, GDM), the variations were as high as 15 nanoteslas.

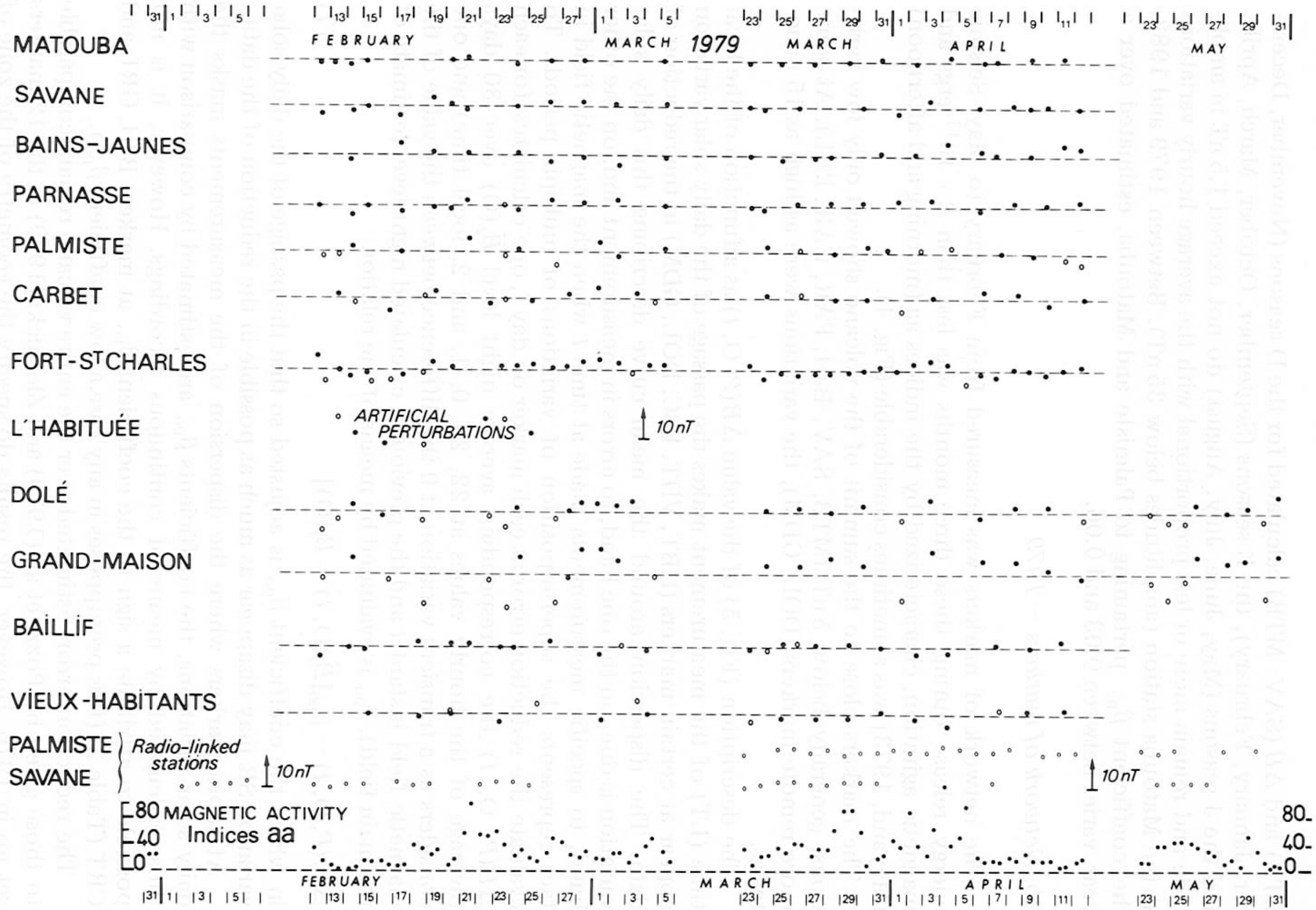
The description (Fig. 5) of the value $\Delta B(P, \mathrm{O}, t)$ as a function of the local time (LT) of the measurement makes the passage of the daily solar variation appear at certain markers (HBT, VHT, FSC, DOL, GDM) in the reduction of data. The dispersion around the mean curve describing this daily solar variation is due, on the one hand, to errors in measurement and, on the other hand, to specific measurements made at time t when the magnetic field in fact represents the superimposition of variations of multiple periods. To evaluate the reduction error at each marker on day j, one calculates, for each $\Delta B(P, \mathrm{O}, t)$ the corresponding average night level $\bar{B}_j(\mathrm{O})$ over 30 days (average of the hourly values at 22, 23, 0, 1, and 2, local time) and one considers as a transient variation at 0 the difference between the value of the magnetic field instant t and the previously calculated night level (estimate of the main field). β_{po} is evaluated by means of the relation:

$$\Delta B_j(P, \mathrm{O}, t) - \beta_{po}[\vec{B}_j(\mathrm{O}, t) - \vec{B}_j(\mathrm{O})] \tag{7}$$

in which the coefficient β_{po} is adjusted so that the passage of the daily solar variation SR may disappear as much as possible in the reduction of the data.

At those markers where the dispersion of the measurements masks the daily solar variations, the coefficients β_{po} are estimated by comparison with those calculated by means of continuous recordings. However, it is not possible to attribute a sign to the coefficient β_{po} at markers BJH, GRI, and CRT (Table 2) (this presupposes in any case a low coefficient β_{po}).

The reduction errors estimated over the entire volcano remain comparable to those given by Pozzi et al. (1979) and Zlotnicki (1979) at the 12 markers set up in 1976. However, the results do specify the geometry of the zone of abnormal conductivity and the intensity of its effects. It is established that the markers close to La Soufrière are only slightly disturbed by these


1979
FEBRUARY
MARCH
APRIL
MAY
MATOUBA
SAVANE
BAINS-JAUNES
PARNASSE
PALMISTE
CARBET
FORT-S^T CHARLES
L'HABITUÉE
ARTIFICIAL PERTURBATIONS
10 nT
DOLÉ
GRAND-MAISON
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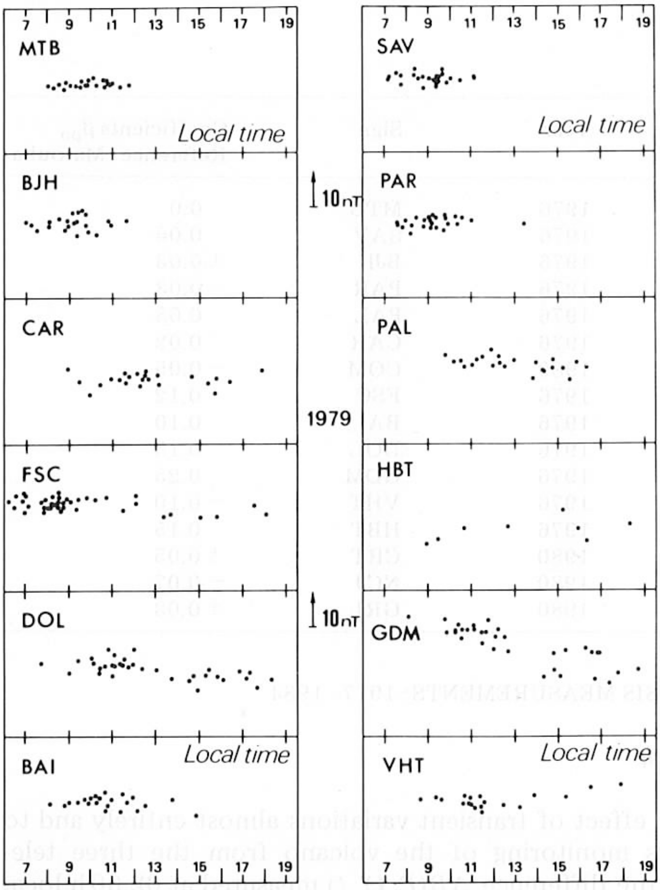


Fig. 5. Differences ΔB measured between February and May 1979 on the network of markers as a function of local time.

anomalies (the hatched zone in Fig. 1), just as the more remote markers located to the west (CRT, VHT, BAI, GRI) and east (NCJ, COM). The SAV and PAL telemetering stations set up in 1978 belong to this zone of low abnormal conductivity.

On the other hand the variations $\Delta B(P, \mathrm{O}, t)$ measured at the markers to the south of the volcano (DOL, GDM, HBT, FSC) remain highly dependent on magnetic agitation. At these markers, daily measurements performed once in the morning and once in the afternoon, can show variations in excess of roughly 10 nT (Figs. 4 and 5).

Fig. 4. Differences ΔB over the network of markers and telemetering network between February and May 1979. The values of the telemetering stations are those measured at 02.00 h local time. The dotted lines represent the mean values: for each day the average indices of magnetic agitation (am and pm) are shown.

TABLE 2

Coefficients β_{po} estimated

Marker	Year	Sign	Coefficients β_{po} Reference: Matouba
Matouba	1976	MTB	0.0
Savane	1976	SAV	0.05
Bains-Jaunes	1976	BJH	± 0.03
Parnasse	1976	PAR	− 0.03
Palmiste	1976	PAL	0.08
Carbet	1976	CAR	0.02
Concession-Manceau	1980	COM	− 0.05
Fort St. Charles	1976	FSC	0.12
Baillif	1976	BAI	0.10
Dolé	1976	DOL	0.15
Grand-Maison	1976	GDM	0.25
Vieux-Habitants	1976	VHT	− 0.10
L'Habituée	1976	HBT	0.15
Les Cretes	1980	CRT	± 0.05
Neuf-Chateau	1980	NCJ	− 0.07
Grand-Rivière	1980	GRI	± 0.03

4. RESULTS OF POST-CRISIS MEASUREMENTS: 1977—1984

4a. Telemetering network

In order to avoid the effect of transient variations almost entirely and to simplify the continuous monitoring of the volcano from the three telemetering stations, only the difference $\Delta B(P, \mathrm{O}, t)$ measured at 02.00 h local time (which roughly corresponds to the night level, see Fig. 3) were taken into account. As the coefficient β_{po} at these stations is sufficiently low (Table 1), the temporal search for a possible volcanomagnetic effect can be conducted easily (Fig. 4).

Figure 6 shows the difference ΔB (PAL—MTB) measured at 02.00 h local time and averaged over 5-day sequences, between 1979 and 1984. From 1979 to June 1982, all the values lie within ± 1 nT of the average annual values represented by dashes. The average annual values seem to show a very slight decrease in the difference ΔB (PAL—MTB) in four years, i.e. a decrease of the order of 0.5 nT (Table 3); but, because of the irregular operation of station SAV and the more substantial dispersion observed on the network of markers (see next paragraph), this result cannot be confirmed. From July, 1982 to May, 1983 many artificial magnetic inputs disturb the Palmiste telemetering station and the measurements could not be resumed until May, 1983 despite a considerable dispersion in the six succeeding months.

Between 1979 and 1984, the only significant event was the seismovolcanic

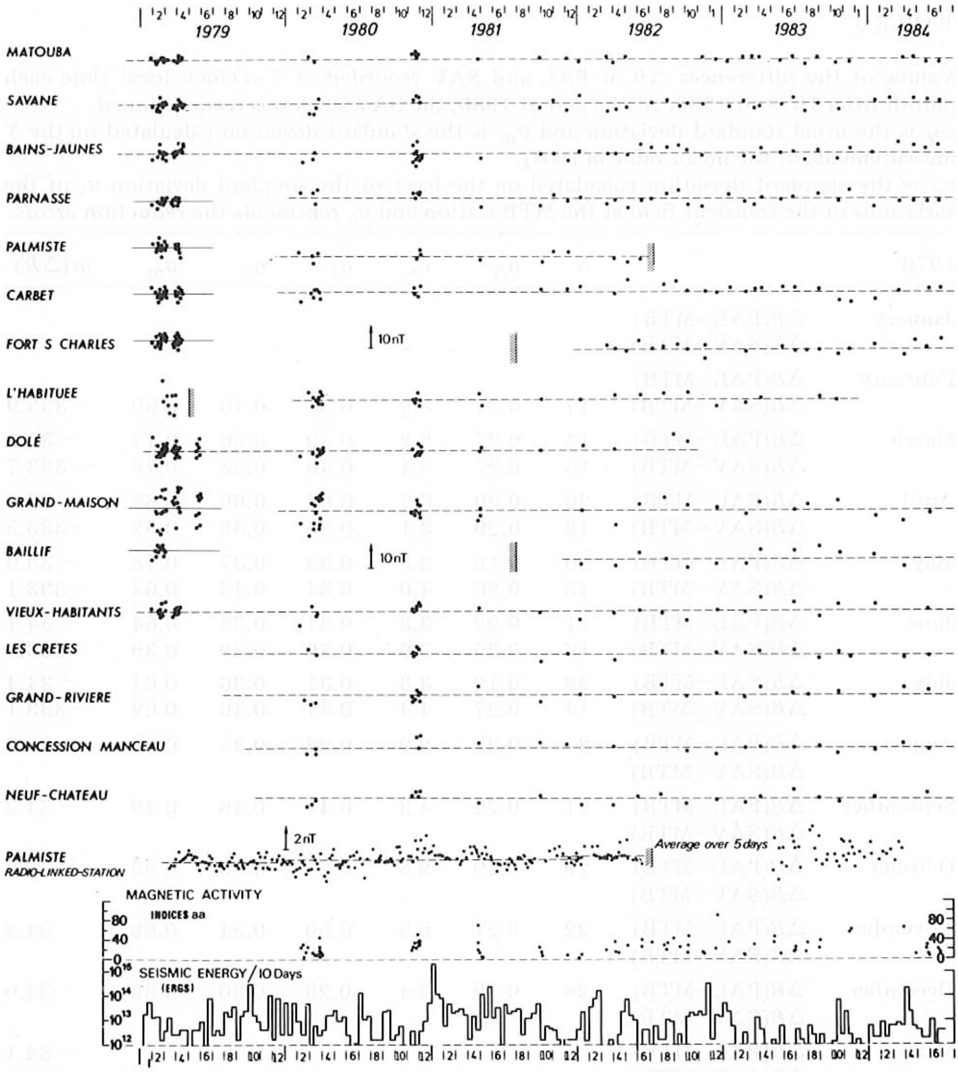


Fig. 6. Differences measured on the network of markers between 1979 and 1984. Averages in 1979 and between 1980 and 1984 (dotted lines). The points representing telemetering stations show mean values recorded over 5 days at 02.00 h local time. The indices of magnetic agitation are indicated at each reading of the network of markers and the seismic energy released per 10-day period is shown.

crisis at the beginning of January, 1981. The seismic energy release reached 3×10^{15} ergs in 10 days (Fig. 6). Though it is difficult to establish a definite limit between the variations ΔB (PAL$-$MTB) observed in December, 1980 and this volcanic crisis, it should be noted however that there is a greater dispersion of measurements (about 2 nT) and that the average level for December is lower by 1 to 2 nT than the average value for 1980.

TABLE 3

Values of the differences ΔB at PAL and SAV recorded at 2 o'clock local time each month from 1979 to 1983. At the end of 1980, the SAV station is reconditioned.
σ_N is the usual standard deviation and σ_m is the standard deviation calculated on the N measurements of the mean value m (ΔB).
σ_c is the standard deviation calculated on the basis of the standard deviation σ_t of the variations of the transient field at the MTB station and σ_ϵ represents the reduction errors.

1979		N	σ_N	σ_t	σ_ϵ	σ_c	σ_m	$m(\Delta B)$
January	ΔB(PAL–MTB)							
	ΔB(SAV–MTB)							
February	ΔB(PAL–MTB)							
	ΔB(SAV–MTB)	17	0.24	3.9	0.32	0.40	0.60	−333.9
March	ΔB(PAL–MTB)	14	0.27	5.2	0.49	0.56	0.77	−34.1
	ΔB(SAV–MTB)	13	0.27	5.4	0.46	0.53	0.75	−333.7
April	ΔB(PAL–MTB)	26	0.20	6.7	0.63	0.66	0.65	−33.5
	ΔB(SAV–MTB)	12	0.29	3.1	0.26	0.39	0.52	−333.5
May	ΔB(PAL–MTB)	30	0.18	3.5	0.33	0.37	0.76	−33.9
	ΔB(SAV–MTB)	15	0.26	4.0	0.34	0.43	0.64	−333.1
June	ΔB(PAL–MTB)	21	0.22	3.3	0.31	0.38	0.64	−34.1
	ΔB(SAV–MTB)	16	0.25	3.6	0.30	0.39	0.39	−333.1
July	ΔB(PAL–MTB)	29	0.19	3.3	0.31	0.36	0.61	−34.4
	ΔB(SAV–MTB)	14	0.27	4.4	0.37	0.46	0.62	−333.1
August	ΔB(PAL–MTB)	21	0.22	2.9	0.28	0.35	0.68	−34.9
	ΔB(SAV–MTB)							
September	ΔB(PAL–MTB)	21	0.22	4.3	0.41	0.46	0.49	−34.3
	ΔB(SAV–MTB)							
October	ΔB(PAL–MTB)	29	0.19	2.8	0.27	0.33	0.93	−33.8
	ΔB(SAV–MTB)							
November	ΔB(PAL–MTB)	22	0.21	5.3	0.50	0.54	0.69	−34.2
	ΔB(SAV–MTB)							
December	ΔB(PAL–MTB)	28	0.19	2.4	0.23	0.30	0.89	−34.0
	ΔB(SAV–MTB)							
Mean	ΔB(PAL–MTB)							−34.1
	ΔB(SAV–MTB)							−333.4

1980		N	σ_N	σ_t	σ_ϵ	σ_c	σ_m	$m(\Delta B)$
January	ΔB(PAL–MTB)	24	0.20	2.61	0.25	0.32	0.82	−34.5
	ΔB(SAV–MTB)							
February	ΔB(PAL–MTB)	25	0.20	3.40	0.32	0.38	0.84	−34.2
	ΔB(SAV–MTB)							
March	ΔB(PAL–MTB)	27	0.19	3.7	0.35	0.39	0.97	−34.0
	ΔB(SAV–MTB)							
April	ΔB(PAL–MTB)	25	0.20	2.7	0.25	0.32	0.95	−34.4
	ΔB(SAV–MTB)							
May	ΔB(PAL–MTB)	26	0.20	2.3	0.21	0.29	0.42	−34.0
	ΔB(SAV–MTB)							

TABLE 3 (Cont.)

1980		N	σ_N	σ_t	σ_ϵ	σ_c	σ_m	$m(\Delta B)$
June	ΔB(PAL–MTB)	21	0.22	5.1	0.48	0.53	0.64	−33.6
	ΔB(SAV–MTB)							
July	ΔB(PAL–MTB)	12	0.29	1.5	0.14	0.32	1.03	−34.0
	ΔB(SAV–MTB)							
August	ΔB(PAL–MTB)	16	0.25	2.8	0.26	0.36	0.20	−33.9
	ΔB(SAV–MTB)							
September	ΔB(PAL–MTB)	22	0.21	2.3	0.22	0.30	0.48	−34.0
	ΔB(SAV–MTB)							
October	ΔB(PAL–MTB)	22	0.21	4.1	0.38	0.44	0.61	−33.4
	ΔB(SAV–MTB)							
November	ΔB(PAL–MTB)	23	0.21	2.8	0.26	0.33	0.55	−33.5
	ΔB(SAV–MTB)							
December	ΔB(PAL–MTB)	25	0.20	4.3	0.41	0.46	0.62	−33.0
	ΔB(SAV–MTB)							
Mean	ΔB(PAL–MTB)							−33.9
	ΔB(SAV–MTB)							

1981		N	σ_N	σ_t	σ_ϵ	σ_c	σ_m	$m(\Delta B)$
January	ΔB(PAL–MTB)	21	0.22	6.9	0.65	0.68	0.34	−33.6
	ΔB(SAV–MTB)							
February	ΔB(PAL–MTB)	12	0.29	5.8	0.54	0.62	0.56	−33.9
	ΔB(SAV–MTB)	7	0.38	2.6	0.22	0.44	0.44	−330.4
March	ΔB(PAL–MTB)	18	0.24	6.8	0.64	0.68	0.51	−33.9
	ΔB(SAV–MTB)	22	0.21	7.1	0.59	0.63	0.41	−330.2
April	ΔB(PAL–MTB)	15	0.26	3.9	0.37	0.45	0.50	−33.3
	ΔB(SAV–MTB)	25	0.20	6.35	0.53	0.57	0.48	−330.4
May	ΔB(PAL–MTB)	25	0.20	8.1	0.76	0.79	0.31	−33.6
	ΔB(SAV–MTB)	16	0.32	7.5	0.63	0.70	0.66	−330.1
June	ΔB(PAL–MTB)	27	0.19	7.1	0.66	0.69	0.44	−34.1
	ΔB(SAV–MTB)	10	0.32	8.5	0.71	0.78	0.47	−330.5
July	ΔB(PAL–MTB)	23	0.21	6.8	0.64	0.67	0.61	−34.1
	ΔB(SAV–MTB)	14	0.27	3.6	0.30	0.40	0.24	−330.8
August	ΔB(PAL–MTB)	29	0.19	5.6	0.53	0.56	0.52	−33.9
	ΔB(SAV–MTB)	29	0.19	3.8	0.31	0.59	0.51	−330.7
September	ΔB(PAL–MTB)	9	0.33	1.3	0.12	0.35	0.41	−34.0
	ΔB(SAV–MTB)	17	0.24	5.3	0.44	0.50	0.35	−330.8
October	ΔB(PAL–MTB)	11	0.30	8.0	0.76	0.81	0.57	−33.6
	ΔB(SAV–MTB)							
November	ΔB(PAL–MTB)	10	0.32	7.0	0.66	0.74	0.42	−33.3
	ΔB(SAV–MTB)	25	0.20	6.0	0.50	0.54	0.50	−330.7
December	ΔB(PAL–MTB)	20	0.22	7.9	0.75	0.77	0.62	−34.0
	ΔB(SAV–MTB)	22	0.21	5.1	0.42	0.47	0.48	−330.9
Mean	ΔB(PAL–MTB)							−33.8
	ΔB(SAV–MTB)							−330.6

TABLE 3 (Cont.)

1982		N	σ_N	σ_t	σ_ϵ	σ_c	σ_m	$m(\Delta B)$
January	ΔB(PAL–MTB)	23	0.21	4.7	0.44	0.49	0.65	−33.6
	ΔB(SAV–MTB)	30	0.18	5.2	0.43	0.47	0.55	−330.9
February	ΔB(PAL–MTB)	19	0.23	4.7	0.44	0.50	1.11	−32.9
	ΔB(SAV–MTB)	23	0.21	5.2	0.43	0.48	0.87	−331.0
March	ΔB(PAL–MTB)	22	0.21	4.2	0.40	0.45	0.24	−33.7
	ΔB(SAV–MTB)	28	0.19	4.4	0.36	0.41	0.64	−330.8
April	ΔB(PAL–MTB)	24	0.20	7.7	0.72	0.75	0.55	−33.5
	ΔB(SAV–MTB)	28	0.19	7.4	0.62	0.65	0.45	−330.8
May	ΔB(PAL–MTB)	26	0.20	6.3	0.59	0.63	0.80	−33.6
	ΔB(SAV–MTB)	30	0.18	6.3	0.52	0.55	0.41	−330.7
June	ΔB(PAL–MTB)	11	0.30	4.4	0.42	0.51	0.41	−34.0
	ΔB(SAV–MTB)	11	0.30	4.6	0.38	0.49	0.43	−330.5
July	ΔB(PAL–MTB)							
	ΔB(SAV–MTB)							
August	ΔB(PAL–MTB							
	ΔB(SAV–MTB)							
September	ΔB(PAL–MTB)							
	ΔB(SAV–MTB)							
October	ΔB(PAL–MTB)							
	ΔB(SAV–MTB)							
November	ΔB(PAL–MTB)							
	ΔB(SAV–MTB)							
December	ΔB(PAL–MTB)							
	ΔB(SAV–MTB)							
Mean	ΔB(PAL–MTB)							−33.6
	ΔB(SAV–MTB)							−330.8

1983		N	σ_N	σ_t	σ_ϵ	σ_c	σ_m	$m(\Delta B)$
January	ΔB(PAL–MTB)							
	ΔB(SAV–MTB)							
February	ΔB(PAL–MTB)							
	ΔB(SAV–MTB)							
March	ΔB(PAL–MTB)							
	ΔB(SAV–MTB)							
April	ΔB(PAL–MTB							
	ΔB(SAV–MTB)							
May	ΔB(PAL–MTB)	17	0.24	4.4	0.41	0.48	1.55	−39.7
	ΔB(SAV–MTB)							
June	ΔB(PAL–MTB	17	0.24	5.2	0.49	0.55	1.75	−38.3
	ΔB(SAV–MTB)							
July	ΔB(PAL–MTB)	6	0.41	3.8	0.36	0.54	2.04	−39.2
	ΔB(SAV–MTB)							

TABLE 3 (Cont.)

1983		N	σ_N	σ_t	σ_ϵ	σ_c	σ_m	$m(\Delta B)$
August	ΔB(PAL–MTB)	21	0.22	5.0	0.47	0.52	1.43	−38.0
	ΔB(SAV–MTB)							
September	ΔB(PAL–MTB)	11	0.30	5.0	0.47	0.55	1.67	−39.5
	ΔB(SAV–MTB)							
October	ΔB(PAL–MTB)	18	0.24	5.5	0.52	0.57	1.09	−39.0
	ΔB(SAV–MTB)							
November	ΔB(PAL–MTB)	18	0.24	4.7	0.44	0.71	0.89	−39.4
	ΔB(SAV–MTB)							
December	ΔB(PAL–MTB)	5	0.45	4.7	0.44	0.63	0.63	−38.8
	ΔB(SAV–MTB)							
Mean	ΔB(PAL–MTB)							−38.9
	ΔB(SAV–MTB)							

In order to refine these results, it is possible to study the slow magnetic variations by calculating the average monthly values of the differences ΔB (P, O, t) measured 02.00 h local time (Fig. 7). For each of the average values $\Delta B(P, \mathrm{O}, t)$, a comparison is made between the standard deviation σ_m of the measurements and the standard deviation σ_c calculated on the basis of the reduction errors (Le Mouël et al., 1984). This last value is computed by means of the N values of the month:

$$\sigma_c = (\sigma_N^2 + \sigma_\epsilon^2)^{1/2} \tag{8}$$

$$\sigma_N = N/\sqrt{N}$$

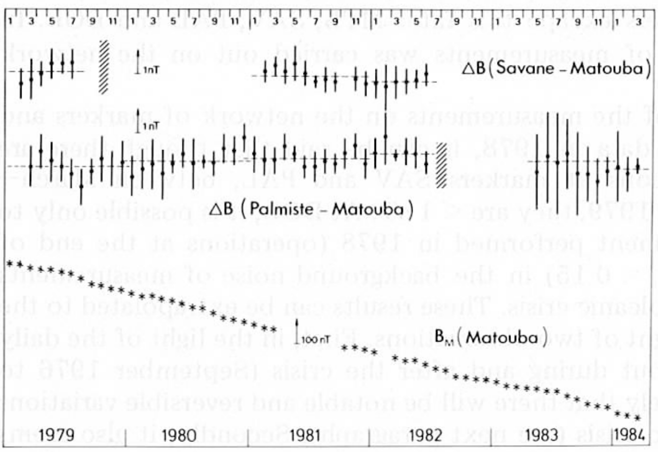


Fig. 7. Mean monthly differences and standard deviations associated with the PAL and SAV telemetering stations between 1979 and 1984. Changes in the mean monthly intensity of the field at MTB.

and σ_ϵ is the standard deviation of the series:

$$\beta_{po}\,[B_j(\mathrm{O},\,2\mathrm{LT}) - \bar{B}_j(\mathrm{O})] \qquad j = 1 \quad \text{to} \quad N \tag{9}$$

β_{po} is the coefficient that characterizes the reduction errors at station P; B_j(O), 2LT) is the value of the magnetic field at 02.00 h local time at reference station O and $\bar{B}_j$(O) is the corresponding average night level on day j (see Table 3).

Until June, 1982, the calculated dispersion σ_c arising from the reduction errors was of the same order of magnitude as the standard dispersion σ_m of the measurements. These dispersions remained close to 0.8 nT and only very rarely exceeded 1 nT. The average monthly values varied only by less than 1 nT around the average annual values, except for the difference ΔB (PAL—MTB) of December 1980, which was linked with the volcanic crisis of early January, 1981.

It is apparent than, that the differential magnetic network which is not significantly disturbed by reduction errors, can bring to light magnetic variations with an amplitude greater than 1 nT over periods longer than one month, and with an amplitude of about 2 nT over periods of several days.

4b. Network of markers

As a result of a new reduction station (MTB) set up in 1978 and located about 200 m from marker MTB, and to the telemetering station used from 1976 to 1977 (Robach et al., 1976), it was possible to connect all the measurements performed between 1976 and 1984. However, many construction projects (roads, buildings, etc.) begun after the volcanic crisis near several markers made it impossible, at these locations, to connect 1976—1977 data with post-1978 data (BJH, PAR, CAR, FSC, BAI, GDM, VHT, HBT). Only four markers excape this fate: MTB, SAV, PAL and DOL. In 1978, only one series of measurements was carried out on the network (Table 4).

Given the accuracy of the measurements on the network of markers and the limited number of data in 1978, it can be said that though there are volcanomagnetic variations at markers SAV and PAL, between March—April 1977 and 1978 or 1979, they are < 1 nT. At DOL, it is possible only to note that the measurement performed in 1978 (operations at the end of 1978) also persists ($\beta_{po} = 0.15$) in the background noise of measurements performed during the volcanic crisis. These results can be extrapolated to the entire network in the light of two observations. First, in the light of the daily measurements carried out during and after the crisis (September 1976 to April, 1977), it is unlikely that there will be notable and reversible variations after the seismovolcanic crisis (see next paragraph). Secondly, it also seems unlikely that there should be magnetic variations at markers as such as GDM, CAR, or BJH without there being detectable variations at respective markers close by DOL, PAL and SAV; it is difficult to imagine very localized

TABLE 4

Mean values $m(\Delta B)$, standard deviations σ_m, and number of values N at each marker between the end of the seismovolcanic crisis (3-4/1977) and 1984 . The series in bold-faced and thin-faced characters bring together homogeneous series (see text). The series in 1977 at MTB and DOL bring together all the values obtained during the crises.

Station	3-4/1977			1978			1979		
	m	σ_m	N	m	σ_m	N	m	σ_m	N
MTB	33.7	1.1	131	31.9		1	33.7	0.8	22
SAV	−205.1	0.9	29				−205.0	1.3	25
BJH				−1119.4		1	−1119.7	2.0	22
PAR							−659.6	1.3	25
PAL	170.6	2.2	17	171.7		1	171.6	1.9	23
CAR				798.2		1	797.6	1.7	23
FSC				**81.7**		**1**	**79.1**	**1.8**	**43**
BAI							−362.8	1.5	23
DOL	**−823.5**	**3.3**	**53**	**820.5**		1	−801.8	2.6	34
GDM				385.8		1	384.6	5.3	32
VHT							1057.2	1.8	19
HBT				**−77.2**	**5.4**	**2**	**−77.1**	**3.5**	**8**
CRT			—			—			—
COM			—			—			—
NCJ			—			—			—
GRI			—			—			—

Station	1980			1981			1982		
	m	σ_m	N	m	σ_m	N	m	σ_m	N
MTB	34.7	1.2	7	33.5	0.2	3	33.1	0.7	6
SAV	−204.7	2.8	8	−205.7	1.6	3	−205.2	1.8	9
BJH	−1121.3	2.5	15	−1122.0	1.2	3	−1119.2	1.5	10
PAR	−657.1	1.6	9	−659.9	1.9	5	−659.6	2.1	10
PAL	168.0	1.6	7	167.4	1.7	4	166.4	0.6	3
CAR	797.5	1.7	13	798.9	0.5	3	799.3	2.0	8
FSC							−60.2	1.9	10
BAI							−332.8	2.1	6
DOL	−802.6	1.7	17	−802.1	0.4	3	−799.1	2.1	4
GDM	383.5	4.4	22	381.7	2.0	3	384.3	3.5	6
VHT	1057.3	2.6	10	1057.2	1.0	2	1055.4	1.9	6
HBT	−91.2	1.9	16	−91.6	2.0	3	−90.5	2.1	6
CRT	750.6	1.3	7	746.7	1.4	2	749.0	1.6	3
COM	−96.7	2.8	6	97.7	0.2	3	94.8	1.3	3
NCJ	−99.4	2.2	7	−101.0	1.6	3	−100.9	1.6	6
GRI	656.2	1.6	10	657.6	3.1	2	658.3	1.8	5

TABLE 4 (continued)

Station	1983			1984			1980–1984		
	m	σ_m	N	m	σ_m	N	m	σ_m	N
MTB	33.2	1.0	10	33.0	0.9	4	33.6	1.1	30
SAV	−205.5	1.6	9	−203.0	0.6	5	−204.9	2.1	34
BJH	−1120.4	1.7	9	−1117.7	1.4	5	−1120.2	2.3	43
PAR	−658.3	1.6	9	−660.2	1.0	5	−658.8	2.1	38
PAL			—			—			—
CAR	797.9	2.2	9	798.0	1.8	4	798.0	2.0	37
FSC	−61.3	0.9	7	−58.9	2.4	5	−60.4	1.8	21
BAI	−328.9	2.5	5	−332.1	2.1	2	−331.2	2.9	13
DOL							−802.0	2.1	24
GDM	385.3	2.0	5	381.1	1.5	3	383.5	3.9	39
VHT	1059.0	2.9	5	1059.1	0.8	2	1057.3	2.7	25
HBT	−91.2	1.0	4				−91.1	1.9	29
CRT	749.6	1.2	6	748.3	0.5	2	749.5	1.6	20
COM	95.9	1.1	5	96.3	1.4	3	96.3	2.0	20
NCJ	−99.7	1.9	5	−98.7	1.0	2	−99.6	1.9	19
GRI	658.7	1.4	5	659.4	1.7	3	657.6	2.2	25

magnetic variations that occur far from the volcano and yet are linked to its activity.

In 1978, at the 8 markers where a measurement of the difference ΔB (P, O, t) is available, all the values lie within the dispersion of the 1979 measurements except for the value of DOL, where construction is in progress. For the period 1979–1984, many measurements are available at each marker and it is possible to make a more detailed study of the variations ΔB(P, O, t) (Fig. 6). This graph indicates the seismic energy released in 10-day periods, the agitation indices aa (Mayaud, 1973), the days when the network is measured, and the values ΔB(P, O, t) at the PAL telemetering station recorded at 02.00 h local time and averaged over 5-day periods. During these 6 years, four markers (HBT, DOL, FSC, and BAI) have been disturbed and the series of measurements must be dissociated temporally. In a period without volcanic activity, the purpose of a network of markers associated with a telemetering network is to search for long-term volcano-magnetic variations over the entire volcanic area. Thus, the 16 markers can be grouped into two subsets. The first (MTB, SAV, BJH, PAR, PAL, CAR, FSC, CRT, GRI, COM, NCJ) show only variations less than 5 nT. The second (HBT, DOL, GDM, BAI, VHT) show variations much greater than 5 nT. The dispersion of the measurements ΔB(P, O, t) made over 5 years, (1980–1984) lies between 1.6 and 2.3 nT at those markers where β_{po} is less than 0.10 and reaches 3.9 nT at marker GDM where β_{po} is equal to 0.25. In spite of these reduction errors, the magnetic variations $\Delta B(P, \mathrm{O}, t)$ fluctuate around the average value over five years (1980–1984) (represented in dashes in Fig. 6) and no long-term variation is detectable (Table 4), (the drift observed at PAL between 1979 and 1980 can be ascribed to crop-

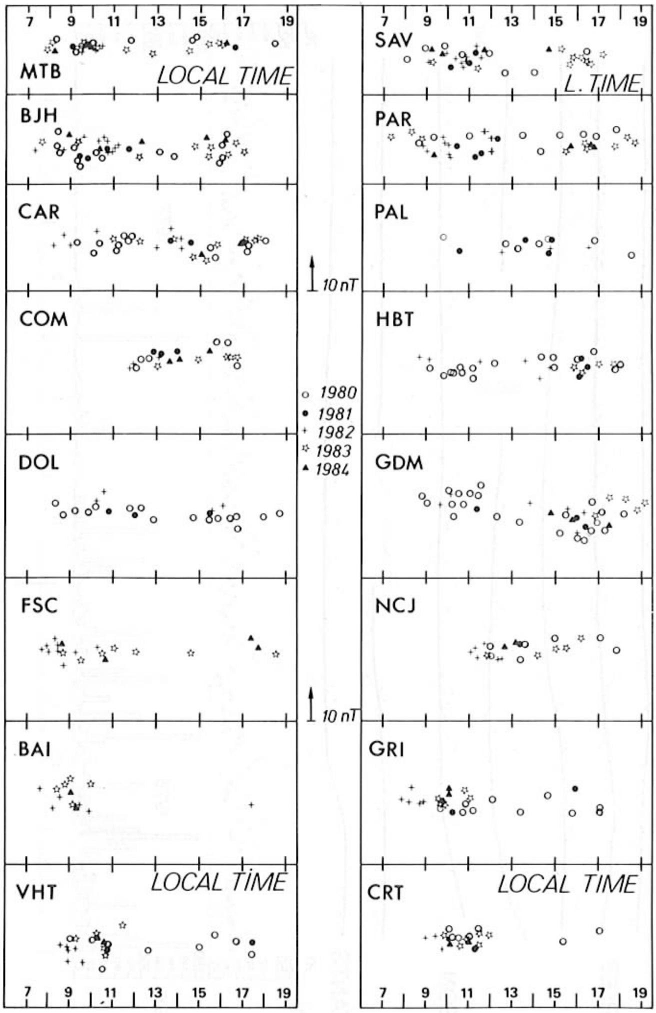


Fig. 8. Differences measured between 1980 and 1984 on the network of markers as a function of local time.

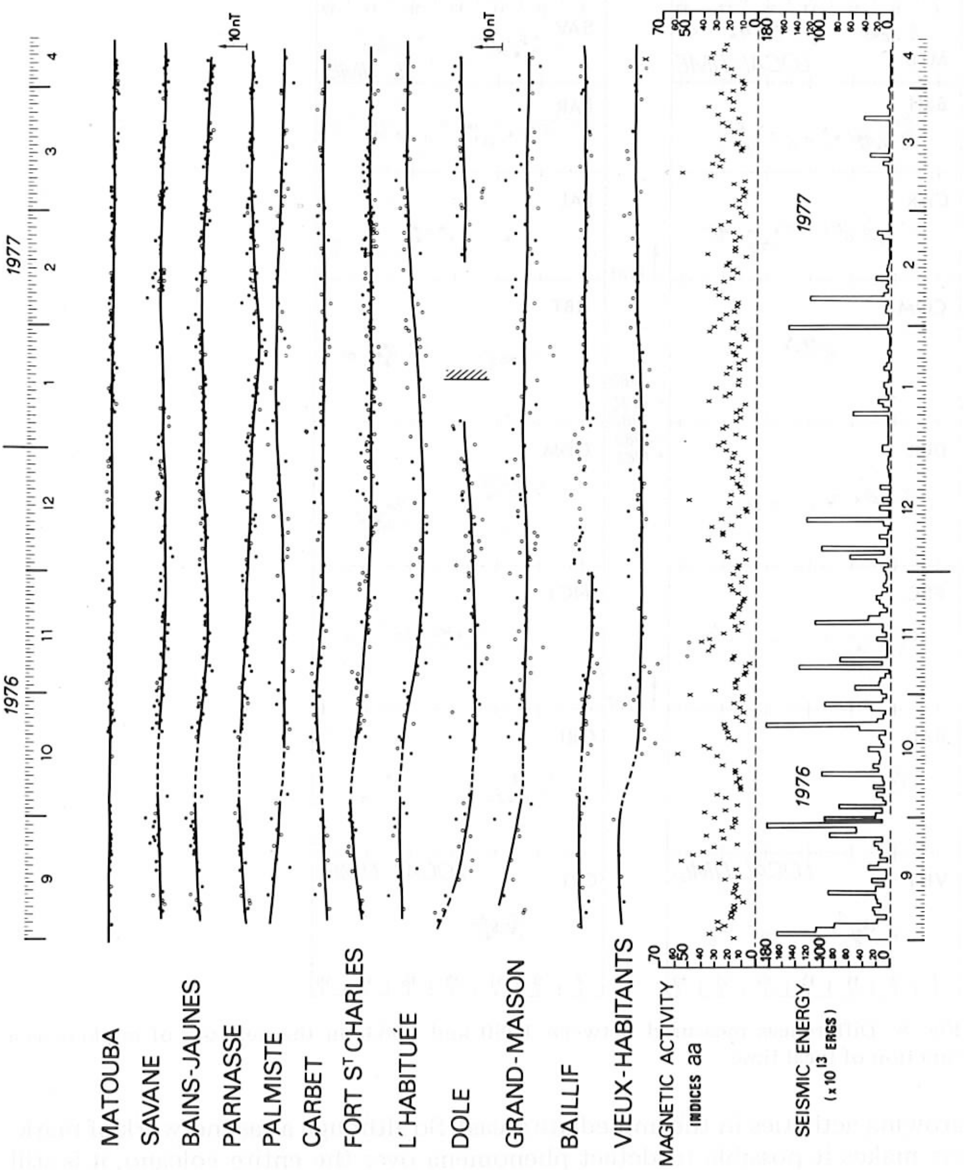
growing activities in the immediate areas). So although a vast network of markers makes it possible to detect phenomena over the entire volcano, it is still much less accurate than a telemetering network (by at least a factor of 2).

5. SEISMOVOLCANIC CRISIS OF 1976—1977

5a. Variations over periods of a few days

It is interesting to compare the results obtained during the seismovolcanic crisis of 1976—1977 with those obtained since that time.

Figure 9 shows the average values, during the crisis, of two series of 10 measurements performed at 10-min intervals at each marker. Great care


1976
1977
MATOUBA
SAVANE
BAINS-JAUNES
PARNASSE
PALMISTE
CARBET
FORT ST CHARLES
L'HABITUEE
DOLE
GRAND-MAISON
BAILLIF
VIEUX-HABITANTS
10nT
MAGNETIC ACTIVITY
INDICES aa
SEISMIC ENERGY
(x10^13 ERGS)

has been taken to distinguish between measurements performed before (full circles) and after (empty circles) 13.00 h local time. The average indices of magnetic agitation aa in the morning and afternoon, as well as the seismic energy released each day, are also shown. As for Figs. 10 and 11, they show, respectively for each marker, the reduced values ΔB as a function of local time (the ΔB-TL curve both during the crisis (September 1976 to April 1977) and when no notable seismic activity is any longer observed (from February 16 to April, 1977).

The variations ΔB with a low amplitude are observed at the markers near the volcano for which reduction errors are low (Tables 1 and 2), namely MTB, SAV, BJH, PAR, and PAL. The markers at which variations are significant are essentially the remote markers at which reduction errors are

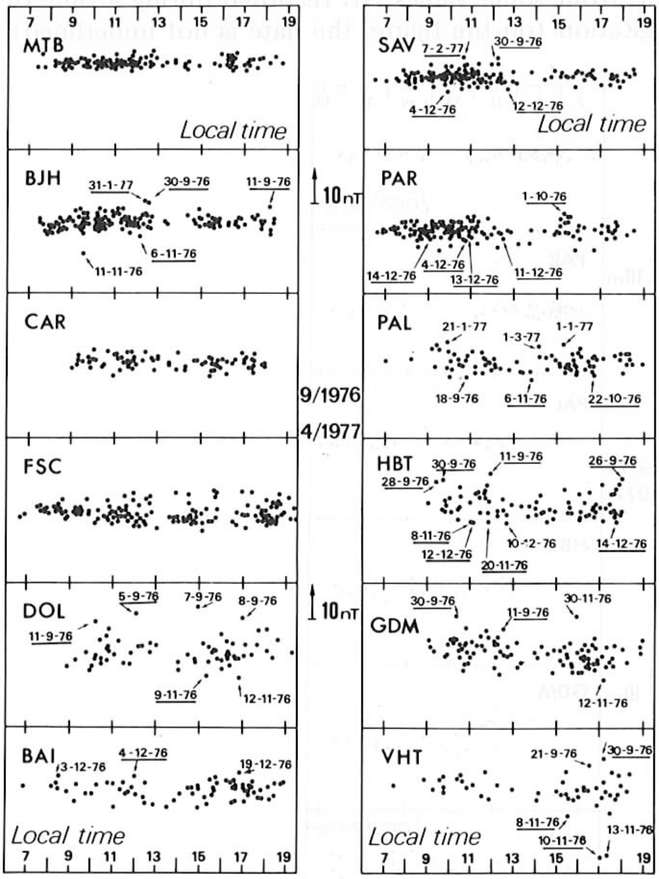


Fig. 10. Differences ΔB measured during the seismovolcanic crisis as a function of local time (9-1976/4-1977).

Fig. 9. Differences measured over the network of markers (○: before 13 TL, ● after 13 TL) during the seismovolcanic crisis (9-1976/4-1977). Indices of magnetic agitation aa (morning and afternoon) and seismic energy released daily. The continuous curves show the long-term variations.

substantial (GDM, DOL, HBT, and VHT). However, the temporal distribution of the points ΔB after the crisis (February—April 1977, Fig. 11) — which is similar to that of 1979 (Fig. 5) — is clearly different from the temporal distribution of the points ΔB during the crisis (Fig. 10); volcanomagnetic variations are superimposed on the transient variations that persist in the reduction of data.

Reduction errors, the sampling interval of the measurements, which is equal to or greater than one day, and measurements made sometimes in the morning and sometimes in the afternoon, do not allow an accurate recording of volcanomagnetic variations over periods less than a few days. However, it is possible to draw some general conclusions (Fig. 10). Not all seismic crises give rise to variations ΔB that differ from the average curve ΔB-LT. On the other hand, aside from some values ΔB recorded during a time of considerable magnetic agitation (on the figure, the date is not underlined),

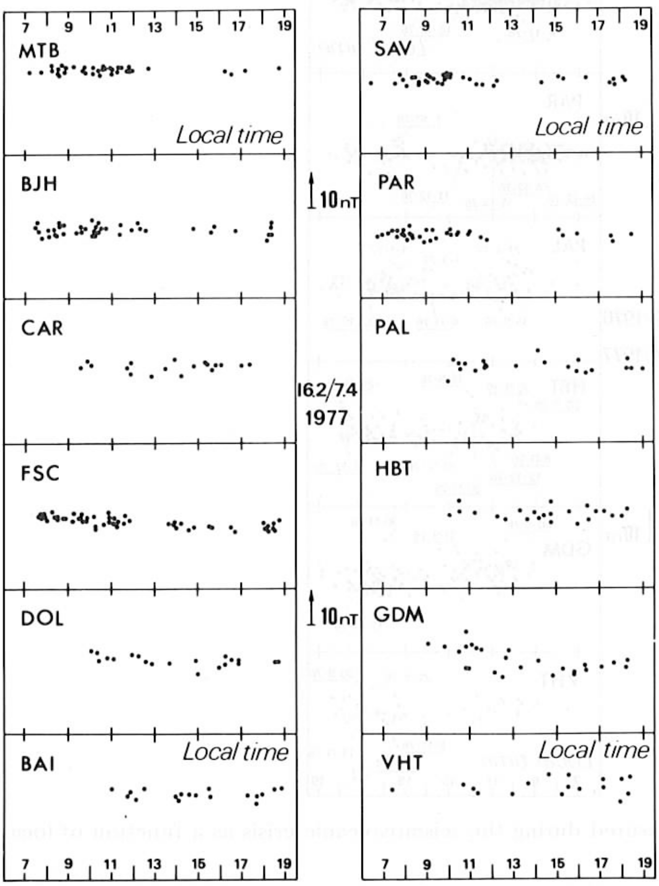


Fig. 11. Differences measured after the seismovolcanic crisis (16.2.77 to 7.4.77) as a function of local time.

all the values ΔB outside the average distribution ΔB-LT correspond to seismic crises (the date is underlined). Dates are omitted when variations over long periods mask rapid variations (e.g., FSC and CAR). The very substantial seismic crisis at the end of September 1976 (with an energy level exceeding 1.8×10^{15} ergs) seems to be sensed by at least 6 markers (SAV, BJH, PAR, HBT, GDM, and VHT), while less substantial seismic crises are not systematically picked up by these same markers. For a given seismic crisis, it seems that the magnetic responses of the markers have the same direction (e.g., the beginning of September, 10.9.76).

Although it is very difficult to hazard a guess as to the order of magnitude of volcanomagnetic variations over short periods, it may simply be noted that these variations do not exceed 4 nT (deviation from the average value) at markers close to the summit (SAV, BJH, PAR, and PAL) and approximately 7 or 8 nT at more remote markers (HBT, DOL, and GDM).

5b. Variations over long periods

It is possible to estimate the volcanomagnetic variations over periods of several weeks on the basis of the coefficient β_{po} (sign and value). At each marker (Fig. 9), the smoothed curve corresponding to long-term variations was superimposed on the reduced value ΔB between September 1976 and April 1977. The curves obtained show variations that cannot be due to reduction errors (see Table 4, for example) or to the sampling interval. They can be grouped into three sets with the same magnetic behavior.

(a) Markers SAV and BJH. The anomalies are low, of the order of 2.5 nT at SAV and 4.5 nT at BJH. The differences ΔB increase from 2 to 3 nT approximately during the month of September and decrease towards the end of October. Beginning in mid-November, the differences ΔB very slowly tend toward a limit that seems to be reached around the end of February 1977.

(b) Markers PAL, DOL and GDM. The volcanomagnetic variations are substantial; they reach 4 nT at PAL, 7 nT at GDM and about 10 nT at DOL. Each of these markers shows a sharp decrease in the difference ΔB from September to October 1976, and then a slow return to a stable level. (February—March 1977).

(c) Markers PAR, CAR, HBT, FSC, BAI and VHT. The variations range between 4 nT (CAR) and 7 nT (HBT, VHT). All these markers show either an increase in the difference ΔB in September or a maximum ΔB value (HBT, BAI, VHT). The ΔB's then decrease from mid-October onward and reach a minimal value between November and December 1976 (January 1977 at PAR). At BAI, the values ΔB measured in December 1976 seem to be out of line by 5 nT with respect to adjacent measurements.

We do not have magnetic data from before the volcanic crisis that could serve as reference base, and thus we cannot assert that the references bases before and after the crisis are the same. But the reversibility of the varia-

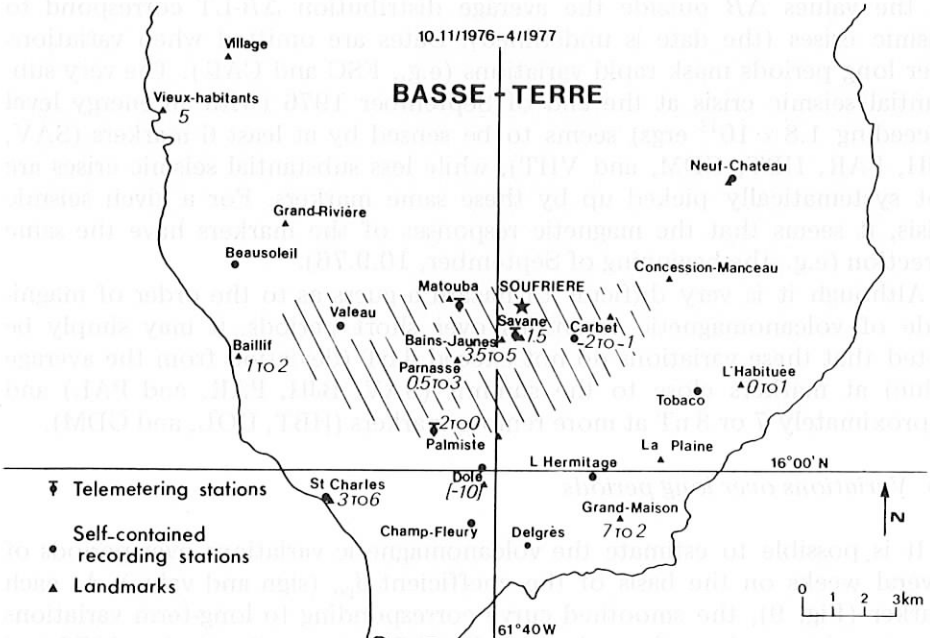


Fig. 12. Volcanomagnetic anomalies from the beginning to the end of September 1976.

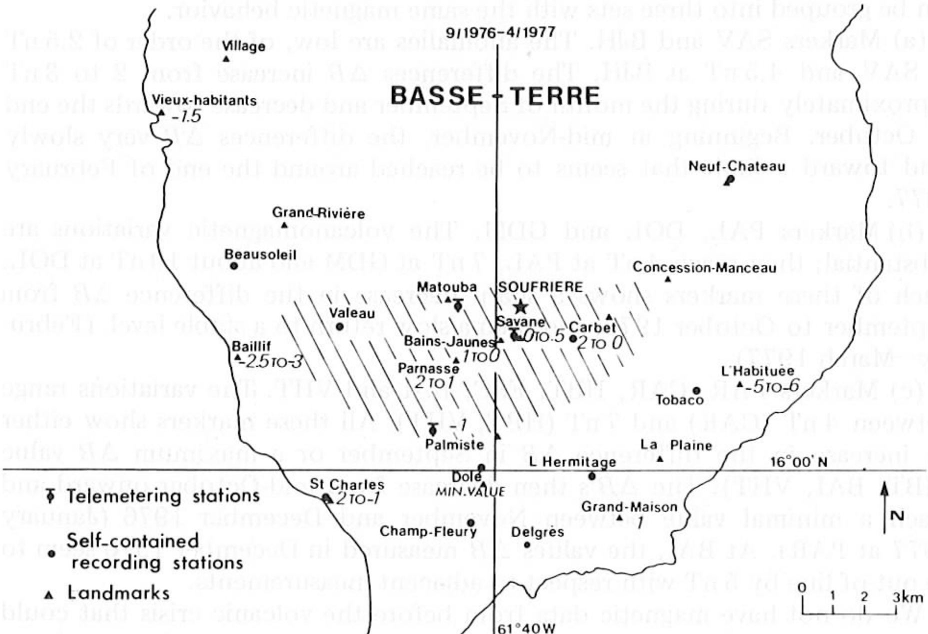


Fig. 13. Volcanomagnetic anomalies from October to November 1976.

tions observed (as far as the accuracy of the measurements allows) during the crisis leads us to take as an approximate reference base the differences ΔB measured beginning in March 1977, when the crisis is virtually over (Dorel and Feuillard, 1980; Feuillard et al., 1983). It is thus possible to map the magnitude of volcanomagnetic anomalies at a given period over the entire volcano. The two selected sequences extend, first, for the beginning to the end of September (Fig. 12) and, secondly, from the end of October to November, 1976 (Fig. 13):

(a) During the month of September 1976 (Fig. 12), the observed variations are positive (except at CAR) and intense. They attain 7 nT at the beginning of the month (GDM) and 6 nT at the end of September (FSC). While anomalies increase during September, over the entire network, they decrease at the three markers south of the volcanic mountain area (PAL, DOL, and GDM).

(b) From the end of October to November 1976 (Fig. 13), the anomalies are much smaller; they do not exceed 3 nT (except at HBT). Either they are positive and close to zero (at markers close to the dome): PAR, BJH, SAV, and CAR) or they are negative (VHT, BAI and HBT).

These two examples show very well that variations over periods close to 1 to 2 months are superimposed on volcanomagnetic variations over short periods. The former variations are sharp and cover the entire network of markers (20 km × 20 km). They show that the phenomena have a deep origin. The data do not make it possible to determine clearly whether the variations at each of the markers are in phase. However, the general behavior of the markers during the crisis can be linked to the three activity phases distinguished by Feuillard et al. (1983) during the crisis. The first phase extends from July 8 — date of the first phreatic eruption — to November 10, 1976. During this period, seismic activity was intense (1257 seisms recorded on August 24) and the many phreatic eruptions were accompanied by tremors (Sheridan, 1980). The magnetic anomalies ΔB are sharp and attain an extreme value between the middle and end of October. The second phase corresponds to a period of 56 days without substantial seismic activity. Only emissions of gas and ashes were observed. Magnetic anomalies decrease, generally pass through a second extreme value, and then slowly increase again in December. The third period begins on January 5 and ends at the end of February 1977. Seismic activity, though much lower than in August 76 resumed and was accompanied by phreatic eruptions. During this sequence, the magnetic anomalies were more diffuse and the variations gradually tended toward a stable level after February, 1977.

5c. Discussion

Several physical processes may explain the observed variations: thermal diffusion, shifts of magnetized matter, electrokinetic effects, piezomagnetic effects, or effects linked to dilatancy.

(a) Thermal diffusion. The phenomenon with its time constants of the order of several years cannot explain the reversible variations observed during the crisis and no long-term variation seems to be detected on the markers of the network.

(b) Shifts of magnetized matter. Peripherical modifications at the dome of the distribution of magnetic masses (emitted ashes cooling in the earth's magnetic field) and in the tectonic structure of the dome during phreatic eruptions (opening of faults) (Feuillard et al., 1983) caused no gravimetric anomaly, deformation, and no irreversible magnetic effect. This hypothesis must therefore be rejected.

(c) The three other hypotheses (electrokinetic, piezomagnetic and dilatancy effects) are closely linked. Indeed, a variation in the stress field in given surroundings can modify not only the induced and natural magnetizations of magnetized structures (piezomagnetic effect) (Ohnaka and Kinoshita 1968; Pozzi, 1973; Martin et al., 1978; Zlotnicki et al., 1981), but also the interconnected pore networks (microcracks, faults, aquifers). In expanding regions, variations in resistivity are linked to modification of the pore network (Brace et al., 1965; Nur, 1972; Scholz et al., 1973) and because of the earth's magnetic field, the structures concerned may induce abnormal magnetic variations. In the same way, in expanding regions where water circulates, the pore pressure gradients create an electrofiltration current that can generate magnetic anomalies (Nourbehecht, 1963; Mitzutani et al., 1976; Fitterman, 1979). These three phenomena may be present at La Soefrière on Guadeloupe. This region is dotted with faults, some of which are active, like the Ty fault (I.P.G., 1976, 1977; Feuillard et al., 1983), with many aquifers, and heat sources (Cormy et al., 1970; Feuillard, 1976).

Assuming a purely piezomagnetic effect, Pozzi et al., (1983) used an axisymmetric elastic model to calculate the piezomagnetic anomaly created by an overpressure of 10 MPa in a magnetic chamber the depth of which is estimated to be 6 km under the dome (Hirn and Michel, 1979; Semet et al., 1982). The piezomagnetic coefficients used are those measured at the laboratory (Zlotnicki et al., 1981) and extended to the entire volcano, assumed to be homogeneous. The influence of pore pressure is not taken into account. The deformations calculated are smaller than the background noise of the measurements (of the order of 3 cm); the piezomagnetic anomaly obtained with an amplitude of 0.9 nT extends over the entire volcanic mountain area (Pozzi et al., 1983) (Fig. 14). To compare the results calculated with the measurements reduced with respect to MTB station O near the dome, one must notice that for any given marker P:

$$\begin{aligned} &\Delta V(P, \mathrm{O}, t) > 0 \qquad \text{for a compression} \\ &\Delta V(P, \mathrm{O}, t) < 0 \qquad \text{for a decompression (O:MTB)} \end{aligned} \tag{10}$$

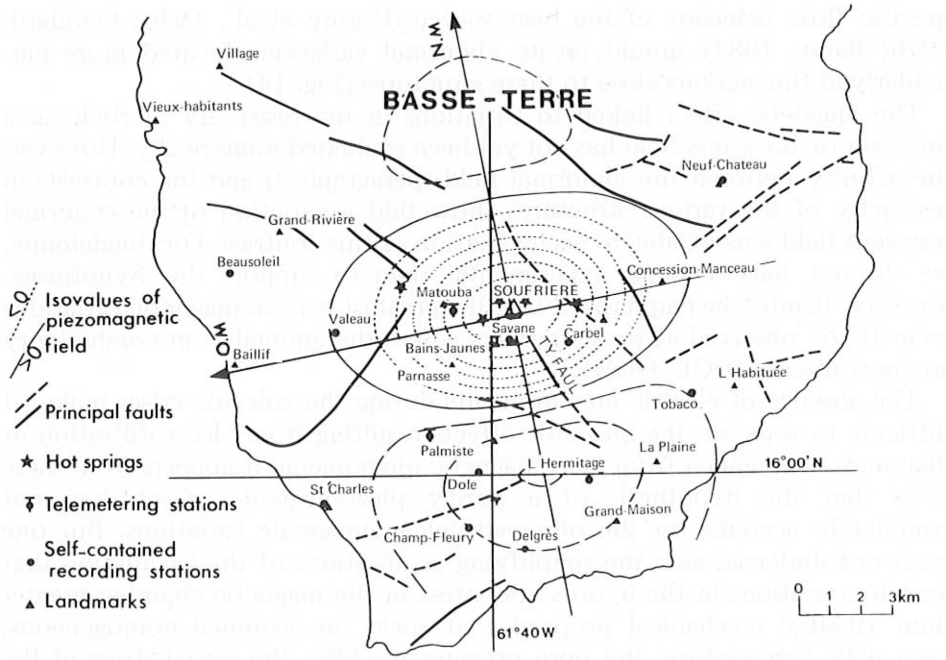


Fig. 14. Major structural lines and superimposition of the piezomagnetic anomalies of the field calculated by Pozzi et al. (1983) for an overpressure of 10 MPa in a magnetic chamber 6 km under the dome.

In all cases, the most remote markers show the most substantial effects.

The comparison with the data makes it possible to emphasize a few points:

(a) Generally (except for HBT), the volcanomagnetic variations are indeed more intense at markers far away from the volcano.

(b) The great magnetic modulations observed over the entire network, correlated with the three activity phases described by Feuillard et al. (1983), can be explained in part according to a scheme of stress field increases and decreases. But, up to November 1976, three markers (PAL, DOL, and GDM) had magnetic signatures opposite to those of the other markers.

(c) Volcanomagnetic variations have an amplitude that may be as high as 7 or 8 nT. But these results, according to the numerical model of the piezomagnetic effect (Pozzi et al., 1983), would lead to high estimates of stress variations (of the order of 70 MPa) and also of ground deformations (of the order of 20 cm).

The spatial distribution and the intensity of electrokinetic effects are highly dependent on those of the conducting channels (microcracks, faults, etc.) and on their response to a pore pressure gradient and to the temperature (Fitterman, 1979; Ishido and Mitzutani, 1981). The great spatial heterogeneity of the superficial faults and the aquifers (Fig. 14) as well as the specific flow behavior of the heat source (Cormy et al., 1970; Feuillard, 1976; Barat, 1984) should create abnormal variations located more particularly at the markers close to these structures (Fig. 14).

The magnetic effect linked to variations in the resistivity of rocks as a function of the stress field has not yet been evaluated numerically. However, the affinity between the abnormal field (paragraph 3) and the contrasts in resistivity of the various structures show that a variation of the abnormal transient field is associated to each variation of this contrast. On Guadeloupe, we do not have enough experimental data to support this hypothesis: however, it must be emphasized that the greatest volcanomagnetic variations seem to be observed at those markers where the anomalies in conductivity are most intense (DOL, GDM).

The absence of electric measurements during the volcanic crises makes it difficult to separate the magnetic effects resulting from electrofiltration or dilatancy phenomena from piezomagnetic phenomena. It appears from these facts that the hypothesis of a purely piezomagnetic effect does not completely account for the observed volcanomagnetic variations. But one must not underestimate the simplifying assumptions of the calculation that lead to variations in the hydrostatic stress in the magnetic chamber greater than 70 MPa; mechanical properties of rocks are assumed homogeneous, even with temperature; the pore pressure modifies the actual stress of the rocks and thus the piezomagnetic coefficients; the entire mountain area is assumed to be magnetically homogeneous, whereas it has been shown (Zlotnicki and Cornet, 1986) that, for a given variation in the stress field, magnetic environments of different characteristics may induce greater piezomagnetic anomalies.

6. CONCLUSIONS

On Guadeloupe, significant contrasts in conductivity of the superficial geological structures create highly nonhomogeneous abnormal transient variations. These abnormal variations are as high as 25% of the transient variations of the earth's magnetic field. It follows that the mere difference of the intensities of the magnetic fields at any two points on the mountain area leaves certain variations linked to the agitation of the earth's magnetic field. Those reduction errors linked to each marker of the magnetic network partly conceal the volcanomagnetic effects and the accuracy of such a network is less than 2 nT. A simplified study of the volcanomagnetic effects by means of telemetering stations set up outside the zone of high abnormal conductivity can be carried out and the accuracy is of the order of 1 nT.

The absence of any notable volcanic activity between 1978 and 1984 makes it possible to re-examine the volcanomagnetic effects observed during the crisis of 1976—1977. Though it may be said that there are volcanomagnetic effects with periods of a few days or less, the network of data reduction and the sampling interval that are used do not make it possible to describe them correctly. On the other hand, long-term variations remain recognizable. Their temporal variations and their amplitudes can be linked to the great phases of eruptive activity described by Feuillard et al. (1983). The fact that no ground deformation or gravimetric anomaly was observed and that no electric measurements (resistivity, spontaneous polarization) were performed during the crisis makes the interpretation of the volcano-magnetic effects a delicate matter. The amplitude of the magnetic anomalies (up to 7 or 8 nT) would seem to indicate overpressures greater than 70 MPa in the magmatic chamber (a numerical model of an elastic, axisymmetric piezomagnetic effect) but because of this the overpressure would generate deformations of the mountain area that would be too great to escape notice. On the other hand, it seems that only the stress variations can generate mechanisms capable of producing the volcanomagnetic variations throughout the entire structure.

A better knowledge of the relation between volcanomagnetic effect and volcanic activity thus seems to require two complementary approaches:

(a) on the sites under study, continuous and digital measurements with telemetering stations in which reduction errors are numerically subtracted (Zlotnicki et al., 1986) should be conducted concurrently with deformation and electric measurements;

(b) numerical simulations should be conducted on the existence of interactions among the various mechanisms: piezomagnetic effects taking pore pressure into account, effects linked with modifications in the resistivity of structures in connection with stresses, and effects linked with electro-filtration.

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REFERENCES

Barat, A., 1984. Etude du rôle des eaux souterraines dans le mécanisme des éruptions phréatiques. Application à la Montagne Pelée de Martinique et à la Soufrière de Guadeloupe. Thèse de spécialité, Université de Bordeaux III.

Beahm, T.J., 1976. Geomagnetic field gradient measurements and noise reduction techniques in Colorado. J. Geophys. Res., 81: 6276—6280.

Brace, W.F., Orange, A.S. and Madden, T.R., 1965. The effect of pressure on the electrical resistivity of water saturated crystalline rocks. J. Geophys. Res., 70: 5669—5679.

Briden, J.C., Rex, D.C., Fuller, A.M. and Tomblin, J.F., 1979. K-Ar geochronology and paleomagnetism of volcanic rocks in the Lesser Antilles Islands Arc. Philos. Trans. R. Soc. London, 291: 485—528.

Cormy, G., Demians d'Archimbaud, J., and Surcin, J., 1970. Prospection géothermique aux Antilles françaises; Guadeloupe et Martinique. Symposium des Nations Unies sur le Développement et l'Utilisation des Ressources Géothermiques, Pisa, 39 pp.

Davis, P.M., Stacey, F.D., Zablotki, C.J. and Ioben, J.V., 1979. Improved signal discrimination in tectonomagnetism — discovery of a volcano-magnetic effect at Kilauea, Hawaii. Phys. Earth Planet. Inter., 19: 331—336.

Davis, P.M., Pierce, D.R., McPherron, R.L., Dzur, D., Murray, T., Johnston, M.J.S. and Muller, R., 1984. A volcanomagnetic on Mount St. Helens. Geophys. Res. Letters 11 (3): 233—236.

Dorel, J. and Feuillard, M., 1980. Note sur la crise sismo-volcanique à La Soufrière de la Guadeloupe, 1975—1977. Bull. Volcanol., 43 (2): 419—430.

Feuillard, M., 1976. Etudes thermiques à La Soufrière de Guadeloupe. Bull. B.R.G.M., 4 (4): 375—388.

Feuillard, M., Allegre, C.J., Brandeis, G., Gaulon, R., Le Mouël, J.L., Mercier, J.C., Pozzi, J.P., and Semet, M., 1983. The 1975—1977 crisis of La Soufrière de Guadeloupe (F.W.I.): A still-born magmatic eruption. J. Volcanol. Geotherm. Res., 16: 317—334.

Fitterman, D.V., 1979. Theory of electrokinetic magnetic anomalies in a faulted half space. J. Geophys. Res., 84: 6031—6040.

Hirn, A. and Michel, B., 1979. Evidence of migrations of main shocks during major seismovolcanic crisis of La Soufrière (Guadeloupe) F.W.I. in 1976. J. Volcanol. Geotherm. Res., 6: 295—304.

I.P.G., 1976. Crise sismo-éruptive de La Soufrière de Guadeloupe (Petites Antilles): 1975—1976. Institut de Physique du Globe, Paris, 161 pp.

I.P.G., 1977. Observations volcanologiques: La Soufrière de Guadeloupe (novembre 1976—april 1977). Institut de Physique du Globe, Paris, 67 pp.

Ishido, T. and Mitzutani, H., 1981. Experimental and theoretical basis of electrokinetic phenomena in rock—water systems and its application to geophysics. J. Geophys. Res., 86, 1763—1775.

Johnston, M.J.S. and Stacey, F.D., 1969a. Volcano-magnetic effect observed on Mt. Ruapehu, New Zealand. J. Geophys. Res., 75: 6541—6544.

Johnston, M.J.S. and Stacey, F.D., 1969b. Transient magnetic anomalies accompanying volcanic eruptions in New Zealand. Nature, 224: 1289—1291.

Johnston, M.J.S., Muller, R.J., and Dvorak, J., 1981. W. Lipman and D.R. Mullineaux (Editors), The 1980 Eruption of Mount St. Helens, Washington. U.S. Geol. Surv., Prof. Pap., 1250: 183—189.

Lacroix, A., 1904. La Montagne Pelée et ses éruptions. Masson, Paris, 2 vol.

Le Mouël, J.L. and Menvielle, M., 1982. Geomagnetic variations anomalies and deflection of telluric events. Geophys. J.R. Astron. Soc., 68: 575—587.

Le Mouël, J.L., Menvielle, M., Pozzi, J.P., Simon, G. and Zlotnicki, J., 1984. Magnetic observations on La Soufrière volcano (Guadeloupe, F.W.I.) over a quiet period. J. Geomagn. Geoelectr., 36: 609—619.

Martin, R.J., Habermann, R.E. and Wyss, M., 1978. The effect stress and cyclingand inelastic volumetric strain on remanent magnetization. J. Geophys. Res., 83: 3485—3496.

Martin-Kaye, H.A., 1969. A summary of the geology of the Lesser Antilles. Overseas Geol. Miner. Resourc., 10 (2): 172.

Mayaud, P.N., 1973. A hundred year series of geomagnetic data 1868—1967: indices aa; storm suddent commencents. I.A.G.A. Bull. 33.

Menvielle, M., 1984. Anomalies de conductivité et phénomènes de canalisation. Thèse, Paris 6.

Mitzutani, H., Ishido, T., Yokokura, T. and Ohnishi, S., 1976. Electrokinetic phenoma associated with earthquakes. Geophys. Res. Lett., 3 (7): 365—368.

Nourbehecht, B., 1963. Irreversible thermodynamic effects in inhomogeneous media and their applications in certain geoelectric problems. Ph. D. Thesis, M.I.T., Cambridge, Mass.

Nur, A., 1972. Dilatancy, pore fluids, and premonitory variations of T_s/T_p travel times. Bull. Seismol. Soc. Am., 62: 1217—1222.

Ohnaka, M. and Kinoshita, H., 1968. Effects of uniaxial compression on remanent magnetization. J. Geomagn. Geoelectr., 20: 93—99.

Pozzi, J.P., 1973. Effets de pression en magnétisme des roches, Thèse, Univ. Paris 6.

Pozzi, J.P., Le Mouël, J.L., Rossignol, J.C. and Zlotnicki, J., 1979. Magnetic observations made on La Soufrière volcano (Guadeloupe) during the 1976—1977 crisis. J. Volcanol. Geotherm. Res., 5: 217—237.

Pozzi, J.P., Philippe, C., Le Mouël, J.L., Rossignol, J.C. and Zlotnicki, J., 1983. Anomalies magnétiques et déformations associées à l'activité de volcans andésitiques. Exemple de La Soufrière de Guadeloupe. Bull. P.I.R.P.S.E.V., 79, 31 pp.

Rikitake, T., 1951. Changes in magnetic dip that accompanied the activity of volcano Mihara. Tokyo Univ. Earthq. Res. Inst. Bull., 19: 499—502.

Robach, F., Le Bof, M. and Rey-Graud, G., 1976. Mesures magnétiques et activité volcanique, Soufrière 1976. Note L.E.T.I./M.A., 76-952 FR/AL C.E.N., Grenoble.

Robson, J.R. and Tomblin, J.F., 1966. Catalogue of the active volcanoes and solfatara fields of the West Indies. In: Catalogue of the Active Volcanoes of the World, part. XX. Int. Assoc. Volcanol., Naples, Italy.

Scholz, C.H., Sykes, L.R. and Aggarwal, Y.P., 1973. Earthquake prediction: A physical basis. Science, 181: 803—810.

Semet, M. and Vatin-Perignon, N., 1979. The 16th century activity of La Soufrière de Guadeloupe (F.W.I.): a model for explosive andesite volcanism in the Lesser Antilles (Abstr.). EOS, Trans. Am. Geophys. Union, 60: 833.

Semet, M., Vatin-Perignon, N., Vincent, P. and Joron, J.L., in prep. Magma mixing once more: its involvement in triggering the 16th century volcanic activity at La Soufrière, Guadeloupe (F.W.I.). J. Volcanol. Geotherm. Res.,

Sheridan, M.P., 1980. Pyroclastic block flow from the september, 1976, eruption of La Soufrière volcano, Guadeloupe. Bull. Volcanol., 43 (2): 397—402.

Uyeda, S., 1961. An interpretation of the transient geomagnetic variations accompanying the volcanic activity of volcano Mihara (Oshima Island). Jpn. Bull. Earthq. Res. Inst. (Tokyo), 39: 579.

Vincent, P.M., Vatin-Perignon, N., Semet, M. and Cheminée, J.L., 1979. Le dôme de la Soufrière (Guadeloupe): son âge et son mode de mise en place. C.R. Acad. Sci. Paris, Sér. D, 288: 51—54.

Yokoyama, I., 1957. Geomagnetic studies of volcano Mihara. The 8th paper (continuous observations of changes in geomagnetic declination during the period 1955—56). Bull. Earthquake Res. Inst., 35: 567—572.

Zlotnicki, J. 1979. Applications du piézomagnétisme à la Géophysique. Contribution à l'interprétation d'effets volcanomagnétiques. Thèse de spécialité, Paris 6, 206 pp.

Zlotnicki, J. and Cornet, F.H., 1986. A numerical model of earthquake induced piezo-magnetic anomalies. J. Geophys. Res., 91: 709–718.
Zlotnicki, J., Pozzi, J.P. and Cornet, J.P., 1981. Investigation on induced magnetization variations caused by triaxial stresses. J. Geophys. Res., 86: 11899–11909.
Zlotnicki, J., Verhille, R., Viodé, J.P., Delmond, J.C., Simon, G. and Guillemant, P., 1986. Magnetic network on the Montagne Pelée volcano (Martinique, Lesser Antilles). A trial to discriminate volcanomagnetic signals. J. Geomagn. Geoelec., 38 (1): 151–164.