42.

JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 89, NO. B13, PAGES 11,216–11,234, DECEMBER 10, 1984

The Volcanoseismic Swarms of 1981-1983 in the Tahiti-Mehetia Area, French Polynesia

JACQUES TALANDIER

Laboratoire de Géophysique, Commissariat à l'Energie Atomique, Papeete

EMILE A. OKAL

Department of Geological Sciences, Northwestern University, Evanston, Illinois

During the years 1981-1983, three intense seismic swarms took place in the Tahiti-Mehetia area of French Polynesia at the presumed location of the Society Island hot spot. The 1981 swarm featured 4000 earthquakes, with a maximum magnitude $M_L = 4.3$, in the immediate vicinity of the island of Mehetia; the 1982 swarm occurred along the flank of the major Teahitia seamount, and involved more than 9000 events; a second swarm occurred at Teahitia in 1983 and involved 3000 events. Although no precise constraint can be placed on the depth of individual events from their travel times to Polynesian stations, features in the evolution of the Mehetia swarm are generally consistent with the probable ascent of a magma body toward the surface. In the case of Teahitia, the recording of abundant tremors of both high and low frequency, particularly intense during the 1983 swarm, is directly similar to cases of documented volcanic eruptions. The swarms are interpreted as episodes of active volcanism, part of the process of building the next island in the chain.

INTRODUCTION

The purpose of this paper is to report three intense seismic swarms which occurred in 1981, 1982, and 1983 in the Tahiti-Mehetia area of the Society Islands, and to analyze them in the framework of the evolution of the underlying hot spot, presently believed to be in the early stages of formation of the next island in the chain. In terms of the importance of volcanic edifices, the Society Islands make up the second best developed among Pacific Ocean hot spot chains. *Duncan and McDougall* [1976] have confirmed that the linear progression of their ages with distance is compatible with the motion of a rigid Pacific plate over a hot spot fixed with respect to the Hawaiian one. More recently, *Henderson and Gordon* [1981] have proposed that the same magma source may have contributed, at least partially, to the formation of islands on the Manihiki plateau and in the Tokelau and Gilbert chains.

A major difference between the Society and Hawaiian islands, however, is that the largest island of the Society Chain, Tahiti, is presently inactive, strongly eroded and fringed by a coral reef; it is estimated that its caldera collapsed about 1 m.y. ago, while the last documented episodes of volcanism go back approximately 400,000 years [*Becker et al.*, 1974]. On the other hand, the easternmost island in the chain, Mehetia, located about 130 km east of Tahiti, is an extremely small and steep cone (only 2 km^2 above sea level but with a maximum elevation of 435 m), along which no coral barrier has yet developed, and where preliminary analyses indicate alkalic volcanism [*Mottay*, 1976]. Thus it is likely that the hot spot is presently in the early processes of building the next island in the chain. In this respect, the Society Islands offer us the rather unique opportunity to look in real time at the transitional period to a new major volcanic edifice, an episode possibly comparable to the early genesis of the island of Hawaii, about 1 m.y. ago. Mehetia's situation is also somewhat reminiscent of Loihi Seamount, southeast of Hawaii [*Clague et al.*, 1981], although the petrology of the Society Islands (and, in particular the general lack of tholeiites on Tahiti) suggests intrinsic differences between the two chains.

Over the past 20 years a systematic study of the seismic activity in the immediate vicinity of Tahiti was made possible by the operation of the high-gain short-period stations of the French Polynesia Network on Tahiti, Moorea, and Rangiroa: results for the period 1963-1979, compiled by *Talandier and Kuster* [1976] and *Okal et al.* [1980], have identified approximately 30 seismic epicenters in a 100,000 km^2 area centered about 60 km east of Tahiti (see Figure 1). Approximately halfway between Tahiti and Mehetia, *Talandier and Kuster* [1976] also identified two sites of repeated seismic swarms, whose characteristics suggest active volcanism. Bathymetric surveys of these two sites later confirmed the existence of seamounts topping at 180 and 2100 m below sea level, respectively, and for which the names "Moua Pihaa" and "Rocard" were proposed.

Since these studies were published, the Tahiti-Mehetia area was the site of three major seismic swarms, whose intensity was by far greater than anything previously recorded in Polynesia: First, in 1981, about 4000 earthquakes occurred on the southeastern flank of Mehetia; then, in 1982, a swarm of more than 9000 recorded earthquakes took place in the vicinity of a known seamount, topping 1600 m below sea level, located 40 km from the Tairapu Peninsula, and for which the name "Te-ahi-tia" (the standing fire) was proposed by the Tahitian Academy (see Figure 2). Finally, in July 1983, a short-lived seismic swarm at Teahitia involved 3000 earthquakes, and was followed by several months of more or less continuous tremors. While the Teahitia area had a history of discrete earth-

Paper number 4B0419.
0148-0227/84/004B-0419$05.00

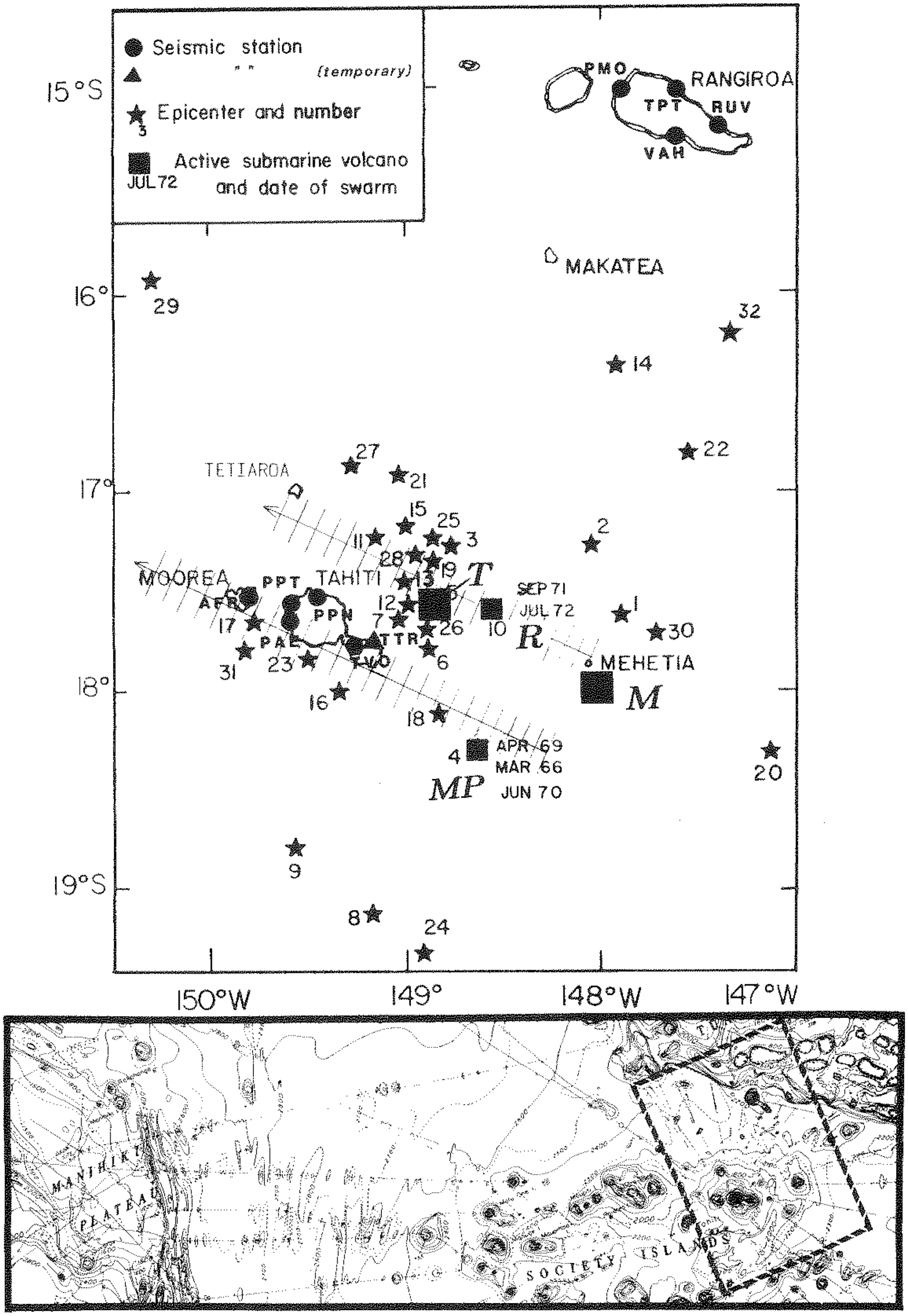


Fig. 1. (top) Map of the instrumental seismicity of the Tahiti-Mehetia area (adapted from *Okal et al.* [1980]). Permanent seismic stations are shown as solid circles (with code); the solid triangle identifies the temporary station at Tautira, operated during the 1982 swarm. Stars are epicenters of seismic activity predating the two major swarms (numbers refer to Table 3 of *Okal et al.* [1980]). The solid squares are the sites of volcanoseismic activity: MP, Moua Pihaa; R, Rocard; larger symbols are used for the three major swarms: M, Mehetia 1981, and T, Teahitia 1982 and 1983. The two faint raked arrows, drawn in the azimuth of absolute motion of the plate, identify potential lineations of volcanic activity. (bottom) Detailed section of *Mammerickx et al.*'s [1975] map replacing the top box into the framework of the Society chain and the nearby Manihiki plateau. The horizontal is oriented along the direction of absolute motion of the Pacific plate.

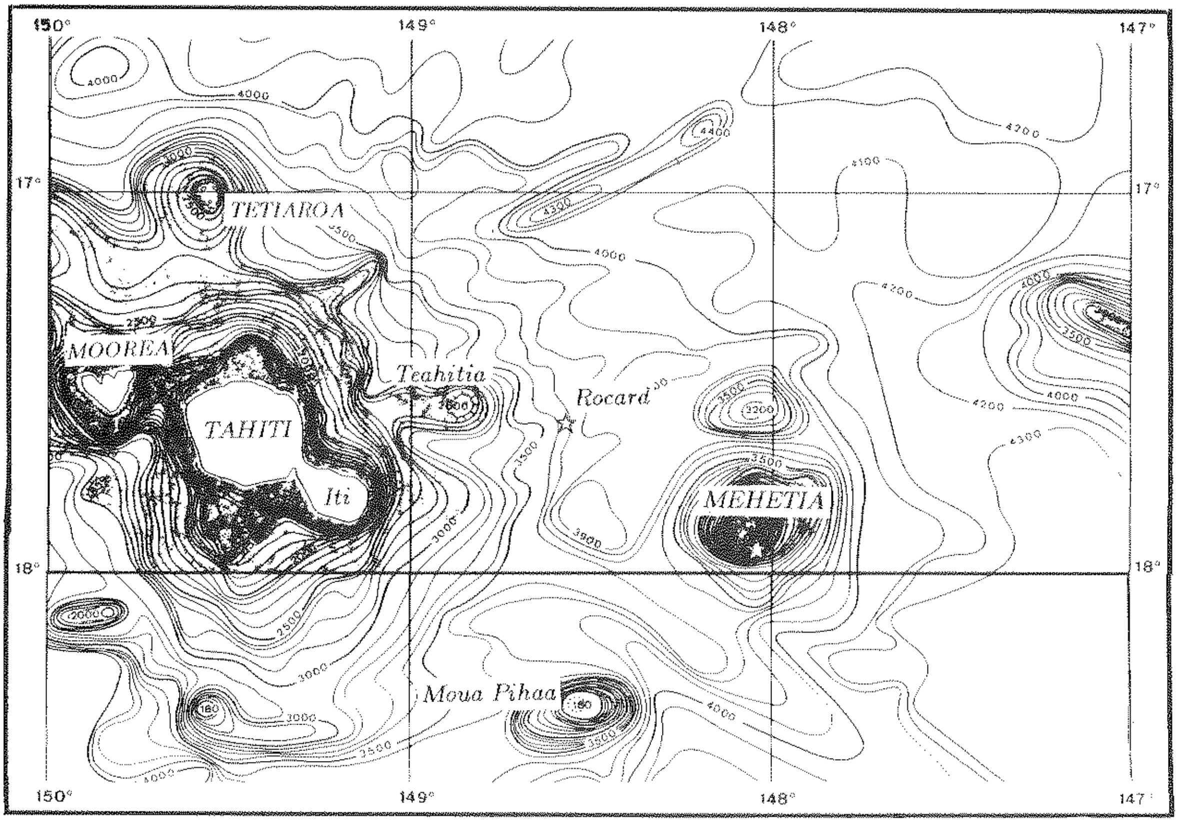


Fig. 2. Bathymetry of the Tahiti-Mehetia area, adapted from GEBCO maps 327 and 358 [*Monti and Pautot*, 1974]. Note that Moua Pihaa and Teahitia seamounts were charted prior to their activity but that Rocard seamount was identified only later (open star). The crater identified by the French Navy survey southeast of Mehetia is shown as a white star.

quake activity prior to the onset of the 1982 swarm, the 1981 swarm at Mehetia occurred in an area which had been seismically quiet, at the detection level of $M_L = 1.5$, since the implementation of the full network in 1965. The study and interpretation of the seismic swarms are the subject of the present paper.

NETWORK AND DETECTION CHARACTERISTICS

Network

The French Polynesia seismic network has been described in detail by *Talandier and Kuster* [1976] and *Okal et al.* [1980]. Its principal characteristics are its multiple subarray configuration and its special instrumentation allowing routine gains of 10^5 at 1 Hz and 2×10^6 at 3 Hz, more than an order of magnitude greater than for standard installations on oceanic islands. For the purpose of detection in the Tahiti-Mehetia area, the only relevant stations are the five stations on Tahiti and nearby Moorea, and the four stations on the atoll of Rangiroa, 350 km to the north. In order to improve the seismic coverage of the 1982 Teahitia swarm, an additional, temporary station (coded TTR) was operated starting March 30, 1982, at Tautira, the easternmost village on the peninsula, only about 40 km from the epicenter (see Figure 1). This station was not operating in 1983. During the 1981 Mehetia swarm, a 2-day expedition to this uninhabited island included the operation of a portable station on March 27 and March 28. Unfortunately, this corresponded to a period of major quiescence of the swarm, and very little information could be gathered from this portable station.

Magnitudes and Energy

Magnitudes are estimated using the formula

$$\log M_L = \log A + \log \Delta + 2.1 \qquad (1)$$

[*Talandier and Kuster*, 1976; *Okal et al.*, 1980], where A is the peak-to-peak amplitude in microns at a period close to 1 s and Δ is the epicentral distance in kilometers. The numerical constants in this formula are designed to lock the upper end of this magnitude scale onto the teleseismic m_b and are similar to those used with the Hawaii Volcano Observatory (HVO) Wood-Anderson instrument for seismic studies of Kilauea and Mauna Loa events.

An estimate of the seismic energy involved in the individual events was obtained using *Gutenberg and Richter*'s [1954] relation

$$\log E_s = 2.9 + 1.9 M_L - 0.024 M_L^2 \qquad (2)$$

where E_s is in joules. This formula was used in previous studies of the regional seismicity, where it was also found to match the relation $\log E_s = 4.8 + 1.5 M_S$ [*Gutenberg and Richter*, 1954]. It provides a comparative basis for discussing the regional output of seismic energy by various episodes of seismicity.

The detection thresholds for seismic activity centered at

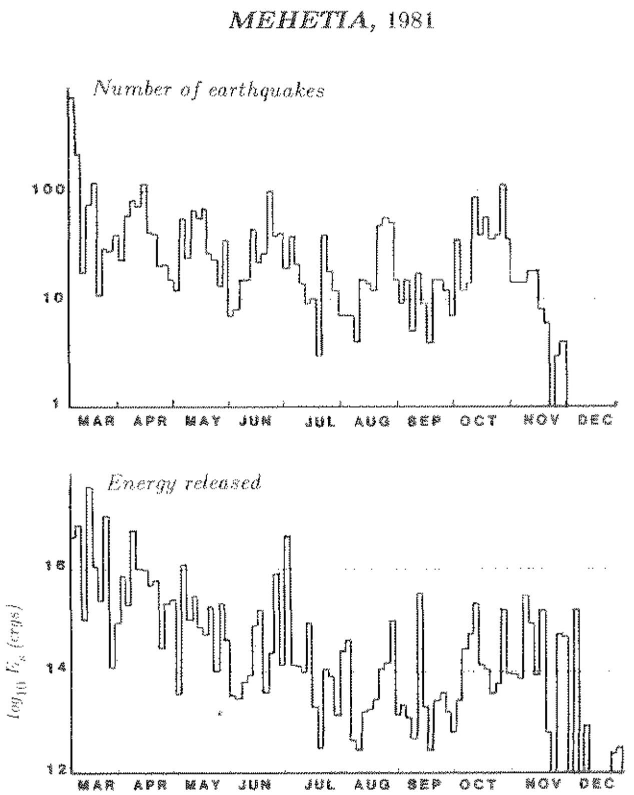


Fig. 3. Histograms of the number of earthquakes (top, total number: 3540; $M_L \geq 0.9$) and seismic energy released (bottom, total energy: 8.6×10^{10} J), using 3-day windows, for the total duration of the 1981 Mehetia swarm (301 days).

Mehetia and Teahitia can be estimated at $M_L = 1.1$ and 0.8, respectively, but detection below magnitude 1.5 at Mehetia and 1.0 at Teahitia is affected by day-to-day variations in the level of background seismic noise; on particularly quiet days, events were detected with $M_L = 0.9$ at Mehetia and 0.5 at Teahitia. Although these thresholds must be considered excellent in the oceanic environment, they remain higher than in the case of densely instrumented island sites, such as Kilauea or Mauna Loa. In particular, it is clear that smaller events, of the type recorded at Kilauea directly on the flanks of the volcano, would, if they exist, totally escape detection. Any comparison between these volcanic edifices must involve either a "magnitude filtering" of the Kilauea data set, or an extrapolation of the frequency-magnitude relations in Polynesia, which would suggest up to 50,000 events at Mehetia (52,000 events at Teahitia in 1982; 25,000 in 1983) at the $M_L \geq 0.1$ level. These figures are then comparable to the 10,000-30,000 events recorded at Kilauea during swarms lasting several months [*Koyanagi*, 1968].

OVERVIEW OF THE SWARMS

The 1981 seismic swarm at Mehetia started abruptly on March 6, 1981 at 00:35 UT, and lasted until December 1981, with some sporadic activity into 1982. Figure 3 describes its history, showing both the number of recorded earthquakes and the energy released, using 3-day windows; Figure 4 emphasizes the initial two weeks of the swarm, using 6-hour windows. It is immediately evident that this swarm features several distinct episodes, which will be more fully described in a later section. The largest event occurred on March 15, with a magnitude $M_L = 4.3$.

Figure 5 similarly presents the history of the number of events recorded and seismic energy released during the 1982 seismic swarm at Teahitia. Activity started abruptly on March 16 at 14:17 UT and increased regularly until March 27, involving mostly low-magnitude earthquakes. After March 27, earthquakes were accompanied by seismic tremors recorded by all five stations of the Tahiti-Moorea subarray, which lasted more or less permanently until April 8 (see Figure 6). Tremor activity then decreased and disappeared on April 18. Earthquake activity decreased regularly until May 19. After that date it dwindled to a number of rare, occasional events. The largest event was an $M_L = 3.4$ earthquake on April 1. Obvious differences with the Mehetia swarm are the shorter duration and more homogeneous character of the Teahitia one.

Figure 7 describes the history of the seismic activity at Teahitia during the 1983 swarm: It started very abruptly on July 12 but was relatively short-lived: after two very active days the intensity of the swarm decayed regularly, and it died off on July 24. The maximum magnitude reached was 2.5 on July 15. From July 12 to July 18, high-frequency tremors were present; very intense low-frequency tremors took over on July 21 and lasted nearly continuously since August 11. This feature is a fundamental difference between the two swarms at Teahitia; its observation could however be the result of a somewhat different epicenter, involving a more favorable path around the magmatic structure.

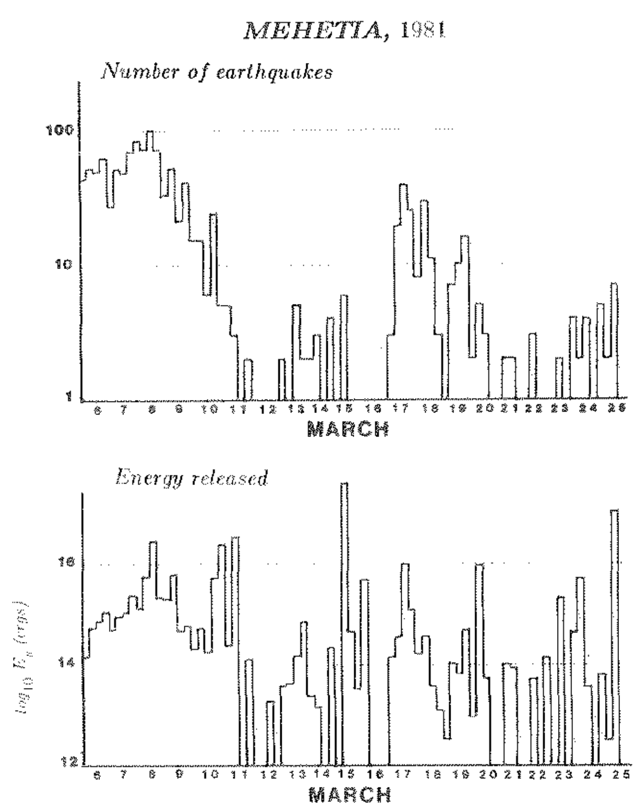


Fig. 4. Same as Figure 3, for the first 20 days of the swarm, using 6-hour windows, totaling 1199 events and 6.5×10^{10} J. Note the change of character of the two curves around March 15, when the number of earthquakes decreases sharply, but the energy released reaches a peak.

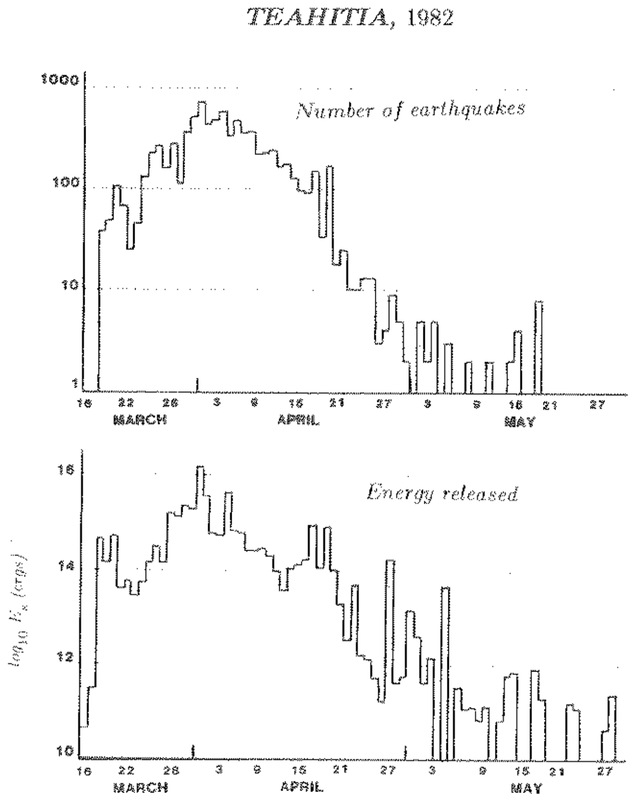


Fig. 5. Same as Figure 3, for the 1982 Teahitia swarm, using 1-day windows, for a total of 8038 earthquakes ($M_L \geq 0.5$), 3.8×10^9 J, over 77 days.

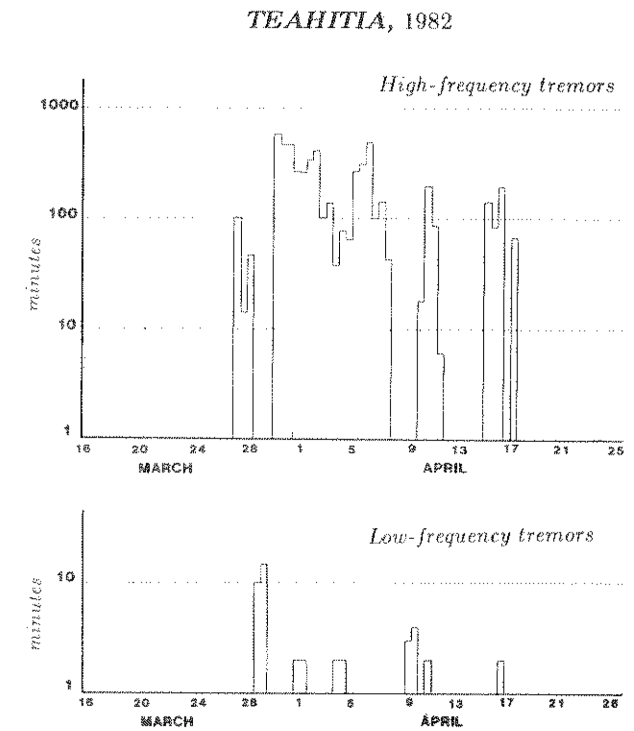


Fig. 6. Histograms of duration of seismic tremors recorded at the Tahiti subarray during the 1982 Teahitia swarm. (top) High-frequency tremors. (bottom) Low-frequency tremors. Note that the tremor activity starts well into the seismic swarm and also ends up rather abruptly well before the end of the seismic activity (see Figure 5). Note also inverse correlation between the two kinds of tremors. Sampling windows are 12 hours long.

Location Techniques

Location techniques are based on *Klein*'s [1978] HYPOINVERSE program. Four different crustal models, each involving three layers over a half-space, are used for (1) oceanic crust in the Tahiti-Mehetia area, (2) the northwestern Tuamotu plateau, (3) the Tahiti and Mehetia volcanic edifices, and (4) the edifice of the Rangiroa atoll. These models were obtained from seismic refraction experiments [*Talandier*, 1982] and are shown on Table 1. Accordingly, first arrivals from the Mehetia swarm are P_n waves at all stations; from the Teahitia area they are always P_n waves at the Rangiroa stations and at AFR on Moorea, always P_g waves at TVO and TTR on the peninsula, and can be either (depending on the exact location of the epicenter) at the remaining stations of the Tahiti subarray (PAE, PPT, and PPN). (We use the symbol P_g for the direct crustal phase, although the nature of the crust is, of course, basaltic in the oceanic environment.) High-speed paper playbacks of recorded signals allow reading errors for impulsive signals of no more than ± 0.025 s. This figure is negligible when compared with other sources of uncertainty, such as station anomalies, and the accuracy of the crustal models used.

The accurate determination of hypocenters in the vicinity of Mehetia suffers both from the relatively large distance to the closest station (TVO on the Tairapu Peninsula, 120 km from Mehetia) and from the repartition of all stations in two subarrays, concentrated around azimuths N10°E and N75°W from the epicenter. In particular, hypocentral depths could not be constrained by travel times alone. The

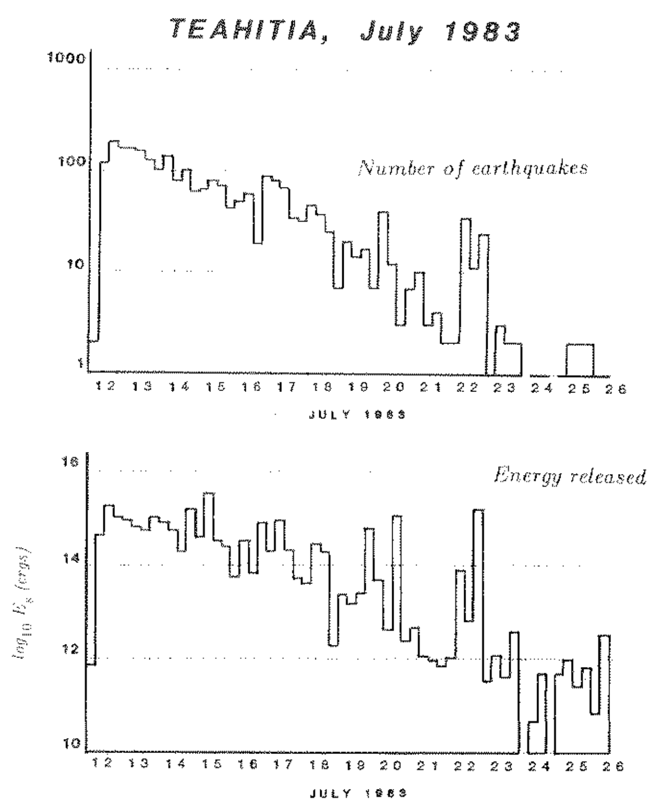


Fig. 7. Same as Figure 3 for the 1983 swarm at Teahitia, using 6-hour windows, for a total of 2471 earthquakes ($M_L \geq 0.5$), and 2.0×10^9 J, over 15 days.

TABLE 1. Seismic Structures Used in Relocations

Layer	Tahiti Thickness, km	Tahiti α, km/s	Oceanic Model Thickness, km	Oceanic Model α, km/s	Tuamotu Plateau Thickness, km	Tuamotu Plateau α, km/s	Rangiroa Thickness, km	Rangiroa α, km/s
Ocean			4.0	1.50	2.0	1.50		
1	0.4	2.00	0.4	2.00	1.4	2.40	2.0	3.30
2	5.8	4.37	1.8	4.37	2.8	4.65	4.8	4.65
3	6.6	7.64	6.6	7.64	24.5	6.83	24.5	6.83
Mantle		8.25		8.25		8.10		8.10

only available depth constraint came from the portable station operated for 2 days on the island of Mehetia in late March 1981: it recorded only one event of low magnitude, which unfortunately went undetected by stations of the permanent network. The $S-P$ interval for this record suggests a depth of 13 km, which according to Table 1 could be representative of the Mohorovičić (Moho) discontinuity. We chose to use this figure as a starting value for all Mehetia relocations. Significantly, HYPOINVERSE then failed to adjust the focal depths and elected to keep their values constrained. Similarly, in the case of Teahitia we used a starting depth of 12.8 km, which the station repartition was insufficient to constrain further. This situation is in contrast with the case of Loihi, where the many stations of the HVO network, providing homogeneous azimuthal coverage over 90°, and at distances as close as 35 km, make it possible to resolve focal depths to a precision of about 5 km [*Klein*, 1982].

We subjected to epicentral relocation 140 events from the Mehetia swarm, with clear arrivals both in Tahiti and Rangiroa, using the HYPOINVERSE routines. Since reading errors can be neglected, station residuals consist of a station correction, resulting from a local deviation of the crustal thickness under the station from that used in the model, and of a possible path effect. For clustered epicenters recorded at a distance large compared to the size of the cluster, these parameters will not vary significantly for individual events and can be modeled as a single station correction. These corrections, listed in Table 2, were obtained by averaging residuals from initial locations of the events, and then used in relocating the 140 events. Table 3 shows the improvement in residuals and hypocentral error parameters resulting from the relocations.

The final locations obtained for the 140 events are given in Table A1[1]. They were computed using between six and nine P_n arrivals as well as the S_n arrival on the horizontal short-periods at the central station PPT (the other stations are not equipped with horizontal instruments). Also included in Table A1 are the standard residuals ("SR") and the vertical and horizontal projections of the standard deviation ($1-\sigma$) ellipsoids ("HCE" and "VCE"). Following *Klein* [1978], we define the great axis of the horizontal ellipse as the largest among the horizontal projections of the three principal axes of the error ellipsoid. Because of the large uncertainty in hypocentral depth this definition actually overestimates its size. Results from Table 3 show the following:

First, standard residuals, already low for the initial locations (average value 0.068 s) are significantly improved by the relocations (average value 0.042 s), confirming that the most random parameters (reading errors) are of a negligible nature.

Second, large semiaxes of the horizontal $1-\sigma$ ellipses ("HCE" in Table A1) are again significantly reduced by the relocations (from an average value of 8.6 km to an average of 5.4 km); despite the general orientation of these axes along the bisector of the vectors pointing to the Tahiti and Rangiroa subarrays, this last value is definitely smaller than the horizontal extent of the epicentral area, as shown on Figure 8: the error ellipses are about the size of the clusters in the bottom box, while the total source area (approximately 40 km in length) is many times larger. This indicates that the source area of the swarm is truly elongated in the NNW-SSE direction.

Finally, despite a similar reduction in their absolute value, the vertical semiaxes ("VCE" in Table A1) remain about twice as large as the horizontal ones. They are much less meaningful, since the program did not readjust the depths and thus the relocations lack depth resolution.

The influence of the use of S times at PPT on the relocation of the earthquakes can be discussed as follows: The simultaneous use of P and S at PPT is equivalent to fixing the total distance traveled by the ray to PPT. For hypocenters located at or below the Moho this grossly results in con-

TABLE 2. Station corrections used in relocations

Station	Correction, s Mehetia	Correction, s Teahitia
AFR	−0.01	−0.11
PAE	−0.16	−0.03
PPT	+0.05	+0.01
PPN	+0.11	+0.07
TVO	−0.13	−0.02
TTR		+0.11
PMO	+0.10	0.00
VAH	+0.02	−0.27
TPT	+0.01	−0.12
RUV	−0.07	−0.48

[1]Appendix tables are available with the entire article on microfiche. Order from American Geophysical Union, 2000 Florida Avenue, N.W., Washington, D.C. 20009. Document B84-005; $2.50. Payment must accompany order.

TABLE 3. Distribution of Three Error Parameters for Mehetia Earthquakes for Initial and Final Locations

Origin time Standard Error			Horizontal Standard Error			Vertical Standard Error		
Seconds	Initial	Final	Kilometers	Initial	Final	Kilometers	Initial	Final
0.00 - 0.01	0	2	0-1	0	0	0-2	0	0
0.01 - 0.02	0	6	1-2	0	2	2-4	0	7
0.02 - 0.03	3	33	2-3	2	17	4-6	4	28
0.03 - 0.04	9	29	3-4	4	31	6-8	7	38
0.04 - 0.05	11	28	4-5	8	27	8-10	18	32
0.05 - 0.06	15	19	5-6	15	23	10-15	70	22
0.06 - 0.07	34	15	6-7	19	19	15-20	25	8
0.07 - 0.08	35	6	7-8	25	4	20-25	4	1
0.08 - 0.09	21	2	8-9	22	4	35-30	5	1
0.09 - 0.10	6	0	9-10	15	2	30-35	3	1
0.10 - 0.11	2	0	10-15	18	6	35-40	1	1
0.11 - 0.12	4	0	15-20	6	3	40-45	0	0
0.12 - 0.13	0	0	20-25	3	1	45-50	1	0
0.13 - 0.14	0	0	25-30	0	0	50-65	0	1
0.14 - 0.15	0	0	30-35	1	0	65-80	1	0
0.15 - 0.16	0	0	35-40	0	0	80-95	1	0

straining the epicentral distance to this station (this is so because of the very low inclination on the horizontal of any mantle ray). On the other hand, for sources whose true location is above the Moho, the observed *S—P* depends both on epicentral distance and depth. Since we constrain the depth at the Moho, the inclusion of *S* wave data improves the relocation of earthquakes below the Moho, while it degrades epicentral relocations of events whose true depth is above it; but it will still improve the relative location of events whose true depths are comparable. This is apparent on Figure 8: the top frame displays relocations using only *P* data, the bottom one uses both *P* and *S*. Two clear clusters become apparent in the latter. It is probable, however, that the apparent distance between the clusters mostly reflects differences in focal depths. A decrease in true focal depth would tend to increase the apparent distance to Tahiti. The two clusters correspond to events separated temporally, with the northwestern ones occurring during the first weeks of the swarm, while the southeastern cluster became active later.

It is possible to conclude that the relative locations of the epicenters of events of comparable depths are probably accurate to better than 5 km. On the other hand, the relative position of the whole ensemble of epicenters is somewhat less well constrained, probably no better than ±10 km.

A similar procedure was used for the Teahitia swarms of 1982 and 1983; we chose to derive a second set of station corrections (listed in Table 2), independently from the case of Mehetia, since these corrections can also be affected by the seismic ray path to the station, which is necessarily shorter for Teahitia. As in the case of Mehetia, initial locations were used to infer station corrections, which were then used to relocate 281 events (142 in 1982 and 139 in 1983), for which clear first arrivals could be picked at all five (or six, including TTR; fewer in 1983) stations of the Tahiti subarray and in at least one station (usually PMO) of the Rangiroa subarray. Table 4 is a parallel to Table 3 and shows the improvement provided by the relocations for events of the 1982 swarm. The fundamental features from this data set are similar to those at Mehetia, namely, the low values of the residuals and the precision in relative relocations of epicenters on the order of ±5 km (assuming common depths). No *S* wave arrivals were used in the Teahitia relocations, since a few attempts showed no improvements over the solutions obtained exclusively from *P* wave data. The extent of the 1982 swarm is shown on Figure 9, with individual epicenters listed in Table A2; Figure 10 and Table A3 present the data for the 1983 swarm. Relocations of the 1983 events are less precise, the average HCE being about 16 km. There were affected by the generally smaller magnitudes of the events, the shutdown of temporary station TTR in late 1982, and the loss of telemetry from TVO following hurricane Reva in March 1983.

Correlation With Bathymetric Features

Following the 1981 swarm, an echo-sounding campaign was conducted along the southeastern flank of the island of Mehetia by the French Navy patrol ship *La Paimpolaise*. This survey identified an elongated, steep trough at a depth of approximately 1700 m, whose presence was confirmed in April 1983 by H. Craig (personal communication, 1983) aboard R/V *Melville*. The location of this presumed crater (17.95°S and 148.04°W), shown as a circled dot on Figure 8, falls in the zone seismically active during March 9-25. Its major axis, oriented east-west is about 1.5 km long.

In the case of Teahitia, and as seen on Figure 2, adapted from the GEBCO maps [*Monti and Pautot*, 1974], a seamount topping above 2000 m below sea level had been charted prior to the swarm. This seamount appears to have grown on the flank of the volcanic edifice of the main island in a situation very similar to that of Loihi with respect to Hawaii. Echo-soundings taken in April 1983 during a cruise of R/V *Melville* have identified two craters at depths of only 1600 m (H. Craig, personal communication, 1983) A dive of the submersible *Cyana* in December 1983 has confirmed their existence and revealed at least one area

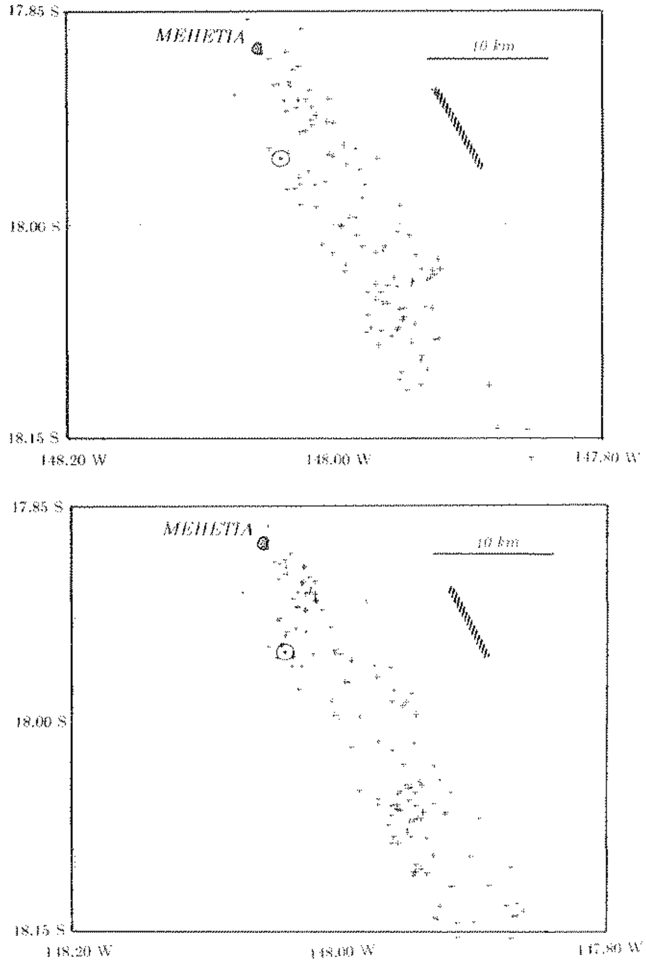


Fig. 8. Effect of the use of S wave arrival times at PPT on the relocations of the 140 major events at Mehetia. (top) Relocations using only P wave times. (bottom) Relocations using S times. Note that the epicenters on the bottom figure tend to regroup into two clusters; Figure 11 further shows the temporal migration of seismicity between the two clusters; on the contrary note that the epicenters farthest to the south are dispersed by the addition of S times, suggesting a shallow character. The striped line provides an estimate of the extent and orientation of the 1-σ "HCE" location error.

of hydrothermal activity (J.-L. Cheminée, personal communication, 1983).

EVOLUTION OF THE SWARMS AND NATURE OF THE EVENTS RECORDED

Seismic activity developing in the form of swarms (rather than of classic foreshock - main shock - aftershock patterns) has been generally observed in areas of active volcanism or extensional tectonics. Swarms directly associated with major eruptions are often (but not always) short-lived (lasting from 1 week to a few months); the seismic activity which preceded the eruption of Mount St. Helens in 1980, of 2 months' duration, would be a typical example of a volcanic swarm [*Endo et al.*, 1981]. On the other hand, swarms known not to be associated with eruptive volcanism have also been widely documented (e.g., the Matsushiro swarm of 1965-1967 [*Hagiwara and Iwata*, 1968]). With some exceptions, they tend to be of longer duration.

Mehetia, 1981

In this case, the systematic analysis of the evolution of the characteristics of the events in the swarm (location, frequency of occurrence, magnitude and spectral content) reveals 4 different periods of activity (See Figure 11).

Period 1: March 6 to March 8. The swarm starts abruptly, with a large number of earthquakes (on the average more than 240 events per day) of relatively low magnitude ($M_L \le 3.3$). This activity is concentrated about 6 km southeast of the island. Seismograms have a high repeatability, with P and S waves featuring high frequencies and simple wave shapes (see Figure 12*a*). As discussed below, we interpret these events as being at least as deep as the Moho, and thus the epicentral locations as probably accurate.

Period 2: March 9 to March 25. The number of events decreases considerably (only 27 per day on the average), but the activity remains substantial, with the two largest events in the swarm occurring on March 15 ($M_L = 4.3$) and March 25 ($M_L = 4.0$), respectively. The March 15 event is not

TABLE 4. Distribution of Three Error Parameters for Teahitia Earthquakes (1982 Swarm) for Initial and Final Locations

Origin time Standard Error			Horizontal Standard Error			Vertical Standard Error		
Seconds	Initial	Final	Kilometers	Initial	Final	Kilometers	Initial	Final
0.00 - 0.01	3	25	0-1	0	0	0-2	1	0
0.01 - 0.02	5	54	1-2	0	5	2-4	0	0
0.02 - 0.03	5	40	2-3	1	23	4-6	0	9
0.03 - 0.04	10	11	3-4	5	23	6-8	0	16
0.04 - 0.05	29	5	4-5	5	37	8-10	3	22
0.05 - 0.06	42	5	5-6	9	26	10-15	9	45
0.06 - 0.07	44	1	6-7	2	10	15-20	20	16
0.07 - 0.08	4	1	7-8	8	8	20-25	24	8
0.08 - 0.09	0	0	8-9	22	4	35-30	22	9
0.09 - 0.10	0	0	9-10	14	0	30-35	16	5
0.10 - 0.11	0	0	10-15	55	4	35-40	6	6
0.11 - 0.12	0	0	15-20	20	1	40-45	4	1
0.12 - 0.13	0	0	20-25	0	0	45-50	3	2
0.13 - 0.14	0	0	25-30	1	0	50-65	9	1
0.14 - 0.15	0	0	30-35	0	1	65-80	13	1
0.15 - 0.16	0	0	35-40	0	0	80-95	4	1

truly representative of this episode of the seismicity, being both shallower and of lower frequency. Apart from this event, seismograms remain homogeneous in their characteristics, while locations move somewhat to the south to the area of the presumed crater.

Period 3: March 26 to May 30. The activity continues to decrease (on the average only 13 events per day) and is characterized by higher-magnitude earthquakes. The signature of the seismograms becomes significantly different: their spectrum evolves toward lower frequencies, and the duration of P_n and S_n increases, leading to occasional ringing, and in certain cases to the occurrence of two distinct arrivals of P_n, separated by about 1.2 s (see Figure 13). The constancy of this figure precludes the development of several sources but rather suggests a multipathing phenomenon. Multipathing of P_n was indeed observed in Rangiroa for seismic refraction arrivals originating in the Mehetia area [*Talandier,* 1982]. Finally, high-frequency surface waves develop after S_n in the seismogram (see record for station PAE on Figure 12*b*). All these characteristics suggest that foci become shallower and are by then located in the crust, which would mean that the epicentral locations could be less accurate. *S* waves are generally of smaller amplitude than *P*. Epicenters remain clustered until April 17 and then disperse somewhat. The cluster is clearly located southeast of the crater; this migration from summit to flank is directly comparable to the pattern observed at Loihi by *Klein* [1982].

Period 4: June 1 to December 31 and into 1982. The level of seismicity decreases with some renewed activity in November and December (on the average 11 events per day). Characteristics of the seismograms vary widely, as do the epicenters, falling in any of the previous three zones as well as outside them. Spectral content of the signals is generally lower frequency, with occasional signals duplicating those of the preceding periods; in such occurrences, epicenters also coincide.

In addition, the temporary station operated on Mehetia at the end of March recorded repeated puffs of seismic noise, whose characteristics could be compared to high-frequency tremors such as those recorded from Teahitia in

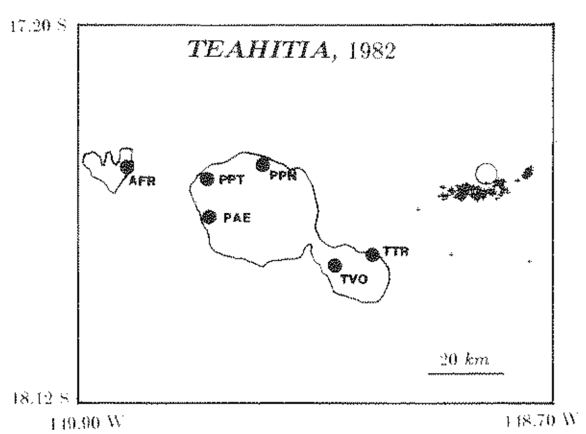


Fig. 9. Spatial extent of the Teahitia swarm of 1982. The islands of Tahiti and Moorea are outlined, with the stations of the Tahiti subarray given by their three-letter code. The light circle sketches the 2000-m isobath, as inferred from the GEBCO map (Figure 2). The 142 events recorded by the whole subarray, and at least one station on Rangiroa, are shown as individual crosses. See Figure 12 for temporal evolution of the epicenters.

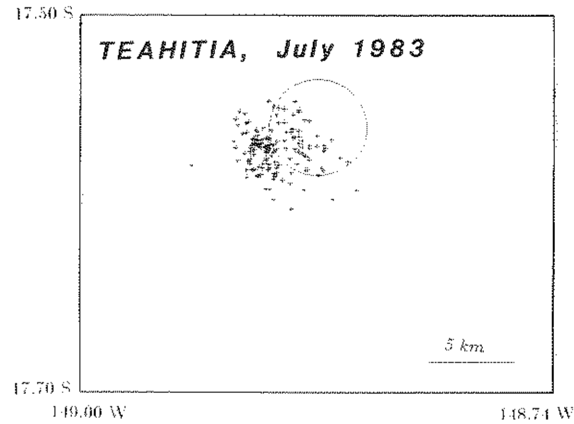


Fig. 10. Location of the Teahitia swarm of 1983. Area depicted is the same as on Figure 14.

1982; they were not recorded by the permanent stations on Tahiti, but our experience of the Teahitia swarm as well as data reported for Kilauea (R. Y. Koyanagi, personal communication, 1981) indicate that volcanic tremors rarely propagate over distances greater than 100 km. Thus it is possible (and the data from the temporary station suggest even likely) that seismic tremors accompanied the Mehetia swarm at least during its third phase. In particular, earthquakes similar to those accompanied by tremors at Teahitia were observed at Mehetia after March 25.

On the other hand, the only *T* waves (propagated in the SOFAR low-velocity acoustic channel in the water column) received from Mehetia were generated by the major earthquakes in the swarm. No additional activity of any nature could be detected on this basis. This situation is fundamentally different, for example, from that of Macdonald volcano 1700 km to the southeast, whose presence and activity have been discovered and monitored exclusively through *T* waves [*Johnson,* 1970; *Talandier and Okal,* 1982].

Teahitia, 1982

Figure 5 shows a history of the evolution of the swarm, both in terms of numbers of events detected, and of seismic energy released. The most intriguing pattern concerning the Teahitia swarm is the apparent migration of the seismicity evident on Figure 14. For about 5 weeks the relocated epicenters move regularly from east to west along the southern flank of the Teahitia Seamount. However, the relocations used in this figure did not adjust hypocentral depth, and the epicentral migration may be an artifact of the change in true depth of the hypocenters. It is easy to show that the shallower the true depth of the source, the closer to the recording network the relocated epicenter, under the assumption of a constant hypocentral depth coinciding with the Moho. Thus the pattern observed on Figure 14 could be a true epicentral migration to the west, an ascent of the hypocenters toward the surface, or a combination of both. In particular, and starting approximately March 29, a pattern of complexity appears in the P_n seismograms recorded at PMO (Rangiroa), featuring two successive arrivals, separated by 1.1 s. Just as in the case of Mehetia, this suggests a shallowing of the source; it is worth noting that this date coincides with the development of strong tremor. Whether or not a vertical component of migration is present, the order

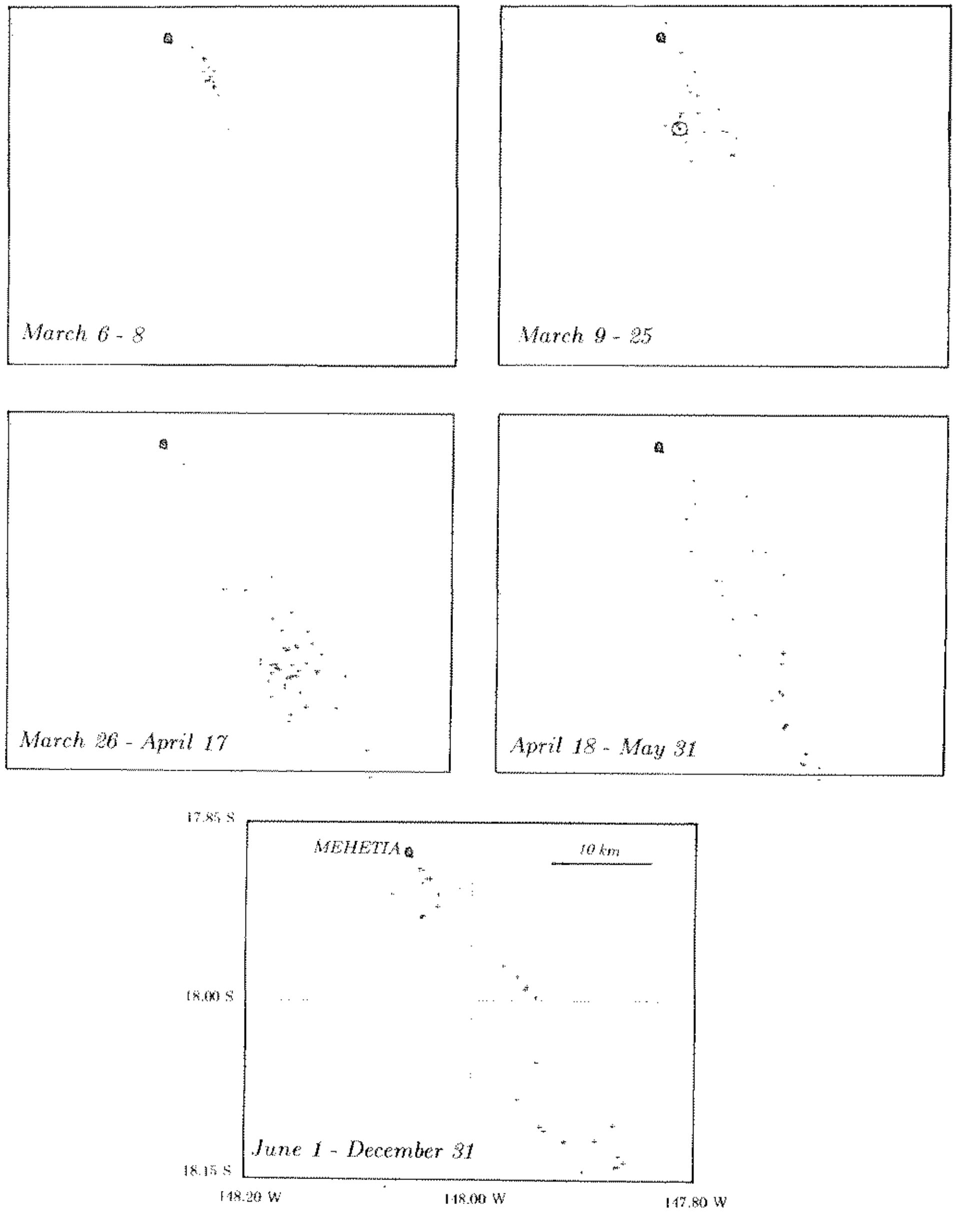


Fig. 11. Migration of seismicity at Mehetia (1981). The 140 events located in Figure 8 are separated in five characteristic time intervals. Note how the swarm begins in the immediate vicinity of Mehetia, then moves to the crater area (shown as circled dot on the second frame), to the southeastern corner during the month of April, and then becomes much more diffuse. Locations plotted as resulting from the use of P and PPT S times.

of magnitude of the rate of migration of seismicity remains 1 km/d, a figure comparable to the progression of seismicity during slow intrusions in moderately active rift zones, such as Loihi or the southwest rift at Kilauea; more active systems such as the eastern rift at Kilauea would be somewhat faster [*Klein,* 1982].

The Teahitia swarm has many differences with Mehetia: first the whole crisis is much shorter-lived, with the activity becoming practically negligible after only 6 weeks; second, the swarm is much more homogeneous in character, with the number of earthquakes growing steadily for about 2 weeks, and then decaying regularly; finally, and most importantly, the two curves shown on Figure 5 are similar to each other, indicating no drastic change in the magnitude distribution of earthquakes, in obvious contrast with phase 2 at Mehetia.

Additionally, a large amount of seismic tremors were recorded from Teahitia during the period March 27 to April 17. Their evolution with time is shown on Figure 6, with typical examples presented on Figure 15. These tremors are basically of two types: high-frequency ones, with seismic energy peaked in the 7 Hz range, of a rather spasmodic nature, following sequences of small but sharp earthquakes, and low-frequency tremors, peaked at 2-3 Hz. It is worth noting that low-frequency tremors occur only during periods of (relative) quiescence of their high-frequency counterparts (e.g., March 29, April 4, April 9, April 16).

The largest event recorded at Teahitia is the $M_L = 3.4$ earthquake of April 1, 1982; this event and three more in early April were reported felt by a few inhabitants on and outside the peninsula. This relatively low level of maximum seismicity is in sharp contrast with the case of the 1981

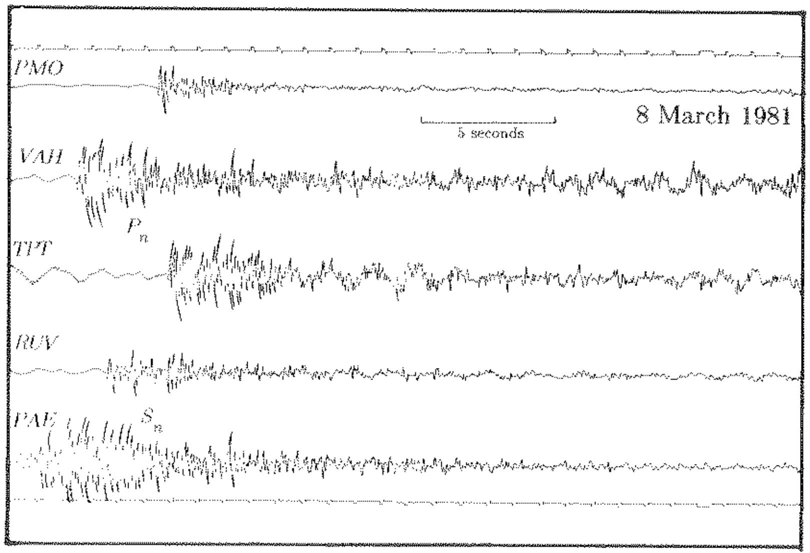


Fig. 12*a*. Short-period records of the M_L = 3.3 event of March 8, 1981 (Mehetia). The four top traces are P_n arrivals from the Rangiroa subarray and the bottom one S_n from PAE on Tahiti. Note high-frequency character of P_n and absence of ringing at Rangiroa or surface waves at PAE (note that low-frequency energy at TPT is noise, since it is present before the P_n arrival). Tick marks are seconds.

Mehetia and of the 1971-1972 Loihi swarm, during which events as large as M_L = 4.3 were recorded. *T* waves were recorded only from the largest Teahitia events and could not be used to identify otherwise unsuspected activity.

Teahitia, 1983

As compared to the 1982 crisis, the 1983 Teahitia swarm had very short-lived seismic activity (only 13 days), lower magnitudes (maximum M_L = 2.4), and the seismic signature of its events was very homogeneous. In particular, the pattern of double P_n arrivals at the Rangiroa stations was present for all events of the 1983 swarm. As explained above, we interpret this as meaning that the whole sequence took place at very shallow depths, probably within a few kilometers of the seafloor. As shown on Figure 7, the activity started abruptly (with about 500 events in each of the first 2 days) and decayed very gradually until July 26. Figure 10 shows that seismic activity concentrated on the western flank of the seamount. A pattern of slight migration from east to west (again probably reflecting shallower sources) may be present but is not significant due to larger uncertainties on the epicenters of these smaller events. High-frequency tremors accompanied the earthquakes during the whole seismic episode; low-frequency tremors developed after the earthquakes ceased, and lasted for several months (see Figure 16). Their cumulative duration is considerably more intense than in 1982, and they are still going on as of April 1984.

A mini-swarm of about 300 earthquakes, preceded by an M_L = 2.6 event took place between December 18 and 21, 1983. As the final version of this paper was being prepared,

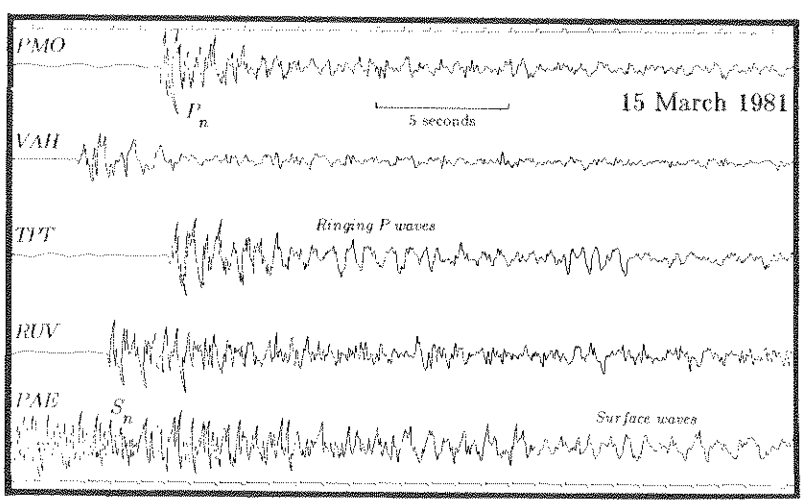


Fig. 12*b*. Same as Figure 12*a* for the M_L = 4.3 event of March 15, 1981, at Mehetia. Note this time the low-frequency components of P_n at Rangiroa stations, the ringing *P* waves lasting for several seconds with a characteristic frequency of 1.3 Hz, and the strong surface wave following S_n at PAE. This surface wave is also present in a later portion of the record at Rangiroa. This event is interpreted as being shallower than the March 8 one shown on Figure 12*a*. Tick marks are seconds.

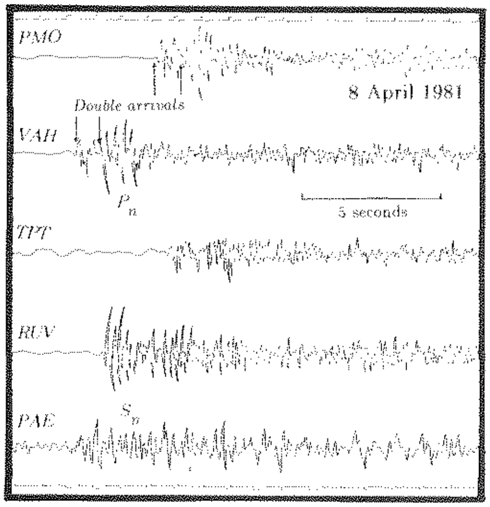


Fig. 13. Same as Figure 12, for the April 8, 1981, event at Mehetia. Note complexity of P_n wave shapes, with at least two distinct arrivals evident at most stations. The time between them remains constant (Δt = 1.2 s) for several events, suggesting a structural origin such as multipathing for this phenomenon. Tick marks are seconds.

a new 30-day swarm of 9000 earthquakes took place starting March 3, 1984. It featured shallow events with a maximum magnitude M_L = 3.7, several of which were felt on Tahiti. They were accompanied by intense tremor, mostly high frequency, and located a few kilometers east of the area active in 1983.

Frequency-Magnitude Relations

A number of frequency-magnitude (b value) investigations were carried out using all earthquakes recorded between March 6 and October 1, 1981, at Mehetia; between March 16 and May 31, 1982, at Teahitia; and between July 12 and 26, 1983, at Teahitia. This technique models the number N of earthquakes with magnitude M with a relationship of the form

$$\log N = a - b\,M \tag{3}$$

[*Gutenberg and Richter*, 1944]. Values of b significantly larger than the worldwide average (b = 0.9) are taken to involve rock undergoing thermal weakening or excessive fracturing [*Mogi*, 1963]. In particular, documented volcanic seismicity has been associated with b values varying from 1.4 to more than 3 [*McNutt*, 1983]. *Talandier and Kuster* [1976] reported b values of 1.0 outside swarms, and ranging from 1.5 to 3.2 during swarms at Moua Pihaa. In the case of Mehetia, our results, which use magnitude windows of 0.2 units, indicate an average b value of 1.13 ± 0.08 for the whole sequence; however, if we restrict the data set to the first 2 days of the swarm, the b value increases to 1.41 ± 0.11. This change of b value illustrates the dramatic increase of the number of larger earthquakes occurring after March 8. These results are summarized in Figure 17. In the case of the 1982 swarm at Teahitia, an investigation for the whole period March 16 to May 31 yields a well-constrained b value of 1.46 ±0.12 (Figure 18). An attempt to use shorter sampling periods failed to unveil significant variations of this coefficient with time. The figure b = 1.46 is in excellent agreement with the value found for the first two days of the 1981 Mehetia swarm. A remarkably similar figure of b = 1.50 ±0.16 was obtained for the 1983 swarm.

While the range of b values obtained at Teahitia and during the first phase of activity at Mehetia falls short of those reported for events of *Minikami's* [1974] B type, during such major eruptions as Pavlof, 1975 (b = 1.9–2.6), for the summit earthquakes at Loihi (b = 2.13), or even for the Moua Pihaa swarms (b = 3.2), they are comparable to results obtained for A-type events at Fuego, Guatemala (b = 1.3), and for the flank events at Loihi [*Talandier and Kuster*, 1976; *Klein*, 1982; *McNutt*, 1983]. In particular, the b values reached during the three Polynesian swarms are significantly higher than found at sites of recurrent tectonic intraplate seismicity such as Regions A and C in the southcentral Pacific [*Okal et al.*, 1980], whose seismic activity was presumably not associated with active volcanism.

While high b values have generally been recognized as indicative of volcanic seismicity, *Okada et al.* [1981] have guarded against the use of b values for short time samplings in the study of the evolution of volcanic swarms: their data at Mount Usu show the late development of large earthquakes and the disappearance of smaller ones, which would lead to negative b. Thus it may not be warranted to assign a sudden decrease in b values to a variation in the physical properties of the rocks involved. However, a general trend toward fewer but larger earthquakes is clearly present at Mehetia after March 8, similar in character to the phenomena reported by *Klein* [1982] at Loihi and Kilauea, and which he describes as an evolution of the activity from rift to flank. A similar pattern was also found at Mount St. Helens and Bezimianny (S. R. McNutt, personal communication, 1983).

In summary, the volcanic nature of the seismicity observed during the 1981 and 1982 swarms in the Tahiti-Mehetia area can be asserted on the basis of (1) the short-lived swarmlike character of the activity, (2) the large b values, (3) the later identification of craters at both sites and of hydrothermal activity at Teahitia, and (4) the evolution of the swarms reminiscent of those at Hawaii and Loihi and the abundant tremors recorded from Teahitia (and suspected at Mehetia on the basis of records from the temporary station).

DISCUSSION

Since the two sites are located under the ocean, our purpose in this section should be to conduct a kind of detective story, using the variation in the characteristics of the seismicity to infer the magmatic processes which may have accompanied the swarms. Unfortunately, two problems hamper our potential insight into these processes. First, our lack of hypocentral depth resolution obscures one of the crucial parameters in the evolution of the swarms, and second, the characteristics and evolution of volcanic swarms are known to vary substantially from one volcano to another. We will nevertheless attempt to draw a parallel between our observations and other examples of volcanic seismicity.

Earthquakes from documented volcanic activity have long been studied. *Minakami* [1974] has reviewed their properties at Japanese volcanoes and established a classification in four types: A-type earthquakes, characterized by a

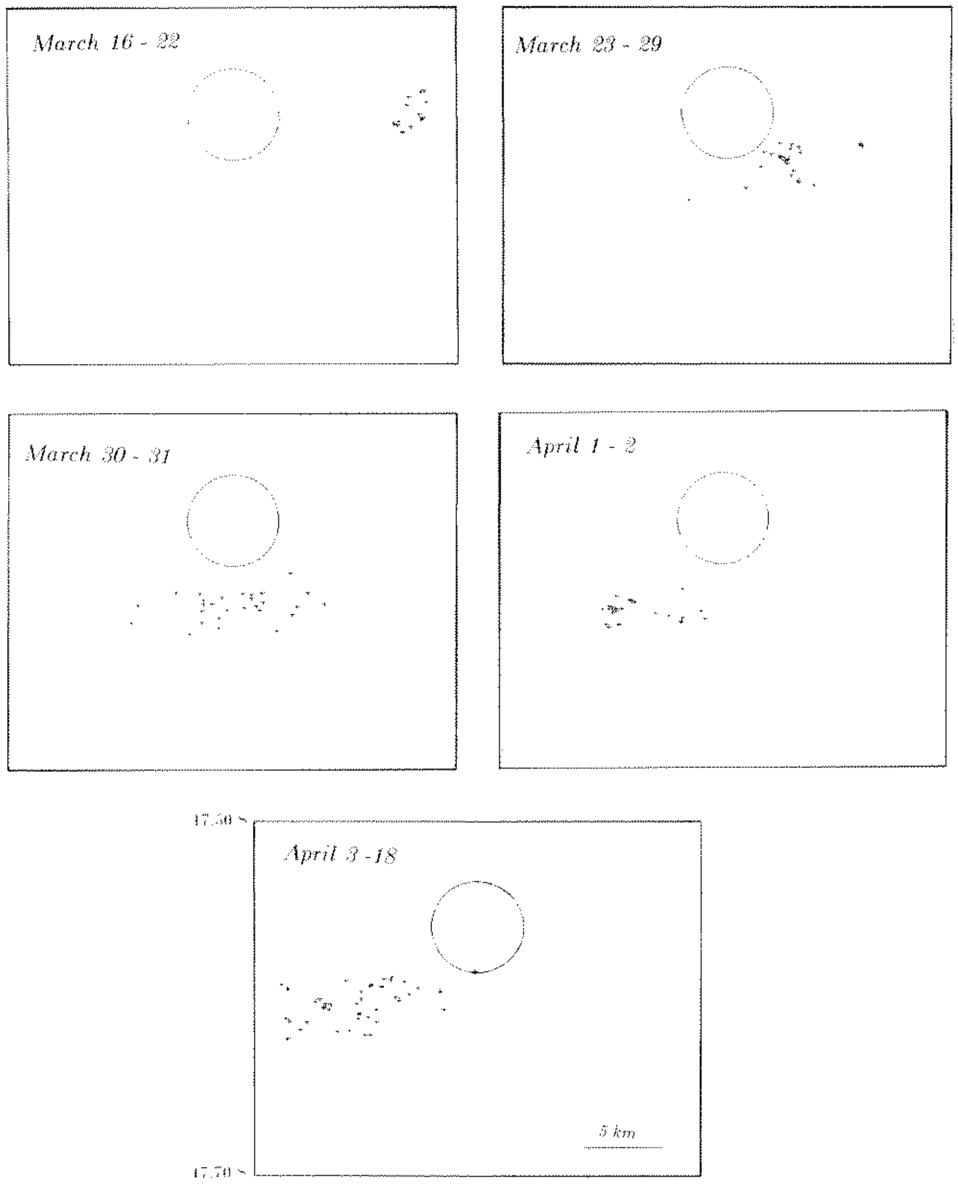


Fig. 14. Migration of seismicity at Teahitia (1982). The 142 relocated earthquakes are separated in five time intervals. Note the general westward motion of the seismicity. See text for a discussion of the possible influence of a change in depth on the extent of migration. The circle sketches the approximate 2000-m isobath, as inferred from the GEBCO map (Figure 2).

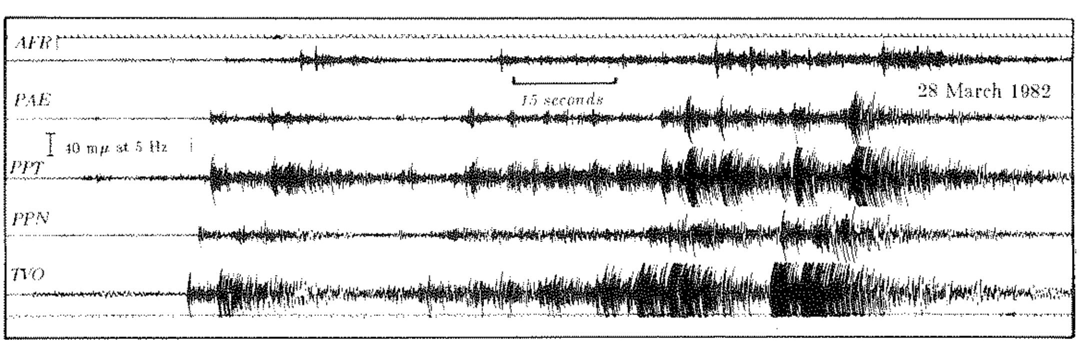


Fig. 15*a*. Typical sequence of repeated small earthquakes, intermixed with high-frequency tremor, as recorded at the Tahiti subarray on March 28, 1982, from Teahitia. Tick marks are seconds.

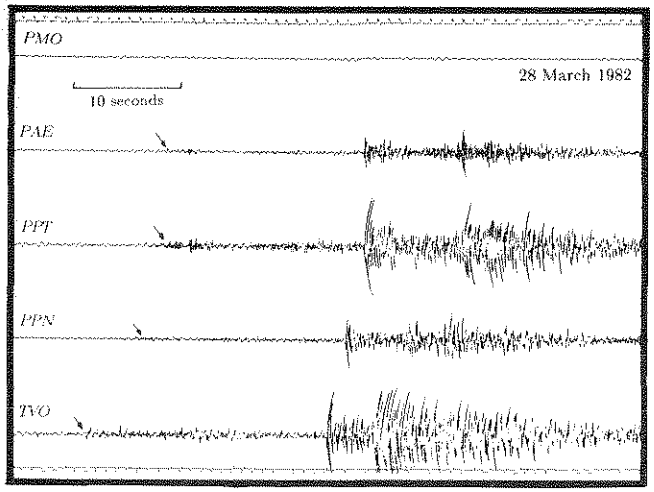


Fig. 15*b*. Detail of the beginning of the sequence shown on Figure 15*a*. Note that the phenomenon starts by an apparent increase in the amplitude and frequency of the ambient noise (arrows), followed by more substantial events. A very similar behavior is reported at Kilauea by *Aki and Koyanagi* [1981]. Tick marks are seconds.

high-frequency spectrum and a seismic signature comparable to that of tectonic earthquakes; *B*-type events, characterized by a lower-frequency spectral content and precursory to the outbreak of eruptive activity; explosive events corresponding to aerial eruption of the volcano; and seismic tremor. (Explosive events, recognized mainly through their strong airwaves, would not be expected from deep underwater locations, where the hydrostatic pressure prevents the release of gases from the magma.) The distinction between *A*- and *B*-type events has been widely used (e.g., *Endo et al.* [1981] at Mount St. Helens), with *A*-type earthquakes being understood as a preliminary, somewhat deeper, step and *B*-type ones being more directly concurrent with eruptions or intrusions. In their early study of the Tahiti-Mehetia area, *Talandier and Kuster* [1976] have distinguished between so-called α and β events, which grossly parallel *A* and *B* types.

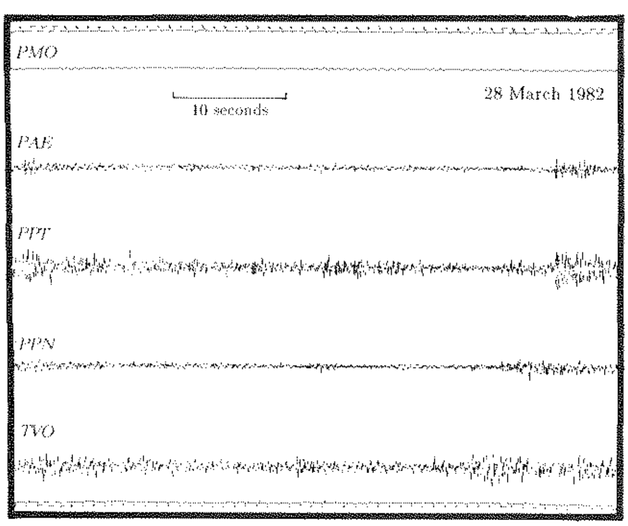


Fig. 15*c*. Typical high-frequency tremor, following record shown on Figure 15*a* by about 15 mn. Note that the seismic amplitude does not fall to either the frequency or the amplitude of the background noise, but rather that seismic energy peaked around 7 Hz is sustained for long periods of time. Rangiroa station PMO has been substituted for AFR on the top trace to highlight the absence of recorded tremor at large distances (350 km). Note on the other hand, that the decrease in amplitude with distance from TVO (45 km) to PPT (90 km) is not substantial, suggesting a deep tremor source. Tick marks are seconds.

On the other hand, Minakami's classification is far from being universal: for example, eruptive cycles at Pavlof Volcano in the Aleutians, totally lack *A*-type seismicity [*McNutt*, 1983], while those of Fuego, Guatemala, lack standard *B* events [*Yuan et al.*, 1984]; also, *Latter* [1981] reports difficulty in applying the classification to Ruapehu and Ngauruhoe in New Zealand. He prefers to introduce the notion of volcanic and tectonic earthquakes, the word "tectonic" simply describing an event not believed to involve rock directly altered or compressed by a magmatic process, but still taking place in the context of the disturbance of the island by the volcanic swarm. Similarly, *Havskov et al.* [1983] identified four different types of events from the 1982 eruption of El Chichon, only one of which is identical with Minakami's. Finally, Minakami himself identified an additional class of events ("*C* type") in a detailed study of earthquakes at Mount Usu [*Minakami et al.*, 1951]. In the present case, and with no definitive control on hypocentral depth, it is rather difficult to assign the seismicity observed at Mehetia and Teahitia to either the *A* or *B* types of *Minakami*'s [1974] classification. Generally speaking, the evolution which we observe from a simple, high-frequency P_n waveshape to a lower-frequency, more complex, seismogram, with concurrent development of tremor at Teahitia, is directly comparable to evolution from *A* type to *B* type. However, patterns such as the decrease in magnitude, and the disappearance of *S* waves in *B*-type events would clearly not apply in the present case.

As discussed above, one of the characteristics of volcanic seismicity most crucial to placing the seismic swarm in the context of magmatic processes, namely hypocentral depth, could not be constrained by our relocations. The only travel time data which can be used to gain an estimate of depth are *S* wave arrival times: their use tends to improve relocations for earthquakes truly located at or below the Moho, while it degrades them if the source belongs to the crust. On this basis, we think that we can interpret the

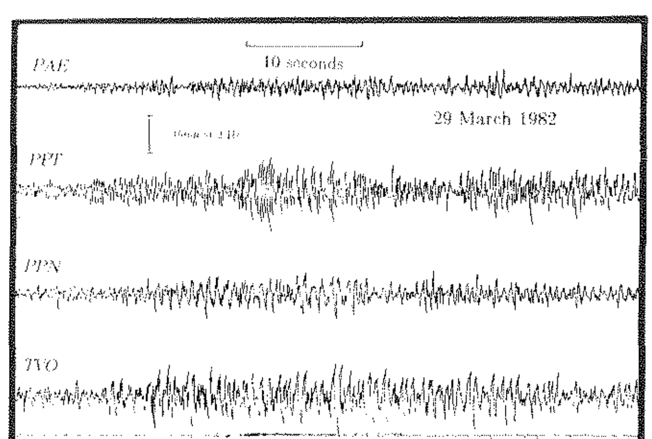


Fig. 15*d*. Example of low-frequency tremor, peaked around 2 Hz, as recorded in the Tahiti subarray on March 29, 1982. Note more pronounced decay of the amplitude at 2 Hz between TVO and PPT, suggesting shallower source than in the case of high-frequency tremor. Tick marks are seconds.

third period in the activity of Mehetia (March 26 to May 31, 1981) and the large March 15 event as being shallower than the earlier events, and the whole 1982 Teahitia swarm as taking place within the crust. Similarly, the ringing of *P* waves found at Mehetia after March 25, 1981 (and also, exceptionally, on March 15), is characteristic of extremely shallow events since it must involve efficient coupling with the water column over depths which cannot greatly exceed a typical P_g wavelength (say 3 km; see *Ward* [1979] for an investigation of the same physical problem at longer periods). The development of substantial surface waves in the 1-Hz range (Figure 12*b*), while totally controlled by details of the structural layering of the volcanic edifices, would, in a flat-layered situation, constrain the focal depth to about $\lambda/4$, or approximately 1-2 km below sea level [*Harkrider and Okal,* 1982]. The entire 1983 swarm at Teahitia is certainly no more than a few kilometers deep.

Volcanic tremor, consisting of more or less continuous seismic agitation accompanying certain phases of volcanic activity, has been reported and studied extensively. Its character can be spasmodic, featuring a repeated number of individual, identifiable seismic events, or harmonic, in which case the frequency content of the signal is predominantly monochromatic. Deep seismic tremors of spasmodic nature, and originating about 50 km below Mauna Loa, were identified by *Eaton and Murata* [1962] as representative of a filling of magma conduits preceding the 1959 eruption by 3 months. Harmonic tremors located at 30-35 km depth were found by *Aki and Koyanagi* [1981] to be caused by magma oscillation in longitudinal cracks, and to represent a continuous, ongoing aspect of the volcano's activity, not directly related to any given eruption. These authors

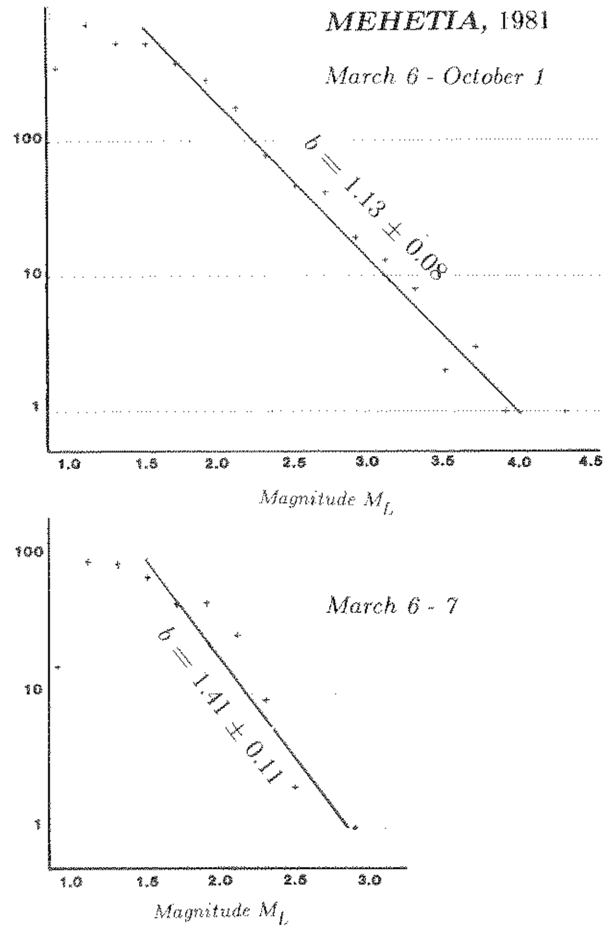


Fig. 17. Frequency-magnitude plots for the 1981 Mehetia swarm. (top) Full swarm. (bottom) Study limited to the first 2 days of the swarm. Windows of 0.2 units of magnitude used in both cases.

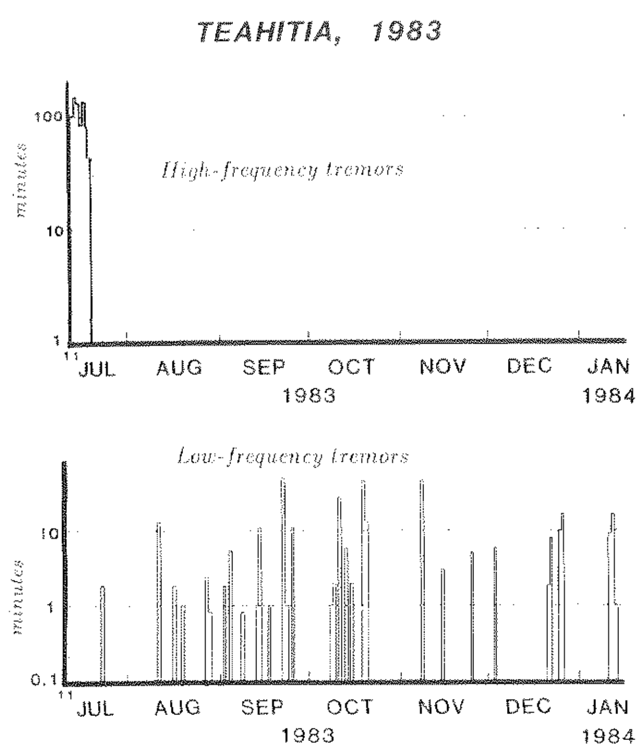
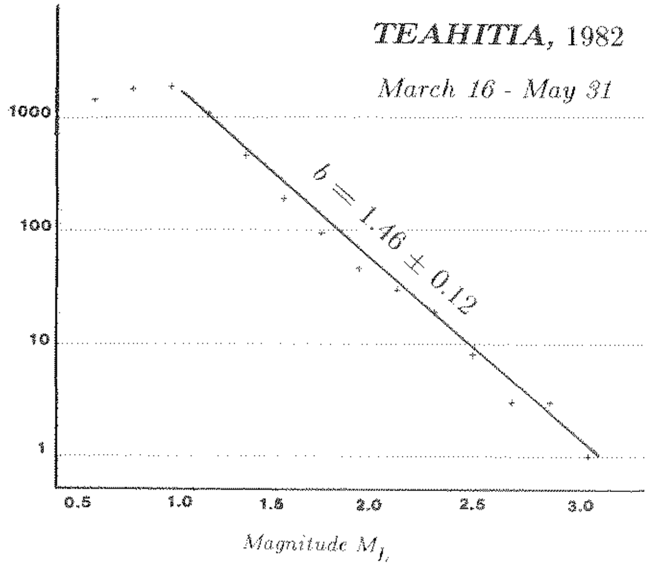


Fig. 16. History of the high-frequency and low-frequency tremors at Teahitia, 1983. This figure is similar to Figure 6 for the 1982 swarm. Note the anti-correlation between the two types and the preponderence of low-frequency tremors, which lasted continuously into the March 1984 swarm.

also noticed an evolution of the dominant frequency of the tremor with time (from about 7 to 3 Hz) during an episode of tremor (typically a few hours), which they interpreted as cracks joining with each other and thus increasing the characteristic length of the oscillator.

The sequence of repeated small earthquakes followed by tremor observed in 1982 at Teahitia and presented on Figure 15*a* is comparable to observations during Pavlof's 1981 eruption (S. R. McNutt, personal communication, 1983) (although Pavlof's tremor was lower frequency); it can be thought to represent a succession of small crack openings, followed by movement and oscillation of the magma in the cracks. This is comparable to *Latter*'s [1981] "intrusion" earthquakes or to *Minakami et al.*'s [1951] "*C*-type" events. As seen on Figure 15*c*, the amplitude of these tremors does not significantly decay with station distance (in the range 40-90 km); this would suggest a deep origin; however, the tremors are concurrent with earthquakes believed to be located in the crust. As seen on Figure 15*b*, the initial waveforms, which start by what could be taken as an increase in the amplitude and frequency of background noise, are reminiscent of harmonic tremors observed at Kilauea by *Aki and Koyanagi* [1981]. The amplitude of the ground motion (about 0.01μm) is also comparable. On the other hand, the pattern of decrease of the wave frequency during an episode of tremor is absent from the Teahitia tremors.



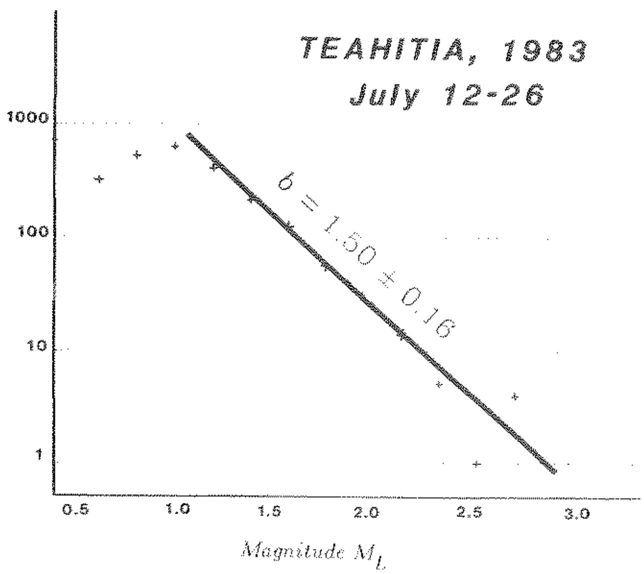


Fig. 18. Frequency-magnitude plots for the (top) 1982 and (bottom) 1983 swarms at Teahitia. Windows of 0.2 units of magnitude are used.

The total duration of the episodes of high-frequency tremor following small earthquakes, as compiled from Figure 6, for the period March 17 to April 17 is 5500 minutes. If we interpret them as representative of crack opening and magma transport in *Aki and Koyanagi*'s [1981] formalism, we come up with a "reduced displacement" of 130 m^2, only 4 times less than the cumulative value over 18 years at the Hawaii volcanoes. Since no swarms of a magnitude similar to the 1981-1983 sequences have taken place since the establishment of the Polynesian seismic array, the cumulative reduced displacements could thus be of the same order of magnitude at the two volcanic chains, over an interval of 18 years. A direct comparison of the two volcanic edifices is, however, difficult, since the characteristics of the swarms at Teahitia and Hawaii are different and the tremors are not directly comparable. In particular, the Kilauea tremors were not connected with a particular eruption. Also, *Aki and Koyanagi* [1981] and *Chouet* [1981] have argued that the quantification of seismic tremors grossly underestimates the amount of lava ejected at Mauna Loa and Kilauea and suggest that part of the transport of the magma through the lithosphere escapes seismic detection. Since the Hawaii volcanoes are the only ones for which this kind of quantification has been performed, it is not clear that exactly identical situations would exist at other locations. In particular, and as we argue below, the volcanic system in the Tahiti-Mehetia area is probably in a much earlier stage of its development than is Kilauea, where the plumbing is well established and the shield-building steady. In this respect, the opening of cracks under Teahitia could involve a higher density of resistive barriers [*Aki,* 1979], leading to the puffs of earthquakes starting the high-frequency tremor sequences.

While high-frequency tremors (peaked at 8 Hz) have also been reported concurrently with the ejection during the 1943 eruption of the Mexican volcano El Parícutin [*Flores,* 1945], fountaining phases of volcanic activity are usually accompanied by intense lower-frequency harmonic tremors, only a few kilometers deep and peaked at 2−3 Hz. This has been widely reported, in particular at Kilauea [*Eaton and Murata,* 1962], Aso [*Kubotera,* 1974], and other sites [*McNutt,* 1984]. At the same time, seismicity and spasmodic tremor activity have been reported to strongly decrease (e.g., at Poás, 1981 [*Güendel,* 1981]). This suggests that the high-frequency tremors are indicative of magma making its way up the plumbing of the volcano, while low-frequency ones accompany the venting and fountaining process. Fountaining probably also acts in a pressure-cooker mode, allowing sudden release, at least in the shallowest parts of the plumbing, of the stresses which otherwise lead to the opening of the cracks generating the high-frequency tremors. It is remarkable on Figures 6 and 10 that a similar pattern of inverse correlation of the two kinds of tremors is present at Teahitia during the months of March and April 1982 and in 1983. Puffs of activity similar to low-frequency tremors were also occasionally observed, although their origin could not be positively localized, as early as January and as late as August 1982.

As seen on Figure 15*d*, the decay of the amplitude of low-frequency tremors with station distance is stronger, suggesting that they originate at shallow depths. We will speculate that they may be associated with eruptive processes, although a direct comparison with the case of Kilauea (possibly again unwarranted) would suggest their total attenuation into the background noise after 90 km, even taking into account the greater sensitivity of the Polynesian stations. Their total duration during the 1982 swarm is very limited (47 mn). On the other hand, in 1983, they lasted for more than 300 mn, about half as long as the high-frequency tremors.

On the basis of the above discussion, we propose the following scenario as an interpretation of the swarms:

Mehetia, 1981

The swarm starts on March 6 with a series of small earthquakes, located deep under the island, presumably below the Moho. By March 9, and in a pattern reminiscent of the Loihi swarm, fewer but larger earthquakes take place; the epicenters move closer to the underwater crater. Additionally, a unique large shallow earthquake takes place on March 15. It may represent tectonic release under the pressure of the deeper magmatic intrusion. After March 25 the poor fit of S wave travel times, as well as the presence of surface waves and ringing P_n waves suggests transition to shallow depths, certainly above the Moho; this ascent of the seismic activity is probably associated with the upward progression of a bubble of magma. Low-frequency events may represent episodes of B-type seismicity, although the evolution of b values is toward lower figures; it is suggested that tremor was taking place at the end of March; submarine eruptions may have occurred: necessarily taking place below the SOFAR channel, they would not have generated T waves because of the impossibility of degassing at 1700 m depth. At least part of the later seismicity may due to the release of intraplate tectonic stress accumulated in the plate, as indicated by the quiescence of the area prior to the swarm. These events are more spread out spatially, and would correspond to *Latter*'s [1981] "volcanotectonic" or "tectonic" classes. Their presence may act to lower the average b value of the whole swarm.

Teahitia, 1982

The swarm begins abruptly on March 16, 1982, by a number of small, simple events, possibly as deep as the Moho; these may represent country rock fracturing under the increased magma pressure; it is likely that the east-west migration of the seismicity also includes a component of shallowing, as suggested by the later development of ringing phases. By March 27 a pattern starts of numerous cracks opening inside the seamount, followed by magma filling giving rise to high-frequency tremor. Occasionally, an eruption takes place, accompanied by low-frequency tremor; the pressure release temporarily shuts off the crack opening process, and thus the high-frequency tremors disappear. Again, no T waves are generated by the eruption, which is too deep. During this phase, three major earthquakes are felt on Tahiti. On April 17, after lasting 3 weeks, the tremors stop, signaling the end of the eruptive process, and since previous seismic activity has drained the intraplate tectonic stress, the seismic swarm dies off quickly during the month of May.

Teahitia, 1983

The volcano awakes again through a series of small, shallow earthquakes concentrated on its western flank. This activity may take the form of a lateral intrusion from the 1982 plumbing, and thus the swarm gets directly into the final seismic stage of shallow events accompanied by high-frequency tremor. This pattern lasts only 2 weeks, but eruptions probably more intense than in 1982 continue for 6 months. In December 1983 a submarine exploration finds evidence of ongoing hydrothermal activity. The interpretation of the small burst of activity in December is unclear, but it may be premonitory to the large outburst of activity in March 1984.

CONCLUSION: VOLCANIC POTENTIAL OF THE TAHITI MEHETIA AREA

We propose that the seismic swarms at Mehetia and Teahitia are representative of magmatic phenomena which have culminated in volcanic eruptions at Teahitia in 1982 and 1983 (and probably also at Mehetia in 1981). This volcanic activity and the much weaker swarms at Moua Pihaa and Rocard Seamounts [*Talandier and Kuster,* 1976] prove that the Society Island hot spot is alive and active east of Tahiti.

Petrological studies of Tahiti have repeatedly failed to identify the tholeiites typical of hawaiian volcanism [*McBirney and Aoki,* 1968]. Tahitian rocks have been found to be generally more alkalic. Studies of Mehetia basalts by *Mottay* [1976], M. O. Garcia (personal communication, 1981) and *Grall and Okal* [1984] have also documented their strongly alkalic character; the steep shape of this small island indicates a viscosity higher than at typical hawaiian volcanoes. On the basis of this evidence, and by comparison with alkalic rocks from posterosional stages of volcanism, such as the nepheline-normative Honolulu series, *Brousse and Mottay* [1979] have argued that Mehetia is an example of late-stage volcanism and, since the island is very small, that the hot spot itself has faded considerably since the time it built Tahiti and is therefore dying or dead.

Recent evidence from Loihi Seamount may however considerably alter this line of thought: *Moore et al.* [1982] have reported both tholeiites and alkali basalts at Loihi, a seamount evidently in the growing stage, and fed by a very healthy hot spot. On the basis of both trace elements and isotopic signatures, *Frey and Clague* [1983] have thus proposed that tholeiites are emplaced only during the steady shield-building episode of the life of a hawaiian island, with more alkalic rocks present both at the early and late stages of its activity. The alkali basalts at Mehetia could then suggest an early stage of island building. This view is also upheld by recent data showing much higher δ^3He values at Mehetia than on Tahiti [*Craig and Rison,* 1983], in direct parallel with values from both Loihi and Maui higher than at Kilauea [*Kurz et al.,* 1983], and by trace element analyses of basalts from Mehetia (island and crater area) and Teahitia which have failed to evidence the trends characteristic of post-erosional series [*Grall and Okal,* 1984]; finally, using a model of diffusive porosity for the building of a volcanic edifice, *Lacey et al.* [1981] have argued that its early stages must be characterized by a steeper slope, all other parameters including viscosity remaining equal.

Thus we interpret the recent activity at Mehetia and Teahitia as episodes in the ongoing process of the building of the next major volcanic edifice in the Society Islands chain. Despite the fact that only Mehetia has succeeded in rising above sea level, three other sites (Moua Pihaa, Rocard, and especially Teahitia) are active, and spread over an area grossly 80 km by 60 km. Figure 1 shows that the spatial distribution of volcanism in the Windward group of the Society Islands is arranged along two lines grossly parallel to the absolute motion of the Pacific plate: the northern one passes through Mehetia, Rocard, Teahitia, and Tetiaroa (this small atoll located 50 km north of Tahiti may represent a stillborn member of a previous stage of activity of the hot spot, later sunk by loading from the neighboring and much larger Tahiti). The second line, about 60 km to the south, links Moua Pihaa, the Tairapu Peninsula (Tahiti-Iti), Tahiti itself, and Moorea. This situation is strikingly similar to the layout of the presently active and recent volcanoes in Hawaii, along two lines 30 km apart and running from Haleakala to Kilauea and from Kahoolawe to Loihi, as reported by *Jackson et al.* [1972]. Although an explanation has yet to be found for this fascinating pattern, it could be a common property of Hawaiian type island chains.

Finally, and in the absence of systematic archives, it is extremely difficult to compile the historical seismicity of the Tahiti-Mehetia area, and a fortiori, to estimate the possible recurrence rate of its volcanic activity. There exist several accounts of earthquakes felt on the island of Tahiti, as well as Polynesian legends mentioning large fires on the island of Mehetia. However, as reported by *Talandier and Okal* [1979], an isolated account of a felt earthquake could be due to a seismic shock as distant as Tonga, felt in Tahiti through its T waves. On the other hand, a unique report of a swarm of earthquakes felt on Tahiti is given by *Lespinasse* [1919], who writes:

> Numerous earthquakes of variable intensity [were felt] starting November 21st [1918], and up to the end of the year. Some days, seismic tremors were felt every hour.

Such a swarm would share the basic characteristics of the 1982 Teahitia activity, being only much more intense and prolonged (only three earthquakes were felt over a period of a few days in 1982, none in 1983). This suggests that the Tahiti-Mehetia area has the potential for large-scale volcanic activity. A period of quiescence of between 20 and 63 years, suggested by this report and the lifetime of the seismic network, falls within the broad range of observed recurrence of eruption at moderately active Hawaiian volcanoes. Needless to say, the Tahiti-Mehetia area ought to become the target of a broad multidisciplinary exploration, involving geophysical, petrological, and geochemical studies and experiments to help solve the many problems which seismology has unearthed.

The volcanoseismic activity of the Tahiti-Mehetia hot spot would have gone undetected, were it not for the seismic instrumentation of the nearby islands. None of the 25,000 events reported here were detected at teleseismic distances. In the absence of systematic monitoring at lower magnitudes the only events reported felt could have been mistaken for isolated earthquakes of tectonic origin. This raises the question of the true level of underwater volcanic activity in remote ocean basins. In situations where the active seamount has grown to shallow depths (e.g., Macdonald Seamount) and penetrates the SOFAR channel, adequate detection is possible through T waves even from large distances [*Talandier and Okal*, 1982].

However, if the seamount is small and eruptions occur at large depths, degassing or water vaporization will be absent and no acoustic signal will penetrate the SOFAR; thus volcanic episodes will be detected only if they feature seismicity above the worldwide detection threshold of $m_b = 4.7$. Fast-spreading ridges such as the East Pacific Rise are an example of underwater volcanism which must be going on but is not routinely detected.

The above discussion suggests that there must exist unsuspected active volcanoes on the floor of the world's oceans. Their number can only be speculated.

Acknowledgments. We are grateful to Bob Decker, Fred Klein, and Bob Koyanagi at Hawaii Volcano Observatory for their hospitality and many discussions on volcanic seismicity. Extensive discussions with Steve McNutt and Kei Aki are also gratefully acknowledged as well as preprint exchanges with many scientists. We thank the French Navy, officers and crew of *La Paimpolaise* for bathymetric work near Mehetia, Harmon Craig for preliminary results from the cruise of R/V *Melville*, Mike Garcia for preliminary results on the petrology of Mehetia, and Jean-Louis Cheminée for discussion of the results of his *Cyana* dive. Fred Klein provided a detailed and careful review of this paper. Tony Watts organized the Lamont-Doherty Seamount Symposium, whose participants contributed greatly to the development and refinement of many ideas expressed in this paper. This research was supported by Commissariat à l'Energie Atomique (France) and by the Office of Naval Research, under contract N00014-79-C-0292.

REFERENCES

Aki, K., Characterization of barriers on an earthquake fault, *J. Geophys. Res.*, *84*, 6140-6148, 1979.

Aki, K. and R. Y. Koyanagi, Deep volcanic tremor and magma ascent mechanism under Kilauea, Hawaii, *J. Geophys. Res.*, *86*, 7095-7109, 1981.

Becker, M., R. Brousse, G. Guille and H. Bellon, Phases d'érosion-comblement de la vallée de la Papenoo et volcanisme sub-récent à Tahiti, en relation avec l'évolution des îles de la Société (Pacifique Sud), *Mar. Geol.*, *16*, M71-M77, 1974.

Brousse, R. and G. Mottay, Basalts more alkaline as younger in the Society linear chain (Pacific Ocean): A declining hot spot?, Paper presented at Hawaii Symposium on Intra-Plate Volcanism, Hilo, July 16-22, 1979.

Chouet, B., Ground motion in the near field of a fluid-driven crack and its interpretation in the study of shallow volcanic tremor, *J. Geophys. Res.*, *86*, 5985-6016, 1981.

Clague, D. A., J. G. Moore and W. R. Normark, Loihi Seamount: Early alkalic eruptive phase (abstract), *Eos Trans. AGU*, *62*, 1082, 1981.

Craig, H., and W. Rison, Helium isotopes and mantle heterogeneity (abstract), *Eos Trans. AGU*, *64*, 348, 1983.

Duncan, R. A., and I. MacDougall, Linear volcanism in French Polynesia, *J. Volcanol. Geotherm. Res.*, *1*, 197-227, 1976.

Eaton, J. P., and K. J. Murata, How volcanoes grow, *Science*, *132*, 925-938, 1962.

Endo, E. T., S. D. Malone, L. L. Noson, and C. S. Weaver, Location, magnitudes and statistics of the March 10-May 18 earthquake sequence, *U.S. Geol. Surv. Prof. Pap. 1250*, 93-107, 1981.

Flores C., L., Investigación geofisico-sismométrica del fenómeno volcánico, in: *El Paricutin*, pp. 23-45, Universidad Nacional Autonoma, México, 1945.

Frey, F. A., and D. A. Clague, Geochemistry of diverse basalt types from Loihi Seamount, Hawaii: Petrogenetic implications, *Earth Planet. Sci. Lett.*, *66*, 337-355, 1983.

Grall, H. M., and E. A. Okal, Petrological arguments in favor of an active Society Islands (French Polynesia) hotspot (abstract), *Eos Trans. AGU*, *65*, 300, 1984.

Güendel, F., Estudio preliminar de las señales sismicas del volcan Poás, I., *Bol. Vulcanol.*, *11*, 5-9, 1981.

Gutenberg, B., and C. F. Richter, Frequency of earthquakes in California, *Bull. Seismol. Soc. Am.*, *34*, 185-188, 1944.

Gutenberg, B. and C.F. Richter, *Seismicity of the Earth*,, pp. 229-230, Princeton University Press, Princeton, N. J., 1954.

Hagiwara, T., and T. Iwata, Summary of the seismograph observation of Matsushiro swarm earthquakes, *Bull. Earthquake Res. Tokyo Univ.*, *46*, 485-515, 1968.

Harkrider, D. G. and E. A. Okal, Propagation and generation of high-frequency sediment-controlled Rayleigh modes following shallow earthquakes in the southcentral Pacific (abstract), *Eos Trans. AGU*, *63*, 1025, 1982.

Havskov, J., S. de la Cruz-Reyna, S. K. Singh, F. Medina and C. Gutiérrez, Seismic activity related to the March-April, 1982 eruptions of El Chichon Volcano, Chiapas, Mexico, *Geophys. Res. Lett.*, *10*, 293-296, 1983.

Henderson, L. J. and R. G. Gordon, Oceanic plateaus and the motion of the Pacific plate with respect to hot spots (abstract), *Eos Trans. AGU*, *62*, 1028, 1981.

Jackson, E. D., E. A. Silver and G. B. Dalrymple, Hawaiian-Emperor chain and its relation to Cenozoic circum-Pacific tectonics, *Geol. Soc. Am. Bull.*, *83*, 601-617, 1972.

Johnson, R. H., Active submarine volcanism in the Austral Islands, *Science*, *167*, 977-979, 1970.

Klein, F. W., Hypocenter location program HYPOINVERSE, *U.S. Geol. Surv. Open File Rep.*, *78-694*, 42 pp., 1978.

Klein, F. W., Earthquakes at Loihi submarine volcano, and the Hawaiian hot spot, *J. Geophys. Res.*, *87*, 7719-7726, 1982.

Koyanagi, R. Y., Earthquakes from common sources beneath Kilauea and Mauna Loa volcanoes in Hawaii from 1962 to 1965, *U.S. Geol. Surv. Prof. Pap. 600-C*, C120-C125, 1968.

Kubotera, A., Volcanic tremors at Aso Volcano, in: *Physical Volcanology*, edited by L. Civetta, P. Gasparini, G. Luongo, and A. Rapolla, pp. 29-47, Elsevier, Amsterdam, 1974.

Kurz, M. D., W. J. Jenkins, S. R. Hart, and D. Clague, Helium isotopic variations in volcanic rocks from Loihi Seamount and the island of Hawaii, *Earth Planet. Sci. Lett.*, *66*, 388-406, 1983.

Lacey, A., J. R. Ockendon, and D.L. Turcotte, On the geometrical form of volcanoes, *Earth Planet. Sci. Lett.*, *54*, 139-143, 1981.

Latter, J. H., Volcanic earthquakes and their relationship to eruptions at Ruapehu and Ngauruhoe volcanoes, *J. Volcanol. Geotherm. Res.*, *9*, 293-309, 1981.

Lespinasse, A., Observations météorologiques faites à l'Observatoire de l'hôpital de Papeete, *Bull. Soc. Etud. Océaniennes Polynésie Fr.*, *5*, 62-63, 1919.

Mammerickx, J., S.M. Smith, I.L. Taylor, and T. E. Chase, Bathymetry of the South Pacific (map), *IMR Tech. Rep. Ser. TR 56*, Scripps Instit. Oceanog., Univ. of Calif., San Diego, 1975.

McBirney, A. R. and K. Aoki: Petrology of the island of Tahiti, in: *Studies in Volcanology, A Memoir in Honor of Howell Williams*, edited by R. R. Coats, R. L. Hay and C. A. Anderson, *Mem. Geol. Soc. Amer.*, *116*, 523-556, 1968.

McNutt, S. R., A review of volcano seismicity (abstract), *Eos Trans. AGU*, *64*, 265, 1983.

McNutt, S. R., Analysis of seismic tremor from Pavlof, Fuego, Pacaya, San Cristobal, and Masaya Volcanoes, *Bol. Vulcanol.*, in press, 1984.

Minakami, T., Seismology of Volcanoes in Japan, in: *Physical Volcanology*, edited by L. Civetta, P. Gasparini, G. Luongo, and A. Rapolla, pp. 1-27, Elsevier, Amsterdam, 1974.

Minakami, T., T. Ishikawa, and K. Yagi, The 1944 eruption of Volcano Usu in Hokkaido, Japan, *Bull. Volcanol.*, *11*, 45-157, 1951.

Mogi, K., Some discussions on aftershocks, foreshocks and earthquake swarms — the fracture of a semi-infinite body caused by inner stress origin and its relation to the earthquake phenomena, 3, *Bull. Earthquake Res. Inst. Tokyo Univ.*, *41*, 615-658, 1963.

Monti, S., and G. Pautot, Carte bathymétrique du Pacifique Sud (Feuille Tahiti, GEBCO 327; Feuille Hao, GEBCO 358), Cent. Océanol. de Bretagne, Brest, 1974.

Moore, J. G., D. A. Clague and W. R. Normark, Diverse basalt types from Loihi Seamount, Hawaii, *Geology*, *10*, 88-92, 1982.

Mottay, G., Contribution à l'étude géologique de la Polynésie Française - Archipel des Australes - Mehetia (Société), Thèse de 3ième cycle, 217 pp., Univ. Paris-Sud, Orsay, 1976.

Okada, H., H. Watanabe, H. Yamashita, and I. Yokoyama, Seismological significance of the 1977-1978 eruptions and the magma intrusion process of Usu volcano, Hokkaido, *J. Volcanol. Geotherm. Res.*, *9*, 311-334, 1981.

Okal, E. A., J. Talandier, K. A. Sverdrup, and T. H. Jordan, Seismicity and tectonic stress in the Southcentral Pacific, *J. Geophys. Res.*, *85*, 6479-6495, 1980.

Talandier, J., Sismique réfraction dans les régions de Tahiti, Mehetia et Rangiroa, Internal Report, Commiss. à l'Energ. At., Papeete, 1982.

Talandier, J., and G. T. Kuster, Seismicity and submarine volcanic activity in French Polynesia, *J. Geophys. Res.*, *81*, 936-948, 1976.

Talandier, J., and E. A. Okal, Human perception of *T* waves: The June 22, 1977 Tonga earthquake felt on Tahiti, *Bull. Seismol. Soc. Am.*, *69*, 1475-1486, 1979.

Talandier, J., and E. A. Okal, Crises sismiques au Volcan Macdonald (Océan Pacifique Sud), *C.R. Hebd. Séances Acad. Sci. Paris, Sér. II*, *295*, 195-200, 1982.

Ward, S. N., Ringing *P* waves and submarine faulting, *J. Geophys. Res.*, *84*, 3057-3062, 1979.

Yuan, A. T. E., S. R. McNutt and D. H. Harlow, Seismicity and eruptive activity at Fuego Volcano, Guatemala: February 1975 - January 1977, *J. Volcanol. Geotherm. Res.*, in press, 1984.

J. Talandier, Laboratoire de Géophysique, B.P. 640, Papeete, Tahiti, French Polynesia.

E. A. Okal, Department of Geological Sciences, Northwestern University, Evanston, Illinois 60201.

(Received September 1, 1983;
revised March 5, 1984;
accepted March 9, 1984.)