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Seismic Detection of Underwater Volcanism: The Example of French Polynesia

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Abstract—We present a review of the principal methods used for the seismic detection and identification of active underwater volcanism, based on our experience in French Polynesia. In particular, we describe the 5-year activity in the Tahiti-Mehetia area, during which more than 32000 earthquakes were detected by the Polynesian network. We discuss the use of the following three types of seismic waves: conventional (mostly body waves), seismic tremor, and T waves propagated in the low-velocity acoustic channel of the ocean. For each of these waves, we discuss the principal characteristics of the signals, their spectral content, the type of information they provide on the activity of the volcano, and the various limitations faced by their use in detection or monitoring of underwater volcanic edifices. We present a review of the principal swarms monitored by the Polynesian network, and discuss their characterization as either volcanic or tectonic.

Key words: Volcanic seismicity, tremor, T waves, Pacific Ocean basin.

1. *Introduction*

Active volcanism has long been known on the ocean floor: apart from providing the mechanism for sea-floor spreading and the formation of new oceanic crust, it is involved in the subduction process near oceanic trenches, and also at mid-plate locations where, given sufficient activity, it can result in the formation of substantial seamounts or even island chains. Quantitative estimates of this latter form of volcanism have recently been revised upwards following such investigations as BATIZA's (1982) and the statistical work of JORDAN et al. (1983). However these estimates have been based entirely on studies of the product of volcanic episodes, namely the morphology of seamounts, as opposed to the direct observation of ongoing volcanic eruptions, as is the case for subaerial volcanoes. As a result, very little is known regarding the present level of volcanic activity in vast areas of the oceanic basins, such as most of the Pacific Ocean.

Since the oceanic column provides an optical, thermal and to a large extent

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chemical, shield for the remote sensing of the planet's surface, careful monitoring of seismic activity on the ocean floor remains one of our very few methods for studying submarine volcanic activity. This line of research goes back more than thirty years, when, following the suggestion of EWING *et al.* (1946), DIETZ and SHEEHY (1954) were able to obtain a detailed history of the 1952 eruption of Myojin Volcano, south of Japan, using teleseismic T waves propagated in the oceanic column over a distance of 8600 km to a receiving array on the California coastline.

However, the situation is made difficult by a number of problems generally connected with the remote character of the underwater environment. Since a permanent deep-basin ocean-bottom seismometer has yet to be developed, all recording must be done using island stations; instrumented islands are few, and their seismic noise characteristics generally poor, resulting in limited coverage and mediocre detection and location capabilities. Furthermore, logistics and cost for seagoing expeditions prohibit the rapid deployment of portable arrays over the epicentral area of a recognized shock, a procedure routinely carried out on land. Finally, our general ignorance of the bathymetry of ocean basins further inhibits geological interpretation.

In this framework, the existence of the Polynesian Seismic Network (Réseau Sismique Polynésien; hereafter RSP) constitutes a notable exception to the generally unfavorable conditions described above. The RSP is a wide-aperture seismic network,

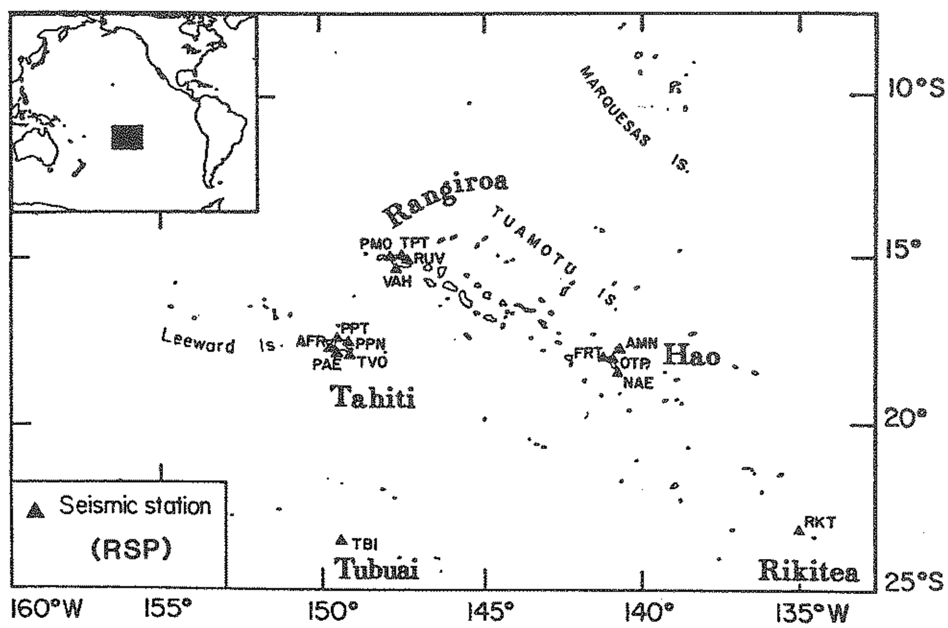


Figure 1
Map of the Réseau Sismique Polynésien (RSP) network in French Polynesia. Inset gives location inside Pacific Ocean.

centered in Tahiti, and specially instrumented to provide detection characteristics comparable to those of the best continental sites: routine magnifications for the seismic displacement are 125000 at 1 Hz, and reach 2×10^6 at 3 Hz. The network has been described in a number of previous papers (TALANDIER and KUSTER, 1976; OKAL *et al.*, 1980) to which the reader is referred. Its location, spanning the Polynesian islands chains (see Figure 1), is obviously favorable to the detection of oceanic underwater volcanism.

The purpose of the present paper is to review the seismological methods used to identify volcanic sources on the ocean floor, and to provide examples drawn from our experience with the RSP in Polynesia. This includes the close monitoring of a major volcanoseismic swarm, lasting 4 years, in the immediate vicinity of the Society Islands, whose first three years have been described in detail by TALANDIER and OKAL (1984a), as well as the monitoring of more distant edifices, such as Macdonald Volcano (TALANDIER and OKAL, 1982; 1984b). Figure 2 is a map of the sites of activity discussed in the paper.

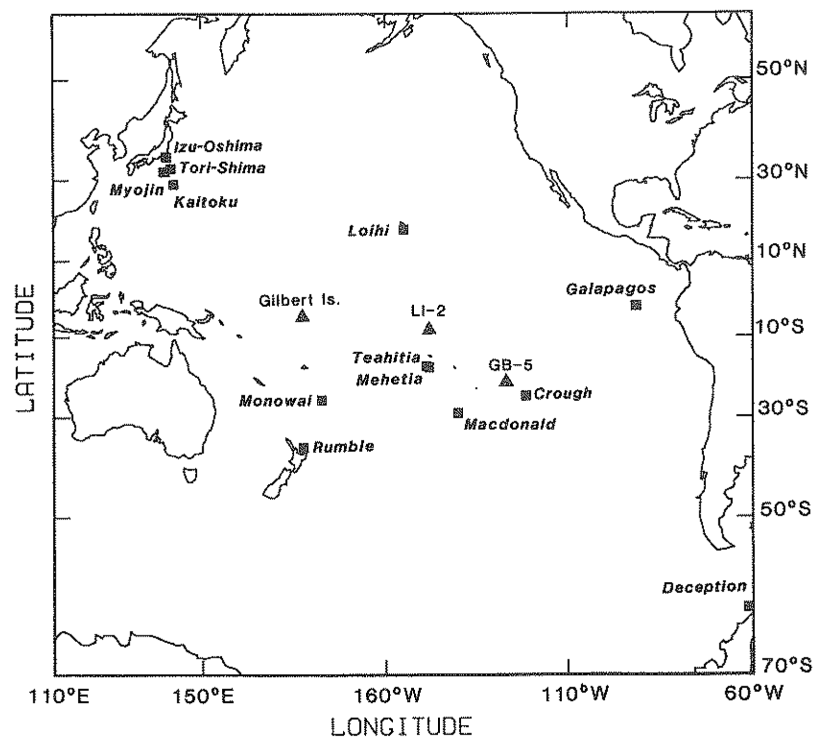


Figure 2
Map of the Pacific Ocean, showing location of principal foci of activity discussed in this paper. Sites shown as squares, and labeled in *italics* are of confirmed or suspected volcanic nature; sites shown as triangles, and labeled in roman type are of confirmed or suspected tectonic nature. Please refer to text for details.

One of the major problems in studying volcanic seismicity is that of discrimination, *i.e.*, the recognition and identification of seismic activity as being of a volcanic origin. In this endeavor, a helpful tool is comparison with data collected at confirmed volcanic sites (in principle subaearial) during documented eruptions, drawing on the considerable progress made in recent years in our understanding of volcanic seismology, thanks to dense instrumentation of continental or island volcanoes. When dealing with Polynesian volcanoes, obvious candidates for comparison are the Hawaiian edifices, whose origin, morphology, and to some extent petrology and geochemistry, are comparable, and which have been extensively monitored and studied for the past few decades (*e.g.*, EATON and MURATA, 1960; DUFFIELD *et al.*, 1982). It must be kept in mind, however, that variations in the mechanism and seismic signature of magma intrusion and ejection are to be expected among mid-plate volcanoes, just as they are widely documented for andesitic subduction systems, even on a short geographic scale (MINAKAMI *et al.*, 1969; MCNUTT and HARLOW, 1983; YUAN *et al.*, 1984).

2. *Seismic Waves Used*

Seismic detection of submarine volcanic activity makes use of three different kinds of seismic waves: (i) *conventional* seismic waves (mostly body waves) which can be individually separated in records as seismic phases, and thus lead to the identification and location of individual seismic events (earthquakes); (ii) *seismic tremor*, consisting of signals of a ringing character which cannot be decomposed into individual events; and (iii) *T waves* propagated in the oceanic SOFAR channel over distances which can be teleseismic in range. It goes without saying that not all types of waves are generated by every phase of every submarine volcanic eruption; in addition, their differing propagation characteristics further affect their eventual detection at an island station. Figures 3, 4, 5 present typical examples of each kind, as recorded at the RSP.

2.1. *Conventional Seismic Waves*

2.1.1. *Example and Background*

We base our discussion of detection by conventional seismic waves on our experience with the Teahitia-Mehetia seismic swarm of 1981–1985. Starting in March, 1981, intense seismic activity developed in the area located approximately 100 km East of Tahiti (see Figure 6) and resulted in the detection of more than 32000 earthquakes, and the precise relocation of approximately 800. As shown in Figure 7, the activity can be decomposed into a number of swarms. The 1981 crisis was located on the flanks of Mehetia Island, while all activity since 1982 moved to the vicinity of a

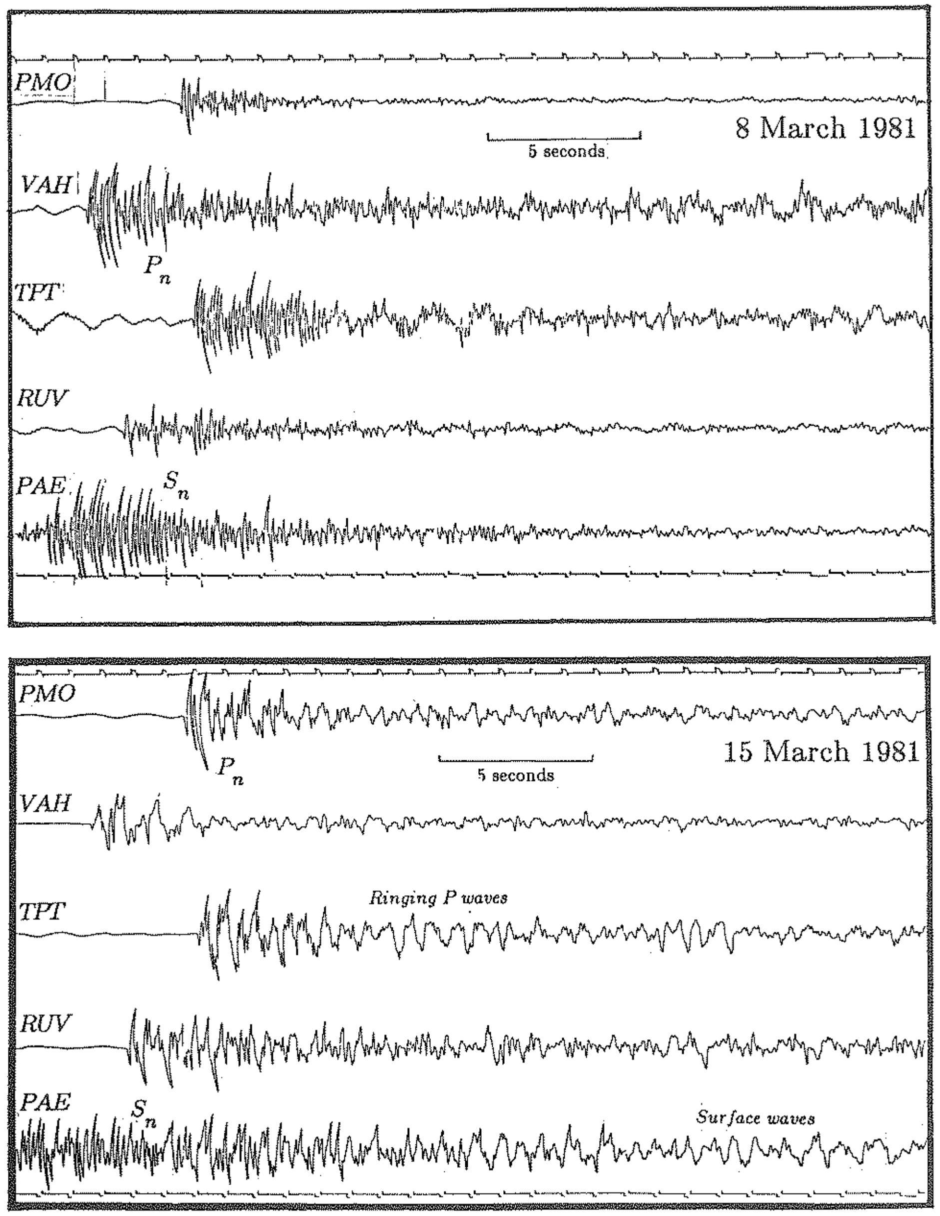
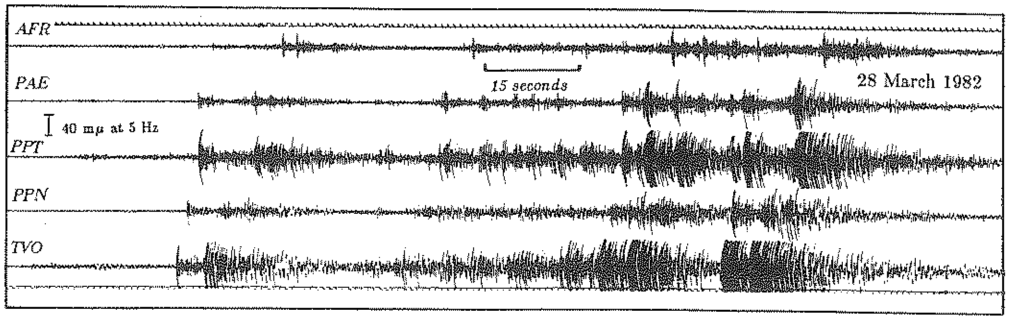


Figure 3
Example of body waves received at stations of the RSP network from the Mehetia 1981 swarm. In each diagram, the top four traces are from the Rangiroa subarray, 300 km to the north, and the bottom from station PAE on Tahiti. Note occasional arrival of surface waves, suggesting shallower source. After TALANDIER and OKAL (1984a).



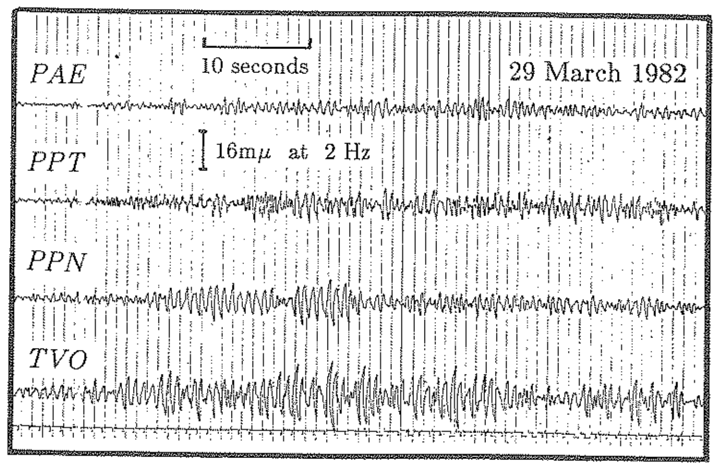


Figure 4
Top: Examples of repeated small earthquakes, intermixed with high-frequency tremor, as recorded in Tahiti from the Teahitia 1982 swarm. *Bottom:* Examples of low-frequency tremor, recorded in a later part of the 1982 swarm. Tick marks are seconds. After TALANDIER and OKAL (1984a).

major seamount structure called Teahitia, 100 km to the west, and only 60 km from Tahiti. Activity has died off to a negligible level (about one event per day) since February, 1985. We refer to TALANDIER and OKAL (1984a) for a complete description of the tectonic framework of the area, and of the history of the first half of the swarm, during the years 1981–1983. Table 1 updates the principal characteristics of the various swarms.

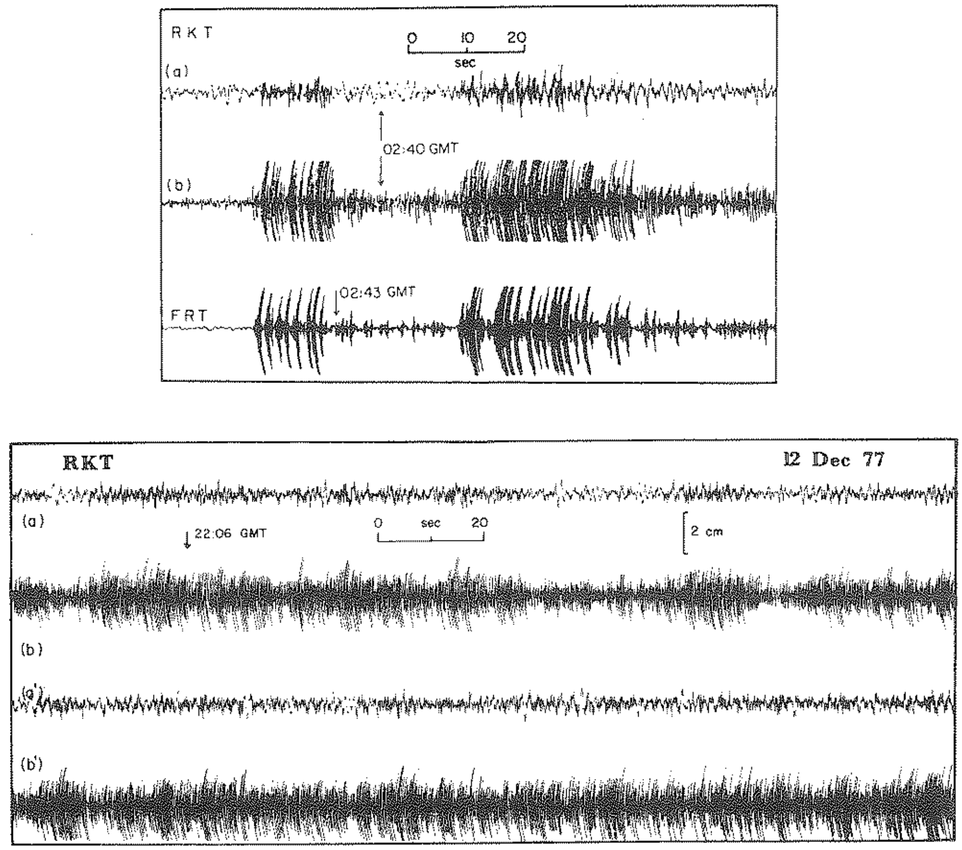


Figure 5
Examples of T waves recorded at RSP stations from Macdonald Volcano. *Top:* Explosives events at the onset of a 1977 swarm, recorded at Rikitea, Gambier (RKT, traces (a), peaked at 1 Hz, and (b), peaked at 5 Hz), and Faratahi, Hao (FRT; peaked at 5 Hz). *Bottom:* Examples of T waves recorded later in the same swarm at RKT (trace a peaked at 1 Hz; trace b peaked at 5 Hz). Note continuous character of the wave. Traces (a′) and (b′) are continuation of (a) and (b) to the right. After OKAL *et al.* (1980) and TALANDIER and OKAL (1982).

2.1.2. Nature of Data

Conventional seismic waves used for volcanic detection at regional distances (tens to hundreds of km) can include direct P and S, propagating in the crust, and/or P_n and S_n refracted at the Mohorovičić (Moho) discontinuity, depending on epicentral distance. In addition, high-frequency Rayleigh waves in the 1–2 s period range were occasionally recognized during the Tahiti-Mehetia seismic swarm.

Table 1

Principal characteristics of the Mehetia-Teahitia swarms of 1981–1985

Location	Year	Duration (days) (†)	Number of earthquakes: Total	Number of earthquakes: $M_L \geq$ 4.0	Maximum M_L	High frequency tremors (minutes) (*)	b-Value (**)	Deep sequence	Tectonic episode
Mehetia	1981	280	3536	1	4.3		1.13 (1.63)	YES	YES
Teahitia	1982	80	8047	0	3.4	5517 (29)	1.44	YES	NO
Teahitia	1983 (July)	15	2487	0	2.8	727 (7)	1.50	NO	NO
Teahitia	1983 (Dec.)	7	328	0	2.6			NO	NO
Teahitia	1984	45 (307)	7795	0	3.7	2051 (20)	1.67	NO	NO
Teahitia	1985	16 (80)	9670	3	4.4	3990 (12)	1.07 (1.43)	NO	YES

(†) Number shows duration of main period of activity; number in parentheses total duration of swarm.
(*) First number is cumulative duration of high-frequency tremors in minutes; number in parentheses number of days with high-frequency tremors.
(**) Number is average b-value for whole swarm (including 'tectonic' episode if present); number in parentheses b-value for early phase of swarm if tectonic episode is present.

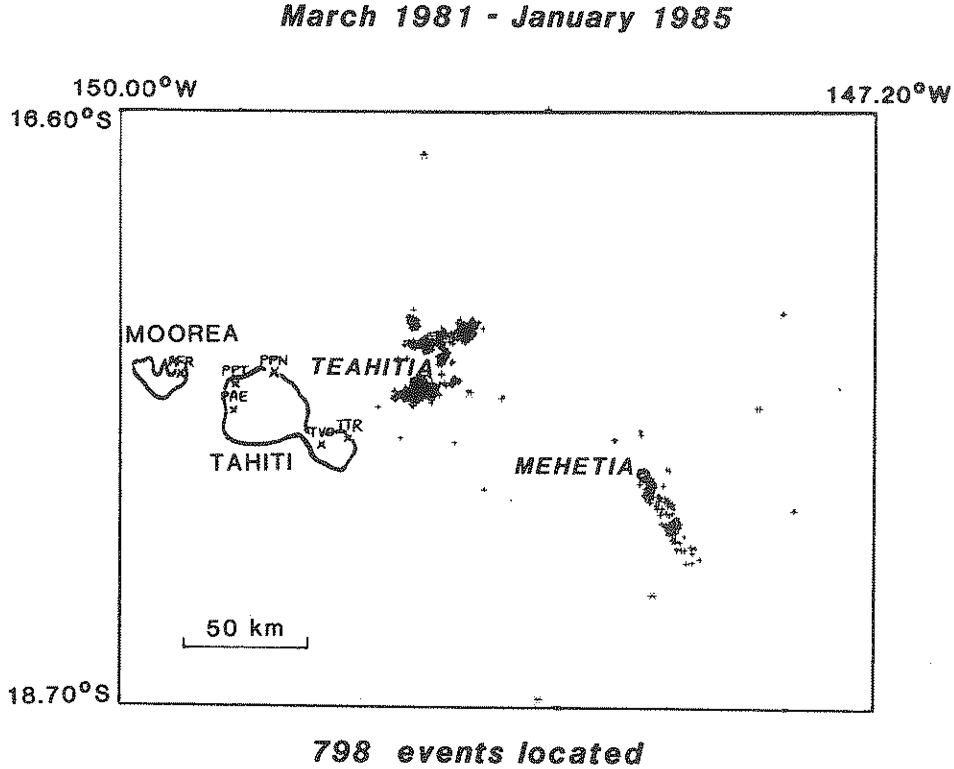


Figure 6
Map of the Tahiti-Mehetia area showing location of 798 events located during the period of activity March 1981–January 1985. Stations of the RSP on Tahiti and Moorea Islands are shown.

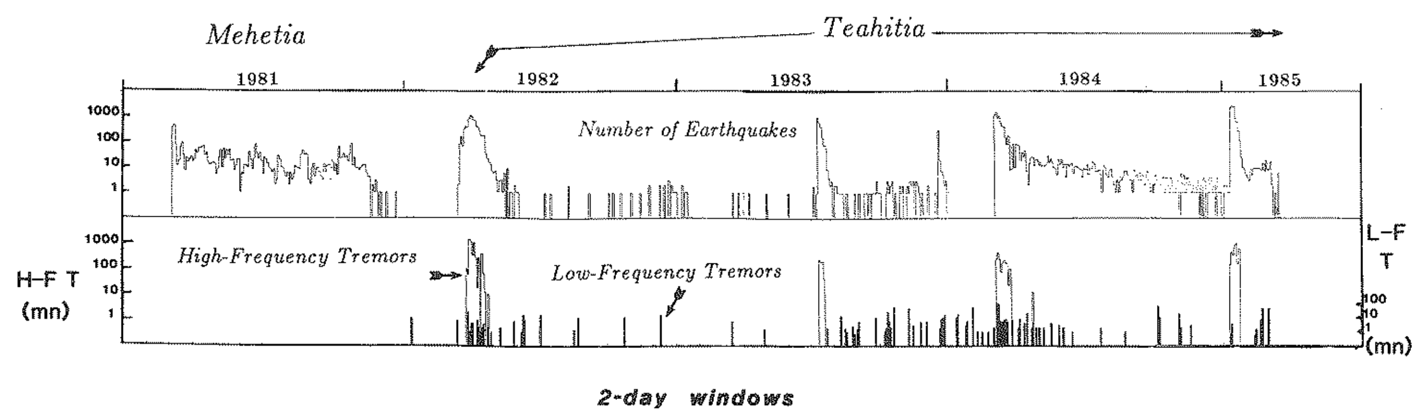


Figure 7
History of the volcanoseismic activity of the Tahiti-Mehetia area, 1981–1985. The top frame shows the number of earthquakes detected by the RSP network, in intervals of 2 days. The bottom frame shows the total duration (in minutes) of high-frequency tremors (open bars) and low-frequency tremors (solid bars) for the same windows. Note that different scales are used to enhance clarity.

2.1.3. *Location Techniques*

The availability of P and S arrival times allows the use of classic location procedures. In our case, we made use of an adapted version of KLEIN's (1978) HYPOINVERSE program, with appropriate models of seismic structures and station corrections, obtained from seismic refraction experiments (TALANDIER and OKAL, 1987). As compared with location procedures for subaerial volcanoes, the distribution of available stations resulted in two major problems, which we believe to be typical of underwater detection. First, the absence of stations in the immediate vicinity of the epicenter made it impossible to achieve depth control from a given set of travel-times alone. In addition, and in the case of the 1981 activity at Mehetia, the azimuthal coverage was concentrated along two directions separated by approximately 95°, and epicentral resolution was poor along their bisector. In the case of Teahitia, its shorter distance to the island of Tahiti resulted in a more adequate azimuthal coverage and in general to more precise epicentral locations, as later verified during on-site surveys.

Since the evolution of hypocentral depth during a volcanoseismic swarm is of major interest for the understanding of the geophysical process, we tried to circumvent the former difficulty in a number of ways. First, we compared the quality of relocations (as measured by the size of one-σ confidence ellipses) for datasets using, as opposed to not using, S times, under the common constraint of a hypocentral depth arbitrarily fixed at the Moho. Under these conditions, it can be shown (TALANDIER and OKAL, 1984a) that the inclusion of an S_n time, in addition to the P_n time at the same station, improves the relocation of an earthquake whose true depth is below the Moho, and deteriorates that of an event truly located above the Moho. While this method gives only minimum information on hypocentral depths, it turned out to be helpful in providing some qualitative insight into the vertical migration of the seismicity.

A somewhat more quantitative measure of an event's depth can come from the occasional observation of surface waves following the *coda* of the S_n wavetrain on vertical short-period seismograms. We interpret those as 1–2 s Rayleigh waves; order-of-magnitude calculations for a point-source double-couple in the layered medium defined by the seismic refraction experiments indicate that their excitation would become negligible for a depth below sea floor comparable to one-half of their wavelength, in the present case $\simeq 5$ km. Thus earthquakes for which a crustal depth is suggested from S times and which do not feature Rayleigh waves are probably confined to the lower crust; those who do show the surface waves are probably within a few km of the top of the edifice.

2.1.4. Migration Patterns

Because conventional waves allow recognition of individual earthquake sources, and as in the case ot subaerial volcanoes, migration patterns of the volcanic seismicity can occasionally be defined. In the case of the Mehetia swarm of 1981, we noticed after a few days a concentration of shallow seismicity approximately 8 km southeast of the island. An echosounding survey by the French Navy revealed the presence of a caldera approximately 1500 m in length; SEABEAM mapping by the *R/V Jean Charcot* in 1986 confirmed these findings, and identified a cratered pinnacle at the eastern extremity of the caldera (OKAL *et al.*, 1987). We tentatively interpret the corresponding sequence in the 1981 swarm as an eruption at this site.

As shown on Figure 8, in the case of the Teahitia swarm, and despite possible artifacts due to poor control on depth, we were able to map the relatively position of the various foci of activity around the seamount structure. During the 1982 swarm, a

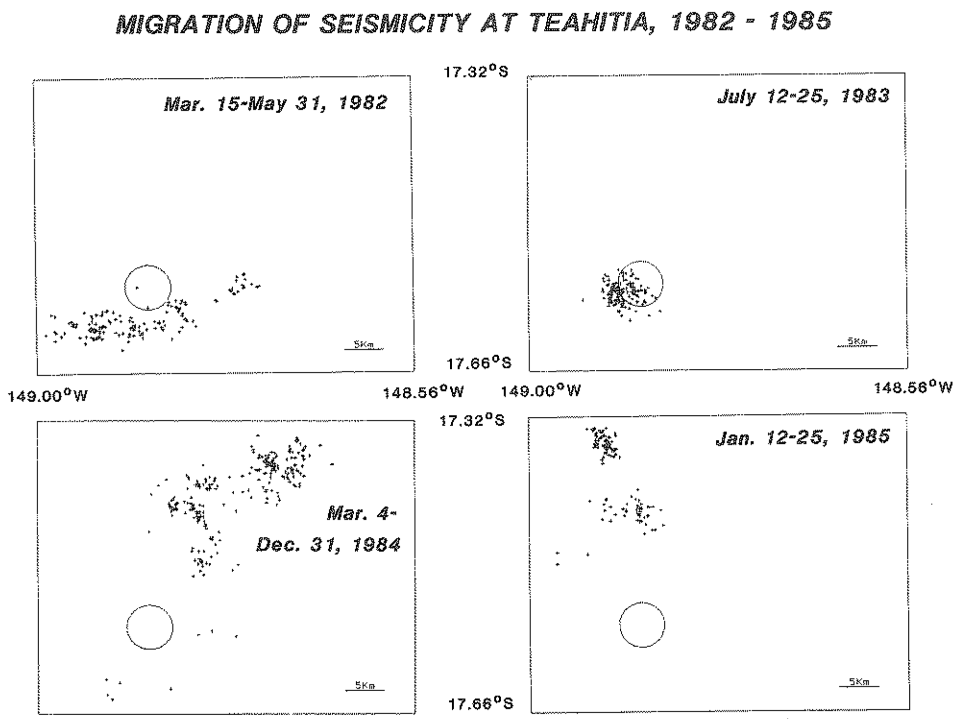


Figure 8

Epicentral migration of seismic activity at Teahitia during the 1982–1985 swarm. On each frame, the circle is a crude representation of the 2000 m isobath around the summit of the seamount. During the 1982 sub-swarm, the activity migrates regularly from East to West on the southern flank of Teahitia (see TALANDIER and OKAL (1984) for details). The dense patch of epicenters to the extreme North of the diagram in 1985 corresponds to the late 'tectonic' phase of activity, starting January 15, 1985.

migratory pattern was clearly recognized along the southern flank of the edifice. Activity in 1982 and 1983 was concentrated to the immediate vicinity of the seamount, but spread over a much larger area in 1984 and 1985, finally covering more than 1500 km^2. SEABEAM mapping in 1986 revealed the presence of cones and pinnacles in the summital area, as well as craters in the area active in 1984–1985, and allowed precise correlation of seismic foci with bathymetric features.

2.1.5. *Interpretation: Origin of Earthquakes*

Volcanic earthquakes have usually been interpreted as related to one of two mechanisms. During the initial phases of a volcanic episode, fluid magma is forced upwards through a series of conduits opening in rock. The increased pressure in the magma is transmitted to the country rock and the resulting stress can be released through brittle fracturing of the rock. This process obviously precedes the eruption of magma at the surface, and forms the basis for the use of volcanic seismicity for the purpose of forecasting eruptions of well instrumented volcanoes (*e.g.*, Kilauea, Mount St. Helens). These events can start as deep as 55 km (EATON, 1962; BUTLER, 1982).

After a major cycle of activity, the newly erupted volcanic edifice can find itself gravitationally unstable, and its mass can readjust itself through earthquakes. These events have occasionally been called 'tectonic' (*e.g.*, KLEIN, 1982) to emphasize that they are not directly related to a magmatic process; they remain of course, related to the presence, and general activity of the volcano. This type of earthquakes would include such phenomena as caldera collapses, and is believed to represent isostacy at work; they can occasionally reach very large magnitudes, as exemplified by the large tensional earthquake on February 8, 1971 ($M_s = 6.9$), which followed by six months a well documented eruption at Deception Island, Antarctica, 60 km away (FARMER *et al.*, 1982). In our particular case, in the Teahitia-Mehetia area, the largest events recorded (March 15, 1981, $M_L = 4.3$ at Mehetia; January 15, 1985, $M_L = 4.4$ at Tehitia) are believed to fall into this category.

2.1.6. *Frequency-Magnitude Relations; Use of b-values*

The study of the frequency-magnitude distribution of earthquakes has classically been taken as indicative of the level of weakness in the rock, and thus has been used extensively for the recognition of volcanic seismicity. This technique models the number N of earthquakes with magnitude M through a relation of the form

$$\log_{10} N = a - bM \tag{1}$$

(GUTENBERG and RICHTER, 1944). Laboratory fracture experiments by MOGI (1962) showed that values of b significantly larger than the worldwide average ($b = 0.9$) are associated with rock featuring thermal weakening or excessive fracturation. Documented subaerial volcanic swarms have featured b-values ranging from 1.4 to more than 3 (MINAKAMI *et al.*, 1969; BRANDSDÓTTIR and EINARSSON, 1979; MCNUTT and HARLOW, 1983). In order to conduct a significant frequency-magnitude investigation, it is important to use a dataset covering several units of magnitude. Therefore detection capabilities become of crucial importance when attempting to use a b-value as an identifier of the volcanic character of a swarm. In our particular case, noise characteristics at the Tahiti Island sites of the RSP allow consistent detection levels of $M_L = 1.6$ for Mehetia (closest station 120 km away) and $M_L = 1.0$ at Teahitia (closest station 60 km away). b-value results are summarized as part of Table 1 and examples shown in Figure 9. We refer to OKAL *et al.* (1980) for a description of the computation of the local magnitude M_L. It is immediately evident that the b-values feature an evolution from high values to lower ones comparable to worldwide averages, not only for the swarm at Mehetia in 1981, as reported by TALANDIER and OKAL (1984a), but also for the last episode of the Teahitia activity, in 1985. In both cases, the return of the b-value to a low figure ($\simeq 1.1$) was followed by the termination of intense seismic activity. Following KLEIN (1982) in the case of Loihi, we propose to interpret these episodes as the 'tectonic'

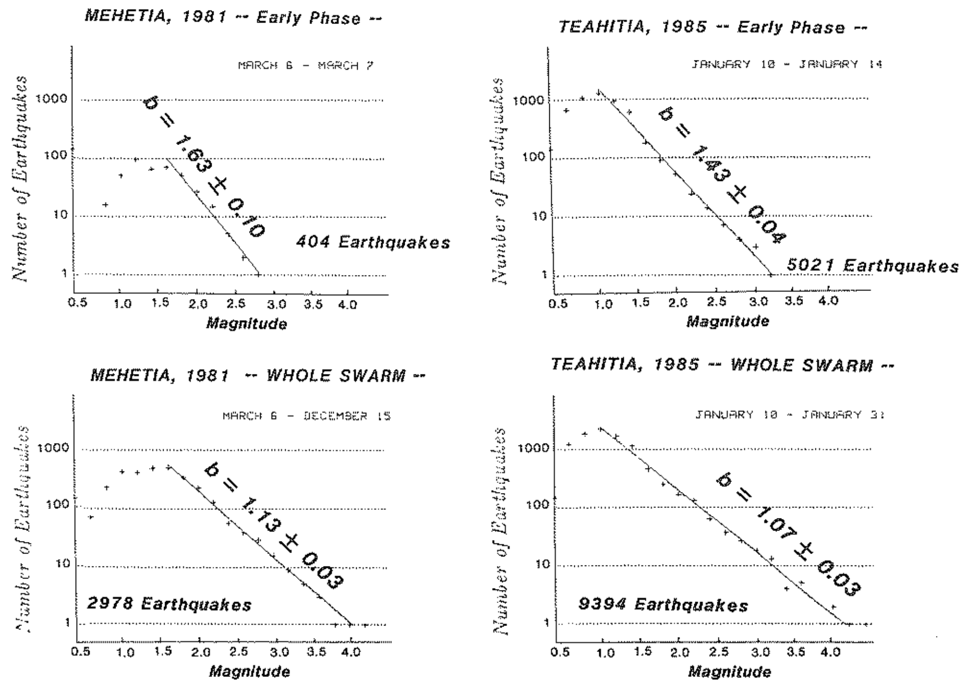


Figure 9

Examples of b-value determinations. *Left:* Mehetia swarm, 1981; *Right:* Final swarm at Teahitia, 1985. Magnitude windows are 0.2 units wide. Note that in both cases, the early phase of activity features a higher b-value, suggestive of magmatic activity, than the later episode of the swarm, which may represent tectonic readjustment. All other swarms at Teahitia (1982–1984) featured only high b-values (see Table 1).

readjustments which followed the eruptive activity. A similar evolution of b-values with time has also been observed at Krafla in Iceland (BRANDSDÓTTIR and EINARSSON, 1979).

2.1.7. *Characteristics and Classes of Earthquakes*

Following studies of densely instrumented subaerial volcanoes, a number of scientists have classified volcanic earthquakes on the basis of characteristics such as depth, spectral content, S-to-P ratio, etc. (MINAKAMI, 1974; LATTER, 1981). In the particular case of the Mehetia-Teahitia swarms and as discussed in detail in TALANDIER and OKAL (1984a), we were only partly successful in applying any of these classifications. This is due in part to the lack of precision on hypocentral depth, and also to the well-documented lack of uniformity in the seismic behavior of various volcanic sites (*e.g.*, MCNUTT, 1983; YUAN *et al.*, 1984). We present in Figure 10 representative spectra of P_n and S_n wavetrains recorded from Teahitia. Both events

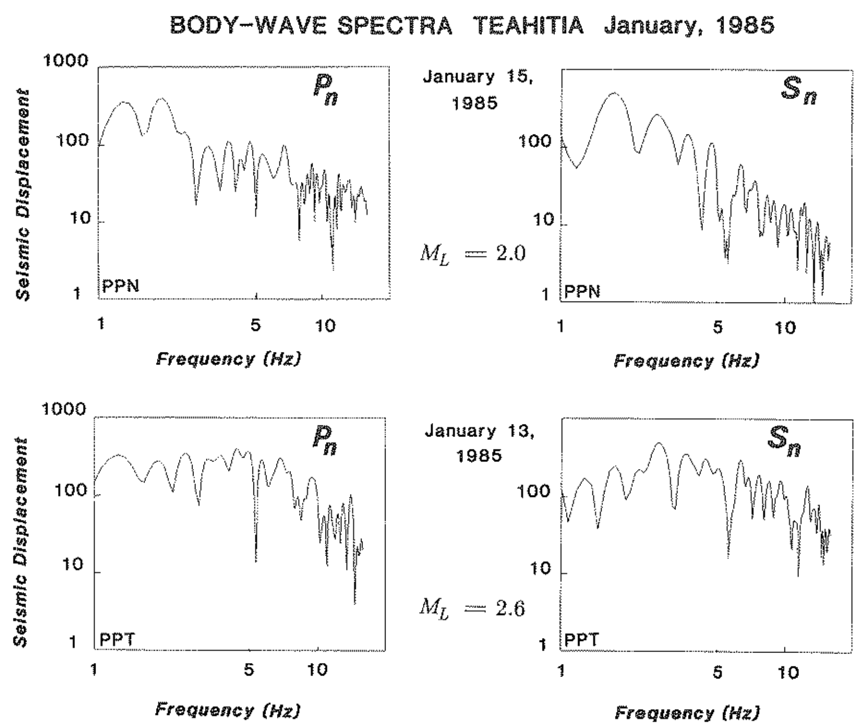


Figure 10
Representative body-wave spectra from Teahitia events: For both events, the left frame is a spectrum of P_n, and the right one of S_n at the same station. The January 15 event is lower-frequency than the January 13 one. Vertical scales are arbitrary.

occurred during the 'magmatic' phase of the January 1985 swarm. Note the difference in frequency content of the two events, particularly evident in their S_n spectra. The magnitudes of the two earthquakes, respectively $M_L = 2.6$ (January 13), and 2.0 (January 15) would, if anything, suggest the opposite behavior.

On the other hand, we were able to identify an evolution of the waveform of P waves recorded at the RSP network between, and occasionally during, swarms. Figure 11 shows P_n waves recorded at station PMO, on Rangiroa atoll, 320 km North of Teahitia, for two events of the 1982 Teahitia swarm. In the first case, the waveform is

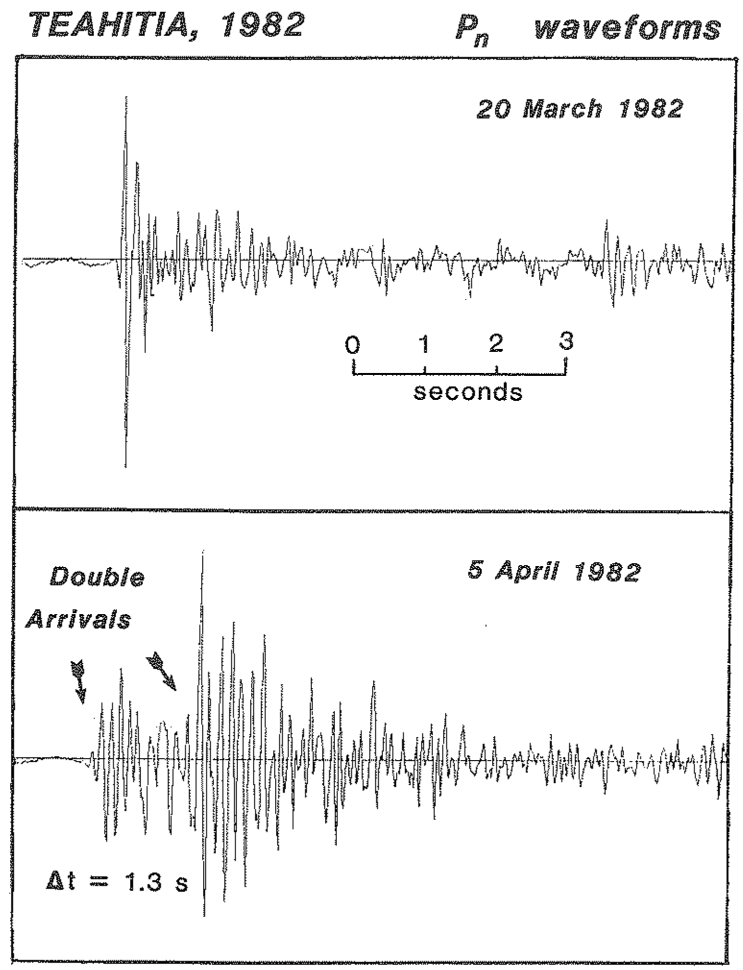


Figure 11
Comparison of P_n waves for two events from the 1982 Teahitia swarm, recorded at station PMO (Rangiroa). *Top:* Early stage of the swarm (presumed deep). Note simple waveform, and short duration of signal; *Bottom:* Later stage of the swarm (presumed shallow). Note second arrival, separated by 1.3 s, and much longer duration of wavetrain. The two events have comparable magnitudes.

extremely sharp and simple, while in the second case, double arrivals separated by 1.3 s are observed, with a much slower decay of amplitude. Significantly, the former type of earthquakes was found only in the early stages of the 1981 Mehetia swarm, and of the 1982 Teahitia one; it was absent from the 1983–1985 activity at Teahitia. Following EATON (1962), we interpret the first type as originating below the Moho discontinuity, and the second type as crustal; in addition, the latter often feature the direct crustal P phase as a later arrival. Thus, the sharper, deeper events would correspond to the initial opening of the magmatic conduits at depth, while the more complex, shallower ones could represent magmatic transfer in the crust and volcanic edifice.

2.1.8. Limitations

There are two limitations to the use of conventional seismological techniques in the study of underwater volcanic seismicity: first, not all episodes of a volcanic crisis give rise to individually recognizable earthquakes. In particular, the eventual eruption of lava on the ocean floor can take the form of a relatively continuous phenomenon, which may not be described properly as a succession of 'earthquakes'. In addition, detection capabilities for the largest parts of the ocean basins are still very limited, and adequate uniform coverage exists only at the level of $m_b = 4.5$. Experience gathered at such volcanoes as Kilauea, Loihi or Teahitia has shown that very few if any of the events at these sites reach this threshold during a single volcanic episode. The fast-opening Mid-Oceanic Ridges, conspicuously absent from the worldwide seismicity maps, are another obvious example of underwater volcanism whose seismicity totally evades detection by presently available permanent networks.

2.2. Tremor

In addition to waves identifiable as individual seismic phases (*e.g.*, P or S), volcanic tremor, consisting of more or less continuous seismic agitation, has been reported during certain phases of volcanic activity, and studied extensively (*e.g.*, AKI *et al.*, 1977; CHOUET, 1981; FEHLER, 1983).

In the case of the Teahitia swarms of 1982–1985, we recorded a considerable amount of high-frequency seismic tremor, of a rather spasmodic nature, and peaked around 7 Hz. We refer to TALANDIER and OKAL (1984a) for a complete description of its characteristics. In addition, we also recorded low-frequency seismic tremor (peaked around 2–3 Hz). Figure 7 gives a history of the occurrence of the two forms of tremor, and Figure 4 shows some typical examples of tremor recorded at RSP during the Teahitia swarms in 1982. It is significant that we did not record any tremor from the Mehetia swarm in 1981. We attribute its absence to the larger epicentral distance, while realizing that it remains possible that none may have occurred.

2.2.1. *Interpretation*

The so-called 'high-frequency' ($\simeq 7$ Hz) tremor has often been described as originating in the resonance of cracks opening in the rock under the pressure of magma (AKI *et al.*, 1977); more recently the role of the oscillation of the fluid body of magma itself inside the crack has been recognized, and it has been suggested that harmonic tremor may be dominated by Rayleigh waves of so-called 'long-period' earthquakes, excited by such an oscillation (CHOUET, 1985), and known to be excited substantially by the so-called '*B*-type' events (MCNUTT, 1986). Alternatively, FERRAZZINI *et al.* (1986) have proposed that tremor may be the expression of slow waves trapped in a fluid layer, and CROSSON and BAME (1985) have presented a model in which seismic energy characteristic of tremor and long-period earthquakes is due to the free oscillation of a spherical cavity filled with fluid magma and gas.

Low-frequency ($\simeq 2$ Hz) tremor has been reported at many subaerial sites in conjunction with the documented observation of the culminating phases of activity, involving deflation and fountaining (*e.g.*, EINARSSON and BJÖRNSSON, 1976; MCNUTT and HARLOW, 1983). Thus we hypothesize that submarine eruptions were taking place at the times when low-frequency tremor was recorded from Teahitia. We must emphasize, however, that some low-frequency tremor may escape detection, due to the relatively large distance to the recording stations.

Low-frequency tremor has also been associated with documented rockslides, for example during recent activity along the dome inside the crater of Mount St. Helens (*E.T. Endo*, pers. comm., 1985).

The interpretation of tremor activity in terms of the eruptive processes of a volcanic edifice is made extremely difficult by the great variability of the tremor patterns among well-instrumented subaerial volcanoes. This probably reflects the fact that the morphology of the magma plumbing, and the kinematics of the ascension of the magma itself, have no *a priori* reason to be similar under all volcanic edifices. For example, at Kilauea, AKI and KOYANAGI (1981) have correlated high-frequency tremor (about 7 Hz) with deep magma progression, occurring weeks to months before the eruption; they have further observed an evolution in tremor frequency with time. On the other hand, at many sites tremor develops only when eruption is imminent (only hours to days away), and lasts for the duration of the eruption (*e.g.*, EINARSSON and BJÖRNSSON (1976); EINARSSON and BRANDSDÓTTIR (1984) at the Icelandic volcanoes; MALONE *et al.* (1981) at Mount St. Helens; MCNUTT and HARLOW (1983) in Central America). In general, this tremor is low-frequency and these authors have not reported high-frequency tremors, despite adequate instrumentation, featuring response curves peaked at 25 Hz in the case of Central America.

In our case, high-frequency tremor is clearly absent from the earliest phase of earthquake activity (which we interpret as deepest) in 1982 at Teahitia, but rather starts when the seismic activity becomes shallow (around March 27, 1982); it readily accompanies the earthquake activity during the 1983, 1984 and 1985 swarms, which do not feature the episode of deep seismicity present in 1981 and 1982; this is of course in clear contrast with the case of Kilauea. Neither do we observe the evolution of frequency of the tremor noted by AKI and KOYANGI (1981) during the course of a tremor sequence. It must be kept in mind, however, that the two edifices are in different stages of their growth, Kilauea being a steady-growing, well-developed volcano, and Teahitia still in its infancy; while this may affect the mechanical state of the plumbing and the kinematics of the transition of the magma through it, it would be speculative to attempt any further interpretation at this stage.

2.2.2. *Limitations*

The principal limitation to the use of tremor for the monitoring of underwater volcanic activity is of course its rapid attenuation with distance. Although tremor has been observed up to distances of 630 km during the May 18, 1980 Mount St. Helens eruption (MALONE *et al.*, 1981), it is not observed at Kilauea at distances greater than 50 km (*R.Y. Koyanagi*, pers. comm., 1984). The fact that we do not observe any tremor from Mehetia at Tahiti may be due to its attenuation over the 120 km distance, rather than to its absence at the source. In practice, and in the marine environment, tremor has not been observed at distances greater than 90 km. At shorter distances. it can provide some insight into the level of magmatic activity, or into the eruptive process, which could be better monitored directly in the case of a subaerial volcano. For edifices concealed under water, the interpretation of tremor on the basis of comparison with documented cases must take into account the extreme variability of the characteristics of individual volcanoes.

2.3. *T Waves*

T waves propagated at great distances in the ocean's low-velocity SOFAR channel, and recorded at seismic stations[1] provide another means of distant detection of volcanoseismic activity. Because of the guided nature of the *T* wave, and of the totally negligible anelastic attenuation in water, *T* waves can propagate over truly teleseismic distances, *e.g.*, across the whole width of the Pacific Ocean, as long as no obstacle such as island or seamount masks their path.

2.3.1. *Example*

The most spectacular example of teleseismic detection of underwater volcanism by *T* waves in undoubtedly the recording of the eruption at Macdonald Volcano (29°S; 140°W) on May 29, 1967 (NORRIS and JOHNSON, 1969), and the subsequent

[1] Strictly speaking, *T* phases can only be recorded at sea by hydrophones; however, the seismic wave generated by their conversion at an island shore can be recorded by a seismometer on the island, and for simplicity we will call this a seismic record of the *T* wave.

discovery of the seamount by R. H. Johnson on July 20, 1969 (JOHNSON, 1970). Since then, Macdonald has been under constant monitoring by RSP stations, and 14 swarms have been documented to date (TALANDIER and OKAL, 1982; 1984b). Figure 5 shows examples of records of T waves from Macdonald eruptions.

2.3.2. *Location*

Precise velocity maps for T waves have long been available for the entire Pacific basin. We refer to JOHNSON (1966), JOHNSON and NORRIS (1968) and NORTHROP (1973) for a description of location algorithms, including the identification of rays reflected on seamounts. These techniques can be readily applied to the case of volcanoseismic events. Because of the comparatively low velocity of acoustic waves in the water, and of their guided character, T wave slownesses across a receiving array are up to one order of magnitude greater than for conventional seismic waves, and the precision of epicentral locations is thus enhanced. The RSP network has a particularly large aperture (1800 km); in addition, for sources such as Macdonald Seamount, located in or around French Polynesia, it provides ample azimuthal coverage; using a regional model of T wave velocities obtained from controlled-source experiments, we estimate an uncertainty of ± 5 km at distances as large as 1500 km. For sources lying at great distances across the Pacific, precise locations must rely on enhanced datasets, including stations outside Polynesia.

2.3.3. *Spectral Characteristics*

Teleseismic T waves efficiently guided by the SOFAR acoustic channel have wavelengths significantly smaller than its width, corresponding to frequencies of 2 Hz and above. Because ambient seismic noise at atoll locations is concentrated at lower frequencies, simple band-pass filtering allows excellent detection and recording capabilities; such special channels are operated between 1.8 and 12 Hz at the RSP stations, with routine magnification of over 2×10^6 at 3 Hz.

As discussed in detail by NORRIS and JOHNSON (1969) and more recently by WALKER *et al.* (1985), the durations and spectral contents of T waves recorded from earthquakes and from volcanic eruptions differ considerably. The comparison of Figures 5 and 12 is a time-domain illustration of these properties using RSP records: earthquake T waves are peaked around 3 Hz, and feature a rapid decay of amplitude with time, whereas volcanic T waves can feature a sharp onset, contain higher frequencies (peaked above 5 Hz), and last considerably longer.

2.3.4. *Detection of Distant Volcanism*

In addition to the monitoring of Polynesian volcanoes, such as Macdonald, T waves believed to have a volcanic origin are regularly received in Polynesia from the

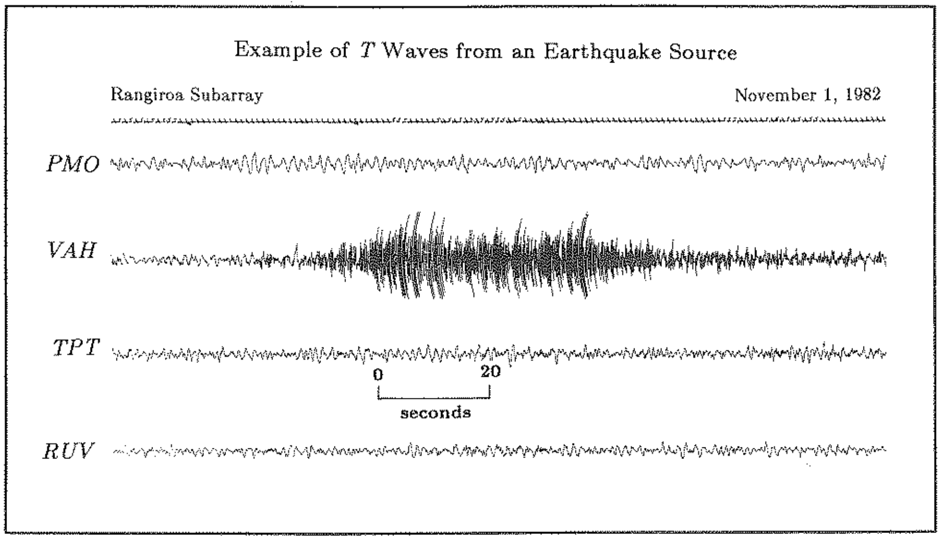


Figure 12
Example of T wave from an earthquake source (in this case a small event from the Balleny Islands area) recorded at the Rangiroa subarray of the RSP. Tick marks on upper trace are seconds; note short duration, and slow, emergent character of the signal, as compared to T waves recorded from volcanic phenomena (see Figure 5).

following regions: Rumble Seamounts (North of New Zealand); Kermadec and Tonga Islands; Volcano, Bonin and Izu Islands; Galápagos Islands.

In the case of Tonga, a particularly active source area has been Monowai Volcano, a recognized active seamount located half-way between Raoul Island (Kermadec) and 'Ata (Tonga) (Davey, 1980). In addition to a particularly intense swarm of explosive T waves received for 10 hours, on January 9, 1980, a seismic swarm took place in 1969, and activity was also detected in 1977 and 1982, in the latter case suggesting a quiet eruption without explosive sequences. Activity further south at Kermadec was detected in August, 1986. In the case of Rumble Seamounts, a swarm of T waves accompanied the seismic activity which culminated with the earthquake of December 30, 1984 ($M_s = 6.8$). This swarm is described more extensively in Section 3.9.

2.3.5. Origin of T waves

The physical origin of T waves recorded as part of a volcanoseismic crisis is not fully understood, and can only be speculated upon. This is due in part to the obvious lack of T waves in the case of subaerial volcanoes, resulting in the impossibility of a direct comparison with a case of documented eruption. In addition, the generation of any wave with a wavelength $\lambda \leq 500$ m is controlled by terrain morphology on a similar scale, which is beyond the precision of our knowledge of the bathymetry of ocean basins. As a result, there is in general no direct relationship between the amplitude of a recorded T wave and the magnitude of the parent earthquake (see TALANDIER and OKAL (1979) for examples and details).

As a first mechanism, it is reasonable to assume that genuine earthquakes occurring during a volcanic episode, can see a part of their radiated seismic energy converted to T waves on a section of the ocean floor either in contact with the SOFAR channel, or featuring adequate dipping. Given such a favourable geometry at the conversion point, the depth of the earthquake is largely irrelevant; indeed T waves are observed routinely from earthquakes as deep as 650 km in the Fiji trench (TALANDIER, 1972). This interpretation is especially appealing in the case of T wavetrains featuring sharp onsets, such as shown on the top half of Figure 5, which allow precise location of the source. However, no conventional seismic wave was ever recorded from the Macdonald site, indicating that any earthquake activity remained below the detection level of $M_L = 3.5$ for this location. We interpret the sharp onsets characteristic of the initial phases of some of the Macdonald swarms as explosive events near the flank or summit of the seamount. We speculate that they correspond to the opening of magmatic conduits inside the edifice or on its flanks.

On the other hand, some T-wave swarms do not feature such explosive events; in addition, explosive events themselves can be followed by a prolonged T wavetrain, lasting up to several hours. Neither observation can be explained by propagation effects, and this rules out interpreting such T waves as the only recorded seismic phases of an otherwise undetected earthquake. These signals are believed to represent acoustic noise generated by a later stage of magmatic activity, which does not involve sudden rock fracture.

Significantly, *no* T waves (other than associated with individual earthquakes) were recorded from either Mehetia (depth to caldera 1700 m) or Teahitia (depth to summit 1450 m). Similarly, no T waves of a magmatic character were recorded in Polynesia from Loihi during the documented swarms of activity (KLEIN, 1982) at this Hawaiian seamount (depth to summit 950 m); the Mid-Oceanic Ridge system (in the present case its East-Pacific and Pacific-Antarctic segments) is another example of 'silent' underwater volcanism, *i.e.*, taking place without giving rise to magmatic T waves. On the other hand, eruptions at Macdonald (depth to summit 29 m) and Monowai (depth to summit 117 m) have generated abundant T waves. Generalizing these results, it appears that the excitation of T waves by the magmatic phases of a submarine volcanic eruption (as opposed to those accompanying documented earthquakes) requires that the seamount structure be at most a few hundred meters below sea level. However, we do not believe, and for several reasons, that this merely expresses the condition that the eruption should take place inside the SOFAR channel (classically taken from 600 to 1800 m in intertropical regions). First, the silent volcanoes Loihi, Mehetia (flank), and Teahitia, as well as the East Pacific Rise, all protrude substantially into the SOFAR. Secondly, while the SOFAR channel acts as an efficient T-wave guide, events exciting substantial T waves are known not to be restricted to its precise depth extent: for example, seismic refraction campaigns, using surface explosions shot at a typical depth of only a few tens of meters routinely generate T waves efficiently recorded at teleseismic distances (OKAL and TALANDIER, 1986). Finally, theoretical investigations of the excitation of T waves, modeled as a superposition of surface modes, indicate that the depth of an explosive source in the water column plays only a moderate role in the amplitude of the resulting T wave (*E. A. Okal, in preparation*).

We are thus led to propose that the existence or absence of T waves during the magmatic phases of a submarine eruption is due to an inherent difference in the physical mechanism of the eruption. Ever since the works of MCBIRNEY (1963; 1971), it has been recognized that water depth plays an important role in governing the eruptive behavior of oceanic volcanism; recently, and on the basis of the petrography and mineralogy of basalts from La Palma Islands, STAUDIGEL and SCHMINKE (1984) have suggested that the maximum depth of explosive underwater volcanism (involving intense degassing, and possibly violent forms such as underwater fountaining or steam venting, as proposed by BATIZA *et al.* (1984)) may be as shallow as 780 m, significantly less than the 2000 m proposed by MCBIRNEY on the simple basis of the critical pressure of seawater. While the argument remains speculative in the absence of a good understanding of T wave excitation, it is interesting to note that this figure of 780 m would explain the observed difference of behavior between 'silent' and 'noisy' underwater volcanoes.

2.3.6. *Limitations*

As mentioned above, T waves can be generated by volcanic earthquakes or magmatic phenomena only under favorable conditions at the source. In attempting to use T waves to recognize underwater volcanism, it is important to record 'magmatic' T waves, not associated with an earthquake. The above discussion has shown that none have been documented from edifices whose summits remain deeper than 950 m.

In addition to this fundamental limitation at the source, the propagation (and hence reception) of the T wave can be affected by masking of the path by a seamount, or island structure. A particular station or small-aperture array can then be shielded from large epicentral areas. In conclusion, T waves can be most valuable for underwater volcano detection of shallow edifices, given a wide-aperture array of hydrophones or island stations.

3. *Discussion*

In assessing the level of volcanic activity of the ocean floor, one of the major problems remains the correct identification of seismic activity as being, or not being, of a volcanic nature. In this section, we examine critically a number of seismic swarms at various locations in the Pacific Ocean (see Figure 2), and discuss the evidence arguing for or against volcanism. For the sake of completeness, we include the Polynesian swarms. We list the various sites in the order of decreasing ability to decide on the nature of the swarms.

3.1. Teahitia-Mehetia Area, French Polynesia, 1981–1985

Ample evidence of confirmed volcanism; see TALANDIER and OKAL (1984a; 1985) and above discussion.

3.2. Macdonald Seamount, Austral Islands Chain, 1967–1986

Ample evidence of volcanic activity; see NORRIS and JOHNSON (1969), JOHNSON (1977), TALANDIER and OKAL (1982; 1984b).

3.3. Kaitoku Seamount, Volcano Islands, 1984

An intense swarm of T waves of mostly explosive character was received during March and April, 1984 from this site located about 8950 km from the RSP Network; 588 separate events were detected. The occurrence of a submarine eruption at Kaitoku, 130 km North of Iwo Jima, was confirmed by similar reports from the Wake hydrophone array (WALKER *et al.*, 1985) and direct monitoring of the site.

3.4. Monowai Seamount, South of Tonga, 1969–1982

Strong T waves of mostly explosive character; volcanic site confirmed in 1978 (DAVEY, 1980).

3.5. Epicentral Location LI-2, Line Islands, 1967–85

This is one of several sites of recurrent seismic activity identified by OKAL *et al.* (1980). Seismicity takes the form of swarms of a few months to 2-yr duration. Since OKAL *et al.*'s (1980) study was published, a few events with magnitude $m_b = 4.3$ occurred in November, 1985. The swarms at LI-2 featured a b-value of only 0.86, comparable to worldwide averages. Focal mechanisms of the "ridge-push" type can be readily interpreted in terms of ambient tectonic stress. Finally, a shipboard expedition to the area failed to reveal any anomalous bathymetric structure (SVERDRUP and JORDAN, 1979). All these observations argue strongly against a volcanic origin for these events.

3.6. Tori-Shima Island, June 13, 1984

KANAMORI et al. (1986) have documented an anomalous seismic event near Tori-Shima Island, for which they propose a compensated linear vector dipole source, rather than a classical double-couple. They interpret this source as water-magma injection into a shallow sedimentary layer. We have recorded abundant T waves from this event (see Figure 13). The T wavetrain starts by an explosive event, and continues for about 15 minutes; in addition, a precursory explosive T wave is observed 70 s before the main event. Additional activity, in the form of T-wave or small explosions, continues for approximately 14 hours. The characteristics of the T wave, both in time and frequency domains, are comparable to those from Macdonald events; their arrival times and frequency suggest little or no body wave path prior to T-wave generation; thus our results are in general compatible with the model presented by KANAMORI *et al.* (1986).

3.7. Gilbert Islands Swarm, 1981–1983

The area south of the Gilbert Islands underwent 225 recorded earthquakes ($2.9 \leq m_b \leq 5.9$) between December, 1981 and September, 1983 (LAY and OKAL, 1983). The b-value is relatively high (1.35). No seamount structure is known from the available (but sparse) shipboard bathymetry, nor can any be inferred from SEASAT altimetry. The Gilbert Islands are masked from Tahiti and Moorea by the Leeward group in the Society Islands, and from Rangiroa by the nearby atoll of Tikehau.

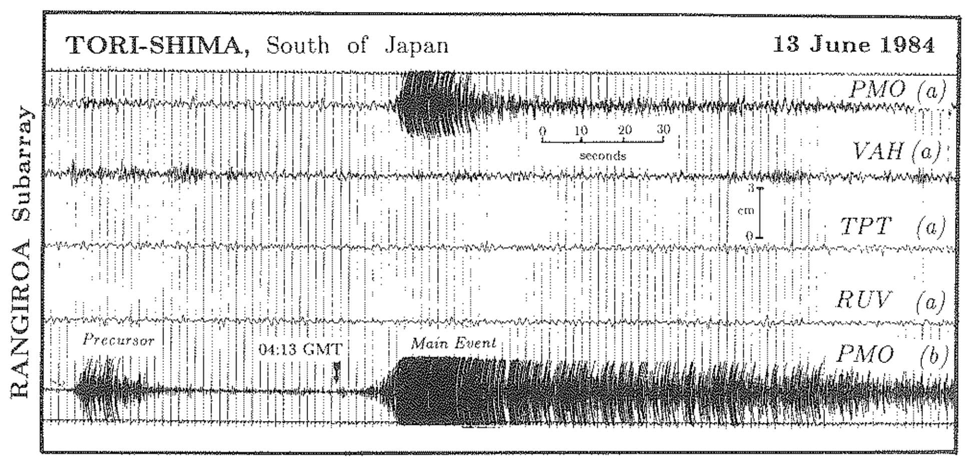


Figure 13
T waves recorded at the Rangiroa subarray of RSP from the Tori-Shima events of June 13, 1984. Traces (a) are standard short-period channels, with a magnification of 125,000 at 1 Hz; trace (b) at station PMO is the higher-frequency channel, with a magnification of 2,000,000 at 3 Hz (see TALANDIER and KUSTER (1976) for details). Note precursor preceding main event by 70s.

However, hydrophone sites to the north did not record any anomalous T wave activity (*D. A. Walker*, pers. comm., 1987). It is difficult to ascribe the Gilbert swarm to volcanism, given the particularly large size of the earthquakes (the largest event having $M_s = 6.0$, and at least a dozen more $m_b \geqslant 5.5$, a feature unrepeated in volcanic swarms). Finally, while focal mechanisms cannot be understood in the context of simple plate tectonics, they show the consistent release of horizontal compressional stress; this is not expected of the largest earthquakes in a volcanic sequence, which express either 'tectonic readjustment' through normal faulting mechanisms (*e.g.*, Fernandina, Galápagos (FILSON, *et al.*, 1987); Deception Island (FARMER *et al.*, 1982)), or a catastrophic eruption involving non-double couple sources (*e.g.*, Mount St. Helens, (KANAMORI *et al.*, 1984); Kalapana, Hawaii (EISSLER and KANAMORI, 1985)). On this basis, we have argued against a volcanic origin for the swarm, and proposed (OKAL *et al.*, 1986) that the occurrence of this swarm may be related to an ongoing relocation of the Vanuatu–Santa Cruz subduction zone to the north, a speculative model also supported by other geophysical evidence (KROENKE and WALKER, 1986).

3.8. Epicentral Location GB-5, Northeast of Gambier Islands, 1976–1979

This other site of considerable seismic activity (97 events detected over a period of 3 years) in the vicinity of French Polynesia was visited by *R/V Jean Charcot* in the Summer of 1980 (*J. Francheteau*, pers. comm., 1981); partial SEABEAM mapping revealed a complex seafloor tectonic pattern. The few available focal mechanisms are compatible with known tectonic stresses; the very low b-value (0.55) obtained by OKAL *et al.* (1980) also argues against volcanism; it could however be biased by the relatively poor detection characteristics at this location. OKAL and CAZENAVE (1985) have proposed to attribute the preferential release of seismicity in the area to the weakness in the plate resulting from a complex history of ridge migration and jumping. On the other hand, during the visit of the *R/V J. Charcot*, an ocean-bottom hydrophone was operated for 7 hours, and recorded 70 events, possibly featuring explosions and rumbling, but whose characteristics (distance, magnitude, origin) could not be determined further; the possibility of active volcanism at the site cannot be totally ruled out.

3.9. Rumble Seamounts, North of New Zealand, 1984–1985

An intense earthquake swarm was recorded from this location starting December 28, 1984, and lasting until January 14, 1985. The earthquakes are located exactly on the line of arc volcanism, between White Island and Rumble III Seamount, both documented active volcanoes. The main shock occurred at 36.66°S and 177.51°E on December 30, 1984, and had magnitudes $m_b = 6.2$ and $M_s = 6.8$; a total of 37 events are listed in the PDE bulletins. T waves, of the earthquake type, were recorded for another 1.5 months, independently of conventional seismic waves. A total of 666

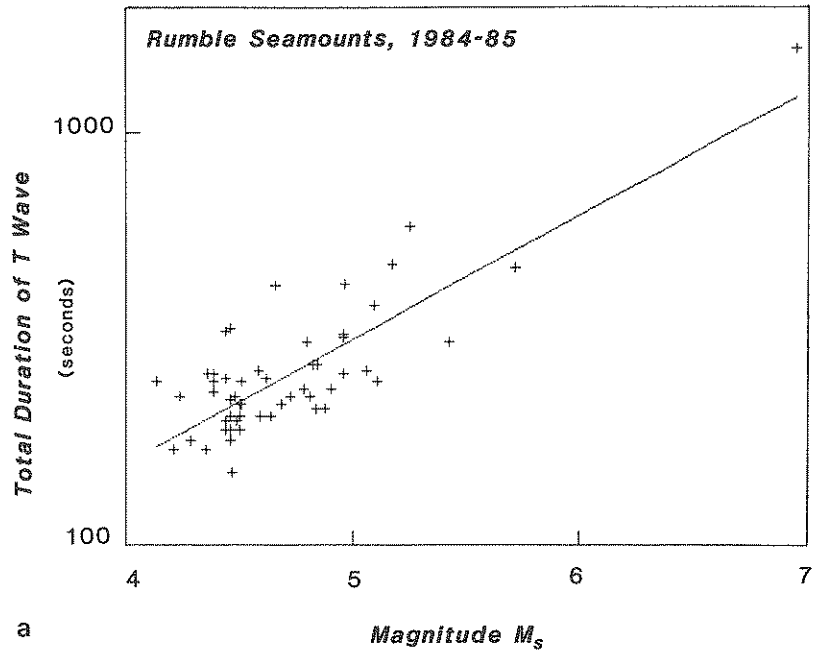


Figure 14a
Total duration of T waves at the RSP network against observed M_s values, for the 55 largest events of the Rumble Seamounts swarm, December, 1984–January, 1985. The solid line is the least-squares fit of $\log_{10}\theta$ to M_s.

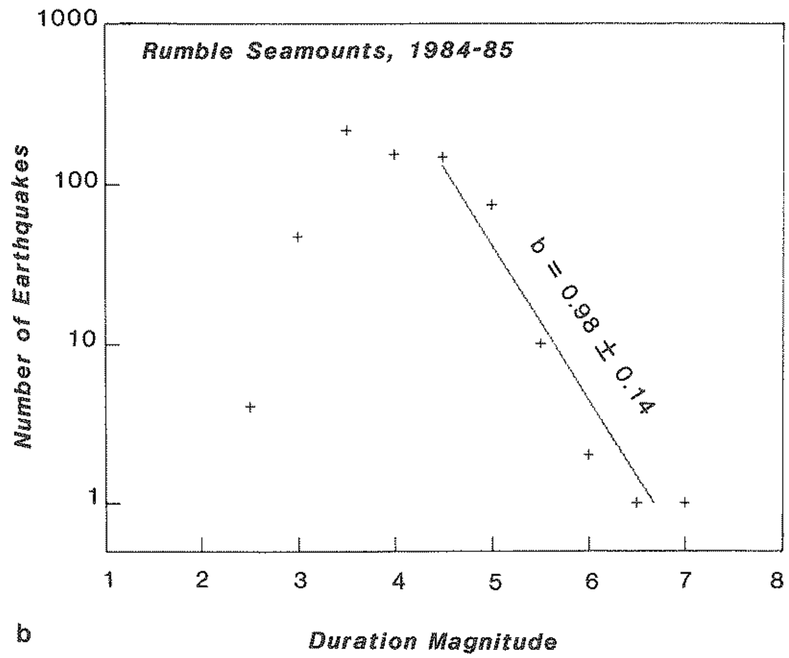
events were thus identified by the RSP Network, 55 of which produced surface waves of sufficient amplitude to allow measuring M_s.

Following the techniques described by OKAL and TALANDIER (1986), we used T waves to define a duration magnitude, calibrated by the 55 events for which a local measure of M_s is available (see Figure 14a). The regressed relationship

$$\log_{10}\theta = 1.00 + (0.30 \pm 0.03)M_s \qquad (2)$$

where θ is the total duration of the T wavetrain in seconds, features a slope in excellent agreement with the expected theoretical value (1/3). We then assigned duration magnitudes to all events detected by their T waves, and conducted a frequency–magnitude analysis (see Figure 14b). The resulting b-value of 0.98 ± 0.14 is within the range of the worldwide average.

The occurrence of major shallow earthquakes arc-wards of a subduction trench is well documented in areas of oblique subduction such as Sumatra, where decoupling of the plate motion takes place and significant strike-slip motion is



b

Figure 14b
Frequency-magnitude relation for all events recorded at the RSP network during the Rumble swarm. Magnitude windows are 0.5 units wide. Equation (2) is used to define the duration magnitude from the T waves. The solid line is the regressed data.

observed inland along the Semangko fault (FITCH, 1972). However, the amount of obliquity of the subduction along the Kermadec trench is only 10 to 15° *vs.* 35° in Sumatra (MINSTER and JORDAN, 1978). Furthermore, a slight component of underthrusting is expected to accompany the strike-slip earthquakes in the Sumatran model; focal mechanisms published for the principal events in the 1984 Rumble swarm show consistently a strong component of normal faulting (see Figure 15).

Except for the low b-value, which is poorly constrained, these characteristics would be difficult to reconcile with a purely tectonic origin of the swarm. We think it is strongly possible that an episode of active volcanism took place in the area during or before the seismic swarm. The absence of T waves characteristic of the magmatic phase could be due to the large depth to the seafloor in the area (2500 m).

3.10. Crough Seamount, Southcentral Pacific, 1955

In this last example, we investigate an isolated earthquake of large magnitude (m_b = 6.8; M_s = 6.2) which occurred in 1955 on the flanks of Crough Seamount, a major structure prolonging the Oeno–Henderson–Ducie island chain in the

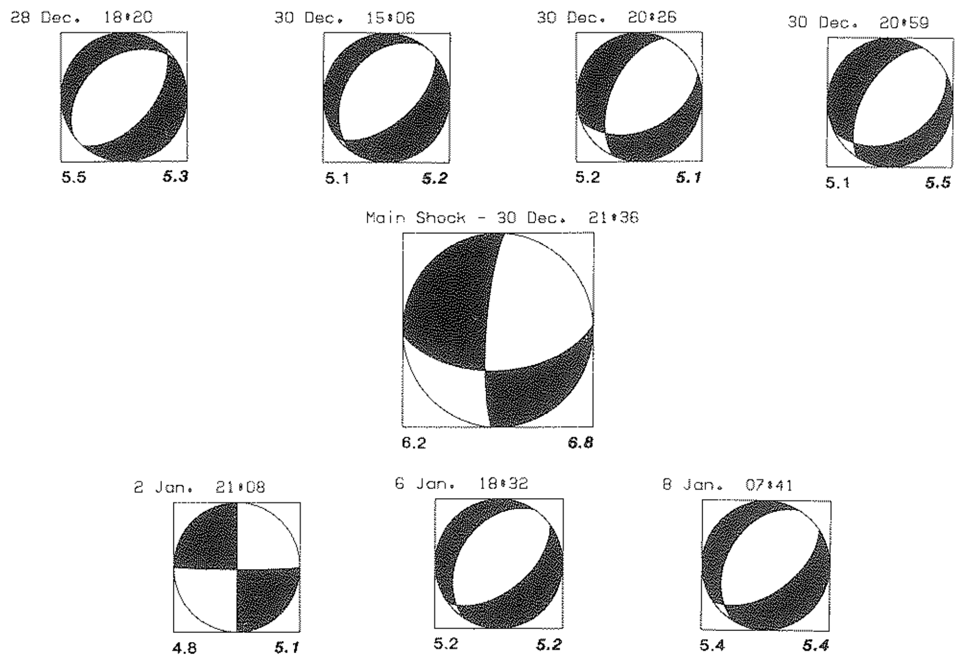


Figure 15
Focal mechanisms of the largest earthquakes in the Rumble swarm, December, 1984–January, 1985. Black-shaded areas are compressional first motions (after the Monthly Bulletins of the Preliminary Determinations of Epicenters). For each event, magnitudes m_b (in roman) and M_s (in italics) are also given.

Eastcentral Pacific around 25°S, 122°W. As discussed in detail by OKAL (1984) and OKAL and CAZENAVE (1985), Crough Seamount has to be a young ($\simeq 3$ Ma) feature which may still undergo sequences of late volcanism, comparable to the 'post-erosional' stages at subaerial sites; the size and focal mechanism of the 1955 shock, as well as the general morphological framework are very similar to those of the 1971 Deception event mentioned earlier. Under these conditions, and despite its definitely speculative nature, it is impossible to rule out the hypothesis that the 1955 event was the only recorded earthquake of an otherwise unsuspected episode of volcanism at Crough Seamount.

4. *Conclusion*

Using the example of the RSP network in French Polynesia, we have shown that the detection of underwater volcanism by high-quality seismic stations located on islands is feasible, but limited in range. Short of direct exploration of an individual site, it remains our only means of monitoring and understanding the activity of the ocean floor.

Conventional seismic waves are the best tool in this respect, but a clear insight into a volcanoseismic swarm requires low magnitude detection levels which strongly restrict the geographical area which can be adequately covered by any network. It is probably futile to try to monitor underwater volcanoes at distances greater than $\simeq 300$ km. Seismic tremor, when recorded, can give a detailed picture of the evolution of magmatic activity inside the volcano; however, it has not been recorded at distances greater than 90 km in the marine environment. T waves offer a very powerful tool, which is not limited intrinsically by distance. However, both the generation of T waves, and their possible masking along a given source-receiver path are controlled by the bathymetry and morphology of the ocean floor on a scale which renders their observation problematic.

On the basis of the cumulative duration of tremor, TALANDIER and OKAL (1985) have estimated that the activity at Teahitia during 1982–1985 may have been comparable to that of Kilauea over a period of 18 years. Yet not *one* earthquake from the swarms was recorded outside Polynesia during this period, and the activity would have gone undetected were it not for the existence of the RSP network (only a few of the larger events in 1982 and 1985 were felt on the island of Tahiti). Thus, one of the most intriguing questions regarding underwater volcanism is 'How much goes undetected?' The perfect scenario for 'camouflage' of an underwater volcanic swarm would involve a maximum earthquake magnitude $m_b < 4.3$, and eruption at depths of 1000 m or more at a site located more than 500 km away from the nearest seismic array: the earthquakes would go undetected by the worldwide network; the regional stations, if any, would have poor location capability, and the water depth would be sufficient to inhibit the generation of T waves from the magmatic episodes of the swarm. Needless to say, the great majority of the vast expanses of the Pacific (and other oceans') basins qualify for this scenario.

Even the detection of a few earthquakes or sufficient magnitudes may shed the wrong light, since they could be mistaken for isolated earthquakes of tectonic origin. Conversely, the many foci of seismicity inside the oceanic basins hold, at least theoretically, the potential for being unsuspected volcanic sites. Fortunately, for many of these sites, we have been able to resolve and interpret the patterns of stress release as tectonic processes which are well understood in the framework of plate theories (OKAL, 1983); shipboard exploration at the epicenter of a major seismic swarm at Region A in the Line Islands area has failed to reveal any bathymetric features suggestive of active volcanism (SVERDRUP and JORDAN, 1979). Nevertheless, almost any isolated seismic focus involving a single event for which no focal mechanism information is available could well be the site of unsuspected active volcanism.

In this respect, the systematic instrumentation of the ocean floor by permanent ocean-bottom, or sub-bottom stations in areas far removed from continents and islands is a highly desirable goal, and a necessary step towards better understanding of the characteristics of mid-oceanic underwater volcanism.

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