105.

Bulletin of the Seismological Society of America, Vol. 86, No. 5, pp. 1529–1544, October 1996

Monochromatic *T* Waves from Underwater Volcanoes in the Pacific Ocean: Ringing Witnesses to Geyser Processes?

by Jacques Talandier and Emile A. Okal

Abstract We analyze swarms of exceptionally intense and sustained *T* waves recorded during 1991 through 1994 at the Polynesian seismic array. The strongest swarm lasted 15 months and originated southwest of the Eltanin fracture zone, in the immediate vicinity of a documented seamount associated with a recognized volcanic chain branching out of the local segment of the mid-oceanic ridge. The Eltanin *T* waves are characterized by an exceptionally monochromatic spectrum featuring a single line (of a few Hertz frequency) in the range 2 to 80 Hz. While no similar characteristics had previously been observed in 25 yr of recording in Polynesia, comparable spectra were recorded during a 1993 swarm at the Revilla Gigedo Islands, following by 4 months a documented eruption of Socorro Island. In both cases, we interpret the *T* waves as evidence of a major underwater volcanic process. The existence of a single resonating frequency, and in particular the absence of any overtones, is generally not compatible with the now classical models of the resonance of a fluid-filled crack, and we speculate that the source of the phenomenon may be the oscillation of a bubbly liquid, which could result from vaporization of seawater in the presence of a large lava lake.

Introduction and Background

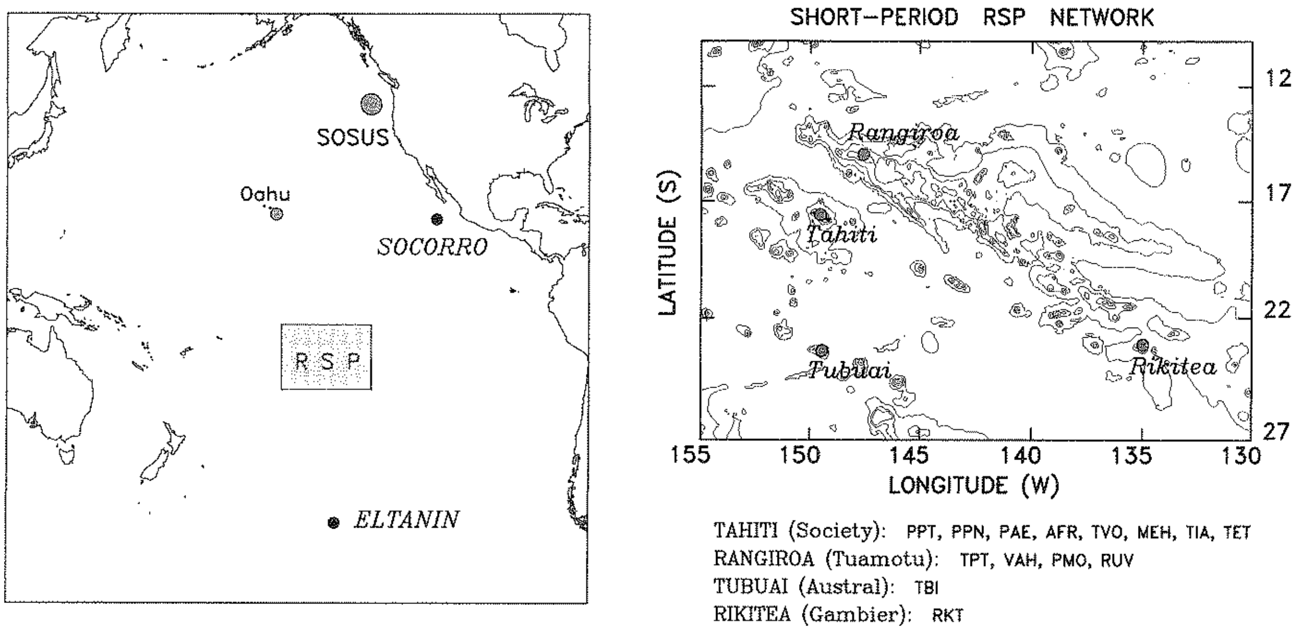
The purpose of this article is to report on a series of remarkable swarms of *T*-wave activity recorded in 1991 to 1993 at the Polynesian seismic array (Réseau Sismique Polynésien; hereafter RSP)[1] and believed to be manifestations of otherwise unsuspected underwater volcanism. By far, the most important swarm took place in the vicinity of the Eltanin fracture zone and lasted 15 months. Another significant swarm, lasting several weeks, took place near Socorro Island in the northeastern Pacific. The most remarkable feature of these swarms is their perfect monochromatism: their spectrum generally shows a single frequency (unmarred by harmonic overtones) over the full range 2 to 80 Hz. This characteristic sets the sources aside from any other volcano-acoustic source previously recorded in Polynesia. We speculate that the source of these monochromatic *T* waves could be the resonance of columns of bubbly liquids associated with exceptionally intense hydrothermal activity during underwater volcanic eruptions.

Ever since the 1952 eruption of Myojin was detected in California by trans-Pacific *T* waves (Dietz and Sheehy, 1954), the latter have been widely used to monitor the activity of underwater volcanoes, and occasionally, to discover unsuspected ones, e.g., Macdonald Seamount in 1967 (Norris and Johnson, 1969). In particular, RSP routinely records activity generated throughout the Pacific basin (Talandier and Okal, 1987; Talandier, 1989). More recently, *T* waves detected on military hydrophone arrays have identified the onset of a burst of volcanic activity along the Juan de Fuca ridge, motivating the dispatch of vessels and additional ocean-bottom equipment for an exceptionally detailed *in situ* look at the process of ocean-floor spreading (Fox, 1995). In this respect, the mere observation of *T* waves generated by abyssal volcanism is not unique; we believe, however, that the spectral characteristics of the sources described in the present study are unique.

The RSP network consists of several subarrays of short-period seismometers deployed on various island chains in French Polynesia. We refer the reader to Figure 1 for its general layout and to Talandier and Kuster (1976), Talandier and Okal (1979), and, more recently, Talandier (1993) for a detailed description of the various channels used for routine monitoring and recording. An important feature for the present study is the availability of the analog "*T*-wave channel," which uses narrow band-rejection filters to eliminate seismic noise due to sea swell and allows amplifications in the range 5×10^5 to 2×10^6 at 3 Hz, depending on the background noise at the stations. Permanent monitoring of the *T*-wave

---

[1] Strictly speaking, a *T* wave can be recorded only at sea, using a hydrophone. Upon reaching shore, it is converted to a seismic wave, which can be recorded by a land-based seismometer. For simplicity, we call this a seismic record of the *T* wave.



TAHITI (Society): PPT, PPN, PAE, AFR, TVO, MEH, TIA, TET
RANGIROA (Tuamotu): TPT, VAH, PMO, RUV
TUBUAI (Austral): TBI
RIKITEA (Gambier): RKT

Figure 1. Left: map of the Pacific ocean showing the location of sources described in this study (solid circles) and *T*-wave receivers. The RSP is detailed at right. The circle for the SOSUS receiver is a rendition of the low-grade estimate of its location, as given by Fox *et al.* (1995). Right: map of French Polynesia showing the location of the RSP subarrays. The individual stations are listed below the map. Additional stations have been operated on a temporary basis on Tuamotu atolls.

channel on paper records can alert the seismologist to events otherwise undetected from body waves. Independently, the full original seismograms are digitized at 50 samples/sec and recorded, with meaningful spectral information thus available for frequencies up to ≈16 Hz.

Among the signals routinely recorded by RSP over the past 4 yr, we earmark five strong *T*-wave swarms of particularly interesting characteristics: The Eltanin swarm (1991 to 1992) and the Socorro swarms (May and June 1993) featured remarkably monochromatic signals; in addition, the eruption near Socorro Island on 19 January 1993 had a broad spectrum more in accord with previously documented volcanic eruptions; it was followed on 19 November 1993 by a short-lived swarm featuring a relatively broad spectrum, on which a strong spectral line was superimposed in the vicinity of 10 Hz. Finally, an additional puff of activity took place on 20 November 1994, which, however, could not be located. In the present report, we emphasize the Eltanin swarm, because of its exceptional strength and duration, which we interpret as evidence of a sustained episode of volcanic activity.

The Eltanin Swarm

The Eltanin acoustic swarm began on 10 March 1991, and the last signals were received on 12 June 1992. During that time, windows of intense or average activity alternated with periods of quiescence (at least, as recorded by RSP), which lasted as long as several weeks. Figure 2 provides a history of activity of the swarm. The strongest activity took place in March and August 1991.

The recordings presented in Figure 3 are remarkable in many respects: first and foremost, the amplitude and duration of the *T* waves are unusually large. The spectrum of a 5-mn window recorded at Vaihoa (VAH; located on the southern shore of Rangiroa atoll) and shown in Figure 4 exhibits prominently the strength of the signal: It towers about 40 dB above the ambient noise, and the deconvolved ground velocities reach 1.4 μm/sec peak to peak, enough to fully saturate the analog (paper) records on the *T*-wave channel. Of course, the amplitude of a *T* wave at an individual station is a complex and poorly understood function of many parameters (such as small-scale bathymetry) that control the seismic-to-acoustic conversion at the source and the acoustic-to-seismic one at the receiver. Nevertheless, the amplitudes achieved at all sites in Polynesia argue for a very strong source. The signals are composed of wave packets of strongly modulated, but usually well-sustained amplitude, which can last anywhere from a few seconds to several minutes. This is in clear contrast to more classical forms of underwater volcanism (e.g., at Macdonald Volcano), where surges of activity can occasionally reach amplitudes larger than recorded here, but typically last only a few seconds, and are separated by long episodes of "noise" at a considerably reduced level (Talandier and Okal, 1987).

The records can be grossly classified into three groups:

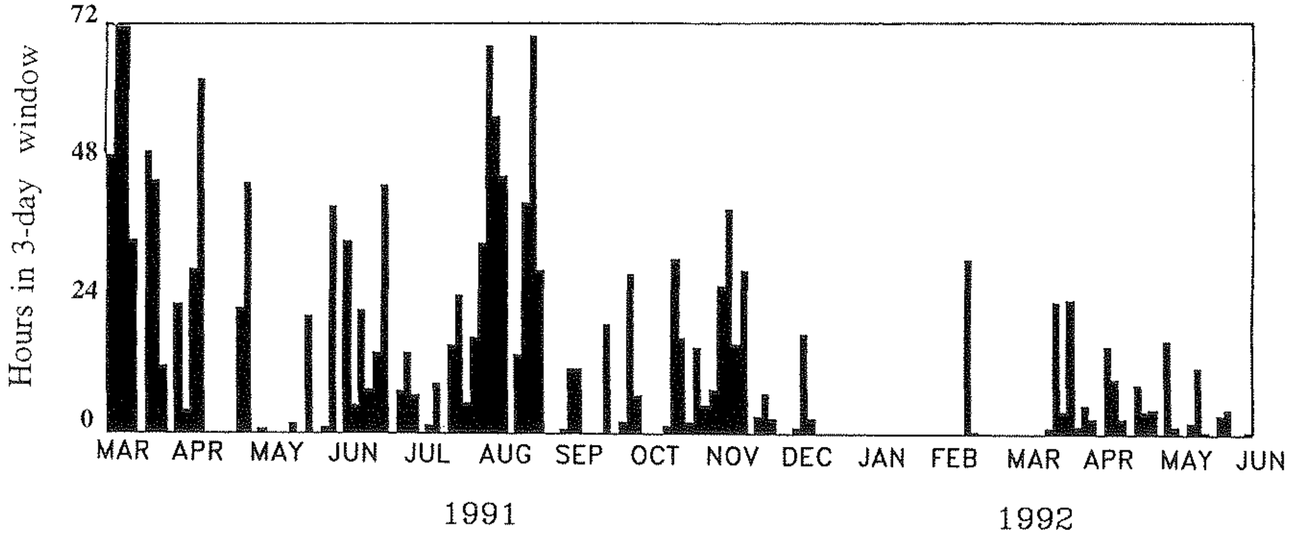


Figure 2. History of activity of the Eltanin source, plotted as the duration of *T*-wave activity recorded at RSP during 3-day windows.

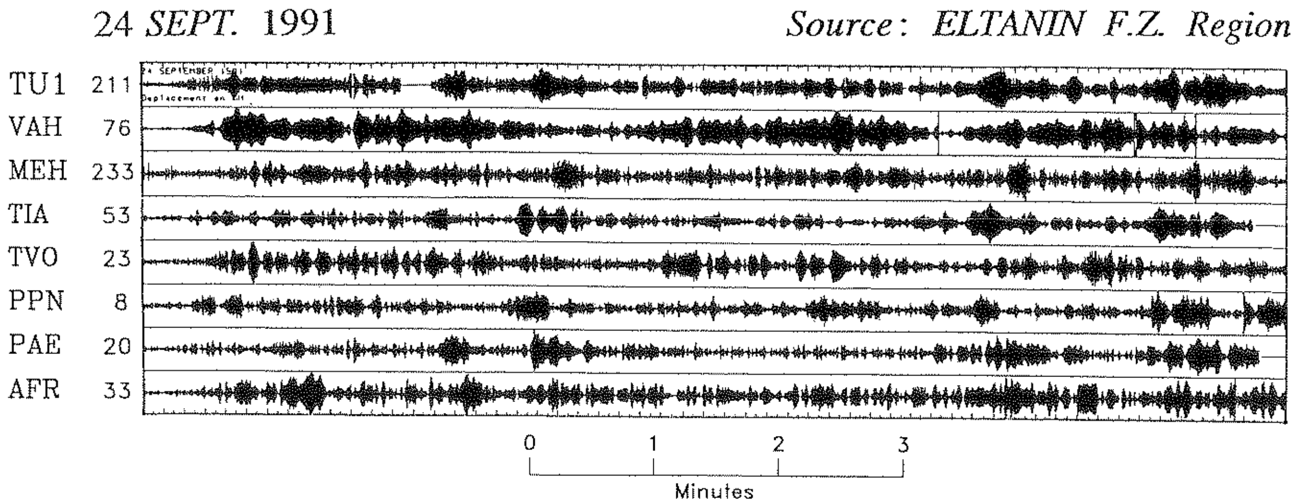


Figure 3. Example of Eltanin *T*-wave swarm, recorded at eight stations of RSP. Note continuous character of signal despite lack of correlation in waveform at the various stations. Time windows have been offset to account for differences in travel times.

(1) short signals, lasting typically from a few seconds to 20 sec—these are usually perfectly monochromatic but occasionally exhibit a slow decay of frequency toward the tail of the signal; (2) intermediate signals, typically lasting on the order of 40 sec; and (3) long-duration signals, which extend typically over several minutes and occasionally up to a half-hour (Fig. 3). Although the amplitude of the signal fluctuates, it remains considerably above noise level.

The most intriguing characteristic of the Eltanin *T*-wave records is their spectral content, with the remarkable monochromatism of the RSP records presented in Figure 5: Their spectrum is composed of a single line, in the frequency range 2 to 15 Hz. As detailed on the sonogram in Figure 6, the value of the resonant frequency fluctuates slightly and erratically with time. The amplitude of the fluctuation is on the order of 10% and can be detected in time series shifted by as little as 10 sec.

Attempts to model the shape of the spectral line around its angular frequency ω_0 with a resonance curve of the type

$$R(\omega) = \frac{2\alpha}{\alpha^2 + (\omega - \omega_0)^2}$$

have resulted in quality factors $Q = \omega_0/2\alpha$ on the order of 10 to 50.

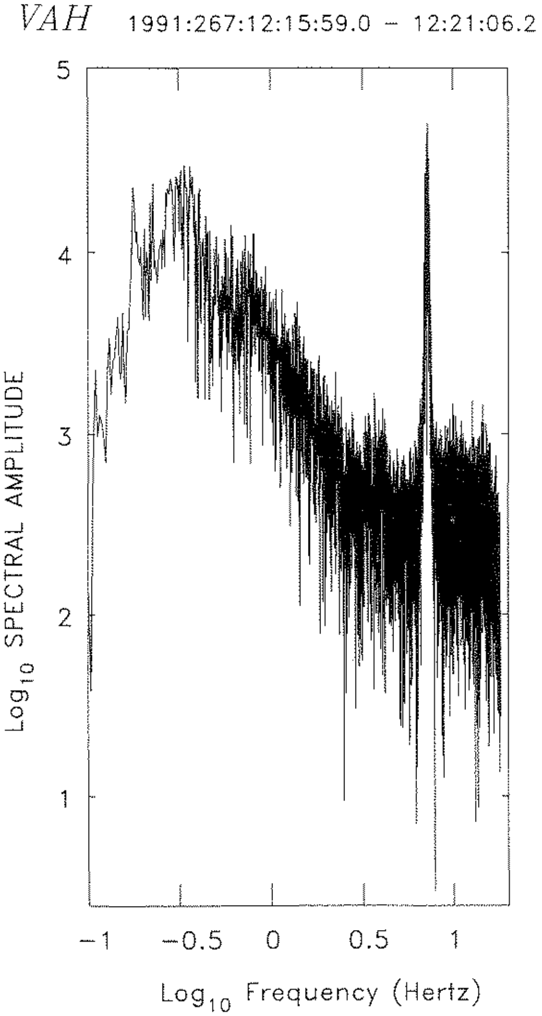


Figure 4. Spectral amplitude of ground velocity at VAH for 5-mn window of *T*-wave signal on 24 September 1991. Note spectral line with ≈40 dB signal/noise ratio.

At the frequencies involved (10 Hz and below), estimates of the anelastic attenuation in seawater (Thorp, 1965) would suggest a loss of only about 0.4 dB along the path to Polynesia, and thus the width of the spectral line is entirely controlled at the source, rather than on the path of propagation, as is the case for traditional seismic waves. Such relatively low values of Q then suggest that the oscillator is not stable for more than a few dozen periods, after which its period adjusts itself (or, conceivably, it could be replaced by another oscillator with a slightly different period). This result confirms the frequency shifts observed over time intervals of 10 sec or more.

Figure 7 shows the variation of the frequency of the spectral line for a 28-mn window as recorded at several Polynesian stations. A remarkable correlation exists between the fluctuation at the various stations, and this observation has two consequences: First, it confirms that the frequency fluctuation is a source, rather than a path or receiver, effect; second, it can be used to refine the differences in travel time to the various stations, through a simple set of intercorrelations of their time series.

Location

The source of the signals was then located by inverting the differences in travel times to the various RSP stations, using the seasonally adjusted LEVITUS database (Levitus *et al.,* 1994).

The resulting epicenter for the Eltanin swarm is found at 54° S, 140° W (Fig. 8), with the best epicentral value for the most intense period of activity (24 September 1991) being 53.87° S, 140.07° W. An error ellipse was obtained by a Monte Carlo method (Wysession *et al.,* 1991), consisting of injecting random noise with Gaussian distribution and a standard deviation $\sigma = 1$ sec into the travel times (the latter value obtained from an estimate of the precision of time picks) and then running the location algorithm using a large number (500) of such altered datasets. The shape of the resulting error ellipses clearly indicates an excellent control on the backazimuth from Polynesia but relatively poorer control on distance. The variation in epicenter during the swarm (illustrated in the case of three events on Fig. 8) is generally insignificant given the error ellipses.

This location is approximately 150 km away (measured along the direction of spreading) from the ridge segment between the Eltanin and Udintsev transform faults. This general area was until recently very poorly charted but exhibits some intriguing bathymetry: As shown on Fig. 8, a seamount (hereafter "location *S*") is present in the SYNBAPS map at 53.9° S, 140.3° W. It tops at 870 m b.s.l. on the SYNBAPS map, which uses a 5′ × 5′ grid, but original soundings from the 1965 Eltanin cruise report a depth of only 260 m. Furthermore, recent shipboard exploration of the area by Natland *et al.* (1995) and (motivated by our detection of the swarm) during a campaign of *N/O L'Atalante* in early 1996, has recognized a 450-km-long bathymetric feature, named the Hollister ridge, and extending from location *S* to the mid-oceanic ridge (MOR); a sounding of 135 m b.s.l. was obtained by *L'Atalante* at 53.94° S, 139.95° W. In addition, Seamount *S* was found to be an integrated part of the Hollister ridge, rather than a discrete bathymetric feature; samples dredged from the area included fresh vesicular basalts (L. Géli, personal comm., 1996).

While in principle, and given the shape of its error ellipse, the epicenter of the swarm could be located at other, uncharted, seamounts, or could have an abyssal location, it is extremely tempting to associate it with the Hollister ridge and more precisely, with its northwestern extremity, mapped as location *S* on Figure 8. At any rate, the epicenter is certainly located no more than 200 km from the center of the Udintsev–Eltanin MOR segment.

That segment of the Antarctic–Pacific ridge is remarkable in that it is one of only two, from Scott Island in the south all the way to the Galápagos triple junction in the north, showing up prominently in the geoid: Figure 9 shows

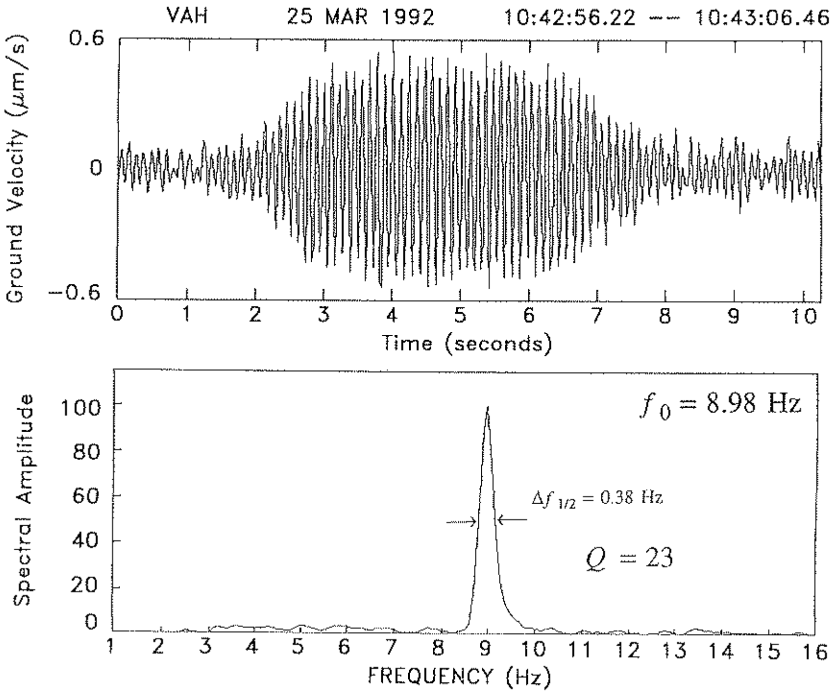


Figure 5. Time series (top) and spectrum (bottom) of a 10-sec window of *T*-wave signal at VAH on 25 March 1992. Note monochromatic character of signal.

a positive geoid anomaly reaching 1.35 m centered on 55° S, 138° W; the only other such anomaly (reaching only 0.95 m amplitude) is located immediately north of the Eltanin system (56° S, 122° W); the rest of the ridge is perfectly compensated, and its geoid signature is everywhere less than 25 cm. This geoid anomaly encompasses location *S*, and the Hollister ridge appears prominently in Figure 9 as the linear "finger" of geoid high pointing NNW from the MOR segment. This suggests that significant upwelling is present on the MOR, with a probable magmatic system leaking away into the Hollister Ridge, in a geometry reminiscent of the observations of Batiza and Vanko (1984) and Shen *et al.* (1993) in the northeastern Pacific. This magmatic system could also be related to the Louisville hotspot, even though most kinematic models put its present location slightly north, rather than south, of the Eltanin fracture zone system (e.g., Henderson, 1985).

The seismicity of this section of the Pacific–Antarctic ridge system is presented in Figure 10. Careful relocation was performed using the algorithm of Wysession *et al.* (1991) for all events initially located between longitudes 140.5° W and 135° W (the stippled band on Fig. 10). The resulting epicenters (with uncertainties generally on the order of a few tens of kilometers) show a significant seismic gap between the southernmost segment of the Eltanin system (offset about 25 km southwest of the Tharp fracture zone) and the short transform segment at 56.5° S, just north of the Udintsev fracture zone; note, incidentally, the anomalous focal mechanism (normal faulting on 17 September 1982) at the extreme western end of the Eltanin system, which may represent a local stress field.

In particular, an event listed by the NEIC on 11 July 1960 exactly at location *S* (54° S, 140.5° W) results from a flagrant typographical error: It relocates at 54.10° S, 140.63° E, along the George V fracture zone, south of Australia, a location also given by the BCIS. In addition, an event (22 May 1962) located at the center of the MOR segment relocates in the Udintsev transform system. We conclude that the source of the *T*-wave swarms is located in an area of aseismic character, at least at the level of the teleseismic detection threshold, estimated to be $m_b = 4.3$ in that part of the world.

Finally, it is worth noting that the soundings obtained at location *S* confirm our previous observation that a shallow water column at the source is required to generate magmatic *T* waves from underwater volcanic edifices: In Talandier and Okal (1987), we distinguished between "magmatic" *T* waves, featuring sustained amplitudes and believed to originate from actual extrusive phenomena on the ocean floor, and "seismic" *T* waves, generated by genuine earthquakes resulting from the country rock failing under overpressure during magma buildup. We went on to report that the observation of the former type required shallow water columns at the source, with volcanoes located at greater water depths remaining silent in this respect. We proposed a depth threshold of 780 m, a figure quoted from Staudigel and Schminke (1984), based on the maximum depth of recognition of explosive underwater volcanism, and emphasized that the entire southern and eastern segments of the mid-Pacific rise were silent for magmatic *T* waves, despite adequate detection capabilities at RSP for the past 25 yr. In the case of the Eltanin *T* waves, our rule of thumb is upheld by the presence at location *S* of a structure topping at 135 m b.s.l., as reported by *L'Atalante,* and which could be an efficient generator of

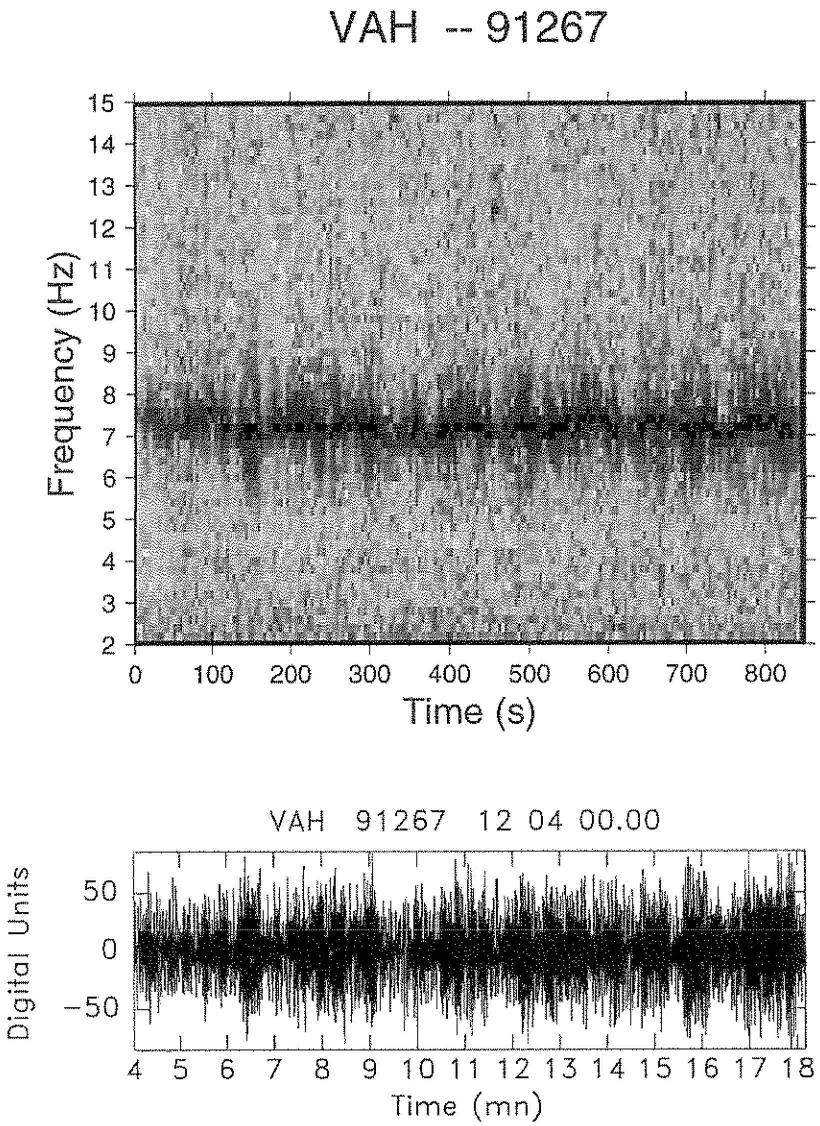


Figure 6. Sonogram of Eltanin swarm recorded at VAH (the corresponding time series is plotted at the bottom), showing the spectral amplitude of the signal contained in a moving window of 3-sec duration, as a function of frequency. The range between black (maximum amplitude) and lightest gray (minimum amplitude) corresponds to 50 dB.

T waves, especially since the SOFAR channel is believed to be particularly shallow (300 to 400 m) at high southern latitudes (Levitus *et al.,* 1994).

Data from Other Sensors

In addition to RSP records, we obtained hydrophone records from the SOSUS Pacific array (Fox *et al.,* 1995) (see Fig. 1). These data were recently partially declassified by the U.S. Navy and made available to the scientific community. However, the precise location of the sensors remains classified, and therefore, the SOSUS data cannot be combined with RSP data in order to improve on the epicentral location of the swarm. A location obtained exclusively from SOSUS data (52 °S, 140 °W) was made available to us (C. Fox, personal comm., 1994) and agrees well with the RSP-based location in the general area of the Eltanin fracture zone.

The SOSUS dataset is of great importance to our study for two reasons: First, it provides the opportunity to check the general characteristics of the spectra against those from the RSP, thus eliminating the possibility that the monochromatic character of the source could be path rather than source dependent; in addition, the hydrophone data, digitized at 250 samples/sec (128 samples/sec starting in 1993), allow the systematic exploration of the spectrum up to ≈80 Hz (40 Hz starting in 1993).

The SOSUS record shown in Figure 11 was selected to be offset about 80 mn from that in Figure 5, which should account for the difference in travel time with the path to RSP [based on the generally available low-grade estimate of the position of the SOSUS receiver (Fox *et al.,* 1995)], and thus should represent energy generated at the source at the same time for both arrays. This was confirmed by cross-correlating the frequency fluctuations in the Polynesian and SOSUS datasets. The spectrum (Fig. 12) is remarkably similar to that obtained at RSP. Furthermore, the spectrum remains fundamentally devoid of energy all the way up to 80 Hz. This

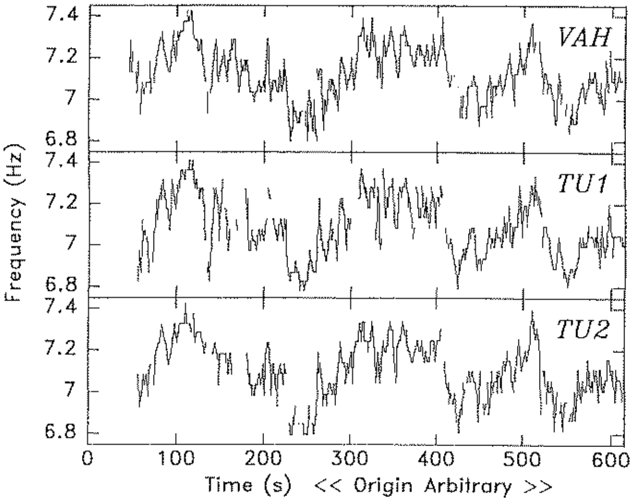


Figure 7. Evolution of the resonance frequency as a function of time for three stations of the RSP network. Absolute time has been shifted to allow direct comparison.

property was verified on many windows of SOSUS data and is also underscored in the sonogram presented on Figure 11.

The source of the Eltanin *T* waves can thus be described as a nearly perfect harmonic oscillator, featuring no observable nonlinearity (which would result in significant peaks at the multiples of the fundamental frequency). This is a very unusual property, especially when considering phenomena with significant amplitudes sustained for extended periods of time: For example, conventional seismic recording is well known to be plagued by microseismic activity generated by sea swell, around periods of 6 and 3 sec, which are prominent overtones of the fundamental period of the swell (12 sec).

The Socorro Swarms

We give here a brief description of the principal characteristics of other *T*-wave swarms recorded since 1993:

- *Socorro, January 1993.* On 19 January 1993, an underwater volcanic eruption took place in the immediate vicinity of Socorro Island (18.79° N, 111.75° W). This eruption was documented a few days later by aerial phenomena, such as floating scoria blocks and sea effervescence encountered by boats a few kilometers off the coast of Socorro (McClelland *et al.*, 1993). While *T* waves were received by RSP, their recording was limited to stations facing the northern and eastern shores of the islands, and the duration of the swarm was limited to a few days. *T* waves were also prominently recorded in Hawaii, and a joint location using both RSP and Oahu receivers was obtained at 18.75° N, 110.8° W (McCreery *et al.*, 1993), a few kilometers away from the island (open circle on Fig. 13). Figure 13 shows as + signs Monte Carlo locations ($\sigma = 2$ sec) computed using only Polynesian data (the higher value of σ used here is due to the lower signal amplitudes and to the less favorable geometry of those stations receiving Socorro *T* waves, as compared to Eltanin; this results in locations being achieved with three arrival times, as opposed to four). These locations are significantly offset to the northeast. The Socorro swarm has characteristics typical of previously observed volcanism in the Pacific Basin: It consists of individual bursts of activity, lasting a few seconds, and with a relatively broad spectrum in the 3- to 10-Hz range; the swarm was also detected by SOSUS, and its spectrum is described by Dziak *et al.* (1996).

  Socorro is an active island located at the intersection of the Mathematicians ridge and the Clarion fracture zone, in the Pacific plate, approximately 250 km southwest of the active plate boundary. The Mathematicians ridge is a fossil segment of the east Pacific rise deactivated about 3.5 Mya (Mammerickx and Klitgord, 1982). The activity on Socorro (and its three neighbors constituting the Revilla Gigedo archipelago) probably emanates from the remnants of the thermal anomaly associated with the fossil ridge. No significant geoid anomaly is present in the Socorro area, except immediately over the four islands (A. Cazenave, personal comm., 1995).
- *Socorro, May and June 1993.* A new period of activity was detected approximately 4 months later, lasting from 11 May to 3 June 1993. *T* waves from this swarm were remarkably monochromatic but generally of lower amplitude than those of the Eltanin swarm, and the duration of the individual sequences was shorter. The locations achieved from Polynesian data cannot be distinguished from those obtained for the January 1993 eruption (Fig. 13). The SOSUS records generally confirm the RSP observations in the 2- to 20-Hz range but exhibit a higher background level at higher frequencies.
- *North of Socorro, 19 November 1993.* Additionally, a short period of activity was detected at Socorro from 16 to 19 November 1993, with the strongest activity occurring on the 19th; in general, the amplitudes were relatively small and the duration of individual sequences short (on the order of 30 sec). In this case, the spectrum consisted of a strong spectral peak at 10 Hz, which was, however, superimposed on a broad background.
- Finally, a short puff of activity (lasting no more than 200 sec) was detected by stations of the Society (Tahiti) subarray on 20 November 1994. Although it featured monochromatic waves, it could not be located in the absence of records at other subarrays.

Discussion and Possible Interpretations

Henceforward, we restrict our study to the case of the Eltanin swarm, given its longer duration and the overwhelming quality of its signals; we will occasionally recall that similar characteristics were observed during the May and

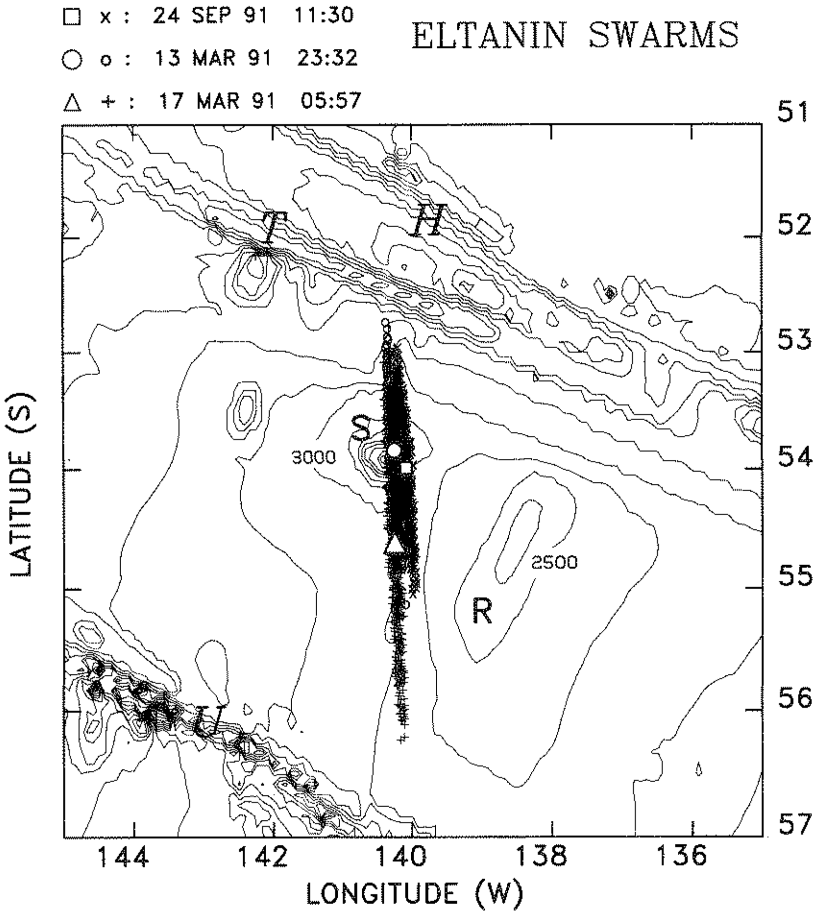


Figure 8. Epicentral area of the Eltanin swarm. Isobaths are plotted from the SYNBAPS dataset, at 250-m interval. *U*, *H*, and *T* identify the Udintsev transform and the Heezen and Tharp segments of the Eltanin fracture zone system, respectively. The Hollister ridge (Natland *et al.*, 1995) runs approximately from Seamount *S* to location "*R*" on the ridge. The large open symbols represent the best epicenters for three episodes in the swarm. The small symbols are the results of the Monte Carlo relocations, with $\sigma = 1$ sec. Note that the 17 March event was located using only three receiving stations, as opposed to four for the other two events.

June 1993 episode at Socorro. In trying to interpret the origin of the swarms as volcanic events, we must first discard other possibilities. We keep in mind that we are motivated by the exceptionally monochromatic character of the Eltanin *T* waves, which had no counterparts among all the volcanic events located by RSP over a period of 25 yr; as such, any interpretation, however tentative, must clearly rely upon a significantly singular phenomenon.

Biological Origin? Cetaceans, principally whales, are known to generate significant low-frequency acoustic signals into the SOFAR channel (Busnel and Dziedzic, 1965). We believe, however, that the present signals were not generated by whales for the following reasons: (1) their frequency band (typically 3 to 12 Hz, with the great majority around 7 to 8 Hz) is well below that characteristic of whales (above 15 Hz); (2) the remarkably monochromatic character of our signals differs from whale-generated *T* waves, whose frequency evolves with time more rapidly than observed here, thus making probable the exchange of information between animals; (3) any biological signal would be expected to have a seasonal character, reflecting the migratory patterns of the species in question, as well as its annual biorhythm, probably related to reproductive cycles; on the contrary, the Eltanin swarm lasted 15 months, spanning an entire seasonal cycle, including two southern autumns; (4) similarly, biological activity would be expected to recur more or less regularly on a yearly basis, due to the annual migration of the species; no signals similar to the Eltanin *T* waves were ever observed before 1991, even though the detection capabilities of the RSP were basically unchanged since 1967. We conclude that the monochromatic *T* waves recorded by the RSP do not have a biological origin.

Artificial (Human) Origin? We must similarly evaluate the possibility of a human origin to our signals, which might be suggested by two of their characteristics: their near perfect monochromatism and their isolated occurrence in 1991 and 1992. However, we regard this hypothesis as extremely unlikely for the following reasons: (1) The combination of low frequencies, high energy levels, slight fluctuations in frequency, and the 15-month duration of the experiment are all in sharp contrast with the procedures used in known experiments in ocean acoustics, e.g., the Heard project (Baggeroer and Munk, 1992), which uses higher frequencies (52 to 61 Hz) that are not only easier to generate but also better channeled by SOFAR, as well as sources with comparatively lower amplitudes and perfectly defined frequencies. (2) The duration of the Eltanin swarm would require the stationing of a ship at the source of the acoustic signals for an extended

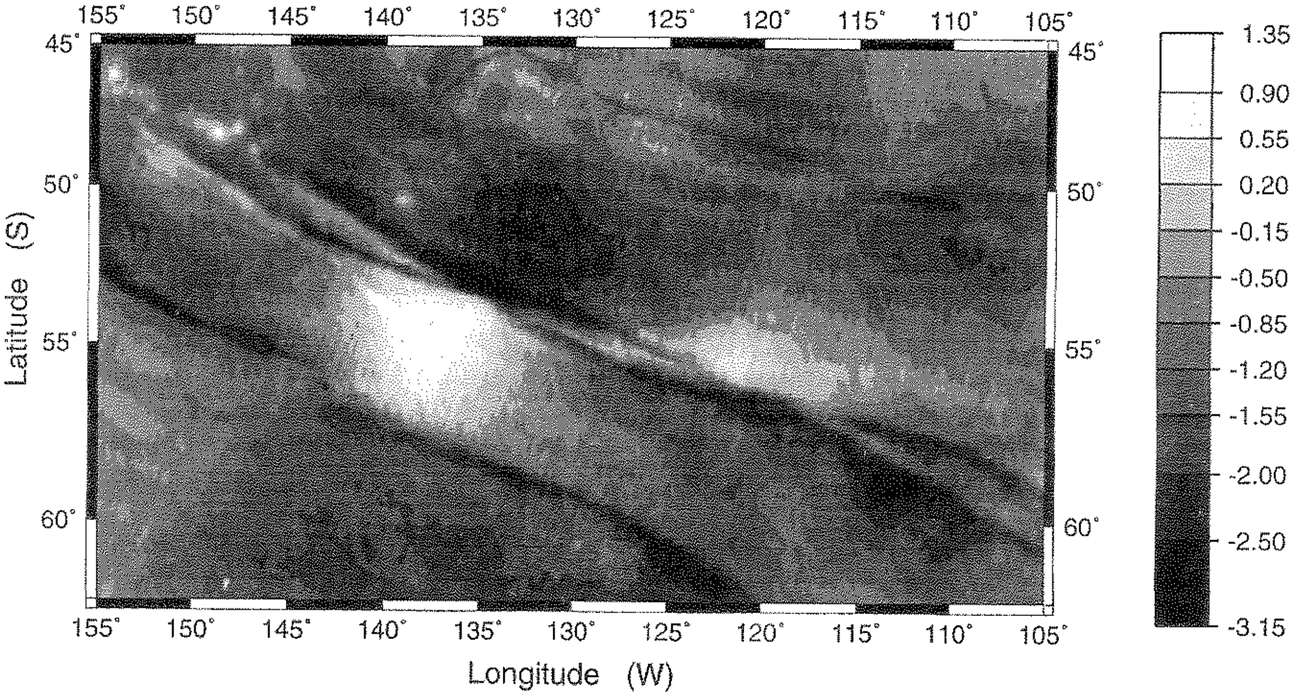


Figure 9. Map of the ERS-1 geoid (filtered for wavelengths $\Lambda \leqq 3000$ km) over the Eltanin transform system. Heights (in meters) given by vertical bar. Note significant bulge on the MOR segment between the Eltanin and Udintsev transforms and signature of Hollister ridge extending from MOR to Seamount *S* (Geoid data courtesy of A. Cazenave).

period of time (notwithstanding the generally inhospitable character of the climate in those Southern latitudes), which was not detected by surveillance satellites during the relevant time window (French Navy, personal comm. to JT, 1993). (3) Similar phenomena were recorded in 1993 from Socorro Island, where the prolonged stationing of a vessel would undoubtedly have been documented.

In view of (1) the striking correlation of the epicenter of the Eltanin swarm with the bathymetric high at location *S* and (2) the temporal correlation of the Socorro swarm with a volcanic eruption documented in the same area just a few weeks earlier, we conclude that the monochromatic *T* waves recorded by RSP were in both instances generated by underwater volcanism, the latter being taken in the broadest possible sense, including possible magmatic eruptions as well as hydrothermal activity.

Comparison with *T* Waves from Other Volcanoseismic Sources

Figure 14 documents examples of *T* waves received by RSP during several phases of eruption of Macdonald Seamount, as well as from an underwater site in the northern Marianas. When compared to the present episode at Eltanin, the difference in spectra is obvious. Even in the sustained, magmatic episodes at Macdonald, where the spectrum is relatively simple (Fig. 14f), the spectral peak never exhibits a purity comparable to its counterpart at Eltanin; the explosive events have a rich, nearly white, spectrum (Fig. 14a). Similarly, the Eltanin spectra differ fundamentally from those recorded during the eruption of the CoAxial segment of the Juan de Fuca ridge in June 1993 [see Fig. 2 of Schreiner *et al.* (1995)] or during the Blanco Transform swarm of 1994 (Dziak *et al.,* 1996). The source of the former was indeed described as individual *earthquakes* triggered in the country rock by the eruption, as opposed to being a direct magmatic process. The latter were characterized by a much broader spectrum covering the 2- to 25-Hz range.

Similarly, well-documented examples of volcanic tremor from subaerial volcanoes—recorded, for example, during so-called long-period volcanic earthquakes at Kilauea, Hawaii—have spectra most often dominated by a strong line (hence the expression "harmonic tremor") but containing, nevertheless, significant overtone energy [see Fig. 45.33 of Chouet *et al.* (1987) for Kilauea or Fig. 12 of Ferrazzini and Aki (1992) for Pu'u O'o]. The challenging nature of the Eltanin spectra resides in the complete absence of overtone energy. To our knowledge, the only comparable reports of monochromatism of approaching quality during a volcanic event are given at La Fossa (Vulcano, Italy) by Montalto (1994) (see his Figs. 3a and 7b), which he attributes to the resonance of fumarolic ducts, an idea first proposed by Anderson (1978), and at Merapi by Schick (1992), which he attributes to degassing processes in the magma. La Fossa's resonance is, however, at higher frequency (12 Hz),

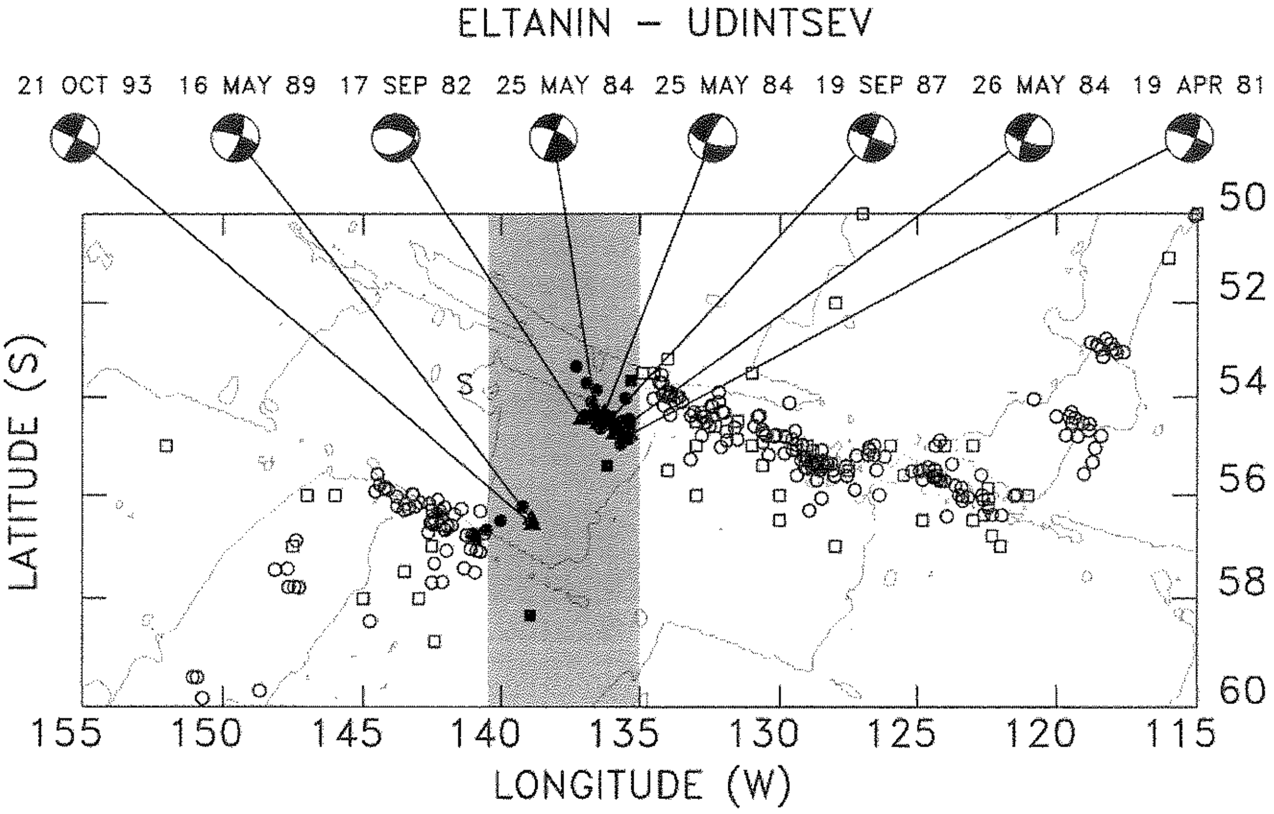


Figure 10. Seismicity of the Eltanin–Udintsev region of the Pacific–Antarctic ridge. The events initially located in the stippled band of longitudes were subject to relocations, and the resulting epicenters are shown as solid symbols. Open symbols are used for unrelocated epicenters outside the stippled zone. Pre-1964 earthquakes are shown by squares; modern ones, by circles. Harvard CMT solutions (only within the stippled zone) are shown by triangles and their mechanisms reproduced above the map. Note that the area of presumed upwelling identified by the geoid anomaly (Fig. 9) is fundamentally aseismic.

and modest overtones are indeed present in its spectrum; Merapi's is at much lower frequencies (1.4 Hz), and significant peaks are present at much shorter periods.

Theoretical Models

Seismic signals emanating from active volcanoes have been the subject of intense study ever since the dawn of instrumental seismology (e.g., Omori, 1912). More recently, Sassa (1935) and Minakami (1974) presented detailed classifications of the various kinds of volcanic earthquakes. Following the pioneering study by Aki *et al.* (1977), an extensive literature exists on the subject of theoretical models of the origin of volcanic tremor; we refer to Chouet (1996) for a comprehensive review of such efforts over the past 20 yr. Signals exhibiting or approaching a character of monochromatism have generally been ascribed to the resonance of a fluid-filled cavity or crack, under the effect of a pressure transient acting upon a part of its wall.

In an alternative model, Leet (1988) has proposed that the tremor energy could be generated by hydrothermal boiling within geyser conduits, an idea supported by Kieffer's (1984) correlation of seismicity and geyser activity at Yellowstone. Using a different approach, Julian (1994) has recently proposed to model tremor as a nonlinear vibration produced by fluid flow in an irregular channel feeding magma through country rock; for an adequate range of the driving pressure of the magma, the resulting spectrum is composed of a sharp peak and its harmonics. However, as already emphasized in the course of the present article, the signals modeled rarely achieve the degree of purity of the Eltanin spectra, expressed both by their relatively high value of $Q \approx 20$ to 50 [as compared with $Q \approx 5$ to 10, as reported by Aki *et al.* (1977)] and the total absence of overtones. These two observations will guide our discussion of the possible nature of the source of the Eltanin swarm.

As detailed by Chouet (1992), the fluid-filled crack model can explain the radiation of predominantly monochromatic seismic waves, with the resonance frequency being a complex function of the geometry of the cavity and of the mechanical properties of the two media (inside and out-

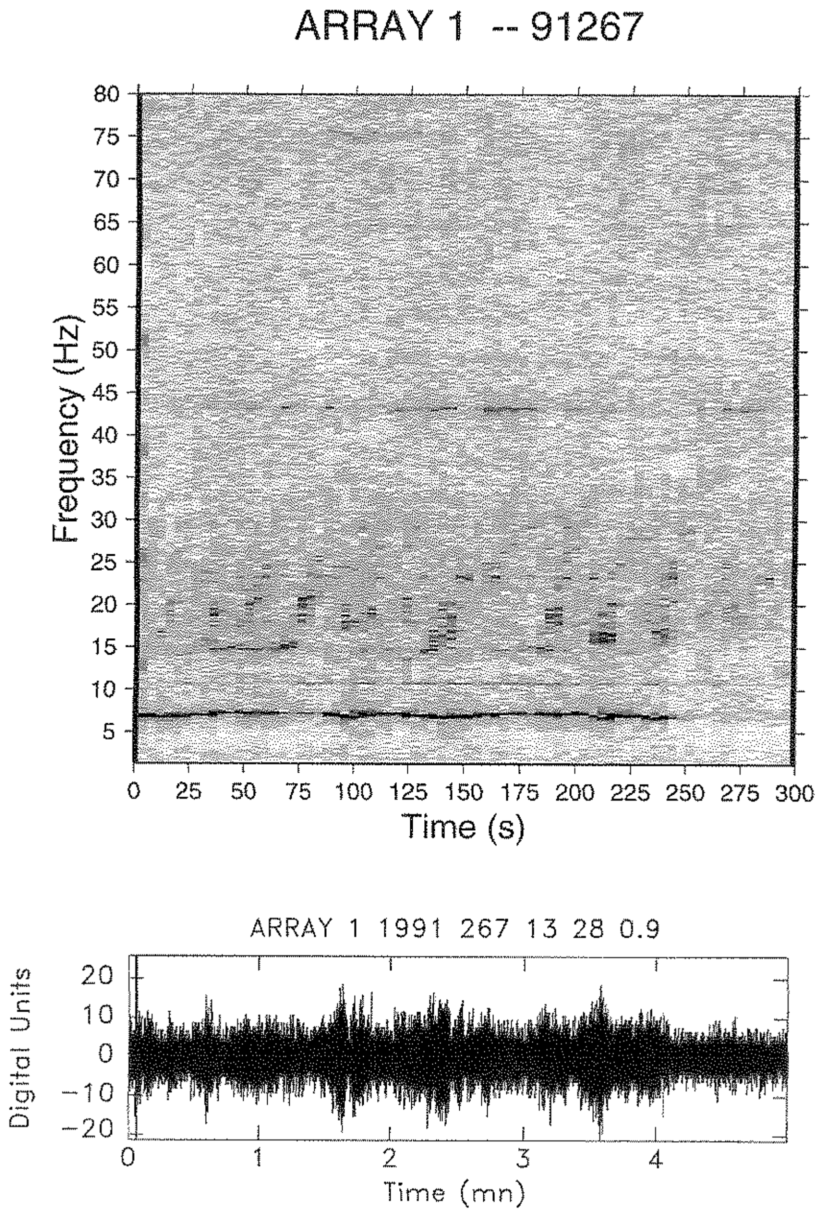


Figure 11. Same as Figure 6 for data recorded on hydrophones of SOSUS array 1. Note that the resonating signal ceases 4 mn into the window.

side the crack), controlled by two dimensionless quantities, the stiffness of the crack, and the impedance contrast at its boundary. Both of these parameters depend critically on the velocity of sound in the fluid filling the crack, which, in turn, can be strongly affected by the presence of impurities, most notably gas bubbles. As a result, "resonance at a long period may be possible in a short crack" (Chouet, 1996), and, indeed, Chouet *et al.* (1994) were able to model the spectra of tremors preceding the 1989 eruption of Redoubt Volcano in the 1- to 5-Hz frequency band by considering cracks with physical properties (including linear dimensions) generally consistent with the concept of intermittent choked flow of magma (Morrissey, 1994).

The Question of the Quality Factor

Regarding the relatively high-quality factor of the spectral lines at Eltanin, we further note that Chouet's (1992) model predicts a strong increase in Q with impedance contrast, which, in turn, argues for the presence of bubbles in the resonator, since they can drastically reduce a liquid's bulk modulus, even in minute concentrations (Kieffer, 1977). While no precise discussion is given of the quality factors of the spectral peaks in Julian's (1994) model, the large number of parameters in the model would probably allow the proper modeling of a quality factor in the range observed at Eltanin.

The Question of the Absence of Overtones

As compared to the Eltanin swarm, we note that most other spectra from volcanic long-period events or tremor (e.g., the 1989 Redoubt events mentioned above) are far from being truly monochromatic; they are generally and expectedly characterized by several resonance frequencies, in-

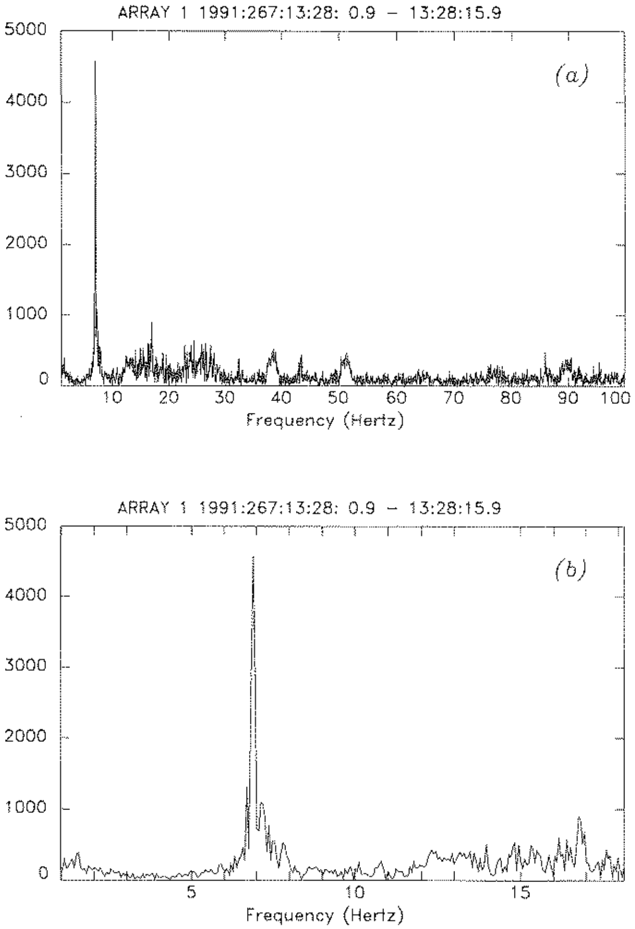


Figure 12. Spectrum of the initial 15-sec window of the SOSUS hydrophone record shown in Figure 11. (a) Full frequency range; (b) close-up in frequency range of RSP records.

cluding a more or less prominent fundamental mode, and by a family of overtones.

In Chouet's (1992) model, the relative amplitude of these modes is controlled by the specific excitation conditions, described as the location and duration of the pressure transient on the surface of the crack. While it is in principle possible to essentially eradicate the overtones from the spectra through an appropriate positioning of the source on the wall of the crack (B.A. Chouet, personal comm., 1995), this situation should remain to a large extent fortuitous, and it is difficult to propose that it would be repeated consistently, in view of the short- and long-term fluctuations in frequency of the Eltanin *T* waves, which indicate a continuous evolution in the parameters of the resonating crack at the source.

In Julian's (1994) model, the oscillation derives from the instability of the nonlinear system, which also results in period doubling, and thus, it is difficult to envision the disappearance of the overtones, although the latter are expected to be of somewhat reduced amplitude. Leet (1988) predicts the excitation of overtones but does not discuss their relative contributions to the spectrum.

In the gas-filled crack model of Anderson (1978), the periodicity of the tremor is a direct expression of the stepwise nature of the opening of the crack, which proceeds in a series of regular, assumably periodic increments. Under those conditions, the absence of overtones may be better explained, but the jerky nature of the development of the crack, which consists of individual ruptures (microearthquakes?) regularly spaced in time, should lead to significant power in the spectrum at higher frequencies, which is not observed in the SOSUS data (Fig. 12).

Another source of harmonic seismic energy described in the literature is the resonance of bodies (either columns or clouds) of bubbly liquids. In the case of volcanic edifices, the candidate bubbly fluids could be either magma or water. We refer to the works of Van Wijngaarden (1972) and more recently Lu *et al.* (1990) for a general description of the generation of sound by collective oscillations of gas bubbles in a liquid. Their main result, confirmed in laboratory experiments (Lu *et al.,* 1990; Yoon *et al.,* 1991), is that even minute fractions (less than 1% in volume) of gas can affect considerably the sound speed in the liquid, so that a bubble cloud (or an ascending jet of bubbly water) can exhibit a strong impedance contrast with the [pure] liquid matrix and resonate in a pattern very similar to that of a fluid-filled crack in a solid matrix. Laboratory experiments confirmed that resonance in the seismic band (say, at a few Hertz) could be explained by physical parameters (volume density of bubbles, size of cloud, etc.) adequate for a volcanic process (Chouet, 1996), in the case of both basalt or water.

As compared with the simple oscillation of a fluid-filled crack, a significant difference arises in bubbly liquids regarding the intrinsic attenuation of the oscillation in the resonating body, in which heat transfer becomes prominent at relatively low acoustic frequencies (Commander and Prosperetti, 1989); as a result, the acoustic wave may be attenuated significantly within the dimensions of the cloud, preventing it from being transmitted away from the source region at frequencies greater than approximately 10 Hz (Chouet, 1996). Such a mechanism, which basically acts as a low-pass filter at the source, could explain the systematic damping of the overtones and their absence from the teleseismic spectra, regardless of the exact value of the fluctuating resonance frequency. If we speculate that the cutoff frequency for eradication of the overtones by this process could be as high as 13 Hz and that, in all cases, the first overtone would have a frequency greater than that threshold, a resonating bubbly column of liquid could be an attractive model of our observations.

The liquid could be basalt, either in a duct or a magma chamber, and the gas bubbles could contain steam; at location *S*, the water depths to the volcanic edifice would not be prohibitive in this respect. An alternative model would be that of a resonating jet of venting bubbly water within the oceanic column itself, which would obviously result in an extremely efficient excitation of the water column and thus in *T* waves of very strong amplitudes.

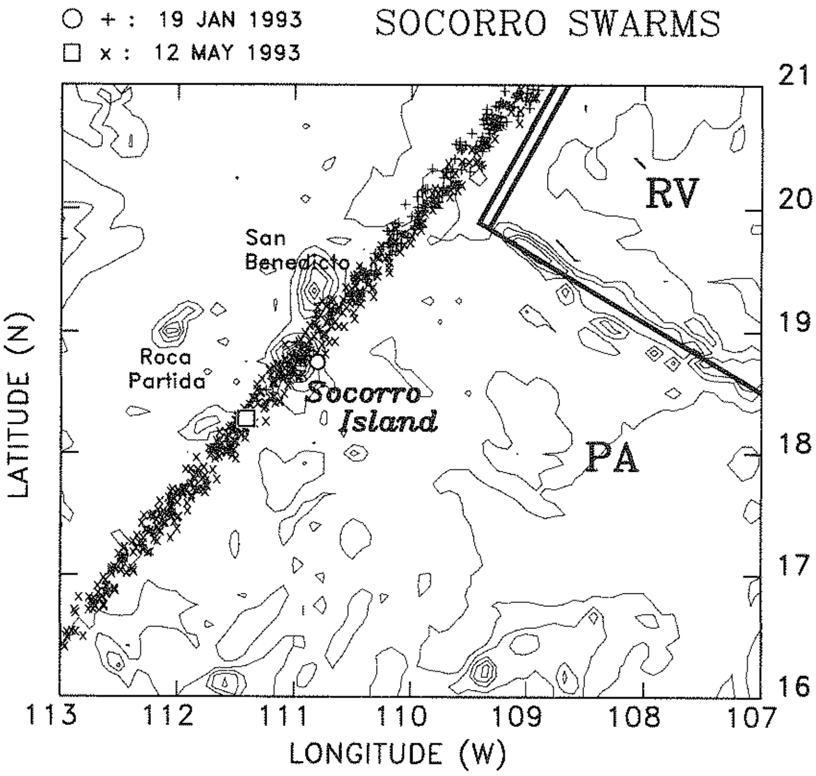


Figure 13. Same as Figure 8, for the Socorro swarms. Isobath interval is 500 m. The boundary separating the Pacific (PA) and Rivera (RV) plates is shown as the thick trace. The location (circle) for 19 January 1993 was obtained by combining Polynesian and Hawaiian data; all other locations use only Polynesian data. See text for details.

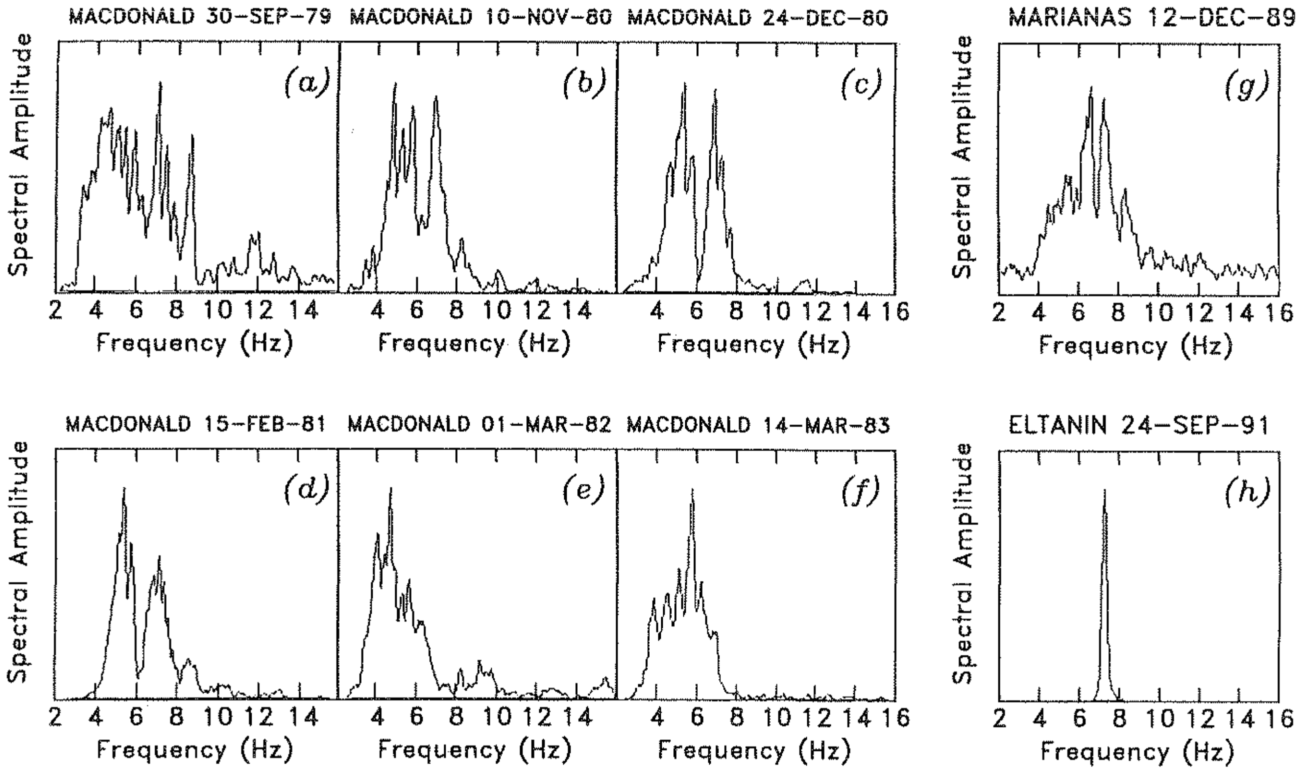


Figure 14. Comparison of the spectra of various underwater sources of volcanic *T* waves. (a) through (f) Various episodes of activity at Macdonald; (g) Northern Marianas; (h) Eltanin source.

In conclusion, it is of interest to note that both Montalto (1994) and Shick (1992) do require a gaseous phase (respectively, fumaroles and bubbles) to model the only other observations (to our knowledge) of *T* waves approaching the monochromatic character of the data described here.

Volcanological Speculations

Because of the fundamentally hidden nature of underwater volcanism, and especially in the case of the remote Eltanin site, we can only offer speculations as to the precise nature of our acoustic sources. An intriguing aspect of our observations is that monochromatic *T* waves have been observed only from two volcanic sites in the Pacific basin: the Eltanin site and Socorro. In seeking an explanation for such exceptional occurrences, it is interesting to note that both locations do indeed feature unusual characteristics for mid-oceanic ridge volcanoes:

- Regarding the Eltanin site, it sits at an exceptional location along the Pacific Ridge, both in terms of bathymetry (only 135 m b.s.l.) and geoid signature (the ridge being otherwise compensated), with probable leakage away from the MOR as expressed by the Hollister Ridge. Preliminary petrological results show a high degree of variability in major element composition (Natland *et al.,* 1995) along the Hollister ridge. Dredging in the immediate vicinity ($\approx$11 km) of our best epicenter has produced samples of fresh vesicular basalts, indicative of significant degassing (L. Géli, personal comm., 1996), which, in turn, would support the development of abundant clouds of gas bubbles. However, this process is relatively common, and this model fails to explain the absence of monochromatic *T* waves at other shallow submarine volcanic sites, where hydrothermal activity was documented or occasionally witnessed (Talandier *et al.,* 1988). Another possibility would be to simply assign the origin of our acoustic energy to hydrothermal activity, i.e., to the fountaining of "smokers"; however, these phenomena are ubiquitous on the MOR system and at other submarine volcanoes (Bougault *et al.,* 1993), which again fails to explain the singular character of our observations. Our preferred (but strongly speculative) scenario would envision the entrapment of seawater below a cooling lava lake of large dimensions (Juteau, 1993), in the general framework of Francheteau *et al.*'s (1979) observation of forests of basaltic pillars on the EPR at 21° N. The water would become vaporized at the ambient high temperatures (due to the presence of the lake) and low pressures (due to shallow water depths) and could then jet through the lake's surface in a series of geysers. This model would have the advantage of explaining the duration of the swarm (a lava lake can subsist for several years at the Earth's surface).
- Regarding Socorro, in addition to its position on the extinct Mathematician Ridge that does constitute an anomaly in itself, petrological studies (Bryan, 1966; Carballido-Sanchez 1994) have shown the presence of pyroclastic flows of peralkaline composition, generally not associated with the basaltic lavas characteristic of mid-oceanic processes. Among the models discussed to explain water depletion in the peralkaline domes and pyroclastic flows of Socorro, Carballido-Sanchez (1994) has suggested exsolution of a water-rich vapor phase during pyroclastic flow eruptions; such a process could be expected to result in significant generation of vapor bubbles, and we speculate that it could be the source of the monochromatic energy radiated into the ocean column.

Conclusions

1. We believe that a major eruptive process took place from March 1991 to June 1992 in the immediate vicinity (200 km at most) of the mid-oceanic ridge segment between the Eltanin and Udintsev fracture zone systems. Although deprived of earthquakes above the teleseismic threshold, this area is characterized by the largest bulge on the geoid along the east Pacific rise (EPR) and Pacific–Antarctic rise, suggesting that major upwelling takes place in the area. The presence of the Hollister ridge suggests the existence of a large volcanic province either leaked away from the MOR or possibly connected to the nearby Louisville hotspot.
2. *T* waves from this eruption exhibit an exceptionally monochromatic character with a single spectral line present in the frequency range 2 to 80 Hz. The resonance frequency does vary—from 4 to 12 Hz over the lifetime of the swarm—and can fluctuate typically 10% or a little under 1 Hz during episodes of activity as short as 10 to 20 sec. Very similar characteristics were recorded during a swarm at Revilla Gigedo Islands, which followed by 4 months an episode of documented volcanic eruption in the immediate vicinity of Socorro Island.
3. The most intriguing aspect of our observations is the absence of secondary peaks in the spectra, in particular when contrasted with the observation of many other volcanic sources during 25 yr of systematic monitoring in the Pacific basin. While it is possible to fine tune the model of a fluid-filled crack in a solid matrix, in order to minimize the excitation of overtones (through an adequate geometry of the pressure transient on its surface), we believe it unlikely that this situation could be regularly and consistently prevalent in the real world, in view of the significant variations in the crack geometry required by the observed fluctuations in frequency. We prefer the model of the oscillation of a bubbly liquid, in which enhanced attenuation through heat transfer at higher frequencies can essentially wipe overtones out of the spectrum, irrespective of the geometrical details of the oscillator. The origin of the gaseous phase could be the vaporization of seawater trapped under an active lava lake.
4. The duration—15 months—and level of sustained activity at the Eltanin location argue for a presumably very significant eruption in 1991 and 1992 at this intriguing site. The fresh basalts dredged by *L'Atalante* would support this

interpretation. We believe that further systematic shipboard and underwater exploration of the site is warranted.

Acknowledgments

We thank Louis Géli for preliminary data from the 1996 cruise of *L'Atalante* and Jean Francheteau and René Maury for discussion. Bernard Chouet sent us a preprint of his review chapter in advance of publication; Chris Fox and Bob Dziak made available the SOSUS records; and we thank René-Guy Busnel, Albin Dziedzic, and Gustave Alcuri for discussions on the choreography of whales. Anny Cazenave kindly provided us with her ERS-1 geoid dataset, from which Figure 9 was constructed. This research is supported by Commissariat à l'Energie Atomique (France). Early portions were also supported by the Office of Naval Research.

References

Aki, K., M. C. Fehler, and S. Das (1977). Source mechanism of volcanic tremor: fluid-driven crack models and their application to the 1963 Kilauea eruption, *J. Volcanol. Geotherm. Res.* **2,** 259–287.

Anderson, O. L. (1978). The role of magma vapors in volcanic tremors and rapid eruptions, *Bull. Volcan.* **41,** 341–353.

Baggeroer, A. and W. Munk (1992). The Heard Island feasibility test, *Phys. Today* **45,** no. 9, 22–30.

Batiza, R. and D. Vanko (1984). Petrology of Young Pacific Seamounts, *J. Geophys. Res.* **89,** 11235–11260.

Bougault, H., J.-L. Charlou, Y. Fouquet, P. Appriou, and P. Jean-Baptiste (1993). L'hydrothermalisme océanique, *Mém. Soc. Géol. France* **163,** 99–111.

Bryan, W. B. (1966). History and mechanism of eruption of soda-rhyolite and alkali basalt, Socorro Island, Mexico, *Bull. Volcan.* **29,** 453–480.

Busnel, R.-G. and A. Dziedzic (1965). Acoustic signals of the pilot whale *Globicephala melsena* and of the porpoises *Delphinus delphis* and *Phocoena phocena,* in *Whales, Dolphins and Porpoises,* K. Norris (Editor), Pergamon, New York, 607–646.

Carballido-Sanchez, E. A. (1994). The geology and petrology of Socorro Island, Revilla Gigedo Archipelago, Mexico, *Ph.D. Thesis,* Tulane University, New Orleans, 253 pp.

Chouet, B. A. (1992). A seismic model for the source of long-period events and harmonic tremor, in *Volcanic Seismology,* P. Gasparini, R. Scarpa, and K. Aki (Editors), Springer-Verlag, Berlin, 133–156.

Chouet, B. A. (1996). New methods and future trends in seismological volcanic monitoring, in *Manual of Volcano Monitoring,* in press.

Chouet, B. A., R. Y. Koyanagi, and K. Aki (1987). Origin of volcanic tremor in Hawaii; Part II: Theory and discussion, *U.S. Geol. Survey Profess. Pap.* **1350,** 1259–1280.

Chouet, B. A., R. A. Page, C. D. Stephens, J. C. Lahr, and J. A. Power (1994). Precursory swarms of long-period events at Redoubt Volcano, Alaska (1989–90): their origin and use as a forecasting tool, *J. Volcanol. Geotherm. Res.* **62,** 95–135.

Commander, K. W. and A. Prosperetti (1989). Linear pressure waves in bubbly liquids: comparisons between theory and experiments, *J. Acoust. Soc. Am.* **85,** 732–746.

Dietz, R. S. and M. J. Sheehy (1954). Transpacific detection of Myojin volcanic explosions by underwater sound, *Geol. Soc. Am. Bull.* **65,** 941–956.

Dziak, R. P., C. G. Fox, R. W. Embley, J. E. Lupton, G. Johnson, W. W. Chadwick, and R. A. Koski (1996). Detection of and response to a probable volcanogenic *T*-wave event swarm on the Western Blanco Transform Zone, *Geophys. Res. Lett.* **23,** 873–876.

Ferrazzini, V. and K. Aki (1992). Preliminary results from a field experiment on volcanic events at Kilauea using an array of digital seismographs, in *Volcanic Seismology,* P. Gasparini, R. Scarpa, and K. Aki (Editors), Springer-Verlag, Berlin, 168–189.

Fox, C. G. (1995). Special collection on the June 1993 volcanic eruption on the CoAxial segment, Juan de Fuca Ridge, *Geophys. Res. Lett.* **22,** 129–130.

Fox, C. G., W. E. Radford, R. P. Dziak, T.-K. Lau, H. Matsumoto, and A. E. Schreiner (1995). Acoustic detection of a seafloor spreading episode on the Juan de Fuca Ridge using military hydrophone arrays, *Geophys. Res. Lett.* **22,** 131–134.

Francheteau, J., T. Juteau, and C. Rangin (1979). Basaltic pillars in collapsed lava pools on the deep ocean floor, *Nature* **281,** 209–211.

Henderson, L. J. (1985). Motion of the Pacific Plate relative to the hotspots since the Jurassic and model of oceanic plateaus of the Farallon Plate, *Ph.D. Dissertation,* Northwestern University, Evanston, Illinois, 312 pp.

Julian, B. R. (1994). Volcanic tremor: non-linear excitation by fluid flow, *J. Geophys. Res.* **99,** 11859–11877.

Juteau, T. (1993). Le volcanisme des dorsales océaniques, *Mém. Soc. Géol. France* **163,** 81–98.

Kieffer, S. W. (1977). Sound speeds in liquid-gas mixtures: water-air and water-steam, *J. Geophys. Res.* **82,** 2895–2904.

Kieffer, S. W. (1984). Seismicity of Old Faithful Geyser: an isolated source of geothermal noise and possible analogue of volcanic seismicity, *J. Volcanol. Geotherm. Res.* **22,** 59–95.

Leet, R. C. (1988). Saturated and subcooled hydrothermal boiling in groundwater flow channels as a source of harmonic tremor, *J. Geophys. Res.* **93,** 4835–4849.

Levitus, S., T. P. Boyer, J. Antonov, R. Burgett, and M. E. Conkright (1994). *World Ocean Atlas 1994,* NOAA/NESDIS, Silver Spring, Maryland.

Lu, N.-Q., A. Prosperetti, and S. W. Yoon (1990). Underwater noise emissions from bubble clouds, *IEEE J. Ocean. Eng.* **15,** 275–281.

Mammerickx, J. and K. D. Klitgord (1982). Northern East Pacific Rise: Evolution from 25 m.y.b.p. to the present, *J. Geophys. Res.* **87,** 6751–6759.

McClelland, L., E. Venzke, and J. Goldberg (1993). *Bull. Global Volcan. Network Smiths. Instit.* **18,** no. 1, 9–12.

McCreery, C. S., D. A. Walker, and J. Talandier (1993). Hydroacoustics detect submarine volcanism, *EOS* **74,** 85–86.

Minakami, T. (1974). Seismology of volcanoes in Japan, in *Physical Volcanology,* L. Civetta, P. Gasparini, G. Luongo, and A. Rapolla (Editors), Elsevier, Amsterdam, 1–28.

Montalto, A. (1994). Seismic signals in geothermal areas of active volcanism: a case study from 'La Fossa', Vulcano (Italy), *Bull. Volcan.* **56,** 220–227.

Morrissey, M. (1994). Magmatic fluids and long-period seismicity: a geological and fluid dynamical perspective, *Ph.D. Thesis,* Arizona State University, Tempe.

Natland, J. H., P. R. Castillo, Y. Niu, S. H. Bloomer, and P. Lonsdale (1995). Dredging exploration of fast-spreading ridges in the far Southern Pacific Ocean: a preliminary report, *Ridge Events* **6,** 1–2 and 26–27.

Norris, R. A. and R. H. Johnson (1969). Submarine volcanic eruptions recently located in the Pacific by SOFAR hydrophones, *J. Geophys. Res.* **74,** 650–664.

Omori, F. (1912). The eruptions and earthquakes of the Asama-Yama, *Bull. Imp. Earthquake Inv. Comm. (Tokyo)* **6.**

Sassa, K. (1935). Volcanic micro-tremors and eruption-earthquakes, *Mem. Coll. Sci. Kyoto Univ. Ser. A* **18,** 255–293.

Schreiner, A. E., C. G. Fox, and R. P. Dziak (1995). Spectra and magnitudes of *T* waves from the 1993 earthquake swarm on the Juan de Fuca Ridge, *Geophys. Res. Lett.* **22,** 139–142.

Shen, Y., D. W. Forsyth, D. S. Scheirer, and K. C. Macdonald (1993). Two forms of volcanism: implication for mantle flow and off-axis crustal production on the West flank of the Southern East Pacific Rise, *J. Geophys. Res.* **98,** 17875–17889.

Schick, R. (1992). Volcanic tremor: seismic signals of (almost) unknown origin, in *Volcanic Seismology,* P. Gasparini, R. Scarpa, and K. Aki (Editors), Springer-Verlag, Berlin, 157–167.

Staudigel, H. and H.-U. Schminke (1984). The Pliocene seamount series of La Palma, Canary Islands, *J. Geophys. Res.* **89,** 11195–11215.

Talandier, J. (1989). Submarine volcanic activity: detection, monitoring and interpretation, *EOS* **70,** 561 and 568–569.

Talandier, J. (1993). French Polynesia Tsunami Warning Center (CPPT), *Natural Hazards* **7,** 237–256.

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

Talandier, J. and E. A. Okal (1979). Human perception of *T* waves: the June 22, 1977 Tonga earthquake felt on Tahiti, *Bull. Seism. Soc. Am.* **69,** 1475–1486.

Talandier, J. and E. A. Okal (1987). Seismic detection of underwater volcanism: the example of French Polynesia, *Pure Appl. Geophys.* **125,** 919–950.

Talandier, J., E. A. Okal, and H. Craig (1988). Seismic and *in situ* observations of Macdonald Seamount Eruption, 11 October 1987 (abstract), *EOS* **69,** 258–259.

Thorp, W. H. (1965). Deep-ocean sound attenuation in the sub- and low kilo-cycle per second region, *J. Acoust. Soc. Am.* **38,** 648–654.

Van Wijngaarden, L. (1972). One-dimensional flow of liquids containing small gas bubbles, *Annu. Rev. Fluid Mech.* **4,** 369–396.

Wysession, M. E., E. A. Okal, and K. L. Miller (1991). Intraplate seismicity of the Pacific Basin, 1913–1988, *Pure Appl. Geophys.* **135,** 261–359.

Yoon, S. W., L. A. Crum, A. Prosperetti, and N.-Q. Lu (1991). An investigation of the collective oscillations of a bubble cloud source, *J. Acoust. Soc. Am.* **89,** 700–706.

Laboratoire de Détection et Géophysique
Commisariat à l'Energie Atomique
91680 Bruyères-le-Châtel, France
(J.T.)

Department of Geological Sciences
Northwestern University
Evanston, Illinois 60208
(E.A.O.)

Manuscript received 19 October 1995.