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Converted *T* Phases Recorded on Hawaii from Polynesian Nuclear Tests: A Preliminary Report

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Abstract—We present a preliminary study of *T* waves from Polynesian nuclear tests at Mururoa, recorded on digital stations of the Hawaii Volcano Observatory network, following their conversion to seismic waves at the southern shore of the Island of Hawaii, and subsequent propagation to the recording stations. We show that seismograms are composed of several packets, which can be interpreted as resulting from $T \rightarrow P$ and $T \rightarrow S$ conversions, and which feature distinct spectral characteristics. As the distance from the shoreline to the station increases, the relative importance of the several wave packets changes; a prominent shadow for $T \rightarrow P$ is found at 8–12 km from the shore. This pattern is affected by the local crustal structure; in a favorable case, propagation in deep, low-attenuation layers resulted in a clear record as far as 76 km from the shoreline. While these results are generally robust, they can be moderately affected by a change of location of the source inside Mururoa Atoll.

Key words: *T* phases, seismic conversions, Nuclear Tests.

1. Introduction and Background

The purpose of this paper is to examine the seismic waves generated on the island of Hawaii (hereafter "the Big Island") upon reception of *T* waves generated by a number of nuclear tests at Mururoa, French Polynesia, over the years 1986–1991. Our goal is to understand the exact nature of the conversion of their energy into seismic waves, and of their subsequent propagation through the structure of the island. In the framework of the monitoring of the Comprehensive Test-Ban Treaty (CTBT), an improved understanding of these mechanisms of conversion and propagation will be necessary to adequately interpret any signals received at the so-called "*T*-wave stations" mandated by the treaty. In addition, this research may help optimize the deployment or relocation of *T*-wave stations in order to maximize their usefulness.

T waves are acoustic vibrations efficiently propagated in the SOFAR channel of the world's oceans. They were mentioned as early as 1930 (ANONYMOUS, 1930), recognized as teleseismic in nature independently by RAVET (1940) and LINEHAN

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(1940), and correctly described as guided waves following intense research during World War II (e.g., PEKERIS, 1948). The combination of an efficient waveguide and the practical absence of attenuation in the seawater at low acoustic (high seismic) frequencies (3–20 Hz) makes them a choice agent for the propagation (and eventual detection) of extremely small signals to extremely large distances in the marine environment.

Upon reaching the shore of an island (or continent), the acoustic energy in a *T* wave is converted to seismic energy, the resulting waves being then able to propagate in the island or continent structure (e.g., CANSI and BETHOUX, 1985; COOK and STEVENS, 1998), be recorded by standard seismographic stations, or even, when sufficiently powerful, felt by the local population (e.g., TALANDIER and OKAL, 1979). While this conversion is simply an illustration of the classic problem of reflection and refraction of elastic waves at a surface of discontinuity, it is made complex by a number of specific circumstances. Most importantly, the conversion is controlled by the exact geometry of the converting slope. In addition to being often unknown on the scale of the relevant wavelengths (e.g., 300 m at 5 Hz in the water), the latter is expected to depart from the simple model of a planar interface. In particular, bays and bights have been shown to provide focusing effects resulting in the preferential radiation of *T*-wave energy during the converse process of seismic-to-acoustic conversion near an earthquake source (OKAL and TALANDIER, 1997).

For this reason, acoustic-to-seismic conversion at island shores has been the subject of intense research, both theoretical (e.g., PISERCHIA *et al.*, 1998) and observational. Among the latter studies, KOYANAGI *et al.* (1995) have investigated the propagation across the Big Island of converted *T* phases to infer their attenuation characteristics. However, the earthquake sources which they used (including the 1989 Loma Prieta event) involved the additional complexity of seismic-to-acoustic source-side conversion. In addition, MCLAUGHLIN (1997) has compared seismic phases converted from Mururoa *T* waves and recorded at San Nicolas Island, off the Southern California coast, to hydrophone records of the same events obtained at Big Sur. His study suffers however, from the lateral distance between the two receiving sites, and from the extreme complexity of the continental shelf off Southern California. Finally, PASYANOS and ROMANOWICZ (1997) have documented the propagation of converted *T* phases from Mururoa events as guided surface waves, up to 200 km inside Northern California.

More recently, TALANDIER and OKAL (1998; hereafter Paper I) used a variety of recording environments in Polynesia (atolls, high islands) and of *T*-wave sources (Hawaiian earthquakes, chemical marine explosions) to study in detail the conversion of *T*-wave energy to and from seismic waves in a number of case scenarios. They concluded that steep segments of ocean floor in the immediate vicinity of a shoreline are efficient converters of *T* waves into seismic *P* waves. As the slope angle decreases, the *T* wave becomes post-critical for a simple refraction, resulting in the development of a shadow zone, and in preferred conversion to *S*, and possibly guided (Rayleigh-type) waves. These results, upheld by the finite-difference calculations of PISERCHIA *et al.* (1998), strongly suggest that the optimal location of a T-wave station may be in the immediate vicinity of a shoreline featuring a steep slope at the depths ($\sim$1200 m) characteristic of the axis of the SOFAR channel.

This geometry can be achieved by locating the station on an atoll, where coral walls are known to offer slopes as steep as 45° to 60° (GUILLE *et al.*, 1993), but it can also be found at the head of fresh basaltic flows, comparable to aerial "palis," and well documented in the offshore bathymetry of the southern flank of the Big Island (SMITH, 1994). In this general context, the present paper gives a preliminary study of the reception of converted phases by seismic stations of the Hawaii Volcano Observatory (HVO) network, following a number of nuclear explosions at Mururoa.

2. *Dataset*

Figure 1 shows the general layout of the present study. Its geometry features several significant advantages: first and foremost, the use of nuclear explosions at Mururoa provides a perfect case study in the framework of the monitoring of the CTBT. The location of the test site on a Polynesian atoll with steep reefs minimizes the complexity of the conversion process at the source (we do not consider here any

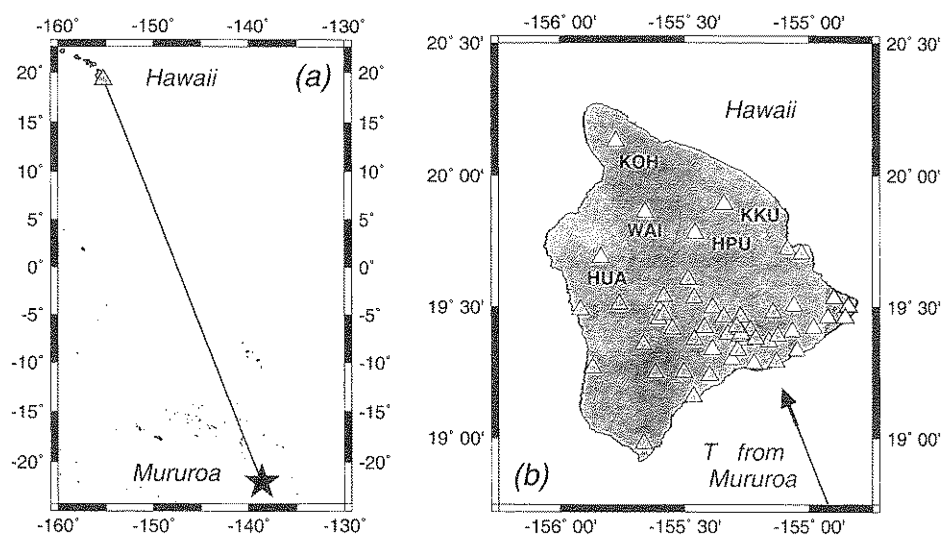

Figure 1

(a): General layout of the study. The star shows the Polynesian test site at Mururoa, and the triangle schematizes the receiving stations on the Big Island of Hawaii. *(b)*: Map of the HVO network on the Big Island. The open symbols are the stations referenced in the text (see also Table 2). Codes for southern shore stations are detailed on frame *(c)*. The large arrow identifies the azimuth of arrival of the T phase.

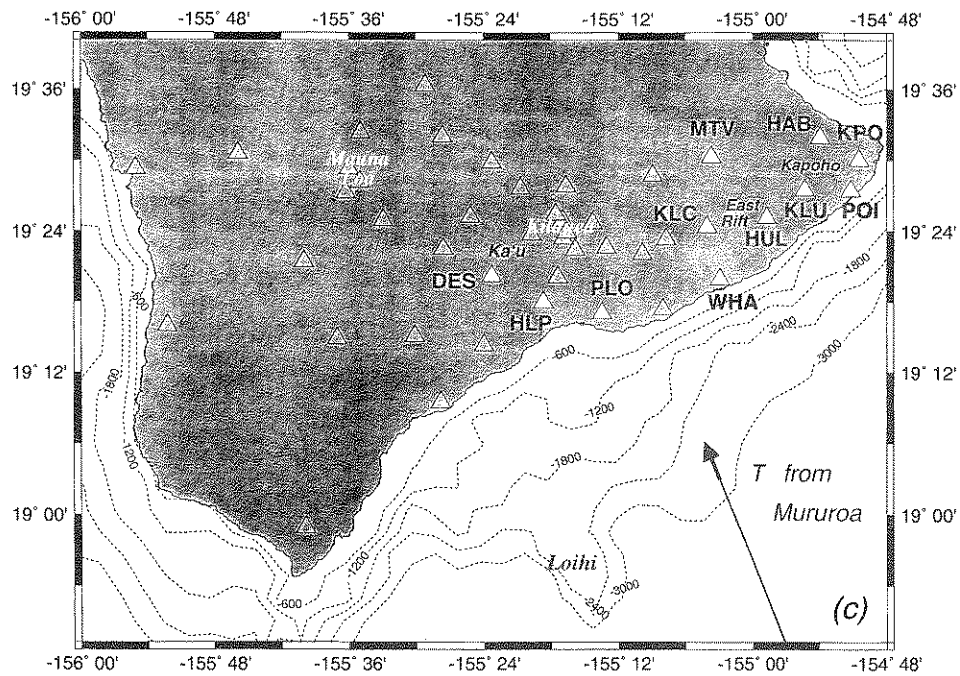

Figure 1*c*
Close-up of the southern shore of the Big Island, showing the stations used in this study (open triangles), and the bathymetry down to 3000 m. The 1200-m contour represents the axis of the SOFAR and is expected to control the conversion. Note the different underwater morphology off stations WHA and HLP.

tests from the southern site at Fangataufa, which entice a more complex generation process at the source). In addition, the shoreline of the Big Island is relatively regular, grossly perpendicular to the direction of arrival of the *T* wave, and most importantly was recently surveyed to a horizontal precision of 20 m (SMITH, 1994). Finally, HVO operates a unique, dense network of short-period seismic stations across the island, and especially on the southern flank of Kilauea Volcano, digitally recorded since 1986. Several of the stations are equipped with three-component seismometers.

While continuous digital recording is in principle available, *T*-wave signals have occasionally been strong enough to trigger the recording algorithm set for local Kilauea events, and are thus accessible at minimal effort from optical disks, with up to 80 channels of 100 sample per second data available.

Table 1 lists the 12 tests at Mururoa (out of a total of 28 for the period 1986–1996) which triggered local recording at HVO. Their epicentral coordinates and origin time are those listed by the National Earthquake Information Center (NEIC). In the present study, we assume that the origin time was at the exact minute, which suggests a systematic error of -1.7 ± 0.4 s in the NEIC source time. Table 2 gives a list of stations used in this study.

Table 1

Mururoa events which triggered HVO digital recording

Date D M (J) Y	NEIC Epicenter			Probable Origin Time GMT	Magnitudes		Number of stations	Peak-to-peak amp. at HUL (digital units)
	Latitude (°N)	Longitude (°E)	Origin Time GMT		m_b	M_S		
12 NOV (316) 1986	−21.911	−139.085	17:01:58.3	17:02:00	5.3		88	1776
5 MAY (125) 1987	−21.900	−139.096	16:57:57.7	16:58:00	4.9		14	835
20 MAY (140) 1987	−21.893	−138.964	17:04:58.3	17:05:00	5.6		181	2799
25 MAY (146) 1988	−21.903	−139.009	17:00:58.4	17:01:00	5.6	4.9	120	3880
23 NOV (328) 1988	−21.991	−138.907	17:00:58.0	17:01:00	5.4		99	2956
11 MAY (131) 1989	−21.853	−139.006	16:44:58.3	16:45:00	5.6		53	1021
3 JUN (154) 1989	−21.835	−138.996	17:29:58.5	17:30:00	5.3		70	1077
20 NOV (324) 1989	−21.851	−138.964	17:28:58.4	17:29:00	5.3		87	1583
2 JUN (153) 1990	−21.877	−138.918	17:29:58.5	17:30:00	5.3		96	739
4 JUL (185) 1990	−21.850	−139.042	17:59:58.7	18:00:00	5.1		69	691
18 MAY (138) 1991	−21.832	−139.014	17:14:58.5	17:15:00	5.1		71	955
15 JUL (196) 1991	−21.877	−138.963	18:09:58.3	18:10:00	5.3		119	3466

Table 2

HVO stations used in this study

Code	Station name	District	Latitude (°N)	Longitude (°E)	Distance to coast (km)
		T waves observed			
WHA	Waha'ula	East Rift	19.332	−155.049	< 1
HUL	Heiheiahulu	East Rift	19.419	−154.979	7
KLC	Kalalua Cone	East Rift	19.407	−155.068	9
MTV	Mountain View	East Rift	19.504	−155.063	18
POI	Pohoiki	Kapoho	19.457	−154.854	< 1
KLU	Pu'u Kali'u	Kapoho	19.458	−154.921	4
KPO	Kapoho	Kapoho	19.500	−154.842	4
HAB	Hawai'i Beaches	Kapoho	19.532	−154.898	12
POL	Poliokeawe	Halape–Ka'u	19.284	−155.225	2
HLP	Halape	Halape–Ka'u	19.299	−155.311	4
DES	Desert	Halape–Ka'u	19.337	−155.388	11
KKU	Keanakolu	Mauna Kea	19.890	−155.343	76
		No T wave recorded			
HPU	Hale Pohaku	Mauna Kea	19.781	−155.311	64
HUA	Hualala'i		19.688	−155.458	71
WAI	Waiki'i		19.860	−155.660	75
KOH	Kohala		20.128	−155.780	114

With the exception of Section 5, we now focus on records from the event of 20 May 1987 (origin time 17:05:00 GMT), which has both the largest body-wave magnitude ($m_b = 5.6$), and the largest number of readings reported to the NEIC. The available dataset consists of vertical seismograms at 48 stations, with horizontal channels at a subset of 12 sites; The resulting 72 triggered channels were extracted and converted to SAC format. Among the many stations available on the southern flank of the Big Island, we first study the record at Heiheiahulu (HUL). While a number of stations are located closer to the shoreline, HUL, being 7 km inland, benefits from reduced background noise (incidentally, it provided the most spectacular records of the 1994 Bolivian T waves (OKAL and TALANDIER, 1997)). This enhanced distance also allows separation of the various converted seismic phases. In addition, HUL is one of the stations most regularly included in triggered datasets, allowing direct comparison of the signals from several Mururoa sources. Finally, all three components of ground motion are available at HUL.

3. *The T Phase at HUL: Identification of Wave Trains*

The crustal structure of the Big Island has been the subject of numerous investigations (e.g., HILL and ZUCCA, 1987). In this study, we use HILL's (1969) model, featuring a 3-layer 11-km thick crust, and directly applicable to the southern flank of Kilauea where the conversion and short propagation to HUL take place. In the framework of Paper I, and using SMITH's (1994) bathymetric data, we then predict that the $T \rightarrow P$ conversion at HUL should be composed of three arrivals, at respectively 17:59:51.8, 17:59:52.8 and 17:59:54.0 GMT.

Figure 2 shows the arrival of the T wavetrain from Mururoa, as recorded on the digital triggered system; this is a 17-second close-up of the seismogram, after band-pass filtering between 1 and 10 Hz. Note that HVO operates on Hawaiian Standard Time (HST), which is 10 hours behind GMT; all times will be given after network trigger time (07:59:42.3 HST or 17:59:42.3 GMT). It is clear that the structure of the wave at HUL is in excellent agreement with that predicted: the T phase is clearly composed of a number of arrivals featuring different characteristics in terms of ground motion polarization and frequency content. The vertical component of the ground motion is dominated by a first high-frequency group of three arrivals, all matching the predicted times within 0.5 s, followed at 14.6 s by a larger, but lower-frequency, arrival. We interpret the former as $T \rightarrow P$ conversions (P_1, P_2, P_3), and the latter as $T \rightarrow S$. This interpretation is confirmed by spectrogram analysis, as shown on Figure 3, indicating energy maxima at 7.5 Hz for P_1, 7.0 Hz for P_2 and 6.5 Hz for P_3, but only 2.8 Hz for $T \rightarrow S$. The vertical seismogram at HUL also bears striking resemblance to the T phase shown on Figure 2 of Paper I, obtained in a reverse geometry (a Hawaiian earthquake recorded at station PMO in Polynesia). Paralleling the study in Paper I, we can characterize the difference in spectrum through the ratio

$$R_{SP} = \frac{X_{T\rightarrow P}(f_{T\rightarrow S})/X_{T\rightarrow S}(f_{T\rightarrow S})}{X_{T\rightarrow P}(f_{T\rightarrow P})/X_{T\rightarrow S}(f_{T\rightarrow P})} \tag{1}$$

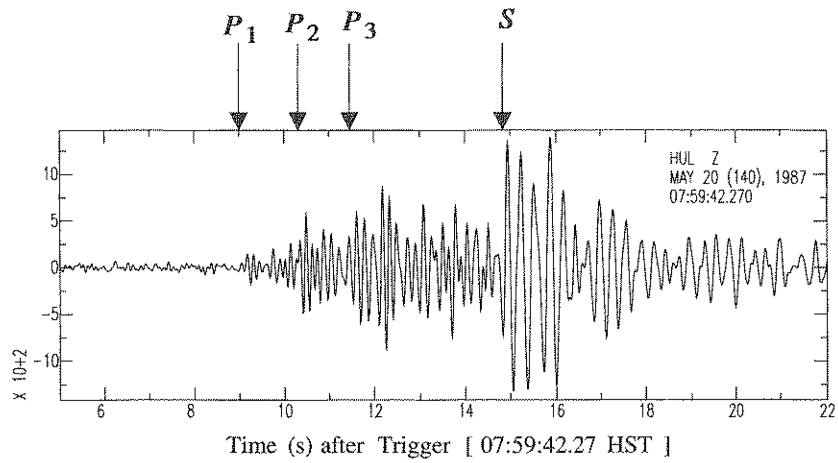

Figure 2
T-phase arrival at Heiheiahulu (HUL); this is a band-pass-filtered window ($1 \leq f \leq 10$ Hz) of the vertical seismogram. Note the several wave packets composing the $T \rightarrow P$ and $T \rightarrow S$ conversions.

of the relevant spectral amplitudes at the peak frequencies $f_{T \rightarrow P} = 7$ Hz and $f_{T \rightarrow S} = 2.8$ Hz of the two wave trains. When applied to two 2-s windows centered on the P_2 and $T \rightarrow S$ phases, $R = 0.025$, which if interpreted as the effect of anelastic attenuation along the path from the conversion point to HUL, yields $Q_\mu = 19$ (we assume here no bulk attenuation Q_K^{-1}). This value is in excellent agreement with that found in Paper 1 ($Q_\mu = 21$) for the reverse situation along a path similarly sampling the flank of Kilauea 15 km to the West, and also generally consistent with the value ($Q_\mu = 30$) proposed by KOYANAGI *et al.* (1995) in the immediate vicinity of the crater. Such values of Q would be regarded as very low in classical seismology, but are indeed documented in the particular setting of an active volcanic structure (AKI *et al.*, 1977; TALANDIER and OKAL, 1996).

Figure 4*a* shows the horizontal ground motion of the presumed $T \rightarrow S$ conversion, for the window comprised between 14.5 and 16.5 s after trigger time. The polarization of the wave train is in the azimuth N120°E. While this could correspond to SV polarization for a seismic S wave converted slightly off the great circle path (the backazimuth from HUL to Mururoa is N157°E), it could also stem from diffraction at a preferential conversion site such as an underwater bay or cove; further interpretation of the wave's polarization would be speculative at this point.

The later part of the horizontal ground motion is dominated (between 18 and 21 s following trigger time) by a wave of relatively low frequency (3 Hz). As shown on Figures 4*b* and 4*c*, this wave is polarized practically north–south and its vertical and NS components feature a $\pi/2$ phase offset. This is characteristic of a surface- or interface-wave; the timing of this arrival, suggestive of a group velocity from the conversion point to HUL of only 0.55 km/s, also supports this interpretation; the

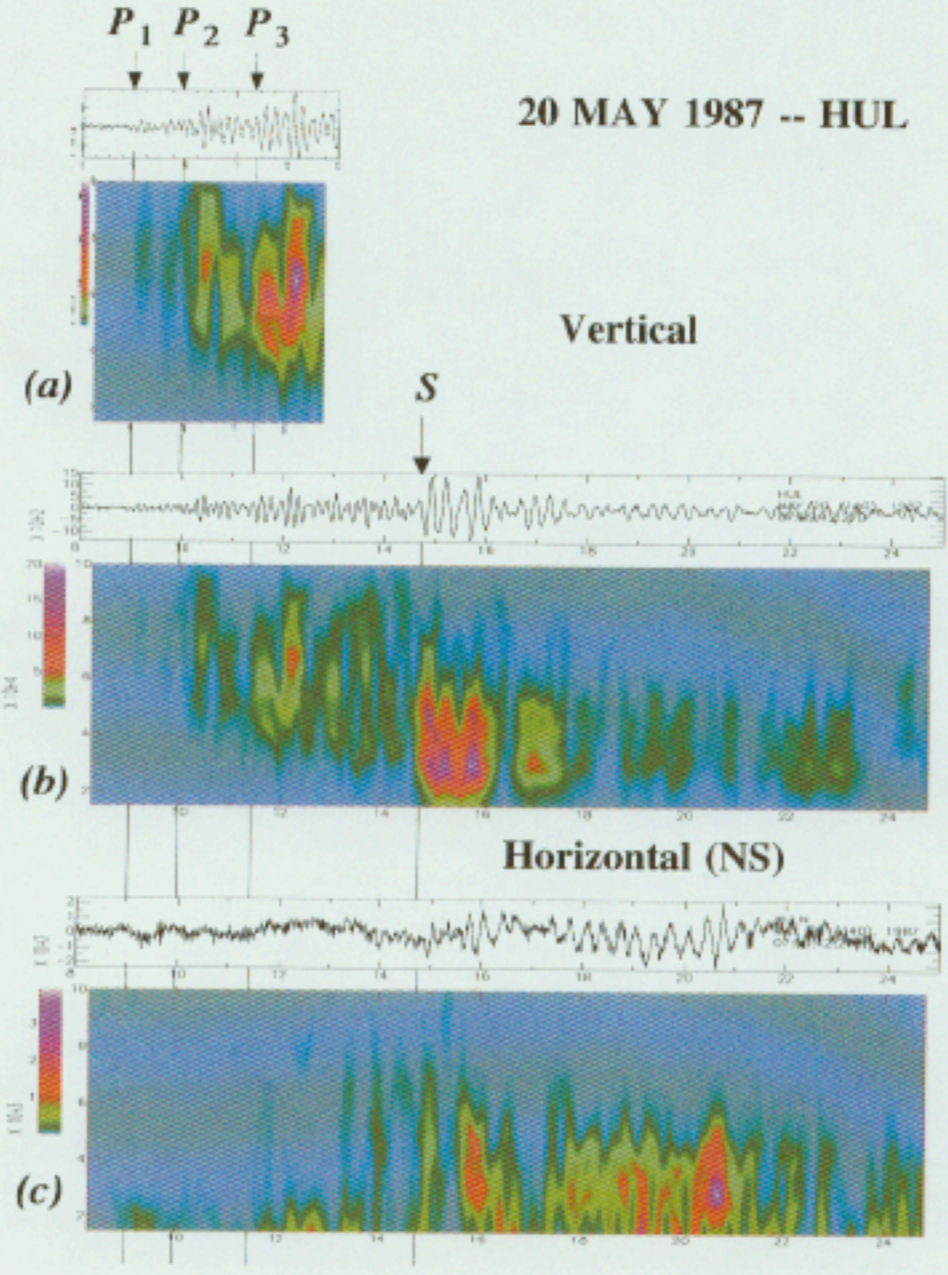

Figure 3

Spectrograms of the T phase recorded at HUL. *(a)* and *(b)*: Vertical component; *(c)*: Horizontal (north-south) component. *(a)* is a close-up of *(b)* detailing the frequency content of the $T \rightarrow P$ arrivals. All times are in seconds after trigger time. The spectrograms use a moving window of 2-s duration, sliding in increments of 0.2 s.

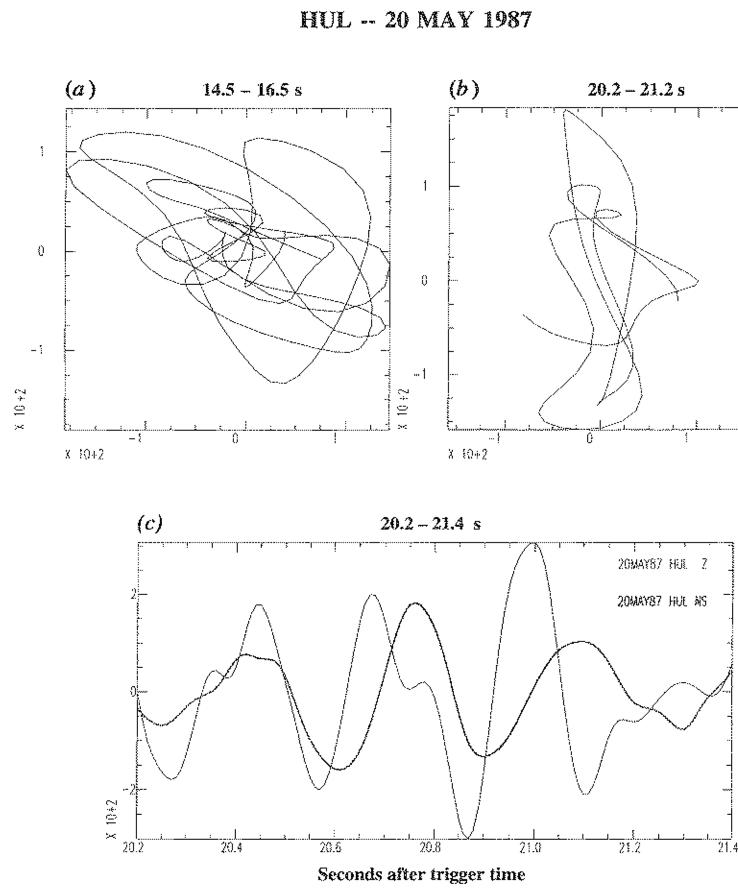

Figure 4

(a) and *(b)*: Particle motion in the horizontal plane for two windows of the later arrivals in the T wave. Note that the $T \rightarrow S$ conversion appears polarized in the N150°E direction *(a)*, while the later arrival *(b)* is polarized north-south. *(c)*: Comparison of the two components of the later arrival. The thin dark trace is the vertical one; the thicker, gray trace is the north-south one. Note that they are in quadrature of phase, a property characteristic of Rayleigh or interface waves (All these records have been band-pass filtered for $1 \leq f \leq 10$ Hz.).

generation of Rayleigh waves upon conversion was demonstrated theoretically by PISERCHIA *et al.* (1998).

4. Evolution across the Network

4.1. East Rift Stations

In this section, we present preliminary data regarding the evolution of the converted waves across the HVO network. We first concentrate on four stations straddling the central portion of the East Rift of Kilauea: WHA only a few hundred meters from the coastline, HUL (7 km away), KLC to the west of HUL along the rift zone, and 9 km away from the coastline, and MTV, 18 km inland. Figure 5 regroups spectrograms of the vertical components of ground motion at the four stations, for a common 20-s time window starting 5 s after trigger time.

As expected, Station WHA on the coast features a much shorter wave train than HUL. In addition, between WHA and HUL, the high-frequency components of the signal (5–8 Hz) are strongly attenuated, and the later arrival, interpreted as the $T \rightarrow S$ conversion becomes prominent at HUL. This reflects both the partition of the $T \rightarrow P$ energy into the several branches corresponding to the various crustal layers, and the resulting development of a shadow zone, as demonstrated in Paper I.

The situation at KLC is more complex; while the lower-frequency part of the wave train remains prominent, it is dispersed over several seconds. This suggests

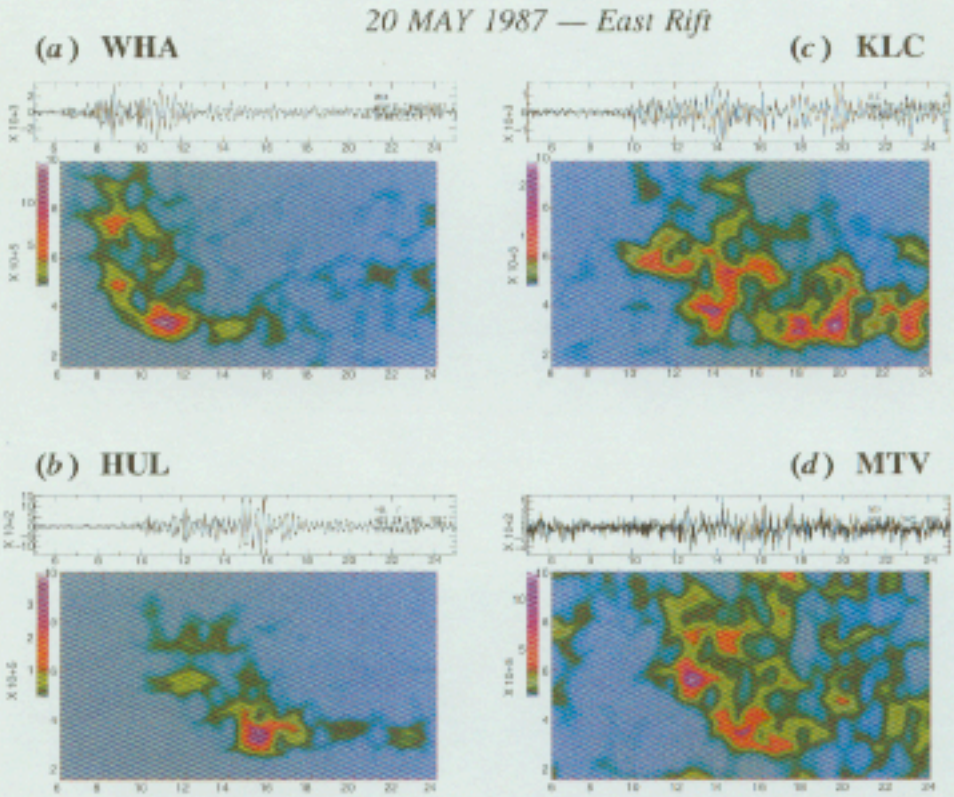

Figure 5
Comparison of spectrograms at the four East Rift stations for the event of 20 May 1987. All four windows sample between 5 and 25 s after trigger time.

multipathing to KLC, with conversions taking place at several points along the coastline. Note also that the $T \rightarrow P$ energy around 6 Hz is not damped as efficiently as it is at HUL.

Finally, when moving further inland to Station MTV, the most energetic part of the signal returns to the high-frequency (5.5 Hz) $T \rightarrow P$ conversion; this high-frequency wave train arrives only marginally later than at HUL, indicating refraction in a deeper, much faster, and probably less attenuating medium, which could be Layer 4 of the model of WARD and GREGERSEN (1973). On the other hand, the prominence of energy at high frequencies illustrates the strong damping of the later, lower-frequency, portion of the seismogram, supporting the interpretation that it corresponds to surface waves, propagating slowly across the shallow, presumably strongly fragmented and thus strongly attenuating, structures of the East Rift.

4.2. *Kapoho District*

We similarly study the case of the four stations located in the Kapoho District, at the easternmost end of the Big Island (Fig. 6). As in the case of WHA, the wave train at Station POI on the coast, is very short (no more than 5 s), but the $T \rightarrow S$ wave train is comparatively less developed, resulting in the strongest energy being carried by the third $T \rightarrow P$ wave packet. The wave trains at the East Rift Station KPO is perhaps most similar to HUL, with a group of high-frequency (6–7 Hz) $T \rightarrow P$ packets followed 3 s later by a low-frequency (2.5–3 Hz) $T \rightarrow S$ wave train. At KLU, located only 8 km from HUL along the East rift, the $T \rightarrow P$ wave packets are somewhat lower frequency (5 Hz rather than 7 Hz at HUL), while the $T \rightarrow S$ wave trains remain less attenuated and richer in higher frequencies (5 Hz). Finally, at HAB, beyond the rift, the distance to the coast (12 km) is insufficient to generate the fast, high-frequency phase previously observed at MTV, and the seismogram remains similar to that at KLU.

4.3. *The Halape Pali-Ka'u Desert Area*

We study here the conversion and propagation of the T phase in the region of the Halape Bight, located due south of Kilauea, using Stations PLO (2 km from the coast), HLP (4 km inland) and DES in the Ka'u Desert, 11 km inland (Fig. 7). With respect to other coastal stations, PLO features strongly attenuated $T \rightarrow P$ packets, with their energy remaining under 5 Hz. Similarly, HLP has all the characteristics of a shadow zone location (comparable to KLC), while DES features a revival of the $T \rightarrow P$ energy along the general lines of the MTV record, as shown in the close-up (Fig. 7*d*). The evolution of the seismogram across this area is generally similar to that along the East Rift, except for shorter distances to the coastline. This is easily explained by noting that the 1200-m isobath, along which the conversion is expected to take place, does not follow the bight, but rather develops a 10-km wide shelf, with the net result of an increase in the path of the various converted (seismic) waves.

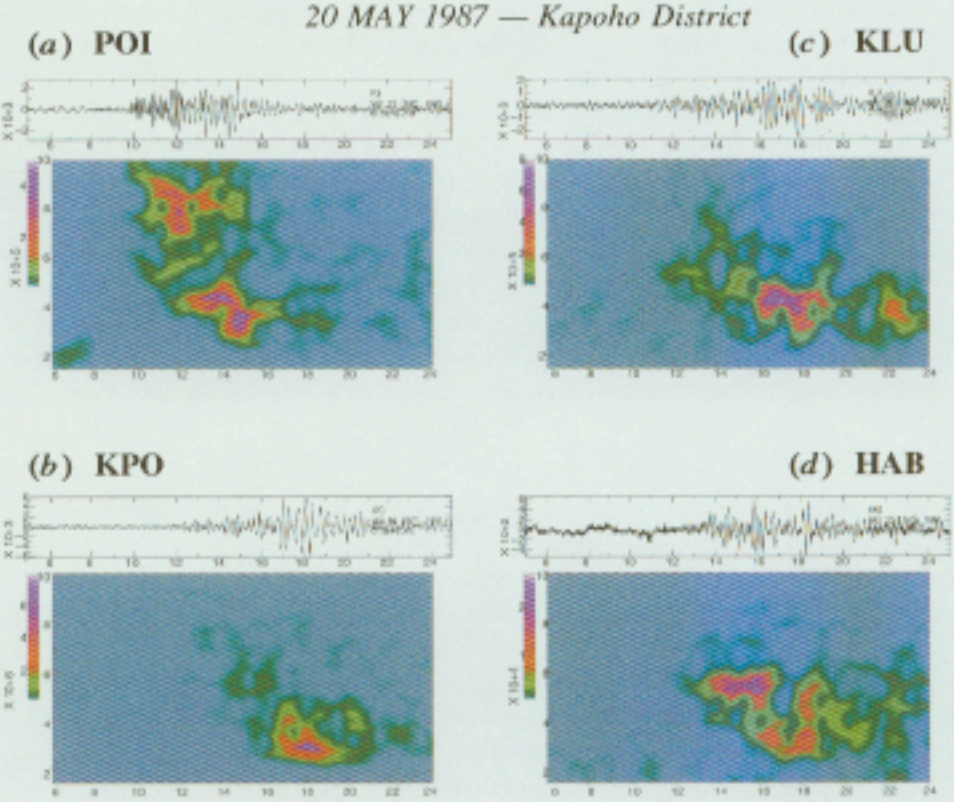

Figure 6
Comparison of spectrograms at the four stations of the Kapoho District for the event of 20 May 1987. All four windows sample between 5 and 25 s after trigger time.

4.4. Keanakolu (KKU)

This station, located on the northeastern flank of Mauna Kea, 76 km from the receiving shoreline, recorded a remarkable *T* wave train (Fig. 8). Despite the presence of background noise, spectrogram techniques easily pull out (i) a weak, high-frequency signal (5.5 Hz), 20.5 s after the trigger time; (ii) two lower-frequency (3.5 Hz) packets 25 and 28 s after trigger time, and (iii) another group at 42 and 46 s with comparable frequency content. In the absence of intermediate stations, it is difficult to identify the first arrival beyond doubt, but the travel-time difference between MTV and KKU would fit propagation in WARD and GREGERSEN's (1973) Layer 4, at 7.7 km/s.

We failed to identify the *T* phase at other distant stations of the HVO network, including Kohala (KOH; 114 km from the coast), Waikii (WAI; 75 km), Hualala'i (HUA; 71 km), and Hale Pohaku (HPU; 64 km). The recording at KKU may be

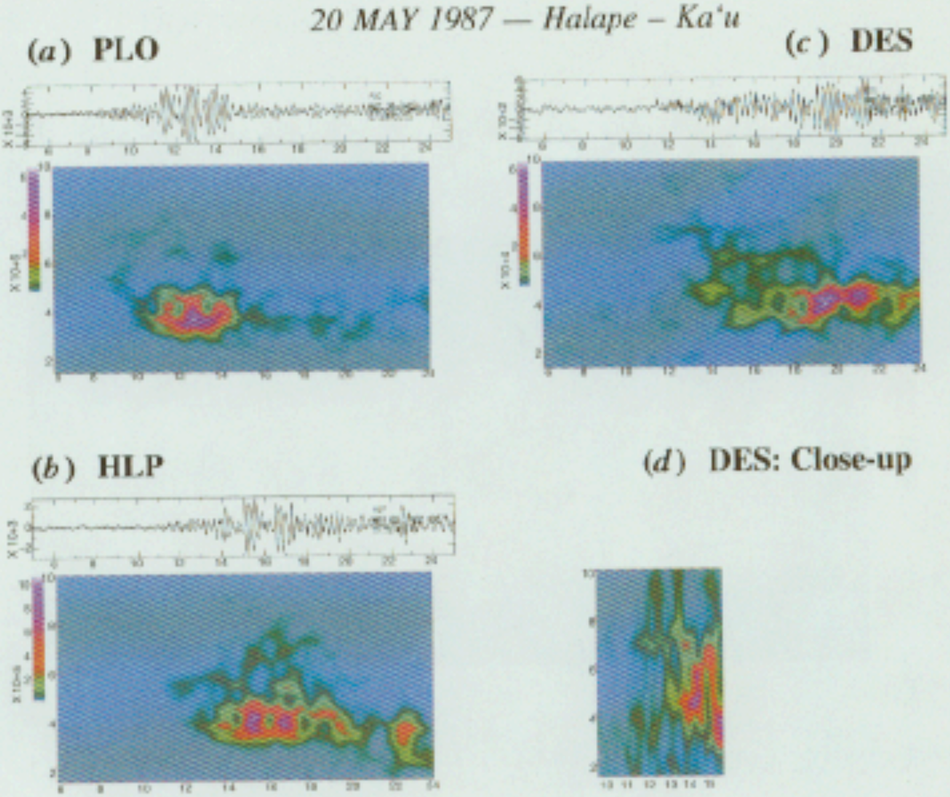

Figure 7

Comparison of spectrograms at the three stations of the Halape–Ka'u District for the event of 20 May 1987. *(a)*, *(b)*, and *(c)* sample between 5 and 25 s after trigger time. Frame *(d)* is a close-up of the seismogram at DES, similarly aligned in time, but emphasizing the re-emergence of the $P \rightarrow T$ conversion at that station, between 8 and 15 s after trigger time.

explained by the preferential location of the station, which high-frequency rays from the coast can reach by diving under the largely horizontal East Rift volcanic system, whereas seismic rays propagating to the other stations are bound to penetrate the deeply extending vertical magmatic systems under Kilauea and Mauna Loa.

5. Other Events

We briefly address here the question of the robustness of the wave shapes of *T* phases recorded at HUL from several sources at Mururoa. Figure 9 shows triggered records from all 12 events listed in Table 1, all plotted on a common vertical scale. The horizontal axes have been lagged to align all origin times (assumed to be on the

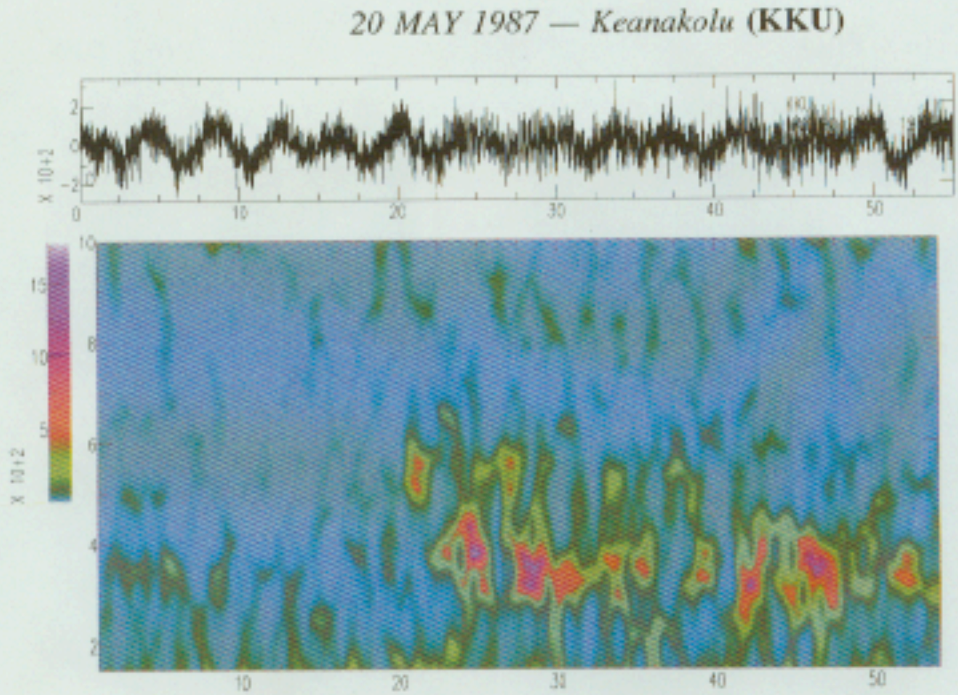

Figure 8
Spectrogram of the T phase at Keanakolu (KKU), 76 km from the shoreline. This window presents the whole triggered record, lasting 55 s.

exact minute). Several features are apparent on this figure. First, there is a confirmed relationship between the amplitude of the T wave and the size of the test, as measured by its reported m_b. While the correlation is mediocre (60%), the strongest T waves were observed from the events with the largest m_b (Fig. 10*a*). A similar correlation is found with the number of stations reporting to the NEIC (Fig. 10*b*). Concentrating then on the four events with the strongest signals (20 May 1987, 15 Jul. 1991, 25 May 1988 and 23 Nov. 1988), we find that the wave shapes of the first three are generally similar, together with their spectral contents (Fig. 11): the signals consist of a series of high-frequency (6–7 Hz) arrivals, that we interpreted as $T \rightarrow P$ conversions, followed by a lower-frequency, but generally higher amplitude $T \rightarrow S$ conversion. The fourth event, however, shows a more "blended" spectrum, in which the later arrival remains of comparatively high frequency (4–7 Hz). The origin of this feature, which is also present at WHA, is currently unclear.

Finally, the arrival times at HUL of the T phases from the 12 tests are shifted within a window of 8 s duration, which most probably reflects the precise location of the source inside the lagoon at Mururoa. In this respect, the NEIC locations are of little if any help, since many of them would plot outside the atoll.

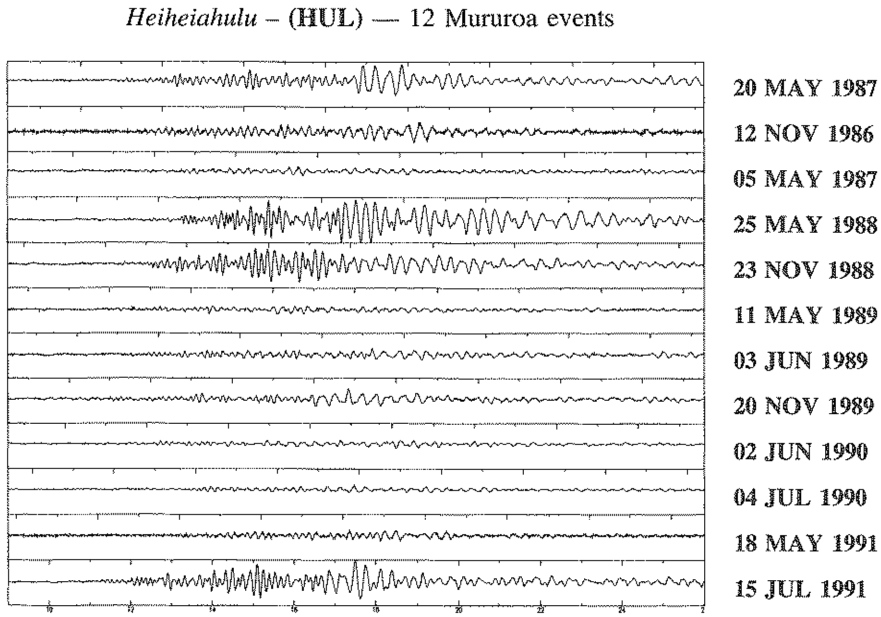

Figure 9
Comparison of the T wave trains from 12 Mururoa events (1986–1991) at Heiheiahulu. A common vertical scale is used, and the seismograms have been lagged in time to adopt a common origin time, under the assumption of a source detonated at the exact minute.

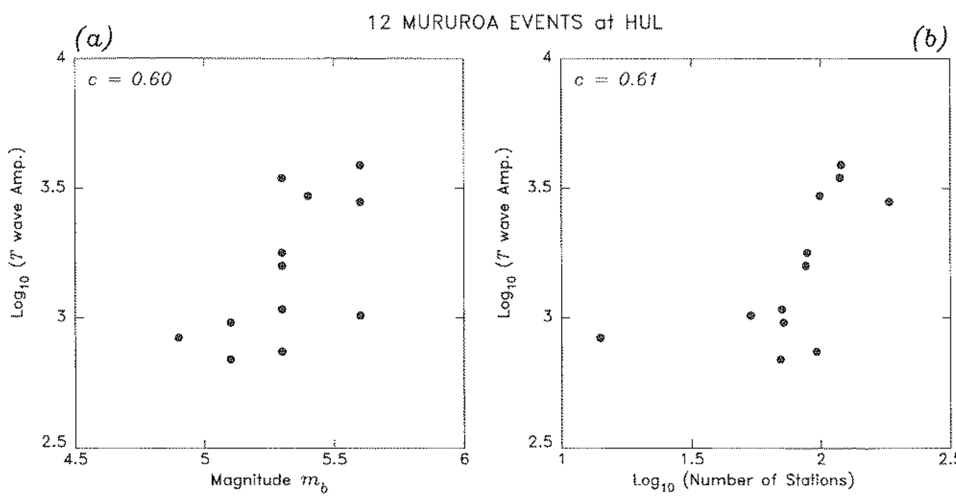

Figure 10
Peak-to-peak amplitude of the T wave trains recorded at HUL for the 12 Mururoa events listed in Table 1, plotted as a function of magnitude m_b *(a)* and number of stations reporting to the NEIC *(b)*. The numbers c at upper left are the correlation coefficients for the various datasets. Amplitude units are digital recording units at HUL.

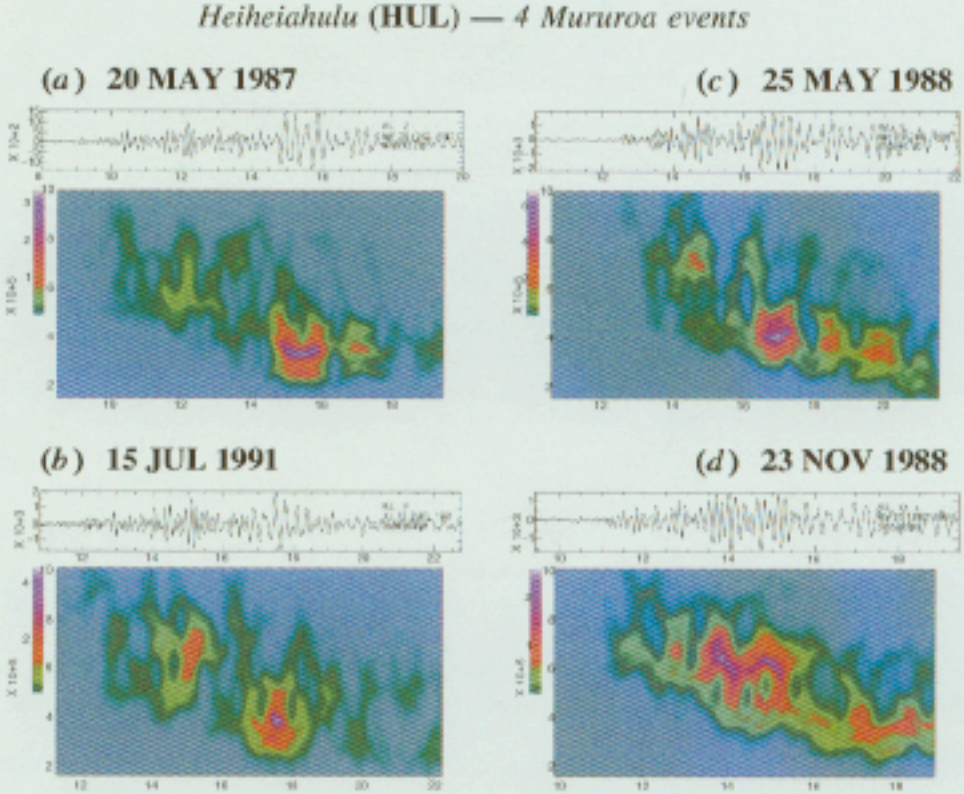

Figure 11
Comparison of spectrograms at HUL for the four largest Mururoa events. The time windows processed are 12 s long, and are lagged to align origin times (assumed to be on the exact minute). See text for interpretation.

6. *Conclusions*

We have examined a preliminary dataset of T phases from nuclear tests at Mururoa, recorded by the HVO network, primarily on the southern flank of the Big Island. The principal conclusions of this study are:

1. The example of the HUL record of the event of 20 May 1987 shows that the T wave is converted into a number of seismic waves featuring distinct and identifiable characteristics. At HUL, we observe separate packets of P and S energy, as well as a later, low-frequency, slow-propagating phase, interpreted as an interface, Rayleigh-type wave.
2. Propagation of the various seismic packets through the surficial structures of the island results in rapid attenuation of the high frequencies initially present at the conversion point. An estimate of the shear quality factor, $Q_\beta = 19$, is in line with the low values found in volcanic environments.

3. As the receiving site moves further inland, the characteristics of the various wave packets evolve rapidly. Most remarkable is the development of a shadow zone for the $T \rightarrow P$ conversion(s); at typical distances of 7 to 12 km, the seismogram is dominated by the $T \rightarrow S$ conversion, and consequently features a lower frequency spectrum. At larger distances, deeper refraction of P waves below the East Rift magmatic system becomes possible, with high frequencies re-emerging at MTV (18 km), and propagating efficiently as far as KKU (76 km from the shore).
4. The local crustal structure plays an important role in the exact evolution of the wave trains with distance away from the shoreline. While the general patterns are robust, the presence of a large shelf offshore can affect the position of the P shadow zone; furthermore, the magmatic systems under the Kilauea and Mauna Loa calderas essentially prevent the propagation of T phases to large distances across the island.
5. Results at HUL are generally robust for other Mururoa tests. Unknown differences in the location and condition of the shots under Mururoa Lagoon are probably the cause of arrival-time shifts of up to 8 s, and of moderate differences in the spectrum of the acoustic wave in the ocean, eventually resulting in wave packets ($T \rightarrow P$, $T \rightarrow S$) with differing frequency contents following propagation and attenuation of the converted wave in the island structure. The case of the event of 23 November 1988, featuring later arrivals with higher frequency content, remains intriguing, and will be the subject of further research.

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