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HUMAN PERCEPTION OF *T* WAVES: THE JUNE 22, 1977 TONGA EARTHQUAKE FELT ON TAHITI

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ABSTRACT

We report human perception of seismic waves generated on the island of Tahiti by exceptionally large *T* waves, following a major earthquake in the Tonga area. The generation mechanism of the *T* waves and their smaller amplitudes at other Pacific stations are explained in relation to source and receiver bathymetry, focal mechanism, and magnitude of the event. It is concluded that a series of favorable path conditions, rather than the size of the event (M_S = 7.8; M_W = 8.1), were responsible for the giant *T* waves felt at Tahiti.

INTRODUCTION

Large seismic events occurring under or near the ocean can generate *T* waves, propagated over large distances in the seawater along the SOFAR channel of minimum sound velocity (Tolstoy & Ewing, 1950). Strictly speaking, *T* phases can only be recorded at sea by hydrophones; however, the seismic wave generated by their conversion at an island (or continent) shore can be recorded by a seismometer on the island, and for simplicity, we will call this a seismic record of the *T* wave.

The location of the island of Tahiti in the South Pacific is very favorable to the regular recording of *T* waves generated at the numerous seismic zones bordering this ocean. *T* phases were observed (but not identified) at Hawaii as early as 1927 (Hawaiian Volcano Observatory, 1930), and associated with the earthquake source (although not correctly interpreted) in 1937 at Tahiti (Ravet, 1940). More recently, Talandier (1972) has reviewed their properties. Since the amplitude and duration of *T* waves generated by an earthquake are an expression of the amount of elastic energy transferred to the seawater, their monitoring is an important aspect of the detection and forecast of tsunamis (Ewing *et al.*, 1950; Talandier, 1972).

The purpose of the present paper is to report human perception of *T* waves of unusually large amplitude (we shall simply speak of "felt" *T* waves although, strictly speaking, only the converted seismic wave was felt on the island), following the June 22, 1977 Tonga Islands earthquake, and to discuss the mechanism of their exceptional generation, in view of the fact that much larger earthquakes (e.g., Alaska, 1964) generated *T* waves of lesser amplitude.

DATA

The epicentral characteristics of the June 22, 1977 Tonga Islands event were preliminarily determined by the USGS as: latitude 22.878°S; longitude, 175.900°W; depth, 65 km; origin time, 12:08:33.4 UTC; magnitudes, m_b = 6.8, M_S = 7.2. A small tsunami was also reported.

The seismic data obtained at the 15 stations of the Polynesian network (Figure 1) make possible a reassessment of some focal characteristics. The hypocentral depth will be a major parameter in our understanding of the generation of the *T* waves. A well-defined pP ($pP - P$ = 12 sec) is present at all stations in the network (Figure 2), yielding a depth of 49 km (Jeffreys-Bullen tables) or 47 km [Herrin *et al.*'s (1968) Tables]. The applicability of standard tables to the area of the Tonga trench is certainly questionable, but Woollard (1975) has shown that the involved portion of the crust is slow in this area; this would further reduce the depth of the event. Independent estimates of the magnitude of the event were obtained from most Polynesian records: $m_b \geqq 7.3$ (Tahiti and Tubuai), $m_b = 7.3$ (Rangiroa), $m_b = 7.2$ (Hao), $m_b = 7.1$ (Rikitea); $M_S = 7.8$ (PPT and TPT), $M_S = 7.7$ (RKT), all significantly higher than proposed by the USGS.

The focal mechanism of this earthquake was studied by Stewart and Given (1978, personal communication). They propose a normal mechanism (slip angle, $\lambda = -63°$) along a steeply dipping (dip angle, $\delta = 73°$) fault plane striking S20°W. This clearly indicates that this event is not directly associated with the subduction, but rather probably represents a decoupling event along a fault located above the slab. On the basis of this mechanism, the seismic moment of the earthquake was computed by comparing the multiple Rayleigh waves R_2 and R_3 recorded at PPT with synthetic seismograms obtained by mode summation for periods larger than 100 sec (Figure 3). A figure of 1.5×10^{28} dyne/cm was obtained both from R_2 and R_3, corresponding

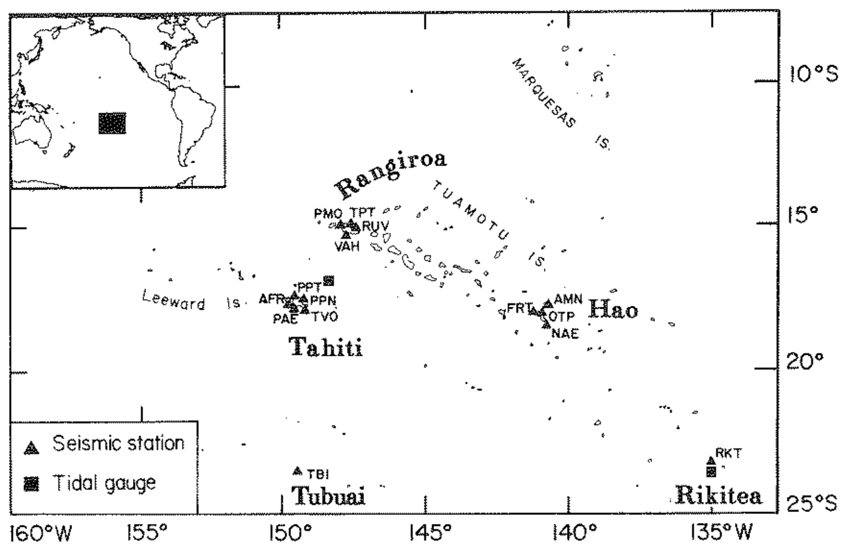

FIG. 1. Map of the seismic network in French Polynesia, showing the three subarrays in Tahiti, Rangiroa, and Hao, and the isolated stations at Rikitea and Tubuai.

to a magnitude $M_W = 8.1$ (Kanamori, 1977). Although this is a large earthquake, we used a point source model since we were dealing primarily with ultra-long periods, for which especially in the particular geometry involved, directivity plays little if any role. The excellent agreement between the figures obtained from R_2 and R_3 constitutes an *a posteriori* justification of this approximation.

The figure of $M_W = 8.1$, the proposed $M_S = 7.8$, and the excitation of a moderate tsunami all indicate that this was a major event, substantially larger than preliminarily reported.

T-WAVE DATA: HUMAN PERCEPTION

Around 2:40 a.m., local time (12:40 UTC), many inhabitants of the West coast of the island of Tahiti were awakened from their sleep by seismic tremors. The following is a typical report from a resident of Paea (see map, Figure 4): "I was awakened by something anomalous at 2:40 a.m. and sat down on my bed. At that time, windows were rattling; the phenomenon lasted approximately one minute".

The time at which the tremors were felt (12:40 UTC) rules out perception of P (arrival time, 12:13) or S (arrival time, 12:17), and is in perfect agreement with the instrumental record of T waves (Figure 5). The long duration of the phenomenon, and the frequencies of several hertz suggested by numerous reports of rattling windows and tinkling glass also agree with this interpretation. Furthermore, as shown on Figure 4, the capital city of Papeete and the Faaa district, where the event was not felt, are clearly masked from the epicenter by the island of Mooréa; nor were the tremors felt at the Club Méditerrannée resort, on the West coast of Mooréa, an area itself masked by the small island of Tubuai-Manu. This again confirms that the felt tremors were due to the T waves following the earthquake.

T waves have been reported felt at teleseismic distances in Hawaii, following the

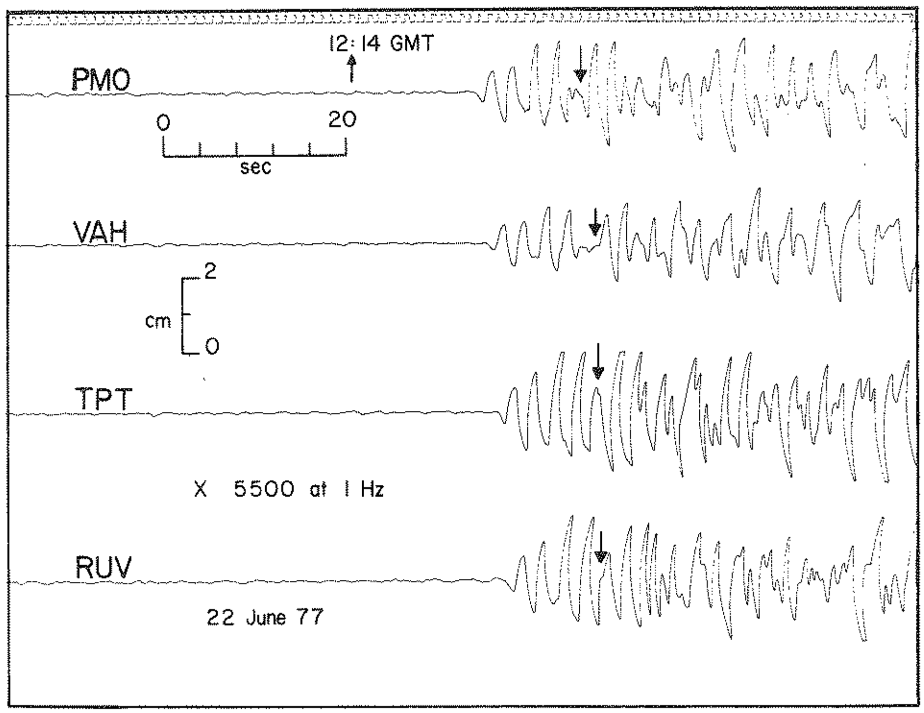

FIG. 2. Short-period vertical records of the initial phases of the June 22, 1977 Tonga event at the Rangiroa subarray. Note the consistent pP arrival 12 sec after P.

large July 10, 1958 Alaskan earthquake, and also following a man-made explosion off the coast of Mexico, on May 14, 1955 (J. P. Eaton, personal communication, 1979). However, this is believed to be the first report of perception of teleseismic T waves in French Polynesia. The epicentral distance at Paea was 2807 km, and the reported phenomena correspond to a degree of IV on the 1931 Modified Mercalli Scale of intensities. The nighttime occurrence of the event was certainly a favorable condition in this respect, but tremors of this intensity would have been felt even during daytime.

Instrumental Data

The exceptional amplitude of the T waves is well documented on Figure 5. With the exception of the Rangiroa stations, masked from the epicenter by Maupelia, Maupiti, and Tupai in the Leeward Islands group, the low-gain (21,000 at 3 Hz) graphic records were off-scale at all stations, and amplitudes at PPT were computed from broadband magnetic tape recordings. Amplitudes at PAE were inferred from a comparison with records of T waves from aftershocks, for which both stations (PPT and PAE) remained on scale.

Table 1 gives peak-to-peak displacements, velocities, and accelerations, both at PPT and PAE. Accelerations on the West coast of the island (close to Paea)

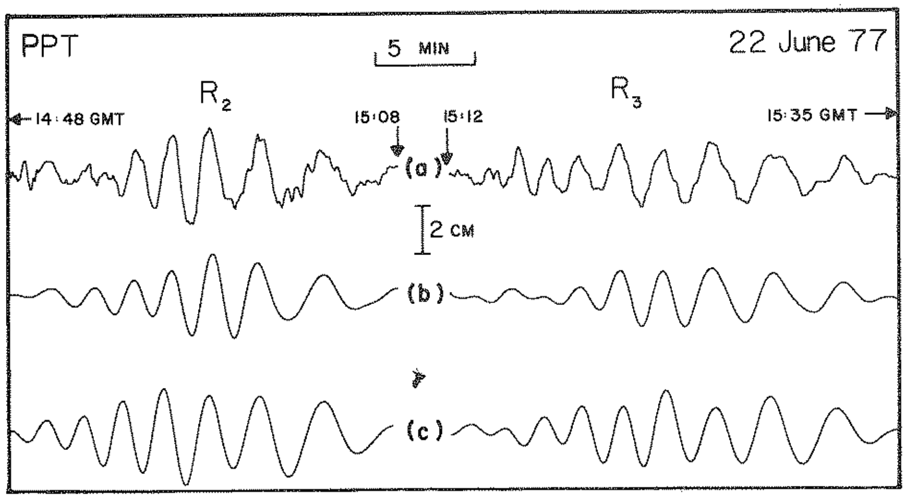

FIG. 3. Broad-band records of the multiple Rayleigh waves R_2 and R_3 at PPT. (a) Original records. (b) Observed records filtered at $T \geqq 100$ sec. (c) Synthetic seismograms obtained by modal summation; a moment value of 1.5×10^{28} dyne-cm is obtained by matching the amplitudes of (b) and (c) both for R_2 and R_3.

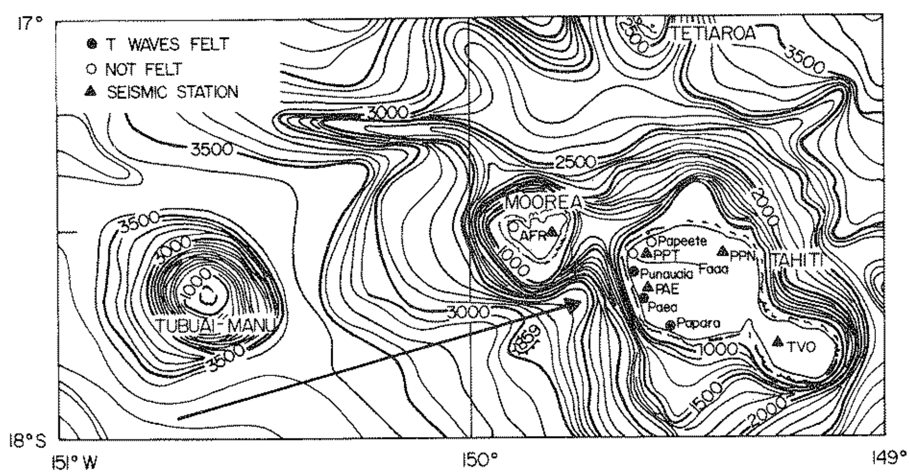

FIG. 4. Bathymetric map of the vicinity of Tahiti and Mooréa. The areas where the tremors were felt are shown as full circles. Open circles indicate populated areas where no felt tremors were reported. Note also the steep slope offshore of Paea, favorable to the conversion of T energy back into seismic waves. The arrow shows the azimuth of arrival of the T waves. (After Monti and Pautot, 1974.)

correspond to an intensity degree of IV on the 1931 Modified Mercalli Scale, in agreement with the phenomena reported in the previous section.

MECHANISM OF GENERATION

The T waves recorded at the Polynesian stations correspond to a propagation in the SOFAR channel (Officer, 1958), which can take place only if the incident ray

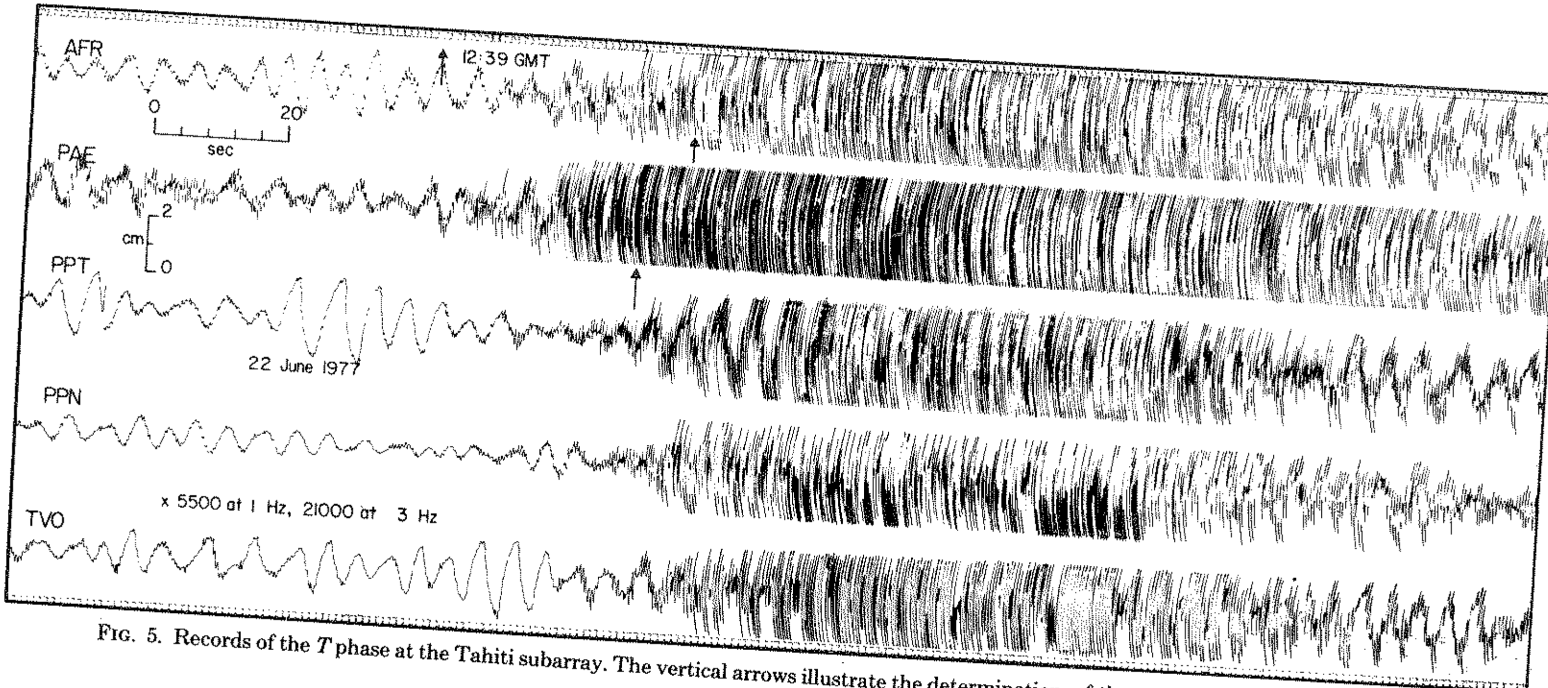

FIG. 5. Records of the T phase at the Tahiti subarray. The vertical arrows illustrate the determination of the arrival times listed in Table 3.

TABLE 1

OBSERVED T-WAVE GROUND MOTIONS AT PPT AND PAE*

Station and Component	Displacement (mm)	Velocity (mm/sec)	Acceleration (mm/sec^2)
PPT Z	0.012	0.17	5.2
PPT EW	0.018	0.34	7.9
PPT NS	0.015	0.22	6.6
PAE Z	0.05	1.0	22
PAE EW	0.08	1.6	35
PAE NS	0.06	1.2	26

* All values are peak-to-peak amplitudes.

penetrates the channel at a horizontal incidence less than 12°. The source of the present T waves can be investigated on this basis, with the help of a precise bathymetric map of the epicentral area. Figure 6, adapted from Mammerickx *et al.* (1974), shows that a favorable slope, dipping approximately 11° to 12° exactly in the direction of the Polynesian stations, is located about 40 km to the east of the epicenter (see cross section on Figure 7). Using Woollard's (1975) crustal model of the island side of the Tonga trench (which includes a well-developed zone of low

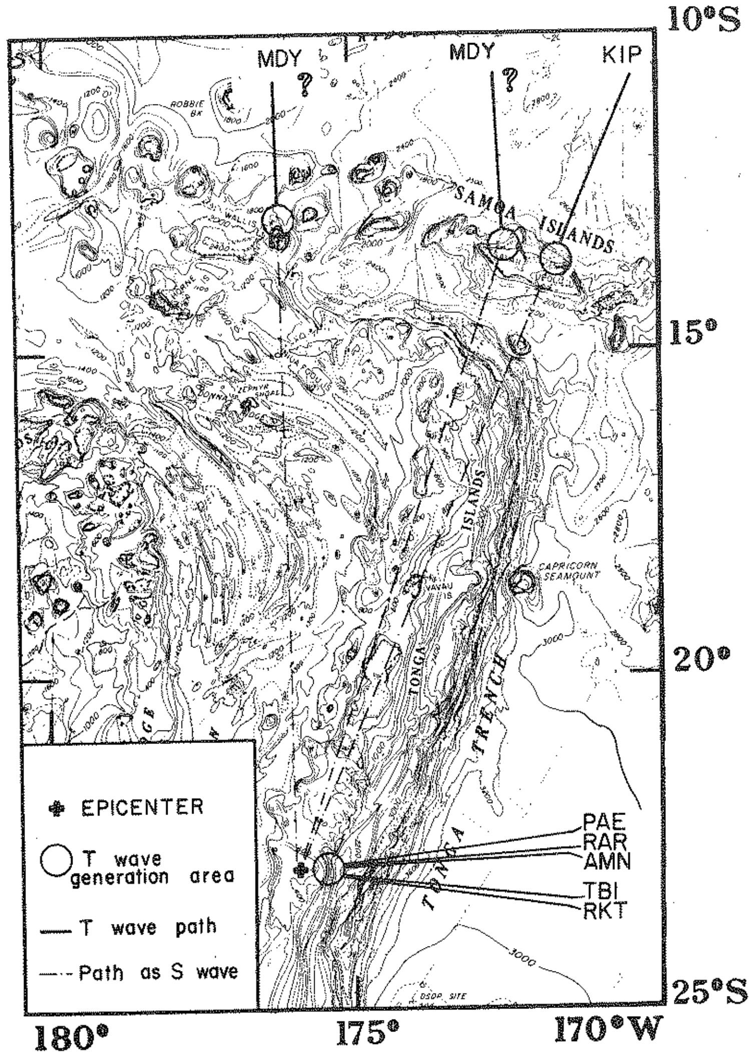

FIG. 6. Close-up of the bathymetry (Mammerickx *et al.*, 1974) in the vicinity of the epicenter, showing the nearby slope, favorable to T-wave generation. The probable generation mechanisms for the T-waves observed at KIP and MDY are also shown.

velocities), we investigated the refraction of P waves into the SOFAR channel. P waves are incident on the ocean at an angle of 40° (with respect to horizontal), and after three water-surface and three ocean-bottom reflections, they can penetrate the SOFAR channel at an incidence less than 12°. In the particular geometry involved, it can be shown that all three ocean-bottom reflections are beyond critical, leading to no loss of energy back to the solid crust. Other crustal models would yield only slightly different results, with a similar conclusion about the penetration of SOFAR.

Both S and the phases Sp and Ps, converted at the Mohorovičić discontinuity, would also lead to the penetration of the SOFAR channel after two or three reflections.

Furthermore, this favorable geometry is accentuated by a favorable source radiation pattern. Using the source geometry of Stewart and Given (1978, personal communication), and a takeoff angle of 130° (computed from Wollard's model), we obtained a radiation pattern coefficient (Kanamori and Stewart, 1976) $R^P = 0.84$, which confirms that a substantial amount of P energy is directed toward this portion of the oceanic slope. The uncertainty of the bathymetric map (Mammerickx *et al.*, 1974) is probably larger than the wavelengths involved ($\lambda = 500$ m in the water), and localized steeper slopes (a common feature of the saw-toothed trench bathymetry) could further enhance the transmission of energy into the SOFAR channel.

It is interesting to note that a similar bathymetry does exist (Chase *et al.*, 1970) in the vicinity of the rupture area of the large July 10, 1958 Alaskan earthquake,

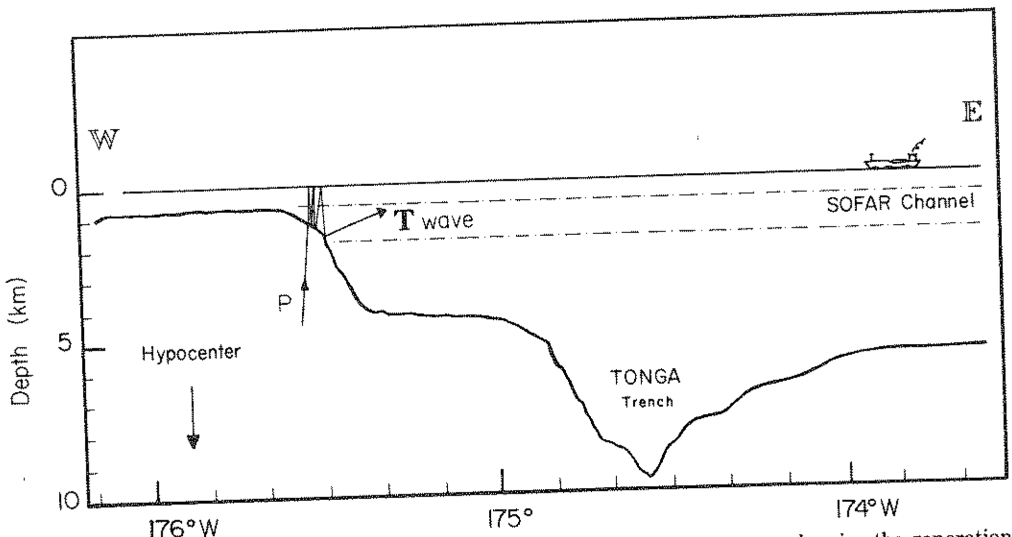

FIG. 7. Cross section of the map on Figure 6, at the latitude of the epicenter, showing the generation of T-waves at the slope of the trench. The vertical scale is exaggerated 10 times.

whose T waves were felt on the island of Hawaii (parameters for this event are given in Ben-Menahem and Toksöz, 1963).

Conversion at the Island Shore

Figure 4 shows the bathymetry in the vicinity of the west coast of the island of Tahiti. Offshore from the districts of Paea and Punauaia, where the tremors were reported felt, the slope of the island is directed precisely toward the direction of the incoming T waves. The geometry precludes horizontal reflection of the T waves, and favors the acoustic-to-seismic conversion of energy.

The average slope of the island is 14° to depths of 1000 m, and 10° between 1000 and 2500 m. An incident ray, propagating in the SOFAR channel at a depth of 1200 m, undergoes a first reflection 5 km from the shore, a second one at 1.25 km, and a third one at 0.8 km. Refraction into the island is possible at any of these points, provided that the seismic velocities in the superficial layer be smaller than, respectively, 1.53, 1.85, and 3.42 km/sec. A seismic refraction profile (Talandier and Kuster, 1976) has indicated the existence of four layers in the island's structure, the top one, which is found only in the immediate vicinity of the shoreline, being particularly slow (see Table 2), and consisting of sediments and coral debris, bordered by basaltic cliffs of more or less weathered nature (de Neufbourg, 1965). The wave observed to travel in that layer at 1.28 km/sec may be either an S wave, or a Love-type wave, the velocity being in this case the group velocity. Even in this latter case, the S-wave velocity in the detritic layer would be on the same order of magnitude and in all cases, conversion of the T wave to seismic energy is found possible: to S waves (possibly channeled) in the top layer at all reflection points of the T wave, and to P waves in the top layer and S waves in the second layer at the third reflection point. Therefore, on the basis of the seismic structure of the island, the conversion from T wave to seismic energy is found to be favorable. Although the S (or Love) waves propagated in the layer of sediments and debris were observed to vanish quickly (after only 10 km) during the seismic refraction experiments, most of the populated area of the island, including all areas which reported felt tremors, is concentrated on

TABLE 2

CRUSTAL STRUCTURE OF TAHITI

Layer Thickness (km)	P-Wave Velocity (km/sec)	S-Wave Velocity (km/sec)
0.1 to 0.2	2.5	1.28*
5.4	4.7	2.3
11.6	7.2 to 7.6	3.7
Mantle	8.1 to 8.5	

* This probably represents the group velocity of a channeled wave.

TABLE 3

COMPUTED AND OBSERVED TRAVEL TIMES OF T WAVES

Station	Azimuth to Station (°)	Distance (km)	Distance Traveled as T Wave (km)	Total Travel Time (sec)	Arrival Time		Residual (sec)
					Computed	Observed	
PAE	83.0	2807	2765	1874	12:39:47	12:39:40	−7
AMN	87.9	3693	3652	2471	12:49:44	12:49:38	−6
TBI	96.3	2707	2665	1806	12:38:39	12:38:50	11
RKT	98.6	4184	4143	2802	12:55:15	12:55:22	7

the ring-shaped coastal plain, bordering the lagoon, whose width is only 300 to 800 m. We therefore propose to identify the waves felt on the west coast of Tahiti as the converted T waves, refracted in the detritic superficial layer.

Our interpretation of the generation mechanism of the T waves can be checked by studying the arrival times of the phase at the various Polynesian stations. The velocity of sound in the SOFAR channel has been investigated by Johnson (1969) and Johnson and Norris (1968). In the area involved in the present study (14° to 23°S), the depth of the SOFAR channel extends from 600 to 1800 m, its axis being at 1200 m, with a minimum velocity of 1484 m/sec. Using this figure, we computed a theoretical travel time corresponding to P-wave propagation from epicenter to generating slope, and T-wave propagation on to the receiving island. The travel time of the converted wave on the island is always extremely small and was neglected. Table 3 compares the observed and computed travel times. T waves being emergent in nature, we define the observed travel time as the onset of the first wave train of maximum amplitude (see Figure 5). Given the uncertainty in this procedure, the agreement between observed and computed times is excellent, and confirms our proposed mechanism and location for the generation of the T wave.

Additional Data and Discussion

Table 4 is a list of additional data obtained from various stations around the Pacific.

The low amplitude of T-wave records at Kipapa and Diamond Head, Hawaii is probably due to the presence of seamounts immediately south of Oahu, lying in the path of the incident T waves and, in the case of Kipapa, to the location of the station several miles inland. However, the records clearly show that the T waves at both

TABLE 4

Additional T-Wave Data at Circum-Pacific Stations

Station		Epicentral Distance (km)	Azimuth (°)	Arrival Time (GMT)	Amplitude (peak-to-peak) (μ)	Remarks*
Code	Name					
MDY	Midway Island	5667	358	13:05	Hydrophone record	7 min early
KIP	Kipapa, Oahu	5279	23	12:59	0.07	8 min early
	Diamond Head, Oahu	5269	23	12:59		8 min early
BKS	Berkeley, CA	8763	41	13:45	0.12	On time
SYP	Santa Ynez, CA	8681	44	13:45	0.30	On time
SBSM	San Miguel Is., CA	8620	45	13:44	0.30	On time
RAR	Rarotonga, Cook Is.	1673	87	12:27	0.27	On time
ECZ	East Cape, N. Zealand	1730	197	12:28	1.2	On time
WTZ	Whakatane, N. Zealand	1810	201	12:28	0.40	On time
RIV	Riverview, Australia	3429	243	12:34	0.11	11 min early
PET	Petropavlovsk, Kamchatka	8774	345	No T wave; masked by Gilbert and Ellice Islands		
LPS	LaPalma, El Salvador	10292	76	No T wave; masked by Marquesas Islands and East Pacific Rise		
GIE	Galapagos Is., Ecuador	9525	89	No T wave; masked by Tuamotu Islands and East Pacific Rise		
NNA	Ñaña, Perù	10401	105	No T wave; masked by East Pacific Rise		

* The timing is with respect to full T-wave propagation from the immediate vicinity of the epicenter to the station.

Hawaiian stations are about 8 min early. We propose that this is the result of the generation of the T waves taking place far from the epicenter, in this case seaward of the Samoa Islands (see Figure 6). A favorable slope exists in this area and T waves generated a large distance away from an epicenter (as far as 47°) have been reported by Shurbet (1955) at Bermuda, and by Talandier (1972), following the 1970 Colombian earthquake. The analysis of the travel time of the T phase suggests that it traveled as SV from the epicenter to Samoa; this is confirmed by the body-wave radiation pattern of the event, which favors SV over P in this geometry.

Similarly, the T waves at Midway, approximately 7 min early, were probably generated either in the area of Wallis Island, after propagating as SV, or may result from refraction (with a change of azimuth) seaward of Savaii, Samoa. T waves at Riverview, Australia, 11 min early, are probably due to $SV \rightarrow T$ conversion taking place on the western slope of Norfolk Ridge, in the vicinity of Norfolk Island; both travel times and radiation patterns confirm this interpretation.

The case of Rarotonga (RAR) is puzzling, since the amplitude of the T waves on this island is on the order of 100 times smaller than on Tahiti, despite a shorter epicentral distance, and a similar azimuth from the epicenter (Figure 6). Since the structure of the island is similar to that of Tahiti, although smaller in size, seismic waves converted from T waves should be of similar amplitude. One possible explanation is the presence of a seamount masking the island, located approximately 100 km west of Rarotonga, shown on Mammerickx *et al.*'s map. It should be noted that the bathymetric coverage in this area is rather sparse, so that other seamounts may exist and deviate T waves.

The amplitudes of the T waves recorded around the Pacific clearly prove that the exceptional amplitude at Tahiti is due to a path effect, namely favorable conditions for generation at the slope, propagation and conversion at the shore, rather than to a source effect. The source of the Tonga earthquake is certainly large, but not gigantic ($M_W = 8.1$). Much larger events have occurred recently around the Pacific Ocean: Chile (1960), Kuriles (1963), Alaska (1964), Aleutian (1965). However, none

TABLE 5

TSUNAMI AMPLITUDES MEASURED AT VARIOUS PACIFIC STATIONS, JUNE 22, 1977

Station	Amplitude (cm)
Papeete, Tahiti	10
Pago-Pago, American Samoa	15
Hilo, Hawaii	10
Kahului, Hawaii	27
Nawiliwili, Hawaii	10
Honolulu, Hawaii	10
Point San Luis, CA	25
Long Beach, CA	25
San Diego, CA	20
Los Angeles, CA	10

of the T waves from these events were large enough to overcome the probably less favorable path conditions, and be felt on shore, despite seismic moments respectively 130, 4, 55, and 8 times larger (Kanamori, 1977). On the other hand, the 1958 Alaskan earthquake was only twice as large as the present Tonga event. The fact that the amplitude of T waves is not in direct relation with the size of the earthquake can be understood from scaling relations. Geller (1976) has shown that Haskell's source model leads to a saturation of the 1-sec m_b scale around $m_b = 6.5$ to 7, corresponding to $M_W \geqq 6.9$, just as the 20-sec M_S scale saturates around $M_S = 8.2$. [Larger values of m_b (up to $7\frac{3}{4}$) are due to measurements taken at longer periods, especially in the case of S waves.] Similarly, a magnitude scale m_T characteristic of the energy of the seismic waves in the frequency band of T waves, and representative of the maximum possible amplitude of T waves, would saturate even earlier, probably around values of $M_T = 5.7$ to 6.0, corresponding to $M_W \geqq 6$. Any earthquake greater than $M_W = 6$ would have maximum possible T waves of similar amplitude, the criterion for their generation and observation resting entirely with the particular geometry and path involved. Conversely, tsunamis reflect the low-frequency behavior of the source, associated with the "static" magnitude M_W (Kanamori, 1972, 1977), in direct relation with the rupture length of the earthquake. Scaling laws (Geller, 1976) would give the present event a fault area S of about 10,000 km^2, and a displacement of a few meters, in agreement with a tsunami of only moderate amplitude, as compared, for example, with the 1964 Alaskan tsunami (see Table 5 and Figure 8). The characteristic length of the source (grossly $\sqrt{S}$) would affect the rupture time, and therefore the duration, not the amplitude, of the T waves. This point, mentioned by Ben-Menahem and Toksöz (1963), is the basis of the use of duration (rather than amplitude) of T waves in Tsunami Warning Procedures (Talandier, 1972). As an illustration of this relationship between length of the phase and moment, it is possible to use the duration of the felt T waves at Tahiti to infer the seismic moment of the event. The duration t of the maximum amplitude of the T waves was 1 min. This corresponds to a fault length (Ben-Menahem and Toksöz, 1963)

$$L = 2\, t v_T/(v_T/v_R + \cos\theta),$$

where v_T and v_R are the velocities of the T wave and of the rupture, and θ the azimuth from the fault to the station. Many parameters of the actual geometry are

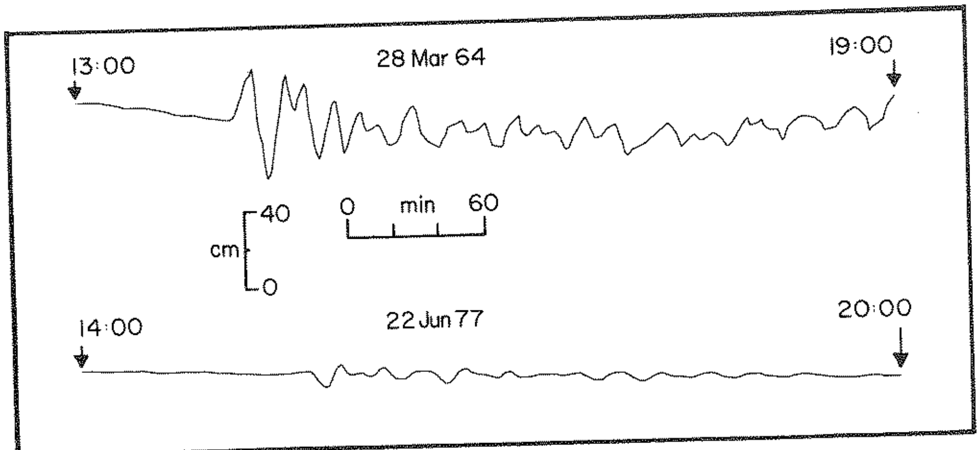

FIG. 8. Comparison of tidal gauge records of the 1964 Alaska and 1977 Tonga tsunamis at the Faré-Uté tidal gauge station in Papeete, Tahiti. This clearly demonstrates the much larger size of the 1964 event.

unknown (such as whether the rupture reached the ocean floor, etc...), but an order of magnitude can be obtained. Using $\theta = 60°$, we obtain $L \cong 180$ km, or $M_W = 8.2$ (Geller, 1976), in good agreement with the seismic figure of 8.1.

An excellent agreement is also obtained by comparison with data from the Chilean earthquake of 1960. Eaton *et al.* (1961) reported a duration of 5 min for the maximum amplitude of the T phase at Hawaii. The ratio of rupture lengths for the two events, then, would be 5, that of the moments (proportional to L^3), 125, yielding $M_W = 8.1$ for the 1977 Tonga event.

CONCLUSION

We have presented a report of human perception of T waves at teleseismic distances, corresponding to a degree of IV on the 1931 Modified Mercalli Scale of intensities. The exceptional amplitude of the waves is found to be the result of favorable conditions at the source, and in the areas of the $P \rightarrow T$ and $T \rightarrow$ seismic conversions. The seismic moment of this earthquake, computed from ultra-long period multiple Rayleigh waves, suggests a length of rupture on the order of 100 km, confirmed by the analysis of the duration of the maximum amplitude of the T phase.

The gigantic amplitude of the T waves in French Polynesia, is not a consequence of the size of the earthquake, as confirmed by smaller amplitudes at other stations around the Pacific Ocean.

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