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DISPERSION OF ONE-SECOND RAYLEIGH MODES THROUGH OCEANIC SEDIMENTS FOLLOWING SHALLOW EARTHQUAKES IN THE SOUTH-CENTRAL PACIFIC OCEAN BASIN

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INTRODUCTION

The purpose of this paper is to report the observation of anomalously long wavetrains of high-frequency seismic energy (0.3 - 1 Hz) following shallow events at a number of epicenters in the South-central Pacific basin, and to model their propagation and generation as due to efficient coupling of the oceanic water column with the solid Earth through a well-developed transitional layer of sediments. The following paragraphs are a short summary of the seismic features of the area, and provide a framework for our study.

The French Polynesia seismic network, consisting of 15 permanent stations, is presently the only large aperture seismic array operating in an oceanic intraplate environment. As such, it can provide exceptional insight into phenomena associated with intraplate seismicity. The network has been fully described elsewhere [Okal et al., 1980; Talandier, 1972]. Among its stations, Rikitéa (RKT), and the four stations on Hao atoll are of primary importance for detection capabilities in the region extending between Polynesia and the East Pacific Rise. Solid triangles on Figure 1 show the location of these five stations, superimposed on the bathymetric map of Mammerickx et al. [1975]. For clarity, the other stations in the network, located on Tahiti, Rangiroa and Tubuai, and irrelevant to the present study, are not shown.

The stations are equipped with short-period vertical seismometers, coupled to a variety of electronic filters [Talandier, 1972] which damp ocean-generated noise, and permit routine magnifications of 10^5 at 1 Hz and 2×10^6 at 3 Hz, the latter allowing detection of

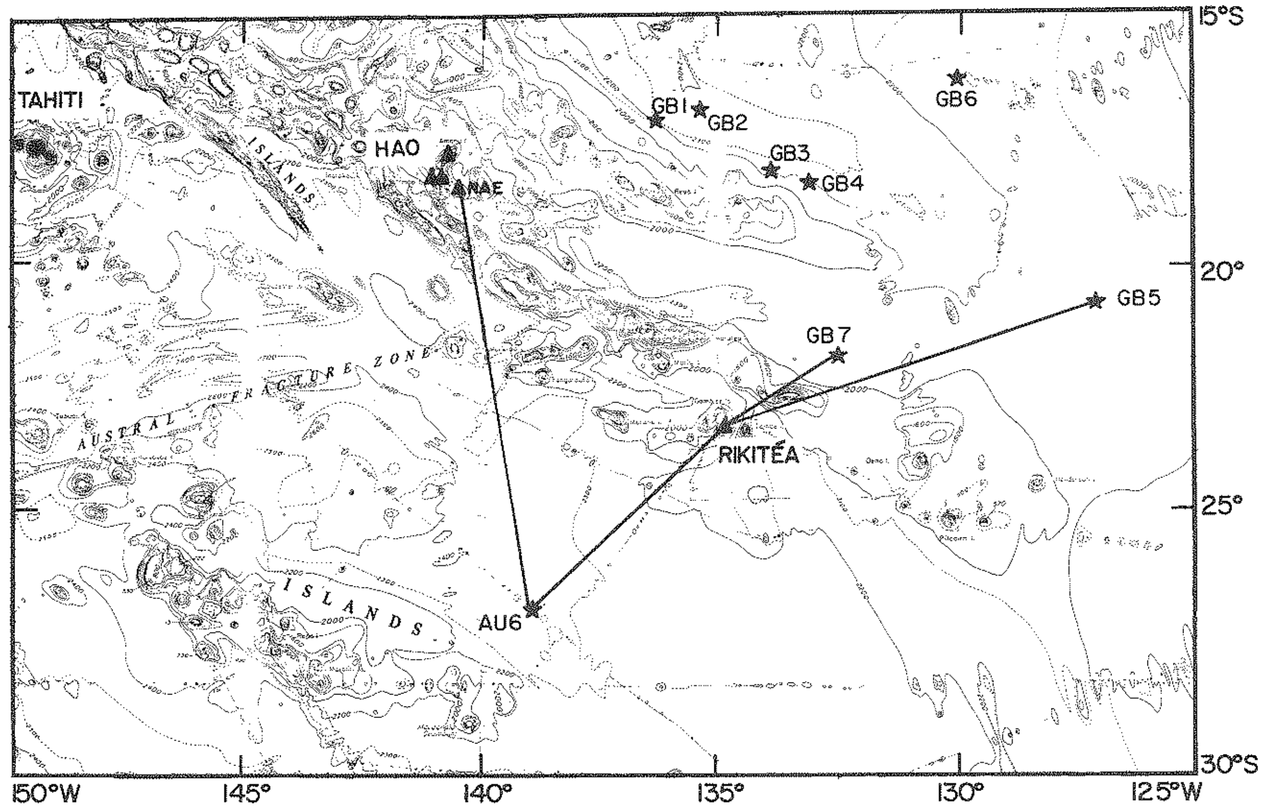

Figure 1 : Map of the bathymetry of the Eastern Tuamotu Islands area (Mammerickx et al., 1975), showing seismic foci East of 140°W (stars) and the seismic stations at Rikitea and on Hao atoll (triangles). Thick lines identify the only paths along which anomalous wavetrains are observed. Note bathymetric features along other paths, e.g. GB1-RKT.

low-amplitude T waves converted at the island shore. The one-second magnification of 10^5 is one order of magnitude above that of WWSSN stations in oceanic environment, and results in considerable boosts in the detection capabilities of the network: most of the area covered by Figure 1 has a detection threshold of $M_L = 3.0$ to 3.5.

Seismicity in the area of the Gambier Islands has been studied extensively by Okal et al. [1980]. Eight epicenters have been identified to the East of 140°W, called GB1-GB7 and AU6 in these authors' classification. They are shown as stars on Figure 1. Over the past 15 years, 116 events have been recorded at these eight locations, with a maximum m_b of 5.5; among the eight epicenters, two (GB4 and GB5) have been particularly active. With 96 events recorded in 1976-79, GB5 is the most active seismic focus inside the Pacific plate (Hawaii excepted). Because of poor station coverage at teleseismic distances, and of the relatively low magnitude of these events, their precise study has been difficult. Since GB5 lies in an area where bathymetric coverage is almost inexistant, the structural interpretation of this seismicity can only be speculative. Nevertheless, Okal et al. [1980] have argued on the basis of the systematic observation of water multiples that these foci must be very shallow, probably within a few km of the water-sediment interface. They have also been able to interpret this seismicity as due to the release of compressional tectonic stress in the oceanic plate.

OBSERVATIONS

Routine recording at RKT of seismic signals from events at GB5 (21°S, 127°W; $\Delta = 7.95$ degrees), GB7 (22°S, 132°W; $\Delta = 2.96$ degrees) and AU6 (26.7°S; 138.9°W; $\Delta = 5.08$ degrees) shows a relatively high-frequency wavetrain lasting up to 20 minutes, whose onset grossly corresponds to the expected arrival time of Rayleigh waves.

Figure 2 is a reproduction of the seismograms obtained from AU6 on November 20, 1979. Focal characteristics for this event are given by Okal et al [1980] as : Latitude 26.75°S, Longitude 138.88°W, Origin Time 06:45:02, Magnitude $m_b = 5.3$. The intermediate and bottom frames on Figure 2 are the continuation of the top one. The trace labeled (a.), which uses a low-gain broad-band amplifier, clearly shows the arrival of P_n, followed by S_n about one minute later. Shortly after S_n, the one-second wavetrain starts arriving with a group velocity of 4 km/s. It lasts for about 13 minutes, before the amplitude dies off to a level comparable to the noise preceding the event. The corresponding group velocity is then 0.6 km/s. This longlasting wavetrain must be opposed to records from events of similar magnitude at similar distances, such as at GB1 or GB4, whose energy dies off to the noise level over a period on the order of 3 to 4 minutes. The wavetrain observed for the AU6

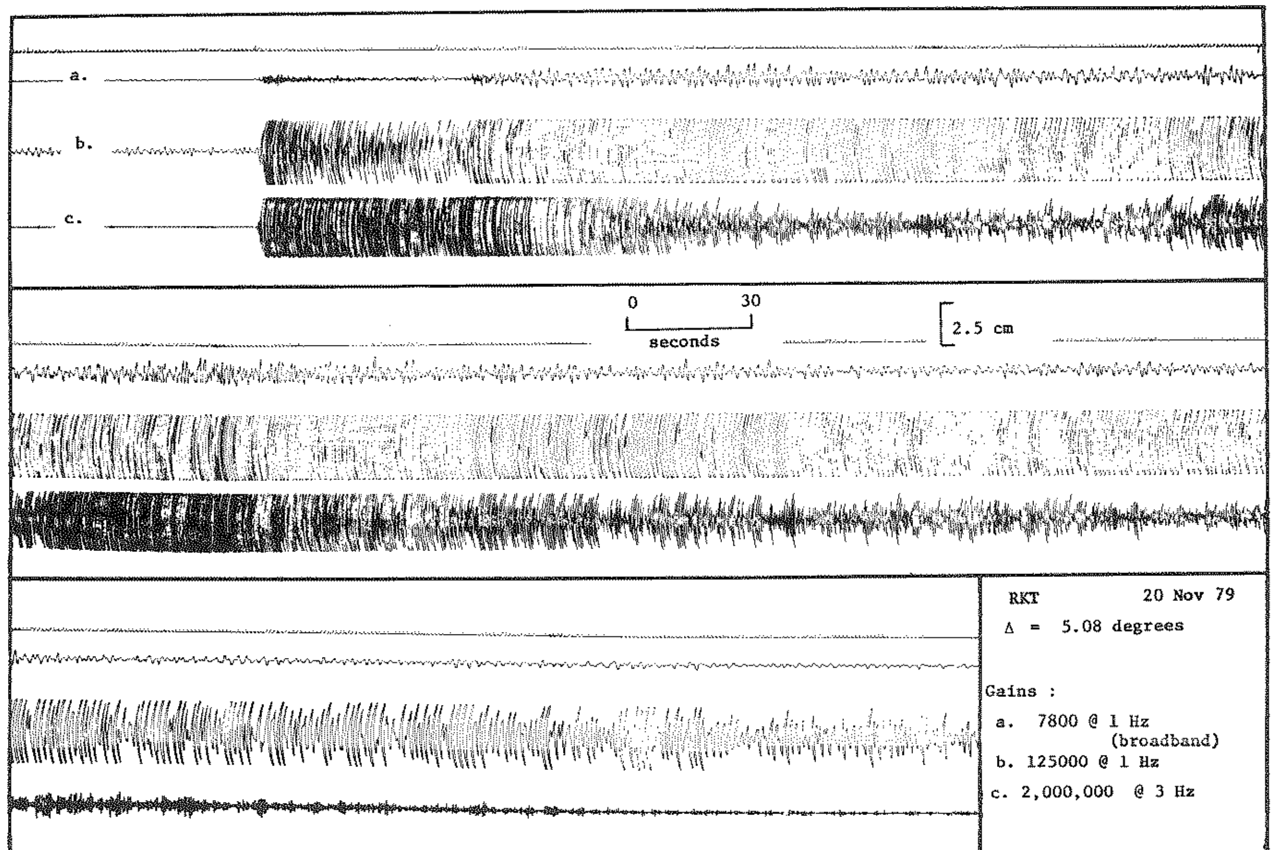

Figure 2 : Seismograms obtained at RKT following event at AU6 on Nov. 20, 1979. The anomalous wave train is best seen on trace (b.), lasting more than 13 minutes. Note low level of signal on high-frequency channel (c.).

Table 1. Summary of observations of anomalous wavetrains

Epicenter	RKT (distance in °)	NAE (distance in °)	Other stations on Hao
GB1	no (5.63)	no (4.21)	no
GB2	no (6.10)	no (5.49)	no
GB3	no (5.14)	no (6.63)	no
GB4	no (5.11)	no (7.42)	no
GB5	YES (7.85)	no (13.22)	no
GB6	no (10.69)	no (13.90)	no
GB7	YES (2.96)	no (8.88)	no
AU6	YES (5.08)	YES (8.47)	no

event saturates the high-gain channel with a magnification of 125,000 at 1 Hz (trace b.); from the low-gain records, we can infer a peak-to-peak ground displacement at this period of 9×10^{-7} m or 0.9 microns, for this $m_b = 5.3$ earthquake. On the other hand, channel (c.), whose response is peaked around 10 Hz, with a gain of 2×10^6 at 3 Hz (see Talandier and Kuster [1976] for details), is saturated only during the arrival of the T phase generated by the event (at the beginning of the second frame in Figure 2), and shows little energy present over background noise in the later part of the seismogram. This, and a direct measurement of the predominant periods on trace (b.), indicate that most of the energy present in the observed long-lasting wavetrain (which, for clarity, we will henceforth call "anomalous wavetrain") is concentrated around one to two seconds in period.

Observations of a totally similar nature are repeatedly made at Rikitéa for events occurring at GB5. The group arrival times characterizing the onset of the anomalous wavetrain and its coda correspond to velocities of 4.0 and 0.7 km/s, and the amplitude recorded at

RKT for major events (e.g. October 31, 1977, $m_b = 5.1$; January 5, 1978, $m_b = 5.5$) indicates a peak-to-peak ground motion on the order of 0.4 microns. For the single event at GB7 (November 6, 1978), similar values are obtained, despite a lower magnitude ($m_b = 4.5$).

Apart from RKT, the only station in French Polynesia having recorded anomalous wavetrains is Naké (NAE), at the southeastern tip of the atoll of Hao (see Figure 1). For the November 20, 1979 event at AU6 ($\Delta = 8.47$ degrees), a wavetrain lasting 18 minutes, with a maximum peak-to-peak ground motion of 0.5 microns, was recorded despite a waveform less regular than at RKT, and a considerably lower signal-to-noise ratio. It is important to note that the signal at the other three stations in the Hao subarray falls down to noise level after about 4 minutes, as in the case of 'normal' earthquakes, and thus no anomalous wavetrain can be identified.

Furthermore, no anomalous wavetrains are observed, either at Rikitéa or Hao, for events located at the other nearby epicenters (GB1-4 and GB6), despite comparable magnitudes; nor are they observed at NAE for GB5 and GB7 events. Finally, no such wavetrains have ever been observed in the Tahiti or Rangiroa stations. Table 1 summarizes observations of anomalous wavetrains in the five Eastern Tuamotu stations.

INTERPRETATION

There can be a priori two causes to the generation of the observed anomalous wavetrains: a source effect or a path effect. We can readily reject a source effect on the basis of three observations:

First, an investigation of teleseismic records of events from GB5 and AU6 has failed to reveal any anomalous waveforms at large distances from the source. Anomalous wavetrains are therefore characteristic of local propagation;

Second, it is clear from GB5 events, which give rise to anomalous wavetrains at RKT but not at NAE, that anomalous wavetrains are observed only along preferential paths;

Finally, the duration of the wavetrain is grossly proportional to the distance traveled, and corresponds to limiting group velocities of 4.0 and 0.6 km/s; a source of long duration would generate a wavetrain of the same duration at all distances; furthermore, it would be totally unrealistic to envision magnitude-5 earthquakes involving a rupture lasting 10 minutes or more: such events have fault dimensions typically on the order of 2 km, and rupture times of no more than 2 seconds [Geller, 1976].

It appears therefore that the observation of anomalous wavetrains is due to a path effect. The possibility that this may represent dispersion involving an unusually steep group velocity

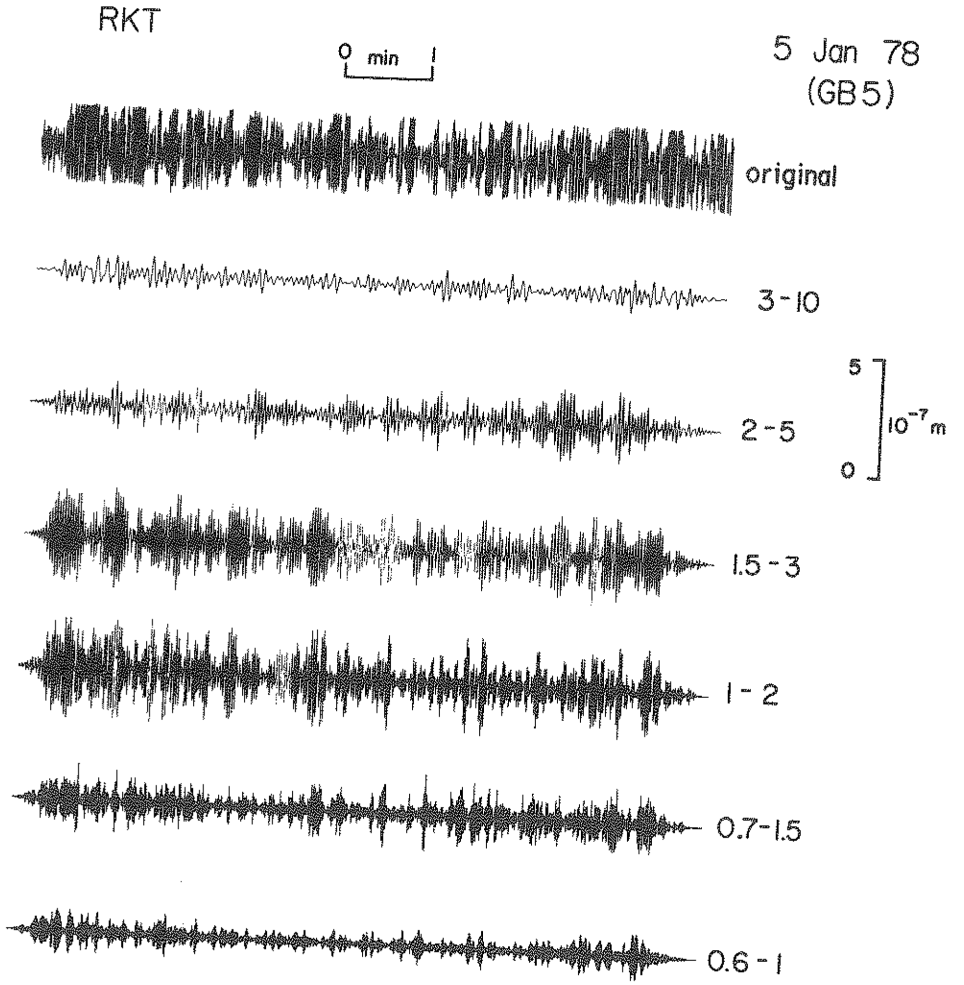

Figure 3 : Selective band-pass filtering of anomalous wavetrain recorded at RKT for event at GB5 on Jan. 5, 1978. Top: original seismogram; other traces: seismogram filtered using cut-off periods given on right-hand side (in s). Note strong decay of signal for frequencies greater than 1 Hz.

Hz)

curve was examined next, through selective band-pass filtering of a high-quality record. We used a stretch of 8 minutes of the high-gain RKT record for January 5, 1978 ($m_b = 5.5$), corresponding to group velocities ranging from 4.2 to 1.3 km/s. After digitizing at a sampling rate of 10 points per second, records were band-pass filtered around 6 frequencies ranging from 0.2 to 1.33 Hz. Results are shown on Figure 3. The top trace is the original record, and for each of the following traces, the numbers at right identify the cut-off periods (in seconds) of the band-pass filters used. All traces are on the same scale, whose ground motion equivalent is also shown.

It is apparent at once that most of the energy in the wavetrain is concentrated between one and three seconds, a result already suggested by a time-domain analysis. Very little energy is present at lower frequencies, and the amplitude also dies off quickly if the frequency is increased beyond 1 Hz. Furthermore, it appears that the anomalous wavetrain cannot be the result of the complex dispersion of a single branch of superficial waves, since the energy at a given frequency (say 1 Hz) is not concentrated at a single group arrival time. Rather, the present situation suggests the complex interference of several high-frequency modes.

Although the bathymetric data available for this portion of the Pacific Ocean is rather poor [Mammerickx et al., 1975], no ridges or seamount chains have been identified along the paths GB5-RKT, GB7-RKT and AU6-RKT, characteristic of the anomalous wavetrains. Along the path AU6-NAE, substantial bathymetric features are found only in the immediate vicinity of the receiver. On the other hand, the epicenters GB1-4, which lie on the northern flank of the Tuamotu archipelago, are separated from the Hao atoll subarray by the northern branch of this island chain (Raroia - Fakahina - Tatakoto - Réao), which forms a continuous plateau at a depth of 1500 fathoms. They are also separated from RKT by the prolongation of this plateau past Réao, where depths remain shallower than 1700 fathoms [Institut Géographique National, 1969; Mammerickx et al., 1975].

We therefore propose that the anomalous wavetrains observed along the paths GB5-RKT, GB7-RKT, AU6-RKT and AU6-NAE result from the interference of a number of branches of modes propagating primarily in the water column and sedimentary transition zone. Any disruption in this structure, such as the presence of a barrier of seamounts, makes this propagation impossible and leads to the disappearance of the wavetrain. A similar situation is encountered if the propagation path crosses an area of irregular, rugged relief, such as a succession of fracture zones. This explains why no anomalous wavetrains were recorded in Tahiti from AU6, a path crossing the Austral Fracture Zone, and whose bathymetry is very irregular. Also, the low attenuation characteristic of modes of the water column favors propagation over great distances at sea, but results in comparatively very inefficient propagation over an island structure, once the energy has been converted to more conventional seismic waves at the island shore. The inability of high-frequency seismic energy to propagate in detritic structures such as those present on oceanic atolls as been described in T wave studies [e.g. Talandier and Okal, 1979], and explains the absence of anomalous wavetrains at the three northern stations of the Hao subarray for the AU6 event, despite a good signal at NAE.

We further propose that the sedimentary and transitional layers are fundamental coupling agents between the water and the solid Earth, and are responsible for the efficient excitation of the modes involved in the anomalous wavetrains by seismic sources in the crust. In the next section, we will study the dispersion of normal modes for two very crude models, in order to justify our proposed interpretation, and discuss this matter more quantitatively. It should be kept in mind however, that this modelling can only be tentative and crude, since the wavelenghts involved are on the order of a few km, several orders of magnitude below the accuracy of our present knowledge of the lateral variations of the bathymetry in this part of the Pacific.

NORMAL MODE MODELLING

In this section, we investigate theoretically the dispersion of the first few Rayleigh overtones in the period range 1-3 seconds and for the two models of oceanic structure described in Table 2. Model O involves no sedimentary layers and a sudden transition from water to crust at a depth of 4 km. Model S includes two transitional layers, which represent the so-called transition zone, and the basement layer. In the absence of any data on the shallow structure of the particular area of the Pacific under study, we chose to adapt Eaton's[1962] oceanic model by incorporating a transition layer, inspired from Spudich and Helmberger's [1979] model H. It should be again emphasized that the purpose of this section is to crudely investigate the influence of such transitional layers on Rayleigh overtones at high frequencies, and not to provide a detailed modelling of the South-Central Pacific Basin.

Spheroidal mode solutions were computed in the range ℓ (angular order) = 3000 - 16000, and T (period) = 1 - 3 seconds, using an eigenfunction function program originally written by R.A. Wiggins. The number of overtone branches computed was $n = 10$ for Model S and $n = 6$ for Model O. These numbers were chosen so as to include all modes with $T \geq 1$ s for $\ell = 10000$. Increments in the value of ℓ between modes computed on a single branch varied substantially depending on the amount of coupling between branches involved [Okal, 1978], from a minimum of $\delta\ell = 10$ to a maximum of $\delta\ell = 1000$.

Table 2. Models used in Normal Mode Investigation

Layer	Thickness km	Density g/cm^3	P-wave vel. km/s	S-wave vel. km/s
MODEL 0				
1. Water	4	1.0	1.485-1.495	0
2. Crust	12	2.7-2.8	6.4-6.6	3.5-3.8
3. Mantle	∞	3.0	8.0	4.6
MODEL S				
1. Water	4	1.0	1.485-1.495	0
2. Transitional	2	1.9	2.5-2.7	0.8
3. Basement	1	2.0-2.05	4.7-4.8	2.0-2.1
4. Crust	9	2.7-2.8	6.4-6.6	3.7-3.8
5. Mantle	∞	3.0	8.0	4.6

Full solutions, including excitation coefficients, were obtained for all modes computed. We will first concentrate on the results concerning angular order (ℓ), period (T) and group velocity (U), presented on Figure 4. It is immediately apparent that a totally different dispersion of the modal energy is involved in the two models. In the pure 'oceanic' model 0 without a transitional layer of sediments, and apart from the fundamental, slow propagating, and poorly excited Stoneley branch, most of the group velocity values are found to lie between 2 and 4 km/s. The only exceptions involve a few 'crossover' periods, such as 1.4 s, for which systematic branch coupling occurs. On the contrary model S, which includes sedimentary layers, has group velocity covering the range 0.6 to 4 km/s more or less uniformly over the period range 1.2 to 2.5 s. This pattern is in much better agreement than the one derived from model 0 with our experimental results, which indicate a wide dispersion of group arrival times at all periods. This proves that anomalous wavetrains are due to efficient coupling of seismic energy into the ocean through the transitional sedimentary layers.

A study of the mode eigenfunctions can give insight into the difference in excitation of comparable modes in both models. For

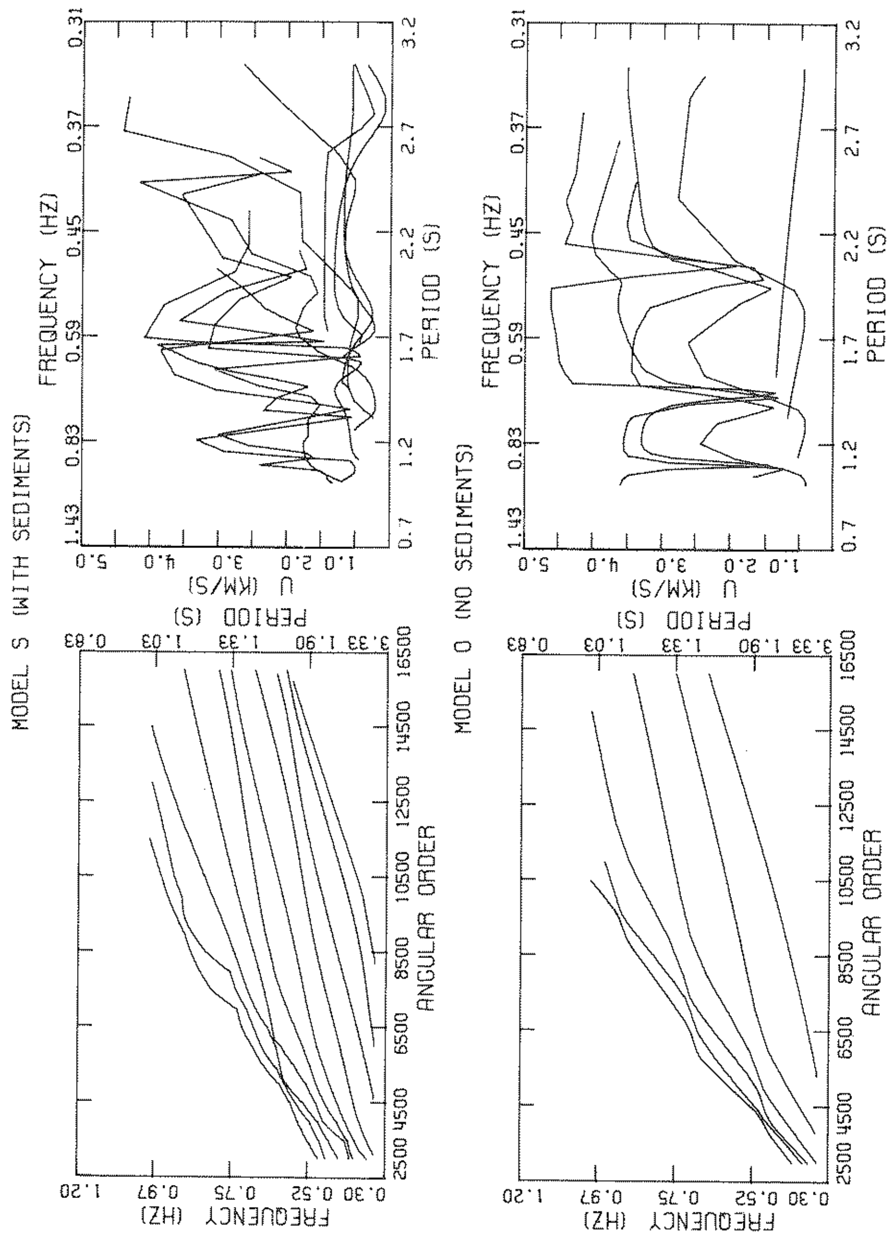

Figure 4 : Normal mode dispersion curves for models S (Top) and O (Bottom). Left: Plot of frequency vs. angular order; Right: Plot of group velocity vs. period.

this purpose, we make use of the excitation coefficients K_0, K_1, K_2 and N_0 [Kanamori and Cipar, 1974; Okal, 1978]. We define

$$K = \sup \{ |N_0| \, \ell^{-0.5}, \; |K_0| \, \ell^{-0.5}, \; |K_1| \, \ell^{0.5}, \; |K_2| \, \ell^{1.5} \},$$

and we focus on the quantity

$$E = I \, K^2 / U^2 ,$$

where I is the Lagrangian integral $I_1 + \ell(\ell+1) I_2$ of the mode involved. From the formalism of Kanamori and Stewart [1976], one easily derives that E is the power spectrum of the energy excited into a single branch of overtones by the most favorable geometry of the moment tensor at the source. A typical comparison of the relevant quantities for a couple of slow propagating modes is given in Table 3. In this example, we investigate modes of very comparable periods and group velocities, which would therefore generate similar contributions to the wavetrain, both in frequency content and group arrival time. The excitation coefficients are computed for a standard earthquake of moment 10^{27} dynes-cm, and a source depth of 7.5 km, corresponding in both cases to a source buried in the crust. The data in this table show that, while values of K are similar in the two models (meaning that surface displacements of the water would be similar), far more energy is channeled into the given overtone branch in the case of the model with transitional layers. Specifically, about 100 times more energy is involved in model S than is in model O. Since the number of overtone branches

Table 3. A Typical Example of the Relative Excitation of Modes in both Models.

	MODEL O	MODEL S
Mode	${}_1S_{10000}$	${}_1S_{16000}$
Period (s)	1.695	1.678
Group velocity (km/s)	0.974	0.955
Lagrangian Integral	1.06×10^{-4}	1.06×10^{-2}
K at 7.5 km (see text)	1.93×10^{-3}	1.69×10^{-3}
Spectral Energy Density ($I K^2 / U^2$)	4.16×10^{-10}	3.32×10^{-8}

covering our area of interest in the T-U plane is also greater in model S, we come to the conclusion that the presence of a well-developed transitional layer favors the excitation of a substantial amount of energy into the seawater. Since the process of conversion of this energy at the receiving shore is a complex one, depending on the particularities of the local topography, the energy power spectrum is probably more representative of the final amplitude of the seismogram than would be the simple displacement at the top of the water column, a quantity commonly used in surface-wave seismology at lower frequencies.

If the source depth is increased, say to 11 km, the difference in energy excitation between our two models is greatly reduced (to a value of only 10 in the example studied above). This reflects the fact that most of the solid Earth is a virtual node for the water-sediment modes involved in anomalous wavetrains. This property strongly suggests that the excitation of anomalous wavetrains requires extremely shallow seismic sources. Although a more quantitative statement will require the systematic use of synthetic seismograms, and therefore a better knowledge of the shallow structure of the area, these results are in qualitative agreement with the suggestions by Okal et al.[1980] that events at GB5 are no deeper than a few km below the water-sediment interface, based on their study of p(n)wP phases regularly generated by these events.

CONCLUSION

In conclusion, we have shown that anomalous wavetrains (observed only along paths involving a smooth bathymetry, and therefore a continuous sedimentary layering), whose energy is concentrated between periods of 1 and 3 seconds, and travels at group velocities of 0.6 to 4.0 km/s, can be interpreted as due to the interference of overtone branches of surface modes. The presence of the transitional layer(s) results in strong coupling of the water column with the upper crust, and in efficient excitation by shallow sources located not more than a few km below the sedimentary layers. The models used in the present study are, of course, very crude, but we hope they will provide a framework for more quantitative studies which could greatly help us understand both the mechanism of the input of seismic energy into the ocean, and the intricate origins of oceanic intra-plate seismicity.

ACKNOWLEDGMENTS

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