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SPECTRAL VARIATION OF THE T PHASE

By
FREDERICK K. DUENNEBIER

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Prepared for

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HAWAII INSTITUTE OF GEOPHYSICS
UNIVERSITY OF HAWAII

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CONTENTS

ABSTRACT

The frequency-time characteristics of the T phase are studied in an effort to isolate the properties of the abyssal T-phase generating mechanism. Early arrivals, or forerunners, are found to have the properties of an abyssally generated T phase. Abyssal T-phase generation is found not to be confined to regions of high latitude where the sofar channel is bounded by the ocean surface, but to also occur where the sofar channel is bounded by a surface channel, as in the equatorial latitudes. Corrections are applied to the observed spectrums to find the frequency characteristics of the abyssal and slope mechanisms. At low frequencies, the slope mechanism is found to be a more efficient generating mechanism than the abyssal mechanism, the two being nearly equal in efficiency at high frequencies. Several possible mechanisms for abyssal generation are discussed, with the conclusion that no satisfactory model for abyssal generation has yet been proposed.

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INTRODUCTION

Johnson et al. (1968) found that abyssally generated T phases (T phases generated in deep water) could be distinguished from slope-generated T phases by their frequency characteristics. New data suggest that forerunners, i.e., low-level signals arriving before some slope T phases, may be abyssally generated. Also, abyssally generated T phases with sources in equatorial latitudes have been identified--although their frequency spectrums differ from those of abyssal T phases from the Aleutian region.

In this paper a study is made of the frequency-time characteristics of the T phase in an attempt to shed some light on the abyssal generation mechanism.

BACKGROUND

The earthquake T phase was first noticed by D. Linehan on a short-period seismograph record (Linehan, 1940). The name, T for Tertiary, was used because the phase followed the primary and secondary waves on many earthquake seismograms; more specifically, the T phase was found on those records where a large portion of the travel path was oceanic. Early discussions concerning the generation and propagation of the T phase were confused due to the large variations in apparent velocity (1.5 to 2.2 km/sec). Tolstoy and Ewing (1950) found that when corrections were made for the ground path from the earthquake hypocenter to the ocean basin and from the ocean basin to the seismometer, the velocity was close to that of the velocity of sound in sea water. They suggested that the T phase is generated at the continental margins where the P wave enters the water as a sound wave and becomes trapped in the ocean sofar channel. Monitoring of sofar channel hydrophones confirmed this theory. In a later paper (1951), Ewing and Press reported that velocities closer to the velocity of sound in the sofar channel were obtained if the arrival time of the most intense point of the phase was used in velocity calculation rather than that of the first arrival.

Early low-level arrivals having apparent velocities greater than the velocity of sound in the ocean are often noticed on hydrophone records of T phases generated by large-magnitude earthquakes. Dispersive properties of sofar propagation (Johnson, 1963; Northrop, 1962), travel paths through the upper sediment layers (Burke-Gaffney, 1954; Leet et al., 1951), and body waves entering the sofar channel from seamounts (Johnson, 1963) have been suggested as possible explanations for low-level forerunners. A suggestion was made by Johnson et al. (1968) that a possible connection between forerunners and abyssal T phases should be studied.

More than 20,000 T-phase source locations were computed from December 1964 through July 1967 from data obtained from seven hydrophone stations in the North Pacific. The majority of these locations are near points where the sloping ocean bottom crosses the sofar channel axis. Johnson et al.

(1964) noted that "*T* phases from earthquakes with hypocenters under deep water were typically weak or not received." In a study of aftershocks of the February 4, 1965 Rat Island earthquakes, Johnson and Norris (1968) found that aftershocks with epicenters at and seaward of the Aleutian Trench generated weak *T* phases with source locations corresponding to the earthquake's epicenter. They recognized that downslope propagation could not be the generating mechanism for these *T* phases. Later, more of these events were found and were termed abyssal or abyssally generated *T* phases; scattering of sound from the ocean surface was proposed as a possible generating mechanism (Johnson *et al*., 1968).

EQUIPMENT AND METHODS

Signals from sofar hydrophones are continuously recorded (frequency modulated) on magnetic tape and on Helicorder paper records by the Pacific Missile Range at four stations in the North Pacific (Eniwetok, Wake, Midway, and Oahu). When an event of interest is noticed on the paper records, the tapes are requested from each station for the time range covering that event.

The frequency-response characteristics of the hydrophone systems vary from one hydrophone to another. The hydrophones themselves are characteristically highly attenuated at low frequencies. Variations in the length of the cables from the hydrophones to the recording stations cause variations in the system responses; the longer the cable, the greater the attenuation of high frequencies. The response of the hydrophones and cable couple such that a system with a long cable has a flatter frequency response than a system with a short cable. The noise level of amplifiers presently in use prevents analysis of signals at frequencies lower than 1 hz. The lack of adequate system-response curves makes analysis of frequencies lower than 5 hz highly questionable.

Frequency-time analysis of the *T*-phase signals is done on a Kay 6061A Sound Spectrograph with the Kay 6070A Contour Display Unit. In normal use, the signals are played into the spectrograph at 160 times the original record speed allowing analysis of frequencies from 0.5 to 50 hz. A single 4- by 12-inch record (sonagram) is a map of frequency versus time with power level contoured in six decibel steps. The sonagrams are not corrected for the hydrophone system frequency response. This correction is applied separately to frequency spectrums (vertical sections through sonagrams). Since frequency response varies with each hydrophone system, sonagrams of the same earthquake recorded at different hydrophones will differ by changes in response as well as by changes caused by differences in the paths to the *T*-phase source.

An important factor in interpretation of a *T*-phase sonagram is knowledge of the source parameters (epicenter, focal depth, magnitude, and bathymetry). To insure the best possible knowledge, only those *T* phases generated by earthquakes listed in the Coast and Geodetic Survey Preliminary Epicenter Determination cards are analyzed. Johnson (1966) has published

published T-phase source locations for the majority of these earthquakes.

OBSERVATIONS AND DISCUSSION

In general, four types of T phases are observed: (1) slope-generated T phases alone, (2) abyssal T phases followed by slope arrivals, (3) abyssally generated T phases alone, and (4) slope-generated T phases preceded by low-level forerunners.

Slope T Phases: Figure 1 is an example of a classical or slope T phase generated by an earthquake in the Rat Islands and recorded at Midway Island (August 31, 1965, 09:46:01 UT, 51.5 N, 175.5 E, magnitude 3.7, depth 33 km). The distance from the epicenter to the hydrophone is 24 degrees. This T phase is typical of the type generated by earthquakes near a continental or island slope. It is characterized by a sharp onset and low (below 10 hz) dominant frequency. The sharpness of the onset and the distinctness of the peak may be attributed to the steepness of the ocean bottom at the T-phase source and to the short distance from the hypocenter to the source. (In this sense, 'T-phase source' refers to the region on the ocean bottom where the T phase is generated. In another sense 'T-phase source' refers to a computed position based on readings of arrival times. Which meaning is implied should be clear from context.) A steeply sloping bottom should produce a sharp T phase as only a short portion of the slope will be in a position to reflect sound energy into the sofar channel. A gentle slope will have a longer horizontal extent over which sofar rays can be produced (Fig. 2). For a given geometry one can compute the limiting depth on the slope for rays to be able to enter the sound channel. This idea was discussed by Johnson et. al. (1964). The distance from the hypocenter to the T-phase source region is a controlling factor in the sharpness of the peak of the T phase. A hypocenter located far from a T-phase source region will insonify a large region with nearly equal energy per unit area; a hypocenter close to a T-phase source will insonify only a small region. The smaller the region insonified, the sharper the T-phase peak will be.

Figure 3 shows the relative levels of energy available along a hypothetical straight coastline for varying distances from the coastline to the earthquake hypocenter. These levels were computed from the formula:

$$L = 10 \log_{10} [h(r^2 + h^2)^{-3/2}]$$

where

L = intensity level

h = depth of focus
r = distance from epicenter to source

given by Johnson et al. (1968). This formula uses spherical spreading and a flat ocean bottom.

Abyssal T Phases: Figure 4 is an example of an abyssal T phase followed by a slope T phase as recorded at an Oahu hydrophone from an earthquake occurring south of the Fox Islands (July 29, 1965, 08:29:21.2 UT, 50.9 N, 171.4 W, magnitude 6.3, depth 27 km). The distance from the hydrophone to the epicenter is 31 degrees. This T phase was studied by Johnson et al. (1968) and used as an example of abyssal generation. The first peak in the figure has a T-phase source corresponding to the earthquake epicenter, the second peak has a source on the slope of the Fox Islands. Two differences between the abyssal and slope T phase are seen in this sonagram: (1) the abyssal T phase is not as strong in the low frequencies as the slope T phase, and (2) the abyssal T phase rises slowly from background noise whereas the slope T phase rises sharply. The lack of high frequencies in the slope T phase is most easily explained by attenuation of body waves in the ground path, the abyssal T-phase source being closer to the hypocenter than the slope T-phase source.

Figure 5 is the sonagram of a T phase off the coast of Oregon as recorded at Midway at a distance of 44 degrees from the epicenter (August 31, 1965, 11:26:23 UT, 4[illegible].3 N, 126.0 W, magnitude 4.3, depth 33 km). In this region the charted water depth, greater than 2500 meters, is many times deeper than the sofar channel axis depth (about 500 meters), thus abyssal generation would be expected to predominate. T-phase sources in this region correspond to the respective earthquake epicenters, not to particular seamounts or radiators (Duennebier, 1968). The similarities between Figure 5 and Figure 1 are striking; both T phases have sharp onsets and maximum strength in low frequencies. Since depth of water prohibits efficient slope generation in a region such as this, any model for abyssal generation should provide for the appearance of events of the type shown in Figure 5. T phases from the Pacific-Antarctic Rise are similar in shape and spectrum to those generated off the coast of Oregon. Water depth along the Pacific-Antarctic rise is also several times the sofar channel depth.

Figure 6 is the sonagram of a slope T phase preceded by a forerunner (Fox Islands, July 2, 1965, 20:58:40.3 UT, 53.1 N, 167.6 W, magnitude 6.7, depth 60 km) recorded at Oahu. The distance from the epicenter to the hydrophone is about 33 degrees. Comparison with Figure 4 shows definite similarities; the forerunner in Figure 6 is weak in low-frequency energy and has a long onset much like the abyssal T phase in Figure 4. Both figures show there was a drop in intensity just before the slope T-phase arrival. Similar features are found on sonagrams of slope T phases preceded by forerunners in regions north of the Volcano Islands on the western rim of the Pacific and north of Washington state on the eastern rim.

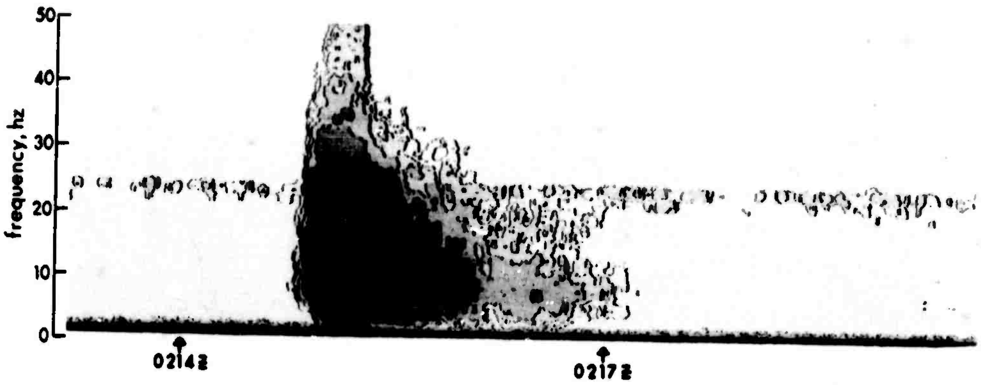


FIG. 1. Sonagram of a slope-generated *T* phase. Note the sharp onset and low peak frequency.

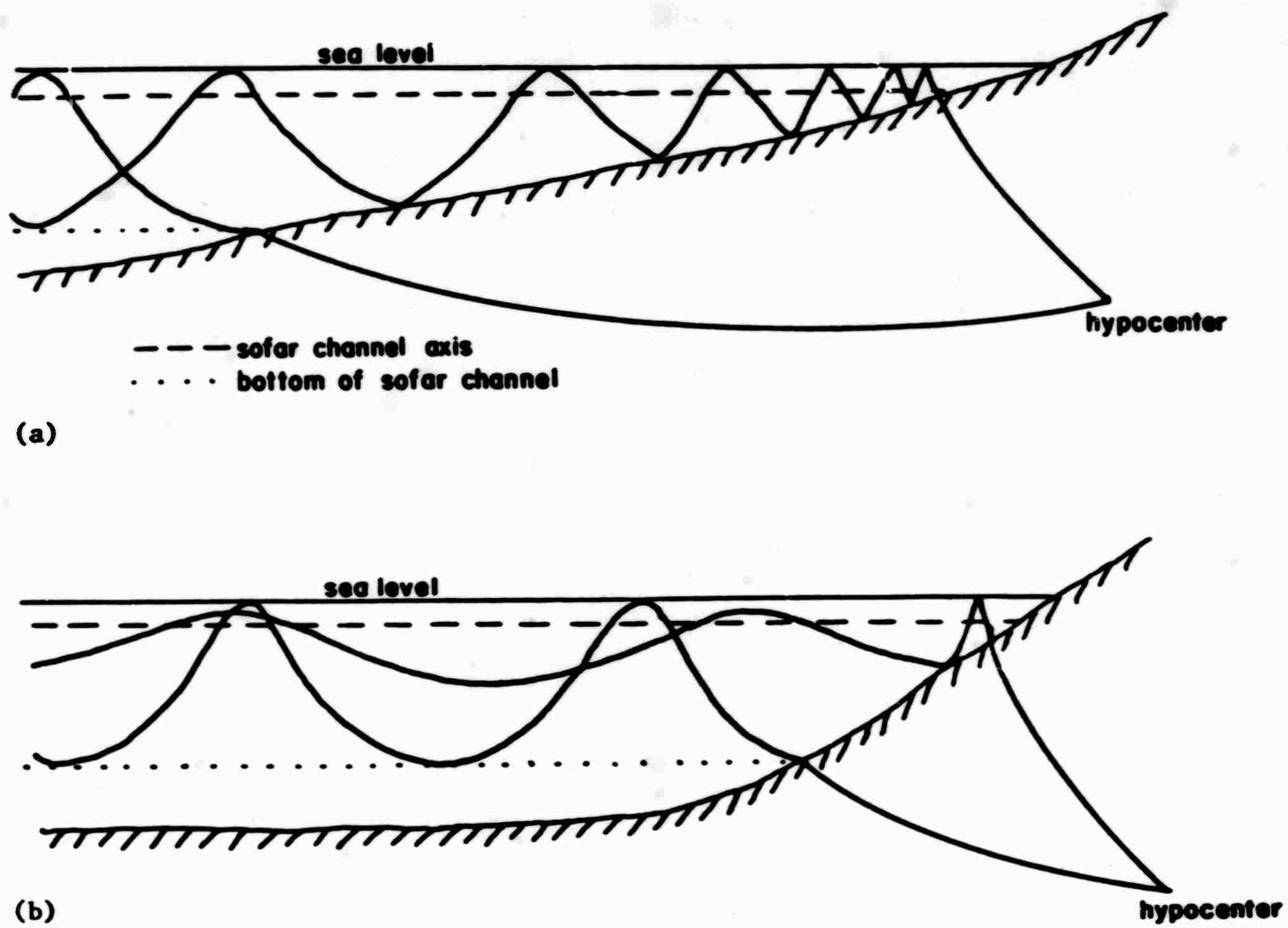


Fig. 2. Downslope propagation of T phases. The length of the generating region increases with decreasing slope of the bottom. A gently sloping bottom (a) produces a broader T phase than does a steeply sloping bottom (b).

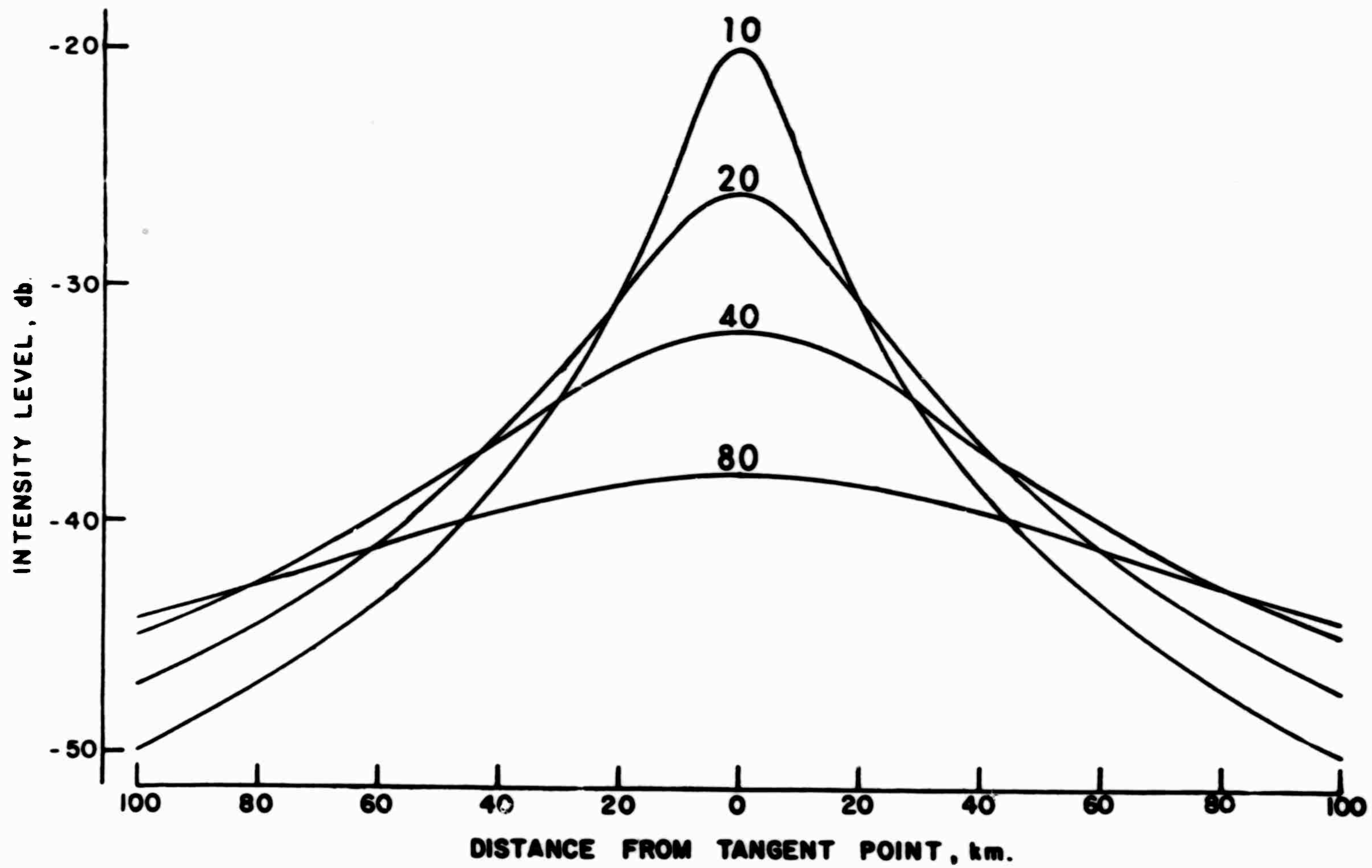


Fig. 3. Intensity level of energy available along a straight line at varying distances (in kilometers) from an earthquake hypocenter.

FIG. 4. Sonagram of abyssal (first peak) and slope (second peak) T phases generated by the same earthquake. Note the gentle onset of the abyssal T phase. The peak frequency of the slope T phase is noticeably lower than that of the abyssal T phase.

FIG. 5. Sonagram of a *T* phase generated off the coast of Oregon.

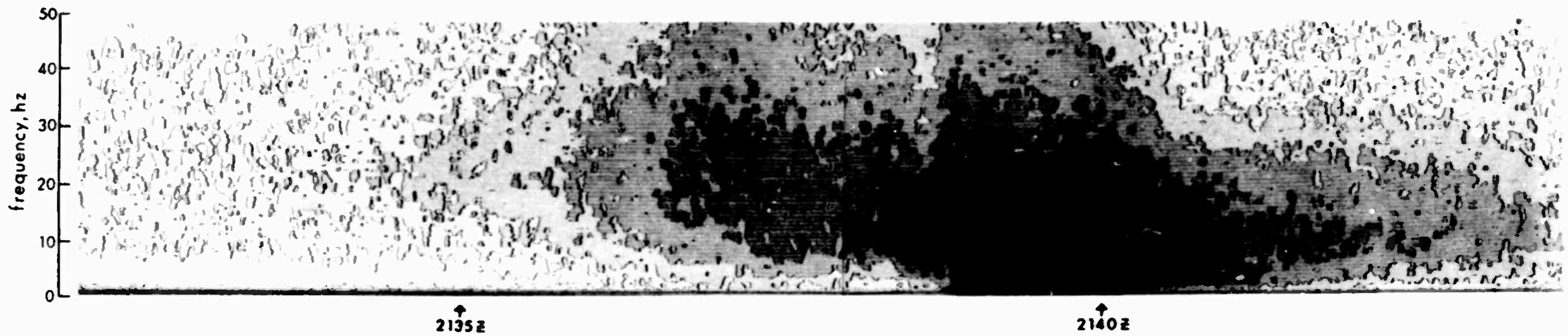


FIG. 6. Sonagram of a slope T phase preceded by a forerunner. Note the similarities between this figure and Figure 4.

Forerunners from the Mariana region are more confused and do not appear to lack energy in the low frequencies. Figure 7 is a sonagram showing the arrivals of the P, S, and T phases from an earthquake in the Mariana Islands (February 10, 1966, 14:21:18.5 UT, 20.8 N, 146.3 E, magnitude 6.2, depth 43 km) as recorded at Eniwetok Atoll at a distance of 18 degrees from the epicenter. Note that the T phase does not have a definite onset and that energy was continuously received at the hydrophone after the arrival of the P wave. The part of the T phase radiated closest to the epicenter arrived at about 1445 UT, although almost equally strong signals were received 5 minutes earlier. Figure 8 is a map of the Mariana area showing arrival times of T-phase signals at Eniwetok referred to the origin time of the earthquake of Figure 7. The strong T-phase arrival near 1445 is at the correct time for the T phase from the epicenter (24 minutes), and the P wave arrival is at the correct time for the P wave travel from the epicenter (4 minutes). The early T phase arrivals appear to have been radiated from the several groups of seamounts as shown in Figure 8. The shallowest of these seamounts are generally twice as deep as the sofar channel axis, thus they should not be efficient slope radiators. From the long duration of this T phase, a deep earthquake focus would be expected (Fig. 3), however the Coast and Geodetic Survey reported a depth of only 43 km.

Deep Hydrophones: In an effort to find out how much T phase energy travels in paths which are not restricted to the sofar channel, hydrophones mounted on the deep-ocean bottom were monitored. Two observations were made from this study. Signals received at deep hydrophones are generally much weaker (nearly 35 db lower) than sofar channel arrivals. The dominant frequency of deep hydrophone arrivals is lower than those of sofar channel arrivals; higher frequencies (greater than 20 hz) are lost in ocean background-noise. If energy in abyssally generated T phases travel in paths which are not confined to the sofar channel (RSR and surface-bottom reflected paths), a relative increase in strength of abyssal T phases when compared to slope-generated T phases would be expected when observed on deep hydrophones. No relative increase has been observed; thus it can be concluded that slope and abyssal T phases travel in nearly the same paths.

Frequency Spectrum: Up to this point no corrections have been applied to the observed spectrums. In an effort to isolate the spectral characteristics of the abyssal and slope mechanisms, corrections to the observed signal were applied for the initial spectrum near the earthquake focus, ground-path attenuation, and frequency response of the hydrophone system.

The near-source spectrum for earthquake waves is poorly known. Most seismic studies are concerned with lower frequencies than are observed in T phases and measurements are most often made far from the earthquake epicenter. To approximate the near-source spectrum, a sonagram of body-wave arrivals from a local (Oahu) earthquake was studied. Since the water path to the bottom-mounted hydrophone was zero, and the ground path was less than 20 km (the P and S waves arrived nearly simultaneously), the spectrum was taken as a standard for near-source seismic waves (Fig. 9).

Since a vertical section through the sonagram of Figure 9 already included the effects of the hydrophone system, the hydrophone correction was also included; applications to other hydrophones need only include relative differences between hydrophone systems.

Attenuation of seismic body waves in the earth is measured by a parameter Q, the reciprocal of the attenuation per wavelength. Sutton et al. (1967) found values of Q between 200 and 1000 across the continental United States where

$$Q = 2\pi f/(2.3Ub)$$

where

f = frequency (hertz)
U = group velocity (km/sec)
b = attenuation (db/km)

If Q in the regions studied is assumed to be within the limits 200 - 1000, an attenuation of from 2 db/1000 km (Q = 1000) to 11 db/1000 km (Q = 200) are expected from absorption at a frequency of 40 hz (U = 6.5 km/sec). If spreading of body waves is approximated as spherical, an additional loss given by L (db) = 20 log D_2/D_1 is suffered in the ground path, where L is the level difference between distances D_1 and D_2 from the hypocenter. The effect of attenuation caused by absorption of seismic waves in the ground path is reflected in greater signal loss at high frequencies as distance from the hypocenter increases. On T-phase sonagrams, the part of the T phase with the highest level at high frequencies should be that part of the T phase which was generated closest to the hypocenter. Loss of energy in the ground path caused by spreading of seismic waves is not frequency dependent, and thus causes lower levels as distance from the hypocenter increases without changing the frequency spectrum. However, since the relative efficiencies of T-phase radiation are variable for different points on the ocean floor, the effects of absorption and spreading in the ground path may be considerably masked.

Attentuation of T phases does not appear to be the same as that of explosives detonated in the sofar channel. T phases corrected for attenuation from the source to the hydrophone using the attenuation coefficient at 10 hz for sofar bombs of 1.6 db/1000 km (Urick, 1963) gain in strength as the distance from the source to the hydrophone increases, thus T-phase attenuation is less than 1.6 db/1000 km. In this paper no correction has been made for attenuation of T phases in the ocean.

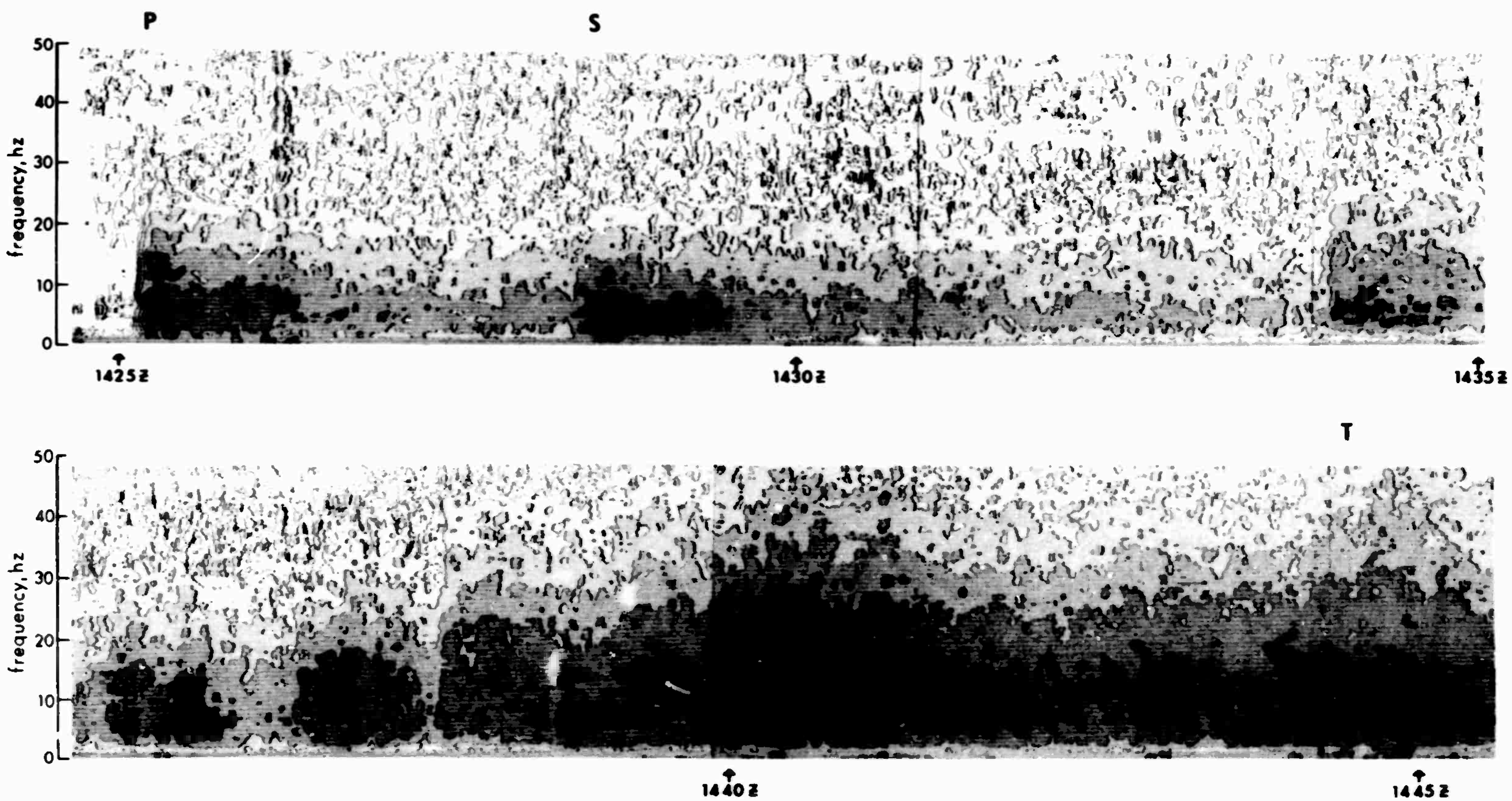


FIG. 7. Sonagram of *P*, *S*, and *T* phases from an earthquake in the Mariana Islands.

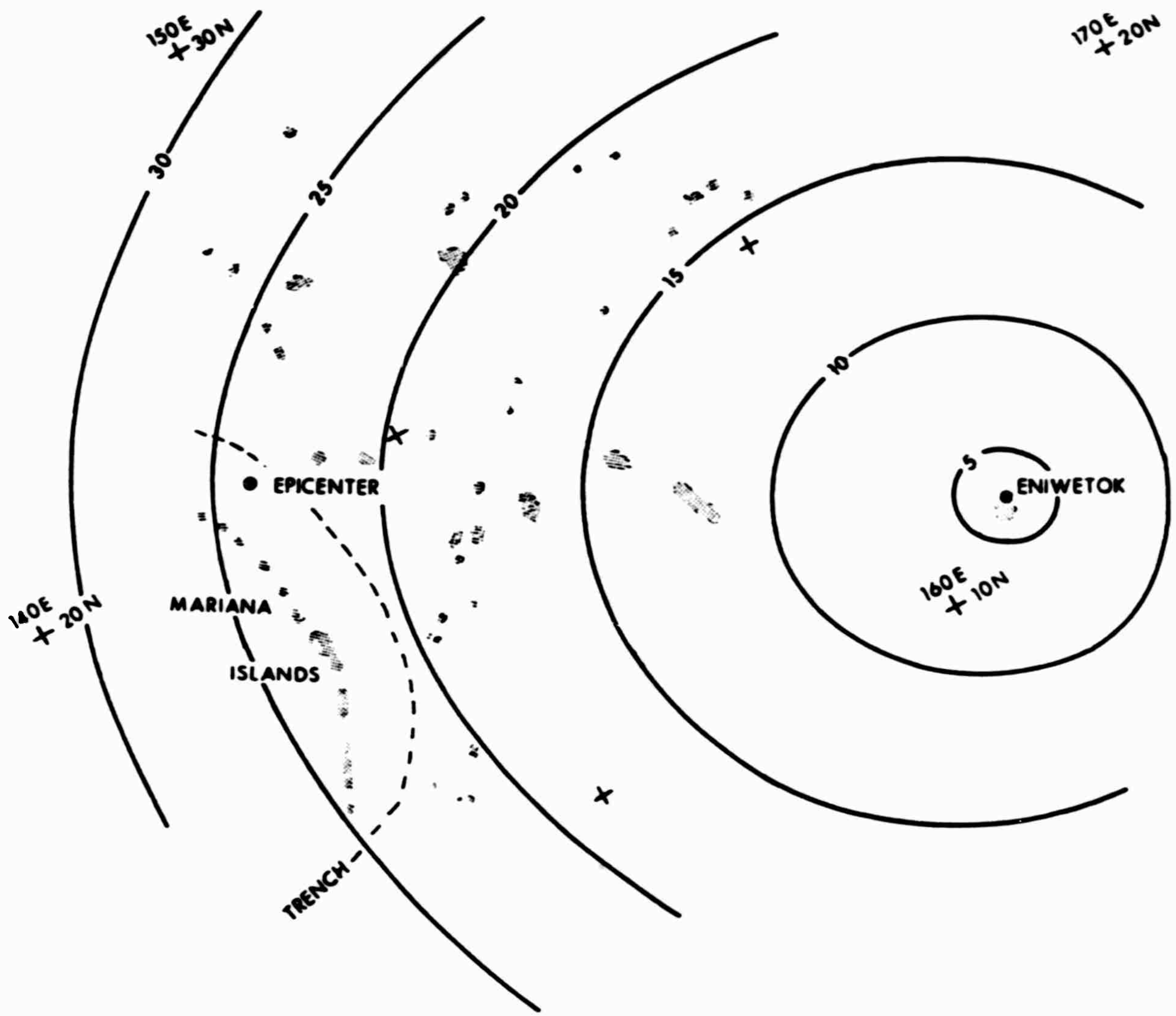


Fig. 8. Map of Mariana Islands region. The contours show the arrival time at Eniwetok, in minutes after the origin time, for the P to T waves generated by the earthquake shown in Figure 7. The shaded areas are regions where the ocean depth is less than 1000 fathoms. The arrival times of the stronger peaks of Figure 7 correspond to arrivals from these shaded areas (seamounts).

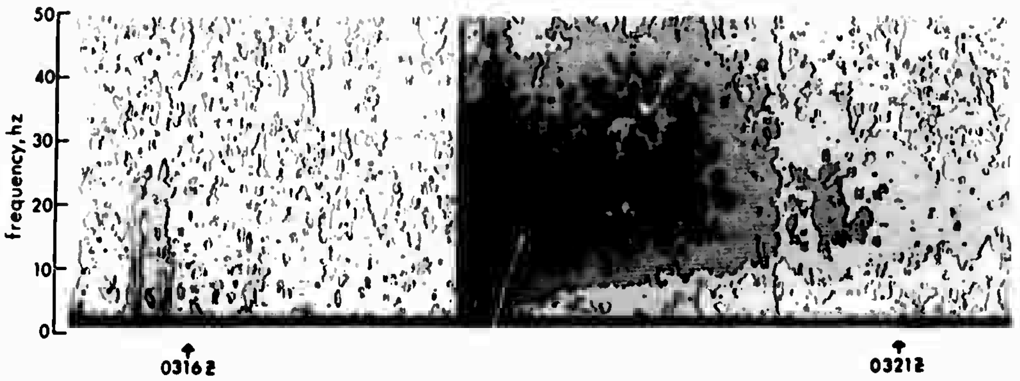


FIG. 9. Sonagram of body waves from a local earthquake recorded on Oahu.

Figure 10 and 11 show various spectrums corrected for seismic wave spectrum, ground-path attenuation, and hydrophone response. The earthquake data for each of the spectrums is given in Table 1. Figure 10 is a comparison between the corrected spectrums of abyssal and slope-generated T phases from the Aleutian region. The levels are adjusted to be equal at 5 hz. The spectrums show that the transfer functions of the two generating mechanisms differ greatly; the slope mechanism favors low frequencies and the abyssal mechanism favors high frequencies. Figure 11a shows pairs of spectrums from earthquakes which generated both abyssal and slope T phases. A ground-path-spreading correction has been added along with the other corrections so that the spectrums may be compared in relative intensity level as well as in shape. In some situations it appears that the abyssal mechanism is more efficient than the slope mechanism in the high frequencies. Figure 11b shows corrected spectrums of T phases generated in regions other than the Aleutian region. As seen, these spectrums are more confused than the spectrums from the Aleutians, and the generating mechanism is not so easily recognized.

MECHANISM FOR ABYSSAL GENERATION

Any mechanism proposed for abyssal generation should conform with the features of observed abyssal T phases. These features include (1) less efficiency than the slope generation mechanism at low frequencies (less than 20 hz) and approximately equal efficiency at higher frequencies, (2) propagation in the sofar channel, (3) a suggestion of time symmetry in the signature of the arrivals, and (4) occurrence in regions where water depth is great in comparison to the sofar channel depth.

The theory of slope generation of T phases is easily handled by ray-path studies. A ray refracted into the ocean at a sloping bottom will be projected into a near-horizontal path after several reflections at the ocean surface and bottom. As the generating slope becomes less steep, more reflections are required before a horizontal path is achieved. If the ray reaches a horizontal path within the vertical limits of the sofar channel, it becomes a sofar channel ray. Abyssally generated T phases are generated in regions where the depth of water is greater than the sofar channel depth and where the ocean bottom is relatively flat. Ray theory does not account for the observation of T phases generated in these regions.

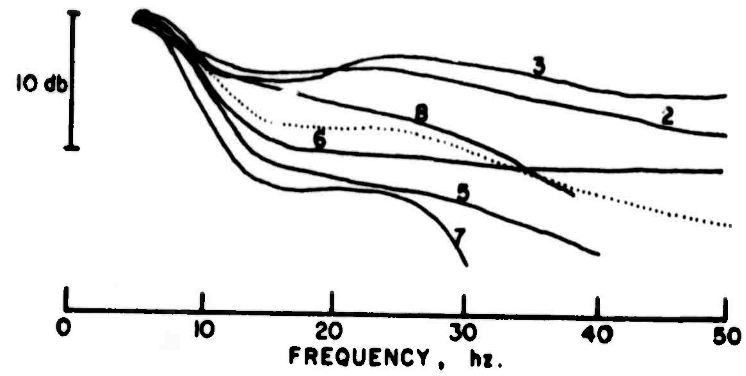
Johnson et al. (1968) suggested scattering of the sound energy by the ocean surface as a mechanism which might account for the spectrum and shape of abyssally generated T phases. Such a scattering mechanism could account for the lack of energy at low frequencies and the generally symmetric shape of the signatures.

Theory of scattering from the ocean surface has been treated by many authors; usually, however, the treatment is limited by the approximation that the wavelength of the ocean-surface wave is much longer than that of the sound waves. For T-phase frequencies, the wavelengths of ocean-surface

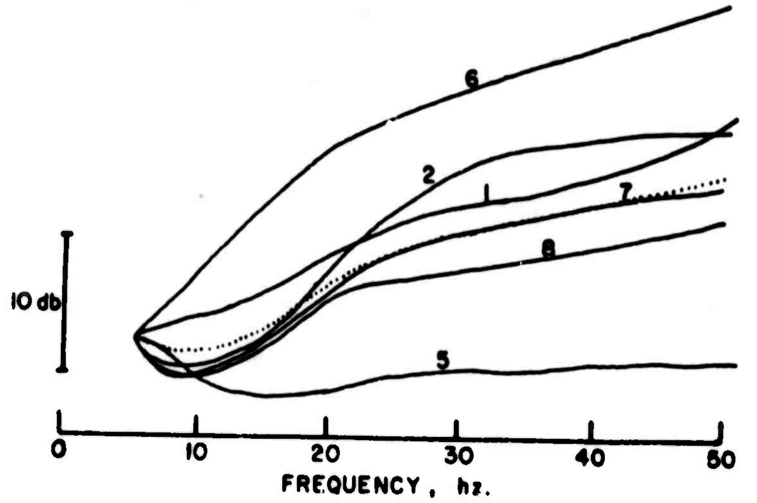
Table 1. Earthquake Data

No.*	Date	Time, GMT	Latitude	Longitude	Magnitude	Depth, km	Location
1.	02 July, 1965	205840.3	53.1 N	167.6 W	6.7	60	Fox Is.
2.	29 July, 1965	082921.2	50.9 N	171.4 W	6.3	22	Fox Is.
3.	31 August, 1965	094601	51.5 N	175.5 E	3.7	33	Rat Is.
4.	31 August, 1965	112623	43.3 N	126.0 W	4.2	33	Off Oregon
5.	01 October, 1965	085204.4	50.1 N	178.2 E	6.3	32	Rat Is.
6.	28 January, 1966	223812.2	51.6 N	157.0 E	5.6	107	Kamchatka
7.	30 May, 1966	095438.3	50.1 N	176.6 E	5.0	30	Rat Is.
8.	07 August, 1966	021305.1	50.6 N	171.3 W	6.5	39	Fox Is.
9.	19 November, 1966	164733	55.1 S	129.4 W	4.9	33	South Pacific
10.	18 October, 1967	231345	44.2 N	129.0 W	4.4	33	Off Oregon
11.	28 April, 1968	041815.7	44.8 N	174.5 E	5.5	39	North Pacific
12.	28 April, 1968	062302	44.8 N	174.7 E	4.3	33	North Pacific
13.	04 May, 1968	175246.2	26.5 S	115.4 W	5.3	34	Easter Isl. Cord.

* Refers to the curves shown in Figures 10 and 11.



(a)



(b)

Fig. 10. A comparison of spectrums of slope-generated (a) and abyssal (b) T phases from the Aleutian region. The numbers correspond to the event numbers listed in Table 1. The dotted line represents the average for each set of curves.

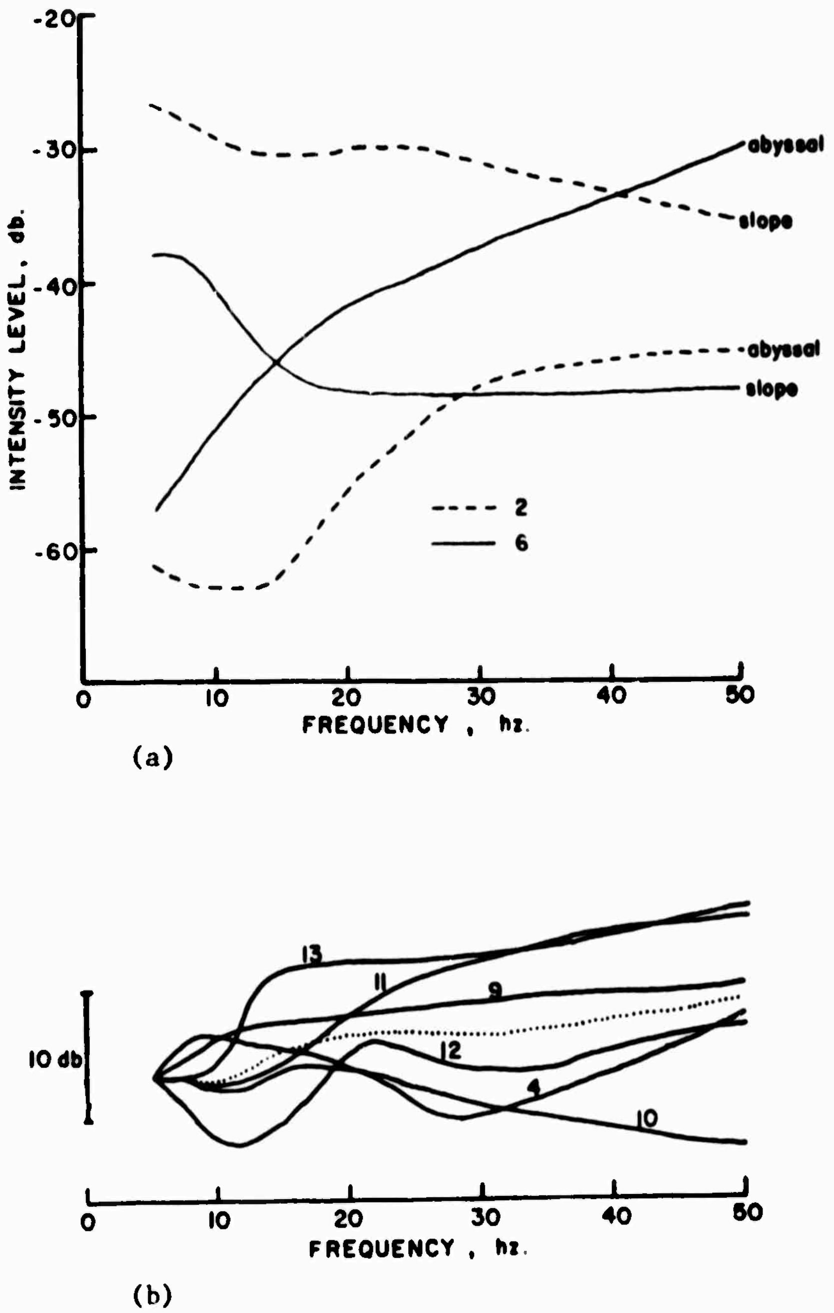


Fig. 11. (a) A comparison of the relative strengths of abyssal and slope T phases. The two events shown represent the observed extremes in the differences in efficiency of the two generating mechanisms. The numbers correspond to the event numbers listed in Table 1. (b) Spectrums of abyssal T phases from non-Aleutian regions. The average curve (dotted line) is similar to the average curve shown in Figure 10 (b). The numbers correspond to the event numbers listed in Table 1.

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waves are equal to or less than the sound wavelengths.

Rayleigh (1945) treated the case of an acoustic wave from a pressure release surface. He found that a normally incident acoustic wave will be reflected without scattering if the wavelength of the corrugated surface is less than that of the sound wave. For short sound-wavelengths (compared to that of the corrugation), the amplitude of scattered spectral orders is found by inversion of an infinite matrix; the number of real scattered spectrums is determined by the ratio of the two wavelengths. The amplitude of the last (most highly scattered) order is smallest, the succeeding amplitude coefficient being imaginary.

A solution to the wave-scattering problem for statistical surfaces developed by Marsh (1961) was used to solve the problem of signal loss from scattering at the ocean surface (Marsh et al., 1961). The abyssal T-phase problem is just the complement, i.e., finding the amount of signal scattered into paths where the sound can be received by a sofar channel hydrophone. From Marsh, the signal-loss per ocean-surface bounce is computed by the formula:

$$(\text{db/bounce}) = -10 \log \left(1 - 0.023\,(fh)^{3/2}\right)$$

where f is the frequency in kilohertz, and h is the rms wave height in feet. For a very rough surface (h = 10) at a frequency of 0.01 khz, the loss per bounce is 0.0035 db. If a non-divergent beam of 0 db level were normally incident at the ocean surface, then the specular reflection level would be -0.9965 db and the upper bound on the scattered radiation level would be -31 db.

E_i = incident energy

E_r = reflected energy

E_s = scattered energy

$$10 \log E_i = 0$$

$$E_i = 1$$

$$10 \log E_r = -0.0035$$

$$E_r = 0.9992$$

$$E_i = E_r + E_s$$

therefore

$$E_s = 0.0008$$

$$10 \log E_s = -31$$

The scattered radiation will spread in all directions; only a small portion of this energy, if any, will reach sofar channel paths. In high latitude regions, such as the Aleutian Islands region, the sofar channel axis is near the surface. Highly scattered energy travelling in refracted-surface-reflected (RSR) paths may enter sofar channel paths as it travels south and the sofar channel axis deepens. However, for sources in lower latitudes, where the sofar channel is not bounded by the ocean surface, surface-scattered rays do not enter sofar channel paths. Energy travelling in RSR paths is detected by sofar channel hydrophones, however to propagate in an RSR path, energy normally incident on the ocean surface at the _T_-phase source would have to be scattered at angles high enough so that contact with the ocean bottom would be prevented by refraction. Observation of deep hydrophones shows that little energy is travelling in paths which reach the ocean bottom. It seems doubtful from current scattering theory that ocean-surface scattering of sound could be the mechanism for abyssal generation.

Johnson _et al_. (1968) were able to generate synthetic intensity signatures similar to observed abyssal _T_-phase signatures using a model which transforms a constant fraction of the seismic energy arriving per unit area at the ocean bottom into abyssal _T_-phase energy. Such a model suggests a scattering mechanism for abyssal generation. However, while the model may predict intensity signatures, it does not predict anything about the frequency spectrums.

Another scattering mechanism mentioned by Johnson _et al_. (1968) is scattering by volume inhomogeneities in the ocean. Large variations in temperature or salinity over a small volume of ocean would change the refractive index of the water and thus act as a possible scattering device. Urick (1963) postulated that scattering by volume inhomogeneities accounts for high attenuation coefficients for signals from 10 to 1000 hz in the ocean sofar channel. According to his work, attenuation of signals near 20 hz is about 2 db per 1000 km. Scattering by similar inhomogeneities over a 4-km path from the ocean bottom to the surface would be negligible. Scattering angles high enough to scatter vertically travelling sound into near-horizontal paths would require extremely large temperature or salinity gradients. Thus, as Johnson _et al_. (1968) concluded, scattering from volume inhomogeneities would not seem to account for abyssal generation.

Johnson _et al_. (1968) felt that a mechanism involving the coupling of seismic surface waves with the ocean sofar channel would not cause the long onset observed in abyssal _T_ phases, nor would it favor higher frequencies. Biot (1951) theorized that Stoneley waves (boundary waves coupled between the ocean water and the bottom sediments) may couple efficiently with the sofar channel at frequencies where the velocity of the Stoneley wave is close to the velocity of sound in water. For the first Stoneley-wave mode, the phase and group velocities are about 0.95 the velocity of sound in the ocean at T-phase frequencies. The amplitude potential decreases exponentially from the bottom to the surface in the first mode. For all higher modes, the phase and group velocities are almost equal to the velocity of sound in water, the velocity increasing with decreasing frequency, possibly causing less efficient coupling in lower frequencies. Davies (1965) observed Stoneley waves which had been produced by an explosion on the ocean bottom and recorded by an ocean-bottom seismograph. The frequencies of the observed arrivals were between 3 and 10 hz.

The main objection to a mechanism involving Stoneley waves is that the source of energy is not close to the ocean-bottom interface. For maximum efficiency in boundary-wave generation, the seismic source should be close to the interface where the displacement is maximum. Abyssal T-phase sources may be tens of kilometers from the earthquake hypocenter. A possible answer may be that body waves arriving at the ocean-bottom interface may cause secondary sources for Stoneley wave generation. This could explain the symmetric shape of abyssal T phases generated on flat ocean floors.

Another clue to the mechanism of abyssal generation may be the apparent dispersion noticed on T-phase records. There appears to be a continuous gain in efficiency at lower frequencies as shallower water is approached. As indicated by the Mariana event, bottom roughness also appears to be a factor in efficiency of abyssal generation.

A comparison of the amount of energy available at the T-phase source with the amount of energy in the T phase would give the absolute efficiency of the generating mechanisms. Unfortunately, a comparison of this sort is not feasible at this time due to the lack of knowledge of the energy in the body waves and to the variability of characteristics of T-phase sources. It is well known that large amounts of energy from earthquakes do enter the ocean; Northrop and Raitt (1963) have recorded seismic wave levels greater than 100 dynes per sq. cm. from earthquakes recorded at sea. Many ships have reported severe shaking as if they had run aground, and sometimes damage from vibrations has later been correlated with known earthquakes (Birch, 1966). In several instances the epicenter was hundreds of kilometers from the ship location. Such high intensities may cause effects in the water column, as yet unknown, which are responsible for the generation of abyssal T phases.

CONCLUSIONS

From the observations reported in earlier papers and in this paper, several conclusions can be reached concerning the properties of T phases.

1. The most noticeable variations in the T-phase intensity signature and spectrum are caused by the bathymetry of the T-phase source region and by the position of the hypocenter with respect to the ocean bottom.

2. The shape and spectrum are not appreciably modified by changes in range from the hydrophone to the T-phase source or by changes in the earthquake magnitude, although changes in magnitude do cause changes in the T-phase strength.

Very large magnitude earthquakes (magnitude 7 or more) generate T phases which are noticeably different in shape and duration from the T phases generated by smaller earthquakes. These changes appear to be an effect of an extended earthquake source region (Johnson and Norris, 1968).

Two mechanisms appear to be important in T-phase generation, slope generation and abyssal generation. Slope generation has been explained previously, and is observed to be the most efficient mechanism for T-phase generation. Abyssal generation has not been fully explained and is defined only by the observation of T phases in regions of flat ocean-bottom in deep water where slope generation cannot occur. The mechanism for abyssal generation is still in doubt, although a possibility is the coupling of Stoneley waves with the sofar channel.

The frequency characteristics of the two types of T phases are usually different enough that they can be told apart. Abyssally generated T phases from the Aleutian region have less strength in the low frequencies than do slope T phases. In regions where the bottom topography is more rugged, the difference between the two types is less obvious.

SUGGESTIONS FOR FURTHER STUDY

Close observation of Figures 1, 4, and 6 shows that in slope-generated T phases there is a tendency for high frequencies to arrive earlier than low frequencies. Johnson, 1963 proposed that such dispersion is caused by normal mode propagation effects at the generating slope. He studied a T phase similar to that shown in Figure 6, also from an earthquake in the Fox Islands, the only difference being that the epicenter (52.4 N, 168.5 W) was in deeper water than that in Figure 6. With new observations of abyssal T phases, it appears that there may be some connection between abyssal generation and observed apparent dispersion in slope-generated T phases. There is some indication that efficiency of T-phase generation at low frequencies increases as water depth decreases.

As an experiment to study the abyssal mechanism, a large explosion could be detonated on the deep-ocean floor. The signal intensities received at various depths and distances should show the paths followed by the sound energy and give an estimate of the efficiency of the abyssal mechanism.

Knowledge of the mechanisms of T-pahse generation would be enhanced if accurate data were available at frequencies lower than 5 hz. It would be advantageous to have a low-noise hydrophone-amplifier system with a flat (or at least well—known) response from 0.1 to 100 hz.

Further frequency analyses of T phases and knowledge of the bathymetry at their epicenters coupled with an intensive source location program should be able to isolate the effects of bottom roughness and water depth on abyssal generation. More data are needed before the rarer phenomena associated with T phases (such as rapid frequency and intensity shifts) can be explained.

Once the frequency characteristics and efficiencies of generation of the T-phase are determined, the phase should be useful in determination of the near-source spectrum and energy of earthquake body waves. Computed sources of abyssally generated T-phases are currently valuable in seismicity studies since the sources correspond in position to the earthquake epicenters. A study of abyssal T phases from mid-ocean rises would help to increase knowledge of the seismicity of these regions.

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13 ABSTRACT

The frequency-time characteristics of the T phase are studied in an effort to isolate the properties of the abyssal T-phase generating mechanism. Early arrivals, or forerunners, are found to have the properties of an abyssally generated T phase. Abyssal T-phase generation is found not to be confined to regions of high latitude where the sofar channel is bounded by the ocean surface, but is found to also occur in southern latitudes where the sofar channel is bounded by a surface channel.

14. KEY WORDS	LINK A		LINK B		LINK C	
	ROLE	WT	ROLE	WT	ROLE	WT
T Phase						
Sofar						
Epicenter						
Earthquake						

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