Seismology - T phase

SPECTRAL VARIATION OF THE

T PHASE

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ABSTRACT

The frequency-time characteristics of _T_ phases in the Pacific Ocean 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 abyssally generated _T_ phases. Abyssal _T_ phases from regions where the sofar channel axis is deep are observed. A _T_ phase generated by an earthquake with an extended source is studied; the fault length computed from the _T_ phase is an order of magnitude greater than the expected length. Corrections are applied to observed frequency spectrums to find the frequency characteristics of the abyssal and slope mechanisms. The slope mechanism is found to be more efficient than the abyssal mechanism in the low frequencies; the two mechanisms being nearly equal in efficiency at high frequencies. Several possible mechanisms for abyssal generation are discussed (scattering from the sea surface, coupling with Stoneley waves, and others) with the result that no satisfactory mechanism has yet been found.

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INTRODUCTION

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 it followed the Primary and Secondary waves on many earthquake seismograms; more specifically, the _T_ phase was found on those records where a large part of the travel path was oceanic. The early discussions concerning the generation and propagation of the _T_ phase were confused due to the large variations in the apparent velocity (1.5 km/sec to 2.2 km/sec). _Tolstoy and Ewing_ (1950) found that when corrections were made for the length of the ground path from the epicenter to the ocean basin and from the ocean basin to the seismometer, the _T_ phase velocity was close to that of sound waves traveling in the ocean sound (SOFAR) channel. 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, or propagates by normal mode across the ocean. Monitoring of the sofar channel with hydrophones confirmed the sofar propagation theory. In a later paper (_Ewing and Press_, 1951), it was found that _T_-phase velocities even closer to the velocity of sound in the ocean were obtained if the arrival time of the most intense point of the phase was used in velocity calculation rather than the first arrival.

Early arrivals having apparent velocities faster than the velocity of sound in the ocean, are often noticed on hydrophone records of _T_ phases generated by large magnitude earthquakes. These low-level forerunners have been explained as being due to dispersive properties of sofar propagation, travel paths through the upper sediment layers

(Leet et al., 1951, Burke-Gaffney, 1954), and as body wave entering the sofar channel from seamounts (Johnson, 1963).

More than 20,000 T-phase source locations were found from December 1964 through July 1967 by continuously monitoring of sofar hydrophones at seven stations in the North Pacific. The majority of these locations are at points where the sloping ocean bottom crosses the sofar channel axis at the rim of the Pacific basin. Johnson et al. (1963), noted that T phases from earthquakes which had epicenters under deep water were weak or not received. In a study of aftershocks of the February 4, 1965 Rat Island earthquake, Johnson and Norris (1968) found that aftershocks with epicenters in and seaward of the Aleutian trench generated weak T phases which had source locations which corresponded to the epicenters. They recognized that down slope propagation could not be the mechanism for generation of these T phases. Later, more of these events were found and termed abyssally generated or abyssal T phases; scattering from the ocean surface was postulated as a mechanism for their generation (Johnson, Norris, & Duennebier, 1968).

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

Acknowledgements

This research was funded by the Advanced Research Projects Agency through contract Nonr 3748(01) with the Office of Naval Research. The University of Hawaii Water Resources Research Center allowed the use of the Kay Sound Spectrograph for production of the sonagrams.

Suggestions and advise from Rockne H. Johnson and Roger A. Norris of the Hawaii Institute of Geophysics have been both helpful and inspiring.

EQUIPMENT AND METHODS

Signals from sofar hydrophones are continuously recorded on magnetic tape (frequency-modulated) 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 the event.

In this study, the primary interest is in frequency analysis of the T phase. This is done on a Kay 6061A Sound Spectrograph with the Kay 6070A Contour Display plug-in unit and Kay 6076 Scale Magnifier. This unit delivers a four by twelve inch paper display, called a sonagram, of frequency (85-8000 cps) vs. time (2.4 seconds per record) with seven power level contours in six decibel steps. In normal use, T-phase signals are played into the spectrograph at 160 times the original record speed allowing analysis of frequencies from 1-50 cycles per second. Approximately 400 signals have been analyzed. The sonagrams presented in this paper are not corrected for hydrophone system response.

This type of analysis has several advantages over other methods of spectral analysis. Earlier work was done by playing the signals through several filters and observing the relative amplitudes on chart paper (Johnson, 1963). While this method supplies the same results as the spectrograph, more work is involved in interpretation. Earlier spectrographs did not have the contouring units and lacked sufficient dynamic range to find frequencies of peak power; such an instrument was used by Northrop (1962). Digital analysis was tried and found to be about three orders of magnitude more costly than the method currently in use, aside from involving more interpretation and time. Digital analysis also has the problem of sampling rates and aliasing, which analog techniques do not have. More expensive analog analyzers were tried with the result that the added flexibility was unnecessary for this work.

An important factor in interpretation of the sonagram is knowledge of the source of the signal. 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 being analyzed. The majority of these also have published T-phase source locations (Johnson, 1966). Comparison of these locations with bottom topography yields information on the source characteristics of the T phase.

OBSERVATIONS

Abyssal generation is known to take place where the water depth is greater than the depth of the sofar channel, such that there exists no ray path which can leave the ocean bottom and directly enter a sofar channel path. A distinction is made in this paper between abyssal _T_ phases and abyssal forerunners. Both abyssal _T_ phases and abyssal forerunners are abyssally generated, the difference being that abyssal _T_ phases are generated by earthquakes which have hypocenters under the deep ocean floor, while abyssal forerunners are generated by earthquakes occurring under land or shallow water. Slope generation refers to the transfer of seismic energy from the ground to the sofar channel from regions where rays from a sloping bottom may enter a sofar channel path (Figure 1). Down slope propagation has been known as a mechanism for _T_ phases generation for some time. A complete explanation of down slope propagation can be found in _Johnson et al._ (1963).

Unlike earthquake sources, which in most cases can be described by a point, (hypocenter), _T_ phase sources are regions of the ocean floor. Before the discovery of abyssally generated _T_ phases, it was thought that a _T_ phase source could be approximately described as one dimensional; the source being along a line where the ocean bottom crosses the sofar channel axis. Since it was thought that _T_ phases could not be generated in deep water, early arrivals were explained by _Northrop_ (1962), and _Johnson_ (1963) as being due to dispersion or travel paths through the sediments. The discovery of abyssally generated _T_ phases placed the problem in two dimensions, that is, a _T_ phase may be generated in any region where a sofar channel exists.

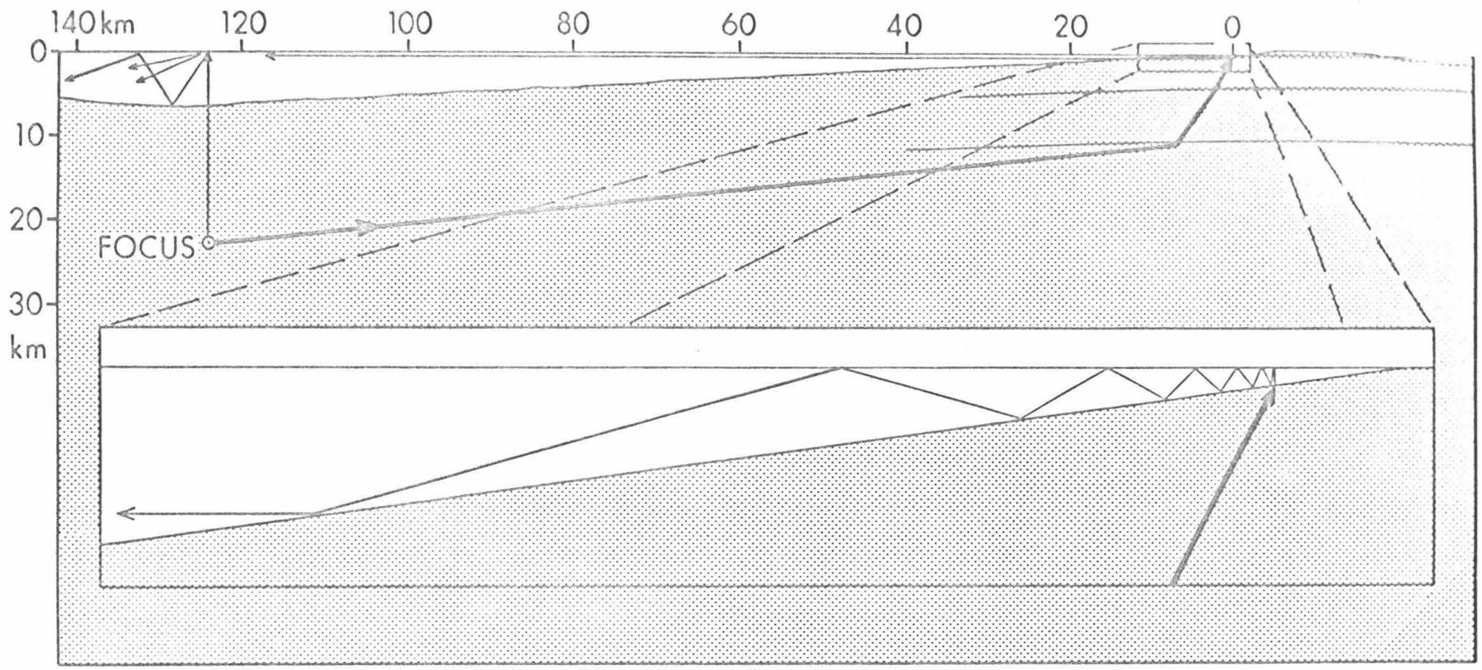

Fig. 1. Downslope or slope generation of the _T_ phase: A seismic wave refracted into the ocean at a sloping bottom is reflected into a near-horizontal path. (From _Johnson et al._, 1967)

In general, four types of signatures are seen in records of _T_ phases: slope _T_ phases, slope _T_ phases preceded by forerunners, abyssal _T_ phases, and abyssal _T_ phases followed by slope _T_ phases. The type of _T_ phase observed is a function of the position of the epicenter with respect to the ocean margin. An earthquake occurring under land or shallow water will produce a slope _T_ phase. If the magnitude is great enough, the earthquake may also generate a noticeable low level arrival (forerunner) before the onset of the slope _T_ phase. If an earthquake occurs under the deep ocean, an abyssal _T_ phase will be generated. If a shallow water region is near, a slope _T_ phase may also be generated by the same earthquake. _T_ phases which are reflected from a land mass before arriving at a receiver are generally weak or not observed. Figure 2 shows hypothetical _T_ phase intensity signatures together with bottom profiles of an ideal source area. The relative intensities of the abyssally generated arrivals and the slope generated arrivals of Figure 2 are based on a relative intensity difference of 11 db for abyssal and slope _T_ phases obtained by _Johnson and Norris_ (1968).

Figure 3 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 z, 51.5 N, 175.5 E, M 3.7, h 33 km). The distance from epicenter to hydrophone is 24 degrees. This _T_ phase is typical of the type generated by earthquakes with hypocenters near a continental or island slope where an efficient _T_ phase radiator is present. The slope _T_ phase is characterized by a sharp onset and an apparent center frequency of less than ten cycles per second. The sharpness is directly related to the slope of the ocean bottom at the source and to the distance from the hypocenter to the source. A steep slope should generate a

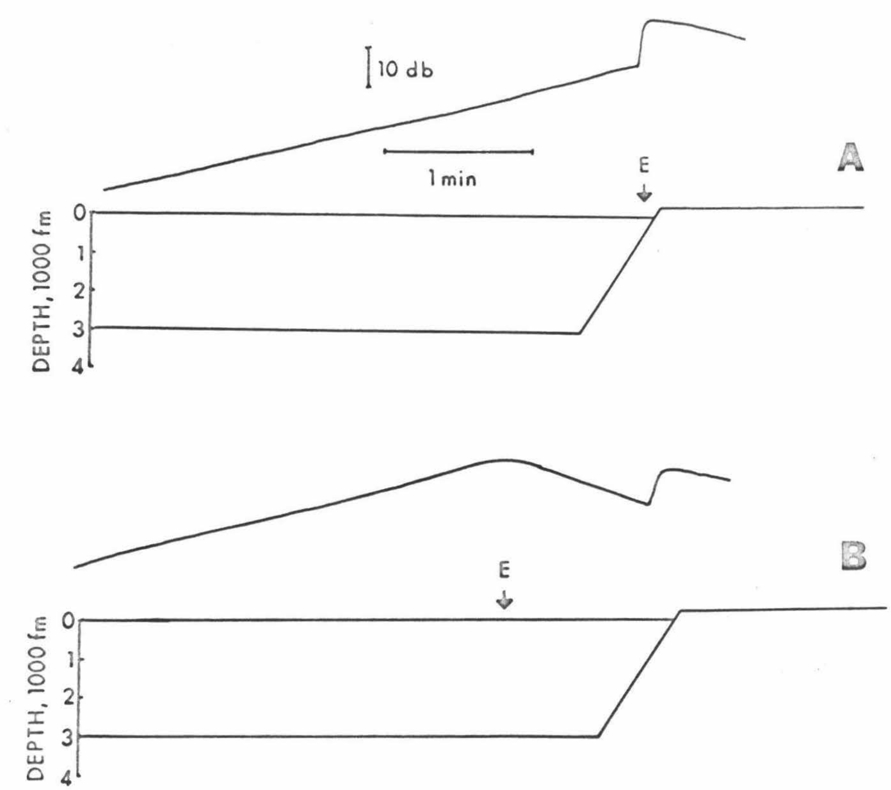

Fig. 2. Expected intensity signatures shown with hypothetical ocean bottom profiles. For Figure 2A the epicenter is in the deep ocean, and for Figure 2B the epicenter (E) is at the ocean margin.

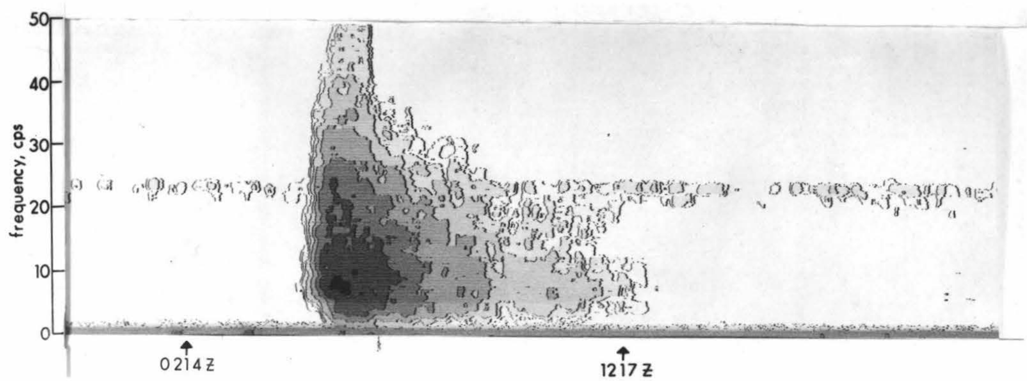

Fig. 3. Sonagram of a slope generated T phase. Note the sharp onset and low frequency of peak intensity.

sharper _T_ phase than a gentle slope. A steep slope should also be more efficient in _T_ phase generation than a gentle slope since fewer reflections with the surface and bottom are required for a steep slope before a sofar channel path is achieved. Each reflection with the ocean bottom at angles of incidence less than the critical angle causes loss due to transmission into the ocean bottom. _Johnson and Norris_ (1966) found that _T_ phases radiated from the same slopes had similar signatures and efficiencies. The hypocenter for this earthquake was under Tahoma Reef, which is an area noted by _Johnson and Norris_ (1968) as being a strong _T_ phase radiator.

The duration and sharpness of the arrival depend upon the distance from the hypocenter to the sloping bottom as well as on the steepness of the slope. The variation caused by the distance from the hypocenter to the slope is the result of spreading of the P wave in the ground. Figure 4 shows the expected intensity of the P wave (assuming spherical spreading in the ground) along various lines normal to a line through the hypocenter. It is obvious that the strongest _T_-phase arrival should have a source closest to the hypocenter, although slope _T_ phases generated far from the hypocenter will have a less noticeable peak.

The angle of incidence of the P wave at the ocean bottom should be considered in estimating the efficiency of slope generation, and may also be a factor in the content of the frequency spectrum. The efficiency of generation is the ratio of the seismic energy available at the ocean bottom to the amount of energy in the _T_ phase. Refraction of seismic waves in the earth's crust causes most waves to arrive at the ocean bottom near normal incidence; at normal incidence, the amount of energy transmitted into the ocean is greatest. A ray refracted into the ocean

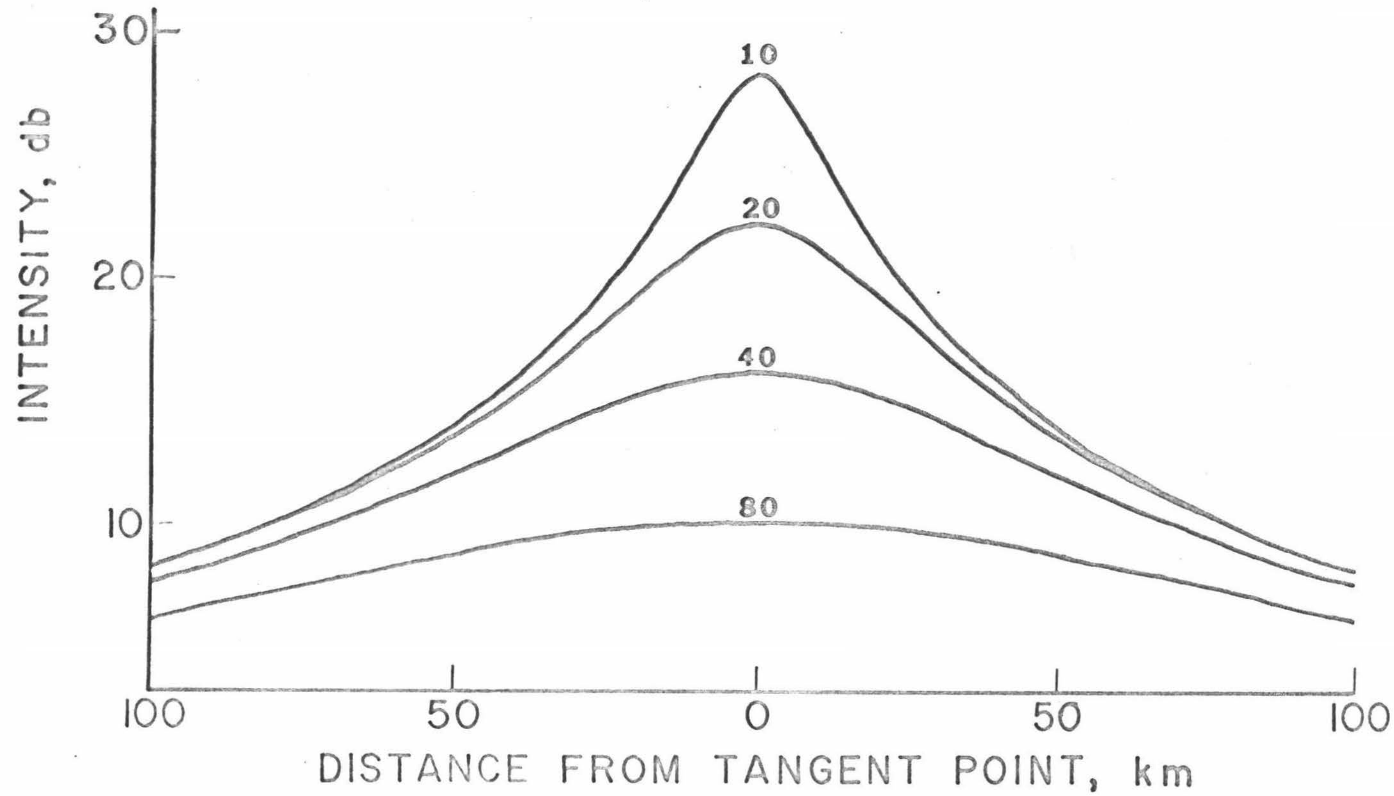

Fig. 4. Intensity variations measured along lines tangent to spheres centered at the earthquake focus caused by spherical spreading of seismic waves in the earth. The tangent points are at 10, 20, 40, and 80 km from the focus.

heading up slope, such as from an earthquake occurring seaward of the slope, will be reflected back down slope after several contacts with the ocean surface and bottom. The greater number of reflections required should cause a greater loss of energy due to absorption into the ocean bottom.

Figure 5 is the sonagram of a slope generated _T_ phase accompanied by a low level forerunner (Fox Islands, July 2, 1965, 20:58:40.3 z, 53.1 N, 167.6 W, M 6.7, h 60 km) recorded at Oahu. The distance from the epicenter to the hydrophone is 33 degrees. It is interesting that this earthquake generated a small tsunami (_Iida et al._, 1967), although from the low intensity and lack of aftershocks it would not appear to be tsunamigenic. The first peak, or forerunner, has been the cause of much confusion and is worthy of discussion. Early workers in _T_ phase study may have chosen the onset of the forerunner for computation of _T_ phase velocity. When the distance from the epicenter to the receiver is used in the velocity calculation, erroneously high velocities are obtained. The low level forerunner was studied by _Northrop_ (1962), who interpreted it as being due to dispersion caused by normal mode propagation. He concluded from frequency analysis that the forerunner was a low frequency arrival. _Johnson_ (1963) studied a _T_ phase originating in the same area as the _T_ phase of Figure 5. While the first peak in the event he studied also lacked low frequencies, he treated it as a slope _T_ phase rather than as a forerunner. The interest in his paper was in the apparent dispersion, that is, the peak intensity in the high frequencies arriving sooner than the peak intensity in the low frequencies. His hypothesis that this was due to dispersion caused by normal mode propagation was based on theory for a flat ocean bottom. It is unlikely that the theory is valid for a

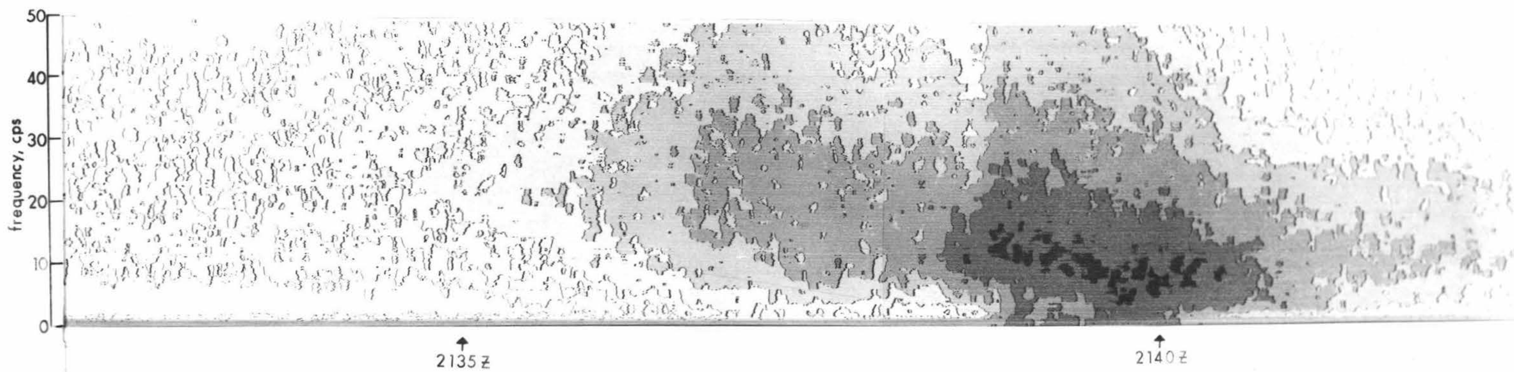

Fig. 5. A slope T phase from the Fox Islands preceded by a forerunner. Note the similarities of this sonagram and Figure 6.

sloping bottom as is found in this region. It can be seen from Figure 5 that the forerunner is not low frequency as suggested by Northrop (1963), but in fact lacks low frequencies and has an apparent peak near twenty cycles per second. This lack of low frequencies is a criterion for identifying abyssally generated T phases (Johnson et al., 1967) on hydrophone records. It appears that the forerunner is also abyssally generated, and that the apparent dispersion noted by Johnson (1963) may be explained by abyssal generation. The intensity of the forerunner in Figure 5 does not rise continuously until the slope arrival as expected from Figure 2, but drops in intensity before the slope arrival. This phenomenon is often noticed in forerunners.

Figure 6 is an example of an abyssal T phase followed by a slope arrival recorded at Oahu from an earthquake in the Fox Islands (July 29, 1965, 08:29:21.2 z, 50.9 N, 171.4 W, M 6.3, h 22 km). The distance from the hydrophone to the epicenter is 31 degrees. This T phase, was studied by Johnson et al. (1967), and used as an example of abyssal generation. The epicenters for the earthquakes of Figures 5 and 6 are both in the Fox Islands, the hypocenter for Figure 6 was under the Aleutian trench, that for Figure 5 was under the island slope. The differences noted between the two sonagraphs are a more sharply defined peak in the abyssal section (first peak) of Figure 6 than in Figure 5, less power in the high frequencies and less pronounced apparent dispersion in the slope arrival of Figure 6. The difference in apparent dispersion may be due to changes in the slope from which the two slope T phases were generated. The differences in frequency content of the slope T phases may have been caused by ground path attenuation of the high frequencies, the ground path to the slope being longer in Figure 6. The difficulty with the

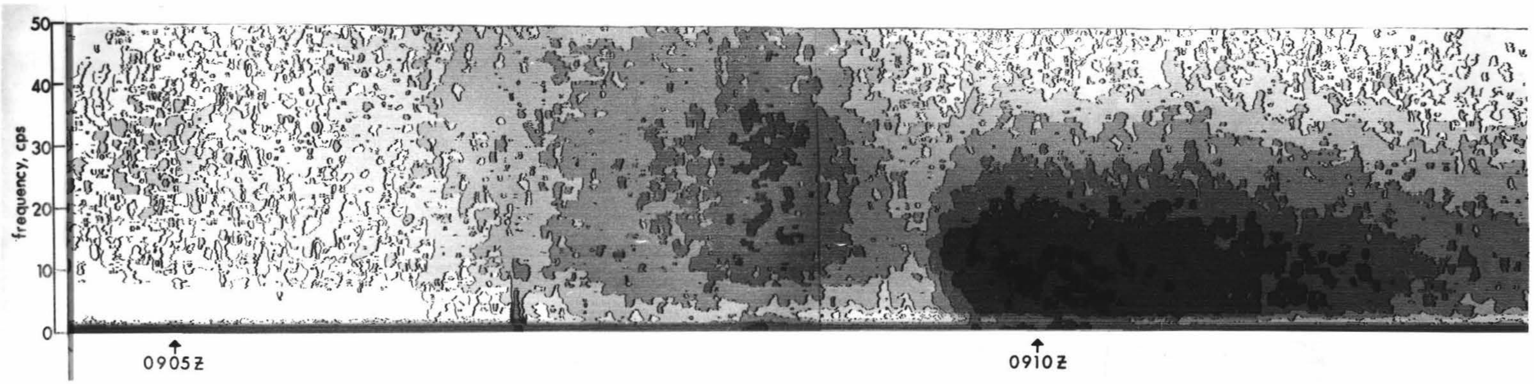

Fig. 6. Sonagram of an abyssal T phase followed by a slope T phase, both generated by the same earthquake. Note the apparent lack of power in the low frequencies in the first (abyssal) peak, and the lack of power in the high frequencies in the second (slope) peak.

hypothesis that ground path attenuation is the cause of the lack of high frequencies is that an equal lack of high frequencies is not seen in the abyssal forerunner of Figure 5 where the ground path was also long. The apparent peak frequency of the abyssal section of Figure 6 is, however, noticeably higher than that of Figure 5. A criterion for choosing the arrival time of the energy radiated into the sound channel from the source closest to the epicenter may be to choose that point which contains the greatest intensity at high frequencies. For earthquakes generating only slope _T_ phases, this should also be the point of maximum total intensity, but since abyssal generation does not appear to be as efficient as slope generation in the low frequencies, an abyssal _T_ phase may have a lower total intensity than an accompanying slope _T_ phase and yet still have higher intensity than the slope _T_ phase at high frequencies.

Johnson et al. (1967) suggested that abyssal _T_ phase signatures should be nearly symmetric in time, however, few symmetric _T_ phases are found. Variations from the expected symmetry are probably caused by topographic effects and changes in the depth of water. Examples of observed signatures compared with bottom profiles at the source are shown in Figure 7. To match the horizontal scale of the bottom profile to the time scale of the _T_ phase record, the horizontal scales of the bottom profiles have been modified by the formula

$$t = \pm \frac{D}{v_t} + T(D)$$

where t is the arrival time of the _T_ phase generated at a distance d from the epicenter, v_t is the _T_ phase velocity, and T(D) is the Jefferys-Bullen p wave travel time from a distance D.

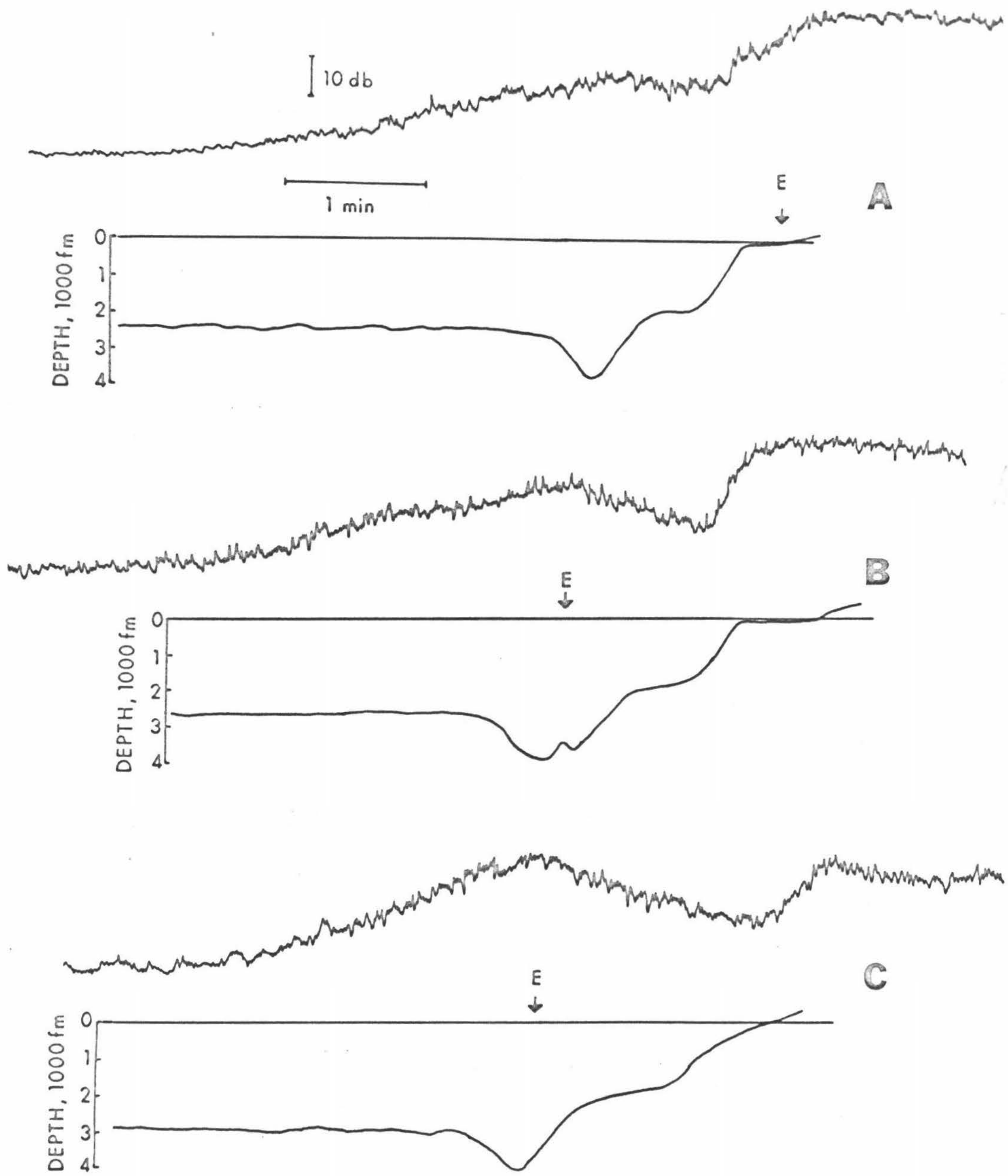

Fig. 7. Comparison of observed intensity signatures with ocean bottom profiles near the source. The point 'E' is above the approximate position of the epicenter in each case. Horizontal distances on the profiles have been modified to correspond with time on the intensity records (see text). Curves A, B, and C correspond to the sonagraphs of Figures 4, 5, and 11A respectively.

In Figure 7 the profiles were aligned such that the epicenters correspond to the points of highest intensity in the high frequencies in the _T_ phase. With the exception of 7C, the slope _T_ phases correspond to the slopes of the bottom profiles. The slope _T_ phase of 7C arrives later than expected possibly because of an error in the location of the epicenter by the Coast and Geodetic Survey. _Duennebier and Johnson_ (1967) found that abyssal _T_ phase sources in this region (Rat Islands) were generally 40 km south of the reported earthquake epicenters. By shifting the epicenter 40 km to the south, the slope arrival corresponds to the slope of the profile. Comparison of Figure 2 with Figure 7 shows the differences between expected and observed signatures. Note that the forerunner of 7A reaches peak intensity at a time corresponding to the arrival generated in the trench, and the intensity drops before the slope arrival. The slope _T_ phases of B and C are more intense than expected from an 11 db difference in efficiency between the abyssal and slope mechanism.

Figure 8 is the _T_ phase of an earthquake which occurred off the coast of Oregon (August 31, 1965, 11:26:23 z, 43.3 N, 126.0 W, M 4.3, h 33 km) as recorded at Midway at a distance of 44 degrees from the epicenter. In this region the water depth, greater than 2500 meters, is many times deeper than the sofar channel axis, about 500 meters, thus slope generation is not efficient and abyssal generation should predominate. _T_ phase sources in this region correspond to the respective earthquake epicenters as abyssal _T_ phases should, not to particular radiators. However, the signatures of many _T_ phases in this region, such as the one shown in Figure 8, have the characteristics of slope _T_ phases, sharp rise times and apparent center frequencies lower than center frequencies of

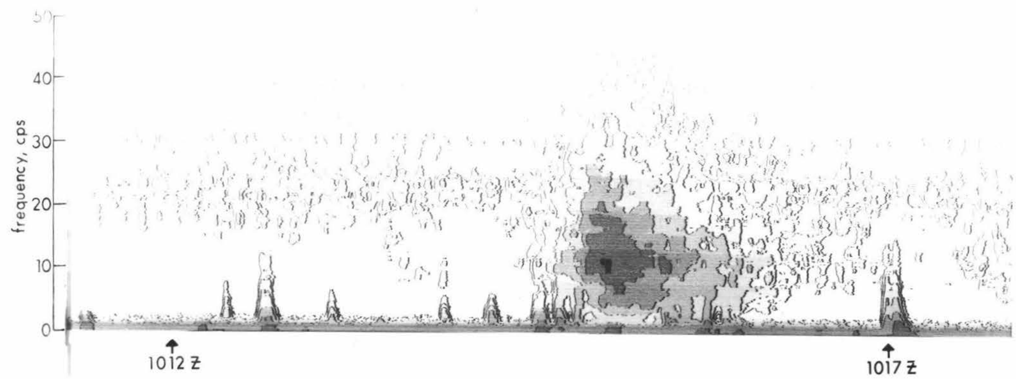

Fig. 8. Sonagram of an abyssal T phase from the East Pacific Rise: Although the depth of water in the source area is several times the depth of the sofar channel axis, the observed T phase resembles a slope T phase more than an abyssal T phase (compare with Figure 3).

abyssal T phases from the Aleutians. This is also seen in other regions, such as the Pacific-Antarctic Rise, where the water depth is great compared to the sofar channel depth. The deviation in the signature from that observed in the Aleutian area may be the effect of rugged bottom topography on the East Pacific Rise, the deeper sofar channel axis in low latitudes, or the absence of the trench and island arc structure present in the Aleutians.

That topography is important in abyssal generation is shown by Figure 9. Figure 9A is a sonagram showing the arrivals of the P, S, and T phases from an earthquake in the Marianas Islands (February 10, 1966, 14:21:18.5 z, 20.8 N, 146.3 E, M 6.2, h 43 km) as recorded at Eniwetok Island at a distance of 18 degrees from the epicenter. Note that the T phase does not have a definite onset, but energy is received at the hydrophone continuously after the arrival of the P wave. It is postulated that this represents energy radiated into the sound channel by the P wave, and possibly the S wave as it traveled towards the hydrophones. The lack of a sharp peak in the main T phase is probably the result of the arrival of energy radiated from several island slopes in the Marianas at nearly the same time as the arrival of the T phase from the epicentral region. Unfortunately the bottom topography in this region is rugged an a comparison with specific features would have little meaning since the azimuth from the hydrophone to the origin of a particular arrival is unknown, though the distance can be computed. Figure 9B 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 9A. The strongest, highest frequency arrival on Figure 9A corresponds to the T phase travel time to the epicenter (24 minutes), and the P wave arrival

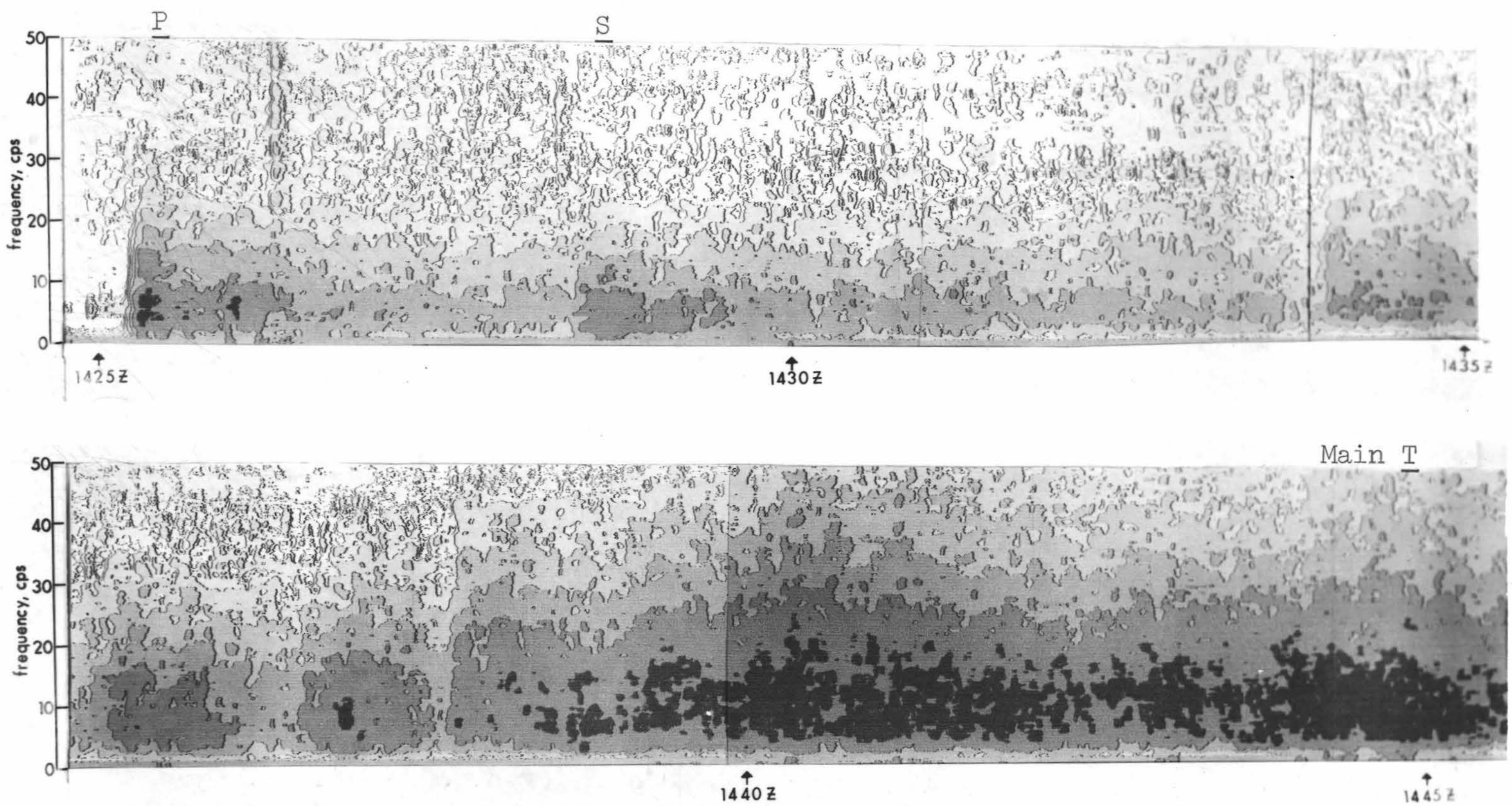

Fig. 9A. P-wave, S-wave, and T-phase arrivals at Eniwetok from an earthquake in the Mariana Islands. The two sonagrams form one continuous record.

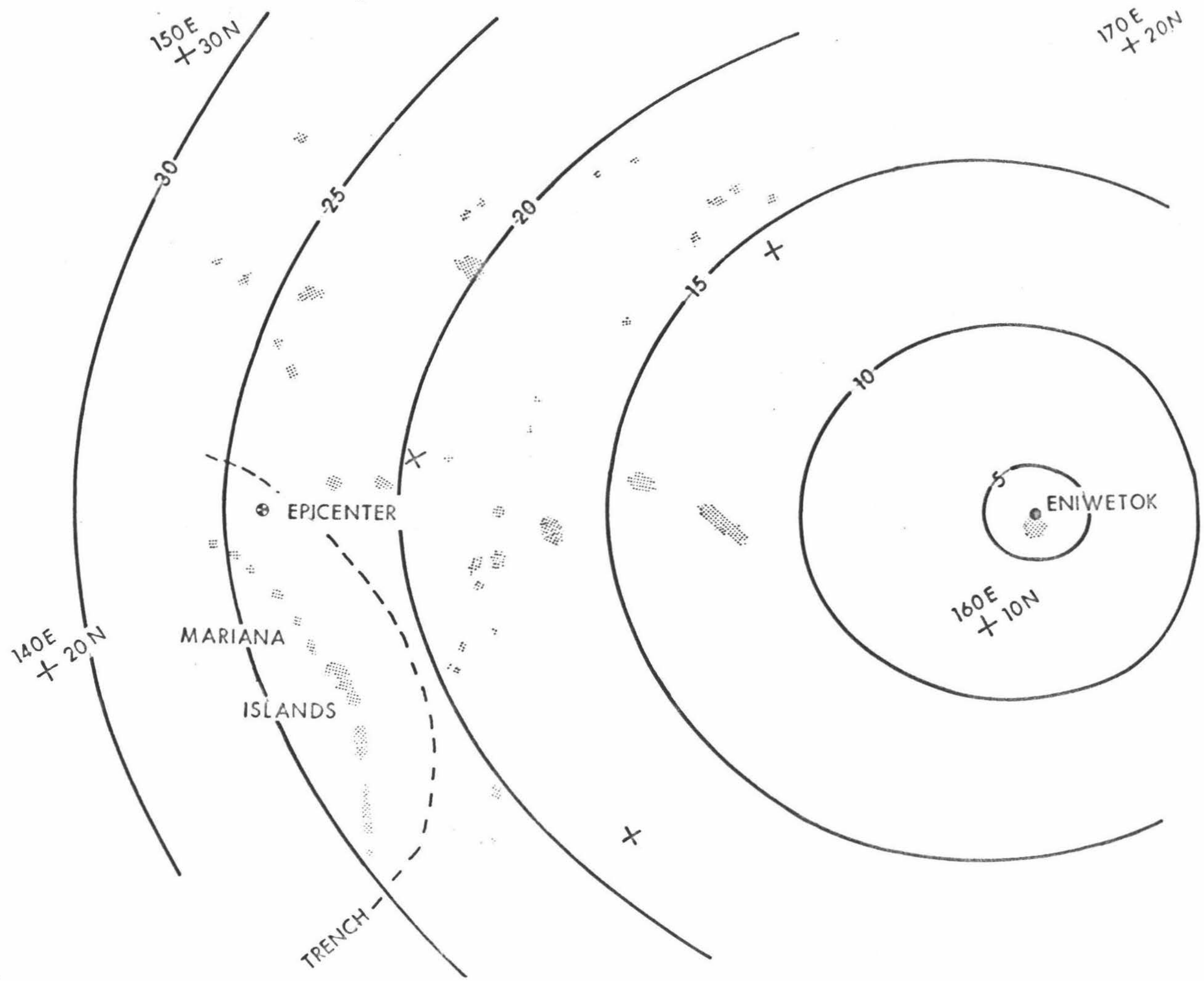

Fig. 9B. Map of Mariana Islands - Eniwetok region. Solid contours are travel times to Eniwetok for P to T phases generated by the earthquake of Figure 9A. Shaded areas are seamounts rising to within 1000 fathoms of the sea surface.

at Eniwetok corresponds to the P wave travel time from the epicenter (4 minutes). The early T phase arrivals appear to have been radiated from the several groups of seamounts shown on Figure 9B. The shallowest of these seamounts are generally twice as deep as the sofar channel axis, thus they should not be efficient slope radiators, yet the T phases generated are almost equal in strength to the T phase radiated at the epicentral region. This is probably the result of a combination of abyssal and slope generation in the seamount region. Note that the depth of this earthquake (43 km) as reported by the Coast and Geodetic Survey does not agree with the rate of decrease of intensity with distance of the T phase source from the epicenter (Figure 4). From the observed rate of decrease of intensity with distance one would expect a much greater depth of focus.

The relative efficiencies of slope and abyssally generated T phases were discussed by Johnson and Norris (1966). They found that the intensity of abyssal T phases were generally 11 db lower than slope T phases generated by earthquakes of similar magnitude when measured through a 15 cycle per second low pass filter. Since more energy is present in frequencies above than below fifteen cycles per second in abyssal T phases, and more energy is present below than above fifteen cycles per second in slope T phases, the total intensity difference should be somewhat lower than 11 db. Observation of the frequency spectra shows that the difference in efficiency is not as noticeable in the high frequencies (above twenty cycles per second) as in the low frequencies. In some regions it appears that the abyssal mechanism is more efficient than the slope mechanism in the high frequencies.

Figure 10 is the T phase of an earthquake which occurred under the Kamchatka Peninsula (January 28, 1966, 22:38:12.2 z, 51.6 N, 157.0 E, M 5.6, h 107 km) recorded at Midway. The distance from the epicenter to the hydrophone is 31 degrees. The abyssal forerunner for this T phase is stronger in the high frequencies than the slope arrival, though the source of the slope arrival is closer to the hypocenter than the source of the forerunner. The method of finding the T phase source closest to the epicenter by choosing the time when the highest power in the highest frequencies arrives is not accurate in this case. This T phase was used by Johnson et al. (1967) as an example of the effect of depth of focus on the abyssal T phase signature. However, the hypocenter is under the Kamchatka Peninsula thus, this is not an abyssal T phase, but a forerunner. Many forerunners are easily confused with abyssal T phases because of the drop in intensity of the signal before the slope arrival.

In the observation of abyssally generated T phases, it is important to observe the event normal to the shoreline nearest to the source of the event. If this is not done, the abyssal arrivals may be masked by slope generated T phases arriving at the same time. Figure 11 is the T phase of a Rat Island earthquake (October 1, 1965, 08:52:04.4 z, 50.1 N, 178.2 E, M 6.3, h 32 km) as recorded at Midway (11A) at a distance of 22 degrees from the epicenter, and Oahu (11B) at a distance of 34 degrees from the epicenter. This earthquake had an epicenter under the Aleutian trench, and, as in Figure 6, a strong abyssal T phase is observed followed by a slope T phase. The abyssal and slope T phases are well separated when observed at Midway (normal to the shoreline), while at Oahu (at an azimuth of about 45° to the shoreline) the abyssal arrival is partially masked by a slope arrival. In this event, the abyssal T phase

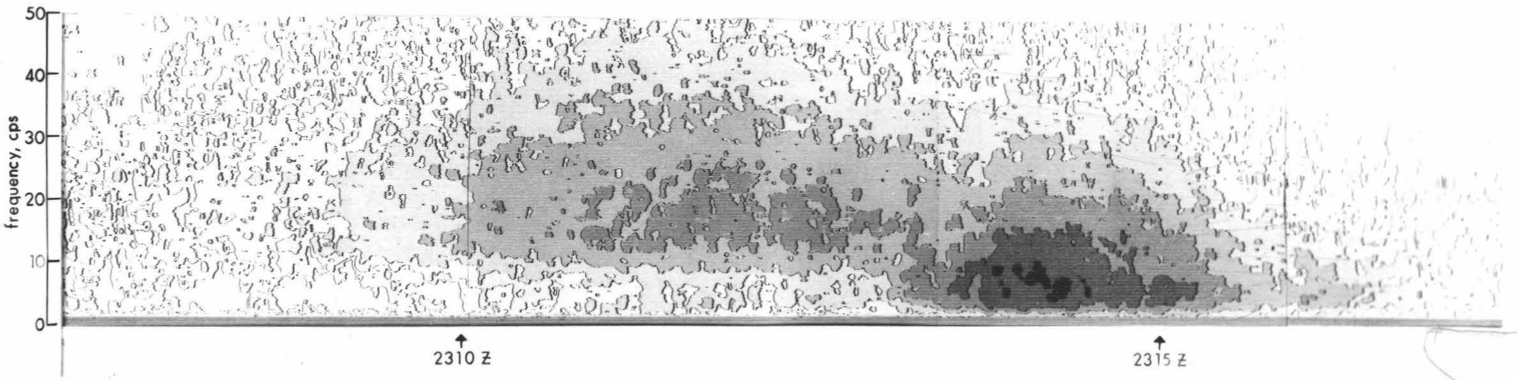

Fig. 10. Sonagram of a slope T phase from Kamchatka preceded by a forerunner: Note that the high frequencies are more intense in the forerunner than in the slope T phase.

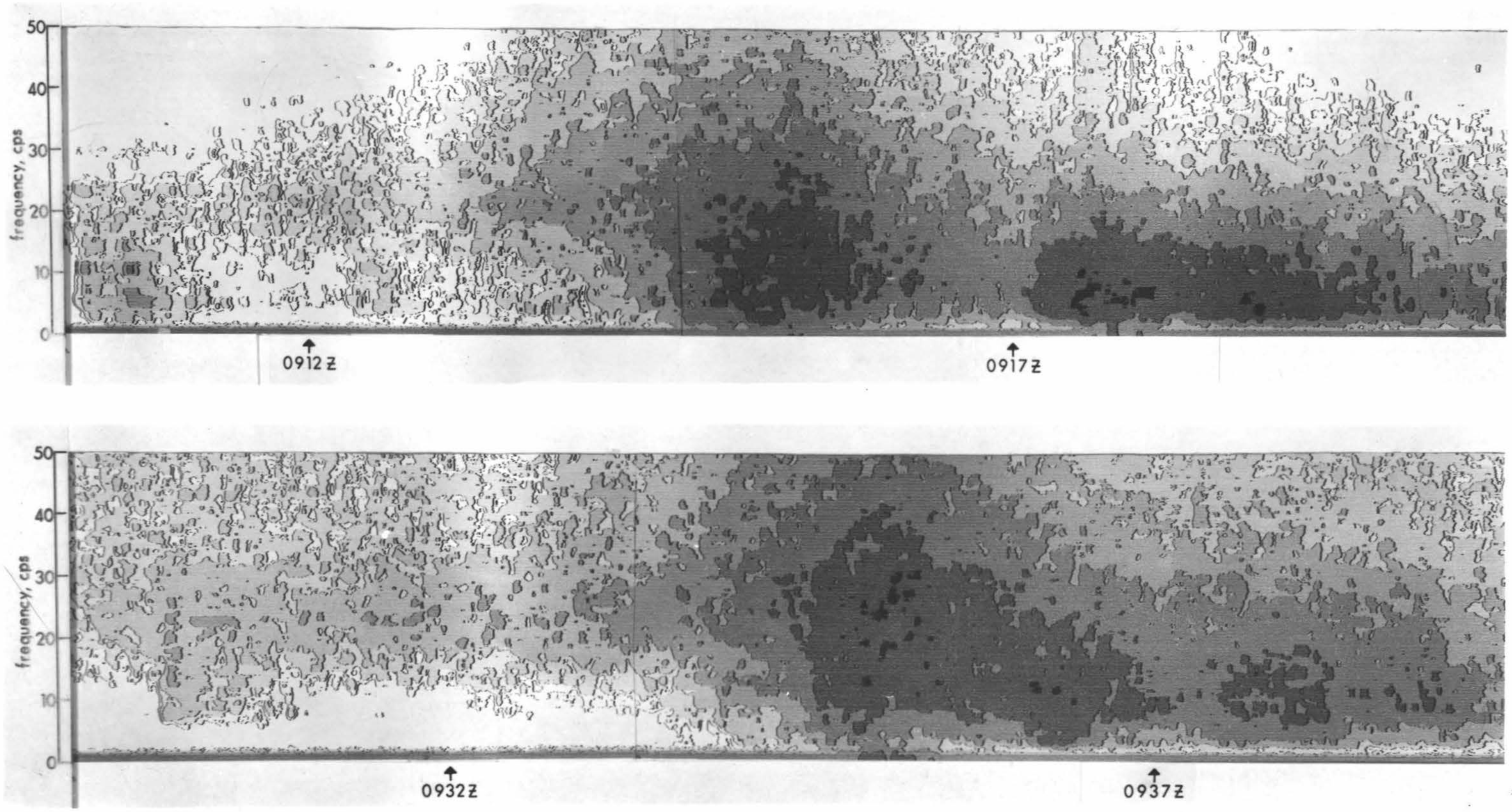

Fig. 11. Sonagrams of abyssal and slope T phases from an earthquake in the Rat Islands as observed normal to the shoreline (A) and at an angle of about 45° to the shoreline (B).

is divided into two parts, the first part lacking power in low frequencies, and the main abyssal T phase which does not lack low frequencies (though its apparent peak frequency, near fifteen cycles per second, is still high compared to the slope arrival). The lack of power in low frequencies in T phases from some areas and not from others is probably due to some property of the environment at the source. The abyssal T phase of Figure 11A might easily be confused with a slope arrival, but the T phase source location for this event coincides with the epicenter, not with the slope.

Very large magnitude earthquakes are postulated to be the effect of rupture along a fault hundreds of kilometers in length such that the point source approximation no longer is valid. T phases generated by such earthquakes show the effects of the extended source. Johnson and Norris (1966), studied the extended source effects for slope T phases, and found that an approximation of the fault length and velocity of rupture could be obtained from the slope T phase. The theory is essentially the same for large earthquakes occurring under the ocean floor generating abyssal T phases. An abyssal T phase is easier to interpret since its main peak should be generated directly over the hypocenter, rather than from a slope near the epicenter.

Figure 12 is the T phase of a tsunamigenic earthquake in the Kurile Island (October 20, 1963, 00:53:07.2 z, 44.7 N, 150.7 E, M 6.0, h 25 km), as recorded at Midway, at a distance of 30 degrees, and Wake Islands at a distance of 28 degrees. Note that the time duration of the abyssal T phase is longer on the Wake Island record than on the Midway record. Figure 13 is a graph of the abyssal T phase duration time vs. azimuth assuming a rupture velocity of 3.5 km/sec and a fault length of 100 km.

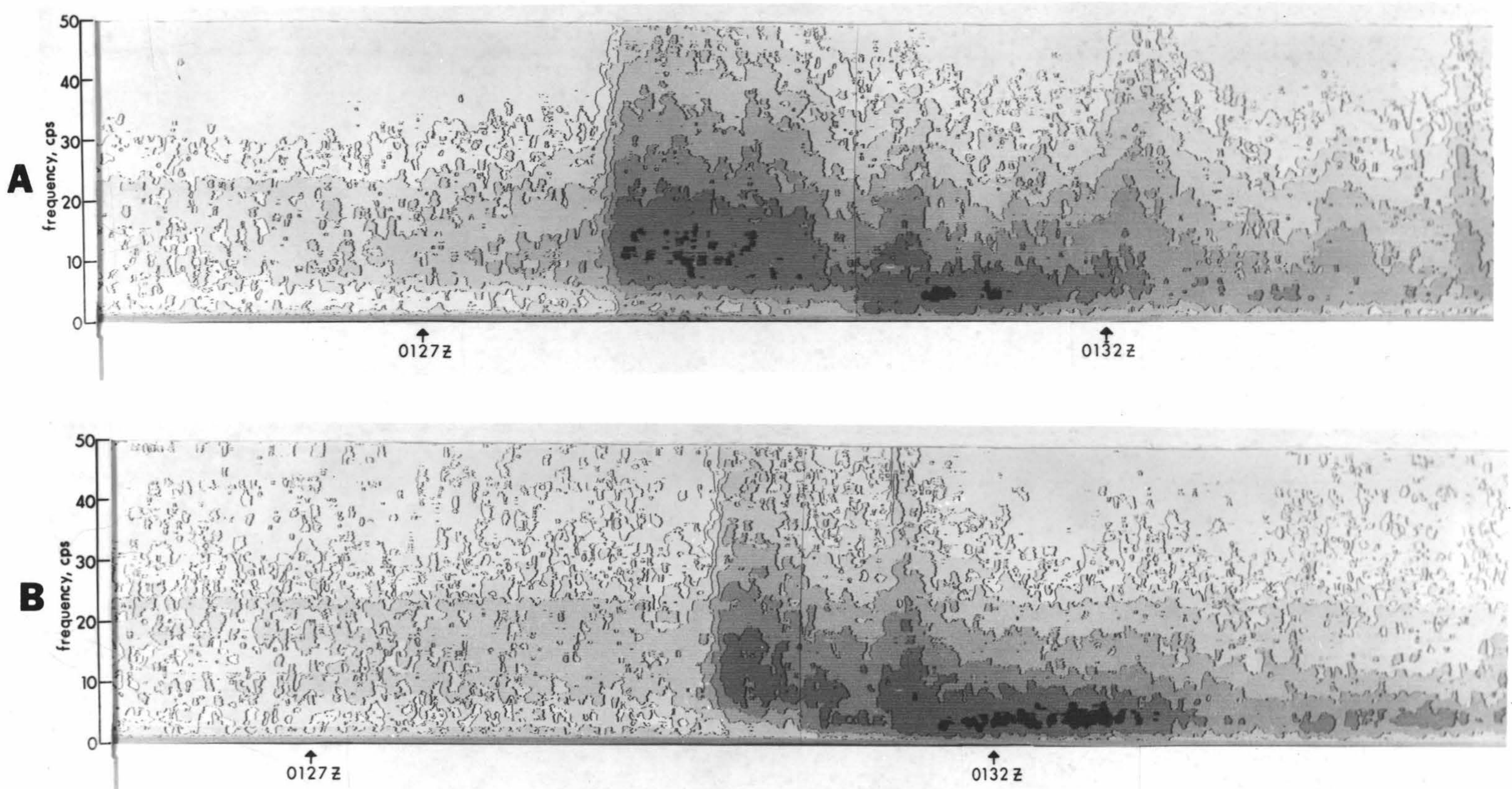

Fig. 12. Sonagrams of a tsunami generating earthquake in the Kurile Islands recorded at Wake (A), and Midway (B). The abyssal T phases are recognized by the lack of intensity at low frequencies.

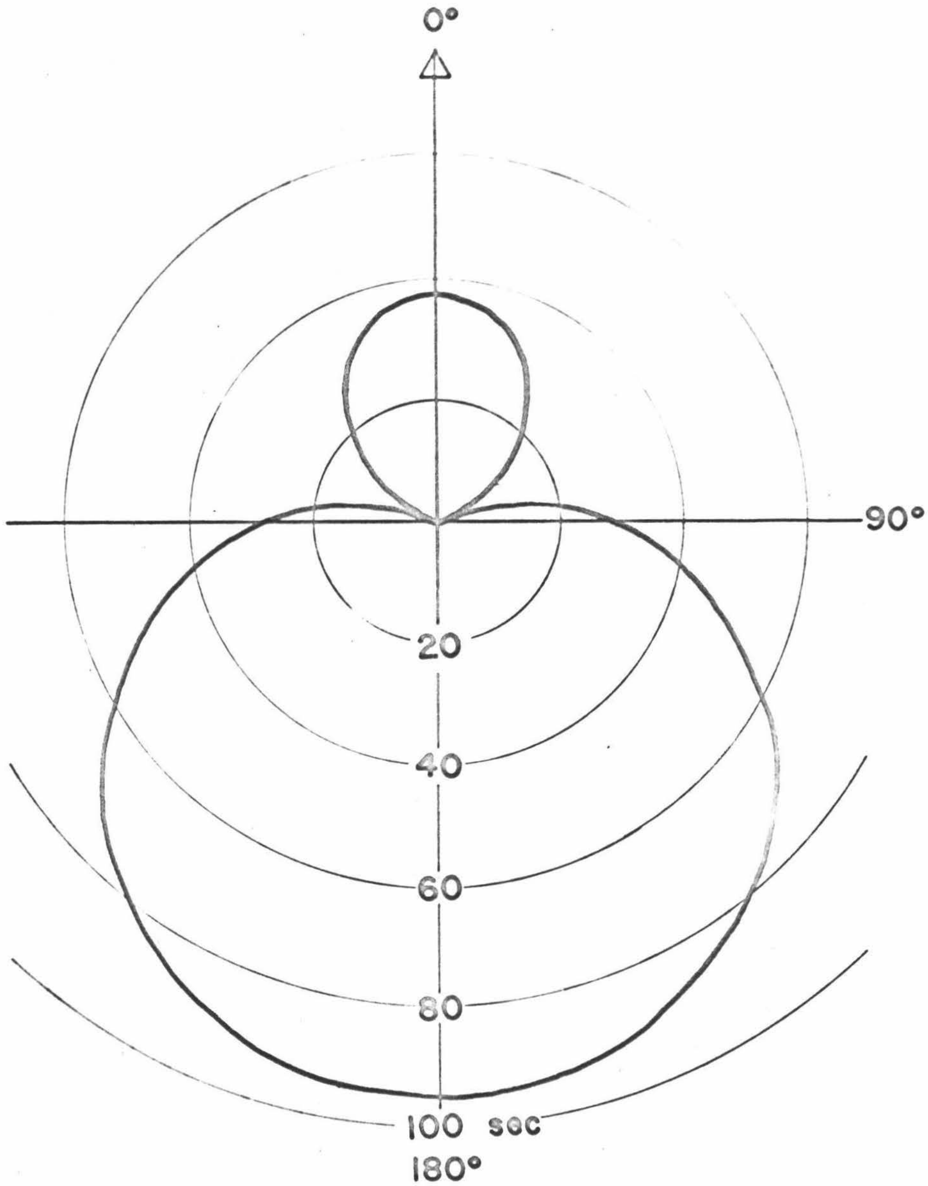

Fig. 13. Graph of duration time of an abyssally generated T phase vs. azimuth from the direction of faulting (0°) to the recording station. The duration time at any azimuth is proportional to the distance from the center of the circles to the heavy black line. This graph is for a fault length of 100 km and a rupture velocity of 3.5 km/sec.

The T phase should have a maximum duration time when observed from an azimuth of 180° from the direction of rupture, and a minimum duration at an azimuth of about 65°; the azimuth where minimum duration is observed being dependent on velocity of rupture and independent of fault length when observation is made at large distances from the source region. If records are available at three stations at different azimuths to the source area, source locations and origin times may be computed for the start and finish of the main abyssal T phase, thus supplying fault length, fault direction and rupture velocity. If two station records are available and the epicenter is known, rupture direction may be estimated by comparison of the two records and fault length may be computed, assuming a rupture velocity, by the formula

$$L = D \left(\frac{1}{v_r} - \frac{\cos\theta}{v_t} \right)^{-1}$$

where D is the duration of the T phase, L is the fault length, v_r is the rupture velocity, v_t is the T phase velocity, and θ is the azimuth from the direction of rupture to the hydrophone. The formula above differs from the formula used by Johnson and Norris (1968) only in that the above formula assumes that the distance from the hydrophone to the epicentral area is much greater than the fault length, while theirs does not. Comparison of the Wake (azimuth: 102°, duration: 15 sec) records in Figure 12 shows that this fault probably ruptured to the northeast parallel to the Kurile Islands. If a rupture velocity of 3.5 km/sec is used, the fault length computed from the duration time of the main abyssal event at Wake Island (90 sec) is 220 km. This estimate of fault length is an order of magnitude greater than expected value of approximately 25 km for a magnitude 7 earthquake (Press and Brace, 1966).

SPECTRAL ANALYSIS

The spectral characteristics of an observed _T_ phase depend upon several parameters; the seismic wave frequency characteristics at the hypocenter, ground path attenuation, transfer from the ground to the ocean, sofar channel attenuation, and the frequency response of the recording system. It is necessary to examine the relative effects of each parameter before a mechanism for abyssal generation can be proposed.

Little is known about the characteristics of the _P_ wave spectrum near the epicenter at frequencies above 5 cycles per second. To approximate the spectrum, a local (Oahu) earthquake body wave arrival at a bottom mounted hydrophone was studied. Since the water path to the hydrophone was zero, and the ground path should be less than 20 km (the _P_ wave and _S_ wave arrive nearly simultaneously), the correction for ground path attenuation should be negligible. The sonagram of the body wave arrivals is shown in Figure 14. An assumption is made that all earthquake body waves arriving at the ocean bottom after short ground paths have similar spectrums; while this assumption is probably not accurate, it is useful as a first approximation.

The attenuation of seismic waves in the earth is measured by a parameter Q, the reciprocal of the attenuation per wave length. _Sutton et al_. (1967) found values of Q for body waves between 200 and 1000 across the continental United States where:

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

f = frequency (cycles per second)

U = group velocity

b = attenuation coefficient in db/km

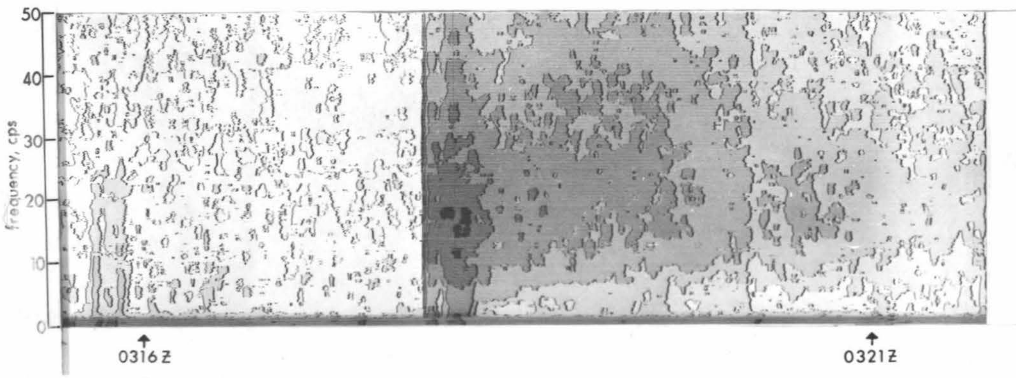

Fig. 14. Sonagram of body wave arrivals from a local earthquake recorded at a bottom mounted Oahu hydrophone.

If the Q in the regions studied is assumed to be within the limits (200 - 1000) found by Sutton et al., an attenuation of from 2 db/100 km (Q = 1000) to 11 db/100 km (Q = 200) are expected from absorption at a frequency of 40 cycles per second. If spreading is assumed to be spherical, an additional loss given by L (db) = 20 log $^{D_2}/_{D_1}$ is suffered in the ground path, where L is the difference in intensity between two distances (D_1 and D_2) to the hypocenter of the earthquake. Thus, the difference in intensity of the _P_ wave due to spreading and absorption when observed at the epicenter of an earthquake with a 33 km depth of focus, and at a point 100 km away at 40 cycles per second would be between 12 db (Q = 1000) and 20 db (Q = 200). Since it is postulated that _T_-phase signals which arrive at different times have sources at different points, these differences should be noticeable on _T_ phase sonagrams; spreading loss should cause changes in intensity with time on the sonagram, and absorption should cause changes in the slope of the power spectrum with time. However, since the relative efficiencies in _T_ phase radiation of different points on the o-ean bottom is variable, the effects of absorption and spreading in the ground path may be considerably masked.

A study of attenuation of sound generated by sofar bombs in the North Pacific sofar channel was made by Johnson (personal communication). A minimum (near zero db/1000 km) attenuation near 50 cycles per second increasing to approximately 6 db/1000 km at ten cycles per second was found. The high attenuation in the lower frequencies may be the effect of leakage of long wave length sound from the sofar channel (Urich, 1963). Attenuation of sound generated by sofar bombs (explosives detonated in the sofar channel) does not, however, agree with the attenuation of earthquake T phases. An attenuation coefficient near one db per 1000 km appears to be high for T phase amplitude studies. The corrections used in this paper are those found for sofar bomb attenuation, only because better information is not available. Since the sofar attenuation correction appears to have the least reliability, it is applied separately to the data.

The hydrophone system response is approximately that of a high pass filter where the corner frequency depends upon the length of the cable from the hydrophone to the recording system. The hydrophone-amplifier system itself attenuates low frequencies; the cable attenuates high frequencies. The coupling causes a system with a long cable to have a flatter response than a system with a short cable.

For a given earthquake T phase, the spectrum can be corrected for the hydrophone response, sofar attenuation, ground path attenuation, spherical spreading, and for the initial spectrum of the seismic energy close to the hypocenter. The resulting curve will reflect errors in the approximations made in the above corrections as well as the effect of the T phase generating mechanism on the spectrum. If all the corrections are accurate and complete, the resulting curve should show just the effect of the generating mechanism.

Figure 15 shows spectrums taken from vertical sections through the sonagraphs of Figures 6, 10, and 11 corrected for the above parameters except for sofar attenuation. The slope T phases (A) appear essentially flat while abyssally generated T phases (B) are attenuated in the low frequencies. Figure 16 includes the correction for sofar attenuation causing the slope T phases (A) to be attenuated in the high frequencies and the abyssally generated T phases (B) to be essentially flat.

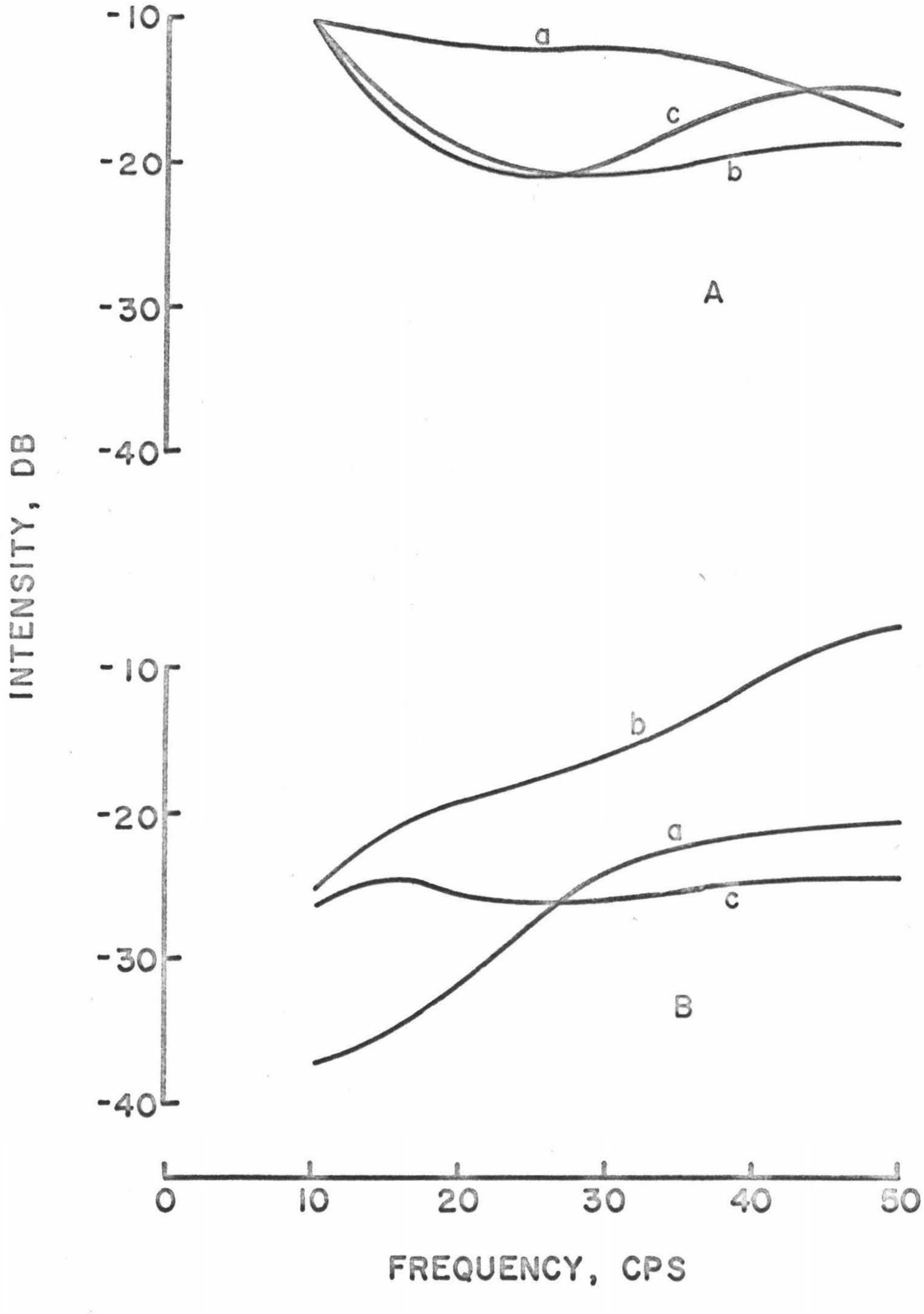

Fig. 15. Intensity spectrums for slope $\underline{T}$ phases (A) and abyssally generated $\underline{T}$ phases (B) corrected for all parameters except sofar channel attenuation. Lines a, b, and c correspond to vertical sections through the sonagraphs of Figures 6, 10, and 11A respectively.

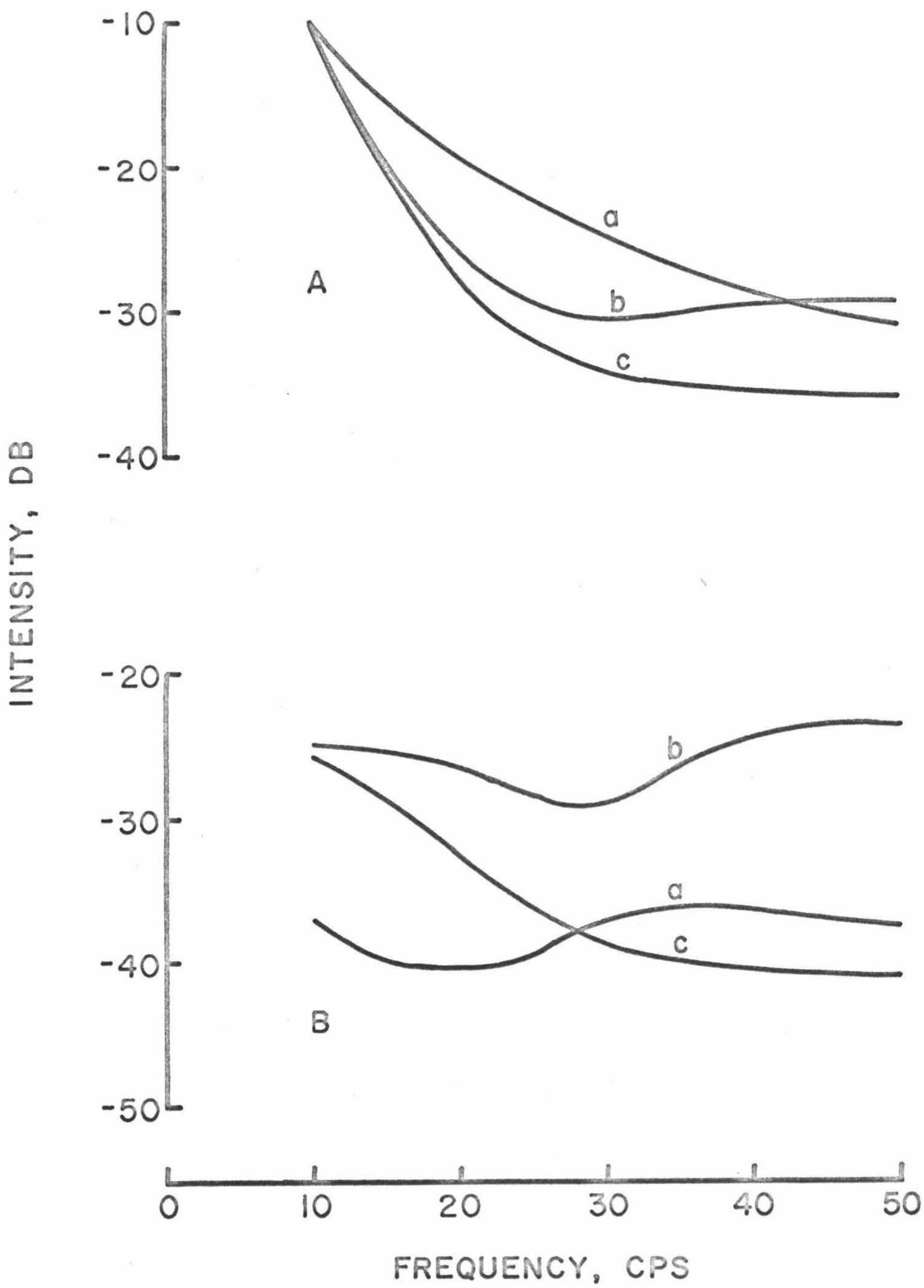

Fig. 16. Intensity spectrums for slope T phases (A) and abyssally generated T phases (B) corrected for source spectrum, ground path attenuation, hydrophone response, and sofar channel attenuation. Lines a, b, and c were made from vertical sections through the sonagraphs of Figures 6, 10, and 11A respectively.

Despite the uncertainties in the corrections applied, it is obvious that the two mechanisms have different frequency characteristics. The slope mechanism is more efficient than the abyssal mechanism at low frequencies, and nearly equal in efficiency to the abyssal mechanism at high frequencies.

According to Johnson et al. (1967), one of the characteristic differences between slope and abyssally generated T phases is the difference in the frequency of peak power. Slope T-phase sonagrams (uncorrected) have peak frequencies from two to ten cycles per second, while abyssal T phases are observed to have peak frequencies above ten cycles per second. Johnson et al. (1967) noted that correction for the response of the hydrophone would shift the peak to a lower frequency. In fact, this correction causes the spectrum of abyssal T phases to be essentially flat down to the limit of the hydrophone calibration curve (10 cycles per second). The apparent (uncorrected) peak frequency of abyssally generated T phases is highly variable, although seldom lower than ten cycles per second. If the mechanism for abyssal generation is scattering of sound from the ocean surface (Johnson et al., 1967), the spectrum of abyssally generated T phases would be expected to vary with changes in sea state, and thus with time. However, the spectra of abyssally generated T phases do not seem to change with time, different events from the same region having approximately the same spectrum. Events from different regions exhibit different spectrums; abyssal T phases from the Fox Islands being usually distinguishable from those from the Rat Islands by changes in the frequency spectrum. Fox Islands abyssal T phases seem to have less intensity in the low frequencies than Rat Islands abyssal T phases.

The low frequency end of the spectrum (<10 cps) is very poorly known. Although signals do appear at these frequencies, the hydrophone system response is not well known. When the hydrophone response curves are extrapolated to frequencies less than ten cycles per second and corrections are applied to observed signals, the results indicate very high intensities at low frequencies for all signals. This is not unreasonable since T phases appear to have peak intensity near 2.5 cycles per second when observed on seismometers. It is unfortunate, however, that the hydrophone systems do not have accurate calibrations at low frequencies.

MECHANISM FOR ABYSSAL GENERATION

Since abyssally generated _T_ phases have been observed from regions where water depth is greater than the depth of the sofar channel, ray path theory does not account for the observation of abyssally generated _T_ phases. The problem amounts to finding a mechanism for transferring seismic energy from its essentially vertical path at the ocean bottom into near-horizontal paths in the sofar channel. The conditions placed on the mechanism are that it be somewhat less efficient than down slope propagation at frequencies less than about twenty cycles per second, and approximately equal in efficiency above about thirty cycles per second. The mechanism must also account for the nearly symmetric shape in time of abyssal _T_ phases when generated above a flat ocean bottom. While abyssally generated _T_ phases are observed from regions of deep water, there is no reason to believe that the mechanism does not also operate in regions of shallow water.

Johnson, Norris, and Duennebier (1968) suggested that a mechanism which might account for the spectrum and shape of abyssal _T_ phases could be scattering of sound from the ocean surface. A study of current theories of scattering of sound by ocean surface waves and comparison with the observed frequency characteristics of the abyssal mechanism shows that surface scattering theory does not account for abyssal generation.

Theory of scattering from the ocean surface has been treated by many authors, however, usually the treatment is limited to the approximation that the wave length of the ocean surface wave is much greater than the wave length of the sound wave. For this case, the mathematics is simplified and the solutions agree with observation in the region covered by the approximation. For T phase frequencies, the wave length of the ocean surface waves is equal to and less than the wave length of the sound wave. Statistical studies of scattering from random surfaces appear to be valid only for high frequencies (above 100 cps).

Rayleigh (1945) treated the case of an acoustic wave scattered from a pressure release surface. According to his work, a wave normally incident on a corrugated surface will be reflected without scattering if the wave length of the corrugation is less than that of the sound wave. The amplitude of scattered spectral orders is found by inverting an infinite matrix. For a sound wave of short wave length compared to that of the corrugation, the number of scattered spectra is determined by the ratio of the two wave lengths. The amplitude coefficient of the last (most highly scattered order) is smallest, the next coefficient being imaginary.

A solution developed by Marsh (1961) for the wave scattering problem for statistical surfaces was used to solve the problem of signal loss from scattering at the ocean surface (Marsh et al., 1961). Our problem is just the complement, finding the amount of signal scattered into paths which can be received by a hydrophone in the sofar channel. According to the results of Marsh, the signal loss per ocean surface bound is computed by the formula

$$\alpha \text{ (db/bounce)} = -10 \log [1 - 0.023 (fh)^{3/2}]$$

where α is the loss in intensity, f is the frequency in kilocycles per second, and h is the rms wave height in feet. For normal states ($h = 3$ ft.), α is already below 0.2 db/bounce at 200 cycles per second, and growing smaller with decreasing frequency. Assuming that the scattering loss at the sea surface is 1/10 db per reflection at T phase frequencies (<50 cps), a +10 db signal reaching the sea surface will have a total of -71 db intensity in the scattered spectrum after one reflection. Only a small fraction of this amount will be scattered at high angles. Thus, the efficiency of scattering at the ocean surface appears to be extremely low at T phase frequencies. The same conclusion can be seen from Rayleigh's work, in that a 5 cycle per second signal traveling 1.5 km/sec would require an ocean surface wave length greater than 300 meters for any scattering to occur. Since analysis indicates a flat spectrum for the abyssal T phase mechanism above 10 cycles per second, and suggests rising intensity at lower frequencies, scattering from the ocean surface seems doubtful as a possible mechanism for abyssal generation.

Another difficulty with a surface scattering mechanism is that unless the sofar channel is bounded by the ocean surface at the T phase source, energy scattered from the surface will not reach sofar channel paths. In far northern latitudes, the sofar channel is close to the surface; in equatorial latitudes the sofar channel axis is much deeper and bounded by the thermocline. Energy scattered by the ocean surface in northern latitudes may enter sofar paths as the sofar axis deepens, however, energy scattered from the surface in equatorial latitudes will not travel in the sofar channel and may be attenuated by continued reflection from the ocean surface. Observation of abyssal T phases having sources where the sofar channel is deep shows that a large amount of the energy is restricted to sofar paths, or at least to paths which would require high scattering angles at the source.

Another possible mechanism for abyssal generation mentioned by Johnson et al. (1967) was 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 tend to scatter sound waves. Urick (1963) postulated that scattering of this sort accounts for high attenuation coefficients for signals from 10 to 1000 cycles per second in sofar channel paths. According to his work, attenuation of signals near 20 cycles per second is about 2 db per 1000 km. Scattering by similar inhomogeneities over a four km path from the ocean bottom to the sea surface would be negligible. Scattering at angles which would become sofar paths would require extremely large temperature or salinity gradients. Thus, as Johnson et al. (1967) concluded, volume discontinuities do not seem to account for abyssal generation.

The suggestion that early arriving T phase signals (forerunners) might be accounted for by travel paths through the upper sediment layers where velocities are only slightly higher than sofar channel velocities does not agree with observation. It has been observed that the intensity of abyssally generated arrivals is much stronger when observed by sofar channel hydrophones than by deep bottom mounted hydrophones. If the forerunners were caused by energy traveling through the sediment, the opposite would be expected. Such a mechanism would also probably severely attenuate higher frequencies. Thus the theory that abyssal generation is caused by energy traveling through the sediments can also be rejected.

Johnson et al. (1967) rejected the possibility that abyssal generation may be caused by the coupling of surface waves with the ocean sofar channel because the symmetric shape of abyssal T phases would not be expected and because higher frequencies would not be favored. Biot (1951) theorized that Stoneley waves, boundary waves coupled between the ocean water and bottom sediments, may couple very well with the sofar channel at frequencies where the velocity of the Stoneley wave is near that of the speed of sound in water. For the first Stoneley wave mode, the phase and group velocities are about 0.95 that of the speed of sound in water at T-phase frequencies, and the amplitude potential decreases exponentially from the ocean bottom. For all higher modes, the phase and group velocities are very close to the velocity of sound in water. Davies (1965) observed Stoneley waves produced by an explosion on the ocean bottom and recorded by a seismograph also on the ocean bottom. The frequency of the observed arrivals were between three and ten cycles per second. The main objection to a mechanism involving Stoneley waves in T-phase generation is that, like all boundary waves, the source must be near the interface where displacement is maximum for high efficiency, thus the source should be at the ocean floor-water interface. Since abyssal generation may be initiated by earthquakes 100 km or more from the source region, the source of energy for abyssal T phases may be far from the ocean bottom interface. A strong P wave, however, may cause the ocean bottom interface to act as a source itself, thus generating Stoneley waves which could couple with the sofar channel. This could explain the symmetric shape of abyssal T phases under flat ocean floors.

Whatever the actual mechanism is for abyssal generation of a T phase, it is not likely to be extremely efficient. Unfortunately, estimates of the T phase intensity compared to the approximate P wave intensity at the T phase source are not feasible at this time due to the lack of knowledge of the energy available at the source and to variability of characteristics of T phase sources. It is well known that large amounts of energy do enter the water from earthquakes. Many ships report severe shaking as if they had run aground, and sometimes even damage from vibrations which are later correlated with known earthquakes (Birch, 1966). In several instances the epicenter is hundreds of kilometers from the location of the ship. Such high intensities may cause yet unknown effects in the water column which are responsible for the generation of abyssal T phases.

CONCLUSIONS

From the observations made in earlier papers and in this paper, several conclusions can be reached concerning the generation of T phases and their frequency characteristics. The most noticeable variations in the T phase intensity signature and spectrum are caused by changes in the characteristics of the T phase source region and the position of the epicenter with respect to the ocean bottom. The shape and spectrum are not appreciably modified by distance from the hydrophone to the T phase source or by changes in magnitude of the earthquake, although changes in magnitude cause changes in intensity of the T phase with the exception that very large magnitude earthquakes with extended sources produce noticeable changes in the shape and duration of the T phase. Early arrivals, or forerunners appear to be abyssally generated.

Two separate mechanisms appear to be important in T phase generation; slope generation and abyssal generation. Slope generation has been expl explained previsouly, 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 fact that it occurs where slope generation cannot occur. The mechanism by which abyssal generation takes place is still in doubt, although it appears that coupling with Stoneley waves may be possible.

The frequency characteristics are usually different enough to separate T phases generated by the differant mechanisms. Abyssally g generated T phases recorded from sofar hydrophones appear to be attenuated in the low frequencies while slope T phases are not. When corrections are applied to the observed signals to obtain the frequency characteristics of the generating mechanisms, two possible sets of characteristics are found; if the uncertain correction for sofar attenuation is included, the slope mechanism attenuates high frequencies, and the abyssal mechanism has little effect on the frequency spectrum of signals. If the sofar correction is not included, the slope mechanism shows an essentially flat response and the abyssal mechanism attenuates low frequencies.

SUGGESTIONS FOR FURTHER STUDY

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 from the source should show what paths the sound is taking and give an estimate of the efficiency of the abyssal mechanism.

Knowledge of the mechanisms of _T_-phase generation would be enhanced if accurate data was available at frequencies lower than ten cycles per second. It would be a great help to have a hydrophone-amplifier system with a flat (or at least well known) response from 0.1 to 100 cycles per second.

Further frequency analyses along with more detailed studies of bathymetry coupled with an intensive source location program should be able to isolate the effects of topography on abyssal generation. More data are needed before the fine structure of abyssally generated _T_ phases, such as rapid fluctuations in intensity and rapid frequency shifts can be explained.

Tsunamigenic earthquakes occur near or in the ocean basin, and thus they generate intense _T_ phases; the relationship between the generation of tsunamis and _T_ phases to worthy of further study.

The fact that abyssal _T_ phases are generated directly over the earthquake epicenter makes source location valuable in that precise studies can be made of epicenter positions by a favorably placed array of hydrophones. Such a study could increase knowledge of seismicity of the ocean rises.

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