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AN INVESTIGATION OF THE RELATION BETWEEN
SOURCE CHARACTERISTICS AND T PHASES
IN THE NORTH PACIFIC AREA

A DISSERTATION SUBMITTED TO THE GRADUATE DIVISION OF THE
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Since the pioneering work on seaquakes by Rudolph (1887) there has been considerable interest in the scientific community concerning ocean floor earthquakes. With present day detection techniques, it has become apparent that many seaquakes do in fact occur beneath certain parts of the ocean floor and its margin, although the deep ocean basins appear to be aseismic. This thesis was undertaken to show how detection of hydroacoustic energy from seaquakes can be used to locate the epicenter, study the seismicity of a selected area (the Aleutian arc), shed some light on the various parameters that limit the threshold of detection, and delineate signal characteristics by which earthquake signals may be distinguished from those of underground explosions on the Aleutian arc. In completing this work I have received assistance from personnel of the T-phase project of the Hawaii Institute of Geophysics (University of Hawaii), the Marine Physical Laboratory (University of California) and the Naval Undersea Warfare Center, San Diego Division. The work has been done in conjunction with U. S. Navy sponsored studies of transient signals in the ocean.
ABSTRACT

Earthquake body waves refracted into the ocean are called T waves. They propagate as acoustic waves in the deep ocean sound channel at frequencies near 10 Hz. Arrival times of T waves, as recorded on widely spaced hydrophones in the sound channel off Eniwetok, Midway, Wake and Oahu Islands in the northern Pacific, are used to locate the T-phase source area. Sources thus located are correlated with published earthquake epicenters.

The amplitude, onset rate, duration, frequency content and number of peaks of T-phase signals differs with earthquake magnitude as well as epicentral location. A study of T phases from earthquake epicenters in a localized source area, the western Aleutian Islands, was undertaken to investigate the cause of variation in T-phase signals. The earthquakes studied occurred in a series of aftershocks following the 7 3/4 magnitude earthquake south of the Rat Islands in the Aleutian Islands on February 4, 1965. The aftershocks occurred in a zone, about 300 miles long and 150 miles wide, roughly parallel to the WNW-ESE trend of the Aleutian arc.

T-phase signals were different for earthquakes on the Aleutian arc, and behind the arc, from earthquakes on the Aleutian insular slope and Aleutian Trench. The T-phase strength, for earthquakes of a given magnitude, was greater for earthquakes on the arc, least for earthquakes in the Trench and outer ridge, and intermediate for earthquakes on the insular slope. The onset rate of T-phase signals varied with earthquake magnitude, but in general was most rapid for
epicenters on the Aleutian arc, intermediate for epicenters on the Aleutian slope and gradual for epicenters in the Trench and outer ridge. Decay rates varied with earthquake magnitude because of reverberation and topographic reflections. Onset and decay rates were symmetrical for earthquakes in the Trench and on the outer ridge. Earthquakes on the Aleutian arc produced multiple peaked T phases due to radiation of T waves from more than one slope in the epicentral area. Earthquakes far out on the Aleutian slope, bench, Trench and outer ridge sometimes produced multiple peaked T phases because of both reflections from the slope behind the epicenter and radiation of T waves from that slope as well as from the epicenter.

The accuracy of source location, as compared with epicentral location using the arrival times of earthquake body waves, was best for earthquakes on the Aleutian arc and in the Trench, and poorest for slope and bench earthquakes. Source solutions for signals from underwater explosions at known locations in the area indicate a source location accuracy of ± 10 miles. The accuracy of earthquake epicentral location by T phase is limited by the relatively broad signal peak of the T phase, and the fact that the point, or points, of radiation of the T phase is usually at an ocean bottom slope near the epicenter instead of at the epicenter itself.
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INTRODUCTION

The T Phase of Ocean Margin Earthquakes

Earthquake energy is propagated radially outward from the point of origin (called hypocenter) by two main types of waves: (1) body waves and (2) surface waves. Body waves travel through the solid earth as short (< 10 sec) period compressional (P) and shear (S) waves in the earth's crust and mantle. Only P waves travel through the earth's core, which does not transmit S waves. P-wave velocities range from about 5 to 13 km/sec and S-wave velocities from about 2.5 to 7 km/sec depending on the type of rock, hydrostatic pressure and ambient temperature. Surface waves are long-period waves that travel in the earth at velocities of 2.5 to 5 km/sec depending on the depth of penetration of the waves, which increases with increasing wave length. There are two main types of surface waves, Love (L) waves and Rayleigh (R) waves. Love waves are propagated only when there is a low speed layer overlying a medium with higher elastic wave speeds. Particle motion in Love waves is transverse in a plane parallel to the surface. Rayleigh waves travel along the earth's free surface and the particle motion is elliptical in the vertical plane. For shallow focus earthquakes, surface waves have a greater amplitude than body waves and make up what is sometimes called the principal part of seismograph recordings.

A special type of surface wave, having a period of about 1/10 second, is associated with ocean margin earthquakes. These short period waves travel in the ocean at the velocity of sound in water,
1.5 km/sec, and are known as T waves. The T phase has been defined as a "train of waves of period between about 1/2 and 1/100 sec propagated across ocean basins from earthquakes having epicenters in the basin or very near its margin" (Ewing, Press and Worzel, 1952). The appearance of a train of T waves on a seismogram is called the T phase.

Previous Work

The T phase was first recognized on seismograms by Linehan (1940). Tolstoy and Ewing (1950) showed that T waves traveled at the speed of sound in water but this was disputed (Leet, Linehan and Berger, 1951). Ewing, Press and Worzel (1952) then proved the underwater propagation mechanism by recording T phases on hydrophones (see Appendix).

The Deep Ocean Sound Channel

A low velocity layer is found in the ocean because the speed of sound in water increases with increasing temperature, pressure and salinity. The relative effects of the three variables controlling velocity are:

Temperature: $1^\circ$ C change = 3 to 5 m/sec change in the velocity, depending on the water temperature.

Salinity: 1 part per 1000 change = 1.2 m/sec change in the velocity.

Pressure: pressure effect of 100 m of water = 1.6 m/sec change in the velocity.

Typical conditions near the Hawaiian Islands are shown in Figure 1.
FIGURE 1. Vertical section of typical ocean temperature, salinity and sound velocity conditions in Hawaiian waters.
1° 5° 9° 13° 17° 21° 25° deg. C
32 33 34 35 °C salinity
1.43 1.45 1.47 1.49 1.51 1.53 1.55 km/sec

SOUND SPEED

SALINITY

TEMPERATURE

depth m
4000
3000
2000
1000
0
The speed of sound in sea water can be computed from the formula of Wilson (1960).

\[ V = 1449.22 + \Delta V_t + \Delta V_p + \Delta V_s + \Delta V_{tps} \]

where the reference velocity \( V = 1449.22 \) m/s was measured experimentally at \( t = 0^\circ C, p = 0.0 \) kg/cm\(^2\) and \( s = 35.00 \) O/00. Corrections to this velocity, \( \Delta V_t, \Delta V_p, \Delta V_s \) and \( \Delta V_{tps} \), are then applied from Wilson's (1960) tables. Earlier, and less precise, computations of sound velocity in the ocean are those of Kuwahara (1939) and Del Grosso (1940).

At depths below the surface mixed layer in the ocean, temperature decreases with depth, and therefore sound velocity also decreases. The sound channel is formed at a depth where the increase in sound velocity with depth due to pressure exceeds the decrease in sound velocity due to temperature decrease with depth. If the surface temperature is cool, the sound velocity minimum is near the surface; if it is warm, the minimum is deep. The depth of the sound velocity minimum, called the sound channel axis, thus varies with latitude and is shallowest at highest latitudes. Depths of the sound channel axis vary from the surface in the Arctic Ocean to 540 fathoms (1000 m) in the central Pacific and 680 fathoms (1250 m) in the central Atlantic. If the assumption is made that the ocean is horizontally stratified, as is entirely reasonable, and that the sound velocity in a given layer is everywhere the same in that layer, then computation of sound ray paths can be made using Snell's law of refraction. This law, applied to acoustics, states that the ratio of sound velocity
to the sine of the ray angle (with the vertical) is constant for a
given ray. (A ray is perpendicular to the wave front, and is defined
for these purposes as the loci of positions occupied by a given part
of a wave front throughout its course.)

Therefore, a ray can be traced through a series of layers using
the formula

\[
\frac{c_1}{\sin \theta_1} = \frac{c_2}{\sin \theta_2} = \frac{c_3}{\sin \theta_3} = k
\]

where \( c_1 \), \( c_2 \) and \( c_3 \) are the sound speeds in three consecutive layers.
Thus, for a given ray, the ray is refracted toward lower velocity
layers. Such a situation is common in the ocean, as shown in Figure 2.

In the deep ocean, sound velocity at the base of the water column
can equal that at the surface, so that the width of the sound channel
is essentially that of the depth of the water. For very deep ocean
basins, the bottom sound velocity exceeds that at the surface, and
sound rays penetrating these depths are refracted upward instead of
being reflected from the bottom. The effect of this phenomenon is
that the steeper rays spend more time in the high velocity section of
the water column than do the shallower rays. The zero degree, or
axial, ray by definition is confined to the sound channel axis,
travels slower than any of the other rays and arrives at the receiver
last. The arrival time of this axial ray is therefore followed by an
abrupt end of the signal.

Because the sound channel axis is below the depth of seasonal
variations in most latitudes, sound velocity in the sound channel is
FIGURE 2. Sound ray diagram for a receiver at the depth of minimum sound velocity at Hawaii in summer. Sound channel depth in summer and winter are shown in the vertical velocity profile at left.
relatively stable (Thor, 1964). Because of this stability, the axial ray arrival times of signals from TNT charges exploded in the sound channel have been used to determine accurately the source of the explosion (Ewing and Worzel, 1948; Frosch, Klerer and Tyson, 1961). Such a method of locating distant explosions, by delay time analysis of arrival times of the axial ray at three or more widely spaced hydrophones placed in the sound channel, has been called SOFAR (SOUND Fixing And Ranging) triangulation. Sound propagation within the channel is called SOFAR propagation, and the sound channel itself is called the SOFAR channel (Ewing and Worzel, 1948). The peak signal level received immediately prior to the sharp cutoff point that follows the arrival of the axial ray is called the SOFAR arrival. The existence of this channel was discovered independently by Dyk and Swainson (1953).

The reasons that low frequency underwater sound can be detected at long range in the SOFAR channel are that attenuation is low and spatial spreading of sound energy in the channel is restricted to two dimensions. Sound energy is subject to only cylindrical spreading loss of $\frac{1}{R}$ instead of spherical spreading loss of $\frac{1}{R^2}$, where R is in units of distance between source and receiver. For events that occur at SOFAR axis depth, some sound rays are continuously refracted between the higher velocity layers which lie both above and below the axis.

**Introduction of T-Phase Energy into the Sound Channel**

Earthquake P waves are refracted into the ocean at the critical
angle of refraction, or less, in accordance with the physical laws of optics. Assuming parallel layering and a crustal velocity at the earthquake hypocenter of 6.6 km/sec, the critical angle as computed from Snell's law is

$$\theta_c = \sin^{-1}\frac{C_1}{C_2} = \sin^{-1}\frac{1.5}{6.6} = 13^\circ$$

where $\theta_c$ is the critical angle of refraction measured from the vertical, $C_1$ is the sound velocity in the ocean and $C_2$ can be taken as an average crustal velocity for shallow focus earthquake P waves.

Some mechanism is required to explain the channeling of these nearly vertical rays into the nearly horizontal SOFAR rays. In terms of ray theory, multiple surface- and bottom-reflected rays can be converted into sound channel rays by the process of down slope propagation (Officer, 1958, p. 159). In such propagation paths, the ray becomes more nearly horizontal by twice the bottom slope upon each bottom reflection. Such an explanation for the transfer of energy from P to T waves was implicit in Tolstoy and Ewing's (1950) paper, and was later restated by Milne (1959) and Johnson, Northrop and Eppley (1963). For earthquakes beneath the deep ocean floor, down slope propagation paths are unavailable. At high latitudes, where the sound velocity minimum is within 200 ft of the surface, sound energy may be introduced into the SOFAR channel by scattering from the sea surface at the point of surface reflection for the critical ray (Johnson, Norris and Duennebier, 1967).

For frequencies below 5 Hz, where the wave length of sound is approximately equal to the sound channel depth, ray theory is no
longer strictly applicable (Tolstoy, 1959). Thus, for frequencies below about 5 Hz some T-phase energy is propagated through the water column via multiple surface and bottom reflections at near grazing angles. This type of energy transmission has been called normal mode propagation. Theoretically, frequencies below 5 Hz are propagated as normal modes and contained in the T phase, but for the recordings discussed here they are discriminated against on two counts: (1) in the deep ocean more energy is present in the higher modes so that the lower modes are masked (Ewing and Press, 1953) and (2) the frequency response of the hydrophone-cable system used in the T-phase recordings, on which this work is partially based, has a peak sensitivity at 100-150 Hz. Below this peak the sensitivity decreases by 6 dB per octave. Therefore, the sensitivity was 30 dB down for frequencies below 5 Hz.

For the purposes of this paper, then, T-phase propagation will be concerned mainly with the ray (optical) theory of underwater sound propagation.

Problem Undertaken

A specific series of earthquake T phases from the same general area (the Rat and Near Islands of the Aleutian archipelago) is studied. The travel paths of T waves from the source area to the detectors are similar for the series of events studied, but the initial starting conditions are dissimilar. The effects of this dissimilarity upon the T-phase signals received are studied in detail.

Objectives of Investigation

The objectives of this investigation are to describe earthquake
T phases from the western Aleutian Islands and determine why T phases from that area differ from one to another. The T phases differ in signal strength, duration, onset, decay and frequency content. The cause of these differences must be at the source or on the path between source and receiver because the same receivers are used for all earthquakes studied. One objective of this investigation is to identify the geologic, oceanographic, and topographic factors that affect both the T-phase signal received and the accuracy of the T-phase source location. Other low-frequency sounds from underwater, or underground, explosions in the western Aleutian Islands would be affected in the same way, but the source characteristics would differ. Indeed, hydroacoustic signals from the 80 Kt Long Shot underground explosion on Amchitka Island in September 1965 was similar to a T phase from a magnitude 5.5 earthquake on Amchitka Island (Daubin and Brumbach, 1966; Johnson, 1966). Through an investigation of the T phase, criteria for differentiating T phases from hydroacoustic signals from explosions may become apparent.

Relation to Present State of Knowledge

Following their initial work on the T phase, personnel of the Columbia University Bermuda Geophysical Field Station have continued research on transient signals from earthquakes and explosions (Ewing, Mueller, Landisman and Sato, 1959). Shurbet (1962) suggested that arrival times of T waves, as noted on recordings from ocean bottom detectors, could be used to locate oceanic earthquakes. Analysis of the T-phase signals as recorded on hydrophones of the Pacific Missile
Range has revealed that numerous T phases (about 1 per hour) are propagated across the Pacific Ocean. The signals received differ considerably for earthquakes from different areas, and also for earthquakes of similar magnitude from the same geographic area. In early studies of the T phase, from land-based seismograph recordings, variations in the T-phase signal were ascribed to geologic conditions at both the source and receiver. In the present work, geologic conditions at the receiver are not pertinent because the detectors are in the ocean. However, bathymetry in the neighborhood of the hydrophones does affect the T-phase signal received.

Relation to Similar Work in Progress Elsewhere

The Advanced Research Projects Agency and the Office of Naval Research support a variety of projects for the purpose of surveillance, detection and identification of telesismic signals. Various aspects of the work involve studies of very low frequency sound propagation because absorption of low frequency sound energy in the ocean is less than that of high frequency sound. Absorption varies as the $3/2$ power of the frequency, as shown in the following formula:

$$\alpha = \text{absorption coefficient} = 0.033 \, (f)^{3/2}$$

where $\alpha$ is in dB per kiloyard and $f$ is in kilocycles per second (Sheehy and Halley, 1957). Propagation of high frequency sound in the ocean, therefore, is clearly limited, whereas transoceanic travel paths are readily attained for low frequency ($< 60$ Hz) sound (Ewing and Worzel, 1948).
Acknowledgements

Acknowledgement is made to Rockne H. Johnson of the T-phase project at the Hawaii Institute of Geophysics, and to the Office of Naval Research (Field Projects Branch) which supported my T-phase work in 1963-64 under contract Nonr 3748 (01) and in 1965-66 under contract N00014-67-A-0109-0001. Personnel of the Marine Physical Laboratory contributed in various ways. I am particularly indebted to F. N. Spiess, M. S. Loughridge, R. W. Raitt and E. W. Werner. The manuscript was read by R. A. Haubrich of the University of California.
FIGURE 3. Location of stations and source area; great circle T-phase propagation paths shown between stations and the source area in the western Aleutian Islands.
GREAT CIRCLE CHART OF NORTH PACIFIC

\(\triangle\) RECORDING SITES

\(\odot\) SOURCE AREA

--- GREAT CIRCLE PATHS
CHAPTER II
MATERIALS AND METHODS

General Statement

The Pacific Missile Range (PMR) operates hydrophone monitoring installations off the islands of Eniwetok, Midway, Wake and Oahu, and near the west coast of North America (Fig. 3). The purpose of these Missile Impact Location System (MILS) hydrophones is to record the arrival times of hydroacoustic signals from SOFAR bombs released from missiles when they land in the water. Ordinarily, the hydrophone recordings are monitored only upon request, and that request must be made in advance to cover the proposed drop time of a missile.

Hydrophone Recordings

The MILS station personnel were asked by R. H. Johnson to monitor hydrophone signals continuously on analog recorders supplied by the T-phase project of the Hawaii Institute of Geophysics. These records form the basis of the T-phase analysis reported on here. A sample record is shown in Figure 4.

The records were mailed to the Hawaii Institute of Geophysics (HIG) for data processing. Because the T phase is strong in frequencies below 15 Hz (Milne, 1959; Northrop, 1962; Johnson, 1963), the hydrophone signal was passed through a 15-Hz low-pass filter prior to recording. This filtering was found to improve the signal-to-noise ratio by 10-20 decibels (Johnson, 1964). Because of the low frequencies, a highly damped pen response was used to give a smoother
FIGURE 4. Sample hydrophone records showing the T-phase signal from an earthquake at 51.5° N, 179.0° E (USC&GS) at 00h50m45s UT, 17 February 1965. Magnitude 4.5, depth 40 km. This earthquake was on Amchitka Island and is number 115 of Figure 5.
EARTHQUAKE T PHASE FROM
AMCHITKA ISLAND
MAGNITUDE 4.5 (USCGS)
DEPTH 40 KM (R)
trace and a more meaningful reading of power level. To reduce the bulk of the records, the slowest available chart speed of 0.25 mm/sec was used. This practice did not compromise reading accuracy as the T-phase peak is typically broad and its position could not be defined more accurately than ± 3 seconds.

All records had two timing traces: one for locally generated time and one for radio (WWVH) time. These were traced from styli operating on additional pens (Fig. 4). The radio time stylus was activated whenever the WWVH-transmitted 440- or 600-Hz cone was received on the radio. The date of the reading was annotated at the beginning and end of each record. Measurements of peak pressure and arrival time of T-phase signals were referred to radio time whenever reception permitted. The pressure level of the event was also measured at this peak arrival. To ensure adequate calibration of the MILS system, a 10-Hz electrical signal was injected instead of the sea noise at the input of the hydrophone amplifier every day. This calibration signal was varied by 10-dB steps throughout the full dynamic range of the system.

Analysis of Records

Three types of characteristic signatures from transient events were noted on the records: (1) short (< 1 min), repetitive, sharply-peaked pulses at individual stations; (2) short (< 1 min) sharply-peaked pulses correlatable between stations; and (3) long (1-6 min) broadly-peaked transient signals correlatable between stations. The first type of pulse was found to be due to a moving source near the
hydrophone and had a seasonal peak (in winter). It was therefore classified as being of biological origin, (see, for example, Schevill, Watkins and Backus, 1964; Weston and Black, 1965; Northrop, Cummings and Thompson, 1967). The second type of signature was found, by comparing discreet arrival times with those computed for known echo-ranging explosions, to be hydroacoustic signals from man-made detonations. These events had, typically, a sudden onset, sharp peak, rapid decay rate and more high-frequency content than earthquake T phases (Northrop, 1962; 1957). The third type of transient signal had characteristically a slowly emerging onset and decay, a relatively broad peak, and was, in many instances, correlatable with the computed arrival time of a hydroacoustic wave propagated from the region of a known earthquake epicenter. These broadly-peaked signatures (Fig. 4) are T-phase signals.

**Correlation of Events**

Correlation was done by delay time-bearing determination between arrival times of events at individual hydrophones at a given island and by delay time-bearing determination between arrivals at different islands. The time of arrival of a T phase was taken as the peak power arrival, as this corresponds to the arrival of the axial ray (Frosch, 1964). If a T-phase signal arrived at the same time and date at Midway and Wake, it must have originated along a locus of points equidistant from both islands; i.e., along the perpendicular bisector of a line joining the two islands. In this case, the bisector would indicate a source area either in Japan, the East
Pacific Rise or South America. Then, looking for an event with similar rise time, decay rate, duration and level characteristics on the record from another island, say, Eniwetok, it could be established that the arrival at Eniwetok was either earlier or later than that at Wake. If earlier at Eniwetok, the event came from the south, eliminating Japan. If it arrived later at Eniwetok, the event came from the north, eliminating South America and the East Pacific Rise. More precise bearings were obtained by determining the azimuth from the array of hydrophones at each station and bringing arrival times at additional stations into play. Using this method of azimuth determination, an accuracy of about 30 could be realized for the PMR sites because the hydrophones were spaced from 60 to 160 seconds of water wave travel time apart. Ambiguity in reciprocal bearings was eliminated at Eniwetok, Midway and Wake because there were four hydrophones at each station. At Hawaii and both California stations the 180° ambiguity was obviated because the hydrophones were close to shore and had an arc of reception limited to about 190° (Johnson, Northrop and Eppley, 1963).

**Data Processing**

After an earthquake T phase had been correlated in this manner, at a minimum of four hydrophone locations, it was given an event number. The events were numbered consecutively throughout the length of a 200 foot roll (two days' data). These analog data were then digitized onto punched cards on a Benson-Lehner Oscar K digitizer which operated in two modes: one for measuring the time and the other
for measuring the level. The punched cards, which give event number, level in dB, date, year, hour, minute, second, island name and hydrophone number, were used as input data for the source location program on an IBM 7040 computer. This program was written by R. H. Johnson and J. S. Sasser (Johnson, 1966). The T-phase source solutions thus obtained were then screened for possible error and printed in lists (Anonymous, 1965) giving the month, day, hour, minute and second of origin time (Greenwich Mean Time), the latitude and longitude of the source area, the geographic area of the source, and the standard deviation of the time residual in seconds (Table I). Also a weighted average of the peak signal level at all stations receiving the signal was used to indicate a measure of T-phase strength, a measurement similar to earthquake magnitude (Johnson and Northrop, 1966). Events correlatable with published USC&GS epicenters are also listed in monthly bulletins of the T-phase project by the Hawaii Institute of Geophysics (Anonymous, 1965).
<table>
<thead>
<tr>
<th>Month</th>
<th>Day</th>
<th>Hour</th>
<th>Minute</th>
<th>Second</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Area</th>
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<td>41</td>
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<td>177.2 E</td>
<td>Rat Island</td>
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<td>17</td>
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<td>8</td>
<td>8</td>
<td>0</td>
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<td>179.3 W</td>
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<tr>
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<td>24</td>
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<td>46</td>
<td>36</td>
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<td>173.8 E</td>
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<td>1 &quot;</td>
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</tr>
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<td>13</td>
<td>6</td>
<td>51.5 N</td>
<td>176.0 E</td>
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<tr>
<td>Feb</td>
<td>11</td>
<td>10</td>
<td>38</td>
<td>42</td>
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<td>169.1 W</td>
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<td>11</td>
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<td>15</td>
<td>15</td>
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<td>11</td>
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<td>11</td>
<td>53</td>
<td>3</td>
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<td>14</td>
<td>45</td>
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<td>Near Island</td>
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<td>22</td>
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<tr>
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<td>8</td>
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<td>173.5 E</td>
<td>Near Island</td>
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<td>13</td>
<td>7</td>
<td>40</td>
<td>49.9 N</td>
<td>177.1 E</td>
<td>South of Aleutians</td>
<td>2 &quot;</td>
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</tr>
<tr>
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<td>10</td>
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<td>175.8 E</td>
<td>Rat Island</td>
<td>2 &quot;</td>
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</tr>
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<td>11</td>
<td>13</td>
<td>40</td>
<td>24</td>
<td>52.3 N</td>
<td>174.0 E</td>
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<td>58</td>
<td>39</td>
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<td>28</td>
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<td>70.3 W</td>
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<td>26</td>
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<td>176.0 E</td>
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</tr>
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<td>49</td>
<td>18</td>
<td>50.8 N</td>
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<td>Rat Island</td>
<td>6 &quot;</td>
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CHAPTER III

EARTHQUAKE T PHASES FROM THE NORTH PACIFIC

General Statement

The rim of the North Pacific is made up of a series of island arcs and associated deep trenches to seaward of the islands. From west to east, these zones are the Japanese Islands and associated Japan Trench, the Kuril-Kamchatka arc and associated Kamchatka Trench and the Aleutian arc and associated Aleutian Trench. Most of the world's earthquakes occur along these, and other island arc structures, as has been well documented by Gutenberg and Richter (1949). Many of the earthquakes in the North Pacific rim that are associated with the Japan-Kuril-Kamchatka-Aleutian arcs cause T phases, and about 25 per day are noted on the T-phase records. The earthquakes often occur in swarms of aftershocks following an unusually large earthquake and are believed to stem from sub-crustal movements beneath the arc (Benioff, 1954). Such a series of earthquakes occurred in the western Aleutian Islands in February, 1965 (Jordan, Lander and Black, 1965).

Geology of the Aleutian Arc

The Aleutian Islands are the exposed part of the Aleutian ridge that extends for 2000 miles between the Alaska and Kamchatka Peninsulas (Fig. 3). The ridge separates the Bering Sea from the Pacific Ocean. The south slope of the ridge deepens rapidly from the 50 fathom shelf that surrounds islands on the crest of the ridge, to the floor of the 4000-fathom deep Aleutian Trench. The Trench is about 60 miles south of the ridge and parallel to it (Fig. 5). The north
FIGURE 5. Earthquake epicenters and bathymetry in the area of the Near and Rat Islands. Contours from Gates and Gibson (1956) and Gibson and Nichols (1953).
EUTIANS IN FEBRUARY 1965

EARTHQUAKE EPICENTERS > MAG. 5.9 (USCG). NUMBERS REFERENCED IN TEXT

EARTHQUAKE EPICENTERS OF MAGNITUDES FROM 5.0-5.9

EARTHQUAKE EPICENTERS < MAG. 5.0

EARTHQUAKE EPICENTERS OF MAGNITUDES FROM 5.0-5.9

GE OF INSULAR SHELF (100 FATHOM CURVE)

HYDROMETRIC CONTOURS. SOUNDINGS IN FATHOMS

CONTOUR INTERVAL 1000 FATHOMS

TOGRAPHY FROM U.S.N. HYDRO. CHART BC 1910N 1ST. EDITION, 1952
slope of the ridge is steeper, but only reaches depths of the order of 2000 fathoms in the Bering Sea. The Bering Sea is divided into an eastern and a western basin called the Aleutian Basin and Bowers Basin respectively. The Basins are separated by a shallow (100 fm) ridge called Bowers Bank. Sediments fill both Bowers Basin and the Aleutian Basin to a depth of 2 km (Ewing, Ludwig and Ewing, 1965). Below the sediment, a normal oceanic crust is present which has been depressed, presumably by the weight of the sediment (Menard, 1967). There is little or no sediment on the south slope of the ridge (Ewing, Ludwig and Ewing, 1965).

The exact age of the Aleutian ridge is unknown, but it is believed to be rather young geologically. A reconnaissance study of the geology of the Rat Islands was made by the U. S. Geologic Survey, which published a series of maps and descriptions entitled "Investigation of Alaskan Volcanoes" (U. S. Geol. Survey Bulletin 1028 A-T). Sections that deal particularly with the western Aleutian Islands are by Coats, Nelson, Lewis and Powers (1961), Coats (1956), Snyder (1959), Nelson (1959), Powers, Coats and Nelson (1960), Lewis, Nelson and Powers (1960), and Nichols, Perry and Kofoed (1964). Geology of the Rat Islands and the Near Islands, including bathymetric studies, were published by Murray (1945), Gates and Gibson (1956) and Perry and Nichols (1965).

The oldest fossiliferous rocks exposed on the western Aleutian Islands are Middle Tertiary Oligocene or Miocene marine sediments interbedded with volcanic rocks on Amchitka and Rat Islands (Lewis, Nelson and Powers, 1960; Powers, Coats and Nelson, 1960). These rocks
overlie older basaltic lavas which are considered to be pre-Middle Tertiary in age. However, except for a single exposure, there are no known pre-Tertiary rocks on any of the Aleutian Islands (Burk, 1965). The islands are made up of more or less flat lying, shield volcano basaltic lava flows and pyroclastic rocks of, presumably, late Tertiary age, overlain by Pleistocene and Recent volcanic rocks in the form of composite volcanoes. Some of the volcanoes are active, and many have been active during historic time. Active or dormant volcanoes in the western Aleutian Islands are Semisopochnoi, Little Sitkin, Segula, Kiska and Buldir (Fig. 5). Erosional evidence of past glaciation in the form of cirques, roches moutonnee, glacial striations and small streams flowing in large U-shaped valleys indicate that the islands were glaciated. The early shield volcanoes were probably much larger both in areal extent and height than the present day composite volcanoes. Pleistocene glaciers apparently covered the islands completely, as there are no subaerially exposed moraines. The islands are surrounded by an insular shelf approximately 50 fathoms deep which has a flat surface presumably due to wave action at a former, lower, stand of sea level.

The Western Aleutian Islands

There are three main groups of islands in the western Aleutian Islands: the Rat Islands, the Near Islands and the Komandorskiye Islands. The Rat Islands are just west of the 180th meridian, between 51° and 53° N lat. (Fig. 3). They are separated from the Near Islands by a submerged section of the ridge near Tahoma Bank (Fig. 5).
division between these three sections of the ridge extends offshore in
the form of deep gullies, or canyons. The two main canyons have been
given the names Murray Sea Valley and Heck Canyon (Gates and Gibson,
1956) (Fig. 5). The axes of these canyons are oblique to the bathy­
metric contours rather than normal to them, which has led to the
belief that they may be of fault origin (Gates and Gibson, 1956). The
faults form an echelon pattern at an oblique angle to the trend of the
arc, and crustal segments between faults move in an echelon pattern
thereby straightening the arc (Menard, 1964).

The trend of major faults and joints in the western Aleutian
Islands is NE-SW (Coats, 1962). The zone between wrench faults is
from 300-400 miles long and 50-100 miles wide. Large numbers of
earthquake aftershocks following large (mag. 7 or greater) earthquakes
occur in zones roughly outlined by the wrench faults. Murdock (1966)
has emphasized that the eastern boundary of the February (1965) after­
shock sequence nearly coincides with the western boundary of the 1957
Andreanof Islands aftershocks. The observed seismicity boundaries
may indicate tectonic features which trend transverse to the island
arc, as is true of the Kuril Island arc (Udintsev, 1955).

Bowers Bank is such a transverse structure, extending north of
Semisopochnoi and the Rat Islands for 200 miles into the Bering Sea,
and is an aseismic ridge (Ewing, Ludwig and Ewing, 1965; Nichols,

Seismicity

The Aleutian arc is one of the more seismically active zones
in the Pacific (Gutenberg and Richter, 1949). Shallow earthquake focal depths ranging from 10-70 km occur throughout the length of the arc, but intermediate depth shocks are restricted to the north and, predominantly, the east side of the arc. No deep (> 300 km) shocks are known to occur in the region, which is unusual for Pacific Island arc structures.

Most of the earthquakes in the Aleutian Islands are of low magnitude. Indeed, earthquakes of magnitude above 7 on the Richter scale are limited to about 10 per year for the whole world, and about one per year occurs in the Aleutian Islands area.

An earthquake of magnitude 7-3/4 occurred in the region south of Rat Island near 51.3° N, 178.6° E (USC&GS) at 05h 01m 22s (UT) on February 4, 1965. An intense series of aftershocks, and one fore-shock, causing T phases occurred in the Rat Island-Near Island area in February, 1965. The number of aftershocks was greater than that following the Prince William Sound earthquake of March 27, 1964 (Jordan, Lander and Black, 1965). In both aftershock series, more than 50% of the total number of aftershocks occurred in the first 5 days after the main shock.

Epicenter Locations

Aftershocks following the main Rat Island earthquake occurred in a zone about 300 miles long and 100 miles wide (Fig. 5). The eastern boundary of the zone was near the 180th meridian, and just east of Amchitka and Semisopochnoi Islands. Brazee (1965) has shown that the sequence of aftershocks following the 1957 Andreanof-
Fox Islands, of the Aleutian Islands, earthquakes occurred in a well-defined zone that terminated just east of the Rat Islands. The western boundary of the Rat Island aftershock sequence lies about 60 miles west of Attu Island, near a shallow (100 fm) shoal called Stalemate Bank, near 170° E long. (Fig. 5). Further confirmation of this seismic boundary is lacking, other than that it is separated from the Komandorskiye (or Commander) Islands by about 100 miles of open water. There were 4 T-phase sources located in the Komandorskiye Islands during February, 1965, and there were about an equal number in the Andreanof-Fox Island area. Earthquakes in the latter zone during February 1965 appear to have no regular temporal distribution as do those in the Komandorskiye area except for a series of 15 earthquakes near 54° N, 166° W within a three hour period on February 8.

The western boundary of the Rat Island earthquake aftershock zone is not as well defined as the eastern boundary, and, indeed, the series of earthquakes in the Komandorskiye Islands on February 8, occurring as they did late in the sequence, may indicate a settling or adjustment of the crust following the Rat Island earthquake.

The northern boundary of the earthquake zone of the Rat Island aftershocks lies near the 1000 fathom curve north of the Aleutian arc, i.e., on the southern boundary of the Bering Sea. This seismic boundary is well marked in that only three epicenters were located north of it (Jordan, Lander and Black, 1965) and there was, in general, a sharp decrease in the number of epicenters north of the crest of the Aleutian ridge.
The southern boundary of the Rat Island earthquake zone is near the 49° N parallel of latitude and extends well south of the axis of the Aleutian Trench in the area of the outer ridge and the Komandorskiye Ridge. The number of epicenters south of the north wall of the Aleutian Trench is less than the number on the Aleutian slope, which is, in turn, less than the number on the Aleutian bench.

The distribution of earthquake epicenters of the Rat Island aftershock series thus is limited to the area between Rat Island to the east, the Bering Sea on the north, Stalemate Bank on the west and the outer ridge on the south. The greatest concentration of epicenters occurred between the 100 and 3000 fathom contours between the Near and Rat Islands and the Aleutian Trench. This pattern of shallow focus earthquakes occurring offshore from the crest of the Aleutian arc has been reported in the Fox Islands area by Murdock (1967).

**Depth of Focus**

Focal depths of the Rat Island aftershocks, as published on the Preliminary Determination of Epicenters (PDE) cards of the U. S. Coast and Geodetic Survey (USCGS) range from 10 to 50 km, with an average of 32 km. There is no perceptible increase in focal depth for events inshore of the Aleutian Trench, in contrast to those reported in the eastern Aleutian Islands (Benioff, 1954; Murdock, 1967). Focal depths are listed on the PDE cards to the nearest km, but these are to be considered no more accurate than ± 25 km (A. S. Furumoto, personal communication).
Crustal Structure

The crustal thickness of the Aleutian ridge is 25 km (Shor, 1964). Shor's seismic refraction measurements, made along a N-S line just east of the Rat Islands along the 177° W meridian, also indicate a crustal thickness of only 6 km under the axis of the Aleutian Trench. These values differ from the crustal thickness of 8 km for normal oceanic areas, 12 km for the Aleutian Basin and about 30 km for typical continental structures. Structure sections for the Aleutian arc, Trench and Basin are shown in Figure 6, along with one for typical oceanic areas for comparison.

The relatively thin crustal section under the Aleutian Trench, and the greater water depth, causes a deficiency in mass from that predicted by the international gravity formula of Bowie (1924). Over the Trench, a free air gravity anomaly of -60 milligals has been measured (Peter, Elvers and Yellin, 1965). The negative free air anomaly indicates a deficiency in mass, which is attributed to the fact that water thickness in the Trench is increased at the expense of mantle material at the bottom (Talwani, 1964) (Fig. 7).

Crustal structure cannot be derived directly from gravity anomalies (Woollard and Strange, 1962), but in areas where independent seismic measurements have determined the crustal layering, density values for the layers may be assigned from the data of Nafe and Drake (1963). Using these values, local effects on the gravity field can be computed following the methods of Talwani, Sutton and Worzel (1959). Peter, Elvers and Yellin (1965) computed a crustal section
FIGURE 7. Gravity anomaly across the Aleutian Trench.

Data from Peter, Elvers and Yellin (1965).
for the Aleutian ridge by using such a technique. Although the section is in the eastern Aleutian Islands, some of Shor's (1964) refraction data near Adak were used in the computations. Thus, the section (Fig. 7) may be considered a generally representative one for the Aleutian ridge and Aleutian Trench. A plot of earthquake focal depths is not included in the cross section because of the uncertainty of their accuracy. However, it is interesting to note that the depth of focus of earthquakes of the Rat Island aftershock sequence, as published on the PDE cards, ranges from 10 km to 50 km, with an average of 32 km for the 517 correlated events. Thus, most of the hypocenters were below the crust and in the upper mantle.
CHAPTER IV
T-PHASE TRAVEL PATHS

Great Circle Paths Between Source and Receiver

Distance between source and receiver was measured along great circle paths, ignoring possible aberrations due to either diffraction or horizontal refraction of sound rays.

The greatest path distance shown is 2500 nautical miles from the western Aleutian Islands to Eniwetok, and the least is 1500 miles to Midway (Fig. 3). Travel paths to the California stations are masked by the eastern Aleutian Islands, and will not be discussed in this paper. Other distances and hydrophone data are shown in Table II.

Topography Between Source and Receiver

Directly south of the source area, the north wall of the Aleutian Trench slopes down to depths of over 4000 fathoms, 60 miles offshore (Fig. 5). The bottom slope ranges from 1° on the insular slope to 12° on the north wall of the Trench. The south wall of the Trench rises from 4000 to 3000 fathoms. A general shoaling to about 2500 fathoms occurs between the Trench and the deep Pacific basin floor. This broad zone of relatively shallow water south of the Aleutian Trench is here called an outer ridge, following the nomenclature developed for the shallow area north of the Puerto Rico Trench (Heezen, Tharp and Ewing, 1959). In general, these features are common to paths from the western Aleutian Islands to all stations of the PMR hydrophone
### TABLE II

Distance and T-phase travel time between source and receiver

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<th>Station</th>
<th>Distance from Source (NM)</th>
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<th>Number of hydrophones</th>
<th>Minimum Distance Between hydrophones (NM)</th>
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<td>1500</td>
<td>31</td>
<td>4</td>
<td>55</td>
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<tr>
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<td>80</td>
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<td>1900</td>
<td>39</td>
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<td>25</td>
</tr>
<tr>
<td>Eniwetok</td>
<td>2500</td>
<td>51</td>
<td>4</td>
<td>90</td>
</tr>
</tbody>
</table>
FIGURE 8. Ocean bottom topography and isovelocity contours between the western Aleutian Islands and the hydrophone stations in winter.
network. However, paths to individual stations differ in detail because the slope is traversed normal to the bathymetric contours (of the Aleutian slope) only on paths to Eniwetok (Fig. 3). Ocean bottom topography is also different along travel paths from the western Aleutian Islands to different hydrophones. The path to Hawaii is interrupted by several seamounts, which cause noticeable acoustic shadows at the Oahu hydrophones (Anderson, Hamilton and Lesser, 1957). The path to Wake is interrupted by seamounts of the Emperor Seamount Chain, as is the path to Eniwetok (Fig. 8). Midway, then, has the only clear propagation path from the western Aleutian Islands, and thus T-phase signals recorded at that station encounter the least interference from topographic effects along the propagation path.

**Sound Channel Depth Between Source and Receiver**

The minimum sound velocity is at the surface in the Aleutian Islands area (Fig. 9) in winter, and increases with decreasing latitude to ~1000 m depth at the MILS hydrophones (Fig. 8). Sound rays refracted into the water from events on the Aleutian shelf and slope propagate downslope by successive surface/bottom reflections until the rays become shallow enough to achieve RSR (Refracted-Surface-Reflected) propagation. For the average bottom slopes (~5°) in the area, only 8 bottom reflections are required for the critical angle ray (sound ray refracted into the ocean at the critical angle, p. 10) to become an RSR ray (Fig. 10). As propagation continues southward toward the detection hydrophones, surface water temperature increases, and the depth of the sound velocity minimum
FIGURE 9. Sound velocity profiles in the Aleutian Islands area, summer and winter.
Sound Velocity, km/sec

- summer
- winter
increases, so that near 40° N latitude the surface sound velocity becomes equal to that (1.5 km/sec) at the depth of the greatest penetration (turn around point) for the critical ray. Rays steeper than critical would require more bottom reflections before becoming RSR rays. Southward from 40° N, RRR (Refracted-Refracted-Refracted) SOFAR propagation is achieved to the detectors.

For earthquakes occurring beneath the Aleutian Trench and outer ridge, there are no downslope propagation paths available. Therefore, some other process for introducing T-phase energy into the SOFAR channel must be invoked. From a study by Johnson, Norris, and Duennebier (1967) it appears that energy is introduced into the SOFAR channel from surface reflections by scattering of sound from the surface at the point of surface reflection of the upward-going refracted rays at the epicenter. After the energy has thus been introduced into the SOFAR channel, it propagates southward to the hydrophones by a combination of RSR and RRR propagation paths, as described above. However, the surface scattering mechanism is less efficient than the down-slope propagation mechanism, as shown by the lower T-phase levels for the former (Fig. 11).
FIGURE 10. Isovelocity contours and downslope propagating ray paths of the critical ray in the Aleutian Islands area in winter.
FIGURE 11. T-phase level vs earthquake magnitude for earthquakes shown on Figure 5. Data for the Rat Island aftershocks in February 1965. Sloping line is representative of the T-phase strength vs earthquake magnitude relationship from Johnson and Northrop (1966).
T-phase strength, dB re. 0.1 microbar at 30° distance

Epicenters

- Insular Shelf (0-100 fathoms)
- S-Slope (100-1500 fathoms)
- B-Bench (1500-3000 fathoms)
- T-Trench (3000-4000 fathoms)

Earthquake magnitude (USC & GS)

0.0
1.0
2.0
3.0
4.0
5.0
6.0
7.0
8.0
9.0
10.0
11.0
12.0
13.0
14.0
15.0
16.0
17.0
18.0
19.0
20.0
21.0
22.0
23.0
24.0
25.0
26.0
27.0
28.0
29.0
30.0
31.0
32.0
33.0
34.0
35.0
36.0
37.0
38.0
39.0
40.0
41.0
42.0
43.0
44.0
45.0
46.0
47.0
48.0
49.0
50.0
51.0
52.0
53.0
54.0
55.0
56.0
57.0
58.0
59.0
60.0

Aleutian Average
S=20M-52
CHAPTER V
TYPICAL T-PHASE SIGNALS

General Statement

A comparison of T-phase source solutions with PDE data was made for the Rat Island aftershocks. Aftershocks of this earthquake accounted for 1861 of the 2137 T-phase sources located at the Hawaii Institute of Geophysics in February (1965). The PDE cards list only 784 aftershocks during the period. Because the T-phase source location system routinely detects lower magnitude earthquakes than the PDE cards for this region (Johnson and Northrop, 1966), it can be assumed that about half of the low magnitude (< 4 mag) events were unreported on the PDE cards. Of the 784 possible correlations, 320 of the epicentral times (as determined by body-wave data) were within one minute of the origin times, as determined by the T-phase source solution program (Fig. 12). This good correlation of the T-phase data with the P-wave data is probably due to good T-phase source solutions being obtained on the North Pacific Rim (Johnson and Northrop, 1966). Even so, it was difficult to make unambiguous correlations between P-wave located events and T-wave located events in the aftershock series because the time between earthquakes was so short that a T phase could be correlated with more than one epicenter.

For the February events, therefore, a further restriction on possible correlations was made in that the PDE epicenters marked with an asterisk (indicating estimated accuracy of 1/2° in latitude and longitude and 50 km depth) were discarded. Furthermore, only those
FIGURE 12. Difference between earthquake origin times computed from P-wave data minus the origin times computed from T-wave data. Rat Island aftershocks in February 1965 stippled, all other areas, from September 1, 1964 to February 28, 1965, dashed.
LEGEND

- **EPICENTRAL TIME EARLIER THAN T-PHASE ORIGIN TIME**
- **T-PHASE ORIGIN TIME EARLIER THAN EPICENTRAL TIME**

**FEBRUARY RAT ISLANDS**

**ALL OTHER AREAS**

**Number of events**

- **0**
- **60**
- **120**
- **180**

**EPICENTRAL TIME EARLIER THAN T-PHASE ORIGIN TIME**

**T-PHASE ORIGIN TIME EARLIER THAN EPICENTRAL TIME**
correlations which were within 1° of arc and one minute of time were allowed. These stringent restrictions were placed on the correlations so that there would be no question but that the proper correlations had been made.

After both the PDE and the T-phase source correlations were checked, it became apparent that discrete T-phase signatures could be roughly correlated with water depth in the source area. Five general depth zones are prominent in the area: (1) shelf, (2) slope, (3) bench, (4) Trench and (5) outer ridge. "Aleutian shelf" is used to refer to areas less than 100 fathoms deep; the term "Aleutian slope" to areas from 100 to 1500 fathoms in depth; the "Aleutian bench" to areas from 1500 to 3000 fathoms in depth and the "Aleutian Trench" to those areas deeper than 3000 fathoms (Gates and Gibson, 1965; Gibson and Nichols, 1953). The term "outer ridge" is used for the shallower (2700 to 2900 fathoms) water area just seaward of the Trench.

T-phase signals were found to vary with station azimuth as well as epicentral location, earthquake magnitude and water depth at the epicenter.

T-phase signals differ in such characteristics as onset, duration, decay, peak level and number of peaks. The signals not only differ from event to event at a given receiver, but for a given event as recorded at different hydrophone sites. Multiple peaked signals present the most complex picture, and it is difficult to correlate peaks between station recordings. Poor correlation of peaks results in poor source location accuracy, so that this aspect of the problem will be treated first.
Because the number of epicenters in the Rat Island aftershock sequence is large, and they are rather evenly distributed east to west along the Aleutian arc and Trench, there are many that give rise to similar T-phase signals. Such signals can be considered as "typical", and, once such types have been delineated, they can be recognized fairly easily on the records. In order to clarify this point, tracings of a number of typical T-phase signals are presented in Figures 13 - 18. In preparing these figures, the envelope of the signature was traced and the peak level marked. The records are aligned according to the individual signal peaks used in reading arrival times, and in Figure 13 the records are arranged from top to bottom in order of increasing azimuth from the source, and, from left to right, in order of increasing epicentral distance from shore. Numbers over the peaks correspond to numbered events on the chart (Fig. 5). Major signal features shown in Figure 13 are: (1) greater signal level and more rapid rise time for nearshore earthquakes of a given magnitude than for offshore earthquakes of the same magnitude; (2) double peaked signals for offshore events; (3) increasing time delay between peaks with increasing azimuth from the source to the station for offshore events; (4) a symmetrical signal peak due to similar onset and decay rates for the initial peak of offshore earthquakes and (5) shorter time intervals between peaks for some T phases (138) than others (126) although their respective epicenters are about the same distance from the 100 fathom curve. With these general features in mind, T-phase signatures from the various geomorphic provinces in the western Aleutian Islands will be discussed.
FIGURE 13. Typical T-phase signatures for (A) shelf epicenter, (B) bench epicenter, (C) Trench epicenter and (D) outer ridge epicenter. Numbers above T-phase peak on record correspond to epicenters on Figure 5.
Epicenters on Islands and Insular Shelves

T phases discussed in this section come from epicenters within the 100 fathom curve of Stalemate Bank, the Near Islands, the Wells Plateau and Tahoma Reef, and the Rat Islands (Fig. 5). Typical signatures from each of these areas are shown in Figure 14. The onset of the signals is relatively steep, the signal duration from 2 1/2 to 5 1/2 minutes, and, for low magnitude earthquakes, only a single peak is present. The ratio of T-phase signal levels to earthquake magnitude is close to the average for the Aleutian Islands (Johnson and Northrop, 1966) (Fig. 11). For large magnitude earthquakes, multiple peaks appear on the records at stations for which the travel path makes an acute angle with the bathymetric contours. These multiple peaks are caused by individual shelf promontories that act as separate radiators of T-phase energy. The time delay between such peaks decreases as the angle between the azimuth and bathymetry approaches 90° (Johnson and Norris, 1966).

Epicenters Behind Islands

No direct hydroacoustic paths to the PMR hydrophones are available for T phases generated by earthquakes having epicenters behind the islands. T-phase energy is therefore radiated from the southward facing slope of the Aleutian arc. Such events were rare in the aftershock sequence, but one in the Aleutian Basin did produce a T phase that was recorded (Fig. 15). The signal exhibits a rapid onset, single peak, relatively short duration and peak level of 32 dB, which is low for an earthquake of that magnitude (4.7) (Johnson and Northrop,
FIGURE 14. Typical T-phase signatures recorded at Eniwetok from earthquakes having epicenters on the western Aleutian Islands and insular shelves. Numbers above T-phase peak on record correspond to epicenters on Figure 5.
Stalemate Bank
53.1°N, 170.8°E
5.7 m 40 km

Agattu Shelf
52.5°N, 173.9°E
5.3 m 39 km

Tahama Reef
51.8°N, 175.4°E
5.2 m 25 km

Amchifka Shelf
51.5°N, 179.0°E
4.5 m 40 km

SCALE
0 1 2 3
-10
-0 dB
FIGURE 15. T-phase signal from an earthquake in the Aleutian Basin recorded on 3 hydrophones at Midway and 1 at Eniwetok.
ALEUTIAN BASIN
54.7°N, 172.8°E
Midway 22
4.2 m 33 km

Midway 23

Midway 24

Eniwetok 44

0 1 2 3 -10 dB
minutes
FIGURE 16. Typical T-phase signals recorded at Eniwetok from earthquake hypocenters beneath the insular slope of the western Aleutian arc. Numbers above the T-phase peak on record correspond to epicenters on Figure 5.
Epicenters on the Slope

T phases from the insular slope south of the Near and Rat Islands exhibit a rapid onset, high signal level vs earthquake magnitude and double peaks. Typical signatures are shown in Figures 16 and 17. For some signals received at a given station, the time difference between the first and second peaks increases with the distance of the epicenter offshore from the 100 fathom curve. The second peak is therefore reflected or radiated from the slope behind the earthquake epicenter, whereas the first peak is generated at the earthquake epicenter. The reflected T phase is identical to the "T-reflected" signals discussed in Shurbet and Ewing's 1957 paper, and the first peak is the abyssal T phase of Johnson, et al., (1967).

Epicenters in the Trench and Outer Ridge

T phases from earthquakes beneath the Aleutian Trench have signal levels well below those of other zones (Fig. 11). The signals have similar onset and decay rates, higher frequency content and the T-phase source is at the epicenter (Johnson, Norris and Duennebier, 1967). Tracings of typical signals are shown in Figure 18. The T phase from the lowest magnitude (4.2 mag) event, shown in Figure 18, has only a single peak. For larger magnitude earthquakes, there are two or three peaks. The first peak of a sequence is due to the abyssally generated T phase, which is recorded first because the epicenter is between the receiver and the insular shelf. The second peak is due to PT* in which T-phase energy is radiated from the insular
FIGURE 17. Typical T-phase signals recorded at Eniwetok from earthquake hypocenters beneath the Aleutian bench. Numbers above T-phase peak on record correspond to epicenters on Figure 5.
Amchitka Bench
122 51.1°N 178.4°E
5.6m 35km

Tahoma Bench
13 51.8°N 174.6°E
5.1m 33km

Kiska Bench
119 51.1°N 177.8°E
5.1m 40km

Agattu Bench
135 51.9°N 173.4°E
5.0m 30km

SCALE
0 1 2 3 -10
minutes -0 dB
FIGURE 18. Typical T-phase signals recorded at Eniwetok from earthquake hypocenters beneath the Aleutian Trench and outer ridge. Numbers above the T-phase peak on record correspond to epicenters on Figure 5.
Ijt

5.0°N, 175.0°E

Tahoma Trench

5.3m 33 km

Agatuu Trench

51.3°N, 174.2°E

4.8m 33 km

50.7°N, 175.5°E

Tahoma Trench

4.6m 33 km

50.5°N, 178.0°E

Rat Is. Trench

4.4m 20 km

50.6°N, 177.3°E

Amchitka Trench

4.2m 33 km

-10 dB

minutes

0 1 2 3
slope inshore from oceanic earthquakes (Shurbet and Ewing, 1957). The third peak is due to reflection of abyssally generated T waves from the insular slope north of the epicenter. Although initial identification of the abyssally generated T phase was made from frequency spectrum analysis of magnetic tapes of T-phase signals recorded at the MILS stations (Johnson, et al., 1967), such recordings were not available to the author for this dissertation. However, once the high (~30 Hz) frequency content of the abyssally generated T phase had been correlated with the low level, symmetrically peaked T-phase signals from the Aleutian Trench and outer ridge, identification could be made initially on the strip chart records. Then, if the source solution for such a symmetrically peaked signal is in the Trench or outer ridge, the standard deviation of origin times low, the convergence high, and the signal level low, the earthquake can be assumed to have been beneath the Trench or outer ridge. The number of such earthquakes in the Aleutian Trench is less than on the Aleutian ridge.

Earthquakes beneath the Aleutian Trench and outer ridge are known to occur at locations other than south of the Rat and Near Islands (Brazee, 1965), are clearly associated with the Aleutian Trench (Burk, 1965) and should not be considered as evidence of seismicity on the deep ocean floor. Several such events occurred in the Aleutian Trench after the Rat Island aftershock series of earthquakes in 1965. Strip charts were available to the author for the period of August 1965 - July 1966 as well as February 1965, and were inspected for symmetrical T-phase signals having source areas in the Aleutian Trench and outer
ridge. Although positive identification of abyssally generated T phases in summer could not be made without frequency analysis of magnetic tape recordings, tentative identification was made in 10 instances during the summer months. It appears, then, that generation of the T phase by surface scattering of refracted waves in the ocean at the latitude (50°) of the Aleutian Trench is not greatly affected by seasonal changes in the sound channel structure.
CHAPTER VI
RESULTS

Types of T Phases Observed

Four main types of T phases were observed from earthquakes in the Aleutian Islands area.

(1) T phases produced by down-slope propagating rays along a sloping bottom between the epicenter and receiver. This type of T phase is called the classical T phase after the original work of Tolstoy and Ewing (1950). For the purposes of this paper, the classical T phase will be referred to as T. The travel path of T is shown diagrammatically in Figure 19.

(2) T phases produced by down-slope propagating rays along a sloping bottom inshore from the epicenter. This type of T phase will be referred to as PT* after the nomenclature of Shurbet and Ewing (1957). The travel path of PT* is shown diagrammatically in Figure 19.

(3) T phases produced by some mechanism such as scattering of hydroacoustic waves refracted into the ocean above the hypocenter. This type of signal is the abysally generated T phase of Johnson, Norris and Duennabier (1967) and will be referred to as T_A. The travel path of T_A is shown diagrammatically in Figure 19.

(4) T phases produced by reflection of hydroacoustic waves from
FIGURE 19. Idealized drawing of near source travel paths of T, PT*, T_A, and T_A-R types of T phases. Not to scale.
submarine topographic slopes. This type of signal is called $T_R$ after Aubrat (1963) and Shurbet and Ewing (1957). In this report, reflections of abyssally generated $T$ phases ($T_A$) from the south slope of the Aleutian arc are the main type of reflected $T$ waves discussed and they are called $T_{A-R}$. The travel path of $T_{A-R}$ is shown in Figure 19.

A given earthquake may produce more than one type of $T$ phase. For example, an earthquake beneath the Aleutian Trench sometimes produces both $T_A$ and $PT\ast$. Additionally, energy from $T_A$ may be reflected from the slope inshore from the epicenter so that $T_{A-R}$ is also produced. In this case, the arrival time sequence at a distant down-slope receiver would be $T_A$, $PT\ast$, and $T_{A-R}$. Both geologic and oceanographic factors influence the type of $T$-phase signal generated from a given earthquake epicenter. Geological factors will be discussed first.

**Geologic Factors Influencing T-Phase Signals**

Hydroacoustic energy from both $T$ and $PT\ast$ is channeled into SOFAR propagation paths by down-slope, bottom-reflected sound rays, and energy from $T_{A-R}$ is reflected off the bottom slope. Thus the received signal level is reduced by the amount of bottom loss from reflection. The total loss due to bottom reflection is the sum of losses due to absorption and scattering at each encounter with the bottom. Bottom slope, bottom roughness and reflectivity of the sediment are important in determining the amount of bottom loss per reflection.
Bottom Slope

At each bottom reflection the angle the sound ray makes with the bottom is decreased by twice the bottom slope (Officer, 1958). Because some energy is lost at each reflection ray paths that encounter the least number of bottom reflections undergo the least loss of energy. For example, a typical bottom slope in the Aleutian Islands area is 5° with the horizontal. With a bottom slope of this type, earthquake body waves refracted (at the critical angle) into the ocean become successively shallower until, after 8 bottom reflections, they no longer impinge on the bottom but are refracted-surface-reflected (RSR) rays (Fig. 10).

Upon each bottom reflection, the ray becomes more grazing, and energy loss per bottom reflection is less. Therefore, downslope propagation over a steeply-sloping bottom is a more efficient process of T-wave propagation than that over a shallow sloping bottom, because there are less bottom reflections required before RSR propagation is achieved in the former case.

Angle of Ray Path to Bottom Slope

The efficiency of transformation of energy from P to T waves is greater for ray paths parallel to the maximum slope than oblique to it (Shurbet and Ewing, 1957). Irregularities in the bottom slope also may contribute to bottom reflection loss by scattering. Irregularities in the reflecting surface have little effect on sound with wave lengths greater than the wave lengths of the surface irregularities (Rayleigh, 1945). For the wave lengths of sound involved in the T phase, we can
compute the order of magnitude of bottom roughness that would scatter the sound upon reflection using the expression

\[ \lambda = \frac{c}{f} \]

where \( \lambda \) is the wave length, \( c \) the speed of sound and \( f \) the frequency. For frequencies of 6 - 10 Hz, the wave lengths are 816 ft and 490 ft respectively. Irregularities in the bottom, of this order of magnitude or larger, are present in the vicinity of submarine canyons when the travel path is across the canyon. Therefore, bottom reflection loss is minimum for travel paths normal to the slope. Greater losses are experienced for paths parallel to the slope. Also downslope propagation turns rays toward the normal to the bathymetric contours (Aubrat, 1963).

For example, the azimuth of the great circle path from the Near Islands to Eniwetok makes an angle of 100° with the normal to the trend of the Aleutian arc. The angle between the trend of the arc and the azimuth of the path to Oahu is 250°. Therefore, lower T-phase signal levels would be expected at Oahu than Eniwetok for earthquakes of a given magnitude in the Near Islands. Such a relationship is apparent in Figure 13 in that the signal level is higher at Eniwetok than at Oahu, even though Eniwetok is 600 miles further away.

Slope Material

Weathered volcanic rock and glacially rafted pebbles on the insular shelves have been locally transported over the edge of the shelf and down onto the insular slope of the Aleutian Islands (Scruton, 1953; Bezrukov, Lisitsyn, Romankevich and Skorniakova, 1960). However, the individual pebbles and boulders are too small to cause scattering
loss for wave lengths involved in the T phase. Ewing, Ludwig, and Ewing (1965) reported no detectable amount of unconsolidated sediments on the slope, using seismic reflection techniques. Therefore, the material upon which the downslope bottom reflected rays probably impinge corresponds to the 3.2 km/sec layer from Shor's (1964; profile CK-8) refraction station. The reflection coefficient between the bottom water and the effective bottom reflector is 0.67.

Topographic Reflectors

Topographic echoes from T phases have been discussed by Aubrat (1963) and Shurbet and Ewing (1957). More recently, bottom reflections from the Aleutian arc have been noted by Northrop (1967) and Daubin (1964). Such reflections are in the frequency range below 20 Hz and appear to be stronger for steeper than for gradual slopes. By these criteria, the Aleutian arc is a good reflector because of the lack of sediment on the slope and the steepness of the slope, as shown in the previous section. It is not surprising, therefore, that TA-R reflections from the insular shelf north of offshore earthquake epicenters are present on the T-phase records. A plot of time intervals between peaks on Eniwetok records vs distance from the 100 fathom curve is shown on Figure 20. The theoretical curve drawn in for travel time was computed from the relationship

$$\Delta T = \frac{2D}{C}$$

where $\Delta T$ equals the time between signal peaks corresponding to arrivals $T_A$ and $T_{A-R}$ on the record, $D$ is the distance from the
FIGURE 20. Graph of time difference between peaks vs epicentral distance from the 100 fm curve. Theoretical relation shown for topographic reflections $T_{A-R}$ (upper line) and $PT^*$ (lower line). Data from this paper shown as dots, data from Johnson, Norris and Duennebier (1967) shown as circles.
epicenter to the 100 fathom curve and \( C \) is the speed of sound in the SOFAR channel. Distances to about 30 miles correspond to reflections from epicenters on the slope, and beyond 30 miles they correspond to epicenters in the Trench and outer ridge. For Trench and outer ridge epicenters, the reflected arrival \( T_{A-R} \) is considerably weaker than the first peak and in some cases (137, 138) is absent. The maximum range of values plotted in Figure 20 therefore appears to be limited to about 120 sec between peaks for earthquakes of the magnitude represented. A minimum time of about 20 sec is also indicated, as shorter intervals are difficult to ascertain on the record because of the emergent character of the peaks. Also it should be noted that for epicenters at or near the 100 fathom curve, the initial T phase is radiated from promontories along the edge of the shelf rather than from directly above the epicenter. If a T phase is radiated from both the insular shelf behind the epicenter (PT*) and from the ocean surface at the epicenter (\( T_A \)), the time difference between peaks is \( 5/6 \) \( D \), as shown by the lower curve of Figure 20 (Johnson, et al., 1967). Indeed, the time lapse between the peaks on the records varies with the cause of the peak.

**Oceanographic Factors Affecting T-Phase Signals**

**Sound Channel Depth**

Depth of the sound velocity minimum, both in the source area and along the path from source to receiver, is a major factor in SOFAR propagation of the T phase from the Aleutian Islands to the PMR receivers. In the Aleutian Islands area, cold surface temperatures
reduce the sound velocity to 1.463 km/sec in winter, and the minimum is at the surface (Fig. 9). In summer, the minimum velocity is at a depth of 150 m, and at the surface the sound speed is increased to 1.483 km/sec (Fig. 9). However, for the long wave lengths of sound in the T phase, the seasonal difference is negligible.

Along the path to the receivers, the depth of the minimum velocity increases to 780 m at 37° N and 900 m at 20° N (Figs. 2 and 9). The corresponding sound channel minimum velocities increase from 1.479 km/sec at 37° N to 1.484 km/sec at 24° N (Fig. 21). The place of changeover from RSR to RRR propagation occurs when the surface sound velocity becomes equal to that at the depth of greatest penetration of a particular ray.

The size of the ray bundle at the source determines the part of the total energy that may be transmitted, which in turn affects the strength of the observed signal level. The bundle of rays involved lies between the vertical and the critical angle (p. 10).

**Propagation Loss**

The T-phase signal level is further influenced by the distance for which propagation is in the RSR mode because of spreading loss. Spreading loss due to RSR transmission is between spherical and cylindrical loss (M. Pederson, personal communication).

Spherical spreading loss is experienced in the water along the initial ray path from the ocean floor to the first surface reflection. According to these theoretical considerations, propagation loss of T-phase energy after the rays have been refracted into the ocean is
maximum before the first reflection, becomes minimum after RRR propagation is achieved and is intermediate for the RSR section in between.

Hydrophone Location

The amount of T-phase energy received at a hydrophone is also dependent upon the size of the ray bundle (number of rays) that is "seen" by the hydrophone. For hydrophones placed on a sloping bottom (as at Oahu), rays striking the bottom seaward of the hydrophone are reflected up to the surface at least once before reaching the hydrophone. Upon each up-slope bottom reflection, the ray angle with the bottom is increased by twice the bottom slope (Luskin, Landisman, Tirey and Hamilton, 1952). At each reflection, some loss is experienced due to scattering and absorption in the bottom, and the amount of loss increases with increasing steepness of the ray. Additionally, irregular topography downslope from the hydrophones may cause acoustic shadowing of the bottom reflected rays, so that for practical purposes only the ray that just grazes the bottom will reach the hydrophone unattenuated by losses due to bottom reflection. In the Kaneohe area, this restriction is more pronounced than at the other stations.

For hydrophones that are suspended in the SOFAR channel, as at Midway #22, #23, #24; Wake #32, #34; and Eniwetok #43 and #44, such restrictions are not placed on the rays "seen" by the hydrophones. For these installations, then, a larger bundle of rays reaches the suspended hydrophones. The limiting ray is the ray that just grazes the bottom (Fig. 2). This ray undergoes successive surface
reflections due to refraction of the rays (Fig. 2). This phenomenon is called 35 mile peaking (Frosch, 1964) and is characteristic of RSR propagation in the deep ocean. However, surface reflection losses at near grazing incidence are negligible (Brown and Ricard, 1960). Steeper rays are bottom-reflected and although energy loss from bottom scattering and absorption may be as low as 1 dB per reflection (Jitkovskiy and Volovova, 1965), a minimum of 60 reflections is encountered between the Aleutian Islands and the Hawaiian area. Therefore, the contribution to the T-phase signal from bottom-reflected rays at the distant PMR hydrophones is minimal.

Bathymetry

The bundle of effective rays is restricted to those that are either RRR or RSR at the hydrophone. In the Hawaiian area in winter, the 16 degree ray is the limiting ray for water depths in excess of 4.8 km. Because the sound speed in water at these depths is greater than at the surface, the bottom limited ray is surface reflected at near grazing incidence with, presumably, no loss of energy.

Water depth between the hydrophone and the source is therefore the controlling factor in determining the number of rays that are seen at a particular detection site. The bundle of RSR rays is slightly larger in winter than in summer because of lower surface temperatures (and resulting lower surface velocities) in winter. Thus, the sound channel is thinner, the bottom of the sound channel is further above the ocean floor and the depth excess (water depth greater than sound channel depth) is greater in winter than in summer (Fig. 2).
difference, however, is small, for in a typical example in Hawaiian waters the 15° ray is the bottom limited ray in summer and the 16° ray is the bottom limited ray in winter.
CHAPTER VII
DISCUSSION

Cause of Errors in T-Phase Source Location

General Statement

For a detailed description of the source location program used, see Johnson (1966). Briefly, the source solution obtained is given by that origin time, latitude and longitude for which the sum of the squares of the station residuals (observed minus computed arrival time) is a minimum. The basic formula can be written

\[ T_0 = T_M - \Delta/C \]

where \( T_0 \) is the computed origin time, \( T_M \) is the measured arrival time, \( \Delta \) is the distance between source and receiver and \( C \) is the speed of sound in the SOFAR channel. The two variables determining the fix accuracy are: (1) the ability to pick and record the arrival time correctly and (2) knowledge of SOFAR velocities between source and receiver.

Comparison of T-Phase Source Solutions with Published Earthquake Epicenters

The difference in origin times of the T-phase source solutions from the body wave determinations studied is shown in Figure 12. A time difference of 1 second indicates a geographic difference of about 1 mile. Some T-phase radiator locations are 60 miles shoreward from the earthquake epicenters and indicate areas from which PT* waves were radiated. The difference between epicenter and T-phase source
is therefore 60 miles for these events. Additionally, because Stalemate Bank, Attu Island, Wells Plateau and Kiska Island all radiate T-phase energy (Johnson and Norris, 1966), the T-phase peak on recordings from different stations may not represent the same T-phase source. Triangulation back to the source then places the T-phase fix behind the row of radiators.

Errors Due to Timing Uncertainties

Because of the slow recording speed used (0.25 mm/sec), and the broadly peaked character of the signals, arrival times were accurate only to within 3 seconds. However, the digitized times, which were used as input for the source location program, were indicated to the nearest second. Therefore an error in arrival times, of the order of ± 1 second, is inherent in the data. Additionally, there are limitations imposed on the timing accuracy by the nature of the recordings. It will be noted on the records (Fig. 4) that there are two timing standards, a WWVH radio trace, and internal clock times. Although the WWVH standard is used in data reduction when available, onset of a five minute mark on the radio trace is not always clear. Sometimes it is blurred so that it can only be estimated to the nearest ± 1 second. In some instances, the radio trace is not present, due to broken pens, electronic failure or fading of radio waves due to changing propagation conditions. In these cases, arrival times must be digitized with reference to the secondary time standard which can be off by a minute. Such poor readings were avoided when possible, but were used when necessary. Even when synchronized with WWVH, the
reading error was about 2 seconds. Therefore, the reading of arrival times on the records due to timing uncertainties alone is certainly no better than about 3 seconds in most cases, and can be as large as 1 minute. Such poor readings, however, are often detected by the iterative least squares T-phase source solution program and poorly timed events dropped from the iteration.

Additional inaccuracies in the reading of arrival times comes from the nature of the signal itself. For example, the signal may be multi-peaked due to various ray paths to a given hydrophone. In these cases, the latest peak in time was read because that would supposedly be a SOFAR arrival (p. 9), and because SOFAR velocities were used in computing the source area, we hoped to read SOFAR arrival times. In some instances it is difficult to pick the SOFAR arrival because sometimes the latest arrival is not the peak arrival, or there may be several closely spaced arrivals due to local station interference, overload, etc. The net result is that the OSCAR operator must use some judgment in picking the correct arrival to read, and different operators read slightly different peaks. The spread of possible values for low level events appears to be near 4 seconds, but for high level events which overload the system, errors of twice that magnitude are present. Accuracy of source location is in general better for low level events than for large ones, other things being equal.

Errors Due to Uncertainties in Velocities

A fundamental assumption made in the T-phase source location
program is that the propagation of energy from source to receiver is via the deep sound channel (Barbour and Woollard, 1949). For the T-phase location work, there is a systematic error introduced into the computation because the events all originate at ocean bottom depths. In areas of the northern Pacific where the sound channel minimum is at or near the surface, the error is negligible for epicenters near the edge of the shelf. An estimate of the error involved can be made from a comparison of the velocity difference between the shallow and deep propagation paths. Near latitude 40° N, for example, the velocity difference is about 100 ft/sec out of 5000 ft/sec, or 2%.

There is an additional error inherent in the source location program due to the velocity function used. The velocity function is the least squares fit to the average SOFAR velocities determined from shot calibration and from velocities averaged across a contour chart of the local velocity to selected zones of the Pacific (Johnson and Norris, 1964; Johnson, 1966). SOFAR velocities are generally known to an accuracy of 1 m/sec, so that errors in fixes due to velocity variations alone are of the order of 1 in 1500, or 0.00067.

Errors Due to Small Azimuthal Distribution of Stations

For source-station geometries in which the azimuthal distribution of stations is limited, poor resolution is obtained. This problem has been treated extensively in the literature (see, for example, Flinn, 1965) and is discussed in some detail by Brumbach (1966). In the T-phase source location, a flat earth and zero parallax are assumed in
the vicinity of the trial solution. In the succeeding iterations, the new source location is compared with the preceding one. If the difference is 0.1 degree or more, the last position is taken as a new trial position and the iteration continues. If the difference is less than 0.1 degree, the solution is retained and the distances and origin time computed. In areas where the azimuthal distribution of stations is small, the distance between successive source solutions may be too large for the flat-earth approximation to be valid. When all stations lie on a line passing through the source, the distance to the source is impossible to compute and the azimuth is poorly defined. A good illustration of such a hyperbola of positions was obtained for a series of shots off the west coast of Washington. A series of 363 explosions in September 1965 was detected on the records as coming from the same location, but the resulting source solutions lay along a hyperbola extending from Midway Island to Montana. This type of error in source positioning is both the greatest in magnitude and the most difficult to detect. Discrepancies of the order of 2000 miles were encountered, although the program output shows a small standard deviation of origin times for individual fixes. These erroneous locations were removed before the final T-phase source solutions were made wherever possible.

Comparison of Source Solutions with Events of Known Origin

It is difficult to compare the epicenter locations based on P-wave data, as those of the USC&GS are, with T-phase source solutions because both locations are subject to similar errors. On the other hand, underwater explosions at known time and place are numerous in the ocean.
Indeed, long series of shots were often noted on the records.

The repeatability of recorded signal level and spacing between events led to inquiries of the Naval establishment as to the originator of the shots. One such series was completed along a great circle route from California to Japan. This shot run provided an excellent calibration for the location accuracy of the hydrophone network in the northern Pacific.

**U.S. Navy Shots in the Aleutian Islands Area**

In November 1965 depth charges were dropped along a great circle route between California and Japan by the U.S. Navy, as a joint effort of the Marine Physical Laboratory of Scripps Institution of Oceanography and Hudson Laboratories of Columbia University. A comparison of the drop locations with 377 source solutions of the hydrophone network shows that source solutions in the central area of the run are within the program accuracy (± 6.0 miles), and are relatively poor at both the eastern and western ends. In this series of events, only the island stations' arrival times were available for computation, except for the set near 120° W - 130° W, where arrival times at a California station were available. The variability in shot location accuracy is thus primarily one of source-station geometry. The relationship of good fixes to large aperture angles between azimuths to different receivers is readily apparent. On a geographic basis, fixes of the order of ± 10 miles error extend from 50° N, 180° to 50° N, 170° E, (Spiess, Northrop and Werner, 1968).
Errors Due to Depth of Focus Uncertainty

An error inherent in locating earthquake epicenters by the T-phase method is that the distance from hypocenter to the point on the ocean floor where P waves are refracted into the ocean as T waves is unknown at the time of computing the T-phase source area. This can best be thought of in terms of computed origin times, i.e., the T-phase source time should always be later than the actual occurrence of the earthquake by the amount of time it would take the body waves to travel from the hypocenter to the ocean floor. Therefore an analysis of the difference in origin times, as computed from seismographic vs hydrophone data, with the T-phase source location times, is of interest. The spread of values for the Rat Island aftershocks appears to be representative of that for other areas, with a slight tendency for the T-phase computed time to be late, as expected on theoretical considerations given above (Fig. 12).

Because the majority of the T phases examined in this report came from shallow focus (< 60 km) earthquakes, the P-wave travel time amounts to only 7.5 seconds (assuming 6.6 km/sec for the propagation velocity of P waves in the upper mantle and crust) and always has the same sense. However, if the depth of focus of an earthquake is known from some other determination, the travel time of P for the mantle-crust portion of the travel path could be subtracted from the origin time, as arrived at by T-phase delay time positioning, to make the T-phase origin time come in closer agreement with the earthquake epicentral time as published by the USC&GS.
Errors Due to Duration of the T Phase

A major cause of error in accurate determination of T-phase source solutions is the duration of the T phase itself. Because a T-phase signature often exhibits multiple peaks caused by multiple radiators near the source (Johnson and Norris, 1966), the duration of the T phase is long. For example, the T-phase signal for the main shock of magnitude 7-3/4 was 7 minutes, and the recorders were overloaded for 3 minutes so that the peak arrival time was not readable. In picking the same arrival on various hydrophone records, only gross similarities in the signature can be used reliably as the fine structure is greatly influenced by bathymetry near both the source and receiver. A seamount near a receiver causes a reflection if the seamount peak is shallower than the sound channel axis. The slope of a seamount near the source may radiate energy as PT*. For great circle paths of different azimuths, respective reflectors and radiators vary.

For epicenters near the ocean boundary the duration of the T phase is largely a function of the size of the area of the ocean floor insonified by the earthquake. In other words, the larger the earthquake, the longer the duration of the T phase (Northrop, 1965; Northrop and Johnson, 1965; Daubin, 1964; Brazee, 1965). Because of the characteristically emergent onset and decay of the T phase, the signal-to-noise ratio of the detection system also affects the apparent T-phase duration. For example, land-based recording of T phases from conventional seismographic installations on islands, which have high microseism noise at frequencies near 1 Hz (Brune and
Oliver, 1959; Bradner and Dodds, 1964), generally show T-phase signatures that are difficult to distinguish from background noise (Shurbet, 1962). Nevertheless, some authors (Ben-Menahem and Toksoz, 1963; Eaton, Richter and Ault, 1961) have tried to relate the duration of a T phase recorded on a land-based seismograph on the island of Hawaii with the duration of faulting. Their interpretation has been disputed by Gupta (1964), and the reliability of making such a comparison has been questioned by Northrop and Johnson (1965).

However, for the majority of T phases from the Aleutian Islands discussed herein, the earthquake magnitude is less than 5.5, and the earthquakes are not known to be associated with any surface expression of strike-slip faults, as were the large (8.0 mag) earthquakes discussed by Ben-Menahem and Toksoz (1963) or Eaton, Richter and Ault (1961).

**Threshold of Detection**

**Comparison with Published Earthquake Magnitudes**

Earthquake magnitude and T-phase strength may be compared for those events in the T-phase list which correspond to events on the PDE cards. A plot of these quantities for all the Aleutian Island earthquakes during the period of August 15, 1964 to January 31, 1965, has been published by Johnson and Northrop (1966). This curve is used as a reference in Figure 11.

In plotting the T-phase strength vs earthquake magnitude for the 144 earthquakes in the Rat Island aftershock sequence studied in this thesis, it was found that some of the earthquakes of magnitude 4-5
produced very low-level (15-20 dB) T phases (Fig. 11). Epicenters which caused the low-level T phases were seaward of the 2000 fm contour, i.e., from the Aleutian bench, within the Trench itself or from the outer ridge. The data points are plotted as B, T and O respectively in Figure 11.

T phases of higher strength were observed from epicenters on the insular shelf and slope. In fact, those on the insular shelf (marked I, Fig. 11) were predominantly of greater strength than those from the slope and bench (marked S and B respectively in Fig. 11).

T Phases from Events not Listed as Earthquakes by the USC&GS

Of the 4900 T-phase source solutions obtained during a six month period from September 1964 to February 1965, only 1360 corresponded with earthquake epicenters published by the USC&GS. The remaining 3540 events are believed to be from low magnitude earthquakes which were below the threshold of detection by the USC&GS network (Johnson and Northrop, 1966). From 10 to 100 times as many events were detected by the hydrophone network as by conventional seismograph stations for the Kuril Islands, Kamchatka, the Aleutian Islands, Alaska, Canada, the west coast of the United States and the East Pacific Rise. In the area of the Rat Islands, the bulk of the smaller shocks not detected by the seismograph network were foreshocks and aftershocks of large magnitude earthquakes. The lower magnitude events often came in "swarms" of aftershocks. The number of smaller aftershocks is known to increase exponentially according to the equation
$\log N = a + b (8 - M)$

where $N$ is the annual frequency per magnitude interval and $M$ is the magnitude on the Richter scale. For shallow earthquakes (which were predominantly the originators of the T phases studied) the world average value of $a$ is $-0.48 \pm 0.02$, and that of $b$ is $0.90 \pm 0.02$ (Gutenberg and Richter, 1949). If this world average is used for, say, $M = 4$, there are about 10000 times as many magnitude 4 events as magnitude 8 events. In other words, for every earthquake of a given magnitude, there are roughly ten times as many of the next lower magnitude class. Because the USC&GS rarely locates events of magnitude less than 4, it would be expected that at least ten times that many magnitude 3 (and 100 times as many magnitude 2) shocks had occurred and remained undetected (Jordan, Lander and Black, 1965; Johnson and Northrop, 1966). Of course there are a great many more smaller earthquakes, of the order of a million each year, but "the frequency cannot go on increasing indefinitely with decreasing magnitude, since a certain minimum stress must exist to produce an earthquake" (Gutenberg and Richter, 1949).
CHAPTER VIII
SUMMARY AND CONCLUSIONS

Earthquake body waves refracted into the water at the ocean floor are called T waves. They are propagated over long distances in the ocean as compressional waves in water and travel at the velocity of sound in the deep ocean sound channel. The distance, both vertical and horizontal, that the earthquake hypocenter is from the intersection of the sound channel axis and the ocean boundary affects the T-phase peak pressure level, duration and frequency content. Earthquakes of magnitude 3.5 - 6.5 in the western Aleutian Islands cause T phases of 1 - 7 minutes duration and signal levels of from 6 - 65 dB re 0.1 microbar at 1800 miles. Although the T phase contains frequencies from 1 - 100 Hz, the effect of the recording system used was to filter out frequencies below 5 Hz and above 15 Hz. The duration of the T phase was longer for large earthquakes than for small ones, all other things being equal. In addition, the duration of the T phase is increased further by the fact that T waves from a given earthquake may be introduced into the sound channel at more than one source area.

In the western Aleutian Islands area, the sound channel is near the surface, the ocean bottom slopes are near 5° and there is little sediment on the slope. For earthquakes on the Aleutian arc, T-waves are introduced into the sound channel by the classical method of down slope propagation paths along the sloping bottom. For earthquakes beneath the ocean floor south of the Aleutian arc, abyssally generated T waves are produced in the near-surface sound channel by
some mechanism such as scattering of sound rays. The former mechanism is more efficient and excites lower frequencies than the latter. Therefore, earthquakes of a given magnitude cause lower signal levels and higher frequency content if they occur in the Aleutian Trench or outer ridge than if they occur on the Aleutian arch.

T phases from the western Aleutian Islands area are often multiple peaked signals. Multiple peaked T phases from earthquakes on the Aleutian arch are caused by generation of T waves at more than one bottom slope near the epicenter. Earthquakes in the Aleutian Trench and outer ridge produce multiple peaked T phases because T waves are introduced into the near-surface sound channel both at the epicenter (T_A) and from the insular slope behind the epicenter (PT*). Additionally, topographic reflection of T waves from the ocean bottom slope behind the epicenter sometimes causes a third peak (T_A-R).

The accuracy of T-phase source location varies with the duration of the T-phase signal received, the sharpness of the signal peak, the station distribution, timing accuracy of the recorders, the accuracy with which the signal peak can be read, knowledge of sound channel velocities and (for multiple peaked signals) correct identification of signal peaks and geographical distribution of bottom slopes causing the separate peaks. Although identification of the T-phase source area is of interest, it may not reveal the actual earthquake epicenter. For earthquakes on the Aleutian arch, and behind the arch, accuracy of epicentral location is poor ( ~ ± 60 mi) because the T-phase source
generating area is along the insular slope within the area disturbed by the earthquake rather than at the epicenter itself. Thus, although the identification of individual peaks on different station recordings may be correctly done, the resulting source solution will indicate the T-phase radiator, not the earthquake epicenter. For earthquakes beneath the Aleutian Trench and outer ridge, accuracy of earthquake epicentral location by T phase is better (~ ± 15 mi) because the surface generated T phase is excited directly above the hypocenter and the signal is sharply peaked, easily recognized, and has a relatively low signal level for earthquakes of a given magnitude.

In the case of large (> 5 mag) earthquakes on the Aleutian arch, the T-phase signal may overload the recording system used. On such records that are overloaded it is impossible to read the peak arrival time. A special study of the T-phase signal, as recorded on magnetic tape, may then be made in which the signal is replayed at a lower amplification onto strip charts from the magnetic tape recording. If such magnetic tapes are not available, however, a somewhat anomalous result of the investigation is that large earthquakes are located with less precision by T phase than small earthquakes. In all cases it should be remembered that the T-phase source solution obtained is the location for which the standard deviation of origin times is a minimum. It does not follow that an earthquake (or explosion) took place at that exact location.

The same qualification applies for source solutions obtained using the hydrophone recordings of hydroacoustic waves from explosions
to locate the detonation point, or for seismograph recordings of body waves used to locate underground explosions (Johnson, 1966).

In the case of underwater explosions, the source solution may be somewhat better than for earthquake T phases because the signal peak of an underwater explosion has a sharper peak. However, underground explosions on islands generate a hydroacoustic signal that is similar to an earthquake T phase. A more sophisticated recording and signal processing technique is required before definite criteria can be made for differentiating hydroacoustic waves due to underground explosions in the western Aleutian Islands from T-phase signals from earthquakes beneath the crest of the Aleutian arch.
APPENDIX

History of T-phase Research

The name T (for Tertiary) was coined by Linehan (1940) because the waves were observed to arrive later than the primary (P) and secondary (S) body waves. Linehan observed that the T-phase signal was rich in frequencies near 10 Hz, had a duration of several minutes and emerged slowly from the background ambient noise. He also noted that even though the amplitude and duration of the T phase were different for different earthquakes, the arrival times of T (at Weston, Mass.) were always 23 minutes later than the arrival times of P for earthquakes in the West Indies, but he gave no explanation of this phenomenon.

Coulomb and Molard (1949) reported T-phase velocities of from 1.58 to 2.36 km/sec from earthquakes in the Caribbean, as recorded on seismographs on islands in the Antilles, and concluded that the T phase was propagated as shear waves in the ocean floor sediment.

Tolstoy and Ewing (1950) showed that the late arrival times of T waves from the Caribbean, as recorded at Weston, Massachusetts, could be accounted for if propagation over the land portions of the travel path was via P waves in the earth's crust at a speed of 6.6 km/sec, and at the speed of sound in water, 1.5 km/sec, over the oceanic portion of the travel path. They pointed out that variations in the average velocity of T waves could be caused by local geological conditions affecting the speed of P waves in the vicinity of the source and receiver.

Leet, Linehan and Berger (1951) contended that T-phase propagation
was via shear waves with a velocity of between 1.6 and 2.7 km/sec over oceanic paths, and 2.1 km/sec over land.

Molard (1952) restated the case for ocean bottom propagation paths of the T phase for earthquakes in the Caribbean as recorded on seismographs in the Antilles. Ewing, Press and Worzel (1952) then reported on measurement of arrival times of T waves as recorded on pressure sensitive detectors (hydrophones) placed in the ocean. Because shear waves do not propagate in liquids, the fact that T waves travel as compressional waves in water was thus firmly established. Furthermore, by effectively eliminating the land travel path of T-wave energy from the ocean basin margin to a land based seismograph recording site, Ewing et al. were able to show that the observed velocity of T waves was $1.47 \pm 0.01$ km/sec, the same as that previously measured for underwater sound propagation in the deep ocean sound channel (Ewing and Worzel, 1948).

The problem of T-phase generation and propagation has been the subject of a number of investigations since the early controversy (Leet, Linehan and Berger, 1951; Ewing, Press and Worzel, 1952).

A brief summary of the T-phase problem is given by Burke-Gaffney (1954), who reported on T phases from the New Zealand region. He concluded that the mean velocity of T waves from Pacific earthquake epicenters to seismographs in New Zealand was 1.53 km/sec. In comparing his data with that of the early workers in the field, Burke-Gaffney concluded that the velocity of T waves was too great to conform to the Ewing et al. theory that T waves are water borne, but at too low a speed to bear out the Leet et al. hypothesis of shear wave velocities near
1.7 km/sec.

Byerly and Herrick (1954) measured travel paths of the T phase from earthquakes in Hawaii to seismographs in the Berkeley, California area. A velocity of 6 km/sec from the earthquake epicenter to the 500 fathom curve, and from the same contour to the seismograph, was assumed for the calculation of P-wave travel time in their analysis. The travel times, which accounted for the portion of the travel path of the T phase in the solid earth, were then subtracted from the total great circle travel time of the T phase between epicenter and seismograph. The remaining time divided by the remaining distance gave a speed of 1.47 km/sec for propagation of the T phase over oceanic paths.

Further studies of the T phase, as recorded on seismograph stations at Iceland (Bath, 1954), Bermuda (Shrubet, 1955; Shrubet and Ewing, 1957), Japan (Wadati and Inouye, 1956; Wadati, 1960), Hawaii (Krivoy and Eppley, 1964) and the Caribbean (Aubrat, 1963), generally confirmed the hypothesis of T-phase propagation within the water column. However, these authors pointed out that irregularities in the observed T-phase propagation velocity, signal duration and amplitude are caused by local geological conditions in the vicinity of the source, as well as in the vicinity of the seismograph station. The background noise level also varies from station to station, and from time to time, at a given station because of local wind and surf conditions (Bradner and Dodds, 1964). Furthermore, T waves are attenuated during the travel path from ocean boundary to seismograph. Both of these phenomena, microseismic noise on islands and attenuation of T waves propagating in rock, reduce the signal-to-noise ratio of
the T phase as recorded on seismographs. The onset time of the T phase is consequently difficult to read accurately on seismograms. This uncertainty in reading the onset time of T waves increases the uncertainty in computing their velocity, just as in the case of reading the P- and S-wave arrival times. In contrast, the T-phase signal as recorded on hydrophones placed in the deep ocean sound channel has a peak which represents the sound channel arrival. If this peak is chosen as the arrival time of the T phase, an accurate measure of the velocity of T waves is possible.

Studies of the T phase as recorded in the Atlantic sound channel were reported by Shurbet (1955: 1962), Shurbet and Ewing (1957), Northrop, Blaik and Tolstoy (1960) and Northrop (1962). Johnson, Northrop and Eppley (1963) discussed T-phase sources in the Pacific. Milne (1959) compared the T-phase signal from an explosion within Eniwetok Atoll with that from an ocean margin earthquake. Later papers by Johnson (1963; 1964; 1965) and Johnson and Norris (1964; 1966) discussed sound channel propagation and the source location problem. Johnson and Northrop (1966) discussed the relation between T phases and earthquake magnitude, and Northrop (1965) studied T phases from the Alaska earthquake of March 27, 1964. Particular aspects of T-phase propagation were studied by Daubin (1964), Brazee (1965), and Johnson, Norris and Duennebier (1967). These works bear out the Ewing-Tolstoy-Press-Worzel theory that T waves propagate as hydroacoustic waves in the deep ocean sound channel.
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