Bulletin of the Seismological Society of America. Vol. 52, No. 2, pp. 333–357. April, 1962

# RAYLEIGH WAVE DISPERSION IN THE PACIFIC OCEAN FOR THE PERIOD RANGE 20 TO 140 SECONDS

BY JOHN KUO, JAMES BRUNE, AND MAURICE MAJOR

ABSTRACT

Rayleigh wave data obtained from Columbia long-period seismographs installed during the International Geophysical Year (I.G.Y.) at Honolulu, Hawaii; Suva, Fiji; and Mt. Tsukuba, Japan, are analyzed to determine group and phase velocities in the Pacific for the period range 20 to 140 seconds. Group velocities are determined by usual techniques (Ewing and Press, 1952, p. 377). Phase velocities are determined by assuming the initial phase to be independent of period and choosing the initial phase so that the phase velocity curve agrees in the long period range with the phase velocity curve of the mantle Rayleigh wave given by Brune (1961). Correlations of wave trains between the stations Honolulu and Mt. Tsukuba are used to obtain phase velocity values independent of initial phase.

The group velocity rises from 3.5 km/sec at a period of about 20 sec to a maximum of 4.0 km/sec at a period of about 40 sec and then decreases to 3.65 km/sec at a period of about 140 sec. Phase velocity is nearly constant in the period range 30–75 sec with a value slightly greater than 4.0 km/sec. Most of the phase velocity curves indicate a maximum and a minimum at periods of approximately 30 and 50 sec respectively. At longer periods the phase velocities increase to 4.18 km/sec at a period of 120 sec.

Except across the Melanesian-New Zealand region, dispersion curves for paths of Rayleigh waves throughout the Pacific basin proper are rather uniform and agree fairly well with theoretical dispersion curves for models with a normal oceanic crust and a low velocity channel. Both phase and group velocities are comparatively lower for the paths of Rayleigh waves across the Melanesian-New Zealand region, suggesting a thicker crustal layer and/or lower crustal velocities in this region.

## INTRODUCTION

Matched three-component, long-period seismographs with high magnification were installed at Mt. Tsukuba, Japan; Honolulu, Hawaii; and Suva, Fiji, during the I.G.Y. Pure, long oceanic paths between these stations and the epicenters of the circum-Pacific belt earthquakes allow Rayleigh wave trains to become well dispersed. Certain of the large earthquakes thus recorded at Mt. Tsukuba, Honolulu, and Suva exhibit the long period waves associated with the inverse branch of Rayleigh wave dispersion as well as the short period waves associated with the normal branch. The period range of the wave trains extends from less than 20 sec to 140 seconds. These excellent data are used in this paper to determine both group and phase velocities. A preliminary interpretation of the structure of the earth's crust and the upper mantle beneath the Pacific is made by comparing the observed dispersion curves with theoretical curves in this period range.

Earlier studies (Oliver *et al*, 1955, Santô, 1960, a,b,c) dealing with the Rayleigh wave dispersion in the Pacific were limited to waves with group velocities of periods less than 40 seconds. Group velocities, if observed over a limited range of periods, do not uniquely determine the phase velocities. Accurate measurements of phase velocity are important for determining physical properties of the earth and are essential for studying the mechanism of the disturbance at the source by surface wave methods.

## Methods

The method of determining the group velocity was discussed in earlier papers by Ewing and Press (1952, p. 377), Pekeris (1948, p. 8), and Oliver *et al.* (1955, p. 922). The periods are determined by measuring the slope of a peak (or trough) vs. arrival time curve. The epicentral distance divided by the group arrival time gives the group velocity.

Because excellent Rayleigh wave data are available it is feasible to apply Brune's simplified method of determining the phase velocity, which was discussed in detail by Brune, Nafe, and Oliver (1960). If a constant of integration related to the initial phase is assumed or can be determined from an independent source, the phase velocity may be calculated from the arrival times of peaks and troughs on the record. His method, referred to here as single-station method, is summarized in the following paragraph.

When the wave train is sufficiently dispersed, the stationary phase approximation may be used to determine the phase velocity from the seismogram at a given time and at a given distance without resorting to direct Fourier analysis, provided that the phase correction $\pm\pi/4$ due to dispersion, and an additional correction $-\pi/4$ due to the non-sinusoidal shape of spherical harmonic waves, are made. If we assume at the origin an initial phase, $\phi_0$[1], independent of period, the phase velocity can be determined from a single station using the following equation:

$$C = \frac{x}{t - \left(\frac{\phi_b - \phi_0}{2\pi} + N - \frac{1}{8} \pm \frac{1}{8}\right)} \tag{1}$$

where $C$ is the phase velocity, $N$'s are assigned to successively observed phases (crests or troughs, for example) of the wave train, $x$ is the distance, $\phi_b$ is the phase observed on the record, and for convenience we read only peaks or troughs, so the $\phi_b$ is either 0 or $\pi$, and $t$ the travel time.

Another method is the two-station correlation method. If at two stations along the path of surface wave propagation at distances $x_a$ and $x_b$ we observe phases $\phi_a$ and $\phi_b$ on an infinite sinusoidal Fourier component of period $T$, the following relation, apart from the polar shift and effects of instrumental phase shift, may be written:

$$C = \frac{\Delta x}{\Delta t - \left(\frac{\phi_b - \phi_a}{2\pi} + N\right)T} \tag{2}$$

where $\Delta x$ is the distance $x_b - x_a$ and $\Delta t$ the difference in time $t_b - t_a$.

Application of either of the above two methods leads to discrete sets of phase-velocity curves corresponding to the different assignment of $N$'s. Ambiguity in assigning $\phi_0$ and $N$ may be resolved by a known initial phase, a known limiting value of $C$, or a known $C$ at some particular period predetermined independently.

[1] The definition of initial phase $\phi_0$ given by Brune (1961) is used in this study. This differs from the apparent initial phase used by some authors (e.g. Aki, 1960) by the phase advance of $\pi/2$ due to the non-sinusoidal shape of the spherical harmonic wave functions near the origin.

In the present study, both the single-station method and the two-station correlation method are used to determine phase velocity. In the single-station method, an initial phase, independent of period, is assumed and from this a phase velocity curve is computed from equation (1). By increasing and decreasing the initial phase by an arbitrary amount, in this case $\pi/2$, two additional discrete phase velocity curves are similarly calculated. The curve which is best matched to the known limiting value of the phase velocity curve at periods of mantle Rayleigh waves previously observed is approximately the correct phase velocity curve. The phase velocity of mantle Rayleigh waves for periods greater than 130 seconds is found nearly independent of the paths (Brune, 1961). The two-station correlation method is next used to determine the phase velocity independently of any assumption of initial phase. The method requires an earthquake occurring on the great-circle which passes through the two stations. Apart from instrumental phase shifts, the phase velocity thus determined involves only the assignment of an integer $N$. By successively assigning integers, discrete sets of phase velocity curves can be obtained by equation (2). The separation between the various possible adjacent curves at long periods is sufficiently great that there is no ambiguity in choosing the correct curve if a comparison is made with the phase velocity curve for mantle Rayleigh waves given by Brune (1961). Discrete sets of phase velocity curves converge toward short periods and diverge toward long periods. Hence the error introduced in phase velocity by a small error in the initial phase will become negligible at short periods.

## Instruments

The instruments used were Columbia-type long-period seismographs. They were not calibrated, but the periods of the pendulum and the galvanometer and the values of the damping and coupling resistances were approximately known, so that instrumental phase shifts are known to better than one twentieth of the period. This source of error would not alter the conclusions reached in this paper.

The type 30-100A (free period of the seismometer pendulum, $T_0 = 30$ sec; free period of the galvanometer, $T_g = 100$ sec) was first used at the Suva station and changed to 15-80B on July 8, 1958. The type 15-80A was used at both Honolulu and Mt. Tsukuba prior to June 3, 1958 and March 29, 1959, respectively, and thereafter changed to 15-80B. (15-80A and 15-80B are the seismographs of $T_0 = 15$, $T_g = 80$ with different constants.) For a further description of instruments, the reader is referred to Sutton and Oliver (1959). Constants of the instruments are as follows:

| Instrument | Constants* | |
|---|---|---|
| 15-80A | $\epsilon_0 = 1.5\ \omega_0$ | $\epsilon_g = 6.0\ \omega_g$ |
| 15-80B | $\epsilon_0 = 3.0\ \omega_0$ | $\epsilon_g = 1.0\ \omega_g$ |
| 30-100A | $\epsilon_0 = 1.5\ \omega_0$ | $\epsilon_g = 6.0\ \omega_g$ |

* Where $\epsilon_0$, $\epsilon_g$ are damping coefficients of seismometer and galvanometer, respectively.

Background noise from microseisms is particularly strong at the Honolulu station, especially during the winter months. The long-period instruments, therefore, have a low gain. Vertical-component seismograms from the Suva station and the Mt. Tsukuba station have a lower level of background noise.

The instrumental group delay times for determination of group velocity and the instrumental phase shift corrections for determination of phase velocity are shown in figs. 1 and 2 and are applied in this study.

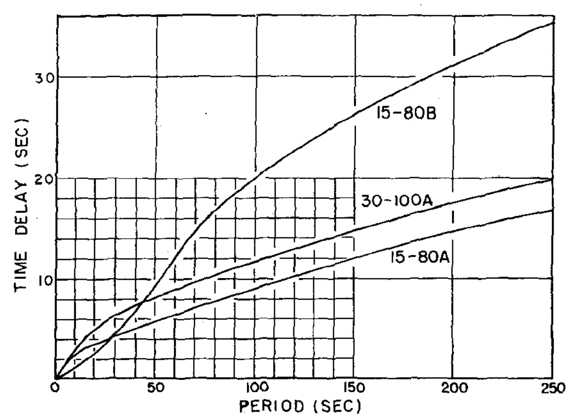


FIG. 1. Group delay time vs. period.

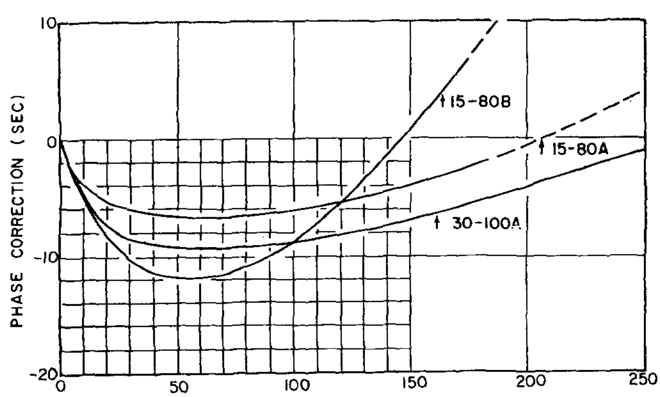


FIG. 2. Phase shift correction vs. period. A phase of $\pm \pi$ (or $\pm T/2$) must be added to the time correction so that the direction of ground motion corresponds to the direction of motion on the seismogram, Espinosa et al. (in press).

## DATA

Altogether 16 earthquakes, including 15 from the circum-Pacific belt and one from China, are analyzed. Figure 3 is the index map showing their geographic distribution. The earthquakes recorded at Suva, numbered consecutively from $P$ 1 to $P$ 14 clockwise, are used to determine the phase velocity by the single-station method. Two other earthquakes from China and Chile, numbered $C$ 1 and $C$ 2 respectively, are used to determine the phase velocity between Honolulu and Mt. Tsukuba by the two-station correlation method.

The selected paths of Rayleigh wave propagation cover the Pacific from Japan, the Kurile Islands, Kamchatka, the Aleutian Islands, through Northern California, the Gulf of California, Mexico, Southern Peru, Chile, Drake Passage, and the Melanesian-New Zealand region. With the exception of $P$ 7, $P$ 8 and $P$ 9, which are mixed paths with less than 4 % continental portions, and $C$ 1, a mixed path with large portions of continent, the paths of all other earthquakes are considered to be purely oceanic. No correction for the small continental portion of the paths $P$ 7, $P$ 8 and $P$ 9 is made. The continental portion of the path $C$ 1 is eliminated by the two-station correlation.

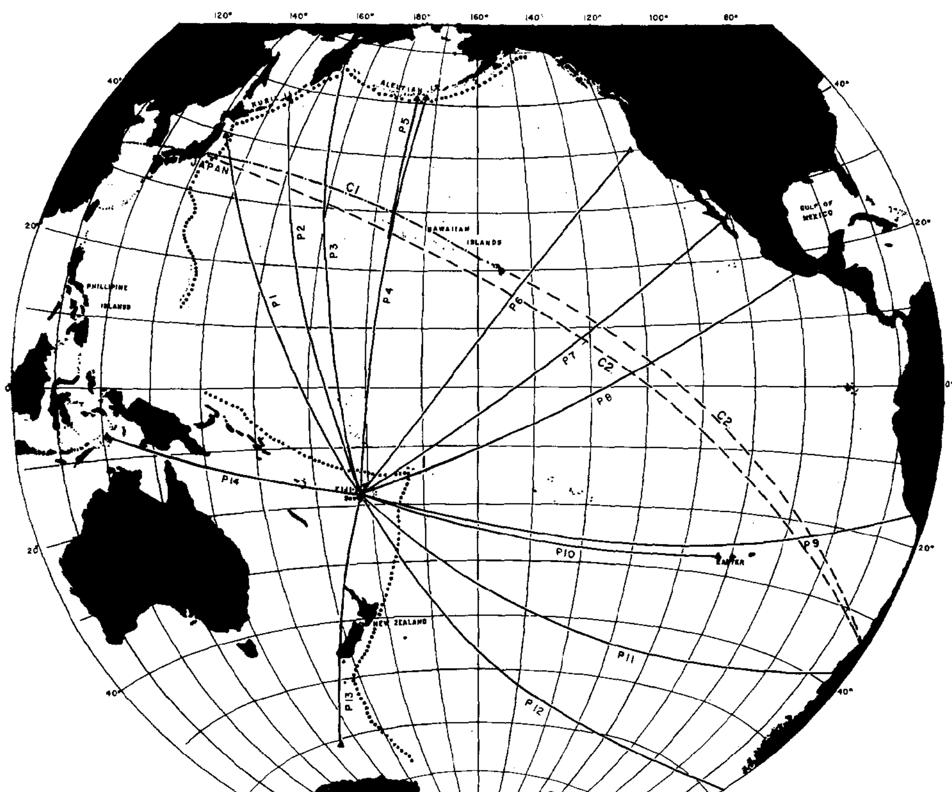


FIG. 3. Index map of the geographic distribution for the earthquakes analyzed. Dotted line indicates the andesite line.

Tables I and II list earthquakes used in this study recorded at Suva, Honolulu and Mt. Tsukuba during the years 1958–1960. The magnitudes of these earthquakes are all greater than or equal to 6. All the earthquakes analyzed are shallow shocks, except $P$ 2, $P$ 6, $P$ 7 and $P$ 9, which have focal depths around 100 km. Epicentral information is obtained from the United States Coast & Geodetic Survey and Bureau Central International Séismologique. Epicentral distances thus calculated range from approximately 3,400 to 17,200 kilometers.

Records used in this study are vertical-component records. Records from the horizontal components are not used for determination of group and phase velocities in order to eliminate the possible mixture of Love and Rayleigh waves; they are used only for identification of wave trains. The particle motion, whenever identifiable, is elliptical-retrograde and is in the general direction of the plane of the propagation.

Typical examples for oceanic Rayleigh waves in this study are the Fox Islands shock of March 20, 1958 and the Andreanof Islands shock of February 22, 1958, shown in figures 4 and 5. The normal branch of Rayleigh waves for both shocks begins at the period approximately 40 sec and decreases to less than 20 seconds. The

TABLE I

EARTHQUAKES OF THE CIRCUM-PACIFIC BELT RECORDED AT SUVA, FIJI (LAT 18° 08′ 56″ S, LONG 178° 27′ 26″ E)

| No. | Date | G.M.T. Origin Time | Lat (Deg) | Long (Deg) | Geographic Location | Depth in Km | Mag | Δ in Km |
|---|---|---|---|---|---|---|---|---|
| P 1 | 3/23/60 | 00:23:22 | 39½ N | 143 E | Japan | ..... | 6½–7 | 7363 |
| P 2 | 10/27/59 | 06:52:50 | 45½ N | 151 E | Kuriles | ~100 | 6½ | 7569 |
| P 3 | 12/27/59 | 15:52:55 | 56 N | 162½ E | Kamchatka | ..... | ... | 8086 |
| P 4 | 3/20/58 | 01:38:04 | 51 N | 173 W | Fox Islands | ..... | ... | 7673 |
| P 5 | 2/22/58 | 10:50:23 | 50½ N | 175 W | Andreanof Islands | ..... | 6¾ | 7627 |
| P 6 | 8/9/60 | 07:39:22.6 | 40 N | 126.6 W | Northern California | ~125 | 6 | 8588 |
| P 7 | 4/12/58 | 11:46:58 | 26½ N | 111 W | Gulf of California | ..... | 6–6½ | 9074 |
| P 8 | 5/24/59 | 19:17:40 | 17½ N | 97 W | Oaxaca, Mexico | ~100 | 6¾–7 | 10050 |
| P 9 | 1/15/58 | 19:14:29 | 16½ S | 71½ W | Southern Peru | ~100 | 6¾–7 | 11460 |
| P 10 | 4/15/60 | 03:25:36 | 27 S | 113 W | Easter Islands | ..... | 6½ | 7035 |
| P 11 | 6/20/60 | 02:01:08 | 38 S | 73½ W | Chile | ..... | 7–7½ | 10283 |
| P 12 | 2/8/60 | 12:45:34 | 58 S | 67 W | Drake Passage | ..... | ... | 9674 |
| P 13 | 3/22/60 | 02:31:17 | 61½ S | 154 E | 400 miles NW of Balleny Islands | ..... | ... | 5189 |
| P 14 | 10/7/60 | 15:18:30.8 | 7.4 S | 130.7 E | Banda Sea | ~45 | 6½–7 | 5297 |

TABLE II

EARTHQUAKES RECORDED AT HONOLULU (LAT 21° 18′ 13″ N, LONG 158° 05′ 44″ W) AND MT. TSUKUBA (LAT 36° 12′ 39″ N, LONG 140° 06′ 36″ E)

| Date | G.M.T. Origin Time | Lat (Deg) | Long (Deg) | Geographic Location | Depth in Km | Mag | Mt. Tsukuba-Honolulu in Km |
|---|---|---|---|---|---|---|---|
| 11/9/60 | 10:43:43.1 | 30.7 N | 103.4 E | China | — | 6¼–6½ | 6136 |
| 9/4/58 | 21:51:08 | 33½ S | 69½ W | Chile | — | 6¾–7 | 6146 |

inverse branch with a period range 40 to 110 sec for the Fox Islands shock and 40 to 120 sec for the Andreanof Islands shock may be seen with the normal branch superimposed. The inverse branch is traced by connecting the mid points of each swing of the shorter period branch (see fig. 4). The Airy phase corresponding to the maximum value of the group velocity for Rayleigh wave propagation across ocean basins occurs at a period of approximately 42 sec, where the two branches coincide. The amplitude for the periods greater than 120 sec generally decreases very rapidly due to low instrumental response. The general characteristics of the Rayleigh wave train for other shocks analyzed are very similar to these two Aleutian Islands shocks, except for some minor variations (figs. 6 and 7). The normal branch of the Rayleigh wave train for the Southern Peru shock of January 15, 1958 (fig. 8) is barely recognizable; and yet the inverse branch is exceptionally well developed. The period range extends from 35 to 120 seconds. The amplitude of the Rayleigh wave trains for the Kuriles shock of October 27, 1959 is unusually high. The first train identified

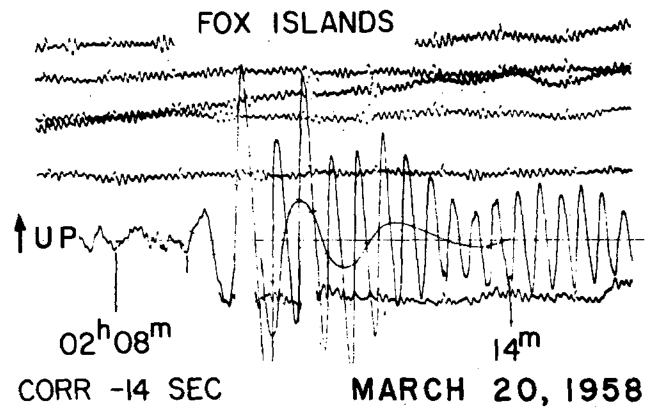


FIG. 4. Rayleigh wave portion of the vertical-component seismogram for the Fox Islands shock of March 20, 1958.

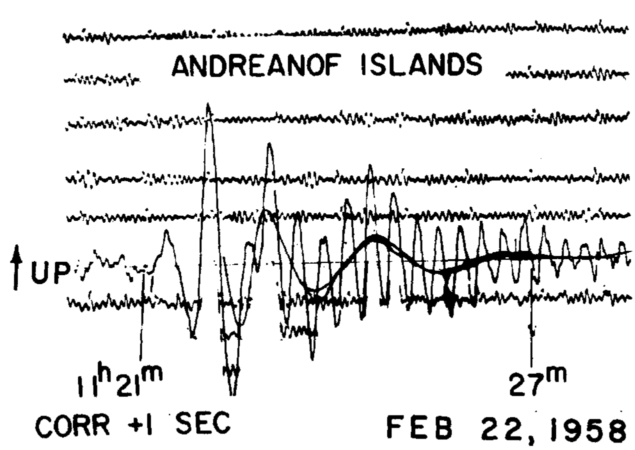


FIG. 5. Rayleigh wave portion of the vertical-component seismogram for the Andreanof Islands shock of February 22, 1958.

as the fundamental Rayleigh mode arrives at approximately $07^h$ $24^m$ (see fig. 9). Even though the inverse branch is much lower in amplitude, it may still be easily traced.

## Dispersion Data

Rayleigh wave dispersion curves including both group and phase velocities are shown in figures 10 to 23. The solid-line curve is the observed phase velocity for mantle Rayleigh waves at periods greater than 80 seconds (Brune, 1961). For periods greater than 120 sec, it is assumed in this paper that this phase velocity curve is nearly the same as for a pure oceanic path. Three discrete phase velocity curves are determined by the single-station method with three different assumptions of initial phase: $\phi_0 - \pi/2$, $\phi_0 - \pi$, $\phi_0 - 3\pi/2$. (See figs. 10 to 23.) The cen-

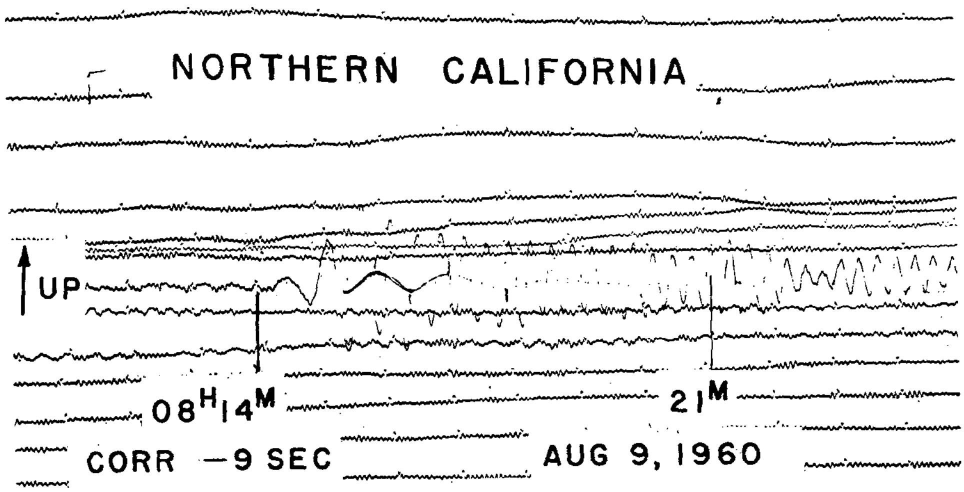


Fig. 6. Rayleigh wave portion of the vertical-component seismogram for the Northern California shock of August 9, 1960.

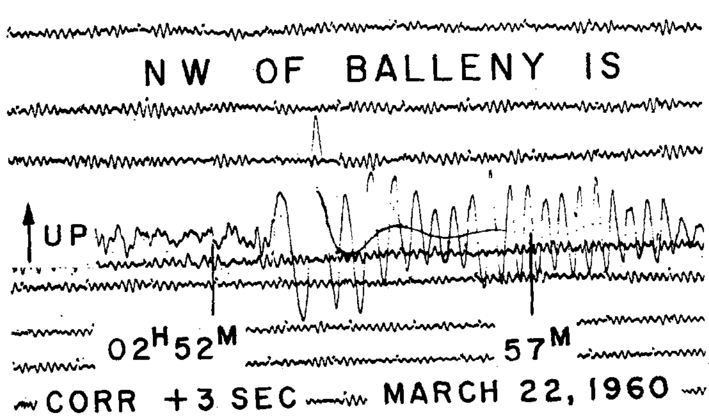


Fig. 7. Rayleigh wave portion of the vertical-component seismogram for the Balleny Islands shock of March 22, 1960.

ter phase velocity curve, which corresponds most favorably to Brune's curve at periods generally greater than 90 sec, is selected to be the correct phase velocity curve for the particular path from the epicenter to the Suva station.

Most of the phase velocity curves show a slight maximum and a minimum. The group velocities are equal to the phase velocities at the phase velocity maximum and minimum. The group velocity approaches a maximum, greater than the phase velocity, corresponding to the Airy phase at the period between approximately 28 sec and 41 sec. Table III summarizes the Airy phase periods for each shock. In

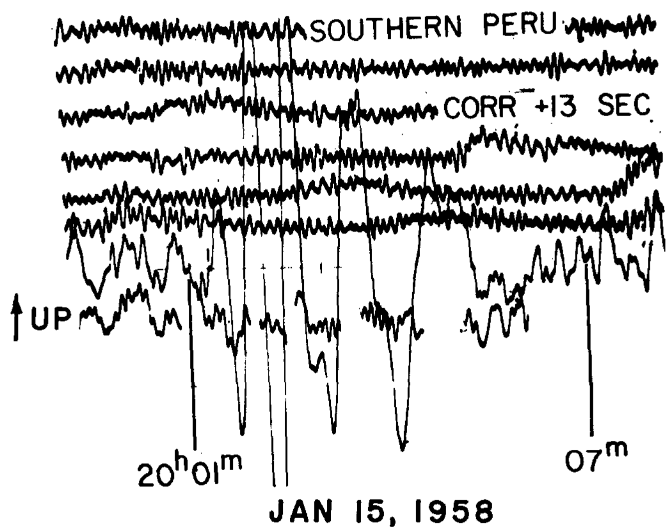


FIG. 8. Rayleigh wave portion of the vertical-component seismogram for the Southern Peru shock of January 15, 1958.

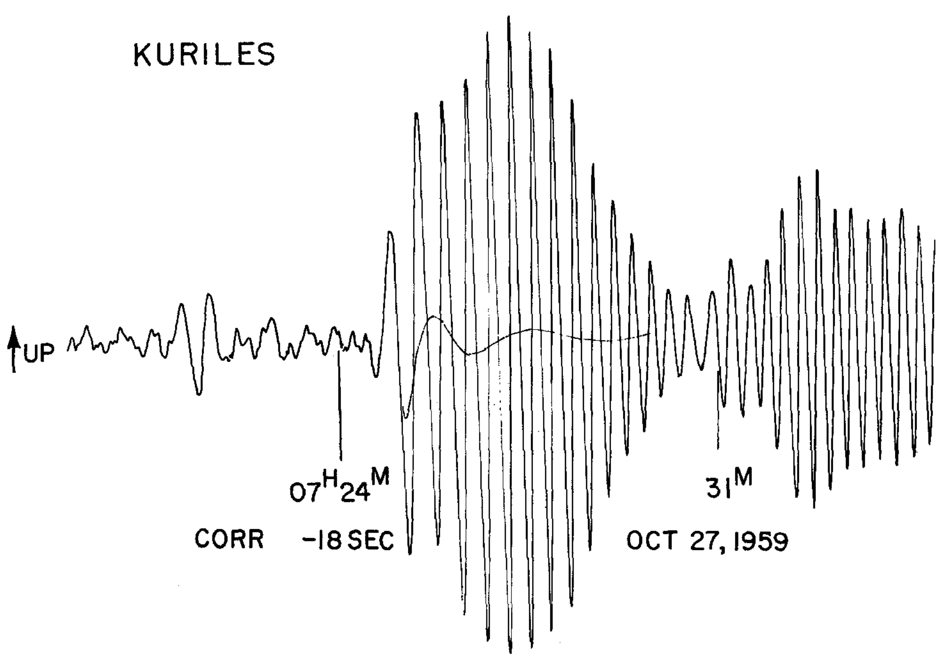


FIG. 9. Rayleigh wave portion of the vertical-component seismogram for the Kuriles shock of October 27, 1959.

general, the periods for the Airy phase and the phase velocity maximum and minimum are greater for the northern Pacific shocks; these have a relatively greater mean water depth along their paths. The Banda Sea shock has the period of the Airy phase of 52 sec approaching a continental type.

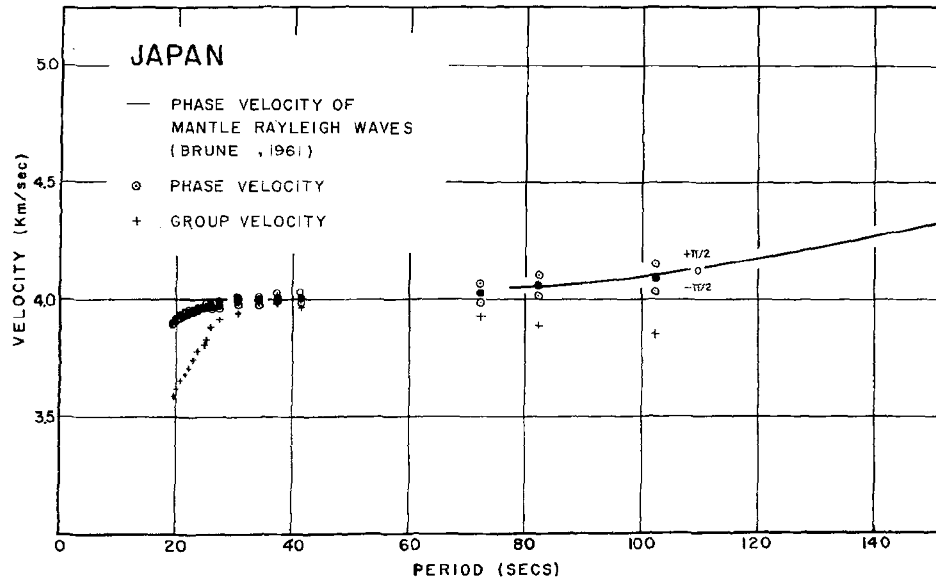


FIG. 10. Rayleigh wave dispersion curves for the Japan shock of March 23, 1960.

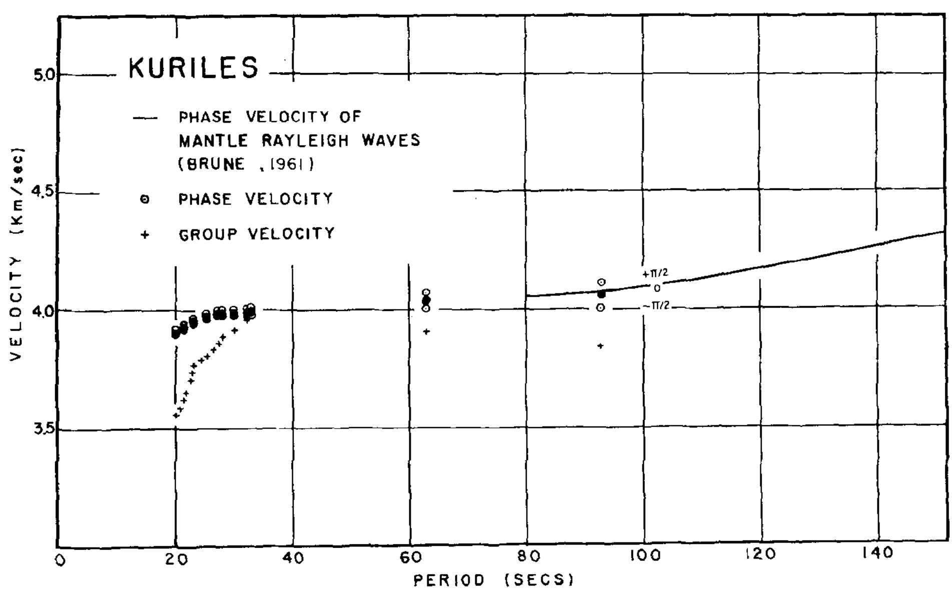


FIG. 11. Rayleigh wave dispersion curves for the Kuriles shock of October 27, 1959.

The paths from the epicenter to Suva for the Fox Islands shock and the Andreanof Islands shock are almost common and of all the paths studied, have the deepest and most uniform water depth in the Pacific. Group and phase velocities of these two shocks are in good agreement, as would be expected. (See figs. 13 and 14.) The center phase velocity curves (the only curves which will be referred to in the follow-

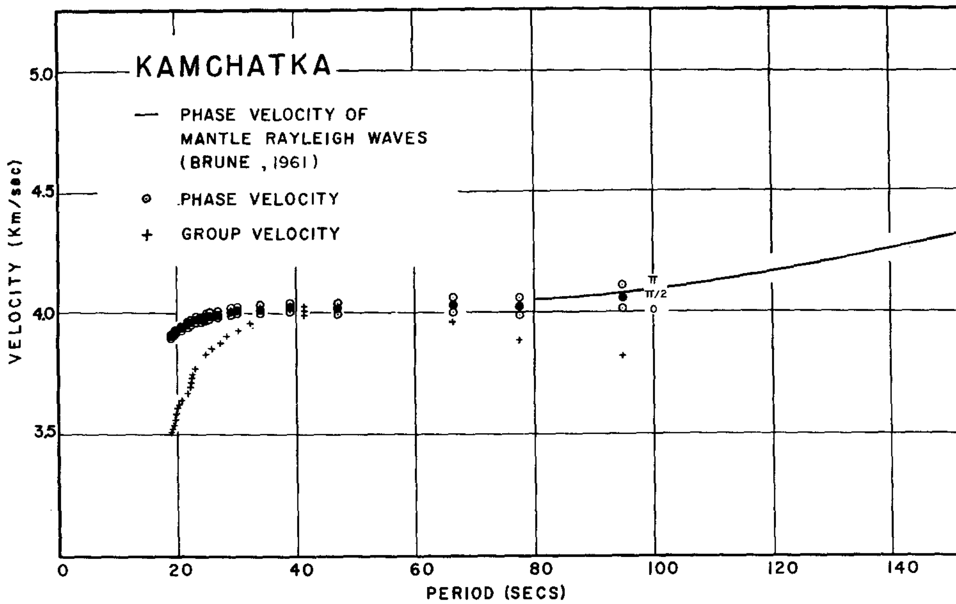


FIG. 12. Rayleigh wave dispersion curves for the Kamchatka shock of December 27, 1959.

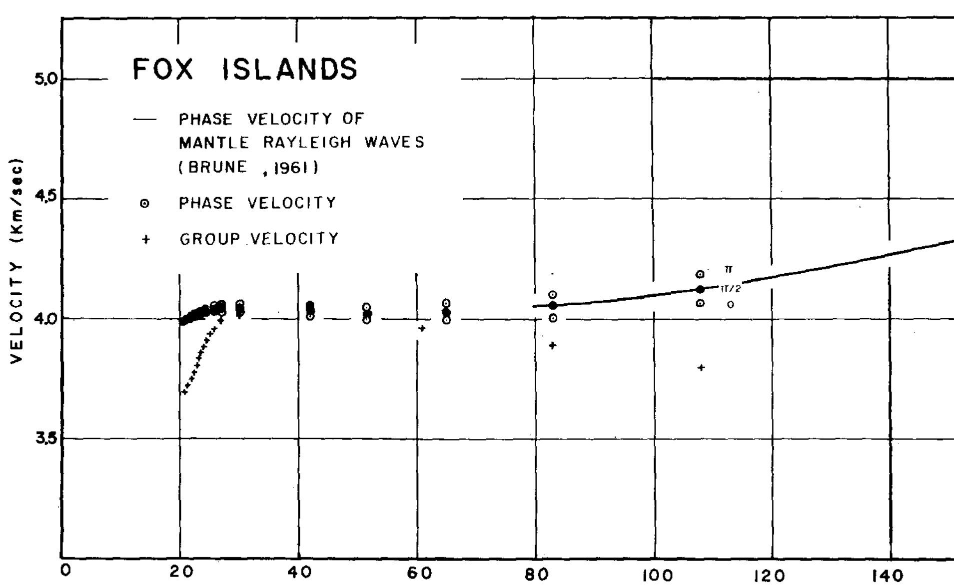


FIG. 13. Rayleigh wave dispersion curves for the Fox Islands shock of March 20, 1958.

ing) for the two shocks give a phase velocity of 4.10 km/sec at 100 sec and a phase velocity of 4.18 km/sec at 120 seconds. The phase velocity reaches 4.25 km/sec at 140 sec and decreases gradually to 4.01 km/sec at 50 sec, where the minimum occurs; increases slightly to 4.04 km/sec, the maximum, at 35 sec and decreases to approximately 3.96 km/sec at 20 seconds. The separation of the adjacent sets of the

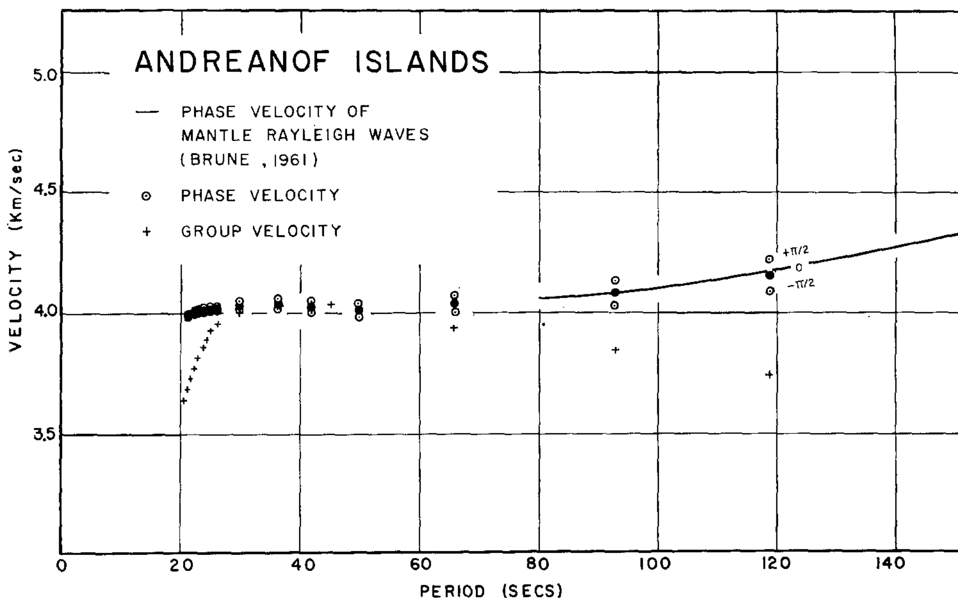


FIG. 14. Rayleigh wave dispersion curves for the Andreanof Islands shock of February 22, 1958.

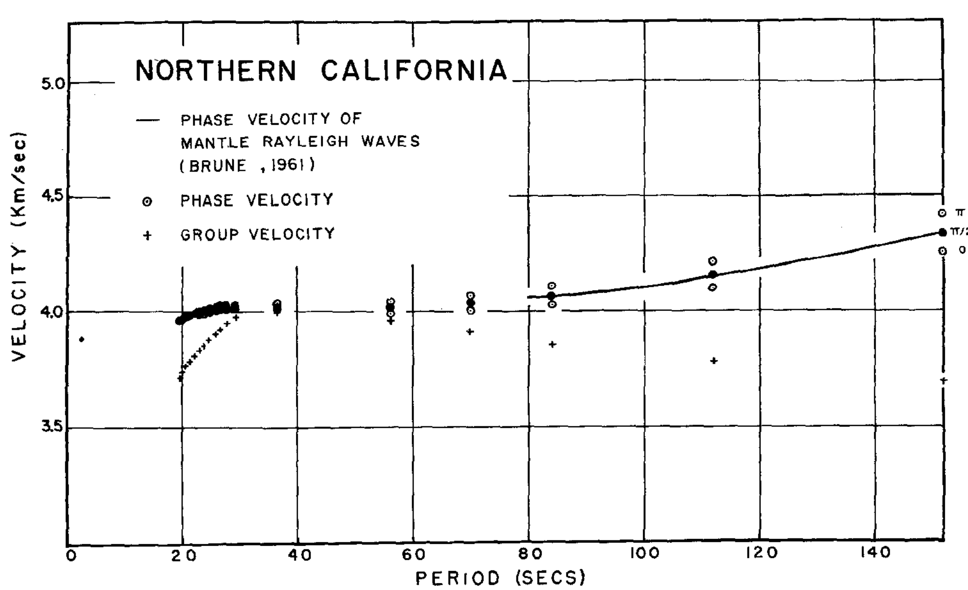


FIG. 15. Rayleigh wave dispersion curves for the Northern California shock of August 9, 1960.

phase velocity curves is greater than 0.05 km/sec at the period 80 sec; experimental errors introduced by the instruments and the epicentral locations are estimated to be less than 0.03 km/sec at the same period with an epicentral distance of 7,000 kilometers. The group velocity maximum in the observed period range is 4.06 km/sec at 42 seconds. For the inverse branch, the group velocity decreases to 3.80 km/

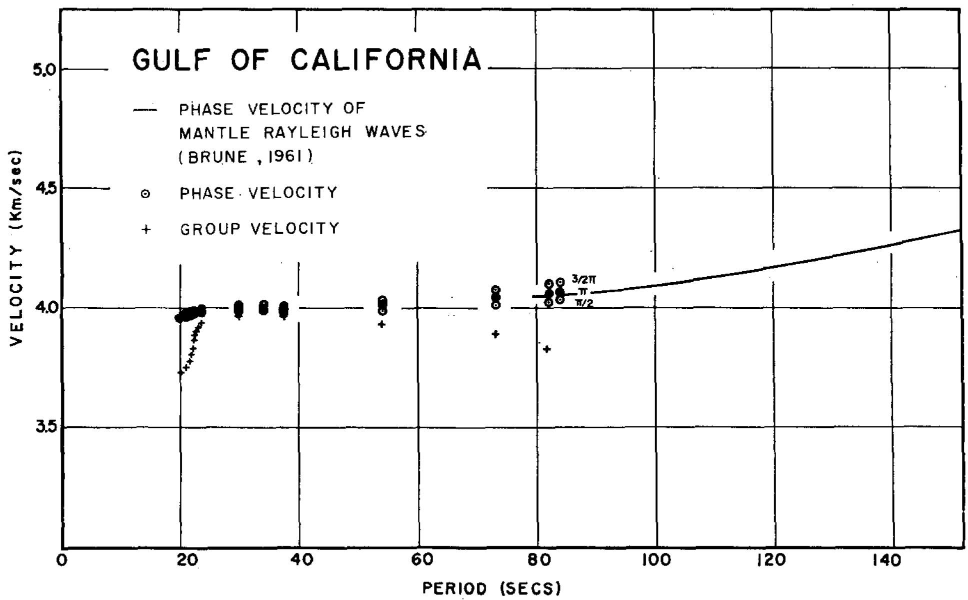


FIG. 16. Rayleigh wave dispersion curves for the Gulf of California shock of April 12, 1958

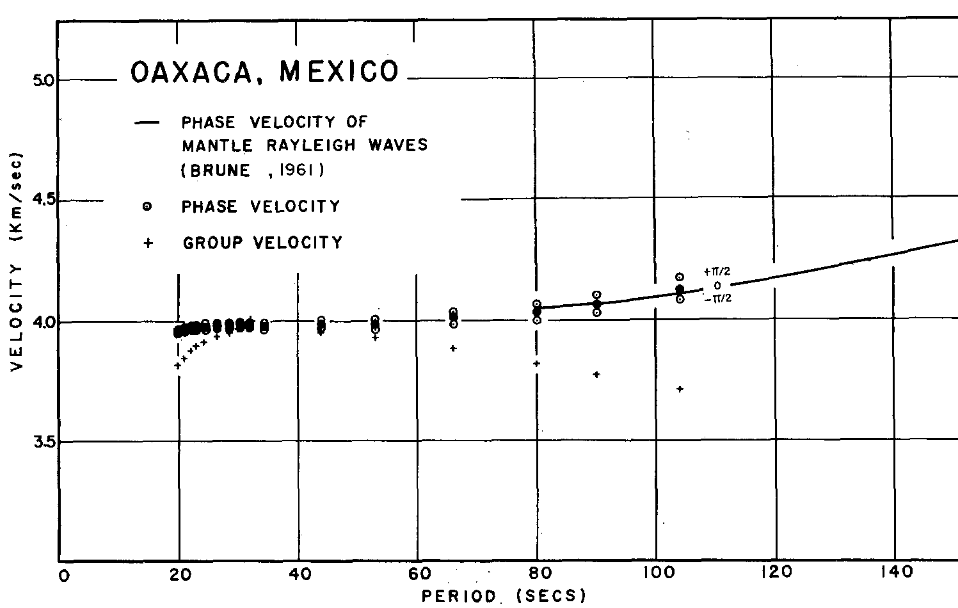


FIG. 17. Rayleigh wave dispersion curves for the Oaxaca shock of May 29, 1959.

sec at 120 seconds. For the normal branch, the group velocity decreases to 3.65 km/sec at a period of 20 sec for the Fox Islands shock and 3.53 km/sec at the same period for the Andreanof Islands shock.

Phase and group velocities for the Northwestern Pacific shocks, the Japan shock of March 23, 1960, the Kuriles shock of October 27, 1959, and the Kamchatka shock

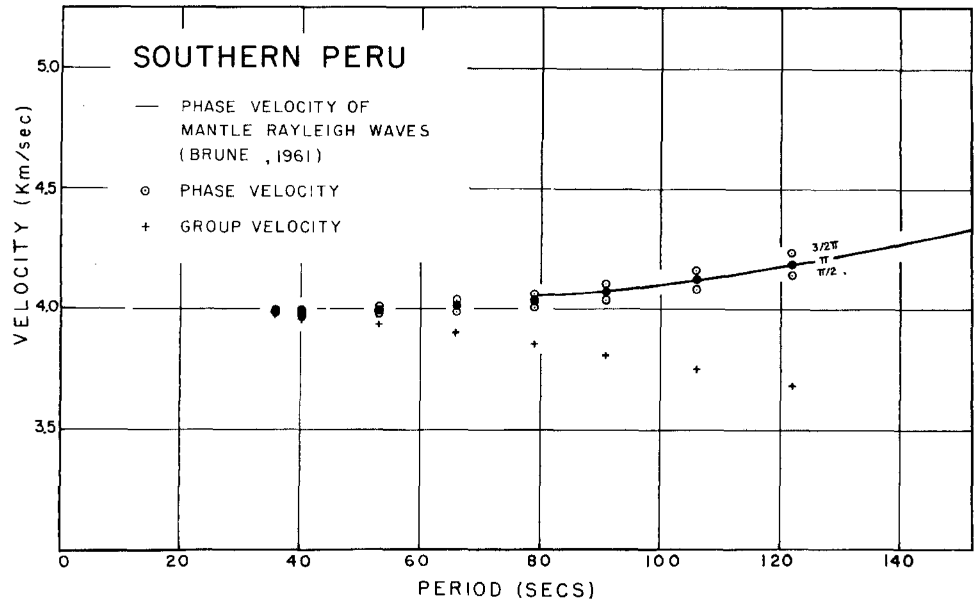


FIG. 18. Rayleigh wave dispersion curves for the Southern Peru sho ck of January 15, 195

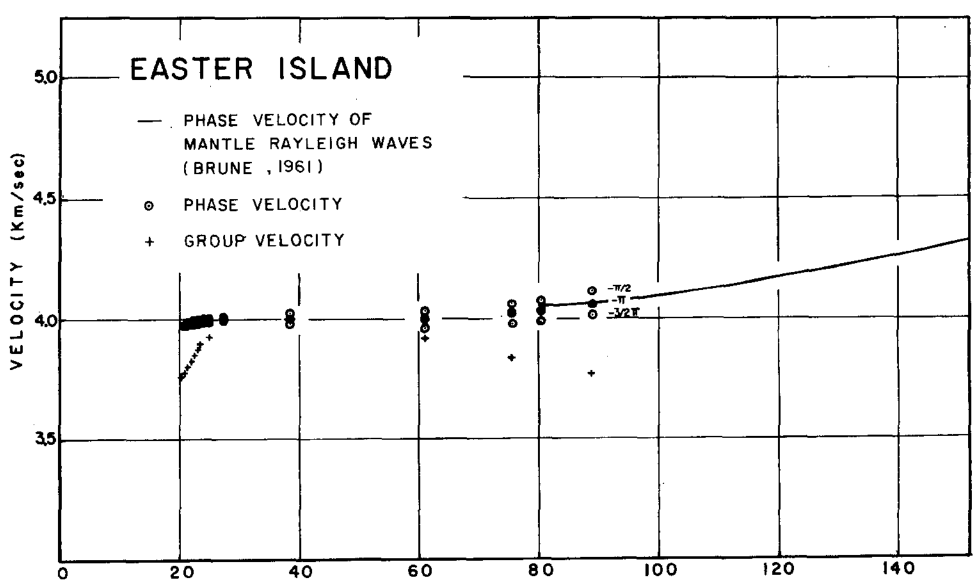


FIG. 19. Rayleigh wave dispersion curves for the Easter Islands shock of April 15, 1960.

of December 27, 1959, are in good agreement with each other at periods greater than 40 seconds (figs. 10, 11, 12). Both phase and group velocities for the Japan shock and the Kuriles shock are in agreement even at periods less than 40 sec, but are comparatively lower than those of the Kamchatka shock, which are in better agreement with those of the Aleutian Islands shocks.

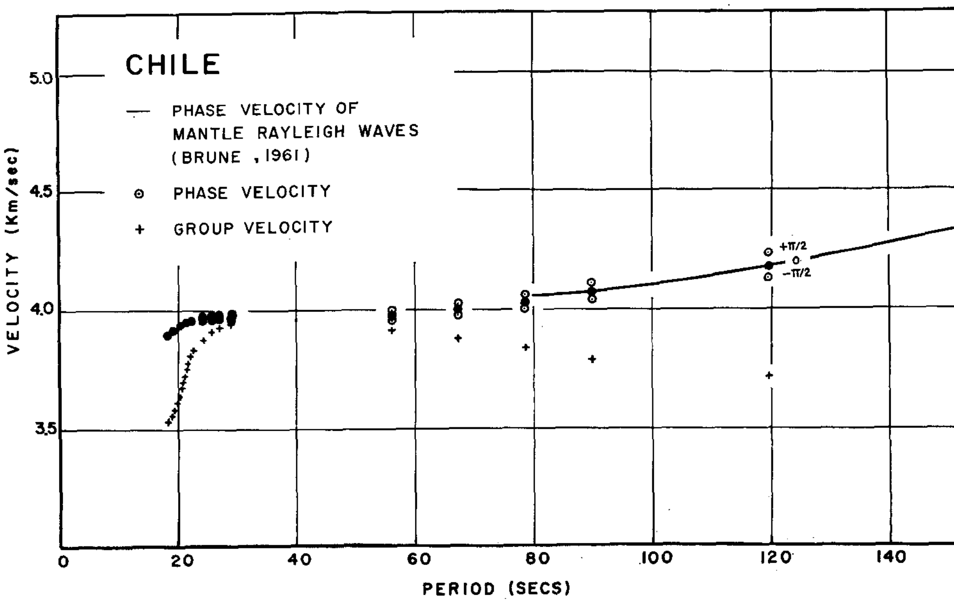


FIG. 20. Rayleigh wave dispersion curves for the Chile shock of June 20, 1960.

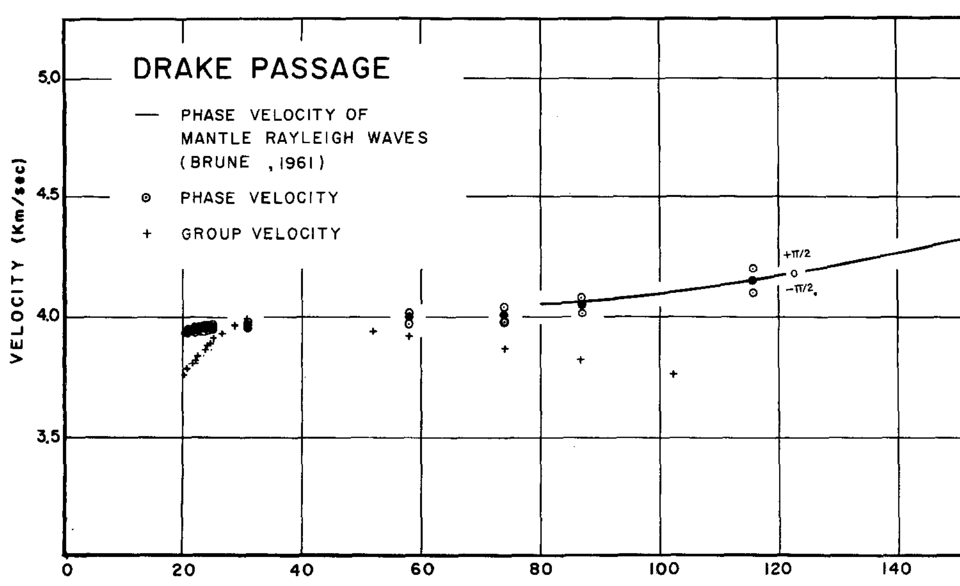


FIG. 21. Rayleigh wave dispersion curves for the Drake Passage shock of February 8, 1960.

For the Northern California shock (fig. 15), the phase velocity curve is slightly lower than that of the Fox (or Andreanof) Islands shock at periods less than 100 seconds. The phase velocity approaches 4.02 km/sec at about 60 seconds. For the Gulf of California shock (fig. 16), phase velocities are lower than those of the Northern California shock at periods less than 40 seconds.

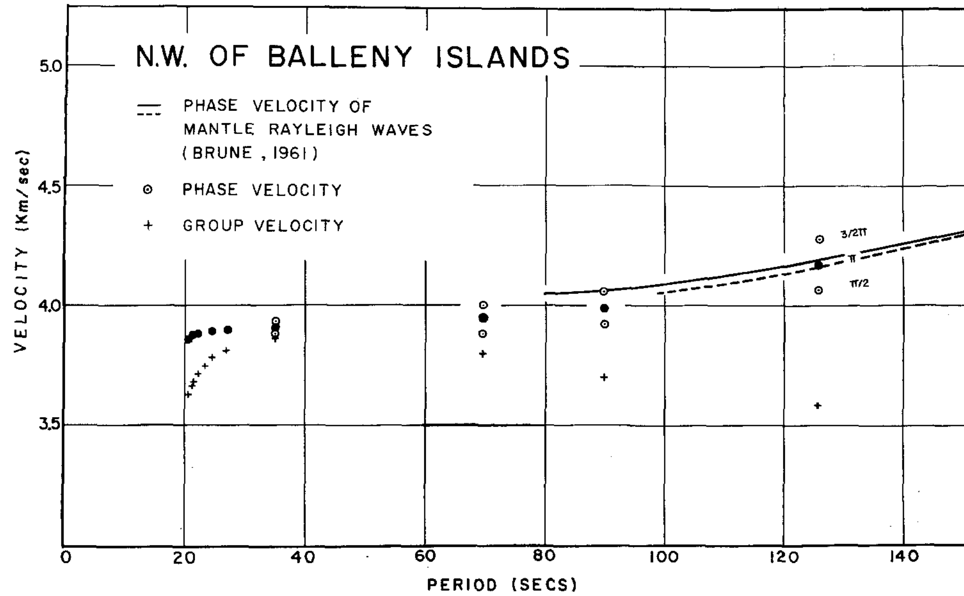


FIG. 22. Rayleigh wave dispersion curves for the Balleny Islands shock of March 22, 1960.

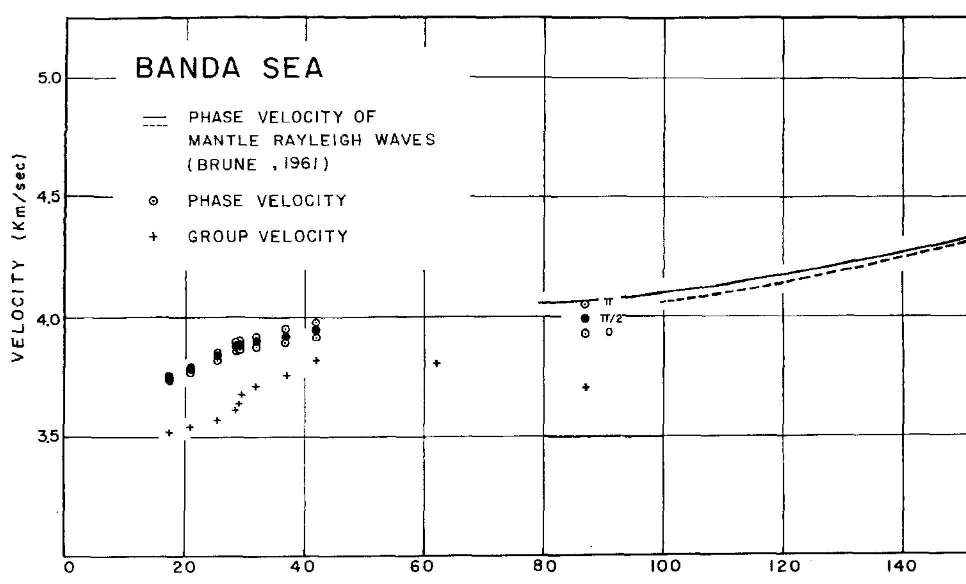


FIG. 23. Rayleigh wave dispersion curves for the Banda Sea shock of October 7, 1960.

The phase velocity curve of the Oaxaca shock (fig. 17), with a path intermediate between the northern and southern Pacific is also lower than that of the northern shocks just discussed (i.e. less than 4.0 km/sec at the periods less than 60 seconds). The group velocity is lower than that of northern shocks for the entire observed period range.

Paths from Southern Peru and from Easter Islands to Suva have a common path from the Easter Islands to Suva. Periods of the inverse branch for the Easter Islands shock do not extend beyond 90 sec, and hence it is not possible to compare the two shocks for periods greater than 90 seconds. However, the observed phase velocities for these two shocks are in good agreement at shorter periods (figs. 18 and 19). The phase velocity reaches approximately 4.0 km/sec at 40 sec for both shocks. The group velocities for the two shocks are also low in comparison with those of other northern shocks. The phase and group velocities for the Chile shock of June 20, 1960 and the Drake Passage shock of February 8, 1960 (figs. 20 and 21), the most southern paths studied, are essentially identical, indicating a very similar structure. The

TABLE III

PERIODS OF THE AIRY PHASE AND INITIAL PHASES FOR P 1 TO P 14

| Shock | | Airy Phase in sec | Initial Phase $\phi_0$ |
|---|---|---|---|
| P 1 | Japan | 37.0 | $\pi$ |
| P 2 | Kuriles | 34.0 | $\pi$ |
| P 3 | Kamchatka | 40.8 | $3\pi/2$ |
| P 4 | Fox Islands | 42.0 | $3\pi/2$ |
| P 5 | Andreanof Islands | 42.0 | $\pi$ |
| P 6 | Northern California | 36.4 | $3\pi/2$ |
| P 7 | Gulf of California | 34.0 | 0 |
| P 8 | Oaxaca, Mexico | 28.2 | $\pi$ |
| P 9 | Southern Peru | 32.5 | 0 |
| P 10 | Easter Islands | 38.4 | 0 |
| P 11 | Chile | 29.4 | $\pi$ |
| P 12 | Drake Passage | 31.0 | $\pi$ |
| P 13 | Balleny | 35.0 | 0 |
| P 14 | Banda Sea | 52.0 | $3\pi/2$ |

phase velocity for these two shocks is about 0.03 km/sec lower than that for the northern Pacific shocks at periods less than 100 seconds.

Comparison of the dispersion data discussed above shows that as the path becomes more southerly, phase and group velocities in general become lower at periods less than 75 seconds. Experimental error due to epicenter locations and instrumental time delay could not account for this variation. The effect of the Easter Island Rise, which possibly involves a thickening of the intermediate crustal layer, and/or an additional layer of relatively lower velocity may cause lower observed phase and group velocities in this period range (Oliver *et al*, 1955).

Dispersion curves for the Melanesian-New Zealand paths are markedly different from all those discussed above (figs. 22 and 23). The phase velocities of the Balleny Islands shock and the Banda Sea shock for the periods less than 120 sec decrease very rapidly. At the periods less than 90 sec, the phase velocity falls below 4.0 km/sec, and continuously decreases to 3.77 km/sec at a period of 20 sec for the Banda Sea shock and 3.86 km/sec at a period of 20 sec for the Balleny shock. The group velocities for these two shocks at periods greater than 60 sec are approximately dentical but are consistently 5 per cent lower than those of the other Pacific shocks.

The group velocity is 3.80 km/sec at a period of 60 sec and decreases to 3.60 km/sec at 120 sec for the inverse branch. At shorter periods, less than 40 sec, the group velocities differ by as much as 0.2 km/sec at a period of 28 seconds. Though the group velocity of the Balleny Islands shock is lower in comparison with other shocks, the general trend of the curve still resembles that of oceanic Rayleigh waves. The group velocity curve of the Banda Sea shock resembles neither that of continental Rayleigh waves nor that of oceanic Rayleigh waves. This difference in appearance

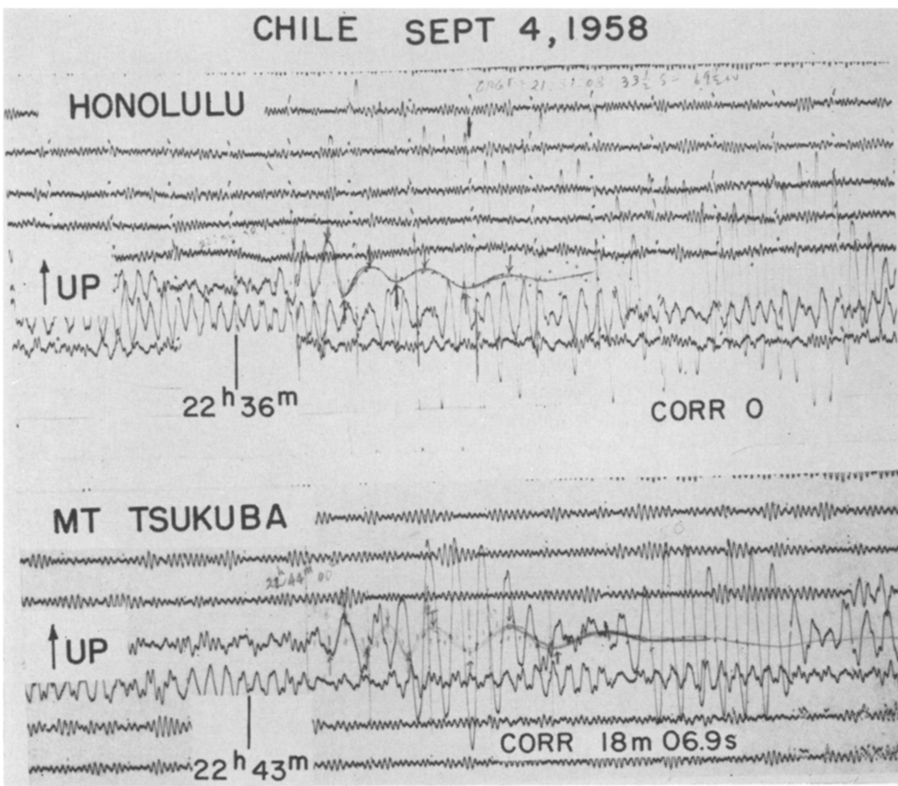


FIG. 24. Rayleigh wave portion of the vertical-component seismogram for the Chile shock of September 4, 1958.

of the group velocity curves for these shocks clearly indicates that the regional structure in the Melanesian-New Zealand region differs from that of the Pacific basin proper. The variations of the phase and group velocities at the short periods for the Balleny Islands shock and the Banda Sea shock further suggest local differences in the crustal structure in the Melanesian-New Zealand region. A more detailed study of the Rayleigh wave dispersion in this region will be made in a later paper.

## Phase Velocity Determined by Two-Station Correlation

The surface wave portion of the seismograms for the Chile shock of September 4, 1958 and the China shock of November 9, 1960 and the phase velocity curves between Mt. Tsukuba and Honolulu are shown in figs. 24, 25 and 26 respectively. The azimuths from the epicenter of the Chile shock of September 4, 1958 ($33\frac{1}{2}°$ S, $69\frac{1}{2}°$ W) to the respective stations Mt. Tsukuba and Honolulu are approximately N 72° W and N 65° W with an azimuthal difference of 7 degrees.

Sufficiently large epicentral distances between the epicenter to the respective two stations and large magnitude of the shock allow both the normal and inverse branches to become well developed. The phase velocity curve determined by two-station correlation by taking into account the instrumental phase shifts at two sta-

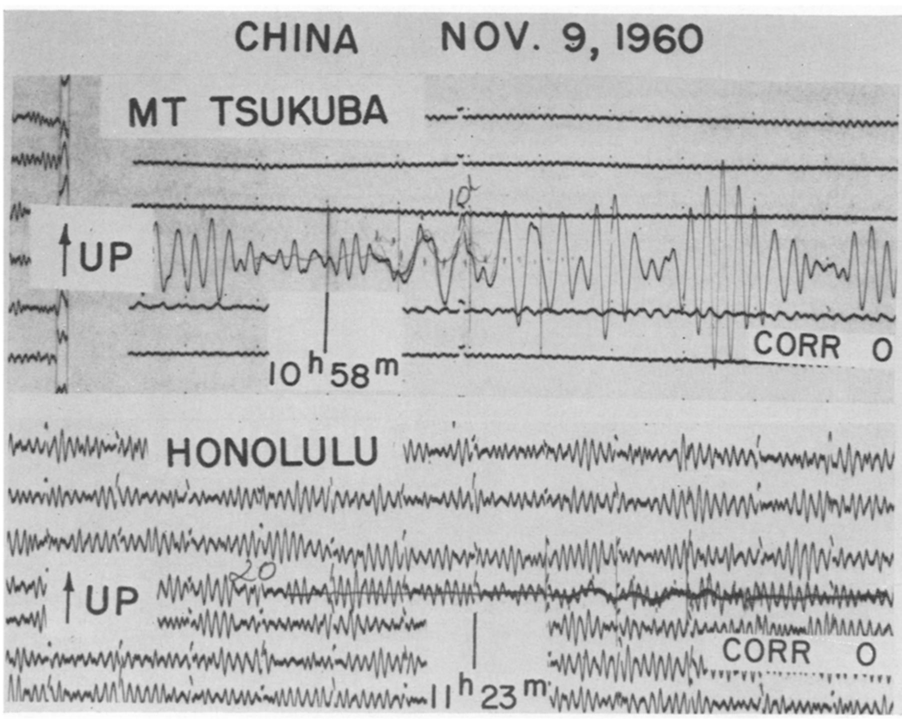


FIG. 25. Rayleigh wave portion of the vertical-component seismogram for the China shock of November 9, 1960.

tions covers a period range 20 to 106 seconds. The phase velocity thus determined is independent of the initial phase. The correct phase velocity curve is shown in fig. 26 and is higher than Brune's curve by 0.02 km/sec at 106 seconds. With an azimuthal difference of 7 degrees, the path of the Rayleigh wave propagation would pass the point approximately 500 km south of Honolulu. Possible errors introduced by assuming two paths lie on a common great circle may arise due to the variations in the underlying structure and variation of initial phase with azimuth. Despite these sources of error this curve agrees fairly well with the curve obtained from the single station method. It is felt by these authors that the curves from the single station method are more reliable.

The epicenter of the China shock of November 9, 1960, the Mt. Tsukuba station and the Honolulu station lie nearly on a great-circle path with an azimuth N 59° E and provide accurate measurements of the phase velocities between these two stations. Since the instruments used at both stations on this date are approximately identical with only minor deviations in constants, no phase shift corrections are applied here in the determination of phase velocity. Only the normal branch of the Rayleigh wave train is identifiable with a period range of 20 to 40 seconds. (See figure 26.) The phase velocity thus determined is 4.05 km/sec at the period 48 sec and in good agreement with the phase velocity of this period obtained from the Chile shock of September 4, 1958. The phase velocity steadily decreases from

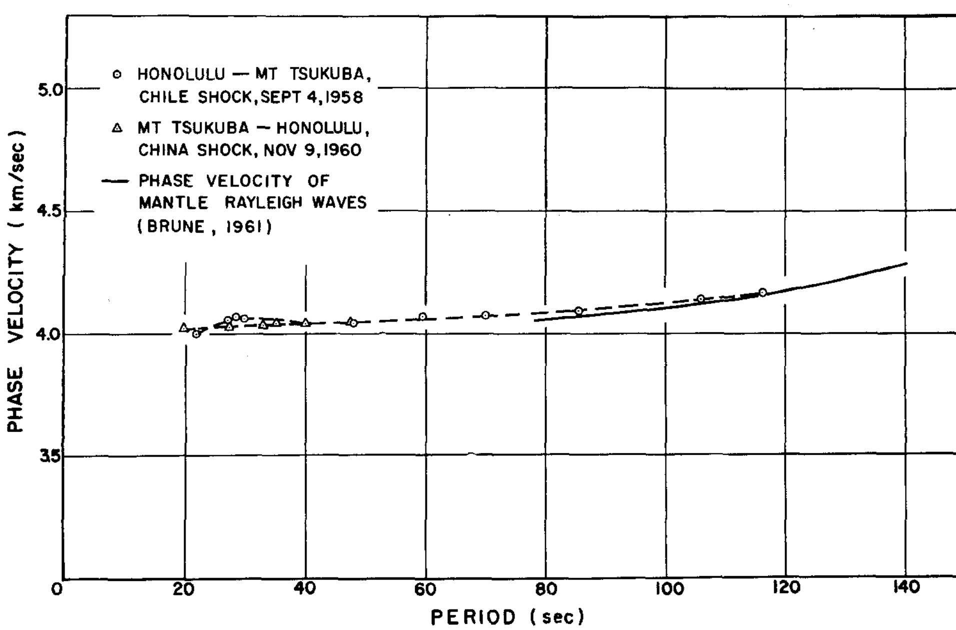


FIG. 26. Phase velocity curves between Mt. Tsukuba and Honolulu.

4.05 km/sec at 48 sec to 4.02 km/sec at 20 sec. and is higher than the phase velocity determined by single station method in this period range.

## Compiled Phase Velocities

Phase velocities for all shocks previously determined are compiled in figure 27. Except *P* 13 and *P* 14, in which the paths of the Rayleigh wave propagations are crossing the Melanesian-New Zealand region, all follow a consistent trend. Phase velocity is nearly constant in the period range from 30 to 75 sec with a velocity slightly greater than 4.0 km/sec; most of the phase velocity curves indicate a slight maximum and minimum at approximately 30 and 50 sec respectively. Phase velocity for longer periods increases to 4.18 km/sec at 120 seconds. Phase velocities are comparatively lower for *P* 13 and *P* 14 in the period range less than 100 sec and fall below 4.0 km/sec at nearly 90 seconds.

The dashed-line curve of 8099S in figs. 27 and 28 is Dorman's 8099 curve (Dorman *et al*, 1960) for oceanic Rayleigh waves approximately corrected for the effect of the earth's curvature according to the following relationship:

$$C \simeq C_h (1 + 0.00016\ T) \tag{3}$$

where $C$ is the corrected phase velocity, $C_h$ is the uncorrected phase velocity and $T$, the period (Bolt and Dorman, in press). Equation (3) is the curvature effect on the phase velocity of continental Rayleigh waves in the period range 20 to 300 seconds. It is assumed that this correction is applicable to the oceanic phase velocity curve for Rayleigh waves in this period range. Observed phase velocities in the period range 20 to 50 sec agree with Dorman's curve 8099S (except the phase velocity curves of $P$ 13 and $P$ 14, which fall below the curve of 8099S significantly). Dorman's curve 8099S lies increasingly above the observed phase velocities for periods greater than 50 seconds.

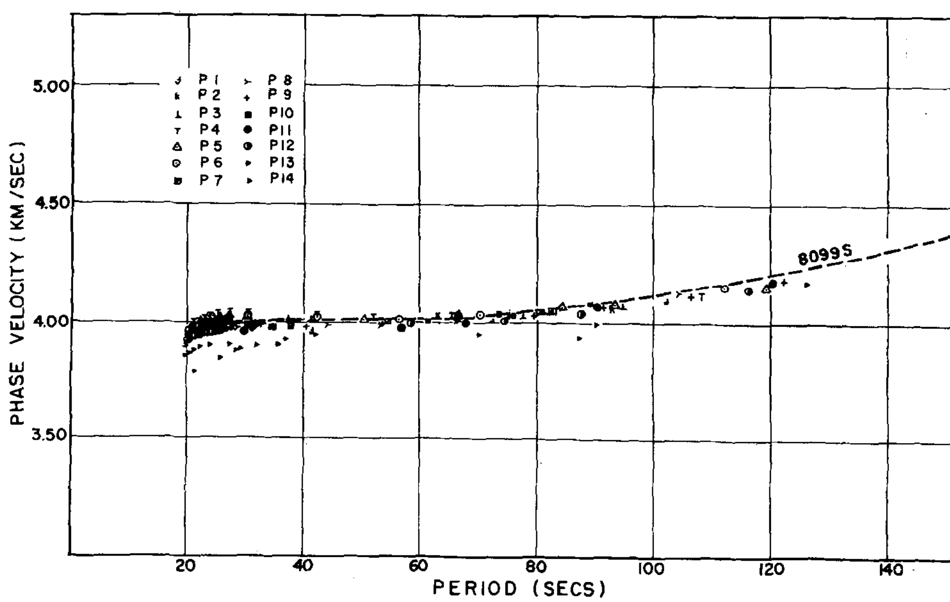


FIG. 27. Compiled phase velocities.

## Compiled Group Velocities

The compiled group velocities of all shocks (no group velocities were determined for $C$ 1 and $C$ 2) are presented in figure 28. Group velocities for Rayleigh waves reach a maximum of 4.0 km/sec at the period approximately 40 sec and decrease to 3.75 km/sec at 120 sec for the inverse branch and 3.50 km/sec at 20 sec for the normal branch. In general, the curve of 8099S gives a good fit to the observed group velocities except the group velocity of $P$ 13 and $P$ 14 are low in the entire period range. The group velocity of $P$ 13 for periods less than 28 sec seems to fall on the curve.

## Discussion

An arbitrary value of initial phase was assumed to make the phase velocity curve match Brune's phase velocity curve at a period of about 120 seconds. It is known that at this period the mantle Rayleigh wave phase velocities are nearly the same for oceanic and continental paths. Measurements of phase velocity over the great circle path between Honolulu and Pietermaritzburg (about 22 per cent continent), the paths between Uppsala, Lwiro, and Pietermaritzburg (purely continental), the complete great circle path including Uppsala and Southeast Alaska (40 per cent continent) and the great circle path between Perth and Ottawa (nearly all in the Indian and Atlantic Oceans) all give almost the same phase velocity value, 4.17 ± 0.01 km/sec at 120 sec period. These measurements are independent of the assumption of initial phase and the effect of the instruments has been eliminated or corrected. The exact difference between oceanic and continental phase velocities at periods of 120 sec is not known, however. If it is found that for Pacific paths the true mantle Rayleigh wave phase velocity is slightly different from that of Brune's

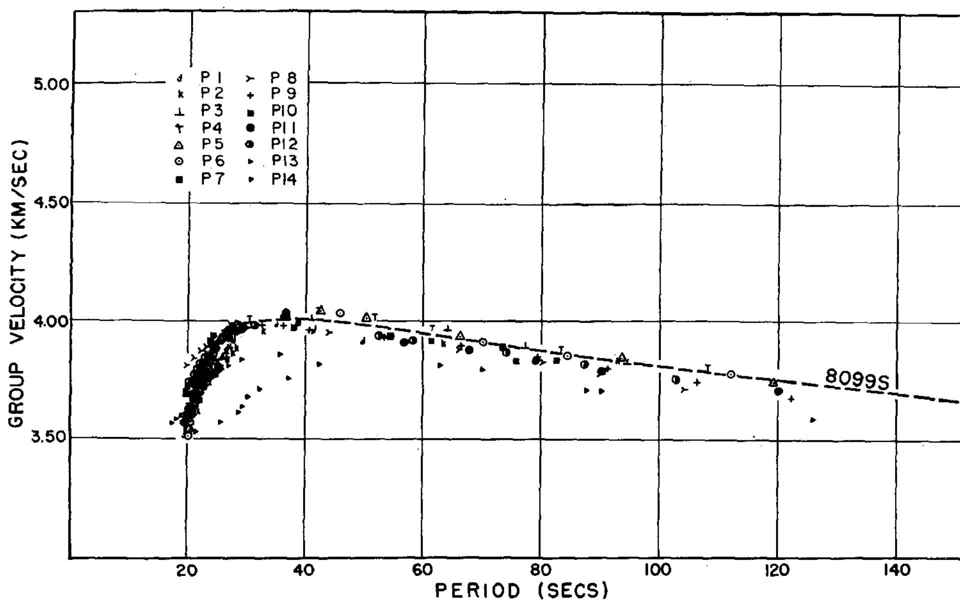


FIG. 28. Compiled group velocities.

curve at periods near 120 sec, then the values for phase velocity presented in this paper will need to be corrected a small amount. Any such correction necessary will be less significant for shorter periods and will be less than other experimental errors for periods less than 60 seconds. When instruments are better calibrated and more closely spaced, it will be possible to make more precise measurements over smaller regions and to establish slight local variations in the phase velocity.

In this analysis, it is assumed that no phase shifts and corresponding group delay times have resulted from the effect of a finite traveling source (Ben Menahem, 1961), which causes the initial phase to vary as a function of period. In terms of interpretations of structure such effects will cause the same error when group velocity is used as when phase velocity is used. This effect is small if the dimensions of the fault are small compared with the wavelengths involved.

A check on the phase velocities presented in the paper might come from examination of initial phases implied if the mechanism of the earthquakes were known. Table III shows the initial phase required for each shock to fit the Brune's curve at long periods. Statistically there are not enough examples to draw conclusions about earthquake mechanism. According to the theory of Lamb (1904) as generalized by Aki (1960b), initial phases of 0 and $\pi$ would be expected from a force couple in the horizontal plane acting as a step function in time. This is a reasonable model for a strike-slip fault. However, the exact nature of earthquake mechanism is not known and any definite conclusions must await more developed theory and experiment.

The seismograms of the circum-Pacific earthquakes recorded at Pasadena and used by Aki (1960b) in the study of earthquake mechanisms will give phase velocity results in agreement with ours if the initial phase is chosen in such a way that the phase velocity agrees with Brune's curve at a period of about 120 seconds. This is guaranteed since his seismograms give group velocity data in good agreement with the group velocity curve for Dorman case 8099, which agrees with the group velocities observed in this paper. Curvature has almost no effect on group velocity (Bolt and Dorman, in press; Pekeris, 1961). However, curvature raises the phase velocity by an amount given approximately by eq. (3) and if the phase velocity curve for Dorman case 8099 corrected for curvature is assumed to be more appropriate for the Pacific, the initial phase results of Aki (1960b) must be corrected by an amount varying directly as the distance. In the period range 40 to 140 sec this amounts to a correction of about $-.75\pi$ in the initial phase per 10,000 km. If a mean phase velocity curve based on our data is assumed, the correction necessary to Aki's results would be about $-.30\pi$ per 10,000 km. Aki found that most of the earthquakes he studied had initial phases of $-\pi/8$ and $7\pi/8$ (in terms of our definition of initial phase). Statistically, then most of the shocks analyzed by Aki would give initial phases of about $\pi/2$ when our revised phase velocity is used.

Two theoretical models have been proposed to explain Pacific Ocean Rayleigh wave group velocity dispersion. These are Dorman's case 8099 (Dorman, Ewing, and Oliver, 1960) and case 6EGHP1 (Aki and Press, in press). Dorman's case 8099 is a modification of the Lehmann's model for continents made by decreasing the depth to the top of the low velocity channel. Case 6EGHP1 is a modification of the Gutenberg's model for continents made by decreasing the shear velocity in the lowest velocity layer of the low velocity channel (4.30 km/sec under the ocean against 4.38 km/sec under the continent). Both case 8099 and 6EGHP1 have the Bullen A density distribution. Case 6EGHP1 has nearly the same structure as 8099 proposed earlier by Dorman, Ewing and Oliver, and as would be expected gives nearly the same dispersion curve. The phase velocities for 6EGHP1 are slightly higher than those for 8099 for periods less than 100 sec (0.01–0.03 km/sec). When the dispersion curves for these two models are corrected for the earth curvature, we obtain phase velocities that lie above the measured velocities in the period range 50 to greater than 200 sec (0.03 km/sec at 100 sec), suggesting a slight decrease in the density of the mantle at depths shallower than 700 km, or a decrease in the shear velocity in the upper mantle.

The consistency of the data of $P$ 1 through $P$ 12, $C$ 1 and $C$ 2 indicates the uniformity not only of the crustal structure but also of the upper mantle in the entire Pacific basin proper. In studying group velocities of Rayleigh waves for periods less than 40 sec in the Pacific, Oliver (Oliver *et al*, 1955) also found there are no major variations in the crustal structure throughout the Pacific basin proper. A similar conclusion was drawn by Santô (1960c).

The data of $P$ 13 and $P$ 14 are very different from those of the Pacific basin proper and suggests crustal thickening or low velocity variations in layers in the Melanesian-New Zealand region of the southwestern Pacific. The boundary of the Pacific which separates the Pacific basin proper from partially submerged continental area in the southwest is commonly drawn on the basis of the petrographic distinction. This boundary is often referred to as the andesite line or the Marshall line, where volcanic islands between continents and the andesite line frequently contain andesite rocks and where those on the ocean side of the line are of olivine basaltic rocks. (See fig. 3.) The dispersion data further support the possibility of the andesite line serving as the appropriate boundary of the Pacific basin proper and the tectonic complex of the southwestern Pacific.

## Acknowledgments

The authors are indebted to Professor Jack Oliver for his criticism and encouragement during the course of the present study and to staff members of the Lamont Geological Observvatory for valuable discussions. Thanks are also due Mrs. Anne Mowat, who has done most of the calculations.

Accumulation of the data under the IGY program was supported by National Science Foundation Grant Y/11.12/54. This work was supported in part by the Air Force Cambridge Research Center under Contract AF 19 (604) 7376, 8485 and the Office of Ordnance Research under Contract DA-30-069-Ord-2431.

## References

Aki, K.
1960a. "Study of Earthquake Mechanism by a Method of Phase Equalization Applied to Rayleigh and Love Waves"; *J. Geophys. Res.*, v.65, p. 729.
1960b. "Source Functions of Circum-Pacific Earthquakes"; *J. Geophys. Res.*, v.65, no.8, p. 2405.

Aki, K. and F. Press
"Upper Mantle Structure Under Oceans and Continents from Rayleigh Waves"; *J. Geophys. Res.* (in press).

Ben-Menahem, A.
1961. "Deriving Fault Length and Rupture Velocity from Seismograms": theory (abstract); *Bull. Seism. Soc. Am.* 51: 15–33.

Bolt, B. and J. Dorman
"Phase and Group Velocities of Rayleigh Waves in a Spherical, Gravitating Earth"; *J. Geophys. Res.* (in press).

Brune, J.
1961. "Radiation Pattern of Rayleigh Waves from the Southeast Alaska Earthquakes of July 10, 1958";, *Sym. on Earthquake Mech.*; Pub. of Dominion Obs., v.24, no.10.

Brune, J., J. Nafe and J. Oliver
1960. "A Simplified Method for the Analysis and Synthesis of Dispersed Wave Trains"; *J. Geophys. Res.* v.65, p.287.

Dorman, J., M. Ewing and J. Oliver
1960. "Study of Shear Velocity Distribution within Upper Mantle by Mantle Rayleigh Waves"; *Bull. Seism. Soc. Am.* 50: 87–115.

Espinosa, A. F., G. H. Sutton and H. J. Miller, S. J. "A Transient Technique for Seismograph Calibration"; *Bull. Seism. Soc. Am.* (in press).

Ewing, M., and F. Press
1954. "An Investigation of Mantle Rayleigh Waves"; *Bull. Seism. Soc. Am.* 44: 127–148
1956. "Rayleigh Wave Dispersion in the Period Range 10 to 500 seconds"; *Trans. Am. Geophys. Union*, 37: 213–215.

Gutenberg, B.
1959. "The Asthenosphere Low-Velocity Layer"; *Annali di Geofisica*, 12: 439–460.
Nafe, J. and J. Brune
1960. "Observations of Phase Velocity for Rayleigh Waves in the Period Range 100 to 400 seconds"; *Bull. Seism. Soc. Am.* 50: 427–439.
Officer, and M. Ewing
1955. "Southwest Pacific Crustal Structure"; *Trans. Am. Geophys. Union*, 36: 449–459.
Oliver, J., M. Ewing, and F. Press
1955. "Crustal Structure and Surface Wave Dispersion, Part 4: Atlantic and Pacific Ocean Basins"; *Bull. Seism. Soc. Am.* 66: 913–946.
Press, F., A. Ben-Menahem and N. Toksoz
1961. "Deriving Fault Length and Rupture Velocity from Seismograms": Experiment (abstract) presented at the Annual Meeting of the Seism. Soc. of Am.
Santô, T. A.
1960a. "Observation of Surface Waves by Columbia-type Seismograph Installed at Tsukuba Station, Japan (Part I)—Rayleigh Wave Dispersion Across the Ocean Basin"; *Bull. of Tokyo Earthq. Res. Inst.*, 38: 219–240.
1960b. "Rayleigh Wave Dispersions Across the Oceanic Basin Around Japan (Part II)", *Bull. of Tokyo Earthq. Res. Inst.*, 38: 385–401.
1960c "Rayleigh Wave Dispersions Across the Oceanic Basin Around Japan (Part III), *Bull. of Tokyo Earthq. Res. Inst.*, 39: 1–22.
Sutton, G and J. Oliver
1959. "Seismographs of High Magnification at Long Periods"; *Ann. Geophys.*, 15, no.4, p.423.
Sutton, G., M. Major and M. Ewing
1958. "Oceanic Rayleigh Wave Dispersion in the Period Range 15–140 sec: The Group Velocity Maximum and Associated Airy Phase", *Bull. Seism. Soc. Am.* (abstract).

Lamont Geological Observatory
Columbia University
Palisades, New York
Contribution No. 526

Manuscript received August 2, 1961.