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THE PROPAGATION OF SHORT-PERIOD SEISMIC SURFACE WAVES ACROSS OCEANIC AREAS PART II—ANALYSIS OF SEISMOGRAMS

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ABSTRACT

The propagation of short-period oceanic surface waves with predominant periods between 5 and 20 seconds was studied for a large number of paths in the Atlantic, Indian and Pacific oceans. In the ocean basins these waves are controlled largely by the sedimentary layer. The amplitudes of the short-period waves are greatly dependent upon the nature of the propagation path and on the properties of the source. Surface waves with periods between 5 and 10 seconds are often observed for paths that cross the continental margins of the Pacific, but are recorded only rarely for similar transmission paths in the Atlantic. Probably this effect is due to the differences in the structural configuration of the margins, particularly to such differences in the sedimentary strata.

Seismic refraction and reflection measurements indicate that most of the oceanic paths for which the 5 to 10 second waves are observed are characterized by sedimentary thicknesses that do not average more than a few tenths of a kilometer. The predominant periods of the short-period wave train are increased to as much as 12 to 20 seconds for paths that traverse the thick sediments of the Argentine Basin. This increase in period, which was predicted theoretically for regions of relatively thick sediments, indicates that the low-rigidity sediments play a prominent role in determining the character of the short-period wave train. The increase in the predominant periods of the waves associated with the first Love and first shear mode also accounts for the absence of the 5 to 10 second waves in areas of thick sediments such as the continental margins of the Atlantic.

Dispersion data for the Rayleigh, first Love and first shear modes were used in conjunction with reflection and refraction results to estimate the average shear velocity in the sediments of the Argentine Basin. The average shear velocity in the upper 0.5 km of sediments is about 0.2 to 0.4 km/sec, and the velocity in the kilometer of sediments below this is about 0.5 to 0.7 km/sec. Sedimentary shear velocities of a few tenths of a kilometer per second were also obtained for paths along which the average sedimentary thicknesses are a few tenths of a kilometer.

For group velocities between 4.2 and 3.4 km/sec the particle motion in the first shear mode is retrograde; ratios of horizontal-to-vertical motion as large as 3.5 are observed at island and coastal stations.

Seismic waves with periods of 10 to 30 seconds and with group velocities of 4.0 to 4.4 km/sec are sometimes recorded on the vertical components of long-period seismographs. These waves may be explained as higher modes of the Rayleigh type. Oscillatory waves with periods of 5 to 10 seconds are observed to follow the P wave in many oceanic areas. These arrivals are attributed to a type of leaking mode that results from the multiple reflection of SV waves in a low-rigidity layer.

Introduction

This paper describes an analysis and interpretation of short-period oceanic surface waves. These waves, which are distinguished by their oscillatory character and by their long duration on the seismogram, have periods between 5 and 20 seconds and are one of the most commonly observed arrivals when the path is predominantly oceanic. Many of the features associated with the propagation of these waves had not been explored adequately prior to this study. These features include the peculiar geographic distribution of the waves and the effect of low-rigidity sediments on the propagation of the waves.

The short-period waves together with the Rayleigh and Love waves of longer periods make up virtually all of the surface wave train for paths in oceanic regions. In this paper data on all of these waves are used in conjunction with calculations for a number of theoretical models (see Part I, Sykes and Oliver, 1964) to obtain information about the physical properties of the sediments, crust and upper mantle. In particular, dispersion data for the short-period waves and information from seismic reflection profiles are employed to estimate the shear velocities in deep-sea sediments.

Previous investigations revealed that the epicenters of the earthquakes that produced the short-period waves have a peculiar geographic distribution. In these previous studies data were used from only one or two stations in the case of a single earthquake. Thus, it was difficult to decide whether the transmission path or the source was responsible for the presence of these waves for certain earthquakes. The distribution of paths, especially those crossing continental margins, was quite limited in these previous investigations. These observational difficulties were overcome in this study through the use of a large number of recording stations located throughout the world. Long-period, high-gain instruments, which were installed in many parts of the world during and after the I.G.Y., provided valuable data of a quality not available for earlier studies.

These recordings demonstrate that both the effect of the propagation path and the nature of the source are responsible for the peculiar geographic distribution of the short-period surface waves. The effect on the waves as a result of transmission across continental margins was studied in some detail. The data show that surface waves with periods of 5 to 10 seconds are propagated readily across the continental margins of the Pacific, but are rarely observed for similar paths in the Atlantic. Probably this effect is due to differences in the structural configuration of the margins, particularly to such differences in the sedimentary strata.

The predominant periods of the short-period wave train are increased to as much as 12 to 20 seconds for paths that traverse the thick sediments of the Argentine Basin. The calculations in Part I of this study show that the predominant periods of waves propagated in the first Love and first shear modes are highly dependent upon the thickness and shear velocity in a low-rigidity sedimentary layer. The association of these longer-period waves with the thick sediments of the Argentine Basin is strong evidence that the sediments play a prominent role in determining the character of the short-period wave train. The increase in the predominant periods of the waves associated with the first Love and first shear modes also accounts for the absence of the 5 to 10 second waves in areas of thick sediments such as the Argentine Basin and the continental margins of the Atlantic. These results may have an important bearing on the propagation of microseisms across oceanic areas, and theories that assume comparable propagation characteristics for all regions may be in error.

In the following sections previous investigations of the short-period waves are reviewed, observational data pertinent to the waves are examined, and these observations are used in conjunction with theoretical calculations and with other geophysical data to examine the nature of the wave propagation in various geological environments.

Previous Studies of the Short-Period Waves

Surface waves with periods of 8 to 10 seconds were observed at Honolulu as early as 1929 (Neumann). Carder (1934) found two predominant periods in the recordings of surface waves for oceanic paths to Berkeley. Ewing and Press (1952, 1953) and Oliver, Ewing and Press (1955) noted a similar duality in the periods of surface waves at Honolulu. The longer-period wave train in the latter three studies was identified as the fundamental Rayleigh mode, but the identification of the short-period waves was incomplete. The shorter-period waves were often observed on all three components of the seismogram, but the amplitudes were normally largest on the transverse component. Ewing and Press (1952, 1953) and Oliver, Ewing and Press also found that these waves exhibited a peculiar geographic distribution in both the Atlantic and Pacific oceans.

Press and Ewing (1952) described the efficient propagation of the surface waves L_g and R_g in continental areas and pointed out the absence of these phases for paths involving even a small segment of oceanic path. Ewing and Press (1952) and Ewing and Donn (1952) called attention to the continental margins as a barrier to the propagation of surface waves and microseisms for periods less than 15 seconds. Oliver and Ewing (1958) also discussed the possible channeling of microseisms by sediment filled troughs such as those found along the continental margins of eastern North America.

Coulomb (1952), in a study of the Love waves of the Queen Charlotte Islands earthquake of 1949, also observed surface waves with periods as short as 8 seconds at Honolulu. Evernden (1954) derived a Pacific crustal structure from the dispersion of Love waves in the period range 7 to 40 seconds. Evernden pointed out the approximate fit of the dispersion curves for the first shear mode and the dispersion data for the short-period waves. Oliver, Ewing and Press (1955) identified the waves at the beginning of the short-period train as Love waves. Nevertheless, neither they nor Ewing and Press (1952) were able to account quantitatively for the waves in the latter portions of the record as either Love modes or shear modes.

DeNoyer (1959) studied Love waves with periods of 8 to 25 seconds and with group velocities of 4.4 to 3.6 km/sec for a path from the Kurile Islands to Berkeley. He was able to fit the observed dispersion by using a model with a low-velocity surface layer. Computations were not given, however, for the first shear mode.

Oliver and Dorman (1961) obtained quantitative agreement between computed and observed dispersion curves for all the waves by taking into account the effect of a small but finite rigidity in the sediments. Oliver and Dorman showed that the short-period waves correspond to propagation in the first Love and first shear modes. Sedimentary shear velocities of 0.25 and 0.50 km/sec were used to fit the dispersion data for a path from Alaska to Honolulu.

These previous studies indicate that the short-period waves might be used to investigate the properties of the sedimentary and crustal layers in many different oceanic areas. The work of Oliver and Dorman (1961) suggests that many of the features of the short-period waves such as the period, velocity, amplitude and attenuation might be dependent upon the nature of the sedimentation along the propagation paths. These features will be investigated in the following sections.

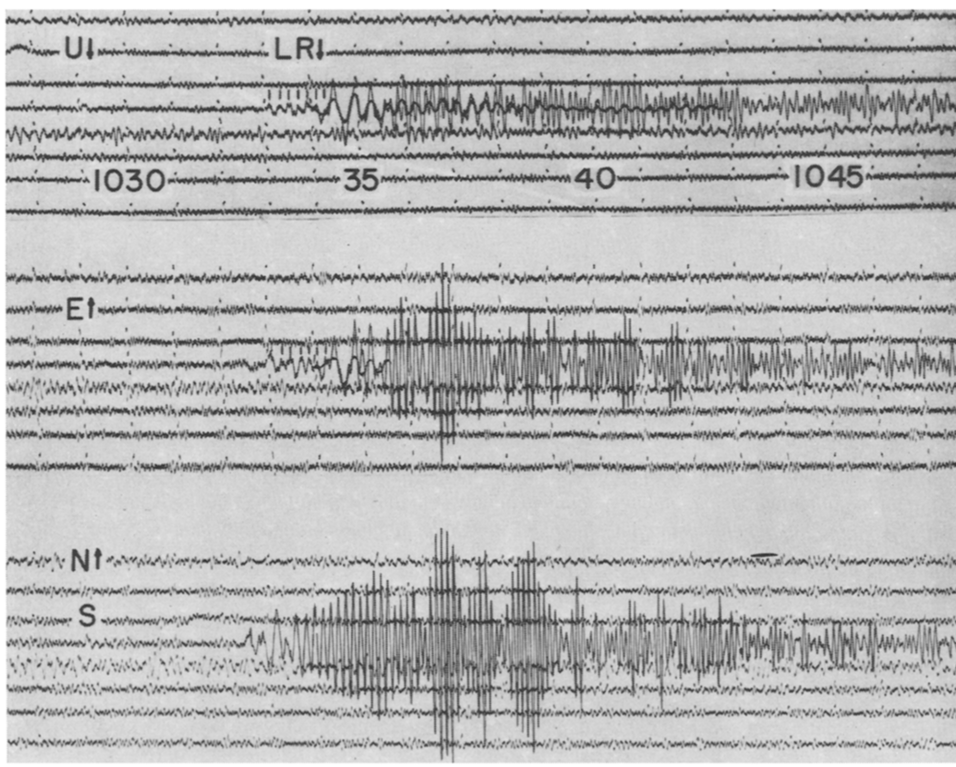

FIG. 1. Honolulu recording of an earthquake off the west coast of Mexico at 19.9N, 109.3W on July 28, 1961, showing short-period surface waves on all three components of the seismogram. Arrows indicate motion up, east and north. Longer-period Rayleigh mode pencilled in on vertical and east components. Seismometer period 15 seconds; galvanometer period 75 sec. Clock correction +0.5 sec. Epicentral and other data in table 1.

HONOLULU RECORDINGS OF SHORT-PERIOD SURFACE WAVES FROM MEXICO

The records shown in figure 1 illustrate the short-period waves as well as any known to the authors. This earthquake occurred off the west coast of Mexico on July 28, 1961 (coordinates and other data in table 1). The waves arrived at Honolulu from nearly due east. The great circle path is shown as a dashed line in figure 2. Large, oscillatory waves with a predominant period of 6 to 10 seconds are present on all three components of the seismogram. The longer-period Rayleigh mode, which begins at about 1034 GCT with waves of periods of about 30 seconds, was pencilled in on the vertical and east components. On these two seismograms waves of the first shear mode begin at approximately 1033 GCT, well ahead of the Rayleigh

TABLE 1

EPICENTRAL AND OTHER DATA FOR EARTHQUAKES ANALYZED IN THIS STUDY

Date	Origin Time			Latitude	Longitude	Magnitude	Station	Distance, km	Azimuth at Station	% Continental Path	Water Depth, km
	h	m	s								
1954											
Oct 21*	00	10	09	41.3°S	80.2°E	7	PIE	4638			
1958											
Feb 28*	09	54	53	$27\frac{1}{2}$°N	$43\frac{3}{4}$°W		BEC	2085			
Apr 12*	11	46	58	$26\frac{1}{2}$°N	111°W	6–$6\frac{1}{2}$	SUV	9074			
Apr 14*	21	32	28	01°N	$79\frac{1}{2}$°W	$6\frac{3}{4}$–7	HON	8787			
Aug 16*	13	17	52	$51\frac{1}{2}$°N	176°W	6–$6\frac{1}{4}$	HON	3692			
Aug 27	02	25	35.4	04.0°S	104.3°W	$5\frac{3}{4}$–6	HUA	3312	283.2°	13	4.1
Nov 14	05	04	24.9	36.0°S	102.8°W	$6\frac{1}{4}$–$6\frac{1}{2}$	HLL	6139	99.6°	6	4.0
1959											
Jul 11	12	01	41.1	36.9°S	78.6°E	$6\frac{1}{4}$–$6\frac{1}{2}$	PIE	4517		5	3.7
Dec 17	16	48	54.7	36.2°S	102.8°W	6	HLL	6123	99.7°	6	4.0
1960											
Jan 8	21	42	41.7	35.5°S	17.2°W		BAA	3736	103.5°	$14\frac{1}{2}$	
Apr 15	03	25	39.3	26.9°S	113.4°W	$6\frac{1}{2}$	HLL	6720			
							SUV	6997		0	3.8
Aug 20	20	08	35.0	35.6°S	15.8°W	6–$6\frac{1}{2}$	BAA	3866	104.2°	$14\frac{1}{2}$	
Aug 30	06	45	12.7	21.0°S	113.9°W		HUA	4220	251.3°	6	4.2
Sept 1*	20	01	57.2	15.9°S	178.9°W	$5\frac{1}{2}$	HON	4696			
Sept 7	01	17	37.3	36.9°S	15.8°W	$5\frac{1}{4}$	BAA	3841	106.4°	$14\frac{1}{2}$	
Oct 7	20	01	11.4	20.3°S	114.1°W	$5\frac{1}{4}$–$5\frac{1}{2}$	HUA	4233	252.5°	6	4.2
1961											
Mar 7*	19	08	36.1	38.4°S	78.1°E	6	PER	3486	247.6°	4	4.2
Jun 10*	20	31	50.9	24.2°S	112.1°W	6	BAA	5268	267.7°		
							HLL	7048	86.0°		
							HON	7075	132.8°		
							HUA	4098	245.7°	7	4.2
							MTJ	13093	100.8°		
							RDJ	6954	253.7°		
							SUV	7171	108.6°		
Jul 28	10	13	45.6	19.9°N	109.3°W	$5\frac{1}{2}$	HON	5064	82.6°	0	4.5

* Asterisks indicate epicentral data from USCGS, BCIS or ISS. All other epicenters were relocated using a program similar to the one described by Bolt (1960).

waves and with a group velocity of approximately 4.4 km/sec and a period of about 14 seconds. Waves of the Rayleigh and first shear modes are superposed on both of these components for group velocities less than 4.1 km/sec. The ratio of horizontal-to-vertical motion in the first shear mode at Honolulu is approximately 3.5 at 8 seconds.

The 6 to 10 second waves on the transverse component (N-S) are nearly always larger than the waves on the longitudinal component (E-W) even when the relative magnifications are taken into account (North = 1.6 × East). This indicates that the waves on the N-S component are probably Love waves. An examination of seismograms for many different oceanic paths indicates that the short-period waves are usually largest on the transverse components and smallest on the vertical components. The 6 to 10 second waves on the vertical component in figure 1, however, cannot be explained as Love waves. Hence, the seismograms for the earthquake off the west coast of Mexico illustrate clearly many features of oceanic surface wave

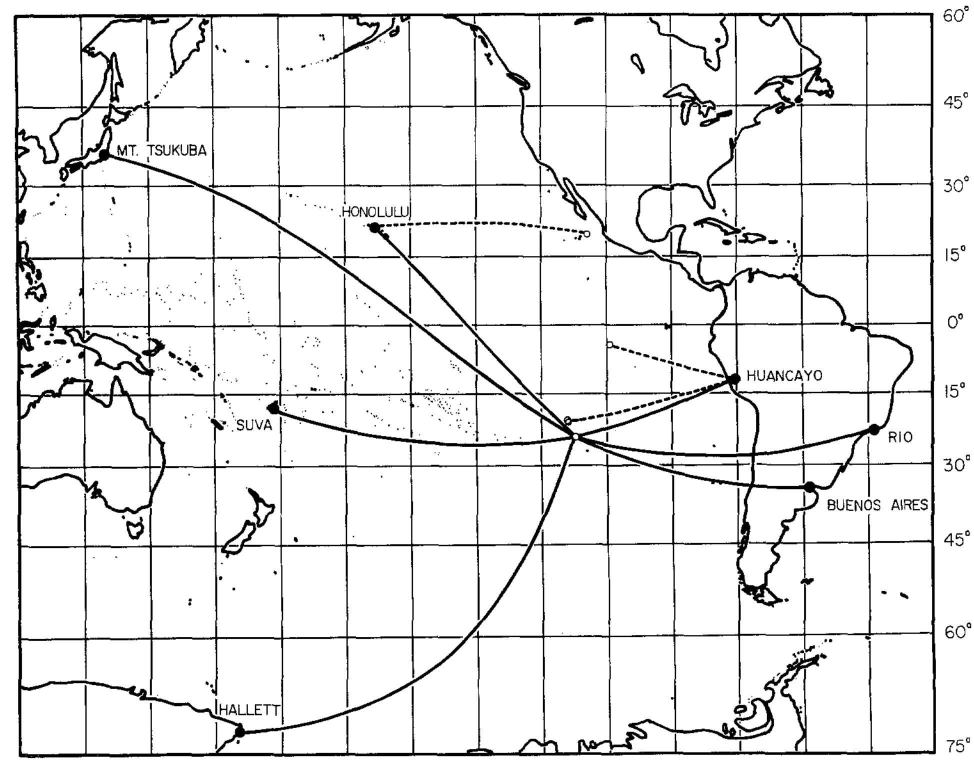

FIG. 2. Map showing surface wave recordings for long oceanic paths. Solid lines—earthquake on June 10, 1961, near Easter Island. Broken lines—great circle paths from west of Mexico to Honolulu and from Easter Island ridge to Huancayo.

propagation and support strongly the hypothesis that the short-period surface waves are propagated in one of the Love modes and in one of the shear modes.

EFFECT OF TRANSMISSION PATH AND FOCAL DEPTH

The effect of the transmission path on the propagation of the short-period waves was studied over a wide range of azimuths for earthquakes in the Atlantic, Indian and Pacific oceans. This effect was well illustrated on a number of seismograms for an earthquake in the southeast Pacific near Easter Island on June 10, 1961. The great circle paths to various stations are shown in figure 2; the seismograms from four of these stations are illustrated in figure 3. Long-period Rayleigh and Love waves were recorded at all of the stations shown in figure 2. Short-period waves with periods between 5 to 10 seconds were only observed at Huancayo, Honolulu, Rio and Buenos Aires. The short-period waves recorded at Rio and Buenos Aires

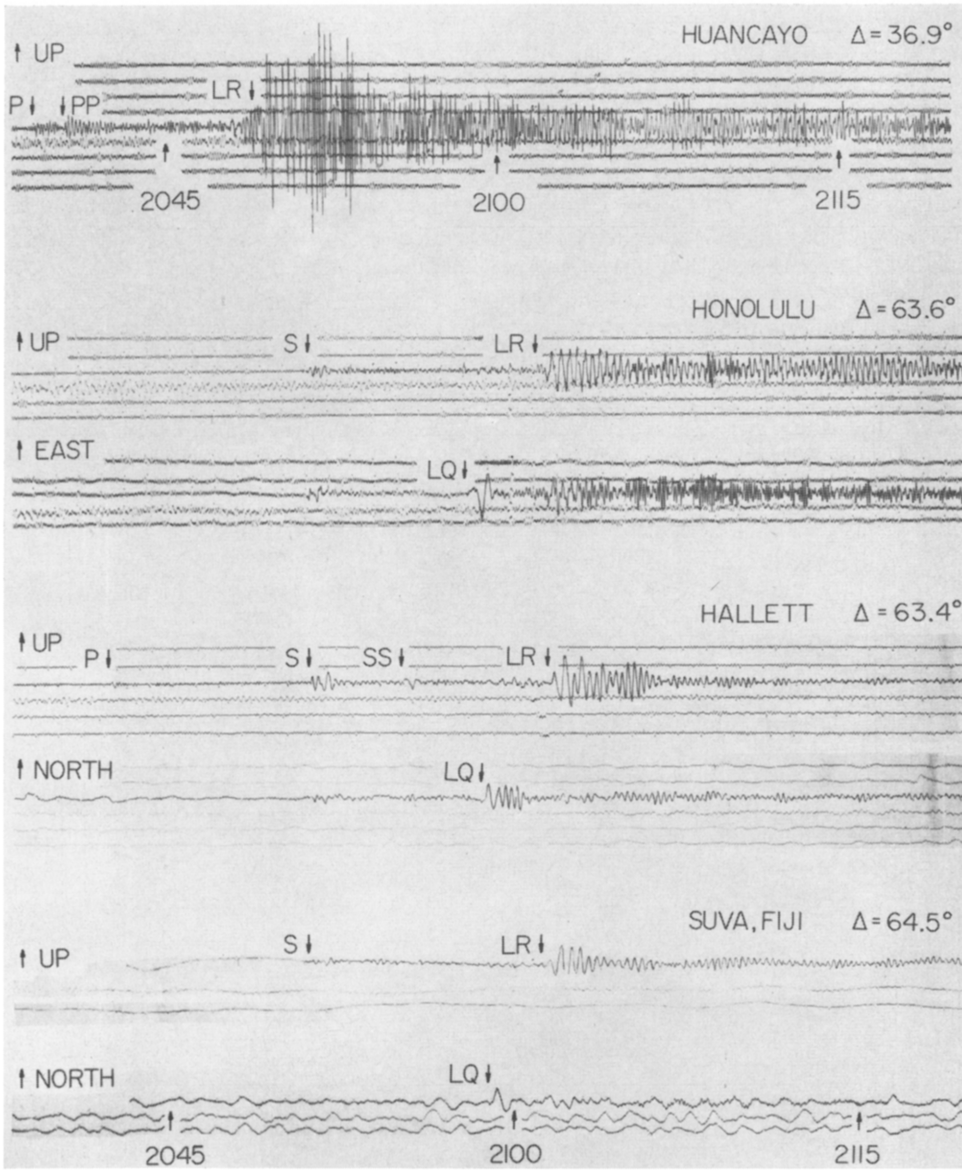

FIG. 3. Surface waves at Huancayo, Peru; Honolulu, Hawaii; Hallett, Antarctica; and Suva, Fiji, for an earthquake on June 10, 1961, near Easter Island. Distances given in degrees. Seismometer period for all instruments is 15 seconds. Galvanometer period is 7 seconds at Huancayo and 75 seconds at the other stations. The seismogram traces were darkened with a pencil to insure adequate photographic reproduction. The 5 to 10 second surface waves were only recorded at Huancayo and Honolulu. Mantle Rayleigh waves with periods between 50 and 100 seconds can be seen on the vertical component of the Hallett recordings.

are Love waves of small amplitude and are only visible on the N-S component. Hallett and Suva did not record the 5 to 10 second waves even though the distance to each of those stations is nearly the same as the distance to Honolulu. The great circle path to Hallett is along the Easter Island ridge; the path to Suva is along the Tuomotu Archipelago. The short-period waves were not observed at Mt. Tsukuba.

The Huancayo instruments (seismometer period, 15 sec; galvanometer period 7 sec) are more sensitive to the short-period surface waves than the other instruments (15 sec seismometer, 75 sec galvanometer). Nevertheless, a closer examination shows that the absence of the 5 to 10 second waves at some of the stations with the less sensitive instruments cannot be ascribed to differences in instrumental response. Likewise, the absence of these waves cannot be attributed to the radiation pattern unless the focal mechanism is completely different for the generation of the long-period and the short-period surface waves, for the long-period waves are well recorded at all stations. Nevertheless, there are no other reasons to suspect such a remarkable radiation pattern near the source, and variations in structure do occur in oceanic areas. Thus, for this earthquake the presence or absence of the short-period surface waves is probably related to the nature of the transmission path and not to the properties of the source. For this event the short-period waves were not observed for paths along oceanic ridges and rises.

Short-period waves are commonly recorded at Bermuda from earthquakes in the West Indies. This study revealed that such waves are not observed at Halifax, Nova Scotia, even though Halifax and Bermuda are situated along nearly the same azimuth from the earthquakes. Oliver, Ewing and Press (1955) reported a similar phenomenon at Bermuda and Palisades from earthquakes on the Mid-Atlantic ridge. Hence, for these events the presence or absence of 5 to 10 second waves must be attributed to the effect of the propagation path and not to the properties of the source. The transmission paths to both Halifax and Palisades cross the continental margins of eastern North America. Certain features of these continental margins are evidently responsible for the drastic attenuation of the short-period surface waves. The effect of the continental margins will be investigated further in another section.

At Honolulu the short-period surface waves are present on some records and are either small or absent on others for epicenters in the Aleutian Islands (Ewing and Press, 1953; Oliver, Ewing and Press, 1955). The short-period waves were observed for all of the earthquakes examined in this study when the epicenters were located on the more seaward (southern) side of the Aleutian arc. When these waves are absent or when only small Love waves are recorded, the source is nearly always located on the northern side of the island arc. Such an effect might be caused by either propagation effects or by a difference in excitation at the source. The latter may result from differences either in the focal depth or in the spectrum of the source. However, since earthquakes in central Alaska (figure 4) make large short-period waves at Honolulu, the effect of the continental margin or island arc on the propagation does not seem to be completely responsible for the absence or smallness of these waves for certain earthquakes in the Aleutians. Hence the properties of the focus as well as the effect of the transmission path must be called upon to account for the presence or absence of the short-period surface waves.

Geographic Distribution of the Short-Period Surface Waves

Carder (1934), Ewing and Press (1952, 1953), and Oliver, Ewing and Press (1955) indicated that the epicenters of the earthquakes that produce the short-period

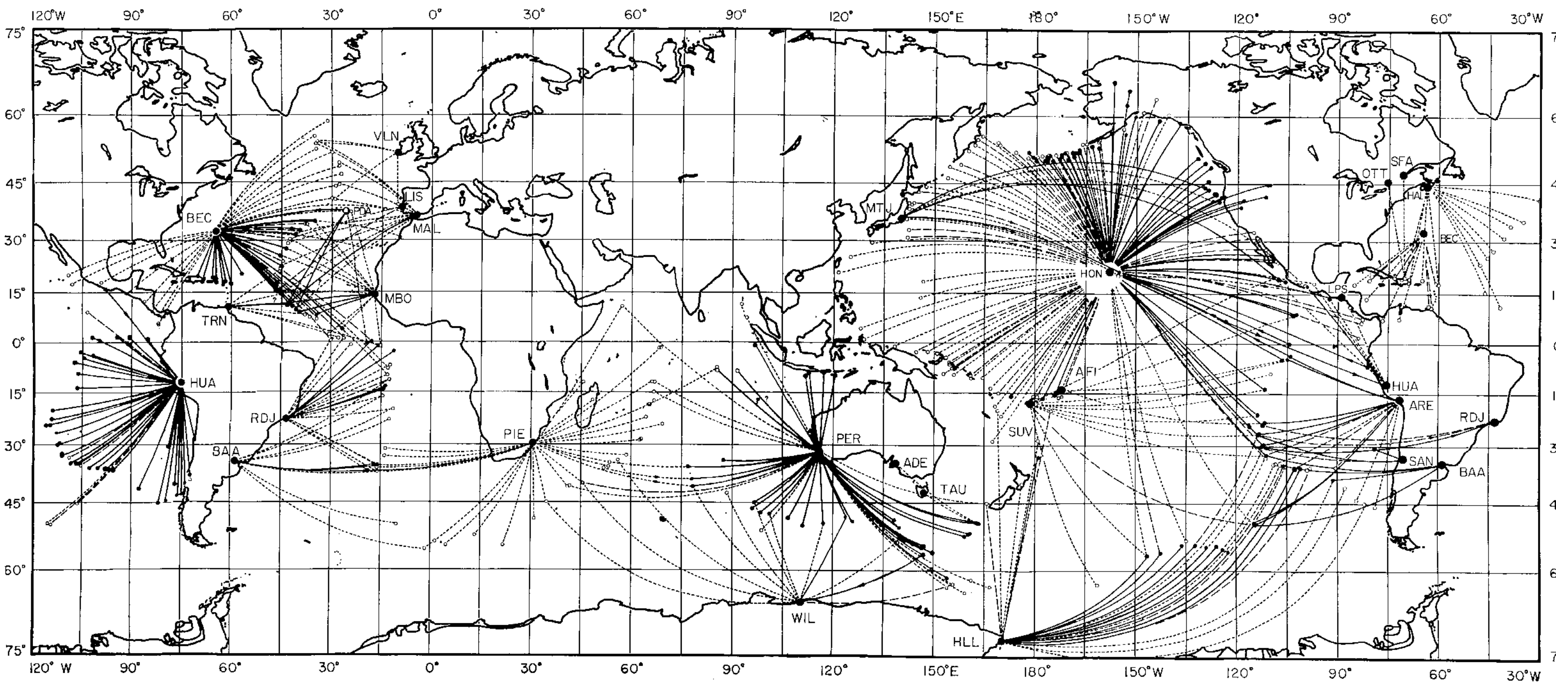

FIG. 4. Geographic distribution of short-period surface waves. Solid lines along paths and solid circles at epicenters correspond to shorter period surface waves on all three components, long dashes along paths and open circles at epicenters—only short-period Love waves (usually of small amplitude and of short duration on records), short dashes along paths and open circles at epicenters—long-period Rayleigh waves observed but short-period surface waves absent. Three-letter abbreviations of U. S. Coast and Geodetic Survey used for stations. Question marks indicate doubtful identification of wave types. With the exception of paths from Tristan da Cunha to Buenos Aires and from Melanesia to Honolulu, the predominant periods associated with the short-period wave train are less than 10 seconds.

surface waves have a peculiar geographic distribution. These waves were observed for some Pacific paths to Honolulu and Berkeley, but were not detected for other Pacific paths to the same stations.

In this investigation the geographical distribution of the short-period surface waves was studied for most of the oceanic areas of the world with the exception of the Arctic Ocean. Seismograms were examined from the Lamont long-period network and from many of the stations of the U. S. Coast and Geodetic Survey network. Seismograms from Lisbon, Portugal; M'Bour, Senegal; Wilkes, Antarctica, and several Canadian stations were also studied. These observations are summarized in figure 4. When short-period waves were recorded by all three components of the seismograph, the corresponding paths are denoted by solid lines and the epicenters are indicated by solid circles. Paths for which only short-period Love waves were recorded are indicated by long dashes; paths for which only long-period surface waves were observed are shown as short dashes. Short-period waves on the vertical component at Perth are also indicated by solid lines since horizontal component seismographs were not available at that station until 1961. Paths along which the water depth did not exceed 3 km were excluded in this study because of the difficulty in differentiating between the Rayleigh and first shear modes for these paths.

It is readily apparent from figure 4 that, as suggested earlier, short-period waves may be recorded at one station from a given shock and not at another. Thus our primary concern should be not over the geographical distribution of the epicenters of the shocks causing the short-period waves but rather over the geographical distribution of the paths over which the waves propagate. Let us now consider the relationship between the geographical distribution of the short-period waves and the following parameters: the amplitudes and periods of the short-period waves, the nature of the continental margin, the thickness of sediments along the transmission path and the depth of the seismic source.

Amplitudes and Periods of the Short-Period Surface Waves

When the short-period waves are observed (figure 4), the predominant periods of the waves are nearly always between 5 and 10 seconds. In a few unusual cases such as the Argentine Basin, the predominant periods of the waves are as great as 12 to 20 seconds. The amplitudes of the short-period waves relative to those of the long-period Rayleigh and Love waves also vary geographically. Large, short-period waves similar to those in figure 1 are recorded at Honolulu from earthquakes near the west coast of North America and from shocks north of Easter Island on the mid-oceanic ridge; at Huancayo, Arequipa and Santiago from earthquakes on the Easter Island ridge; at Bermuda from shocks in the West Indies and on the Mid-Atlantic ridge; at Perth from events near Java and from epicenters to the south and southeast; and at Hallett from shocks in the southeast Pacific.

The short-period waves are not as well developed for the other oceanic paths studied. The short-period arrivals are small and occur late in the surface wave train on the Perth recording (figure 13) of an earthquake in the central Indian Ocean. Other recordings of the short-period waves, notably those at Rio and Pietermaritzburg, are similar to the waves on the Perth seismogram. The short-period waves are also small for paths that follow the Nasca Ridge to Huancayo.

The short-period wave trains are usually best developed for oceanic paths that are structurally simple in the sense that the paths cross a single ocean basin to reach the seismograph stations. The short-period waves are normally small or absent for paths that cross oceanic ridges.

Nature of the Continental Margin

Short-period surface waves with periods between 5 and 10 seconds are conspicuously absent for nearly all of the paths that cross the continental margins of the Atlantic Ocean even though the waves may have propagated well to Bermuda. Surface waves with periods between 5 and 10 seconds were observed on all three components of the seismogram at Honolulu from earthquakes in Montana, Utah, Nevada and central Alaska (figure 4). In each case the short-period waves were propagated across the continental margins of the Pacific and across a considerable length of continental path. Short-period waves were also observed at Berkeley and at Palisades (although weakly) from earthquakes in Hawaii (Oliver, Ewing and Press, 1955). The 5 to 10 second waves were recorded at stations on both the eastern and western coasts of South America from earthquakes in the southeast Pacific. The mechanism of propagation in the continents is not known; however, the group velocity for the continental portion of these paths is apparently about that of surface waves and is certainly not as fast as the velocity of P waves.

The relative ease with which the short-period waves cross these continental margins of the Pacific is in marked contrast to the difficulty with which these waves traverse the continental margins of the Atlantic that were studied here. The continental margins of these oceans differ in several important respects. The Circum-Pacific zone, unlike the continental margins of the Atlantic, is characterized by a high level of seismic activity, by the occurrence of tectonic movements and mountain building and by the presence of deep-sea trenches. Seismic refraction and reflection work shows that great thicknesses of sediments are found along the continental margins of the following areas: eastern North America (Drake, Ewing and Sutton, 1959; Ewing, 1963), the western approaches of the English Channel (Day and others, 1956) and Argentina (Ewing, 1963; Ewing, Ludwig and Ewing, 1963, 1964). Similar measurements along the eastern margins of the Pacific [South America: Fisher and Raitt (1962), Ewing (1963); Central America: Fisher (1961), Shor and Fisher (1961); Baja, California: Fisher and Hess (1963); Southern California: Uchupi and Emery (1963); Northern California: Menard (1960a, describes some of Raitt's unpublished results); and the Gulf of Alaska: Shor (1962)] indicate that the sedimentary thicknesses are much smaller than those found along many of the continental margins of the Atlantic. Abyssal plains adjacent to the continental margins are also more numerous in the Atlantic than in the Pacific (Heezen and Loughton, 1963). Thus the smaller thicknesses of sediments near the continental margins seem to occur at places where the propagation of 5 to 10 second surface waves across the margins is good. Nonetheless, other differences in the structure of the margins may also be partly responsible for the presence of the 5 to 10 second waves for paths across the continental margins of the Pacific and for the absence of the waves for similar paths in the Atlantic.

The 5 to 10 second waves were recorded for a few paths that cross the continental margins of the Atlantic. These include arrivals at M'Bour from earthquakes along the Mid-Atlantic ridge between 7° and 11°N, arrivals at Rio from shocks along the Mid-Atlantic ridge between 0° and 15°S, and one recording at Pietermaritzburg, South Africa, from an earthquake on the Mid-Atlantic ridge near Tristan da Cunha. Drake, Ewing and Sutton (1959, after Warren and Ewing) reported smaller sedimentary thicknesses for a profile northeast of Recife, Brazil. These refraction measurements indicate that thick sediments are not always found along the continental margins of the Atlantic. This evidence is nowhere in disagreement with the data on short-period waves at a few stations near the borders of the Atlantic, however. Since reflection and refraction data are not available in the areas where the 5 to 10 second waves are observed, a direct comparison with the structure cannot be made.

Short-period waves with a predominant period of approximately 7 seconds were recorded with a hydrophone on the ocean bottom (G. Latham, Lamont Observatory, personal communication). This measurement demonstrates that a continental margin is not required for the generation of the short-period surface waves as supposed by Crenn and Metzger (1959). Likewise, the presence of an island at the recording end of the path is also not a requisite for the short-period waves. The limited observations available do not disagree with the mechanism of propagation discussed here for the short-period waves.

Thickness of Sediments along the Transmission Paths

Oscillatory wave trains with predominant periods of 12 to 20 seconds were observed on all three components of ground motion at Buenos Aires, Argentina, from earthquakes on the Mid-Atlantic ridge near Tristan da Cunha. The great circle paths to Buenos Aires cross the Argentine Basin, one of the largest deposits of sediments in the ocean basins (Ewing, Ludwig and Ewing, 1963, 1964; Ewing, 1963). The transverse component of one of the Buenos Aires recordings is shown in figure 5. The surface waves at Buenos Aires are practically identical to the typical waves recorded at Honolulu (figure 1) from a shock off Mexico with but one exception—the predominant periods of the waves are much greater at Buenos Aires. This indicates that the mechanism of normal mode propagation is quite similar in the two areas. The calculations presented in Part I of this study show that the predominant periods of waves propagated in the first Love and first shear modes are highly dependent upon the thickness and shear velocity in a low-rigidity sedimentary layer. These calculations also indicate that the predominant periods of the waves should increase as the sedimentary thickness becomes very large. The association of the 12 to 20 second waves with the thick sediments of the Argentine Basin indicates that the sediments play a predominent role in determining the character of the short-period wave train.

Many of the oceanic paths for which the 5 to 10 second waves are observed are characterized by sedimentary thicknesses of a few tenths of a kilometer. Refraction and reflection measurements indicate average sedimentary thicknesses of a few tenths of a kilometer in the Gulf of Alaska (Shor, 1962), in the central and eastern Pacific (Raitt, 1956; Shor, 1959; Ewing, Ewing and Talwani, 1964), on the flanks of the Mid-Atlantic ridge (Ewing and Ewing, 1963; Ewing, Ewing and Talwani,

1964) and in the region between the West Indies and Bermuda (Ewing and Ewing, 1962, 1963). The difference between the average sedimentary thickness along the great circle paths in the Argentine Basin (approximately 1.5 km) and the average thickness in each of these regions is sufficient to account for the change in the predominant period of the waves. The increase in the predominant periods of the waves associated with the first Love and first shear modes also accounts for the absence of the 5 to 10 second waves in areas of thick sediments such as the Argentine Basin and the continental margins of the Atlantic. Such results have obvious application to the propagation of 5 to 10 second microseisms across oceanic areas. Many of the so-called "microseismic barriers" may be related to the presence of great thicknesses

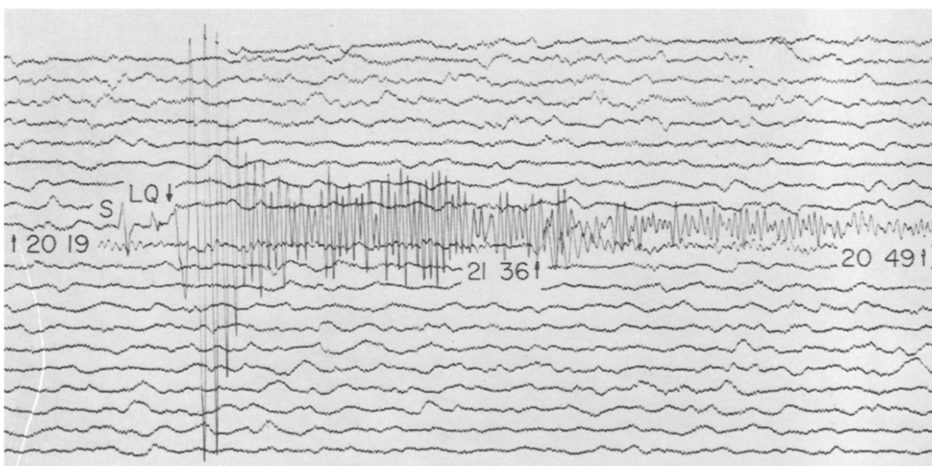

FIG. 5. Surface waves recorded at Buenos Aires from an earthquake on the Mid-Atlantic ridge near Tristan da Cunha on August 20, 1960. Path traverses thick sediments of Argentine Basin. North-south component shown; upward motion on the figure represents movement to the north. Azimuth nearly east (104.2°). Predominant period of highly dispersed Love waves is about 12 to 20 seconds. Arrival between S and LQ may be SS. Constants of instruments: seismometer period 15 seconds, galvanometer period 75 seconds. Clock correction −91 sec. Surface waves from an aftershock shown at 2136 GCT.

of sediments or to differences in the structural configuration of the various continental margins. Theories that assume comparable propagation characteristics for all regions may be in error.

Seismic reflection profiling shows that sedimentary thicknesses can be quite variable in some oceanic areas. On the western flank of the Mid-Atlantic ridge near 30°N, the sediments are irregular in thickness and are absent in some areas (Ewing and Ewing, 1963; Ewing, Ewing and Talwani, 1964). The short-period waves are recorded at Bermuda for paths that cross these regions of variable sedimentary thickness. Thus the 5 to 10 second waves can traverse areas where the sediments are absent or are of variable thickness.

The theoretical calculations in Part I indicate that for group velocities greater than about 1 km/sec most of the energy in the first Love and first shear modes is propagated in the crust and upper mantle and not in the sediments. Thus these calculations suggest that waves corresponding to transmission in the first Love and

TABLE 2

SHEAR VELOCITY MEASUREMENTS FOR LOW-RIGIDITY MATERIALS

Material	Location	Method	Compressional Velocity, km/sec	Shear Velocity, km/sec	Reference
Globigerina ooze	deep-sea core from NE Atlantic basin	Artificial compaction	2.68	1.20	Laughton (1957)
			2.89	1.42	Laughton (1957)
			3.06	1.57	Laughton (1957)
Welded tuff (Ignimbrite)	Vhakamaru, New Zealand	Direct timing	2.08	0.08	Evison (1956)
Pierre Shale	near Limon, Colorado	Refraction	2.28	0.85	McDonal and others (1958)
		Refraction	2.36	0.86	McDonal and others (1958)
Shale	near Mounds, Oklahoma	Surface wave analysis		0.3 to 1.5	Jolly (1956)
Loose sand	near Sand Spring, Oklahoma	Refraction	0.55	0.27	Jolly (1956)
Eagle Ford Shale	near Dallas, Texas	Refraction	2.46	0.61	White and Sengbush (1953)
			1.84	0.38	White and Sengbush (1953)
Austin Chalk	near Dallas, Texas	Refraction	3.07	1.07	White and Sengbush (1953)
			2.61	1.20	White and Sengbush (1953)
Alluvial Layers	near Tokyo, Japan	Spectra of multiple reflections		0.10	Shima (1962)
				0.32	Shima (1962)
				1.15	Shima (1962)
Fine Sand	Shallow water	*in situ*	1.68	(0.5)	Hamilton (1956) and Hamilton and others (1956)
Very fine sand	Shallow water	*in situ*	1.66	(0.5)	Hamilton (1956) and Hamilton and others (1956)
Medium silt	Shallow water	*in situ*	1.47	(0.0)	Hamilton (1956) and Hamilton and others (1956)
Clayey silt	Shallow water	*in situ*	1.46	(0.0)	Hamilton (1956) and Hamilton and others (1956)

first shear modes may propagate across areas in which the sediments are absent or are of irregular thickness without suffering drastic attenuation. Most of the dispersion in these wave trains would result from propagation across regions that have appreciable sediments. Hence, sediments along at least a portion of the path are necessary to explain the oscillatory character of the short-period surface waves.

In some ways this effect is analogous to certain features of the propagation of 15 to 20 second Rayleigh waves. Rayleigh waves in this period range are strongly affected by the water layer and become highly dispersed when the waves traverse oceanic regions. However, these waves can propagate across regions of varying water depth as well as across continental areas and transition zones. The amount of dispersion is small in continental areas, and a segment of oceanic path can produce as much dispersion in these waves as a segment of continental path ten times as long. In analogy with the short-period surface waves, most of the energy in the 15 to 20 second Rayleigh waves is propagated in the crust and upper mantle rather than in the water layer. Thus, it is not surprising that the short-period waves propagate across areas where the sedimentary thickness varies from zero to a few tenths of a kilometer.

Short-period waves were not recorded at M'Bour, Senegal, for earthquakes on the Mid-Atlantic ridge north of 15°N. Ewing, Ewing and Talwani (1964) found

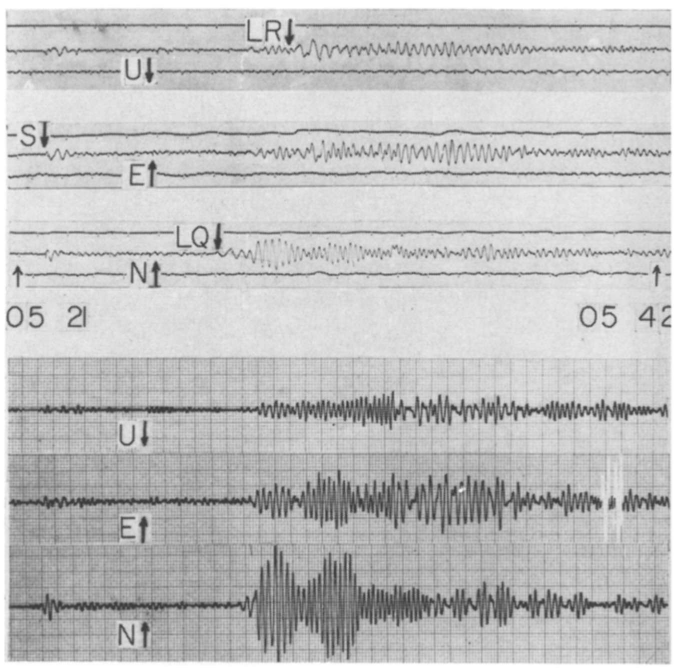

Fig. 6. Hallett, Antarctica, seismograms for earthquake on West Chile rise, November 14, 1958. Azimuth at Hallett nearly east (99.6°). Top—original records. Short-period waves superimposed with the longer-period Rayleigh waves on vertical and east components. Bottom—bandpass of 7 to 14 seconds used to enhance short-period surface waves. Clock correction −1.9 sec.

practically no sediments on the eastern flank of the Mid-Atlantic ridge near 30°N. The absence of the short-period waves at M'Bour from these azimuths may be caused by the absence of sediments. However, some other feature of the propagation path such as the configuration of the continental margin may also be responsible for the absence of the waves. Nonetheless, the absence of the short-period waves is not in disagreement with the reflection results cited above.

The short-period waves are not recorded at Hallett, Antarctica, for paths along the Easter Island ridge. However, these waves are observed (figures 4 and 6) for paths from the West Chile rise that pass to the east and south of the Easter Island ridge. This demonstrates once again that the transmission path has a large influence on the propagation of the short-period waves. The transmission paths along the ridge may be characterized by an absence of sediments or by a crustal structure that is more complex than that found in the basin to the southeast of the ridge. Either of these factors may account for the absence of the short-period waves for paths along the Easter Island ridge.

Short-period waves were observed at Wilkes, Antarctica, for two earthquakes in the Indian Ocean, but were not recorded for ten other events there. Perth recorded the 5 to 10 second waves from many of these ten events. Either some feature of the transmission path is responsible for the absence of these waves at Wilkes or else the waves have a most remarkable radiation. Thus it is often difficult to determine which of several mechanisms is responsible for the absence of the short-period waves at a given station.

Association with Shallow Focal-Depths

The short-period surface waves are usually, and perhaps always, associated with earthquakes with a shallow focal depth, i.e., less than 70 km. These include earthquakes along the mid-oceanic ridges, the Circum-Pacific belt from Mexico to the Aleutians, and the West Indies. Although deep earthquakes (depth greater than 70 km) occur in some of these regions, there is no indication that these shocks generate the short-period surface waves.

Particle amplitudes as a function of depth were computed for the first Love and first shear modes in Part I of this study (Sykes and Oliver, 1964). These amplitudes are very small for depths exceeding 50 km for periods less than 10 seconds. This indicates that a source below 50 km will not excite these modes very efficiently (Satô, Usami, and Landisman, 1963).

Particle Motion and the Ratio of Horizontal-to-Vertical Motion

The particle motion and the ratio of horizontal-to-vertical motion in the first shear mode are examined for three different sets of records in this section. These recordings are as follows: west of Mexico to Honolulu, July 28, 1961 (figure 1); West Chile rise to Hallett, Antarctica, November 14, 1958 (figure 6); and Easter Island ridge to Huancayo, Peru, August 30, 1960 (figure 7). Since the arrivals at Honolulu and Hallett were from nearly due east, the Rayleigh, Love and shear modes could be separated. The particle motion in the first shear mode was retrograde for group velocities between those corresponding to the initiation of the waves (4.4 to 4.2 km/sec) and 3.4 km/sec. The particle motion could not be discerned with confidence for group velocities less than 3.4 km/sec. On the Honolulu records the ratio of horizontal-to-vertical motion is slightly greater than unity for group velocities between 4.4 and 4.0 km/sec, but increases to 3.5 at a group velocity of 3.7 km/sec (1037 GCT). The instrumental magnifications at Honolulu were taken into account in these measurements. At 10 seconds they are as follows: Up 470, East 700, North 1100 (Miller, 1963).

On the Hallett records (figure 6, top) the Rayleigh waves are considerably larger than the waves propagated in the first shear mode. However, the noise level on the records is very low, and the short-period waves can be distinguished easily from the microseismic noise. In order to separate the waves propagated in the Rayleigh and shear modes, the original records were transcribed onto magnetic tape using a machine available at the Lamont Observatory. The signals were played twice through a filter with a bandpass of 7 to 14 seconds in a manner such as to eliminate phase shifts and to attenuate the Rayleigh waves (Sutton and Pomeroy, 1963). The filtered records are shown at the bottom of figure 6. The time scales, but not the amplitudes, are the same on the original and filtered records. The filtered records show that the first shear mode is a continuous train of waves on both the vertical and east components for group velocities between 4.2 and 3.4 km/sec (0529 to 0533 GCT). For these group velocities the particle motion is retrograde,

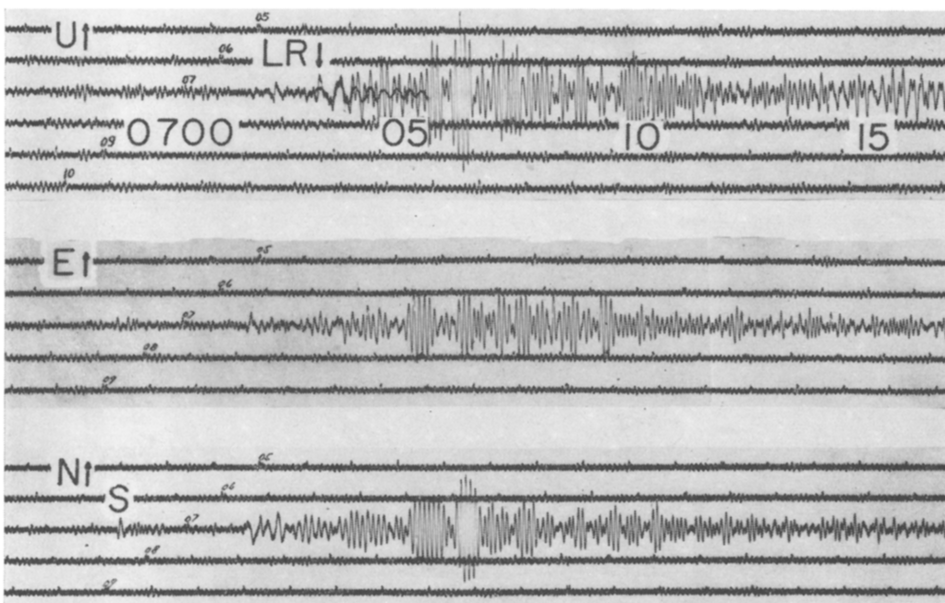

FIG. 7. Huancayo recordings of short-period surface waves for an earthquake on August 30, 1960, on the Easter Island ridge. Longer-period Rayleigh mode pencilled in on vertical component near 0705 GCT. Clock correction +1.0 sec.

and the ratio of horizontal-to-vertical motion is about 2. The later figure could not be determined exactly since calibration curves were not available for Hallett.

The ground motion of the short-period waves at Huancayo (figure 7) is approximately the same on the three components when the magnifications are taken into account. At 7 seconds the gains are as follows: Up 3500, East 2000, North, 1800 (Miller, 1963). Since the azimuth at Huancayo was nearly WSW, it was not possible to separate the first shear and first Love modes on the horizontal components. However, the ratio of horizontal-to-vertical motion in the first shear mode cannot be greater than unity even if all of the motion on the horizontals represents propagation in the first shear mode.

Of the three sets of records examined in this section, the ratio of horizontal-to-vertical motion in the first shear mode is largest on the Honolulu seismograms and is smallest on the records at Huancayo. The influence of the continental portion of the propagation path also becomes greater in the same order—Honolulu, Hallett, Huancayo. Propagation over continental paths may lead to a reduction in the ratio of horizontal-to-vertical motion. Experimental data and theoretical computations for continental paths show that, for periods greater than 5 seconds, the ratio of horizontal-to-vertical motion is always less than unity in the Rayleigh and first two shear modes (Oliver and Ewing, 1957; Oliver, Dorman and Sutton, 1959) if a prominent low-rigidity surface layer is not present (Lee, 1932, 1934).

Thus, in the absence of three-component recordings on the ocean bottom, the observations at Honolulu are probably the best indication of the motion in the first shear mode on the ocean bottom. The theoretical calculations in Part I for a model containing a low-rigidity layer indicate that the first shear mode is characterized by retrograde motion and by a ratio of horizontal-to-vertical motion greater than three at the water-sediment interface. These theoretical computations are in close agreement with the observations of the particle motion and the ratio of horizontal-to-vertical motion at Honolulu. The relationship of the particle motion on an island platform to that on the ocean bottom is not readily evident, and it is surprising that the observed and computed values are, in fact, quite similar.

Dispersion Data

Dispersion curves for the Rayleigh mode, the first Love mode, and the first shear mode were determined experimentally for the following paths: Tristan da Cunha to Buenos Aires, west of Mexico to Honolulu, Easter Island ridge to Huancayo, and West Chile rise to Hallett. These data are compared with theoretical dispersion curves so that additional information may be learned about the propagation of short-period waves. With this knowledge inferences can be made about the physical properties of the sediments, crust and upper mantle along the transmission paths. In particular, the shear velocity in the sediments can be deduced from the dispersion data when the thickness of the sediments is known. At the present time this is the only method that has been used to make a direct, *in situ* determination of the shear velocity in deep-sea sediments.

The epicenters of many of the earthquakes used in this study were determined by the author using a computer program similar to the one described by Bolt (1960). The remaining epicentral locations were taken from the *Seismological Bulletins* of the U. S. Coast and Geodetic Survey, from the *Bulletin Mensuel* of the Bureau Central International de Séismologic and from the *International Seismological Summary*. The accuracy of the locations made with a computer is about 10 to 20 km (Bolt, 1960; Gunst and Engdahl, 1962; Sykes, 1963). For the surface waves studied in this paper, the error in the computation of group velocity that arises from errors in epicentral location is very small, i.e., less than about .05 km/sec.

Results of calculations for case 8043 of Oliver, Dorman and Sutton (1959) were used to make a correction for the continental portion of the propagation path. These corrections were only applied to the Rayleigh mode and to the longer-period portion of the first Love mode. The correction for the continental segment of the path is not known for the short-period waves. However, since the short-period wave trains are nearly sinusoidal in character, such a correction is relatively unimportant for these waves. Since the continental portion of the path is less than 7% of the total distance in all cases with the exception of paths from Tristan da Cunha to

Buenos Aires and for one path to Huancayo, the choice of velocities for the continental correction is not critical. With the exception of the above mentioned paths, the corrections to the group velocities for the Rayleigh mode and for the first Love mode are less than .08 km/sec. A small correction was also made for the response of the instrument. Nevertheless, these corrections are both small and do not significantly affect the conclusions drawn from the dispersion data.

The dispersion calculations were performed using the program PV7 of Dorman (1962). The earth-flattening approximation of Alterman, Jarosch and Pekeris (1961) was used to correct the flat-layer computations for the effect of the sphericity of the earth. In the computations for the fundamental Rayleigh mode and for the various Love modes, the mantle structure is the same as that in case 8099 of Dorman, Ewing and Oliver (1960) and case 122 of Sykes, Landisman and Satô (1962). This theoretical structure of the mantle provides a good fit to mantle Love and Rayleigh wave data. A homogeneous mantle was used for the computations of the first shear mode since precision problems were encountered with program PV7. However variations in the properties of the mantle do not affect the calculations for the first shear mode appreciably for phase velocities less than 4.0 to 4.2 km/sec. The portions of the first shear mode of principal interest in this paper correspond to phase velocities less than 4 km/sec.

Tristan da Cunha to Buenos Aires

Surface waves with a predominant period of 12 to 20 seconds were present on all three components of the Buenos Aires seismograms for earthquakes on the Mid-Atlantic ridge near Tristan da Cunha on January 8, 1960, August 20, 1960, and November 29, 1961. Approximately 48% of the transmission path to Buenos Aires lies in the Argentine Basin, one of the largest deposits of sediments in the ocean basins (Ewing, 1963; Ewing, Ludwig and Ewing, 1963, 1964). For the paths across the Argentine Basin, the sedimentary thickness is greater and the predominant periods of the first Love and first shear modes are larger than those in the other areas investigated in this paper.

Rayleigh waves and longer-period Love waves were also recorded for the three events that generated the 12 to 20 second waves. Rayleigh waves and Love waves with periods greater than 18 seconds were the only surface waves observed at Buenos Aires for an earthquake on September 7, 1960, at 36.9°S, 15.8°W. Although this earthquake occurred about 150 km south and east of the three events that generated 12 to 20 second waves, the transmission paths to Buenos Aires for the four earthquakes are not known to differ in any important details. Thus differences in either the focal depth or the source spectrum are evidently responsible for the differences in the surface waves at Buenos Aires.

Dispersion data for these three earthquakes are presented in figure 8. For the short-period waves, which were detected only in the case of two of the shocks, no differences could be discerned in the dispersion curves for the three components of motion. These waves are interpreted as corresponding to the first shear and first Love modes. The absence of any detectible differences in the dispersion curves indicates that the *SH* and *SV* velocities are identical to within the precision of the measurements.

It is also evident from figure 8 that the frequency of the first shear mode is not an integral multiple of the frequency of the fundamental Rayleigh mode. Cleary and Peaslee (1962) interpreted the 6 to 8 second waves of Oliver and Dorman (1961) as an integral multiple of the frequency of the Rayleigh mode. Such an explanation cannot be used for the data for the paths from Tristan da Cunha to Buenos Aires nor for the other paths studied here.

The sediments in the vicinity of these great circle paths to Buenos Aires were surveyed extensively using the technique of the seismic reflection profiler, which was developed at the Lamont Geological Observatory by J. Ewing, M. Ewing and

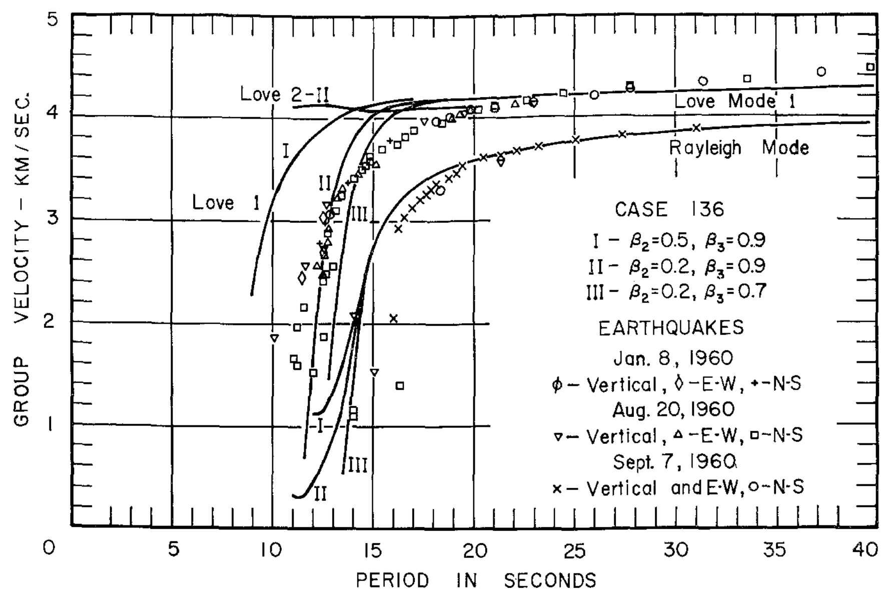

Fig. 8. Dispersion data for three paths from Mid-Atlantic ridge near Tristan da Cunha to Buenos Aires. Path crosses thick sediments of the Argentine Basin. Predominant period of first Love and first shear modes is 12 to 20 seconds. Theoretical dispersion curves for Rayleigh and Love modes were calculated using the model described in table 3. β_2 β_3 are shear velocities in two sedimentary layers.

others (Ewing and Tirey, 1961). The sediments of the Argentine Basin and the Argentine continental margin were described by Ewing (1963) and by Ewing, Ludwig and Ewing (1963, 1964). The average sedimentary thickness along the portion of the path that lies within the Argentine Basin is approximately 1.5 km. Between Tristan da Cunha and the Argentine Basin, the sedimentary thickness is approximately 0.2 km (Ewing, Ewing and Talwani, 1964).

Since the distribution of sediments is known for the paths to Buenos Aires, the average shear velocity in the sediments of the Argentine Basin can be deduced from the dispersion data for the first Love and first shear modes. As noted in Part I, the dispersion curves for these modes are sensitive to the ratio of thickness to shear velocity in the sediments. A knowledge of one of these parameters is necessary before the other can be determined from the dispersion data.

The great circle paths from the epicenters to Buenos Aires were divided into three segments in the theoretical model. Theoretical dispersion curves (case 136) were derived for the entire path from the dispersion curves for the three segments. The theoretical model and the dispersion results are given in table 3. The continental case 8043 of Oliver, Dorman and Sutton (1959) was used for the segment (14.6% of total path) from Buenos Aires to the 1000 fathom contour at 35.9°S, 52.5°W. For the path across the Argentine Basin to 37.1°S, 32.0°W (47.5% of total path), an average sedimentary thickness of 1.5 km and an average water depth of 4.7 km were obtained from the study of Ewing, Ludwig and Ewing (1964). For the eastern portion the path (37.9% of total path), an average sedimentary thickness of 0.2 km and an average water depth of 3.8 km was determined from the records made during *Vema* Cruise 18 (Ewing, Ewing and Talwani, 1964) and cruise II of the Argentine research vessel *Zapiola*. The sounding depths were corrected using the tables of Matthews (1939). Although refraction data were not available along these reflection profiles, refraction measurements were made further south in the Argentine Basin (Ewing, Ludwig and Ewing, 1964). These results were used in the theoretical models and are indicated by asterisks in table 3.

The properties of the thinner sediments along the eastern portion of the path do not affect the calculations noticeably for periods greater than 10 seconds. The correction of the theoretical dispersion curves for the effect of the continental path is probably a close approximation for the case of the Rayleigh mode. However, a low-rigidity layer was not included in the computations for continental case 8043. Thus additional dispersion may occur along the continental portion of the path for the first Love and shear modes if a low-rigidity layer is present. However, this should not cause large errors in the estimate of the shear velocity in the sediments since the continental portion of the path is only 14.6% of the total distance.

In the theoretical computations the shear velocity in the thick sediments was varied (curves I, II and III of figure 8) to fit the dispersion data for the first Love and first shear modes. For simplicity only the Love mode calculations are shown in figure 8. It is interesting to note that in the theoretical computations most of the dispersion for Love waves takes place in the region of thick sediments and not along the other two portions of the path. Several interfaces within the sediments were detected on the reflection records (Ewing, Ludwig and Ewing, 1964). However, in the theoretical computations the average sedimentary column was divided into two layers (layers 2 and 3 in table 3) 0.5 and 1 km thick, respectively.

The use of two plane homogeneous layers for the sediments of the Argentine Basin is admittedly a simplification that permits theoretical computations to be performed. A somewhat better approximation might be that of a wedge of sediments that gradually thickens to the west. However, in view of the approximations already made, such as the division of the path into three homogeneous segments, the use of a more complicated model for the sediments does not appear justified. Thus the shear velocities derived using the simple layered models represent only average values at best. Nonetheless, such calculations are useful since *in situ* values of shear velocity cannot be obtained presently in any other manner.

The first Love mode was originally computed (curve I) using shear velocities comparable to those estimated by Nafe and Drake (1957, Oliver and Dorman, 1961) for deep-sea sediments. These velocities are too high to fit the dispersion data.

TABLE 3

Layer Parameters				
N	THKNS	ALPHA	BETA	RHO
Case 124—No Sediment				
1	4.00	1.52	0.00	1.03
2	1.34	5.09	2.94	2.54
3	4.62	6.76	3.90	2.90
4	8099 mantle			
Case 124—0.6 km Sediment				
1	4.00	1.52	0.00	1.03
2	0.60	2.00	BETA2	1.90
3	1.34	5.09	2.94	2.54
4	4.62	6.76	3.90	2.90
5	8099 mantle			
Case 135				
1	4.50	1.52	0.00	1.03
2	0.20*	1.70	BETA2	1.80
3	1.54*	5.37*	3.10	2.60
4	5.08*	6.80*	3.70	2.90
5	8099 mantle			

Rayleigh Mode—BETA2 = 0.1			
T	*C*	*U*	*R*
40.00	4.03794	4.03956	0.758907
35.00	4.03873	4.04700	0.755689
30.00	4.03952	4.03824	0.742140
28.00	4.03906	4.02599	0.731523
26.00	4.03744	4.00582	0.716132
24.00	4.03385	3.97505	0.693865
22.00	4.02683	3.92548	0.661432
20.00	4.01235	3.85225	0.613590
19.00	4.00248	3.79395	0.580860
18.00	3.98871	3.71601	0.540045
17.00	3.96914	3.60734	0.488499
16.00	3.94033	3.44663	0.422390
15.00	3.89521	3.18989	0.336008
14.50	3.86199	2.99938	0.282675
14.00	3.81670	2.74427	0.220742
13.50	3.75201	2.40214	0.148554
13.00	3.65460	1.96244	0.064373
12.50	3.49824	1.48834	−0.032855
12.00	3.27080	1.11568	−0.144436
11.50	2.98488	0.91667	−0.274280
11.00	2.70158	0.85213	−0.438332
10.50	2.45821	0.84905	−0.669850
10.00	2.25563	0.85192	−1.034978
9.50	2.07949	0.83082	−1.696428
9.00	1.90182	0.67697	−3.215910
8.50	1.60163	0.29060	−9.848966

Rayleigh Mode—BETA2 = 0.2			
T	*C*	*U*	*R*
12.50	3.50202	1.51465	0.089010
12.00	3.28228	1.15171	0.047039
11.50	3.00849	0.95785	0.013334
11.00	2.73792	0.90060	−0.013474
10.50	2.50756	0.91487	−0.037837
10.00	2.32159	0.95448	−0.063682
9.00	2.05137	1.04732	−0.132619
8.00	1.86877	1.13080	−0.252846
7.00	1.73754	1.19243	−0.498877
6.00	1.63509	1.21321	−1.129939
5.50	1.58710	1.19351	−1.910826
5.00	1.53097	1.07692	−3.866971
4.50	1.42200	0.70832	−13.295100
First Shear Mode—BETA2 = 0.20			
8.80	4.16009	3.51599	−0.748096
8.40	4.12675	3.54317	−0.877028
8.00	4.09476	3.54313	−1.019142
7.60	4.06245	3.51887	−1.175509
7.20	4.02813	3.47103	−1.347303
6.80	3.98993	3.39825	−1.535951
6.40	3.94552	3.29743	−1.743539
6.00	3.89180	3.16404	−1.974024
5.60	3.82449	2.99262	−2.236952
5.20	3.73751	2.77652	−2.560474
4.80	3.62150	2.50012	−3.056442
4.60	3.54731	2.32040	−3.527936
4.40	3.45491	2.07449	−4.698341
4.20	3.30216	1.29643	−21.333449
First Love Mode—BETA2 = 0.2			
20.00	4.44458	4.39005	
18.00	4.43883	4.38531	
16.00	4.43247	4.37910	
14.00	4.42516	4.36992	
13.00	4.42090	4.36193	
12.00	4.41572	4.34473	
11.00	4.40759	4.28179	
10.80	4.40496		
10.60	4.40166		
10.00	4.38575		
8.40	4.29971	3.77458	
8.00	4.26624	3.70989	
7.60	4.23223	3.64157	
7.20	4.19377	3.56748	
6.80	4.15017	3.48657	
6.40	4.10048	3.39649	
6.00	4.04334	3.29165	
5.60	3.97620	3.15792	

TABLE 3—CONTINUED

First Love Mode—BETA2 = 0.2—*Continued*

T	*C*	*U*	*R*
5.20	3.89317	2.95556	
4.80	3.77560	2.55155	
4.70	3.73424	2.37609	
4.60	3.68321	2.14599	
4.50	3.61645	1.84253	
4.40	3.52151	1.45023	
4.30	3.36912	0.96816	
4.20	3.07628	0.47200	
4.10	2.38626	0.12909	
Second Love Mode—BETA2 = 0.2			
20.00	4.60029	4.31370	
18.00	4.56996	4.31372	
16.00	4.53965	4.30641	
14.00	4.50752	4.27925	
13.00	4.48932	4.25005	
12.00	4.46815	4.20671	
11.40	4.45353	4.18116	
11.20	4.44839	4.17652	
11.00	4.44322	4.17640	
10.80	4.43812	4.18306	
10.60	4.43326	4.19842	
10.40	4.42882	4.22233	
10.00	4.42172	4.27920	
9.50	4.41580	4.32674	
9.00	4.41167	4.34724	
8.00	4.40532	4.35941	
7.00	4.39970	4.36131	

Case 136

Layer Parameters—Segment A

35.65S, 15.75W to 37.1S, 32.0W

0.3790 of total distance

N	THKNS	ALPHA	BETA	RHO
1	3.80	1.52	0.00	1.03
2	0.20	1.80	0.10	1.85
3	1.04*	4.74*	2.70	2.54
4	3.94*	6.46*	3.64*	2.80
5	8099 mantle			

Layer Parameters—Segment B

37.1S, 32.0W to 35.9S, 52.5W

0.4754 of total distance

1	4.70	1.52	0.00	1.03
2	0.50	1.80	BETA2	1.85
3	1.00	2.20	BETA3	2.10
4	1.04*	4.74*	2.70	2.54
5	3.94*	6.46*	3.64*	2.80
6	8099 mantle			

TABLE 3—CONTINUED

Layer Parameters—Segment C

(Case 8043)

35.9S, 52.5W to Buenos Aires (34.61S, 58.48W)

0.1455 of total distance

N	THKNS	ALPHA	BETA	RHO
1	2.105	3.98	2.30	2.340
2	17.895	6.15	3.55	2.817
3	20.000	6.58	3.80	2.922
4	inf	8.14	4.70	3.300

UI = group velocity of Segment I

U = group velocity for total path

$1/U = 0.3790/UA + 0.4754/UB + 0.1455/UC$

Rayleigh Mode

Period T	Curve I BETA2 = 0.5 BETA3 = 0.9 U	Curve II BETA2 = 0.2 BETA3 = 0.9 U	Curve III BETA2 = 0.2 BETA3 = 0.7 U
40.0		3.944	
35.0		3.913	
30.0		3.857	
25.0		3.758	
24.0		3.732	
23.0		3.703	
22.0		3.671	
21.0		3.633	
20.0		3.588	
19.5		3.561	
19.0		3.530	
18.5		3.493	
18.0		3.450	
17.5		3.398	
17.0		3.332	
16.5		3.248	
16.0		3.134	
15.5		2.973	
15.0		2.732	2.752
14.5		2.347	2.288
14.0	1.974	1.768	1.463
13.5	1.578	1.201	0.549
13.0	1.333	0.842	
12.5	1.239	0.623	
12.0	1.224	0.442	
11.5		0.306	
11.0		0.333	

First Love Mode

40.0		4.279	
35.0		4.258	
30.0		4.236	
25.0		4.210	

TABLE 3—CONTINUED

First Love Mode

Period T	Curve I BETA2 = 0.5 BETA3 = 0.9 U	Curve II BETA2 = 0.2 BETA3 = 0.9 U	Curve III BETA2 = 0.2 BETA3 = 0.7 U
20.0		4.191	4.185
19.0		4.185	4.176
18.0		4.174	4.158
17.0	4.176	4.156	4.122
16.0	4.163	4.118	4.037
15.0	4.140	4.023	3.811
14.0	4.087	3.784	3.284
13.4		3.518	2.614
13.0	3.974	3.242	1.879
12.4		2.500	0.567
12.0	3.812	1.646	
11.8		1.126	
11.4		0.322	
11.0	3.589		
10.0	3.204		
9.6	2.928		
9.0	2.324		

Second Love Mode

(Assuming travel in Second Love in 2 oceanic segments and as First Love Mode in Continental Segment)

20.0		4.123	4.114
19.0		4.117	4.105
18.0		4.105	4.089
17.0		4.089	4.069
16.0		4.069	4.053
15.0		4.062	4.073
14.0		4.095	4.119
13.0		4.129	4.140
12.0		4.133	4.138
11.0		4.107	

N = layer number. THKNS = layer thickness in km. ALPHA = layer compression velocity in km/sec. BETA = layer shear velocity in km/sec. RHO = layer density. T = wave period. C = phase velocity. U = group velocity. R = ratio of horizontal particle amplitude in the solid medium at water-solid interface to vertical particle amplitude at water surface (Rayleigh and first shear modes only). $H1$ = water depth in km. BETA2 = shear velocity in sediments in km/sec. 8099 mantle—Indicates mantle structure of case 8099 (Dorman, Ewing and Oliver, 1960) used for Rayleigh mode and Love modes. Asterisks indicate layer parameters taken from refraction results. The earth-flattening approximation of Alterman, Jarosch and Pekeris (1961) was applied in all of the calculations in Table 3.

In a later section it will be shown from other evidence that the shear velocities estimated by Nafe and Drake are probably too high. Curves II and III fit the data better than curve I. A single layer with a gradient in shear velocity could also be used to fit the data. Nevertheless, it is evident that a low-rigidity layer is necessary to account for the large amount of dispersion in the first Love and first shear modes.

The theoretical group velocity curves I, II and III (figure 8) are consistently above the data in the period range 15 to 20 seconds. This discrepancy could be reduced in a number of ways. The contrast in shear velocity that was used for curve III could be reduced. Shear velocities $\beta_2 = 0.3$ to 0.4 and $\beta_3 = 0.5$ to 0.6 would fit the data better than the velocities used for curve III. On the flank of the Mid-Atlantic ridge between Tristan da Cunha and the Argentine Basin, the shear velocities in the crust and upper mantle are probably somewhat lower than those assumed in the theoretical model. The velocities used in the model are more appropriate to ocean basins than they are to portions of the Mid-Atlantic ridge. This

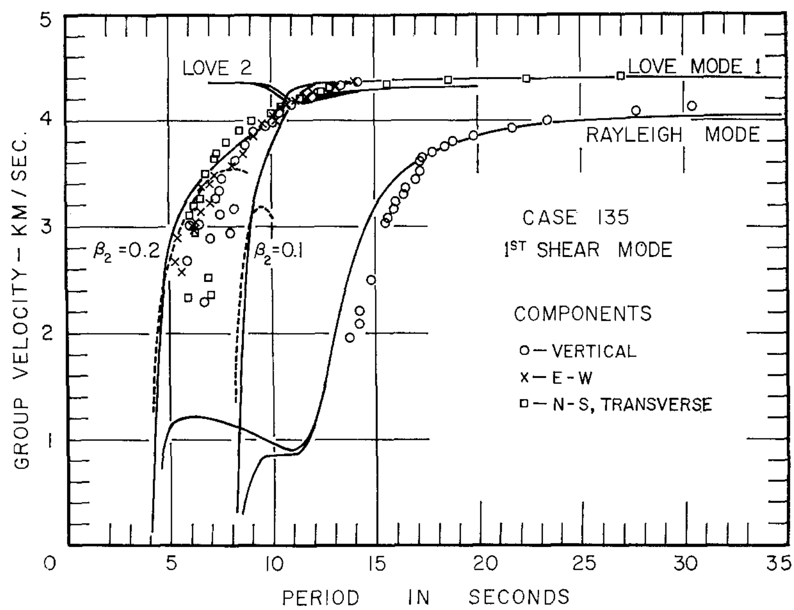

FIG. 9. Dispersion data for path from west of Mexico to Honolulu, July 28, 1961. Theoretical computations performed for a sedimentary thickness of 0.2 km and shear velocities of 0.1 and 0.2 km/sec. A water depth of 4.5 km used in computations. Thicknesses of sedimentary, basement and crustal layers are an average for Raitt's (1956) refraction stations M1, M2 and M9. The group velocity curves for the first and second Love modes cross near 11 seconds. The corresponding phase velocity curves, however, do not cross.

could also account for part or all of the discrepancy between the data and the theoretical computations for curves II and III.

Short-period surface waves with predominant periods between 10 and 20 seconds are sometimes recorded at Rio de Janeiro and Buenos Aires from earthquakes in the South Sandwich arc. Since the arrival at Buenos Aires is from the southeast, it is difficult to separate these wave trains from the somewhat longer-period waves of the Rayleigh mode. However, the existence of the 10 to 20 second waves for these paths, which also cross the thick sediments of the Argentine Basin, is additional evidence that the dispersion of the short-period waves is highly dependent on the properties of the sedimentary layer.

West of Mexico to Honolulu

Dispersion data for the earthquake on July 28, 1961, are presented in figure 9. Seismograms for this event were shown in figure 1 and were discussed in earlier sections. For a given period the group velocities are somewhat higher for the transverse component than for the other two components. This situation was not encountered for any of the other data analyzed in this paper. The discrepancy between the group velocities may indicate that the shear velocities are not completely isotropic along this path. However, Love waves may be propagated with faster group velocities (figure 10) if sediments are absent along part of the path. A knowledge of the sedimentary thickness along the entire path would help to resolve this ambiguity. However, the observed dispersion curves themselves and not the

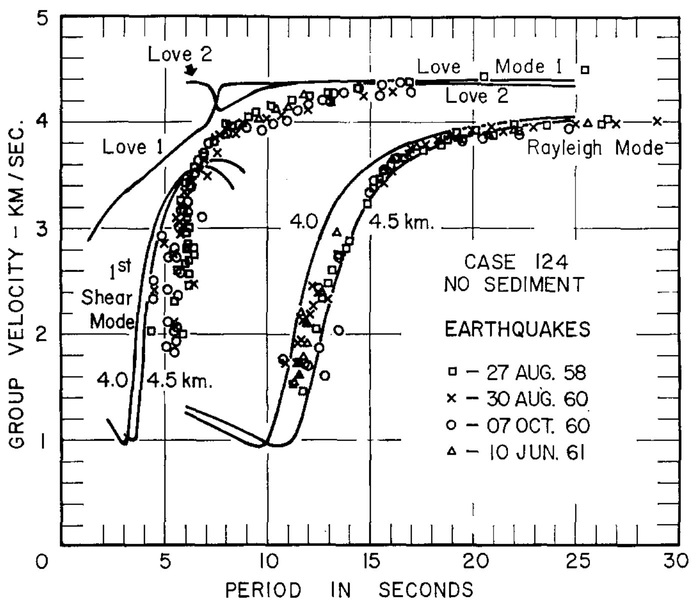

FIG. 10. Dispersion data for four paths from the Easter Island ridge to Huancayo, Peru. Computations for Rayleigh, first Love, second Love, and first shear modes of case 124. A low-rigidity sedimentary layer was not included in these calculations. Water depths of 4.0 and 4.5 km used in computations. Small correction made to long-period data for effect of continental portion of path.

small differences among these curves are the most prominent features which must be explained.

Calculations for the first two Love modes, the first shear mode and the Rayleigh mode are shown in figure 9 for case 135, a model containing a low-rigidity layer. The distribution of sediments along this path is not as well known as it was in the Argentine Basin. However, data from three refraction stations were available in the vicinity of the propagation paths. The sedimentary thickness and the thicknesses and compressional velocities of the crustal layers used in the theoretical calculations are an average of the corresponding parameters for Raitt's refraction stations M1, M2 and M9. The average water depth was estimated as 4.5 km from the Monaco bathymetric charts. Menard's (1960b) more recent bathymetric map and the data for the Rayleigh mode indicate a water depth somewhat greater than 4.5 km for this path. The layer parameters and dispersion results are presented in table 3.

The computations for the various dispersion curves were performed for sedimentary shear velocities of 0.1 and 0.2 km/sec. The shear velocity of 0.2 km/sec fits the data better. However, if sediments are present along only half of the path, a shear velocity between 0.1 and 0.2 km/sec would also fit the data. Thus sediments along at least a portion of the path are necessary to fit the data. It is interesting to note that the values of shear velocity estimated for this path are approximately the same as the ones obtained for the upper layer of sediments in the Argentine Basin. This indicates that the variation in the sedimentary thickness is the principal factor controlling the periods of the short-period waves. The similar values of shear velocity also indicate that the mechanism of the wave propagation is substantially the same in the two areas.

The theoretical curves for the first shear mode of case 135 (figure 9) are nearly coincident with the curves for the first Love mode for group velocities less than 3 to 3.5 km/sec. However, the shear modes, unlike the first Love mode, exhibit group velocity maxima between 3 and 4 km/sec. The positions of these maxima can be changed considerably by making small changes in the layer parameters. A homogeneous mantle was also assumed in the calculations for the first shear modes since precision problems were encountered with the program PV7 when a more complicated mantle structure was introduced. Variations in the properties of the mantle do not affect the calculations for the first shear mode appreciably for phase velocities less than about 4.0 to 4.2 km/sec (group velocities less than 3 to 3.5 km/sec). When mantle structure is taken into account in the theoretical models, it does not seem unlikely that dispersion curves can be found for the first shear mode that will fit the data more closely over a wider range of group velocities.

Easter Island Ridge to Huancayo, Peru

Dispersion data are presented in figures 10 and 11 for four paths from the Easter Island ridge to Huancayo. The propagation paths are illustrated in figure 2, and the records for one of these events are shown in figure 7. Data for the Rayleigh mode and for the short-period surface waves were available over a wide range of periods and group velocities. The scatter in the dispersion data is relatively small compared to that for some of the other events analyzed in this paper. No differences could be detected in the dispersion curves for the three components. Thus no evidence of anisotropy was found.

The average water depth along the four great circle paths from the epicenters was estimated as 4.1 to 4.2 km using the Monaco bathymetric charts. An average water depth of 3.9 km (corrected) was determined for a profile between Callao, Peru, and the oceanic ridge near 17°S, 113°W, which was made during *Vema* cruise 19. The thicknesses and compressional velocities for the crustal layers were taken from the study of Raitt (1956) and represent average values for the Pacific exclusive of Melanesia. Shear velocities were calculated using a Poisson's ratio of 0.25. Along the great circle paths to Huancayo only one refraction station was available that was located west of the Peru-Chile Trench. The results for this station (number 23 of Fisher and Raitt, 1962) are very close to the average for the Pacific. A sedimentary thickness of 0.3 km was reported for this station.

Only very small amounts of sediments were detected with the reflection profiler along the path from Callao, Peru, to the Easter Island ridge at 17°S, 113°W (M. Ewing, personal communication). Hence, an attempt was first made to fit all of the dispersion data without the use of a low-rigidity sedimentary layer. Theoretical dispersion curves are shown in figure 10 for the Rayleigh, first Love and first shear modes of case 124 for water depths of 4.0 and 4.5 km. A low-rigidity layer was not used. Estimates of the total liquid thickness from Rayleigh waves vary from 4.0 to 4.5 km (figure 10). These results are thus in accord with a sedimentary thickness of 0 to 0.5 km although somewhat greater thicknesses are possible when shear velocities greater than about 0.5 km/sec are introduced in the sediments. Kovach and Press (1961) also used Rayleigh mode data and interpreted the average sedimentary thickness as 0.57 km for other paths from the Easter Island ridge to Huancayo.

The periods of the short-period waves for the paths used in figure 10 are shorter than the periods of any of the other data studied in this paper. The computations of the first shear mode for the model without a low-rigidity layer are just within the range of the short-period data. Thus it is possible to explain the data for both the Rayleigh and short-period waves without introducing a low-rigidity layer into the computations if the short-period waves propagate in only the first shear mode. Such an explanation is not possible for the other sets of data in this paper since the dispersion data do not coincide with the computations for the first shear mode.

There are several indications that the short-period waves are propagated in both the first Love and first shear modes. Waves with group velocities between 4.4 and 4.0 km/sec on the horizontal components (figure 7) can be identified as Love waves from their particle motion. For the earthquake shown in figure 7, the azimuth of the great circle path at Huancayo was approximately WSW. The short-period waves are larger on the north-south component, especially in the earlier parts of the wave train (group velocities between 4.4 and 2.5 km/sec). These observations indicate that some of the short-period disturbance is propagated as Love waves. Large Love waves were even more apparent for earthquakes near 9°S, 110°W (November 10, 1958) and 13.5°S, 111.5°W (September 20, 1959), which produced large short-period waves on the north-south component at Huancayo. The azimuth of these arrivals at Huancayo was nearly due west. The small amount of scatter in the dispersion data, the presence of large amplitudes on the north-south component early in the surface wave train, and the fact that the great circle paths are nearly perpendicular to important geological boundaries all indicate that reflection and refraction of these waves are not important for group velocities greater than 3 km/sec. The increased scatter in the data for group velocities less than 3 km/sec may perhaps be attributed to some or all of the following factors: refraction and reflection of the waves, differences in the physical properties of the medium, or errors in the determinations of the periods of the waves. Nevertheless, it must be concluded that the short-period waves are propagated, at least in part, as Love waves and that the data in figure 10 do not fit the dispersion curves of Love waves for the oceanic structure used in case 124, a model without a low-rigidity layer. Thus some change must be made in the theoretical model to satisfy the observed dispersion of the short-period Love waves.

For the curves of figure 11, computations were performed for the Rayleigh and first Love modes assuming a sedimentary thickness of 0.6 km. The data fit the Rayleigh mode computations for an average water depth of 4.1 to 4.2 km. Sedimentary shear velocities of 0.5 and 1.2 km/sec bracket the short-period data. Dispersion curves for the first shear mode are nearly coincident with those for the first Love mode. Thus all of the surface wave data may be satisfied by introducing a low-rigidity layer into the calculations. If the sedimentary thickness is taken as 0.3 km shear velocities of approximately 0.6 and 0.25 km/sec would also satisfy the surface wave data.

If a low-rigidity layer is not included in the theoretical model, the dispersion

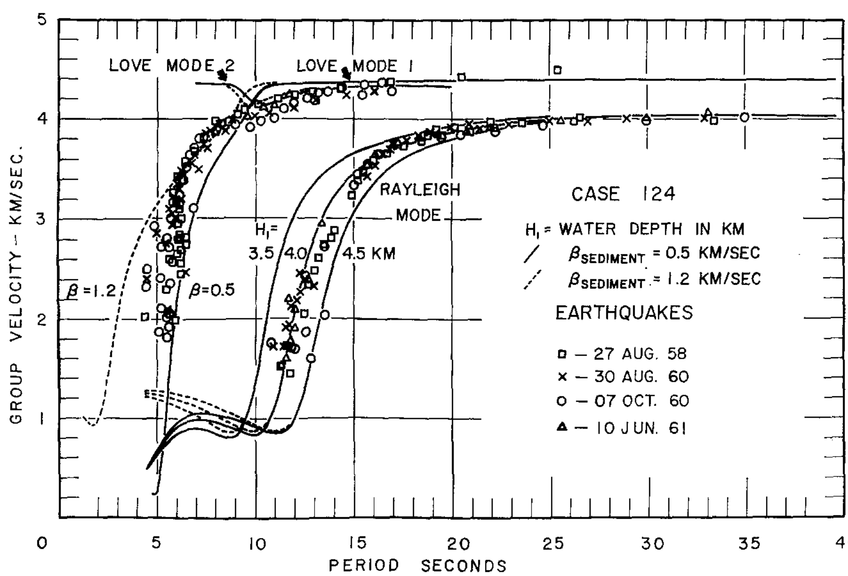

FIG. 11. Dispersion data for four paths from the Easter Island ridge to Huancayo, Peru. Theoretical dispersion curves for Rayleigh mode and first two Love modes of case 124. Average water depth along paths is 4.1 km. Thickness of sedimentary layer taken as 0.6 km. Calculations performed for sedimentary shear velocities of 0.5 and 1.2 km/sec. Small correction made to long-period data for effect of continental portion of path.

curve for the first Love mode may be varied to satisfy the short-period data by changing either the shear velocity or the thickness of the uppermost crustal layer. In such a model the predominant periods of the first shear mode are controlled mainly by the properties of the water layer while the predominant periods of the first Love mode are governed principally by the properties of the uppermost crustal layer. Although the Huancayo data might be satisfied by the calculations for a model that does not contain a low-rigidity layer, the coincidence of the dispersion curves for such a model is fortuitous. It does not seem probable that such a model can account for the numerous other instances in which the dispersion curves are coincident or are nearly coincident.

A model that contains a low-rigidity sedimentary layer is not in agreement with the reflection results for the profile from Callao, Peru, to the Easter Island ridge near 17°S, 113°W. Only very small amounts of acoustically transparent sediments were detected along this path (M. Ewing, personal communication). There are no reports of earthquakes on the Easter Island ridge between 20°S and 13.5°S (Sykes, 1963). Hence, there is no surface wave data along the reflection profile. Whether the reflection profile is typical of the entire area between the Easter Island ridge and the coast of Peru is not known. Sediments of greater acoustic opacity may not have been detected by the reflection profiler. Heezen and Tharp (1961) indicate areas of low relief to the west of the Peru-Chile trench. The relief is less than that normally observed in areas of abyssal hills or on the flanks of the mid-oceanic ridges, but is greater than that associated with modern abyssal plains. Further research

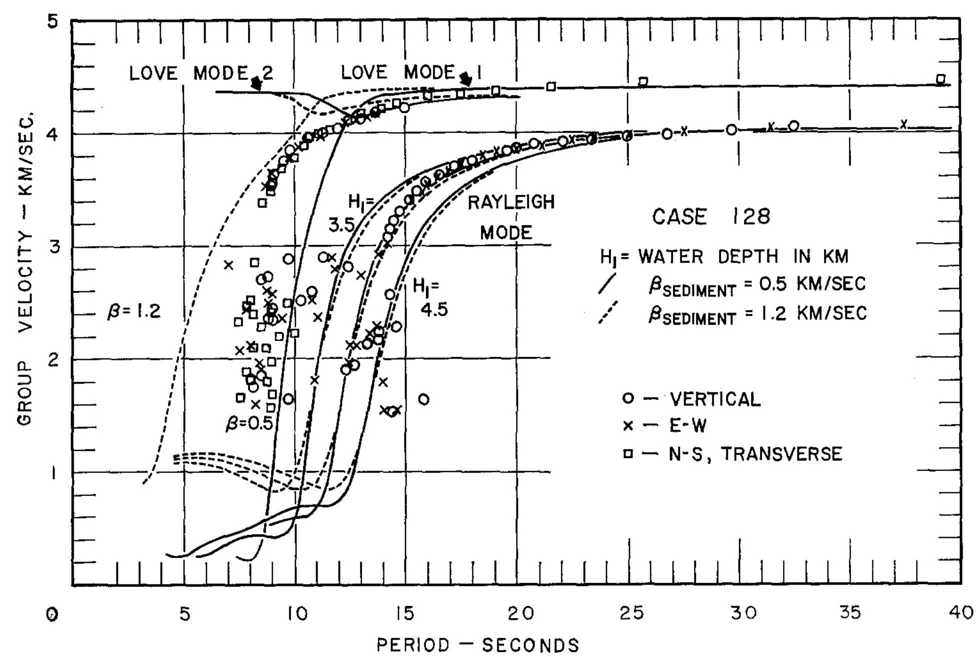

FIG. 12. Dispersion data for paths from West Chile rise to Hallett, Antarctica, November 14, 1958, and December 17, 1959. Sediment thickness taken as 1.0 km. Small correction made to long-period data for effect of continental portion of path.

should help to clarify the nature of the crust and sediments between Huancayo and the Easter Island ridge and to resolve the apparent discrepancy between the surface wave data and the results of reflection profiling.

West Chile Rise to Hallett, Antarctica

Dispersion data for two earthquakes on the West Chile rise are presented in figure 12. Seismograms for the event on November 14, 1958, are shown in figure 6. The dispersion data exhibit very little scatter for group velocities between 4.4 and 3 km/sec. The particle motion in the Rayleigh and first shear modes is very regular. The general simplicity of the seismograms and the dispersion data argues in favor of normal mode propagation in these portions of the record.

For group velocities less than 3 km/sec (times greater than 0534 in figure 6), the data for all of the waves scatter very badly. Rayleigh waves with periods of 16 to 20 seconds, which normally propagate with group velocities greater than 3 km/sec, are present in the later portions of the record and apparently travelled distances in excess of the great circle distance to Hallett. Reflected and refracted arrivals of this type are possible since the great circle path near Hallett is not perpendicular either to the coast of Antarctica or to the mid-oceanic ridge. Hence, caution must be used in interpreting the waves in the later portions of the records.

Bathymetric data and refraction stations are very sparse in the vicinity of the great circle paths to Hallett. Heezen and Laughton (1963) indicate an abyssal plain of uncertain extent in the Bellinghausen Basin. The profiles published by Zhivago (1962) also show large regions of very flat topography south of 60°S. These reports indicate that sediments are probably present along at least a portion of the great circle paths to Hallett.

The water depth was estimated as 4.0 km from the Monaco bathymetric charts, but may be in error by 0.5 km or more as a result of the sparse number of soundings in the southeast Pacific. Hence, any estimates of sedimentary thickness that are made using the Rayleigh mode are subject to large uncertainties.

Computations are illustrated in figure 12 for a model (case 128) with a sedimentary thickness of 1.0 km. Shear velocities of 0.5 and 1.2 km/sec were used to bracket the data. The parameters of the layers below the sediments in case 128 are identical to those in case 124. As well as satisfying the short-period data, the computations also fit the data for the Rayleigh mode and the estimated water depth. Reflection data and additional bathymetric data are needed to make more extensive use of the surface wave data for these paths.

Summary of Dispersion Data

The predominant periods of short-period surface waves vary considerably for the various transmission paths discussed in figures 8 through 12. At group velocities of 3 km/sec the predominant periods of the waves are as follows: Argentine Basin, 12.5 to 13 sec; Alaska to Honolulu (Oliver and Dorman, 1961), 8 to 9 sec; West Chile rise to Hallett, 8 to 8.5 sec; west of Mexico to Honolulu, 6 to 6.5 sec; and Easter Island ridge to Huancayo, 5.5 to 6.5 sec. The predominant periods do not correlate with the average water depth along the various transmission paths. Although data on the sedimentary thicknesses are incomplete for many of the paths, it can be observed that the waves of longest periods are associated with the thick sediments of the Argentine Basin. The waves of shorter periods are usually found in areas where the average sedimentary thicknesses are a few tenths of a kilometer. These results indicate that the sedimentary thickness plays a dominant role in the propagation of the short-period surface waves.

Shear Velocity Measurements for Low-Rigidity Materials

No *in situ* determinations of shear velocity are available for deep-sea marine sediments. Laughton (1957) measured the compressional and shear velocities of artificially compacted *Globigerina* ooze. The compressional velocities determined by Laughton, however, are typical of velocities at depths of several kilometers below the ocean bottom (Nafe and Drake, 1963). Hence, the shear velocities in these materials (table 2) are probably considerably greater than those in the upper one or two kilometers of sediments, the depths of principal interest here.

Other measurements of the velocities of various low-rigidity materials are summarized in table 2. Some of these determinations were tabulated previously by Nafe and Drake (1963). The measurements of White and Sengbush and McDonal and others (1958) pertain to semi-consolidated, water-saturated materials that are found below the low-velocity layer (weathered layer). The compressional and shear velocities of these materials are similar to those estimated for the deeper sedimentary layer (layer 3) in the Argentine Basin. Even lower values of shear and compressional velocity are often found on land in the weather layer (White and Sengbush, 1953; Jolly, 1956).

Hamilton and others (1956) made *in situ* determinations of compressional velocity in shallow water marine sediments. Values of rigidity were computed by attributing the discrepancy between the *in situ* determinations and the emersion (Wood) equation as resulting entirely from rigidity. These rigidities were used to compute upper limits to the shear velocity (table 2). The range of these shear velocities (0.0 to 0.5 km/sec) includes the velocities obtained for the upper layer of sediments in the Argentine Basin and the shear velocities estimated from the other surface wave data.

Arrivals with velocities of a few tenths and occasionally a few hundredths of a kilometer per second are sometimes observed in shallow-water seismic surveys. These waves are known as granddaddy waves because of their slow velocities and low frequencies. Apparently the velocities are controlled largely by the low shear velocity in the ocean bottom sediments although the effect of gravity may not be negligible for these waves (Worzel and Ewing, 1948, p. 23; Ewing, Jardetzky and Press, 1957, p. 203; Oliver and Isacks, 1962). The measured velocities for these waves are very similar to the estimates of the shear velocity made in this paper. Thus neither these observations nor the measurements on low-rigidity materials are in conflict with the shear velocities estimated from surface wave data. *In situ* measurements of shear velocity are urgently needed to check the surface wave results.

Nafe and Drake (1957) estimated the distribution of shear velocity as a function of depth in deep-sea sediments from experimental determinations of compressional velocity as a function of depth. These values of shear velocity are higher than those obtained from the surface wave data for the Argentine Basin. There are two other indications that the shear velocities of Nafe and Drake are too large in the upper one or two kilometers of sediments. Houtz and Ewing (1963) found a compressional velocity-depth function that was several tenths of a kilometer per second lower than that given by Nafe and Drake (1957). The compressional velocities of Nafe and Drake were compiled before velocity determinations at the water-sediment interface had been made from studies of the phase change of reflected waves. Hence, the estimates of shear velocity as a function of depth must be reduced since these are computed from measurements of compressional velocity as a function of depth. Nafe and Drake's (1957) estimates of Poisson's ratio at a given porosity are lower than the values reported by Laughton (1957) and others (Nafe and Drake, 1963, p. 808). The higher values of Poisson's ratio (lower shear velocities) proposed by

Nafe and Drake in 1963 are in close agreement with the surface wave results reported in this paper.

First Shear and First Love Modes for Periods Greater Than 10 Seconds

Waves with periods between 10 and 30 seconds are sometimes observed on vertical component seismographs prior to the arrival of the fundamental Rayleigh mode. The most common arrival is a train of normally dispersed waves that begins with a period of approximately 14 seconds and with group velocities between 4.2 and 4.4

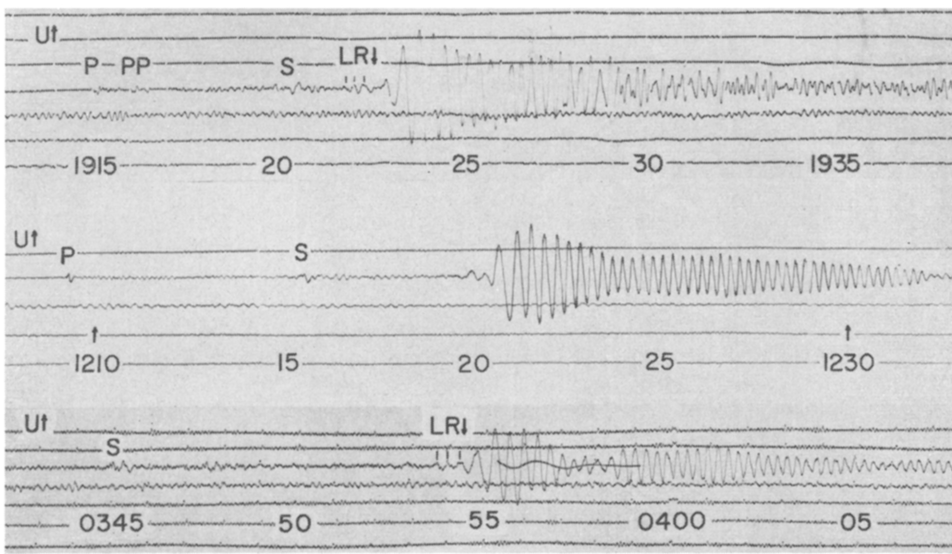

Fig. 13. Seismograms showing 15 to 30 second forerunner to Rayleigh mode recorded on vertical component instruments (15 sec seismometer, 75 sec galvanometer). Top—Indian Ocean to Perth, Australia, March 7, 1961, distance 31.3°. Center—Indian Ocean to Pietermaritzburg, South Africa, July 11, 1959, distance 40.6°. Bottom—Easter Island to Suva, Fiji, April 15, 1960, distance 62.9°. Clock corrections are 0.0, +11, and 3 sec, top to bottom.

km/sec. When the 5 to 10 second waves are absent or small, waves with apparent periods of 15 to 30 seconds are occasionally recorded by the vertical component instrument. Some of these arrivals are illustrated in figure 13. Up to 4 or 5 cycles of these waves are observed at distances ranging from 30 to 85 degrees. The apparent periods of these arrivals are shown in figure 14 along with some typical data for the normally dispersed portion of the first shear mode. For periods greater than 14 seconds, the observed wave form is more pulse-like in character than the waves of shorter periods, and the measurements of period and velocity are only an approximation to the actual dispersion of the waves.

The 15 to 30 second waves shown in figure 13 evidently correspond to higher modes and are not part of the Airy phase pulse that is associated with the Rayleigh mode. Several lines of evidence support this conclusion. Many seismograms of Rayleigh waves do not show this 15 to 30 second arrival. Synthetic seismograms for the oceanic Rayleigh mode (Kovach and Press, 1961) do not exhibit the type of forerunner illustrated in figure 13. The period of most of the waves is too short to be related to the oscillations associated with the Airy phase (Sutton, Major and Ewing, 1959) of this mode.

These arrivals do not agree with the travel times of any single body phase such as *SS* or *SSS* over the observed distance range 30 to 85°. Kovach and Anderson (1964) computed dispersion curves for several of the higher Rayleigh modes and

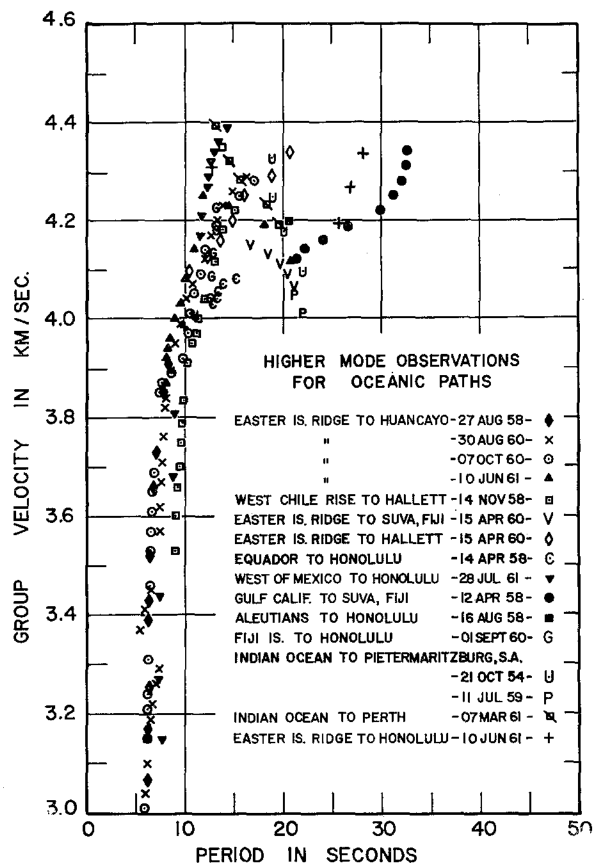

FIG. 14. Summary of higher mode observations for oceanic paths. All observations made from vertical component recordings. The 12 to 20 second waves recorded at Buenos Aires are not included.

showed that the group velocities corresponding to many of these modes are between 4 and 4.5 km/sec. The dispersed nature of the waves observed in figures 13 and 14 and the existence of theoretical higher-mode curves with group velocities between 4 and 4.5 km/sec indicates that the observed arrivals may be explained as higher modes. The wavelengths of the shear and Love modes for periods greater than 10 seconds indicate that these waves are affected by the physical properties of the upper mantle. The velocities and dispersion of these waves should give additional information concerning the distribution and lateral variation of shear velocity in the upper mantle.

Love Waves

For figures 10 and 11 dispersion curves were calculated for the first two Love modes of case 124. Sykes, Landisman and Satô (1962) showed that the first Love mode of their oceanic case 122 develops large amplitudes in the low-velocity channel in the upper mantle for periods between 10 and 40 seconds. Cases 122 and 124 have the same mantle structure. For periods shorter than 10 seconds the motion is negligible in the low-velocity channel and is largely confined to the upper 50 km of the earth.

Near 10 seconds the phase velocity curves for the first and second Love modes of case 122 (or case 124) nearly intersect; the group velocity curves do cross near 10 seconds. The configuration of the phase velocity curves is similar to that shown in

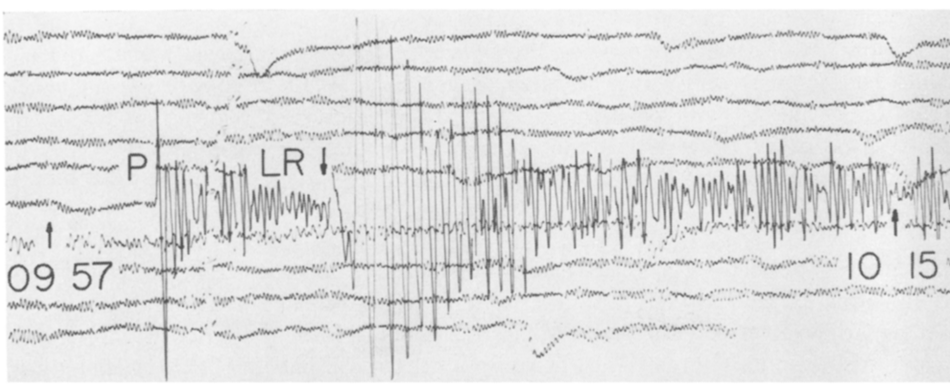

FIG. 15. Bermuda vertical component seismogram illustrating 7 to 10 second waves following the arrival of the *P* wave. Earthquake on Mid-Atlantic Ridge near $27\frac{1}{2}$°N, $43\frac{3}{4}$°W (BCIS), February 28, 1958. Distance to Bermuda $18\frac{3}{4}$°. Clock correction −2.5 seconds. The seismogram traces were darkened with a pencil to insure adequate photographic reproduction. Seismometer period is 15 sec; galvanometer period, 75 sec.

figure 9, Part I, except that in this case the two subwave guides are (1) the low-velocity channel in the upper mantle and (2) the near-surface layers. For periods between 10 and 15 seconds, the second Love mode is more sensitive than the first to changes in the sedimentary and crustal parameters. For these periods the second Love mode, unlike the first Love mode, does not have large amplitudes in the low-velocity channel. Thus in this period range the second Love mode would be more easily excited by a surface source; the first Love mode would be more readily excited by an earthquake in the low-velocity channel. For periods greater than about 40 or 50 seconds, energy considerations show that the first Love mode is more readily excited than the second Love mode by a surface source. For periods between 10 and 25 seconds, the group velocities for the first Love mode of case 122 (Sykes, Landisman and Satô, 1962) are larger than the group velocities for most of the experimental data. However, nearly all of the observations in this period range are bracketed by the dispersion curves for the first and second Love modes. Thus the dispersion curves for the first two Love modes of case 122 are in close agreement with the data in this period range.

Leaking Modes with Predominant Periods Between 5 and 10 Seconds

The existence of leaking modes with periods between 5 and 10 seconds was discussed in Part I. A seismogram illustrating this phenomenon is shown in figure 15. The recording was made at Bermuda at a distance of 18.8° from the epicenter on the Mid-Atlantic ridge. Oscillatory waves with periods of about 7 to 10 seconds can be seen following the *P* wave. Surface waves with these periods as well as Rayleigh waves with longer periods are also present on the seismogram. Short-period waves are also evident in figure 3 following the *PP* arrival at Huancayo. These waves follow various body-wave phases on many other seismograms. The 7 to 10 second waves at Bermuda are interpreted as a type of leaking mode that corresponds to the multiple reflection of *SV* waves in the sedimentary layer. Leaking modes corresponding to the multiple reflection of *P* waves in the water and sedimentary layer were observed previously at Bermuda for oceanic paths from the West Indies (Oliver and Major, 1960). However, these waves (*PL* waves) are easily distinguished from the 7 to 10 second waves by their longer periods (12 to 16 seconds). The observation of the 7 to 10 second waves and the prediction of leaking modes in the same period range (Part I) are another indication that the sediments play a large role in the propagation of certain normal and leaking modes.

Summary

The short-period surface waves are commonly observed for paths crossing oceanic areas. These waves are found on hundreds of seismograms, and some of the best records were selected for analysis in this paper. The short-period waves exhibit normal dispersion and are often present on all three components of the seismogram. When best recorded this surface wave train is initiated with periods of about 14 seconds and with group velocities of 4.2 to 4.4 km/sec. The short-period train is recorded with periods as short as 5 seconds and group velocities as low as 1 km/sec. Normally, the predominant periods of the waves are between 5 and 10 seconds; however, in unusual cases such as in regions of thick sediments, the predominant periods are as great as 12 to 20 seconds.

The short-period waves are propagated in the first Love and first shear (first higher Rayleigh) modes. Waves corresponding to propagation in the first shear mode exhibit retrograde motion for group velocities between 4.2 and 3.4 km/sec. The sense of the motion is usually not distinguishable experimentally for smaller group velocities. The ratio of horizontal-to-vertical motion may be as large as 3.5 at island and coastal stations. The observed periods, velocities, particle motion and ratio of horizontal-to-vertical motion are in close agreement with the parameters computed for models that contain a low-rigidity sedimentary layer.

A map showing the presence and absence of the short-period waves was prepared for the Atlantic, Pacific and Indian oceans. An examination of this map indicates that the presence or absence of these waves may be caused by differences in both the nature of the transmission path and the properties of the source. Normally, the short-period waves are best recorded for paths that are structurally simple such as propagation paths across a single ocean basin; the waves are usually not recorded for paths along oceanic ridges. The short-period wave trains are usually, and perhaps always, associated with shocks of shallow focal depth.

Short-period surface waves with periods between 5 and 10 seconds are usually absent for the paths that cross the continental margins of the Atlantic. In contrast, these surface waves are normally present when the paths traverse the margins of the Pacific. These differences are apparently related to the larger thicknesses of sediments that are found near the continental margins in the Atlantic although perhaps other differences in the configuration of the margins are also important. In the Argentine Basin, a region of very thick sediments, the predominant periods of the short-period waves are 12 to 20 seconds. The calculations presented in Part I of this study indicated that the predominant periods of waves propagated in the first Love and first shear modes should increase as the sedimentary thickness becomes very large. The association of the 12 to 20 second waves with the thick sediments of the Argentine Basin is strong evidence that the sediments play a prominent role in determining the character of the short-period wave train. The increase in the predominant periods of waves associated with the first Love and first shear modes also accounts for the absence of the 5 to 10 second waves in areas of thick sediments such as the Argentine Basin and the continental margins of the Atlantic. Many of the so-called "microseismic barriers" may be caused by the presence of great thicknesses of sediments.

Surface wave data were used in conjunction with reflection and refraction measurements to estimate the shear velocity in the sediments of the Argentine Basin. The average shear velocity in the upper 0.5 km of sediments is about 0.2 to 0.4 km/sec, and the velocity in the kilometer of sediments below this is about 0.5 to 0.7 km/sec.

Dispersion data were also examined for the following paths: west of Mexico to Honolulu; Easter Island ridge to Huancayo, Peru; and West Chile rise to Hallett, Antarctica. Shear velocities of a few tenths of a kilometer per second were obtained for the sediments along these paths; however, these values are more uncertain than those for the Argentine Basin since detailed information was not available concerning the sedimentary thicknesses. Nevertheless, since the shear velocities estimated for these areas are comparable to those obtained for the upper layer of sediments in the Argentine Basin, the mechanism of normal mode propagation is probably the same in all regions.

Arrivals with periods between 10 and 30 seconds and with group velocities between 4.0 and 4.4 km/sec were observed on the vertical components of long-period seismographs. The velocities and periods of these arrivals indicate that these waves probably correspond to certain portions of higher modes of the Rayleigh type.

Oscillatory waves with periods of 7 to 10 seconds were recorded following the *P* wave and other body phases in oceanic areas. These waves are interpreted as corresponding to leaking modes that result from the multiple reflection of *SV* waves in a sedimentary layer.

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