Journal of Geophysical Research Volume 66, No. 11 November 1961

A Telemetering Ocean-Bottom Seismograph[1]

John Ewing and Maurice Ewing

Lamont Geological Observatory, Columbia University
Palisades, New York

Abstract. Successful tests of a telemetering ocean-bottom seismograph have been made on three occasions. In all cases, the seismograph was resting on the ocean botton or planted in the sediments, sending its information to the surface by frequency modulation of a supersonic beam. The use of cables connecting the instrument on the bottom to the recording ship was avoided so that the level of background noise would not be influenced by shaking the instrument by a long cable. These first tests were designed to help determine what frequencies should be recorded, at what levels, and what method is best for transmitting to the recording instrument at the surface the earth's vibration in these frequencies. Neither the instrument used nor the method was the optimum for obtaining all of the seismological data from the ocean bottom, but they have demonstrated the feasibility of ocean-bottom seismographs and have helped to determine the criteria for the more complicated instruments and methods of transmittal which will ultimately make up a world-wide system. Data from such a system are expected to settle the question of the origin and propagation of microseisms, provide detailed information about the sedimentary layer and about the earth's crust and upper mantle, and, most important of all, may greatly increase the radius over which a single station can monitor small earthquakes or explosions. This will materially increase our ability to investigate the seismicity of the entire earth and also to monitor nuclear explosions.

In the preliminary tests body waves from one earthquake and several seismic refraction profiles were recorded. The earthquake record indicates reasonably good signal-to-noise ratio in the short-period range. The refraction profiles give indications from *P* and *S* waves of important regional and local variations in the character of the crust-mantle interface.

Introduction

This paper describes the history and present status of the development of seismographs to record earth vibrations on the ocean floor. The installation of seismographs on the ocean floor offers a number of attractive possibilities for new investigations. Since much of the seismic background noise registered at land stations is attributed to meteorological and man-made sources, it is reasonable to expect that the deep ocean floor, far from the coast, might be exceptionally quiet. If so, we might hope to operate instruments of far greater sensitivity than those used heretofore, thereby increasing our ability to monitor small earthquakes or large test explosions at much greater distances than at present. Fundamental studies of seismicity, which ideally require that data on magnitude, depth, location, focal mechanism, etc., be as nearly complete and uniform as possible, and for shocks as small as possible, will be greatly facilitated by the use of more sensitive seismographs as well as by a more uniformly spaced world-wide network of receiving stations, requirements which may readily be met if ocean-bottom stations are available. The effectiveness of world-wide monitoring for large explosions would likewise be increased.

With stations on the ocean floor it will be possible to determine whether there are waves whose propagation is confined to oceanic structure and cannot be recorded at land stations because of distortion or attenuation at the continent-ocean boundaries. Pertinent data about the character of the boundary and information on oceanic structure which can be derived from studying earthquake waves having entirely oceanic paths may be obtained. At present, observations on islands offer the nearest approach to this capability, and they are limited by noise level and by the effects of the island structure itself on the propagation of waves.

In explosion seismology, receiving stations on

[1] Lamont Geological Observatory Contribution 511.

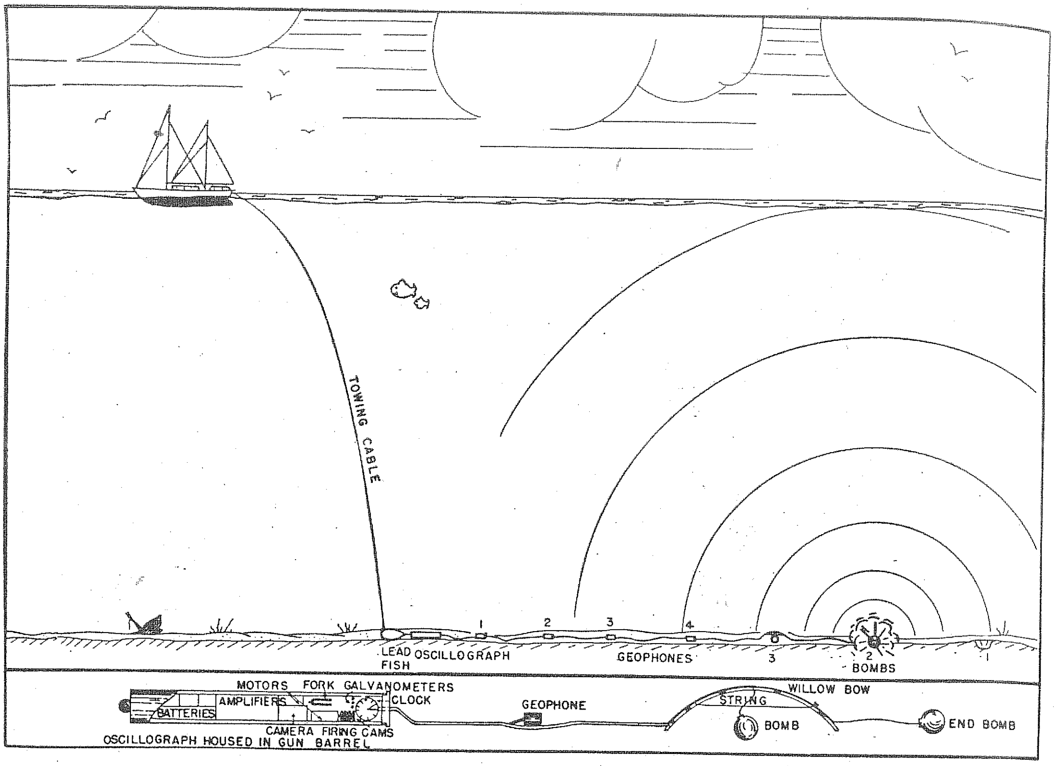

Fig. 1. Cable method of making deep seismic measurements.

the ocean floor will permit a more precise study of all the outer layers of the earth, including the sediments, crust, and upper mantle.

History of the Development of the Ocean-Bottom Seismograph

In 1937 and 1938 seismographs were put on the ocean floor at depths exceeding 2000 meters in several attempts to shoot seismic refraction profiles. Only one of these tests, in 3000-m depth, about 350 km south of Newport, R. I., gave data about ocean-bottom layering. In these experiments an automatic oscillograph, four geophones, and four bombs, all distributed along an electrical cable about 1 km long (as shown in Fig. 1) were lowered to the bottom on the end of a heavy steel wire rope, laid out to form a straight line, and left undisturbed for about 15 minutes while the profile was shot and recorded automatically [*Ewing and Vine*, 1938; *Ewing, Woollard, Vine, and Worzel*, 1946].

In 1939 and 1940 a new system was developed; seismographs and bombs, each containing a timer to make it function at the correct time, were sent to the bottom in deep water, according to the plan illustrated in Figure 2, and a small amount of seismic refraction data was obtained at two stations, one 250 km south of Cape Cod, Massachusetts, in 2600 m, and one 550 km northwest from Bermuda in 4800 m [*Ewing, Woollard, Vine, and Worzel*, 1946]. These instruments were allowed to drop to the bottom under ballast which was detached automatically to allow them to float to the surface after the tests were completed.

At this time the program was interrupted by World War II. When work was resumed in 1947, the availability of wartime apparatus and techniques, the information from earlier tests that the sediment layer was at least several hundred meters thick, plus a desire to shoot long profiles in order to investigate the thickness of the oceanic crust led to the adoption of a technique for using shots and receivers near the sea surface. The effort to operate seismographs on the floor of the deep sea was postponed for several years.

The renewal of this effort was announced in

1951 at the seminar on microseisms held under the auspices of the Pontifical Academy of Sciences in response to the need for observations on the ocean floor ~~to~~ resolve some uncertanties about the generation and propogation of microseisms [*Ewing and Press*, 1952]. The program proceeded at a slow pace, owing primarily to lack of financial support; however tests and preliminary designs were made of seismographs which could be recovered after recording on the ocean floor for extended periods of time.

In 1958, after additional requirements for seismic recording on the ocean floor developed, J. Ewing and B. Luskin joined the project and it was decided that the best approach for the first tests would be to telemeter the information to a surface ship by modulation of a supersonic signal sent out from the ocean-bottom seismograph.

Successful tests with the telemetering ocean-bottom seismograph were made in November 1959, in September 1960, and in January 1961. The problem of keeping proper position for the vessel which received the telemetered signals was greatly simplified by using on the second attempt a trainable transducer for receiving the supersonic signals from the bottom seismograph. In this way the direction toward the seismograph could be determined and corrections for the drift of the ship made before contact was lost. Continuous monitoring thus became possible under all conditions under which the receiving ship could maintain her position.

The unit now being used on the ocean bottom contains a short-to-medium-period vertical seismometer and an amplifier whose signals modulate the frequency of a 12-kc/s acoustical source that sends signals in a broad beam toward the sea surface. The battery is adequate to operate this unit for 7 or 8 days.

The signals received on the monitoring ship are recorded on magnetic tape. The signals from seismic refraction shots are also recorded as conventional seismograms, which include time markers and shot-instant signals received by radio. In some experiments the monitoring ship has also used a near-surface hydrophone to record, for comparison with the ocean bottom seis-

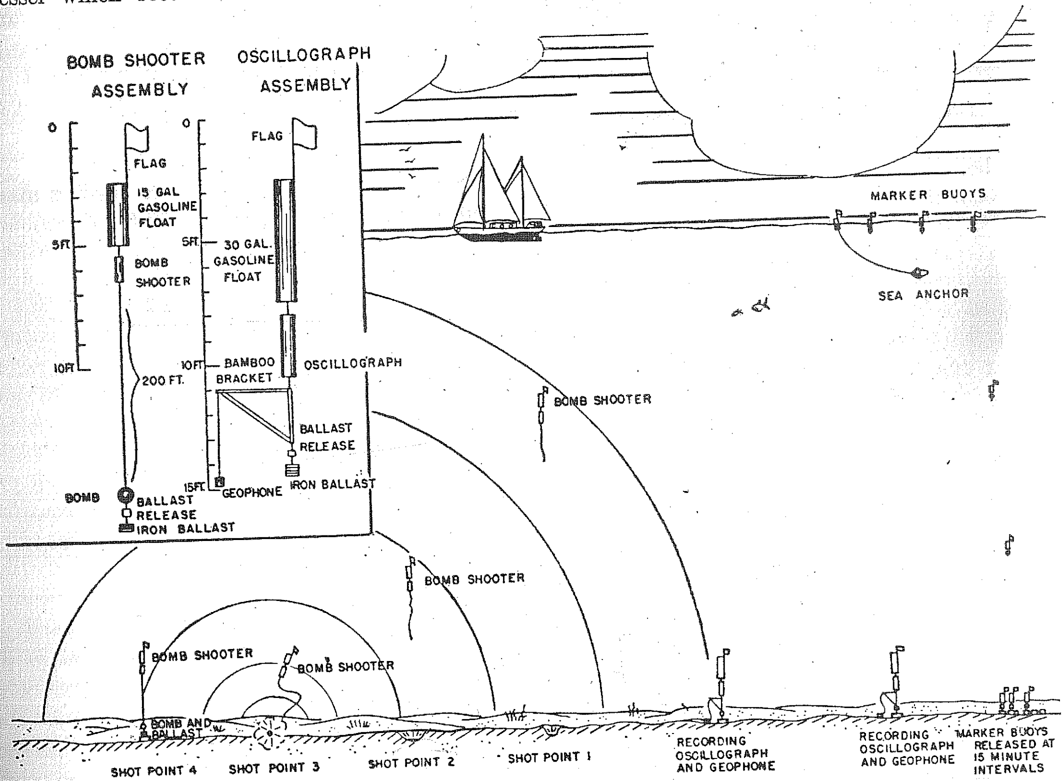

Fig. 2. Float method of making deep seismic measurements.

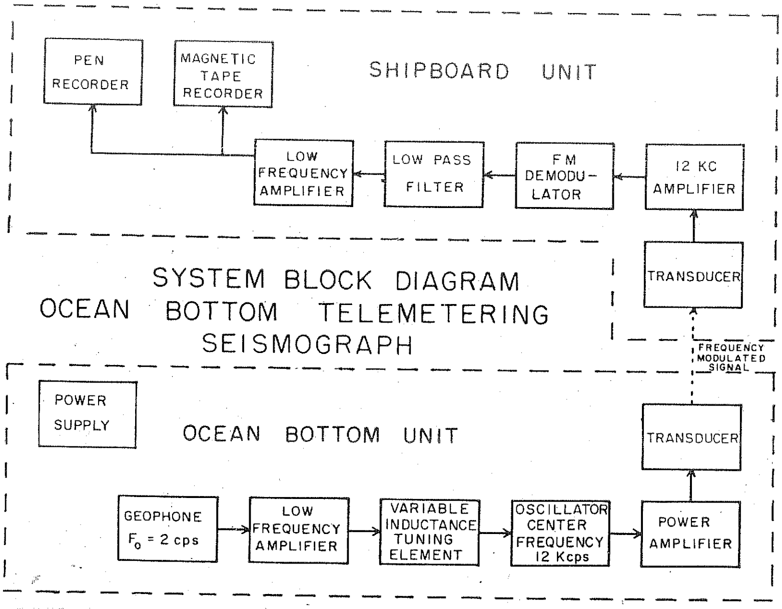

Fig. 3. System block diagram of telemetering ocean-bottom seismograph.

mograph data, seismic refraction data of the type which has been used extensively in our work during the past 15 years.

Some Details of the Instruments Used

A block diagram of the telemetering ocean-bottom seismograph is shown in Figure 3. A geophone with natural period 2 cps is operated at about 0.5 critical damping as the detector of the ocean-bottom unit. The signal from the geophone is amplified and used to drive a variable-inductance element which controls variations in frequency of the oscillator about 12 kc/s mean. The frequency-modulated output of the oscillator is amplified to deliver about 1 watt of power to the transducer.

The transducer is a small free-flooding magnetostrictive unit, whose transmitting response

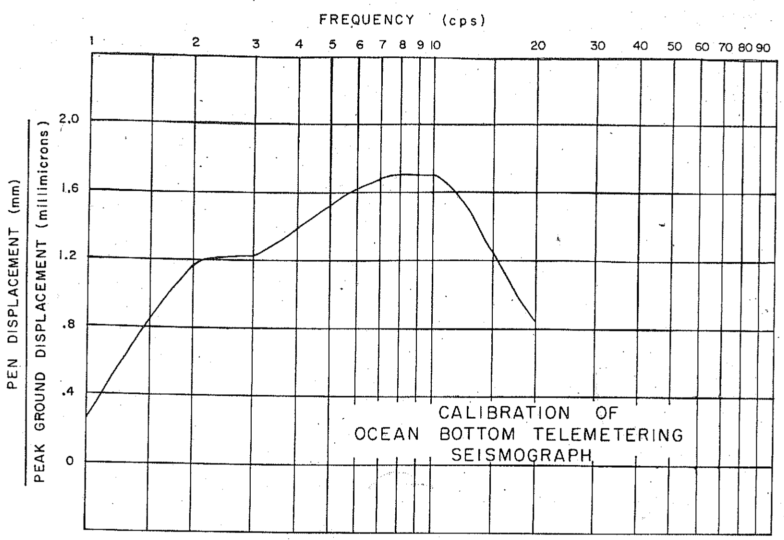

Fig. 4. Calibration of telemetering ocean-bottom seismograph.

TABLE 1. Bottom Seismograph Stations

Station	Date	Monitor Ship	Water Depth, m	Position Lat.	Position Long.	Azimuth Shot Line, deg	Length Shot Line, km
1	Oct. 10, 1959	*Vema*	4800	35°55.5′N	68°37.5′W	165	83
1*a*	Oct. 14, 1959	*Vema*	4995	35°02′N	66°43′W		
2	Sept. 17, 1960	*Grace*	5000	30°17.5′N	65°55.5′W	000	52
						180	268
2*a*	Aug. 10, 1960	*Grace*	5000	30°18′N	65°55′W		
2*b*	Aug. 11, 1960	*Grace*	5000	30°18′N	65°55′W		
2*c*	Sept. 10, 1960	*Grace*	5000	30°18′N	65°55′W		
3	Jan. 26–28, 1961	*Vema*	3700	24°00′N	91°27.5′W	090	235
						300	172
4*a*	June, 1961	*Vema*	5600	44°09′S	53°27′W		
4*b*	June, 1961	*Vema*	5600	44°27′S	53 35′W		

is about 10 to 14 kc/s. The ocean-bottom unit is completely transistorized. Its power supply is a battery, estimated to provide for 7 to 8 days' continuous operation.

The telemetered acoustic signal is received on shipboard with a standard UQN-1 echo-sounder transducer, which is tuned broadly to 12 kc/s and provided with a band width of about 2 kc/s. It has a beam width (between 10 db points) of about 45°. An amplifier covering the band 10 to 14 kc/s is used to amplify the signal from the transducer and to pass it into the FM demodulator (a commercial frequency meter). The output of the demodulator is filtered, amplified, and recorded on a pen recorder and on magnetic tape.

The FM telemetering system was chosen to permit a calibration of the entire system which would be independent of losses in the telemetering link. The calibration curve in Figure 4 was obtained by use of manufacturer's constants for the geophone and electrical measurements made on the circuits. The same calibration curve was obtained in several runs in the laboratory at temperatures from 27° to 0°C and in a run on shipboard just before the main unit was placed in operation.

Three types of battery supply and launching were used. At station 1*a* (see Table 1) the unit was assembled on the framework of the deep sea coring equipment as shown in Figure 5, the geophone being in a pressure case at the bottom of a steel pipe about 3 m long and 7 cm in diameter. The signal leads from the geophone passed first through pressure seals and then through the pipe to a platform at the top. On the platform were the battery, the transducer, and a pressure case containing the amplifier and telemeter circuits. The lead-acid battery was insulated from sea water by rubber sheets which permitted pressure equalization. The transducer was free-flooding. The pipe was mounted beneath the 550-kg weight of the standard coring apparatus by means of a 'weak link' which would allow the weight to drive the lower part of the assembly into the sediment and leave it there. As in the standard coring operation, the assembly was lowered to about 5 m above bottom on the trawl wire, then released by the action of a bottom trigger weight, and allowed to fall freely, as illustrated in Figure 5.

At station 1, a second type of assembly was used, smaller and lighter than the first so that it could be lowered on the hydrographic wire. This apparatus, shown in Figure 6, was to be seated on the ocean floor, instead of being driven into it. The geophone was mounted on a cartwheel base about 1 m in diameter to assist it in maintaining an upright position and was used as the trigger weight in the coring apparatus. The transducer, the electronic circuits, and the battery, which consisted of 48 No. 6 dry cells potted in roofing tar, were placed in the coring head position. When the geophone was seated on bottom the trigger arm released the other package, allowing it to fall and seat itself separately.

At station 2, occupied in September 1960, a third type of bottom unit was used. This was a free-fall model, the design being dictated by the lack of a deep sea winch aboard. Figure 7 shows the arrangement of components and design of the 'missile.' The total length from the transducer to the tip of the nose pipe was about 4 m. Its weight in air was approximately 300 kg. This unit reached bottom in 10 minutes in water 4800 m deep. It struck bottom at a speed of about 8 m/sec without causing damage to any of the components.

At stations 3, 4*a*, and 4*b*, the standard piston coring weight and release mechanism were the basic unit used in planting the seismometer, which was set gently on the bottom by release from the trawl wire. The ocean bottom unit, in which transducer amplifier and seismometer or hydrophone were mounted together, was attached below the core weight by a line some 20 m long. The trigger weight for release was

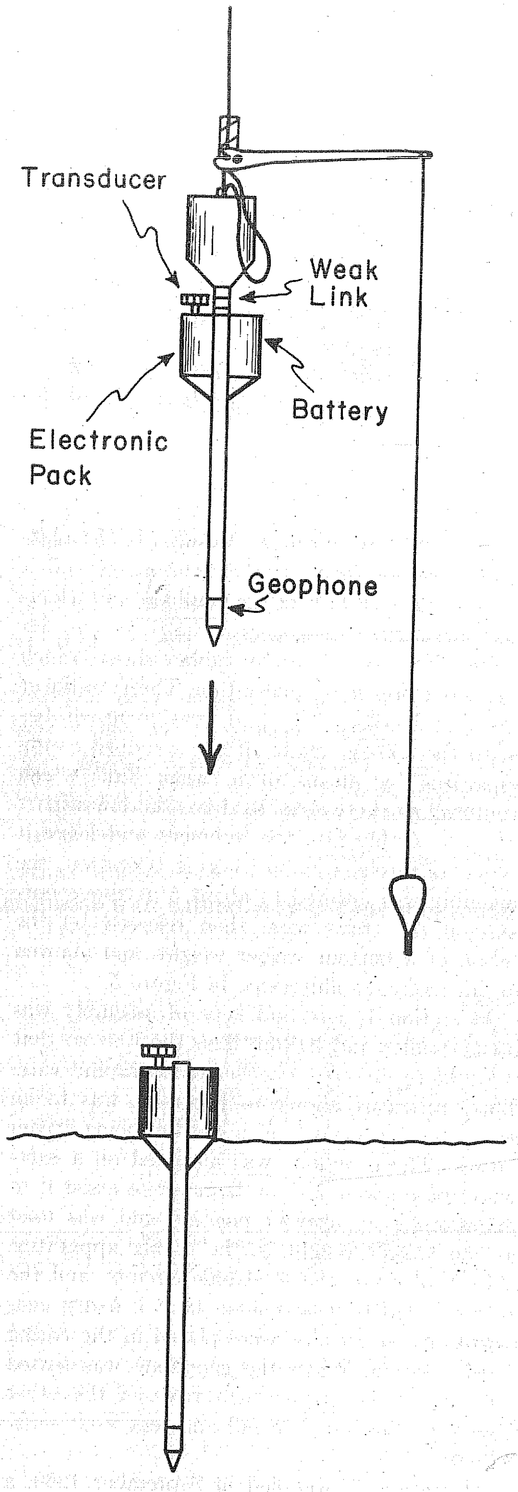

Fig. 5. Seismograph unit designed for placement on bottom with coring apparatus.

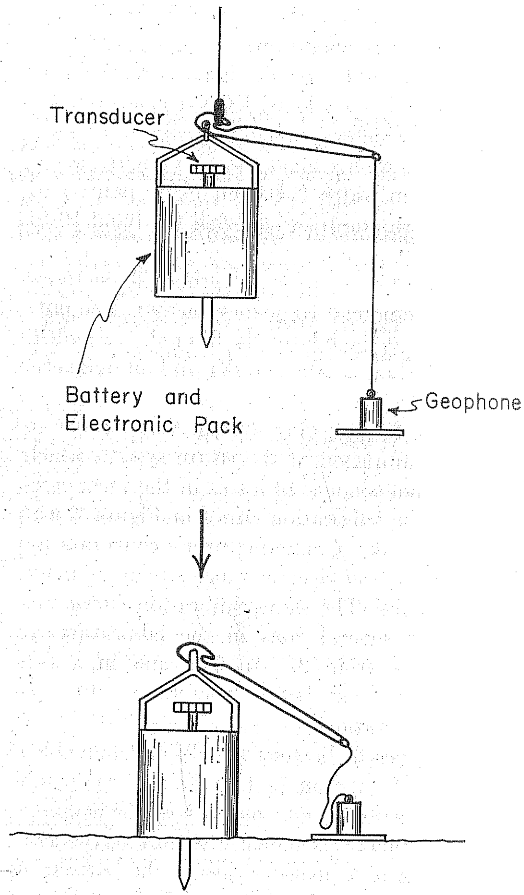

Fig. 6. Seismograph unit designed for placement on bottom with small-diameter hydrographic wire.

about 10 m below the release, or about 10 m above the unit, and displaced about 2 m laterally. It was possible to set the unit on the bottom, judging contact by the telemetered signal, and to monitor it for proper operation for a few minutes with neither the main weight nor the trigger weight in contact with the bottom. If the performance was satisfactory, the unit was lifted and held a few meters above bottom until a 'messenger' could be sent down the trawl wire to put the trigger mechanism in condition to operate. The trawl wire was then payed out slowly until first the seismometer unit and then the trigger weight touched bottom. Contact of the trigger weight with the bottom allowed the main weight to drop about 1 meter and to sever all the connections with the seismometer unit.

Except for the method of placement on the ocean floor, the units described have been very similar functionally. The differences have been in minor details of electronic circuitry, power supplies, or detector sensitivity. Each has used a short-period vertical seismometer modulating a 12-kc/s carrier frequency. In the earlier part of the work, the transducer on the monitoring ship was rigidly attached to the hull, and the only indication of the direction toward the telemetering unit was obtained by correlating the variation in carrier intensity with the roll of the ship. Later, provision was made for training and tilting the transducer in any desired direction, and the problem of maintaining contact with the unit on bottom was greatly simplified.

The demodulator system requires some improvement. On several occasions the carrier itself, and the modulation of it by shots, could be heard clearly, but were apparently below the threshold level for satisfactory operation of the demodulator recorder.

The system described here has many obvious and serious disadvantages, some of which will be discussed. The system was, of course, never intended to serve for long-continued monitoring. For such monitoring, a cable connected to shore or extended along bottom to such a distance that it could then be led to the surface without introducing mechanical noise in the seismograph would be one solution. Another solution would be to store the data at the seismograph, either on magnetic tape or on film, for periods up to possibly one month. Upon command from a monitoring vessel, the accumulated data could be delivered, either by accelerated telemetered playback or by transport of the record to the surface. Both of these systems are being investigated with a view to installing permanent or semipermanent stations. However, further studies with the instrument described here are essential in order to gain sufficient information about noise level, optimum sensitivity, etc., particularly in the longer-period region of the spectrum, so that the permanent installations can be properly designed.

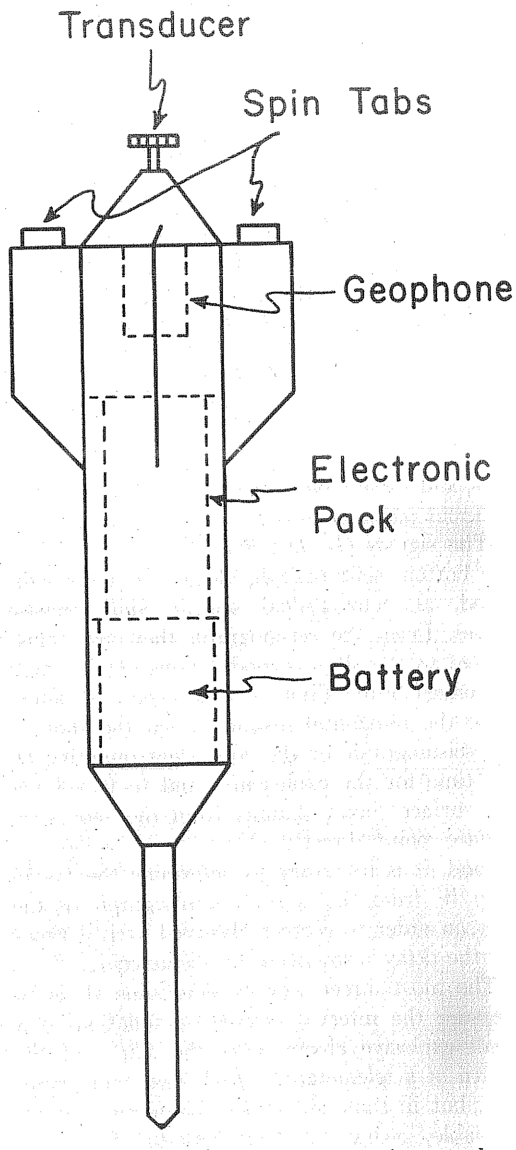

Fig. 7. Free fall ocean-bottom seismograph.

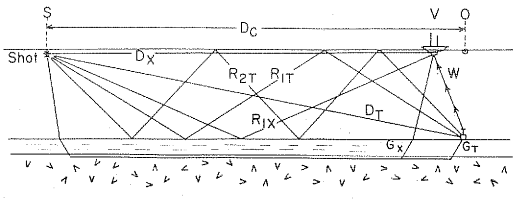

Fig. 8. Ray paths for water waves.

Method of Reduction of Bottom Seismograph Refraction Data

This method of operating and telemetering an ocean-bottom seismograph depends heavily on sound propagation through the water. All signals are received at the ship as acoustic signals, and these have traveled along a number of different paths (Fig. 8).

The signals G_T, D_T, R_{1T}, R_{2T}, . . . travel to the bottom seismograph, along the paths indicated, as conventional seismic and acoustic waves. From the seismograph they are transmitted to the ship as modulations of a 12-kc/s acoustical beam. First, it is necessary to determine the horizontal distance from the shot to the seismograph, or the equivalent quantity D, the time for the explosion sound to travel via the surface sound channel from the shot to a surface point directly above the seismograph. Second, it is necessary to determine the travel time W from the bottom seismograph to the ship, in order to correct observed arrival times for the delay introduced by telemetering.

The most direct way to determine W is to measure the interval between detonation of a small explosive charge near the ship and the return of a telemetered signal from that explosion, but in some situations this method is not available, owing to the unsuitable frequency response of the bottom seismograph or to blanketing of the telemetered signal by a strong echo from bottom.

There is also a comparison method for estimating W, which involves subtraction of W_1, an approximate value, from all telemetered arrival times to obtain a first approximation to the true arrival times at the seismograph. The arrival times are then compared with those for paths such as G_x, D_x, R_{1x}, R_{2x}, . . . directly to a hydrophone near the ship, as shown in Figure 8. The method of comparison must be chosen in a particular way because the ray paths, which are shown as straight lines, are actually curves, owing to variations in velocity with depth.

From the paths shown in Figure 8 it may be seen that the following relations hold between travel times to the seismograph and those to to hydrophone at distances chosen to keep the inclinations of the comparable paths the same:

$$D_T = 1/2R_1(2\ D)$$

$$1/3R_{1T} = 1/4R_2(4/3\ D) = 1/2R_1(2/3\ D) \quad (1)$$

$$1/5R_{2T} = 1/6R_3(6/5\ D)$$

$$= 1/4R_2(4/5\ D) = 1/2R_1(2/5\ D)$$

where D_T, R_{1T}, and R_{2T} are direct and reflected waves to the bottom seismograph, D is the direct wave to point 0 at the surface directly over the seismograph, and R_1, R_2, . . . are reflected waves arriving at that point. The expression $\frac{1}{2}R_1(2D)$, for example, signifies one-half the reflection time (surface-bottom-surface) for a ray at twice the range D. These equations are exact only for conditions of constant water depth. Corrections are required for sloping or uneven topography.

At station 1, in the first long refraction profile recorded by a bottom seismograph (Fig. 9), the quantity W was not measured directly; hence it was necessary to use the water wave comparison method to determine shot-detector distances. This was accomplished as follows:

1. Water waves recorded by a hydrophone suspended near the surface from the receiving ship, yielding data for D_x, R_{1x}, R_{2x}, . . . were used to determine the curves D, R_1, R_2, and R_3 in Figure 9. The observed points are omitted, to improve legibility.

2. From equations 1 and from the above curves, the relationship between direct and reflected travel times for arrivals at point 0 via the bottom seismograph were obtained. These curves are labeled D_T, R_{1T}, and R_{2T} in Figure 9.

3. Approximate values for D_T, R_{1T}, and R_{2T}, the travel times to the seismograph, were determined for each shot by subtracting an amount W_1 = 3.20 sec from all telemetered water wave arrivals. This amount corresponds to the travel time from the bottom seismograph to a point on the surface approximately 1 km from point 0. W_1 was chosen on the basis that an area of radius 1.5 km about point 0 is judged to be the area of detectability of the acoustic beam (based

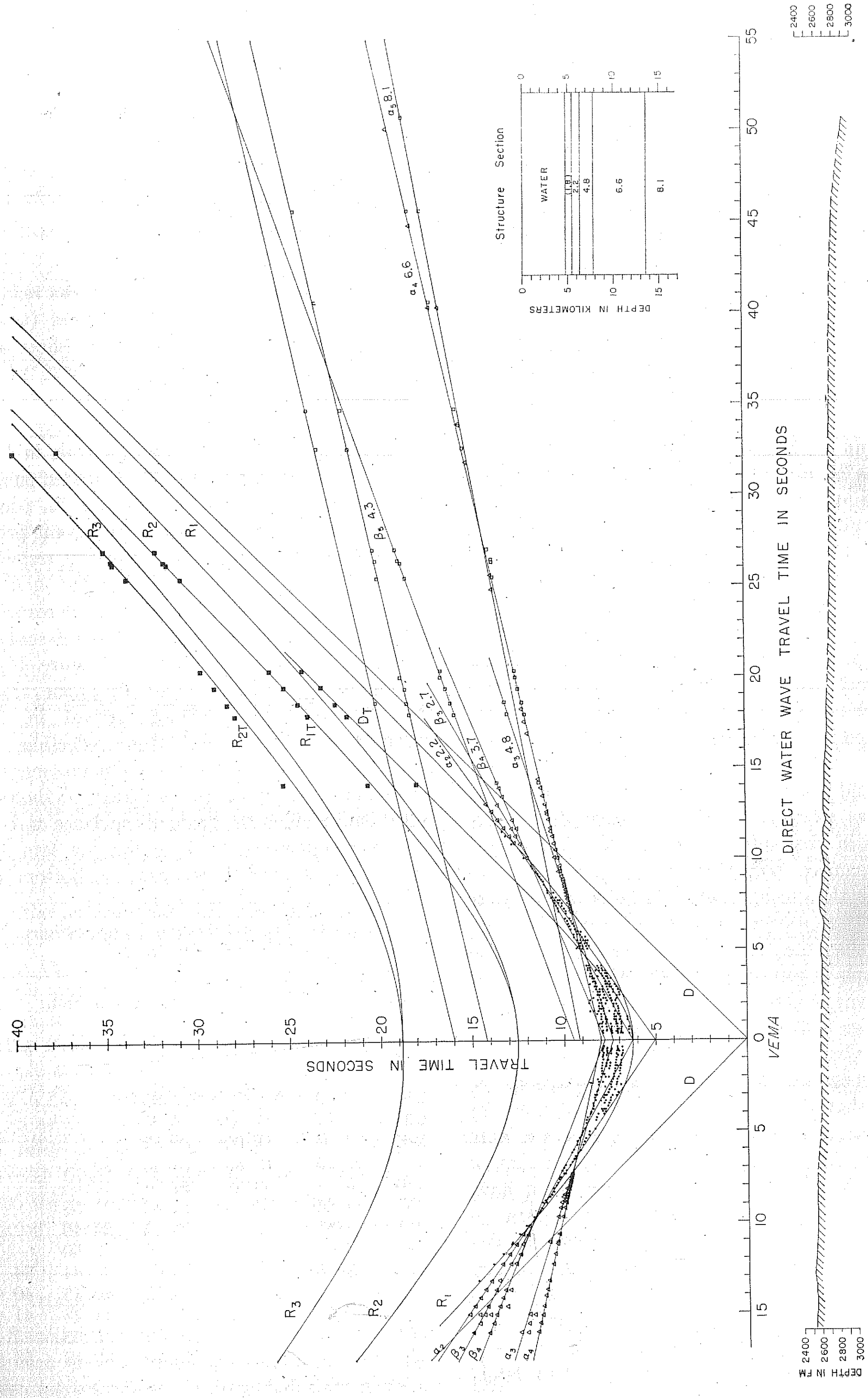

Fig. 9. Time-distance graph, ocean-bottom seismograph profile 1.

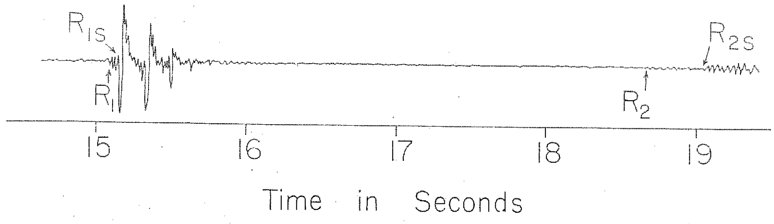

Fig. 10. Oscillogram showing bottom reflected and sub-bottom reflected arrivals.

on the directionality of the receiving transducer). The proper value of W would be between 3.30 and 3.15, the latter being appropriate if the receiving ship were at 0.

4. The resulting travel times (to the seismograph) were then used to determine a value for D from each shot, using equations 1 and the curves D, R_1, R_2, and R_3. The values thus obtained are shown in Table 2 and compared with the values D_x, the direct water wave received at the ship.

5. The travel-time data for water waves and ground waves were plotted as squares in Figure 9, using the distances derived in this fashion, to permit comparison with the curves D_T, R_{1T}, R_{2T} and with the ground-wave data recorded by standard surface hydrophone technique (represented by triangles).

The agreement of the bottom seismograph data and those obtained in the conventional way is close, although there are systematic discrepancies in the water wave arrival times. These discrepancies indicate that the range from the shot to the receiving ship was less than that to the bottom seismograph for all shots. This suggests that the transducer on the bottom seismograph is somewhat directional and that the unit was slightly tilted. A similar transducer has been sent to a testing facility to determine the beam pattern, but the results have not yet been received. From the field observations we know that a sufficiently strong carrier signal for effective demodulation could be received only in a small area; hence even a small amount of directionality in the beam pattern could have produced the range discrepancies shown in Table 2. The fact that hydrophone-recorded ground waves plotted against hydrophone-recorded water waves produced essentially the same refraction lines as the equivalent quantities received by the bottom seismograph is a good indication that the method used to determine shot-to-seismograph distances was reasonably accurate. The errors in distances (travel times of the direct waves) between shots and the bottom seismograph are estimated to be ±0.2 sec for ranges up to 35 km and no greater than ±0.3 sec at the range of the longest shot.

The telemetered water wave arrivals in Figure 9 plot somewhat above the computed curves of D_T, R_{1T}, and R_{2T}. The fact that the R_{2T} points at short ranges are well above the curve is probably the result of poor high-frequency response in the bottom seismograph circuitry and of sedimentary structure. As shown in the response curve (Fig. 4), the sensitivity falls off sharply above 10 cps, and as shown in Figure 10, a tracing of a low-frequency hydrophone trace, the strongest low-frequency arrival in the reflected wave is a sub-bottom reflection, labeled R_s and corresponding to the points marked with small dots in Figure 9. Figure 10 shows a reflection in which the angle of incidence on bottom is comparable to those for the late R_{2T} arrivals. The interval between the bottom re-

TABLE 2. Shot Distance Computations, Telemeter Station 1

Shot	D_x Measured at *Vema*	D, Computed from: D_T	R_{1T}	R_{2T}	R_{3T}
241	13.19		14.25		
242	16.97	17.92	17.95		
243	17.56	18.60	18.57		
244	18.25	19.39	19.35	19.77	
246	19.17	20.39	20.24	20.65	
247	24.85		25.40	25.58	
248	25.67		26.15	26.40	
249	32.02		32.46	32.50	
250	33.96		34.65	34.64	34.77
251	39.67		40.53	40.45	40.48
252	40.31		41.31	41.23	41.48
253	44.77		45.44	45.35	45.42
254	50.07		53.40	53.48	53.40

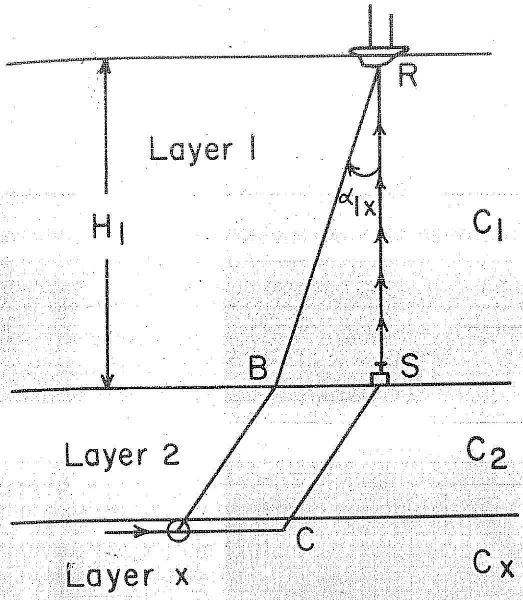

Fig. 11. Ray paths for arrivals to the surface directly and via the bottom seismograph.

flection and the sub-bottom reflection is comparable to the amount that the R_{2T} arrivals are late. At the same range, owing to the difference in angle of incidence of the rays on the bottom, the sub-bottom reflected arrival in R_{1T} is almost coincident with the bottom reflection, and the observed arrivals fit the curve more closely.

The corrections required to reduce the telemetered ground-wave arrivals to the same surface of reference (sea level) as those recorded by the hydrophone are obtained as follows. Figure 11 shows the ray paths corresponding to a signal received via the bottom seismograph and one received at the surface directly above the instrument. If T_1 is the travel time along OBR and T_2 is that along $OCSR$, it is simple to show, with Snell's law, that $T_2 - T_1 = H_1/C_1\ (1 - \cos \alpha_{1x})$, where $\sin \alpha_1 x = C_1/C_x$.

In addition to the corrections for the surface of reference and shot depth, all ground-wave arrival times (telemetered) were adjusted by 0.05 sec to account for the receiving ship's estimated horizontal distance from the seismograph.

Preliminary Results of Measurements

Seismic refraction profiles have been recorded in each of the three successful tests that have been made to date. Pertinent information about them is given in Table 1. The data have been completely analyzed only for station 1. The results are discussed below. Also described are certain observations from stations 2 and 3 which were obvious in a less thorough examination of the data. Travel-time graphs are not shown for the last two stations but probably will appear in a later paper.

Station 1. Figure 12 shows tracings of three records from this station made with shot-detector distances of 25, 37, and 48 km. The charge size for shots 243 and 247 was 4 kg of TNT; shot 249 was 10 kg. Three ground-wave phases (P^*, R_1P^*, and mantle shear waves) are indicated and their correlation shown. The water waves indicated are the direct wave to the transducer on the ship's hull, the first-order bottom-reflected wave received at the transducer, the direct wave to the geophone, and the first- and second-order reflected waves to the geophone. The direct wave to the geophone was received only for the shots at ranges of about 35 km or less, owing to refraction in the water. Presumably, they would have persisted to somewhat longer ranges had the high-frequency response of the system been better. Figure 8 shows the ray paths for reflected and refracted waves received at the geophone and at the hydrophone and transducer.

The water waves which traveled directly or by reflection from the shots to the ship's transducer were identified by comparison with records made simultaneously with hydrophones floated near the surface. These signals represent only the portion of energy in the frequency range near 12 kc/s and appear as sharp impulses of short duration. They have little character other than a sharp beginning. The records for the water waves which travel to the geophone and then up to the transducer are sinusoidal if the amplifiers are not overloaded. Their characteristic frequency at the onset is in the neighborhood of 40 cps, the approximate upper cutoff frequency of the system. The frequency decreases with time owing to bottom penetration by the low-frequency sound.

The ground waves appear at frequencies between 5 and 15 cps. They are similar to waves received by hydrophones suspended near the surface. In the three records reproduced, ground motion associated with the first phases was about 10 mμ at 5 cps.

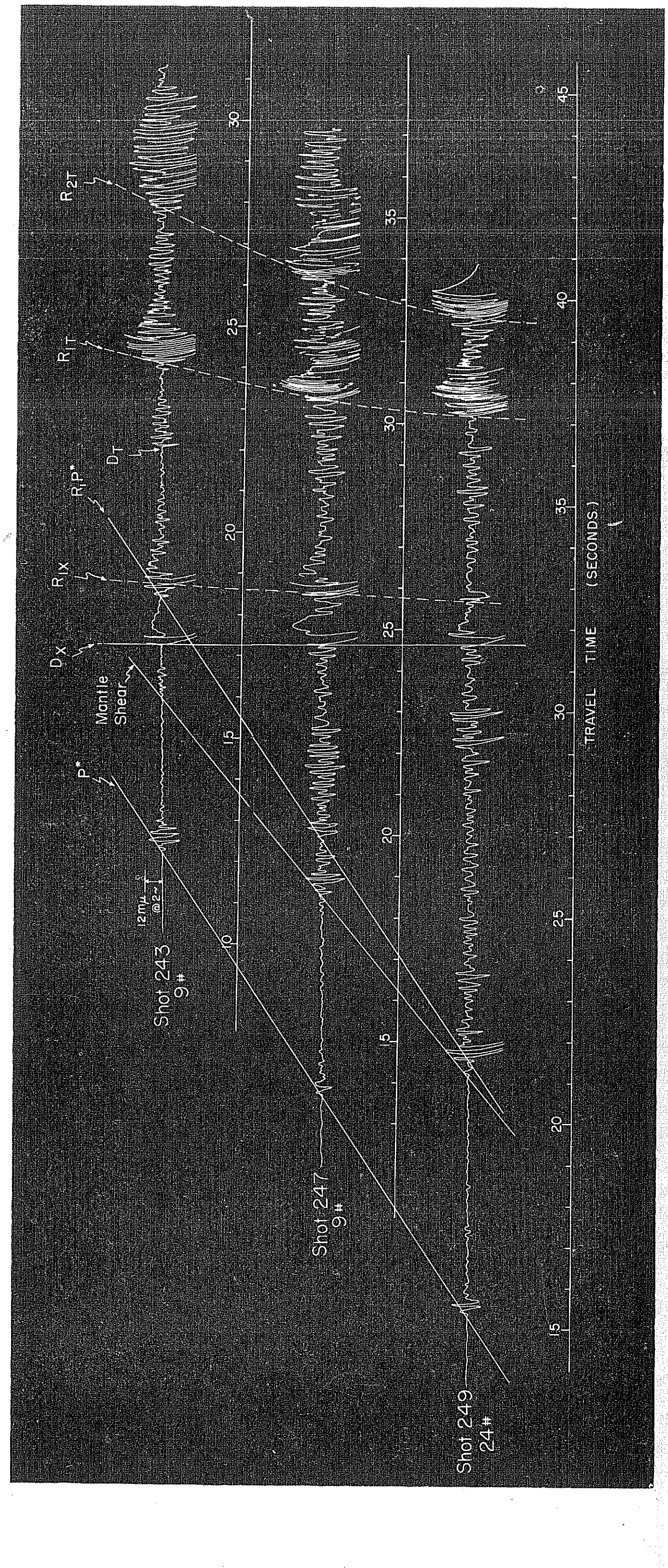

Fig. 12. Typical oscillograms from explosions, profile 1.

Figure 9 shows the time-distance graph and structure section deduced from the data. The telemetered arrivals were received at the same position as a short end-to-end standard profile and are plotted as the extension of one leg of it (ranges beyond 15 sec D time). Triangles denote arrivals at a suspended hydrophone; squares denote arrivals at the geophone on the bottom. Sub-bottom reflections are indicated by dots. All arrivals were corrected with sea level as the surface of reference.

The seismic velocity determined by each line is given (in km/sec) on the time-distance graph and in the structure section. In addition to refracted and reflected sedimentary arrivals, compressional and transformed shear waves were observed for each of the three high-velocity layers (4.8, 6.6, and 8.1 km/sec compressional velocity). Shear waves propagating horizontally in each of these layers underwent transformation at the top of the 4.8-km/sec layer. Transformed shear waves in the 6.6-km/sec layer (β_4) have been observed frequently in standard marine refraction measurements [*Hersey, Officer, Johnson, and Bergstrom,* 1952; *Katz and Ewing,* 1956; *Ewing and Ewing,* 1959] giving about 3.7 km/sec as an average shear velocity in this layer. Mantle shear waves have been recorded only rarely, and in general it has not been possible to determine an accurate velocity from the measurement. In the present experiment, distinct mantle shear wave arrivals were recorded by the geophone on the bottom over a range of 27.5 to 50 km. In some cases, this arrival was one of the most prominent on the record, e.g., shot 247 in Figure 12. It is not yet possible to say whether these arrivals were received strongly owing to some particular features of structure, or whether their reception was enhanced by receiving on the bottom rather than near the surface of the water. Transformed crustal shear waves are frequently received strongly at the surface, however, and both crustal and mantle arrivals cross the water-sediment interface as compressional waves. These facts tend to lessen the possibility that receiving on bottom was an important factor in the strength of the shear waves.

Regardless of the explanation of the strength of these arrivals, there is little doubt that they are mantle shear waves. The slope and intercept of the line through them is in general agreement with that predicted by the structure derived from computations of compressional arrivals. The apparent velocity of mantle shear waves from this measurement is 4.3 km/sec. Although this is an unreversed measurement and may be affected by slope, indications from the compressional wave data are that any slope correction would lower, rather than raise, the shear velocity. The value of 4.3 km/sec is lower than that indicated by the study of earthquake surface waves [*Dorman, Ewing, and Oliver,* 1960]. It is difficult to attach any significance to this discrepancy on the basis of a single measurement; it will suffice here to point out that if there is a discrepancy it may be explained on the grounds that these seismic arrivals sampled only the uppermost part of the mantle where the velocity may be appreciably lower than at a slightly deeper level. It is certain that these arrivals did indeed sample the upper part, because critical distance for shear waves here is 26 km, and the first arrival was received at a range of 27.5 km.

Measurement of background noise at this station shows less than 1-mμ displacement in the frequency range 2 to 10 cps. This agrees qualitatively with the results given by *Brune and Oliver* [1959] from a study of seismic noise at land stations. They found minimum displacements of 1 mμ at 1 cps and less than 0.1 mμ at 10 cps.

Later tests at stations 2 and 3 were made with detectors having somewhat increased response to longer periods, and these showed displacements of up to a few microns at periods of 3 to 5 sec. These periods are in the range of storm microseisms, and it should not be surprising that relatively strong motion would be recorded. Unfortunately, however, the measurements are not wholly reliable because of electronic difficulties encountered at these stations. It appears now that in these two units the power transistors were not properly connected to the wall of the pressure case and as a result overheated the amplifier B batteries, causing some instability.

Station 2. During the monitoring period between seismic shots at this station, body waves from an earthquake at 69°N, 77.5°W, near the Panama-Colombia border, were recorded. The shock occurred on September 19, 1960, and was of magnitude 6 to 6¼ on the Richter scale.

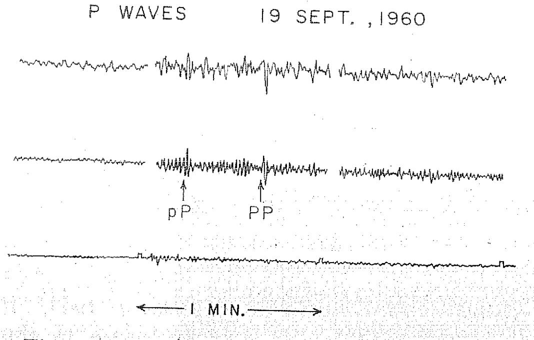

Fig. 13. Earthquake recorded by bottom seismograph (upper trace unfiltered; middle trace filtered 0.75 to 1.5 cps) and by Palisades short-period vertical instrument (lower trace).

Figure 13 shows the bottom seismograph record, unfiltered and filtered 0.75 to 1.5 cps, compared with the record from the Palisades 1-sec vertical instrument. (A standard Benioff short-period vertical seismograph, $T_o = 1$ sec, $T_g = 0.2$ sec, both critically damped, maximum magnification 50,000 at 0.2 sec). The epicentral distance was 2800 km from the bottom seismograph and 3700 km from Palisades.

In recording the bottom seismograph data the minute marks were made by interrupting the signal for a few seconds, and unfortunately the beginning of the P wave occurred during one of these periods. The arrival shortly after P comes at about the proper time for pP or sP, and the arrival at about 40 seconds after P is apparently PP. The sharpness of this phase is noteworthy; it suggests that P would have been very clear and prominent had a more refined timing system been employed. Unfortunately, the end of the reel of magnetic tape arrived a few seconds before S.

Two refraction profiles were recorded at this station—one line shot north 52 km and the other shot south 268 km from the seismograph position. The structure determined by these profiles is of the type which has been shown to be typical of ocean basins by many investigators. The principal crustal layer (6.5 km/sec) and the mantle (8.0 km/sec) were clearly recorded. Exceptionally strong crustal shear waves were recorded, giving a velocity of 3.65 km/sec. A few arrivals which are probably mantle shear waves can be picked in the long profile. The indicated apparent velocity is 4.6 km/sec. The crustal velocities recorded here (both compressional and shear) are determined by end-to-end profile arrangement and are probably close to true velocities. Although the mantle velocities are from an unreversed profile, the values found do not indicate any appreciable effect of slope.

It is noteworthy that very strong crustal shear waves were recorded at this station, whereas only a few rather weak mantle shear waves could be picked. At station 1 the mantle shear waves were much stronger than those from the crustal layer. This difference is apparently the effect of the character of the various interfaces. *Ewing and Ewing* [1959] noted that a marked difference in the strength of crustal shear waves existed between profiles which were quite similar in other respects and likewise attributed this to horizontal variations in the upper crustal interface.

Another observation of some significance in the long profile recorded here is the absence of any noticeable focusing or unusual diffusion of energy in the mantle P waves as the shot distance was increased. The mantle arrivals were never exceptionally strong; and, from the range at which this phase became a first arrival (about 45 km) out to the maximum range, the amplitudes decreased at a relatively even rate.

Station 3. At this station the bottom seismograph was put down in the southern part of the Sigsbee deep, in the Gulf of Mexico. Shots were fired on azimuths of 090° and 300° from the receiving position. In the first of these profiles, ground-wave arrivals were measured at ranges out to 230 km (153 sec of water-wave time). In the second profile, with the same charge size being used, ground-wave arrivals could not be recorded beyond a range of 55 sec. In this area, 45 to 50 sec is about the range at which mantle arrivals can be seen as first events on the seismograms, and the ray paths associated with these arrivals would include only the upper part of the mantle. Hence it would appear that these results indicate significant variation in the crust-mantle boundary over relatively short horizontal distances.

In the first profile at this station (090° azimuth) a pronounced focusing of energy was observed in the mantle arrivals at a range of 150 to 170 km. This can be seen in Figure 14, which shows records of mantle P waves from shots at ranges between 100 and 235 km. All shots were 120-kg depth charges, and all records

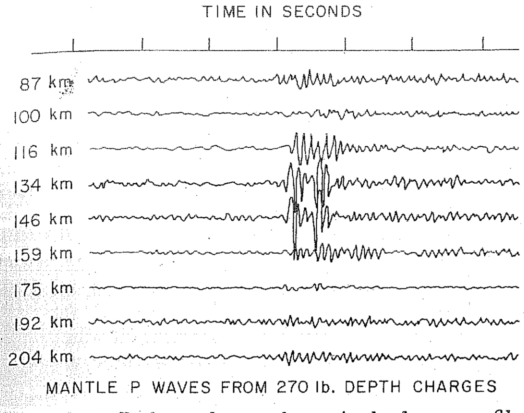

Fig. 14. Refracted mantle arrivals from profile 3 in the Gulf of Mexico showing focusing of energy at ranges of 150 to 170 km.

were made with the same amplifier gain settings. Therefore there is no doubt that the focusing is a structural effect such as might result from a velocity gradient in the upper mantle or a focusing configuration in the crust-mantle interface.

From previous refraction studies in this area [*Ewing, Antoine, and Ewing,* 1960] it is known that the crustal thickness along this profile is appreciably greater than that along the second profile, which was shot northwestward across the main basin of the Gulf of Mexico. The first is, in fact, entirely in the transitional region between oceanic and continental structure, whereas the major part of the second profile crosses structure that is more typical of ocean basins. Station 2, also in an oceanic area, did not give any evidence of focusing, so there is some reason to attach significance to these results, limited though they are. Because amplitudes as well as arrival times can be used, additional measurements of this type should give more definitive information about crustal and upper-mantle structure than has been available previously.

Summary

The principal results derived from these tests are the following:

1. The system of placing a receiving installation on the sea floor and acoustically telemetering the information to a surface ship has proved satisfactory.

2. Body waves from an earthquake have been recorded, indicating that a good signal-to-noise ratio was achieved by receiving on the sea floor.

3. Useful data on background noise in the short-period range have been obtained.

4. Long refraction profiles have given additional measurements of crustal and mantle P and S waves.

5. Variation in the amount of energy propagated along various refraction paths indicates appreciable differences in interface characteristics from one area to another.

Acknowledgments. The work reported here was supported by the U. S. Navy under contracts NObsr 64547, Nonr-266(48), and Nonr-266(65). G. B. Tirey, J. F. Hennion, D. E. Koelsch, S. M. Gerber, A. C. Hubbard, J. L. Worzel, and D. D. Prentiss have made major contributions to many phases of the work. We wish particularly to acknowledge the important assistance rendered by C. Hartdegen and G. R. Hamilton of the Columbia University Geophysical Field Station in Bermuda, both in personal participation and in making available the services of R. V. *Sir Horace Lamb* and her crew. For the Gulf of Mexico operation, we wish to thank H. McLellan and J. W. Antoine for their support with R. V. *Hidalgo*.

Mr. Charles M. Grace's gift of the schooner *Grace* (formerly *Crishaza*) greatly facilitated the work, for it was aboard *Grace* that much of the testing of the instrument was done.

Also participating in one or more phases of this work were A. Stockel, R. Zaunere, and P. Kunsman of Lamont Observatory and F. S. Wendt of R. C. A. Princeton Laboratories.

B. Luskin played a leading part in the design and construction of the telemetering system and contributed effectively to the test at station 1.

References

Brune, J. N., and J. Oliver, The seismic noise of the earth's surface, *Bull. Seism. Soc. Am., 49*(4), 349–353, 1959.

Dorman, J., M. Ewing, and J. Oliver, Study of shear-velocity distribution in the upper mantle by mantle Rayleigh waves, *Bull. Seism. Soc. Am., 50*(1), 87–115, 1960.

Ewing, J., J. Antoine, and M. Ewing, Geophysical measurements in the western Caribbean Sea and in the Gulf of Mexico, *J. Geophys. Research, 65,* 4087–4126, 1960.

Ewing, J., and M. Ewing, Seismic-refraction measurements in the Atlantic Ocean basins, in the Mediterranean Sea, on the mid-Atlantic ridge, and in the Norwegian Sea, *Bull. Geol. Soc. Am., 70,* 291–317, 1959.

Ewing, M., and F. Press, Propagation of elastic waves in the ocean with reference to microseisms, *Pontificiae Academiae Scientiarum Scripta Varia, 12,* 121-127, 1952.

Ewing, M., and A. C. Vine, Deep-sea measurements without wires or cables, *Trans. Am. Geophys. Union,* part 1, 248-251, 1938.

Ewing, M., G. P. Woollard, A. C. Vine, and J. L. Worzel, Recent results in submarine geophysics, *Bull. Geol. Soc. Am., 57,* 909–934, 1946.

Hersey, J. B., C. B. Officer, H. R. Johnson, and S. Bergstrom, Seismic refraction observations north of the Brownson deep, *Bull. Seism. Soc. Am., 42,* 291–306, 1952.

Katz, S., and M. Ewing, Seismic-refraction measurements in the Atlantic Ocean, 7, Atlantic Ocean basins west of Bermuda, *Bull. Geol. Soc. Am., 67,* 475–510, 1956.

(Manuscript received July 21, 1961; revised August 24, 1961.)