SEISMIC INVESTIGATION OF GEOLOGICAL STRUCTURE
BORDERING THE CARIBBEAN ISLAND ARC

PART II

By

David Garrison Harkrider

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C. B. Smith
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In order to ascertain exactly what role the island arcs play in fundamental geologic processes, it is desirable to determine the subsurface structure of these island arcs and their associated topographic features. The deep sea refraction techniques, which had been used successfully in obtaining information on the basement and crustal structure and velocities under the ocean basins and in the vicinity of oceanic islands, were used in this investigation. Between 1949 and 1955 there were three crustal seismic investigations over island arc structures; a single profile by J.B. Hersey in 1949 in the Puerto Rico trench and a series of eight profiles in 1951 across the Eastern Caribbean, interpreted by G.A. Sutton, both sets reported in Ewing and Worzel (1954) and Ewing and Heezen (1955); and a series of three profiles by Raitt et al (1955) across the Tonga Island arc.

In the spring of 1955, an extensive seismic investigation of the Caribbean Island Arc Area was conducted by Officer et al (1956), during a cruise of the research vessels Atlantis and Caryn of the Woods Hole Oceanographic Institution. In all, 47 seismic-refraction profiles were taken in an area including the interior Venezuelan basin, island arc of the Lesser Antilles and Puerto Rico, Puerto Rico trench, Barbados ridge and the Atlantic basin.

In the summer of 1956, the 1955 investigation in this area was continued by Officer et al on a cruise of the research vessels Atlantis and Bear of the Woods Hole Oceanographic Institution. The purpose of this cruise was to obtain better control on the 1955 data and cover the area more thoroughly. This investigation consisted of 30 seismic refraction profiles in the same general area as the 1955 cruise. Besides equipment
for refraction work, both cruises had facilities for sedimentary reflection study.

This paper presents five deep sea refraction profiles from the 1956 cruise in the study of the submerged structure of the Island Arc and Trench. These are profiles numbers 1, 3, 5, 18 and 29. These profiles along with profiles numbers 2, 4, 6, 17 and 20, interpreted in Part 1 by D.E. Miller, are located in an area containing the Atlantic Basin approach to the Puerto Rico Trench, the eastern extension of the Puerto Rico Trench and the southern slope of this extension. (See figure 1 for profile locations.)
II. PROFILE TECHNIQUE AND EQUIPMENT

In making the deep sea refraction profiles the two ship reversed station method was used. In this method both ships are equipped to shoot and record. On the first half of the profile, one ship heaves to and records sound waves generated by explosives dropped by the second or shooting ship, and detonated at a depth below the surface as it proceeds along the profile. The shooting ship continues along the profile, dropping explosives at intervals, until the end of the profile is reached. The shooting ship then heaves to and prepares to listen or record. On the reversed half, the ship that was listening on the first half gets underway and shoots up to the ship that is now listening on the other end of the profile.

The technique and equipment for shooting was similar on both ships. The shooting ships kept a continuous record of the bottom topography over which it went by means of an Edo Corporation echo sounder which was fed into a Woods Hole Oceanographic Institution helical depth recorder. The depth recorder was equipped with phasing switches which enabled it to record the incoming echo without interference from a later transmitted pulse. The recorder gave a visual record of depth versus time.

The shots or explosives ranged in size from 3 pound blocks of TNT to 300 pound Navy depth charges. The explosives were fired with a safety fuse ranging in length from 12 inches for small shots to 72 inches for the 300 pound depth charges. The shooting time interval ranged from 3 minutes to 40 minutes. The small charges and short time intervals were used when shooting close to the listening ship.
The usual procedure used in making a shot went as follows. The shooting ship when a shot was desired, would give the listening ship a one-minute warning by radio. The warning also contained a short description of the shot size and fuse length. At approximately the time for shooting, the fuse was lit and the charge was thrown over the side. The echo pulse transmitter of the Edo was turned off while the echo pulse receiver was left operating. The listening ship was given an "Over the side" warning by radio. The radio transmitter was then switched into parallel with the depth recorder. The time interval, $\Delta t$, between the time when the shot was thrown over the side of the ship until the sound of the explosion reached the shooting ship was obtained by a stop watch.

The sound energy from the explosion traveling through the water to the shooting ship was picked up by the activated sound head of the Edo echo sounder receiver and was simultaneously transmitted by radio to the listening ship, and recorded on the depth recorder and the shooting ships shot record. The shot record, which is connected in parallel with the depth recorder, is a galvanometer excited stylus which makes a record of the sound energy on a moving roll of paper. Standard time intervals of one second are recorded continuously on the shot record paper by means of a second stylus.

All sound energy arrivals from the explosion, including the various bottom to surface reflections, were recorded on the shot record and transmitted to the listening ship. These recognizable arrivals, recorded on the shot record, were useful in determining the shot instant time, which is the expression used to denote the time of arrival of the direct sound energy from the explosion to the shooting ship, when the
listening ship missed the first part of the shot radio transmission from the shooting ship. The time interval on the shot record between shot instant and the first bottom reflection was used to determine the depth of water under the shooting ship when the depth was not recorded on the depth recorder due to weakness of the Edo echo or interference with noise.

The only difference in shooting equipment on the two ships was that on the R.V. Bear the shots were recorded on a Sanborn hot wire stylus recorder, and on the R.V. Atlantis the shots were recorded on an ink stylus recorder. The radio transmitters were medium range, frequency modulated short wave transmitters.

The equipment set up for listening on the two ships was slightly different. Before describing the individual ship arrangements, an overall description of the listening technique is given.

In order to discriminate against noise, each listening ship recorded the output of two independent hydrophones. One at the forward end of the ship and the other near the stern. The hydrophones are listening devices which transform the arriving sound pressure waves in the water into an electric potential. The hydrophones were connected by cables to amplifiers aboard the listening ship, whose outputs were recorded by a photographic galvanometer recording oscillograph.

Floats were attached to the hydrophones and their cables so that their over all buoyancy was slightly negative. After each shot was received by the listening ship, the hydrophones were raised by hand to just below the surface. At a predetermined time shortly before the sound waves reach the listening ship, the hydrophones cables were slacked or played out so that the hydrophones and cables due to their slightly
negative buoyancy behaved like a free falling body when the sound waves reached the listening ship. This isolated the hydrophones from any noise which was generated by ship movement, or mechanical vibrations aboard the ship, and eliminated noise, which would have resulted from vibrations in the hydrophone cables if they had been under tension between the hydrophone and the ship. Using this technique there was still a time delay at the beginning of each cable slack before the hydrophones quieted down. This delay was dependent upon the condition of the sea and the individual slacking of the cable. Therefore on each profile before any shots were fired, the hydrophones were tested to determine the delay time, so that the cables could be slacked at the right time for minimum noise.

When the one-minute warning was received by the listening ship, the listening equipment was energized and the men at the cables were notified. At the time, determined by the hydrophone noise delay time, the hydrophone cables were slacked. Shortly after receiving "Over the side," the paper drive of the oscillograph recorder was started. After the sound energy from each shot had reached the listening ship, the listening equipment was turned off, and the hydrophones raised to the surface. The Edo echo sounder and helical depth recorder were then energized just long enough for a depth reading under the listening ship. The seismic records were then developed and read between shots so that a rough travel time plot could be kept. From the rough travel time plot, the size and time interval for the next shots could be estimated and relayed to the shooting ship. Since the ships motors and generator created noise on the oscillograph galvanometers, the entire listening leg of the profile was obtained under "silent ship" conditions. In other
words, all unnecessary motors and generator, including the ship's engines were stopped, and the electrical power for the listening ship was taken from the ship's batteries.

Brush Development Company type AX-58 hydrophones were used on both ships. Each hydrophone consisted of two Rochelle Salt piezoelectric crystals connected in series and immersed in castor oil for proper impedance matching to the water. The electrical output of the crystals was matched to cable impedance by means of a cathode follower preamplifier. The cables were about 350 feet long and were covered with rubber insulations.

On the R.V. Atlantis each hydrophone cable was connected into a three-channel amplifier and analyzer, similar to the type designed and described by Sutton (1952). Each amplifier analyzer consisted of three separate units. The units had a frequency range built in their circuits. The first unit was a low pass amplifier with a frequency range of 3.5 to 135 cycles per second. The second unit was an "all pass" amplifier with a range of 30 to 250 cps. The third unit was a high pass amplifier with 620 to 14,000 cps range. Each amplifier unit had two outputs, high and low gain, which were 180° out of phase with each other. The outputs were connected to the galvanometers of a Century Geophysical Corporation photographic recording oscillograph.

The shot instant, which was radio transmitted by the shooting ship to the listening ship, and standard time intervals of one second are recorded on the seismic record by means of two additional galvanometers in the oscillograph. The one second intervals were interpolated by means of timing lines photographically produced by a rotating, constant speed, slitted drum in the oscillograph. There were approximately 100 lines
The chronometer impressed a voltage of short duration every second on the standard time trace. Once every 60 seconds the voltage was not impressed on the galvanometer. This missing second on the time trace was called the "skip" second and was used for time correlations between the ships.

During the shooting and receiving phase, navigational fixes, both celestial and loran, were taken as often as possible.
III  ANALYSIS OF DATA

A. General Theory

This section contains the derivation of the refraction formulas used in the solution of the multiple sloping, constant velocity, seismic layer problem. First, consider four layers of different constant velocities, bounded by planes, Figure 2. The possible refraction paths along the layer interfaces between two points A and B on the uppermost surface have been indicated in the figure. The velocity relations are as follows:

\[ V_1 < V_2 < V_3 < V_4 \]

Each angle represents the angle of a given interface with respect to the one above it, and the angles \( \alpha \), \( \beta \), and \( \gamma \) are defined by Figure 2. Let \( h_1a \) be the perpendicular distance from A to the interface between layers 1 and 2; \( h_2a \) that from the foot of \( h_1a \) to the \( x, z \) interface.

Any plane wave traveling in a layer with velocity \( V \), and incident upon the surface of the layer at an angle \( \phi \) with the surface normal, will trace a wave shape on that surface whose velocity along the surface is \( \frac{V}{\sin \phi} \). This velocity is called the trace velocity. (See Figure 3). Using Snell's Law to determine the ratio of the sines of the incident and refraction angles at the velocity discontinuities, and the definition of the trace velocities, \( V_{1a} \) and \( V_{2b} \) at points B and A respectively, for a particular wave path that has been refracted along a layer \( n \) interface, the following formulas are obtained.

\[ V_{2a} = \frac{V_i}{\sin(\beta_2 - \omega_2)} \quad \text{and} \quad V_{2b} = \frac{V_i}{\sin(\gamma_2 + \omega_2)} \]

\[ \frac{V_i}{V_2} = \sin \gamma_2 \]

\[ V_{3a} = \frac{V_i}{\sin(\gamma_3 - \omega_2)} \quad \text{and} \quad V_{3b} = \frac{V_i}{\sin(\beta_3 + \omega_2)} \]
\[
\frac{V_1}{V_2} = \frac{\sin \alpha_3}{\sin \left( \alpha_3 - \omega_3 \right)} = \frac{\sin \alpha_1}{\sin \left( \alpha_1 + \omega_3 \right)}
\]

\[
\frac{V_2}{V_3} = \sin \alpha_3
\]

\[
V_{4a} = \frac{V_1}{\sin \left( \alpha_4 - \omega_2 \right)} \quad \text{and} \quad V_{4b} = \frac{V_1}{\sin \left( \alpha_4 + \omega_2 \right)}
\]

\[
\frac{V_1}{V_2} = \frac{\sin \alpha_4}{\sin \left( \alpha_4 - \omega_3 \right)} = \frac{\sin \beta_4}{\sin \left( \beta_4 + \omega_3 \right)}
\]

\[
\frac{V_2}{V_3} = \frac{\sin \alpha_4}{\sin \left( \alpha_4 - \omega_3 \right)} = \frac{\sin \beta_3}{\sin \left( \beta_3 + \omega_3 \right)}
\]

\[
\frac{V_3}{V_4} = \sin \alpha_4
\]

From the above formulas, it is evident that all the angles in Figure 2 will remain constant if \( \beta_4 \) is changed, or if \( \beta_4 \) is moved relative to \( \alpha_4 \) and vice versa. Now let point \( B \) be moved a short distance toward \( A \), which is kept stationary for the time being. Since the wave refracted from the interface along which it traveled is a plane wave arriving at point \( B \), one can consider the difference in time for the wave to travel to the new point, \( B' \), along a wave path parallel to the original wave path to \( B \) from \( A \), as the difference in time.
FIGURE 2. REFRACTION WAVE PATHS

FIGURE 3. TRACE VELOCITY

a. PLANE WAVE FRONT

b. WAVE NORMAL OR WAVE PATH
that is required for the original plane wave front to travel from B to B'.

The ratio of the change in distance \( x \) to the change in wave travel time is identical to the trace velocity as long as A is stationary. (See Figure 3). Since the angles in Figure 2 remain constant with a change in \( x \), the trace velocity is therefore constant for any change in \( x \), and the ratio of the change in distance to the change in travel-time, which is called the apparent velocity, is constant for every distance \( x \), between A and B as long as A is stationary.

If the travel time of the wave from A to B, A stationary, is plotted against distance, it is apparent from the above discussion that the wave travel-time for each layer, \( M \), is represented by an equation of the form
\[
T = \frac{L}{V} + T_{ma}
\]
where \( T_{ma} \) is the time-intercept at \( x = 0 \) for that layer.

The time-intercepts do not represent actual travel-times, since the refracted wave ceases to exist as such, when the path along the layer interface on which it traveled is equal to zero. This happens at a certain value of \( x \), which is dependent on the layer velocities and their relative thicknesses and dip. For smaller values of \( x \), the wave to that layer will no longer be refracted along the interface, but exists as a reflection only. Although the time intercepts are purely graphical or mathematical concepts, they are useful in calculating the depths of layer interfaces.

Since the time intercept is the extrapolation of the travel-time line from the \( x \) where the refraction waves cease to exist to equal to zero, the apparent velocity must remain constant. Therefore the angles with the interface normals must remain the same as in Figure 2. Now the time of travel along the layer interfaces for this fictitious
path associated with $X$ equal to zero represents a negative time and is equal to the path length along the interface divided by the velocity of the lower layer at the interface. The rest of the path represents positive time. Figure 4 illustrates the fictitious wave paths associated with these intercepts. Comparing this figure with Figure 2 the following equations for the time intercepts are obtained.

$$T_{za} = \frac{(AC + AF)}{V_1} - \frac{CE}{V_2}$$

$$T_{za} = \frac{(AF + AC)}{V_1} + \frac{(FA + CE)}{V_2} - \frac{HF}{V_3} \quad (2)$$

$$T_{za} = \left(\frac{AH + AF}{V_1} + \frac{(FM + EM)}{V_4} + \frac{(MD + ND)}{V_5} \right)$$

Upon substituting trigonometric relations obtained from Figure 2, and Snells Law, $T_{za}$ becomes

$$T_{za} = \frac{2h_1}{V_1 \cos \theta_2} - \frac{2h_1 \tan \theta_2}{V_1} \quad (3)$$

$$= \frac{2h_1}{V_1 \cos \theta_2} \left(1 - \sin^2 \theta_2\right) = \frac{2h_1 \cos \theta_2}{V_1}$$

The trace velocities along the $V_1$-$V_2$ interface for the $V_3$ refracted wave at points F and G are (See Figure 5).

$$V_{3f} = \frac{V_2}{\sin (\theta_3 - \omega_3)} = \frac{V_1}{\sin \theta_3}$$

and

$$V_{3c} = \frac{V_c}{\sin (\theta_3 + \omega_3)} = \frac{V_1}{\sin \beta_3} \quad (4)$$
FIGURE 4. FICTICIOUS WAVE PATHS ASSOCIATED WITH TIME INTERCEPTS

FIGURE 5. FICTICIOUS PATHS SHIFTED AT $V_1 \cdot V_2$ INTERFACE
By using these trace velocities, the path in Figure 5, \( \overline{DE+FS-RT} \), may be shifted to \( \overline{DE+DS-RT} \), the time for which is known by comparison with equation (3)

\[
2h_2 \cos \alpha_2 / V_2
\]

Therefore

\[
T_3a = 2h_2 \cos \alpha_2 / V_2 - FD \sin \alpha_3 / V_2 - DF \sin \beta_3 / V_2
\]

\[
+ (AF + AB) / V_2
\]

\[
= 2h_2 \cos \alpha_2 / V_2 - HF \sin \alpha_3 / V_2 - HF \sin \beta_3 / V_2
\]

\[
+ h_1 / (V_1 \cos \alpha_3) + h_1 / (V_1 \cos \beta_3)
\]

\[
= 2h_2 \cos \alpha_2 / V_2 + h_1 (\cos \alpha_3 + \cos \beta_3) / V_2
\]

The relation for \( T_{fa} \) is obtained by the same process. The resulting time-intercept formulas are:

\[
T_{fa} = 2h_1a \cos \alpha_2 / V_2
\]

\[
T_{fa} = 2h_2a \cos \alpha_3 / V_2 + h_1a (\cos \alpha_3 + \cos \beta_3) / V_2
\]

\[
T_{fa} = 2h_3a \cos \alpha_4 / V_2 + h_2a (\cos \alpha_4 + \cos \beta_4) / V_2
\]

\[
+ h_1a (\cos \alpha_4 + \cos \beta_4) / V_2
\]

All of the above formulas were derived by keeping A stationary and changing the position B. Now if the arrival times are plotted versus distance as B moves away from A for refracted ions along these four layers, one will obtain four straight lines of the form \( \ell_m = \frac{x}{V_m} + T_m \) with inverse slopes of \( V_m \). The \( \ell_1 \) line will be a straight line representing the direct wave from B to A and its inverse slope will be the true velocity of the first layer. If B is then stopped, at say a distance \( R \) from A, and one moves A up to be B plotting arrival times versus distance from A to B, a graph similar to the above is obtained. This graph will
consist of four straight lines of the form \( t_n = \frac{y_n}{V_{nb}} + T_{nb} \)

The inverse slopes of these straight lines will be equal to the apparent velocity of the arrival times at A, \( V_{nb} \), which is determined from the angles \( \beta \) instead of \( \alpha \) as above. The time intercepts at B are governed by equation (7) with a substitution of \( h_{ib}, h_{2b}, h_{3b} \) for \( h_{ib}, h_{2a}, h_{3a} \).

At the distance R, the wave paths from A to B are the same as from B to A, and therefore the arrival times at distance R for both graphs are equal.

Figure 6 shows the characteristic travel-time graph, which is composed of the two graphs for stationary points A and B superimposed at distance R, which is called the reverse distance. The data which are taken from this travel-time graph are: (1) the time intercepts \( T_{2a}, T_{2b} \), \( T_{3a}, T_{3b}, T_{4a}, T_{4b} \) and (2) the apparent velocities \( V_{a}, V_{b}, V_{2a}, V_{2b}, V_{3a}, V_{3b} \).

From this data and equations (1) and (7), one can determine the true velocities of the four layers, the relative angles between layers, and the layer thicknesses at the stationary points, A and B, for the first three layers. The theory can be easily extended to more layers by using the same analysis.

**B. Application to Deep Sea Refraction Profiles**

In order to apply the general theory to the sedimentary and crustal layers below the ocean, we must make four assumptions:

1. Each seismic layer is a constant velocity layer bounded top and bottom by planes.

2. At the interface between two seismic layers the path of the seismic waves obeys Snell's Law of Refraction.

3. The waves generated by a refracted wave traveling along an
FIGURE 6. CHARACTERISTIC TRAVEL-TIME GRAPH
interface are in phases so that they form a plane wave leaving the interface at an angle with the interface normal equal to the angle of incidence of the initial wave, which was refracted along the interface.

(4) Any travel time will be unchanged if the shot point and the recording point are interchanged.

Deviations of the actual from the assumed conditions will be indicated by differences between the observed travel time data and a straight line drawn through these times.

The relative angles between interfaces which are computed give the apparent dip of the layers along the profiles.

In figure 6 the first refraction arrivals observed on the seismic records are represented by heavy lines. Since the reception of the seismic energy for each refracted wave is generally extended, second arrivals behind the first arrival may be hidden in the first arrival sequence. At greater distances though second arrivals may be more apparent. Thus it is sometimes possible to fill in the rest of the graph by use of later arrivals in order to get a better determination of the refraction lines. When the lines are determined by first arrivals only, the line will represent the ray path along only a small part of the interface. This section of the interface does not coincide with the section indicated by the reversed profile. If the interface deviates from a plane and the sections observed on the graph are not coplanar, the extrapolated lines for a given layer will not reverse (i.e. same arrival time at the reverse ends of profile). But if the deviation is small, the calculation of thicknesses will give an average regional angle for the interface, since the thickness or time intercept equation (7) is
fairly insensitive to the changes in interface angles and most interface angles encountered are usually small by the correct choice of direction of profile.

The general theory assumes that the layer velocities increase with depth, since by Snell's Law it is not possible for a wave to be refracted along a layer interface if the lower layer at the interface has a smaller velocity than the layer above it. Therefore a low velocity layer between two higher velocity layers will not be indicated on the travel time graph, but will add to the calculated thickness of the seismic layer above it. There is another situation in which a layer might be missed. This is what is known as a masked layer. It is represented by a thin layer bounded on top by a thick low velocity layer of large velocity contrast and on the bottom by a higher velocity layer. In this situation, it is apparent (See figure 7) that the arrival time line for this layer will be a first arrival for only a short interval of distance and could be very easily overlooked. This is another inherent difficulty in being able to determine the arrival lines by first arrivals only.

In order to plot a travel time graph, besides knowing the arrival times, one must determine the distance between the shot point and the recording or listening point. At sea, this is done primarily by the measurement of the direct travel time of sound in sea water. The velocity structure of the sea is such that at the sea surface, there is a constant velocity layer of about 200 feet thick followed by a decrease in sound velocity with depth until a certain depth at which the velocity starts increasing. The constant velocity layer acts as a sound channel and its velocity is primarily dependent on the water temperature. By use of a
FIGURE 7. TRAVEL TIME GRAPH SHOWING MASKED ARRIVAL
bathythermograph, the thickness and velocity of this layer is determined and with the direct wave travel time the distance between the shot and receiver can be calculated.

The direct wave travel time for each shot is also plotted against the time of day that the shot was thrown over the side to give a navigation plot. This subsidiary plot will be a straight line for constant ship speed and course. This plot is valuable for determining the speed of the shooting ship, necessary for the corrections described below.

Before plotting corrected arrival times, it is necessary to find the reverse point or the distance from listening station to listening station. The reverse distance is determined from the navigation plot and the time when first the shooting ship stopped to listen and the first listening ship started along the profile to shoot.

The first layer in the deep sea refraction profiles is, of course, the ocean. The thickness of this is determined by the depths obtained from the echo sounder. In general, the ocean bottom is very irregular. The depths obtained from the shooting ship are plotted versus distance. In order to conform to general theory which requires that the layer interface be planes, a mean base line was drawn through the bottom topography, so as to closely represent the bottom trend and corrections were made to the arrival times, reflection and refraction, to compensate for the reduction of the bottom to a plane. The theory and the method of calculating the corrections are taken up in a later section. (See Raw Data Corrections). As mentioned in the previous paragraph, the sound velocity in the ocean is not constant. But for the purpose of calculation in equations (1) and (7), the mean vertical velocity from
sea level to the base line is used for \( v \). This mean vertical velocity is defined by the relation,

\[
\bar{v} = \frac{H}{\int \frac{dH}{C(z)}}
\]

and is that velocity which when multiplied by the vertical reflection time will give the true depth of water. This is used instead of the usual average vertical velocity because in the refraction formula, thicknesses or distances are determined by differences in time and not vice versa. Although the average vertical velocity is not theoretically correct for use with waves not perpendicular to the bottom, the usual angle of incidence for the refracted wave in water is usually not large enough to warrant the use of a more accurate velocity. The average vertical velocity is obtained from "Tables of the Velocity of Sound in Pure Water and Sea Water," published by the Hydrographic Department, Admiralty, London (1939).

At the critical angle where the angle of refraction is \( 90^\circ \) and the wave travels along the layer interface, the travel time path for the refracted wave is equal to that for the corresponding reflection. At greater distances the time for the refraction path is shorter than that for the corresponding reflection path. If the bottom reflections are also plotted on the travel time graph, the reflections will plot as a hyperbola which approaches the direct wave line asymptotically as distance is increased. The refraction for that layer will be tangent to the reflection at the critical angle and diverge from the reflection as distance is increased. The reflection from the sea bottom is easily detected on the records and if a refraction along the water-bottom
interface exists, it will appear as a refraction line tangent to this bottom reflection line on the travel time graph. Since this refraction arrival was very rarely noticed on these profiles, an estimate had to be made to determine the velocity of this material in order to determine its thickness from equation (7). First, the thickness was estimated by constructing a line tangent to the water reflection parallel to the $v_f$ refraction lines and this intercept was noted. The difference between this constructed intercept and the $v_f$ time intercept was halved and multiplied by a reasonable velocity for that layer, usually composed of low velocity sediments. The depth obtained was then compared with a composite sediment mean vertical velocity graph which had been determined from a separate investigation by T. Lawhorn. The velocity associated with this depth or thickness was then multiplied by the half intercept differences, giving a new thickness which was compared with composite velocity graph. By this method of successive approximation a mean vertical velocity was obtained that corresponded to the approximate thickness of the velocity graph. This velocity was used in equation (7) for $v_2$.

Whenever there was good reason to believe that a layer was masked due to an unusually thick layer above it, a fictitious refraction line for this layer was drawn on the travel time graph with an inverse slope equal to the velocity of this layer. The fictitious line was drawn through the points that might have been mistakenly associated with another line. This was done in only one of the profiles discussed in this paper and there was a very good reason for assuming that this interface or layer existed. The reason is stated in the individual profile results.

C. Obtaining the Raw Data from the Seismic Records
Figure 8 illustrates a typical seismic record taken while listening on the R.V. Atlantis and on the R.V. Bear. The bottom trace on both records is the chronometer trace discussed in Part II of this paper. The trace just above it is the shot instant trace; it records the instant of detonation transmitted from the shooting ship. The six center traces on the Bear record and the four center traces on the Atlantis record are the high frequency traces. On the Atlantis record the frequency traces from one hydrophone are on the top half and the traces for the other hydrophone are on the bottom half. On the Bear the frequency trace for one hydrophone alternates with the trace for the other hydrophone. Therefore, every two traces represent a single frequency range on the Bear for the two hydrophones.

The energy from the shot explosion contains a wide frequency spectrum, and the refractive arrivals are dominantly low frequency. The surface sound channel arrivals are dominantly high frequency. The bottom reflections cover a broad spectrum.

Since the direct wave travels in the surface sound channel, one would expect to detect the direct wave at short distances from the shot with both low and high frequency components. At greater distances, the wave is composed of high frequency components only and appears as a high frequency spike on the high frequency trace. Therefore the "D" or direct wave is usually read from the high frequency trace. As long as the sound channel exists, the direct wave can be recognized up to distances of 50 miles.

The refracted wave path lengths are extremely long and the geometric path is through great thicknesses of material. Therefore
one might expect the refracted wave to be composed primarily of low frequency components. This arrival is found on the low frequency traces with a frequency range of from about 5 to 30 cps.

The bottom reflection arrival is the most intense signal and contains a broad spectrum. It is easily detected on all traces at every distance used in these profiles. The reflected, refracted, and direct waves are indicated in figure 8. All arrival times are read at the beginning of each arrival.

D. Corrections to and Reductions of Raw Data

Before the raw data taken from the seismic records can be used to obtain the final values for the travel time plots, several corrections and reductions must be made. They include a zero time correction, a shot instant correction, a correction reducing all travel times to sea level, and a correction for irregular topography.

The zero time correction is due to the fact that all arrival time information taken from the seismic records is read from the first chronometer tick following the shot instant. Therefore the time difference between the shot instant and the zero second tick must be added to the time data. This correction is made on the record.

Since the shooting ship is moving along the profile away from the shot charge, which was thrown over the side of the shooting ship, a shot instant correction must be made. This correction is equal to the time required for the sound to reach the detector aboard the shooting ship from the explosion. In order to do this it is necessary to know the depth of the shot and the distance that the shooting ship has traveled from the
time that the shot was thrown over the side to the time of explosion, \( \Delta t \).
The distance that the ship has traveled is obtained by multiplying the ships speed by the time interval, \( \Delta t \).
The time \( \Delta t \) is recorded by the shooting ship for each shot. The depth of shot is determined from the time of the first bubble pulse, i.e., second detonation from the oscillation of the gas bubble. The bubble pulse is a function of charge size and pressure (depth). A set of bubble pulse time versus depth graphs have been empirically determined for various charge sizes and types. These two distances determine the legs of a right triangle, with the hypotenuse being the distance from the shot to the shooting ship at the time of explosion. (See Figure 9). Knowing the velocity of sound in the water from bathythermograph observations, the shot instant correction can be then calculated.

Since the shots explode at various depths due to charge size and fuse length, it is desirable to reduce the various arrival time data to a surface of reference. This surface of reference is chosen as sea level. Also since the arrivals are detected by hydrophones at about 100 feet below the surface, a time correction is needed to reduce that end of the wave path to sea level. The total correction for the first reflection is the time it would take an essentially plane wave front to travel in water from A to B plus the time from C to D (see Figure 10).
This time is obtained by multiplying the depth of the shot plus the depth of the hydrophone, expressed in seconds, of vertical water travel time, by \( \cos \theta \), where \( \theta \) is the angle of incidence of the reflected wave on the bottom. The correction for the refracted waves is equal to the product of the sum of these depths and the cosine of the angle, \( \alpha \).
FIGURE 9. SHOT INSTANT CORRECTION

FIGURE 10. 1ST BOTTOM REFLECTION REFERENCE
LEVEL CORRECTION
whose sine is the ratio of the surface water velocity, $c_0$, to the velocity in the refracted layer, $v_n$, (see Figure 11). The value of the sine of $\alpha$ is based on the assumption that the layer interfaces are parallel. For nonparallel interfaces there would be a correction for it. But for the interface angles usually encountered in this type of work, it would mean a much smaller correction to an already small reference level correction and is therefore negligible. Since the direct wave path is almost parallel to sea level, the correction for it is negligible. The values for $c_\theta$ versus direct wave travel time for different depths and the values of cosine $\alpha$ versus the ratio $v_n/c_\theta$ are found in "Reduction of Deep Sea Refraction Data", Technical Report No. 1, Lamont Geological Observatory, by C.B. Officer and P.C. Wunschel.

As stated earlier, in order to conform to the general theory which requires that the layer interfaces are planar, a topographic correction must be made in which the effects due to the irregularities in the bottom topography are removed from refraction and first reflection travel times. The first reflection travel time was corrected by a correction similar to the reference level correction. The point of reflection was determined on the bottom topography plots and its height above or below base line was obtained from the plot. The correction was the time it would require an essentially plane wave front to travel from A to B plus the time from B to C. (see Figure 12). This is obtained by multiplying the height above or below the base line, expressed in seconds of vertical water travel time at that depth, by the cosine of $\theta$, where $\theta$ is again the angle of incidence of the reflected wave on the bottom.

The topographic correction for refracted waves is derived on
FIGURE 11. REFRACTION REFERENCE LEVEL CORRECTION

FIGURE 12. 1ST REFLECTION TOPOGRAPHIC CORRECTION
the assumption that the layer interfaces are parallel and that the part of the bottom and the surface of relief that the wave passes through are parallel to the interfaces. (See Figure 13). The travel time for the undeformed case (dashed lines) is given by

\[
T_i = \frac{V}{V_m} - \frac{2}{V_m} \left[ h_1 \tan \alpha_{15} + h_2 \tan \alpha_{25} + h_3 \tan \alpha_{35} + \tan \alpha_{45} \right]
\]

\[+ 2 \left[ \frac{h_1 \sec \alpha_{15}}{C_v} + h_2 \frac{\sec \alpha_{25}}{V_2} + h_3 \frac{\sec \alpha_{35}}{V_2} + h_4 \frac{\sec \alpha_{45}}{V_4} \right] \tag{9}
\]

The travel time in the deformed case is given by

\[
T_2 = \frac{V}{V_m} - \frac{2}{V_m} \left[ h_1 \tan \alpha_{15} - \tan \alpha_{15} + h_2 \tan \alpha_{25} + h_3 \tan \alpha_{35} \right]
\]

\[+ 2 \left[ \frac{h_1 \sec \alpha_{15}}{C_v} - \frac{h_1 \sec \alpha_{15}}{C_v} + \frac{h_2 \sec \alpha_{25}}{V_2} + \frac{h_3 \sec \alpha_{35}}{V_2} + h_4 \frac{\sec \alpha_{45}}{V_4} \right] \tag{10}
\]

Subtracting and making the substitution from Snell's Law

\[
\sin \alpha_{15} = \frac{C_v}{V_m}, \quad \sin \alpha_{35} = \frac{V_2}{V_m} \tag{11}
\]

We have

\[
AT = T_1 - T_2 = \frac{\Delta h}{C_v} \left[ \sqrt{1 - \frac{C_v^2}{V_m^2}} - \frac{C_v}{V_m} \sqrt{1 - \frac{V_2^2}{V_m^2}} \right] \tag{12}
\]
which is the correction in travel time, $\Delta T$, of a refracted arrival due to a height, $\Delta h$ above or below the base line. The correction is positive for topography above the base line and negative for topography below it. When computing the correction it was necessary to know the point at which the refraction wave was incident on the bottom. The distance of the incident point from a point directly below the ship is given approximately by:

$$\Delta D = h \tan \alpha \sin \alpha = \sqrt{\frac{g}{V_n}}$$

(13)

The correction, $\Delta T$, plotted versus $\frac{V_n}{V}$ for various values of the ratio is found in "Topographic Correction Curves," Technical Report #3 by G.H. Sutton and G.R. Bentley. $\overline{h}$ is known from the depth of water to the base line, and $V_n$ is determined approximately from the refraction line. $V_n$ the velocity of the layer causing the topography was taken as $\sqrt{g}$ since $\sqrt{g}$ in most cases consisted of unconsolidated sediment.

Whenever the depth under the shooting ship was unobtainable from the Edo depth records, the depth was calculated from the shot records. This was done by reading the time interval between the shot instant and the first bottom reflection to the shooting ship. The shot instant and depth of shot correction, expressed in seconds were added to the time interval and this sum was divided by two. This time represents the required time for the reflection to travel from the bottom to the shooting ship. Using the method of successive approximations with depth and mean vertical velocity, a depth was obtained that corresponded to the mean vertical velocity for that depth.
IV Geophysical Results and Conclusions

A. Individual Profile Results and Description

The following is a discussion of each profile, including the navigational location, the important features, and the seismic velocity and thickness results.

For the sake of comparison with the 1955 results, the results of this paper are tabulated using the terminology and velocity groups used by Officer et al (1956). The first group, A, with a range in velocity of from 1.6 - 2.0 km/sec. represent unconsolidated sediments. Group B, with a velocity range of 2.2 - 3.0 km/sec., represent more consolidated sediments up to lithified sedimentary rocks. Group C, from 3.2 - 4.2 km/sec. represent velocities which could be other sedimentary or volcanic rock. Group D are velocities which could represent either sedimentary, igneous or metamorphic rocks. Group E represents the velocity layer which is found above the major velocity discontinuity in the ocean. In keeping with the terminology used for the continents, this layer of material is called the crust. Finally, Group F represents the velocity found beneath the major discontinuity. This major discontinuity is identified by Officer et al (1956) as the Mohorovicic discontinuity found beneath continents. A summary of the resultant seismic velocities and thicknesses is given in Table I.

Refraction Profile 1: - Bn, 24° 12' N, 62° 15' W; Ag, 23 - 29' N, 62 - 04' W

This profile was taken in Atlantic Basin approach to the eastern Caribbean. The depth of water was of normal Atlantic basin depth around 3000 fathoms. The bottom topography was level except on the southern end.
The topography from both ships shows an abrupt peak of about 2000 feet high and 10 kilometers in diameter (see Figure 14). Except for this peak the topographic corrections were negligible. The records were unusually devoid of noise so that on only one shot was there any question as to the time of the refraction arrival. On the southern end three refraction layers were evident, but only the two higher velocity layers were detectable on the northern end. (see Figure 15). Due to this and the value of the apparent velocity on the southern end; it was decided that the peak was a topographic expression of a volcanic stock. Since it was impossible to exactly determine how much of the profile this stock covered, the profile was calculated by assuming that the apparent velocity on the southern end of the profile represented the true velocity of the volcanic material. The layer thicknesses of the northern end were then calculated by assuming that the second layer was unconsolidated sediment using the estimated sediment velocity method discussed in Section III. The layer thicknesses at the southern end were then calculated by assuming that the second layer was made up entirely of the volcanic material. Since the time intercept equation (7) for calculating thicknesses is relatively insensitive to velocity and wave layer angles, this is a valid method for determining the order of magnitude of layer thicknesses. The determined true velocities from the two calculations for the crustal velocity and for the velocity below the major velocity discontinuity were essentially identical and these velocities and thicknesses are tabulated in table I. The thickness of the crust shows an increase of 5.8 km to 7.11 km at the southern end with a velocity of 6.87 km/sec. The velocity below the major discontinuity is 8.21 km/sec. The thickness and the velocity of the crust
DIRECT WATER WAVE TRAVEL TIME IN SECONDS

ATLANTIS-BEAR REFRACTION PROFILE 1
JUNE 29, 1956

Figure 15
and the subcrustal velocity are higher than generally associated with the Atlantic Basin.

**Refraction Profile 3:** B_{SE}, 20° - 31° N, 61° - 52° W; A_{NW}, 20° - 16° N, 61° - 18° W

Profile 3 was taken on the northern flank of the topographic rise or ridge just north of the eastern extension of the Puerto Rico Trench. The profile paralleled the trend in that area. The bottom topography was extremely irregular especially at the northwest end of the profile, with a maximum variation in relief of about 3000 feet. (See Figure 16). The assumption that the bottom relief was an expression of topography in the first velocity layer measured reduced scatter to a minimum. The effects of the topographic corrections are shown in the detailed enlargement in Figure 17. The closed circles represent the arrival times after the topographic correction had been applied and the open circles represent the arrival times before the correction. An interesting side light to topographic reflection corrections is evidenced in this figure. Refraction arrivals that arrive at a time later than the first bottom reflection are usually hidden in this reflection. The first refraction arrivals in this figure are shown inside the reflection curve indicating that they came later than the reflection. This is not the case though, but it appears this way in the figure since the reflection curve has been corrected topographically to a base line lower than the bottom relief. Therefore these first few refraction arrivals would have been hidden in the reflection if it were not for the presence of this topographic high in the bottom relief. The exact time of the refraction arrival was questionable on only two shots due to noise.
JULY 1, 1956

DIRECT WATER WAVE TRAVEL TIME IN SECONDS
Three corresponding apparent velocities were measured on each end of the profile. (See Figure 17). The thickness of the unconsolidated sediment was determined from the time intercept of the first high velocity layer and the average velocity depth relation deduced from Lawhorn's reflection investigation. The rest of the thicknesses and true velocities were determined using the standard layer calculations. The third layer or the layer under the sediment had a velocity of 5.70 km/sec. The thickness of this layer decreases from 3.01 km to 1.67 km underneath the southeast end of the profile. The next or deepest layer had a velocity of 7.11 km/sec. This layer was taken as the crust by correlation with profiles 1 and 2, the latter being discussed in Part I. The thickness of the crust decreased from 4.94 km to 3.31 km at the southeast end. The velocity below the major discontinuity was 8.44 km/sec. The depth below sea level to this discontinuity decreased from 13.80 km to 10.97 km at the southeast end. The velocities of the crust and the material under the major discontinuity are higher than the corresponding velocities in Profile 1 and normal oceanic areas.

**Refraction Profile 5:** B, 18° - 01' N, 60° - 19' W; A, 18° - 33' N, 60° - 19' W

Profile 5 was taken near the axis of the eastern extension of the Puerto Rico Trench. The profile approximately paralleled the trend of the trench. The southeast end of the profile was slightly displaced over the southwest flank of the trench. The bottom topography showed moderate relief compared to the other profiles in this area. The southeast end was approximately 2000 feet higher than the northwest end. Topographic corrections using the lowest measured velocity were very effective in reducing scatter along the refraction lines. The seismic records were
BOTTOM TOPOGRAPHY
ATLANTIS-BEAR REFRACTION PROFILE 5
JULY 3, 1956

Figure 18
ATLANTIS-BEAR REFRACTION PROFILE 5
JULY 3, 1956

Figure 19
moderately noisy although on only one shot was the arrival time hidden in the noise. Three corresponding apparent velocities were measured on each end of the profile. (See Figure 19). The thickness of the unconsolidated sediments were determined by the method described previously. The three measured velocities and thicknesses were determined using the standard layer calculations. The third layer had a true velocity of 5.44 km/sec. The thickness of this layer decreased 4.03 km to 1.03 km at the southeast end. The deepest layer, which was taken as the crust, had a true velocity of 6.98 km/sec. The thickness of the crust decreased from 5.09 km to 4.94 km at the southeast end. The velocity below the major discontinuity was measured as 8.64 km/sec. The true velocity measured for the material below the major discontinuity here is the highest velocity measured in this area.

Refraction Profile 18: Bhp 19° - 24' N, 62° - 01' W; A by 18° - 43' N, 62° - 30' W

Profile 18 was taken on the southern flank of the deepest part of the eastern extension of the trench. The profile was approximately normal to the trench axis. The bottom topography was extremely irregular. The regional dip of the sea bottom was 1.63°. The south end of the profile was about 8000 feet higher than the north end. The maximum relief from the assumed base line is 4000 feet. (see Figure 20). The topography correction was applied assuming that the bottom relief expressed topography in the first velocity layer measured at both ends. This correction effectively reduced scatter on all refraction lines. (See Figure 21). A residual deviation is seen on the Vhb line. This probably represents buried topography in this layer, i.e. topography not reflected
on the sea bottom. The arrivals were frequently weak especially for the
deeper layers. The records were somewhat noisy, but fortunately the noise
and the arrivals were frequently spaced as to not coincide with each other.
Three apparent velocities were measured on the south end, but only two were
measured on the north end. The velocity corresponding to the crust in this
area was not measured. Since it seems a valid assumption that the crust
does not disappear, it was then assumed that the crust was a masked layer
due to an unusual thick layer above it of a much lower velocity. This is
indeed the case, and a normal thickness crust would appear as a masked
arrival. The crust layer was then constructed for calculation purposes
on the travel time graph using the measured velocity for that layer
obtained from nearby profiles (see dashed line Figure 21). The first
measured velocity on the south end did not have a corresponding velocity
on the deeper northern end. This is interpreted as indicating a thinning
of that layer at the north end. In addition there was a slight thickening
of the unconsolidated sediment above this layer toward the north. Since
the north end was located in the bottom of the trench, it seemed logical
that the unconsolidated sediment would be thicker there. By means of the
usual calculations for the unconsolidated sediment and assuming a small
thickness of the first refraction layer beneath it at the north end, the
thicknesses of the two layers were calculated using equation (7). Using
the measured velocities and time intercepts, and the constructed velocity
and intercept for the masked crustal layer, the thicknesses and the true
velocities were calculated for the remaining layers. The apparent velocity
for the thinned layer was measured as 3.48 km/sec. The true velocity for
the next lower layer was 4.62 km/sec. This layer thickness increased
from 6.21 km to 13.37 km at the south end. The constructed layer of crustal velocity, 6.73 km/sec, decreased in calculated thickness from 3.19 km to 1.18 km at the south end. The true velocity of the layer below the major discontinuity was 8.30 km/sec. The thickness for the crust and the layer above it cannot be taken as accurate, but the value for the depth to the major discontinuity will not change appreciably for any construction assumed for the masked crustal layer and can be accepted with confidence. From this value one can see that there is an appreciable amount of material above the discontinuity and that the discontinuity is lower at the south end of the profile than at the north end which is located over the deepest part of the trench.

Refraction Profile 29: B, 19° 05' N, 64° 05' W; A, 19° 07' N, 64° 53' W

Profile 29 was taken on the south flank of the Puerto Rico Trench approximately along the 2000 fathom bottom contour. The profile was parallel to the general trend of the trench in that area. The bottom topography was very irregular. The maximum change in relief was about 2500 feet. The western end of the profile was shot on a plateau of about 35 km in length. On the reversed leg of the profile the listening ship at the west end drifted off the edge of the plateau with a subsequent drop in the sea floor of about 1500 feet. (see Figure 22). Topographic corrections using the first measured velocity reduced arrival scatter to a minimum. The seismic records were noisy but all arrivals were readily visible. Three corresponding apparent velocities were measured at each end of the profile. The first velocity measured on the west end was seen as a first arrival for only three shots, but fortunately the velocity
BOTTOM TOPOGRAPHY
ATLANTIS-BEAR REFRACTION PROFILE 29
JULY 30, 1956

Figure 22
could be better determined from good second refraction arrivals. (See Figure 23). The thickness of the unconsolidated sediment layer was determined by the usual calculation. The layer below the unconsolidated sediment had a measured true velocity of 4.96 km/sec. The thickness of this layer decreased from 2.64 km to 1.65 km at the west end. On this profile two velocities were measured that represented crustal velocities. The upper velocity was 6.05 km/sec. The thickness for this layer decreased from 8.30 km to 3.93 km at the west end. The largest velocity measured on this profile was 7.03 km/sec. and is considered crustal also. The major velocity discontinuity of around 8 km/sec. was not measured on this profile. The depth to the last velocity discontinuity decreased from 15.40 km to 10.02 km.

B. Correlation of Results with 1955 and 1956 Cruise Profiles in This Area

The results of Profiles 1, 3, 5, 18 and 29 from this paper and the results of Profiles 2, 4, 6, 17 and 20 interpreted by D.E. Miller (1957) from the 1956 Cruise are shown in Tables 1 and 2 and Figures 24 and 25, respectively. The results of Profiles 42, 43, 44, 45, 46 and 47 from the 1955 Cruise are shown in Table 3. (see Figure 1 for profile locations).

The Atlantic Basin in regions which have not been deformed shows a uniform crustal structure. The depth to the major discontinuity is about 10 km below sea level. The velocity below the discontinuity is 6.0 km/sec. with small variation. The crustal material above this discontinuity has an average thickness of 4 - 5 km and velocity of 6.5 km/sec. Between the crust and the ocean is a layer of material with a
Figure 24. Calculated Layer Velocities And Thicknesses, Part II
Profile #2  Profile #4  Profile #6  Profile #17  Profile #20

Depth in Kilometers

5

10

15

20

Note: Underlined velocities are estimated. Velocities in parentheses are averaged. Velocities in brackets were assumed.

Figure 25. Calculated Thicknesses And Depths, Part I
thickness of a kilometer or less. This layer generally has a velocity associated with sediments.

Profiles 1 and 2 were taken north of the eastern extension of the Puerto Rico Trench on a line from Bermuda. Previous profiles north of Profile 1 have shown little variation from the above Atlantic Basin structure. Both Profiles 1 and 2 show an increase in depth to the major discontinuity of about 2 km. The velocity below this discontinuity in Profile 1 is 8.21 km/sec, which is higher than the velocity associated with ocean basins. While in Profile 2 this velocity is 7.96 km/sec, which is within the variation seen in ocean basins. Both profiles show an increase in crustal velocity to 6.87 km/sec. for Profile 1 and 6.93 km/sec. for Profile 2 over the ocean basin crustal velocity of 6.5 km/sec. Besides the upper low velocity layer, which is less than 1.5 in thickness in both profiles, Profile 2 shows an extra layer between the crust and the uppermost low velocity layer. This layer has a velocity of 5.96 km/sec, and increases in thickness in the direction of the trench from 1.3 km to 1.7 km. The crust in Profile 1 shows a much greater thickness than Profile 2, but the combined thickness of the crust and the 5.96 km/sec. velocity layer of Profile 2 is approximately equal to the crustal thickness in Profile 1. The velocity of 5.96 km/sec. is characteristic of either extremely lithified sedimentary rock, igneous, or metamorphic rock. This suggests to the writer two possible explanations of the above results: (1) The two profiles together represent a transition zone from the typical ocean basin structure to the structure found on the northern edge of the trench and the 5.96 km/sec. layer was missed on Profile 1, or (2) Profile 1 represents a thickening of the crust caused by the intrusion of material
from deep in the mantle and Profile 2 represents a later stage where the more volatile constituents of the intruding magma have worked their way to the top of the crust and 5.96 km/sec. layer represents this combination of crustal material, igneous intrusives and the metamorphism of the deeper sediments by the volatile intrusives.

The Puerto Rico trench has a maximum depth of 4400 fathoms north of Puerto Rico. The axis of the trench follows the general trend of the island arc to the southeast with diminishing depth. The actual topographic relief of the trench extends just beyond the series of profiles seen on Figure 1 as profiles 4, 5, 6, 7 and 17. At the top of the northern flank of the extension of the trench north of Anguilla, there is a ridge which extends in a general westerly direction from the deepest extension of the trench to beyond the limits of Figure 1.

Profile 3 was singular in that it was located just in the northern slope of this topographic high or ridge and no previous profiles have been taken here or on the ridge itself to the knowledge of the writer. The southeastern end of the profile was located farther on the ridge than the northwest end of profile. The end of the profile that is farthest on the ridge shows a slight decrease in thickness for the crustal layer and a larger thinning for the 5.70 km/sec. layer above the crust over the northwest end of the profile. The velocity below the major discontinuity has increased to 8.44 km/sec. in this profile, and the depth to this major discontinuity has decreased toward the ridge. This suggests to the writer that the ridge is a topographic expression of upwarp in the crust with perhaps thinning of the velocity layer above the crust. A horizontal compression on the crust in the trench and island arc area as postulated by Officer et al. (1956) could produce the crustal upwarp observed here.
Profiles 4, 5, 6 and 7 are a series of profiles taken essentially parallel to the axis of the eastern extension of the trench near the extreme southeastern end of the trench. At the time this paper was written Profile 7 had not been interpreted. Profile 17 is normal to the axis of the trench and approximately bisects Profiles 5 and 6 which are located on the southwestern flank of the trench. Profile 4 is located about midway on the northwestern flank of the trench coming off the ridge. Profiles 4, 5 and 6 show an increase in depth to the major discontinuity as one proceeds down the flank of the trench from the ridge and up the southwestern flank toward the island arc to the 3000 fathom bottom contour. Profile 17 shows this increase in depth also. The maximum depth indicated by the profiles in the area is 21 km at the southwestern end of Profile 17. Profiles 4, 5 and 6 show that the crust is thicker in this trench area than is found north of the ridge. The average thickness here is about 6.0 km. The decrease in thickness under the southwest slope shown in Profile 17 cannot be taken as accurate since this crustal thickness was calculated from a constructed masked crustal layer. The layer above the crust in Profiles 4, 5, 6 and 17 shows an increase in thickness from 2.0 km at the northwest end of Profile 4, which is the highest station on the northeastern flank of the trench, to 7.4 km on the southwest end of Profile 17 which is highest station in this vicinity on the southwestern flank of the trench. The layer above this shows the same general trend with the greatest thickness measured at the southwest end of Profile 17. Now if a crustal downwarp was created by the compression forces postulated by Officer et al (1956) to explain the trench, and if intense sedimentation and volcanism took place from the island arc which was also formed by the compressive forces, one would expect the original trench or
downwarp to be filled by these processes from the island arc side of the
trench with a subsequent apparent shifting of the topographic axis of the
trench away from the island arc flank, and the accumulation of great
thickness of material underneath the flank of the trench on the island arc
side. This is identical to what was seismically measured in Profiles 4,
5, 6 and 17. These profiles indicate that the axis of the crustal downwarp
was underneath the southwest flank of the trench which is the flank of the
trench on the island arc side and that the greatest thickness of material
is above the axis. Profiles 4, 5 and 6 also show a decrease in the depth
to major discontinuity and the thickness of the crust and the layers above
the crust at the southeast end indicating that the decrease in depth of the
topographic trench in that direction is an expression of the structural
trend of the structural downwarp, i.e., the structural downwarp is decreas¬
ing in that direction also.

The velocity of the material below the major discontinuity in
Profiles 4, 5, 6 and 17 was considerably higher than the 8.0 km/sec. found
in ocean basins. This velocity ranged from 8.35 km/sec. to a maximum of
8.64 km/sec. under Profile 5 which is located over the deepest part of
the topographic trench. The crustal velocities measured in Profiles 4, 5,
6, and 17, show an average velocity of about 6.8 km/sec. which is well
above the oceanic crustal velocity of 6.5 km/sec. The lowest crustal
velocity measured was 6.62 km/sec. in Profile 6. The rest were from
6.92 km/sec. to 7.03 under Profile 17. The velocity in the layer above
the crust in Profiles 4 and 5 increased from 5.44 km/sec. under Profile
4 to 5.66 in Profile 5. The velocity for this layer in Profile 17 was
5.48 km/sec. On the other hand, the velocity for this layer in Profile
6 was 4.75 km/sec.
This abrupt change in velocity between Profiles 5 and 6 is difficult to explain. The decrease in velocity in the layers above the crust is also seen between Profiles 20, 5.55 km/sec. and 29, 4.96 km/sec. where both profiles are on the island arc flank of the Puerto Rico trench and of the two profiles, 29 is closest to the island arc. It is possible that this change in velocity is somehow related to the fracture zone indicated from earthquake seismology as a zone of major earthquakes, which is well defined for the numerous Pacific Island Arcs and the Antilles Island Arc. This zone as determined by earthquakes starts at the surface between the trench and the island arc and dips at an angle of approximately 45° beneath the island arc. Another, and perhaps more likely possibility is that this change in velocity represents a change in the stratigraphy of the sedimentary and igneous materials.

Profile 18 was taken normal to the axis of trench from the deepest part of the eastern extension of the trench to a point near the 3000 fathom bottom contour on the southern flank of the trench. Its purpose was to correlate the data obtained on Profiles 42, 43 and 44 of the 1959 cruise, which are approximately parallel to the general trend of the trench. (see Figure 1). In this area, the major discontinuity of around 8.0 km/sec. or higher was seen on only Profile 18 and Profile 44 which is located almost at the bottom of the trench. On Profiles 42 and 43 the highest velocity obtained was 5.49 km/sec. and 5.47 km/sec. respectively. The depth to the major discontinuity is seen to increase in Profile 18 from 17.5 km to 22.1 km at the south end, with an increase in the total thickness of the crust and the velocity layer above it of from 10.9 km to 14.6 km at the south end. The fact that neither the crust nor the major discontinuity was seen in Profiles 43 and 44 indicates
to the writer that there is an extreme thickness of material between the top of the 5.5 km/sec. layer down to the discontinuity giving further evidence that the axis of the crustal downwarp is beneath the southern or island arc flank of the topographic trench. The velocity below the major discontinuity is 8.0 km/sec. in Profile 44 and 8.30 in Profile 18. The crustal velocity in Profile 18 is 6.73 km/sec. and in Profile 94, it is 6.33 km/sec. The velocity, 5.5 km/sec. corresponds to the velocity for the layer just above the crust found in profiles north of the trench axis. This layer has a velocity of 4.82 km/sec. on Profile 18 and is missing entirely on Profile 44.

In the Puerto Rico Trench area, five profiles were taken in all during the 1955 and 1956 Cruises. They were Profiles 45, 46 and 47 in the bottom of the trench and Profiles 20 and 29 on the southern flank of the trench. (see Figure 1). Profiles 46 and 47 are west of the series of Profiles 45, 20 and 29, which parallel the axis of the trench.

Profile 46 shows a depth to the major discontinuity of 16.8 km at west end and 19.4 km at the east end. The thickness of the crust increased from 6.8 km to 8.5 km at the west end. As seen in Figure 1 the east end of the profile is in the deepest part of the trench and the west end is a short distance up the northern flank. In profile 47 the deepest velocity obtained was the 5.95 km/sec. layer. The sedimentary and igneous layer above it increased in thickness to 3.8 km at the southern end. To the east of profiles 46 and 47, profile 45 showed a depth to the major discontinuity of 9.9 km at the west end and 18.3 km at the east end of the profile. The crustal thickness in profile 45 changed from 2.2 km to 10.3 km at the east end. The sedimentary and
Igneous layer above the crust is not seen at the west end, but has a thickness of 1.9 km at the east end. This apparent decrease in the sedimentary-igneous layer and the crustal layer and the rising of the major discontinuity between profiles 45 and 46 is evidenced in the bottom topography as a decrease in trench depth between profiles 45 and 46 and subsequent narrowing of the trench there. The velocity below the major discontinuity in the deepest parts of the trench is seen to range from 7.90 to 7.95 km/sec. The crustal velocity is from 6.24 to 6.32 km/sec, which is a little lower than the average of 6.5 km/sec. for the Atlantic.

Proceeding from profile 45 up the south flank to profile 20 the depth to the major discontinuity is seen to increase to 22.2 km at the eastern end of profile 20. The thickness of the crust is seen to decrease from profile 45, average thickness of 6.2 km, to an average thickness for profile 20 of 3.7 km. The thickness of the sedimentary to igneous layer is seen to increase from a maximum 1.9 km for profile 45 to a maximum of 10.1 km for profile 20. The highest velocity seen in profile 29, which is the highest profile on the southern flank is 7.03 km/sec, and is taken to represent the crust by this writer. Therefore no values can be given for the depth of the major discontinuity or thickness of the crust on this profile. The sedimentary to igneous layer shows an average thickness of about 6.5 km, which is a decrease in the average thickness of this layer from profile 20 to profile 29 of from 10.5 km to 6.5 km. The above indicates that in this area also the axis of the crustal downwarp is under the island arc flank of the trench with a large thickness of material above it.

The velocity of material beneath the major discontinuity.
shows an increase from profile 45 to profile 20 of from 7.90 km/sec. to 8.43 km/sec., respectively. The crustal velocity shows an increase in velocity from 6.24 km/sec. for profile 45 to 6.79 km/sec. for 20 to 7.03 km/sec., for 29, as one proceeds up the southern flank. The velocity layers above the crust also show an increase.

C. Conclusion

All of the seismic results obtained on the 1955 and 1956 cruise indicate that there is, indeed, a crustal downwarp to a depth of about 20 km of which the trench is a topographic expression, and that its axis is located under the island arc flank and not under the trench axis. Evidence other than seismic can be given for this offset of the crustal downwarp to the island arc side of the trench. From the gravity measurements of Vening Meinesz (1948) there is a large negative gravity anomaly associated with the trench, and a positive anomaly associated with the adjacent island arcs. The axis of the negative anomaly is offset from the axis of the trench to beneath the flank of the trench toward the island arc. (see Figure 1).

Vening Meinesz interpreted this anomaly as representing a great plastic buckling of the crust which formed a large downbuckle of lighter material into the more dense material beneath the crust. None of the 1955 or 1956 seismic results show the large crustal downbuckle necessary to give this result. The maximum downwarp of the crust measured was to a depth of 22 km. This represents only an increase in depth of about 12 km over the normal oceanic depth of 10 km. However, Ewing and Worzel (1954) demonstrated that the large negative anomaly
could be explained also by a smaller thickness of lighter sedimentary material near the ocean bottom. From this, Officer et al (1956) concluded that the large thickness of the lighter 5.5 km/sec. layer above the crust on the island arc flank might contribute appreciably to this negative anomaly. This layer of around 5.5 to 4.5 km/sec. was found on the 1956 cruise also and was evidenced by extremely large thicknesses. The maximum thickness was found under the island arc flank of the trench and was approximately 10 km under profile 20. It seems conclusive from these results that the large thickness of 4.5 to 5.5 km/sec. material contributes appreciably to the negative gravity anomaly.

In general, the seismic results of 1955 and 1956 cruises show that, as one proceeds from the Atlantic Basin to the trench, there is an increase in depth and layer thickness of the normal oceanic crustal structure with the addition of 5.5 km/sec. layer in the vicinity of profiles 1 and 2. This trend continues to a point beneath the island arc flank, where the depth of the crust begins to decrease to the depth found beneath the island arc. The maximum thickness of material above the crust is found at that point.

The results of the 1956 cruise show that the velocity of the material found under the major discontinuity increases from 8.0 km/sec. to a maximum of 8.64 km/sec. beneath the trench as one approaches the trench from the Atlantic Basin. This overall increase in velocity is also seen in the crust although there is some variation from profile to profile. The 8.64 km/sec. velocity found in profile 5 for the material below the crust is unusually large, but even discounting this velocity the rest of the 1956 profiles measure around 8.3 to 8.4 km/sec. in the
trench area.

The velocities for the material below the crust, 7.9 km/sec. and the crustal velocity 6.3 km/sec., measured on the 1955 profiles in this area are almost identical with the average velocities found in the ocean basins of 8.0 km/sec. and 6.5 km/sec. respectively. If anything, the 1955 results in the trench are lower than those found in oceanic layers. This inconsistency in velocity results between the 1955 and 1956 cruises is difficult to explain, unless the differences are due to inconsistencies in interpreting the 1955 and 1956 profiles, or that the inconsistencies are due to intrusions from below the crust of variable composition or in different stages of development.
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*Table 3. Summary 1955 Seismic Results, Puerto Rico Trench And Vicinity*
REFERENCES


