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Seismic Refraction Studies in Baffin Bay: An Example of a Developing Ocean Basin*

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Summary

Sonobuoy and tape recording buoy seismic refraction measurements were carried out in Baffin Bay and Davis Strait to study the extent and tectonic development of the oceanic region and the structure of some of the major features of the surrounding continental shelves. Both the oceanic and continental shelf areas are occupied by thick sequences of sediment, 3–7 km, the sediment pile being thicker in the north. A refraction profile in Davis Strait shows that it is underlain by a crust similar to that beneath Iceland, with a total crustal thickness of over 20 km. The main difference is that a normal mantle velocity is measured beneath the Davis Strait. The results, supported by a seismic reflection profile in the region, suggest that Davis Strait may once have been the site of a hot spot or upwelling mantle plume. The main oceanic crustal layers, layers 2 and 3, under the central basin are thin. A total main crustal thickness of 4 km was measured (omitting the sediment, that is) compared with a mean thickness of 7 km for the major ocean basins. The crust is underlain by mantle rocks exhibiting normal mantle velocities, the mean being $8 \cdot 0 \text{ km s}^{-1}$. Seven refraction lines distributed over the east–west extent of oceanic crust show no detectable median ridge. This accords well with models of the decrease of ridge topography after spreading has ceased as the thermal anomaly beneath decays and supports the hypothesis that the area has not been spreading for about 50 My. The anomalously thin crust can also be related to its age. Thin crust is found near the active mid-ocean ridges at distances corresponding to ages between 50 and 80 My. We postulate that the crust in Baffin Bay is not fully developed and in time a thicker crust will form. These results support the hypothesis that at least the lower part of layer 3 of the ocean basins is composed of altered mantle material.

1. Introduction

This paper describes the results of seismic refraction experiments in Baffin Bay and Davis Strait and discusses their significance in relation to the tectonic development of this area. Previous geophysical studies in Baffin Bay have shown that the deep central basin of the Bay is underlain by oceanic crust and therefore that the area may have been formed by sea-floor spreading (C. Keen *et al.* 1972). However, no evidence of a central buried ridge, such as that in the Labrador Sea, or of magnetic lineations

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normally associated with the spreading process was found (C. Keen *et al.* 1972; M. Keen, Johnson & Park 1972). Heat flow is normal in the central part of the Bay (Pye & Hyndman 1972). The structure beneath the Davis Strait sill, separating the Labrador Sea from Baffin Bay, had not been defined.

A tentative hypothesis for the evolution of the Baffin Bay area was suggested by C. Keen *et al.* (1972). They proposed that the central basin was created by sea-floor spreading between 80 and 50 My ago when the Labrador Sea was formed (Laughton 1971). A reconstruction of the initial positions of Greenland and North America was presented which involved two phases of spreading, about two poles of rotation. The Davis Strait sill, they suggested, may have originated during a change in spreading direction about 60 My ago. The reconstruction also shows that the Melville Bay graben, a major structural feature on the west Greenland continental shelf, may once have been joined to one of two similar structures between the Canadian Arctic Islands, Lancaster Sound or Jones Sound.

Further geophysical measurements were necessary to support the above ideas and therefore from August to October 1971, CSS *Hudson* carried out a survey in the area. The observations consisted of bathymetric, magnetic, gravity, seismic reflection and refraction measurements and of several dredge, core and bottom camera stations. Satellite navigation was used for positioning. The main objectives of the cruise were:

(1) To determine if a basement ridge exists beneath the central basin. Previous seismic reflection and refraction measurements show that there is about 4 km of sediment in the centre of the Bay. The first multiple on the reflection records obscures any basement structures beneath. Therefore a series of seismic refraction lines across the area, defining regional variations in the depth to basement, is the best means of detecting a ridge.

(2) To define the crustal structure beneath the Davis Strait sill.

(3) To trace the Melville Bay graben northward, to determine the thickness of sediment within it and to compare sedimentary structures found there to those in Lancaster and Jones Sounds.

(4) To determine if any structural features in Smith Sound were offset or truncated by a possible transform fault running through Nares Strait.

(5) To define the offshore extent of the Tertiary basalts which occur on land in the vicinity of Cape Dyer, Baffin Island. A similar study of the volcanics off Disko Island, Greenland, has been made by Park *et al.* (1971).

This paper is concerned with the seismic refraction measurements collected during the cruise with infrequent references to the other geophysical observations. The locations of the refraction lines are shown in Fig. 1.

2. Experimental methods

Three types of refraction measurements were made. First, reversed lines, up to 55 km long, were obtained using telemetering sonobuoys manufactured by G. & E. Bradley Ltd in a similar manner to that described by Hill (1963). Second, an unreversed line, 155 km long, was obtained in Davis Strait with a buoy which contained a tape recorder, programmed to switch on and off at predefined intervals. This system removes the limitation on shot-receiver distance imposed by the transmission link. Finally, expendable sonobuoys were used in conjunction with a seismic reflection system to obtain short, unreversed lines.

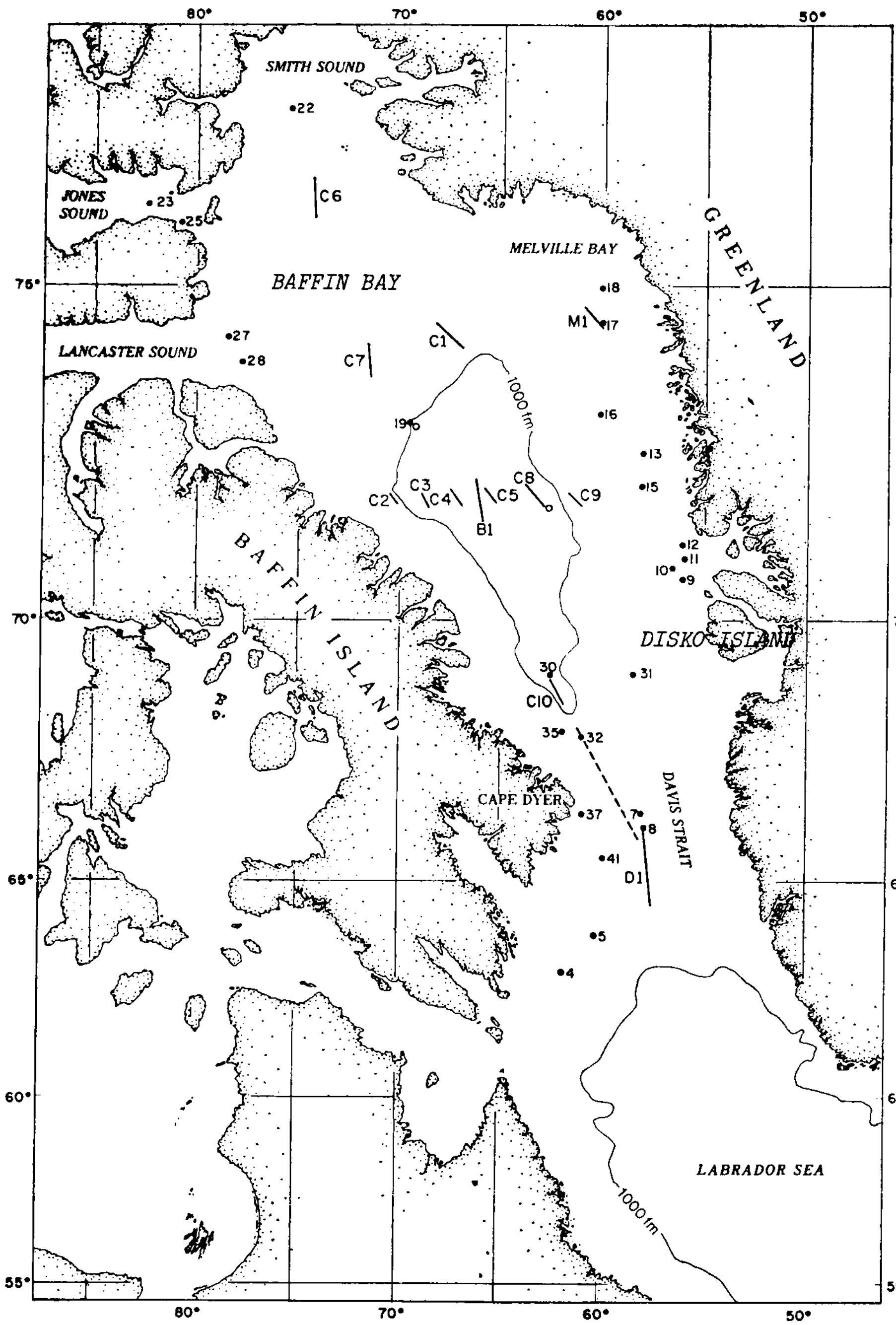

Fig. 1. Location of the seismic refraction lines. The solid lines are reversed refraction profiles. The black dots are expendable sonobuoy positions and the open circles, the positions of the sound velocimeter stations. The dashed line is the seismic reflection profile referred to in the text. The 100 fm contours are shown in Baffin Bay and the Labrador Sea.

One or two buoys were laid at each end of the reversed lines. These were free-floating to reduce flow noise at the hydrophones. When two buoys were laid they were spaced about 2 km apart. Two sound sources were used: a Bolt air gun with a 1000 cubic inch chamber and with a pulse shaper, and explosive charges. The air gun was used to define the velocities of the upper layers beneath each end of the lines by firing along them every 16 s. Explosive charges were shot between the buoys, or pairs of buoys, to obtain reversed velocities for the deeper layers. The charge sizes ranged from 25 to 450 lb and were fuse-fired at a 5- or 10-min interval. We found that the combination of the air gun and explosive sound sources gave more information than could be obtained with explosives alone. Barrett *et al.* (1971) described the first refraction line in Baffin Bay and it is evident in their discussion of the results that

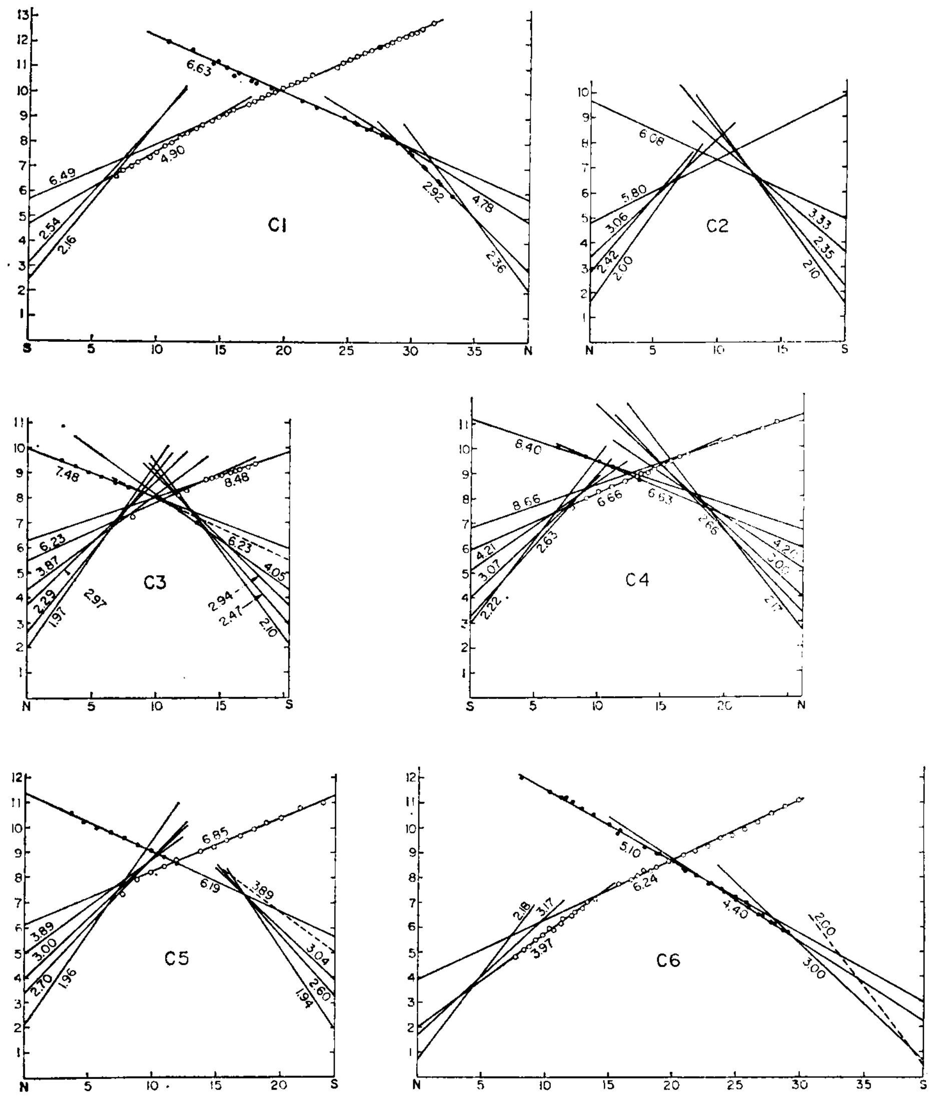

FIG. 2. The time-distance plots for some of the reversed refraction profiles. Both the time and distance scales are in seconds. Other numbers are velocities in kilometres per second. The lines with no data points are refraction results using the air gun. These were originally recorded on a seismic profiler recorder.

some layers are not well defined by first arrivals. In most cases, second arrivals did not appear to correlate sufficiently well from shot to shot to be useful. This earlier experiment was performed using explosives. With the rapid firing rate of the air gun many velocities are apparent which are only defined by later arrivals.

The drift of the buoys during an experiment was typically 4 km. However, in one case the buoys drifted about 7 km. Drift is not considered to be a serious source of error because all the lines were obtained in areas of smooth bottom topography and corrections for variations in the depth of water under the moving buoys would not exceed 0·05 sec. These corrections were however, applied for bathymetric variations beneath the shots along each line. The shot instant was recorded by a geophone on the ship and a correction was made to allow for the distance between the ship and the shot at the time of detonation.

The tape recording buoy was used for receiving long range (over 60 km) explosive shots, fired every 20 min. The tape recorder was a Uher 4000 Report L. A crystal clock with an accuracy of 1 part in 10^7/day was used for timing, and for switching the recorder on and off. The clock was compared with the ship's clock and a WWV time signal just before the buoy was launched and immediately after its recovery to ensure that its drift had been negligible over the period of the experiment. Corrections to the travel times were applied as described above.

Two types of expendable sonobuoys were used, modified Hermes AN/SSQ–41A and Aquatronics SM–44. An Eddystone f.m. receiver and a seismic profiler recorder were used to record the signals. The maximum useful range of these systems was typically 10 km. The Bolt air gun was used as the sound source, firing at a rate of 16 s. The lines were unreversed. Because of the short distances over which signals were received the variations in bottom topography were negligible and no corrections were applied.

Two sound velocimeter stations were completed in Baffin Bay (Fig. 1), both of which gave similar results. These results are also in agreement with a previous measurement made at one station in central Baffin Bay (Barrett *et al.* 1971). The velocity of sound in sea water used to calculate distances from the water wave arrival times is 1·47 km s^{-1}.

3. Interpretation

The time–distance plots corresponding to the reversed lines and to the one long unreversed line are presented in Figs 2–4. They include the results from both the explosive charges and the air gun; most of the shallow layers being defined only by the short air-gun lines near each buoy or set of buoys. Two of the original time–distance plots, displayed on a continuous seismic recorder, are shown in Fig. 5 to indicate the quality of the air-gun data. Several sources of error are present in the interpretation of the air-gun lines. Many of the velocities, particularly the higher ones, are measured over short distance intervals, typically 2 to 3 km. Structural variations with wavelengths greater than or equal to distances of this magnitude will cause errors in the measurement of these velocities. Also, these lines do not allow the shallow layers to be properly reversed. When the same layer was detected by both the explosive sources and the air-gun along the same line both results produced similar velocities, to within 0·2 km s^{-1}, but a higher time intercept was observed for the air-gun results. This difference suggests that as many as two full cycles of the arriving signals were not sufficiently large to be detected on the display of the air-gun results. When possible, a correction between 0·10 and 0·25 s was aplied to the air-gun time intercepts to make them compatible with the explosive measurements. However, some of the lower velocities apparent in the air-gun results were not observed on the explosive records so that the estimated errors for the intercepts of these upper layers may be as much

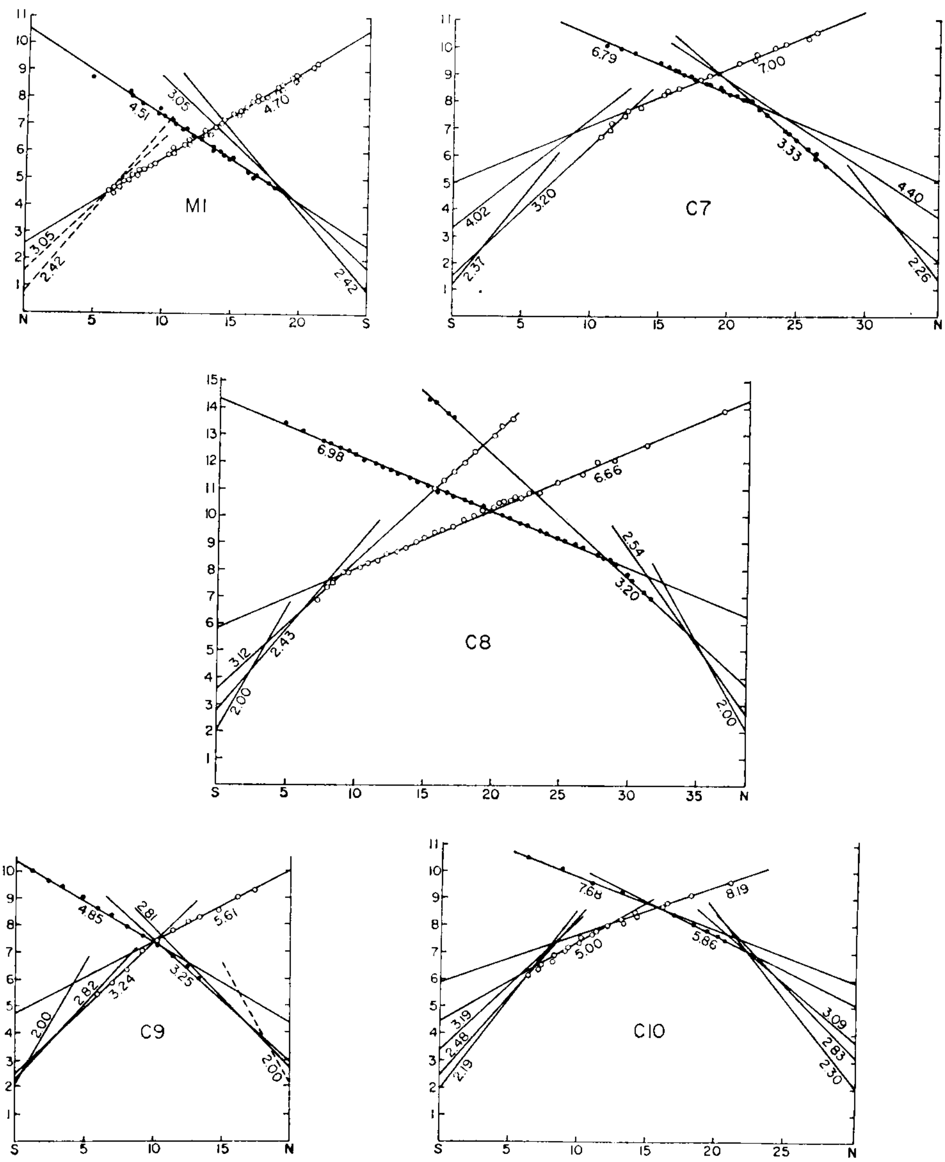

FIG. 3. The time–distance plots for some of the reversed profiles. Notation is the same as in Fig. 2.

as 0·1 s. On some of the air-gun lines (Fig. 5), it was possible to fit the refraction velocities to be tangential to the corresponding reflection curves. This affords additional confidence in the interpretation of the results.

The complete time–distance plots (Figs 2–4) are presented as reversed lines and were interpreted as though all of the layers were properly reversed. Although this treatment is not strictly justified, there were in general no substantial differences in the apparent velocities and time intercepts observed at the two ends of the lines. Least squares straight lines were fitted to the explosives results to obtain the best fitting velocities and intercepts. The explosive data are well reversed; the reversal points not differing by more than 0·2 s. Occasionally, a velocity was observed at one

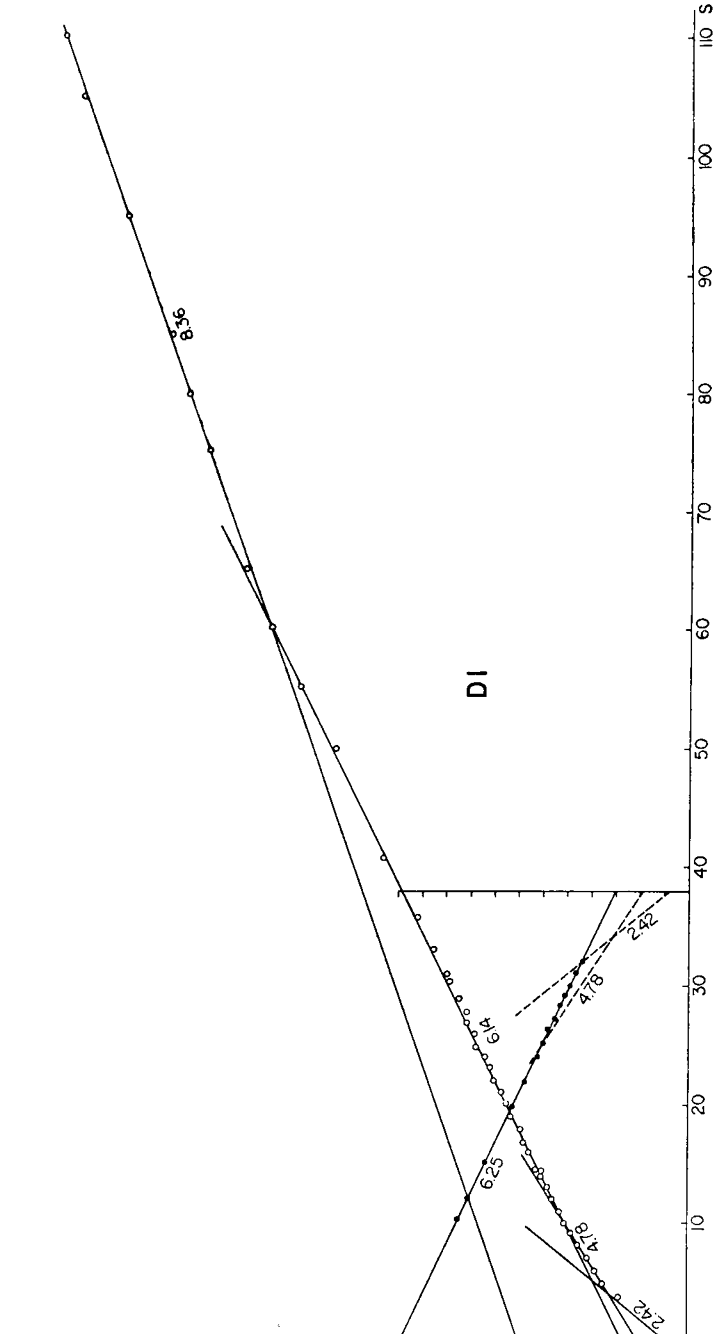

FIG. 4. The time–distance plot for line D1. Notation is the same as in Fig. 2.

end of a line, but not at the other. It seems reasonable to assume that this is not due to its absence but to the complexity of the air-gun, time–distance plots which often made it difficult to resolve two lines, representing two layers where velocities differ by about 1 km s^{-1}. Therefore, if a layer was observed at one end of a line, the same layer was assumed to be present beneath the other end of the line and the same apparent velocity and intercept were used in the calculations. When a layer was assumed in this way, the corresponding line on the time–distance plot is dashed in Figs 2–4. The lowest velocity, corresponding to unconsolidated sediments directly beneath the sea floor, was often not observed. In this case a velocity of 2·0 km s^{-1} was assumed for the uppermost layer. Calculations of layer velocities and thicknesses were made using the equations of Ewing (1963).

Two examples of the results obtained with the expendable sonobuoys are shown in Fig. 6. In interpreting these data refraction arrivals were mainly used. Some attempt was made to use reflected arrivals as others have done (Le Pichon, Ewing & Houtz 1968). However, in many cases no clear reflection curves could be seen on the records. Standard techniques for interpreting unreversed lines were used.

4. Results

4.1 *Central Baffin Bay*

The positions of the lines C1 to C10 are shown in Fig. 1, and the time–distance plots, in Figs 2 and 3. All of these lines, except C6, were in deep-water areas, greater than 1000 m, and it is these which we shall discuss below. Also, they are in the area occupied by oceanic crust as defined by the gravity measurements (C. Keen *et al.* 1972; Ross & Manchester 1972). The lines were obtained to define the general properties of the Baffin Bay oceanic basin and to determine if a ridge could be detected in the centre of the Bay. To accomplish the latter, lines C2 to C5 and lines C8 and C9 were shot at about 50-km intervals from west to east across the deep central region. The refraction line, line B1, described by Barrett *et al.* (1971) can be added to the results of this traverse. Its position is also shown in Fig. 1.

The results along this section are presented in Fig. 7 and Table 1. On three of the lines including B1 a depth to the *M* discontinuity was obtained and on most of the other lines a minimum depth to this interface is shown. The measured mantle velocities range from 7·7 to 8·5 km s^{-1} which implies that there is no evidence for ‘ anomalous ’ mantle material in the central part of the Bay as many investigators have shown to occur beneath some active and inactive mid-ocean ridges (Talwani, Le Pichon & Ewing 1965; Drake *et al.* 1963). The remaining velocities can be divided into the following four groups:

Unconsolidated and semi-consolidated sediment: V = 1·9 to 3·2 km s^{-1};

Consolidated sediment: V = 3·9 to 4·2 km s^{-1};

Oceanic Layer 2: V = 5·0 to 6·3 km s^{-1};

Oceanic Layer 3: V = 6·5 to 6·9 km s^{-1}.

This division may seem somewhat arbitrary, particularly the association of velocities between 3·9 and 4·2 km s^{-1} with consolidated sediment and of velocities as high as 6·3 km s^{-1} with layer 2. It is difficult to correlate these measurements with rock types or layers found from refraction measurements in other oceanic areas (Ludwig, Nafe & Drake 1970). However, it seems unlikely that velocities of about 4 km s^{-1} correspond to layer 2 as it is commonly observed in the Atlantic, for example, because the top of this layer should then be visible as a strong, rough reflector above the multiple reflection from reflection profiling results across Baffin Bay. This

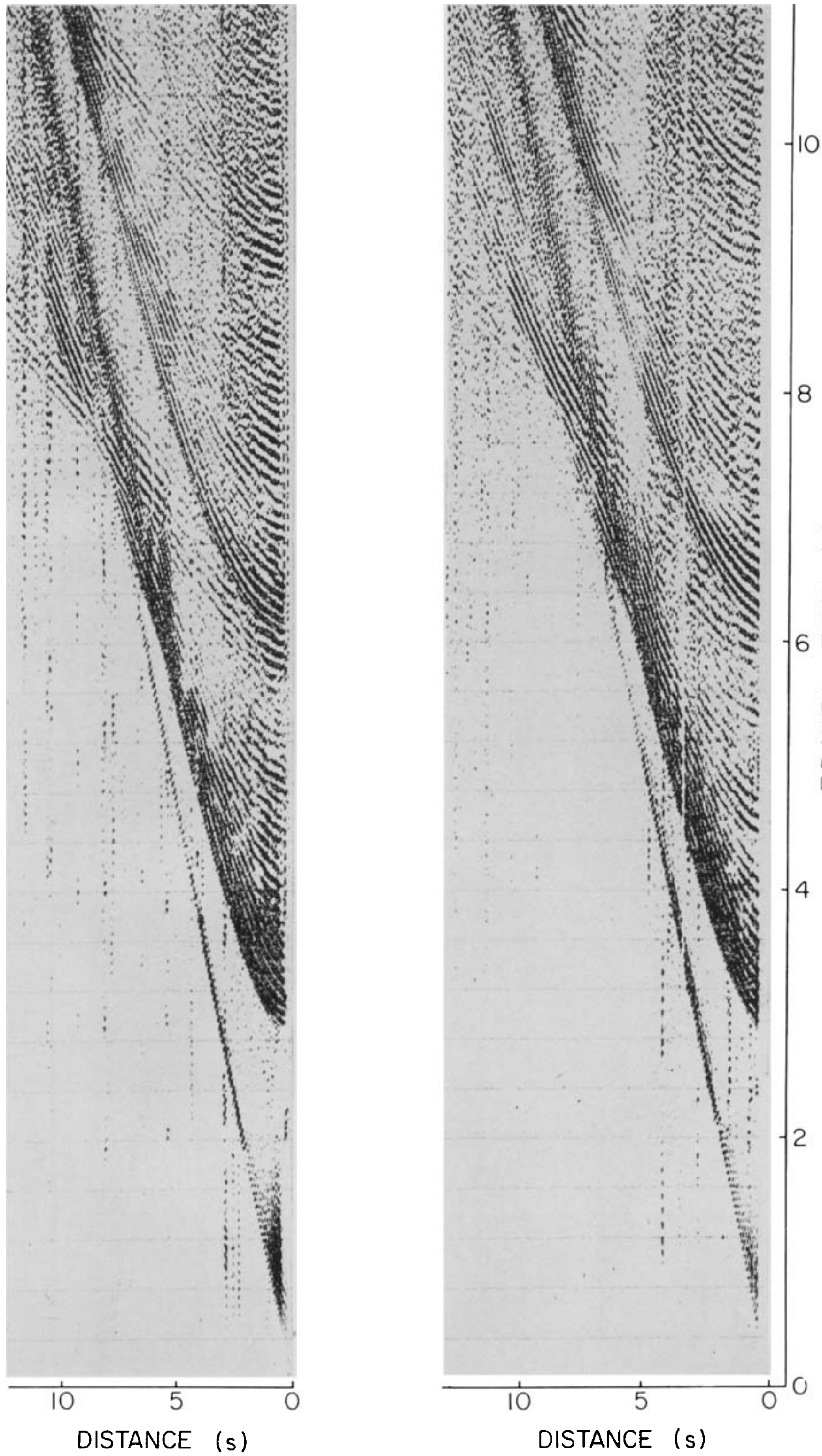

FIG. 5. Two examples of the original air gun lines. The examples are taken from lines C5 and C9.

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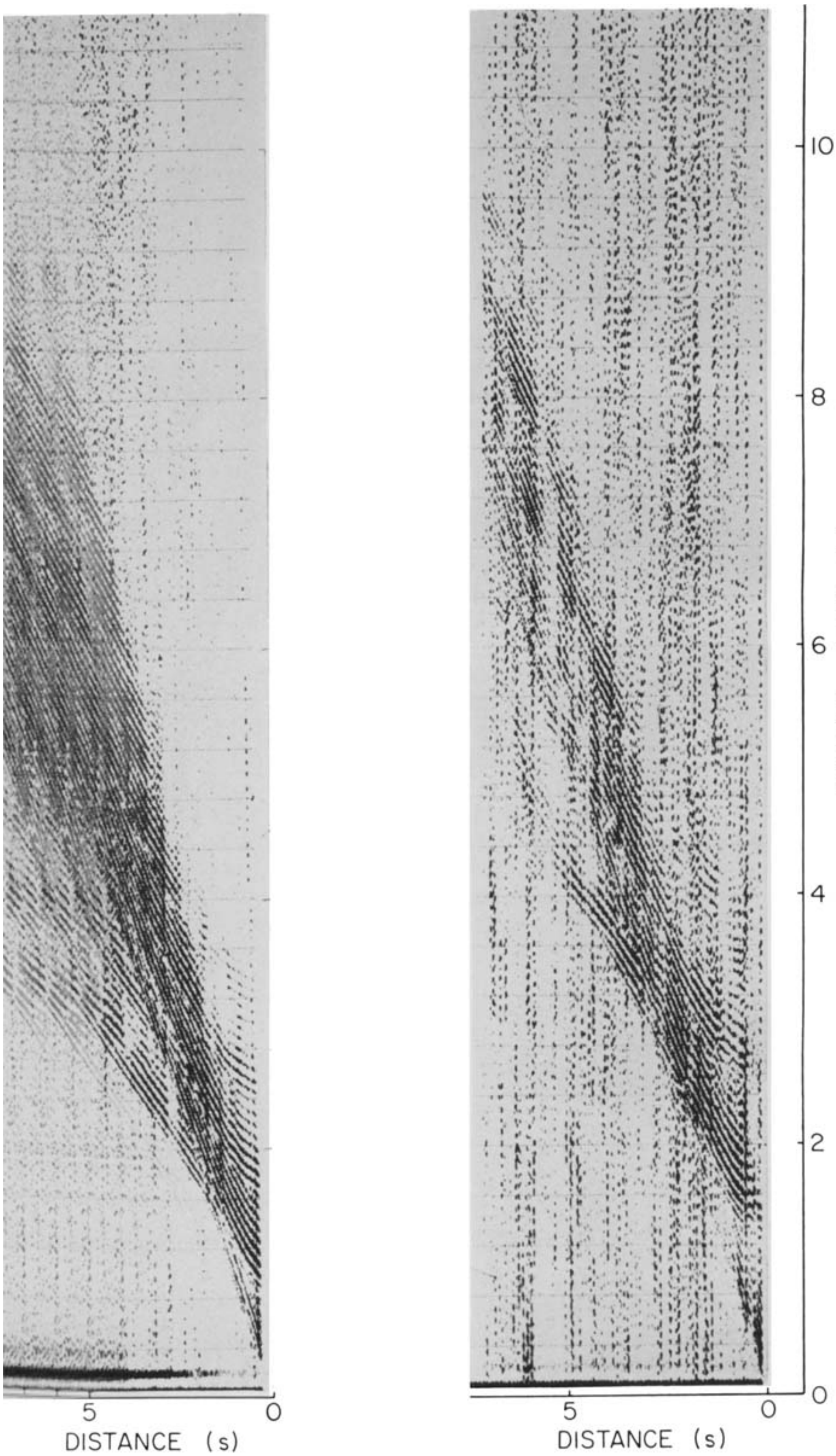

Fig. 6. Two examples of the original recordings of expendable sonobuoy lines 22 and 27.

Table 1

Refraction results from reversed lines

Line		Velocity (km s^{-1}) A	B	C	D	E	F	Thickness (km) Water	A	B	C	D	E	Location Lat.	Long.
C1	N	2·20	2·70	4·81	6·54			2·09	0·92	2·30	2·26			74° 32′	68° 08′
	S							2·09	1·37	1·40	2·77			74° 10′	66° 46′
C2	N	2·10	2·38	3·18	5·95			1·68	0·77	1·67	1·30			72° 06′	70° 27′
	S							1·68	1·94	0·00	1·71			71° 56′	70° 03′
C3	N	2·10	2·37	2·95	3·95	6·22	7·90	2·08	0·78	1·43	0·00	2·21	3·30	72° 06′	68° 54′
	S							2·08	1·32	0·55	0·55	2·11	0·93	71° 52′	68° 31′
C4	N	2·20	2·64	3·03	4·20	6·64	8·53	2·30	1·51	0·74	1·32	0·80	2·94	72° 10′	67° 29′
	S							2·30	1·29	1·13	1·01	0·99	3·82	71° 53′	66° 53′
C5	N	1·95	2·65	3·03	3·89	6·50		2·31	1·00	1·01	1·38	1·32		72° 11′	65° 50′
	S							2·31	0·92	1·06	1·58	0·06		71° 55′	65° 09′
C6	N	2·20	3·08	4·16	5·51			0·41	1·85	0·07	5·59			76° 24′	74° 10′
	S							0·41	0·23	3·49	1·15			75° 53′	74° 08′
C7	N	2·30	3·26	4·20	6·88			1·10	1·08	3·87	1·50			74° 18′	71° 30′
	S							1·10	0·41	4·33	2·38			73° 48′	71° 24′
C8	N	2·00	2·48	3·16	6·81			2·23	0·87	0·79	3·45			72° 15′	63° 47′
	S							2·23	0·60	0·89	2·91			71° 52′	62° 49′
C9	N	2·00	2·81	3·24	5·19			1·64	1·13	0·14	2·23			72° 06′	61° 37′
	S							1·64	0·54	0·25	4·11			71° 55′	61° 03′
C10	N	2·25	2·63	3·11	5·37	7·88		1·80	2·45	0·25	1·10	2·19		68° 59′	62° 30′
	S							1·80	0·90	1·75	0·36	4·50		68° 31′	61° 59′
M1	N	2·10	2·42	3·05	4·60			0·76	0·10	1·43	1·25			74° 45′	60° 55′
	S							0·76	0·10	1·50	0·95			74° 28′	59° 55′
D1	N	2·10	2·42	4·78	6·19	8·36		0·66	0·18	1·39	2·13	18·43		65° 59′	57° 54′
	S							0·66	0·18	1·39	3·56			65° 29′	57° 46′
B1	N	2·30	3·47	5·72	6·93	7·68		2·30	2·00	2·30	2·10	2·10		72° 18′	65° 13′
	S							2·30	2·10	2·00	2·00	1·90		71° 39′	65° 55′

(Note: Water depths are the mean water depths along each line.)

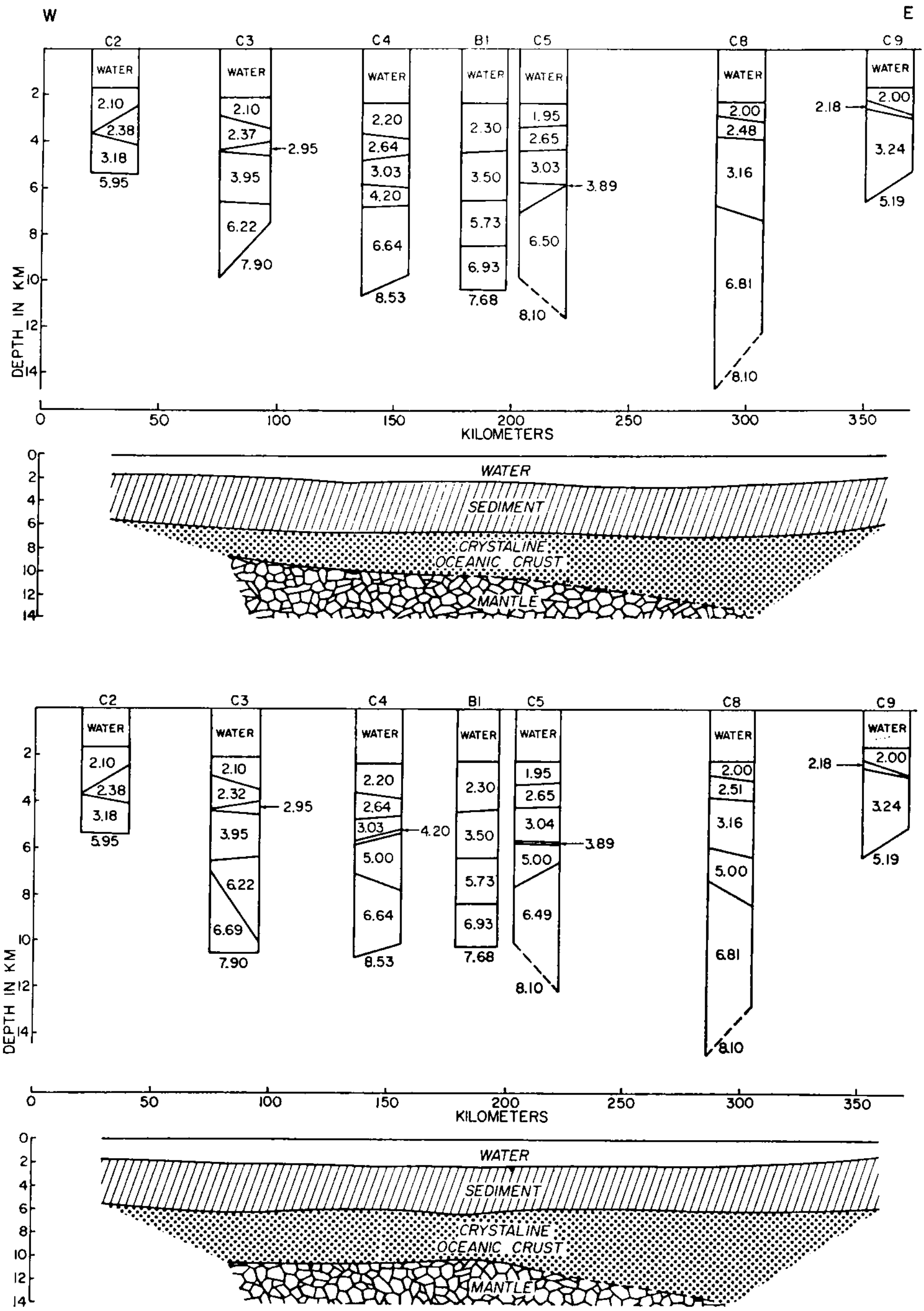

FIG. 7. Crustal sections across Baffin Bay from west to east, and a schematic diagram of the structure across the Bay. The upper portion shows the results, with no assumptions regarding hidden layers. The lower portion (see text) depicts the structure when a $5\cdot0\ \mathrm{km\ s^{-1}}$ layer is assumed to occur. The dashed lines indicate the minimum depth to the M discontinuity. Velocities are in kilometres per second.

is not observed (C. Keen *et al.* 1972). Furthermore, these velocities are lower than those normally associated with layer 2 (Christensen & Shaw 1970; Ludwig *et al.* 1970). Beneath the 4 km s^{-1} material only one high velocity crustal layer is measured and this varies from 5·2 to 6·9 km s^{-1}. These velocities encompass those associated with layers 2 and 3 in other oceans. We have made a somewhat arbitrary grouping so that velocities less than 6·3 km s^{-1} correspond to layer 2. This derives some support from velocity measurements on oceanic basalts (Christensen & Shaw 1970) which show that basalts, thought to form layer 2, may exhibit velocities as high as 6·3 km s^{-1} at appropriate pressures. Alternatively, as Raitt (1963) has observed, failure to detect layer 2, a not uncommon occurrence, is usually combined with a reduced layer 3 velocity measurement. Therefore velocities between 6·2 and 6·5 km s^{-1} are difficult to assign to a particular layer.

It is probable that both layer 2 and layer 3 occur in the Baffin Bay basin and in fact they were both detected on line B1. Because one of the purposes of this set of refraction lines was to map the depth to the top of layer 2, it is important to determine how the sediment thicknesses change if we assume that layer 2 occurs beneath all the lines. A fictitious layer was therefore assumed with a velocity of 5·0 km s^{-1}. The corresponding time intercept was chosen so that this layer would have the maximum thickness without appearing as a first arrival on the time–distance plots. The structure was then recalculated for lines C4, C5 and C8, and the results shown in Fig. 7. Along these lines layer 2 was not observed. It is evident that the main effect of assuming a hidden layer 2 is to reduce the thickness of sediment by about 1 km. Also, there is more variation in the depth to layer 2, about 0·75 km. These changes are random and no regional variation is observed. If another intercept for layer 2 were chosen the resulting structure would fall between the two models shown in Fig. 7.

Fig. 7 also shows the increase in depth to the *M* discontinuity beneath line C3 if a layer with a velocity of 6·7 km s^{-1} is assumed to occur. The calculations were made in the same way as those described above. The crustal thickness is increased by about 2 km.

In neither case is there evidence for a basement high in the centre of the Bay. The sediment thickness varies from 3 to 4 km with no significant increase near the margins. The topography of the basement is flat, within the resolution of the refraction method and the line spacing. Although the measurements of depth to basement may be in error by as much as 1 km, it is unlikely that the errors are systematic across the traverse. Also, as shown above, the most likely source of error on the most central profiles, C4, C5 and C8, does not affect this conclusion. The depth to the *M* discontinuity and the thicknesses of layers 2 and 3 are variable. A total depth of 10 km to the *M* discontinuity which is overlain by 4 to 5 km of layer 2 and layer 3 appears to be the most reasonable estimate of total crustal thickness for the central part of the Bay. Excluding the sediment cover, this makes the total crustal thickness less than in other well known ocean basins (Ludwig *et al.* 1970).

Lines C1 and C7 (Figs 2 and 3) were obtained near the transition from oceanic to continental crust in the northern part of the Bay. Although the *M* discontinuity was not detected beneath these lines, the gravity and bathymetry suggests that they were located over oceanic crust (Ross & Manchester 1972). Both lines exhibit high crustal velocities, typical of oceanic layer 3 (Table 1, Fig. 8). However, velocities of 4·2 km s^{-1} beneath line C7 and 4·8 km s^{-1} beneath line C1 are lower than those which we have previously associated with layer 2. If layer 2 has not been detected and layers exhibiting velocities less than 5 km s^{-1} are sedimentary, the sediment thickness is approximately 7 km beneath both lines. The 4·8 km s^{-1} velocity does, however, fall within the range of layer 2 velocities measured elsewhere (Raitt 1963) and it is possible that the sediment thickness beneath line C1 is only 3 km. Unlike the traverse of refraction lines across the Bay, there is no additional evidence from reflection profiling to aid in resolving the ambiguity in the interpretation of this line.

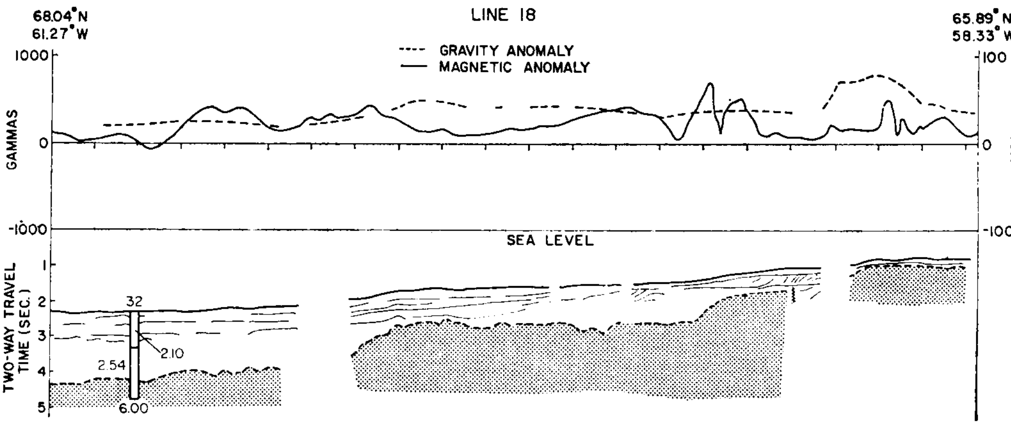

FIG. 8. Line drawing of the seismic reflection profile and the gravity and magnetic measurements along the dashed track shown in Fig. 1. The shaded area of the reflection profile indicates basement. The column denotes the results of expendable sonobuoy (line 32), the numbers being velocities in kilometres per second. The ticks on the horizontal scale are divisions in hours so that each division represents approximately 11 km.

Line C10 in the southern part of Baffin Bay was carried out to define the structure near the transition between the deep basin and the Davis Strait sill to the south. The crustal structure (Table 1, Fig. 8) is oceanic, similar to that measured further north with a depth to the *M* discontinuity of 9 km. It is possible that a high velocity crustal layer corresponding to layer 3 is present but was not detected. As mentioned above, the presence of this layer would increase the calculated depth to the *M* discontinuity. The sediment thickness, 3 km, is slightly less than that measured along the traverse further north.

Four expendable sonobuoy experiments were carried out in the deep basin (Lines, 19, 30, 32 and 35, Fig. 1). Line 19 which lies in the northern part of the area is similar in some respects to the results along the traverse shown in Fig. 7. Velocities and thicknesses are presented in Table 2. Approximately 5 km of sediments are underlain by a layer with a velocity of $6 \cdot 0\ \mathrm{km\ s^{-1}}$. The thickness of unconsolidated sediment, 0·6 km, however, is much less than that measured beneath other lines in the area. Furthermore, no reflector corresponding to the interface between the $2 \cdot 9\ \mathrm{km\ s^{-1}}$ and $3 \cdot 7\ \mathrm{km\ s^{-1}}$ layers appears on the seismic reflection record. This makes one somewhat suspicious of the results. Lines 30, 32 and 35 are located in the southern part of the deep basin. All these shown between 2 and 3 km of unconsolidated and semi-consolidated sediment which is consistent with the results from C10. The velocities below the sediments vary from 3·5 to $6 \cdot 0\ \mathrm{km\ s^{-1}}$ and it seems probably that the top of layer 2 was not detected on line 30 where $3 \cdot 5\ \mathrm{km\ s^{-1}}$ was the highest velocity measured. The $4 \cdot 5\ \mathrm{km\ s^{-1}}$ velocity measured on line 35, the $6 \cdot 0\ \mathrm{km\ s^{-1}}$ on line 32 and the $5 \cdot 3\ \mathrm{km\ s^{-1}}$ layer detected beneath C10 probably represent the top of layer 2. The depths to the top of these layers approximately correspond to a rough, strong reflector observed on the reflection record whose position is shown in Fig. 1 and the results in Fig. 8.

4.2 *Davis Strait*

The refraction line in Davis Strait (D1) is a 55 km reversed profile which was extended beyond its southern end to a total distance of 155 km using the tape recording buoy. This allows reversed velocities for the main crustal layer and an unreversed measurement of the velocity and depth to the *M* discontinuity to be obtained (Fig. 4). The results are presented in Table 1. About 1·5 km of unconsolidated and semi-consolidated sediment overlies a layer with a velocity of 4·8 km s^{-1} and a thickness of 2 km. The main crustal layer is about 18 km thick with a velocity of 6·2 km s^{-1}. The total depth to the *M* discontinuity is about 22 km.

Two expendable sonobuoy lines (lines 7 and 8) near the northern end of line D1 also measure a velocity between 4·2 and 4·7 km s^{-1}. The sediment cover over this layer varies from near zero beneath line 7 to 1·3 km under line 8. The magnetic anomaly measurements show that this layer is highly magnetic and is therefore probably basaltic in composition, the equivalent of layer 2 in the deep ocean basins. The seismic reflection profile (Figs 1 and 8) indicates that layer 2 can be traced from the deep portion of Baffin Bay, known to be underlain by oceanic crust, to the top of the Davis Strait sill. Therefore the results support the hypothesis that Davis Strait is not underlain by rocks of continental origin but is composed of a thick pile of oceanic crustal material.

Two expendable sonobuoy measurements, lines 37 and 41, were made near Cape Dyer, Baffin Island. Tertiary basalts (Clarke & Upton 1971) outcrop on land in this area and their seaward extent was traced by observing the characteristic high intensity, high frequency magnetic anomalies associated with basic rocks, near the surface (Grant 1972). Sonobuoy line 37 was obtained directly over the basalts; in the area of high magnetic anomalies. The results show virtually no sediment cover and a near surface velocity of 5·1 km s^{-1}. Sonobuoy line 41, in the area of relatively smooth magnetic anomalies, shows a thin cover of unconsolidated sediment and a layer, 4·9 km thick, of material presumed to be consolidated sediment underlain by rocks with a velocity of 6·0 km s^{-1}.

Table 2

Refraction results from expendable sonobuoys

	Velocity (km s^{-1})				Thickness (km)				Location	
Line	A	B	C	D	Water	A	B	C	Lat.	Long.
4	2·83				0·27				62° 58′	61° 50′
5	2·28				0·71				63° 49′	60° 14′
7	2·42	4·64			0·57	0·01			66° 23′	57° 57′
8	2·36	4·19			0·66	1·30			66° 04′	57° 53′
9	2·20	3·68			0·57	0·42			70° 41′	55° 58′
10	2·30	4·40			0·63	1·03			70° 53′	56° 30′
11	2·31	4·37			0·51	0·53			71° 00′	55° 54′
12	2·11	4·40			0·41	0·01			71° 13′	56° 00′
13	4·35	5·20			0·22	0·93			72° 40′	57° 59′
15	2·32	3·54			0·27	0·87			72° 10′	58° 00′
16	2·24	3·00	4·32		0·28	0·07	4·07		73° 14′	60° 05′
17	2·30	3·12			0·78	1·19			74° 29′	60° 00′
18	5·80				0·51				74° 29′	60° 00′
19	2·30	2·91	3·69	6·00	1·90	0·22	0·39	4·42	73° 07′	69° 26′
22	2·10	3·02	4·64		0·59	0·06	1·51		77° 09′	75° 18′
23	2·10	2·93	5·23		0·79	0·31	1·19		76° 03′	82° 25′
25	2·10	2·52	3·12		0·56	0·01	0·46		75° 49′	80° 41′
27	2·12	2·65			0·70	0·52			74° 21′	78° 25′
28	2·03	2·54	4·14		0·88	0·56	1·29		74° 01′	77° 44′
30	2·10	2·29	3·50		1·90	0·77	1·41		70° 02′	62° 29′
31	2·10	2·46	2·86	3·68	0·34	0·04	0·91	1·10	69° 00′	58° 25′
32	2·10	2·54	6·00		1·67	1·08	1·82		67° 53′	61° 03′
35	1·91	3·00	4·49		1·54	1·14	1·00		67° 59′	61° 56′
37	2·42	5·08			0·40	0·06			66° 23′	60° 51′
41	2·10	3·35	6·00		0·47	0·75	4·91		65° 30′	59° 49′

The arrivals from the $6 \cdot 0 \mathrm{~km~s^{-1}}$ layer were poor and it is possible that the calculated thickness of the sediments is too great. However, the results together with those in the central part of Davis Strait suggest that the basalts are continuous, but down-faulted by several kilometres from Cape Dyer across the Davis Strait sill.

4.3 *Continental shelves*

Two reversed lines, C6 and M1, were obtained on the continental shelf surrounding Baffin Bay (Fig. 1). Line C6, over the northern shelf area, defines three sedimentary layers with velocities corresponding to unconsolidated, semi-consolidated and consolidated sediment above a high velocity ' basement ' layer (Table 1, Fig. 9). The latter is probably composed of carbonates although it may represent pre-Cambrian basement. The total thickness of sediment down to the high velocity layer is about 7 km. Comparable thicknesses of sediment are known to exist elsewhere north of the Baffin Bay basin, for example, in Lancaster Sound (Barrett 1966). Expendable sonobuoy line 22, to the north of C6, exhibits a similar velocity layering although the high velocity ' basement ' was not detected (Table 2).

It would be unwise to relate the velocity structure to the age of the sediments. Overton (1970) and Sander & Overton (1965) have made seismic refraction measurements in the Canadian Arctic Islands and found that rocks of Devonian age and older exhibit velocities between 5 and $7 \mathrm{~km~s^{-1}}$. Permo-carboniferous and Mesozoic rocks are associated with velocities ranging from $5 \cdot 2$ to $4 \cdot 0 \mathrm{~km~s^{-1}}$. These results show that one should be cautious in relating high velocity layers, over $5 \cdot 5 \mathrm{~km~s^{-1}}$, with lower Palaeozoic or Pre-Cambrian rocks.

Line M1 was obtained to define the nature and thickness of sediments filling in Melville Bay graben (C. Keen *et al.* 1972). A similar velocity layering was obtained beneath this line as beneath C6 (Table 1, Fig. 9). However, the highest velocity observed is $4 \cdot 6 \mathrm{~km~s^{-1}}$. If one assumes a velocity of $5 \cdot 8 \mathrm{~km~s^{-1}}$ for the basement rocks the minimum thickness of sediment is 7 km. Expendable sonobuoy lines 16

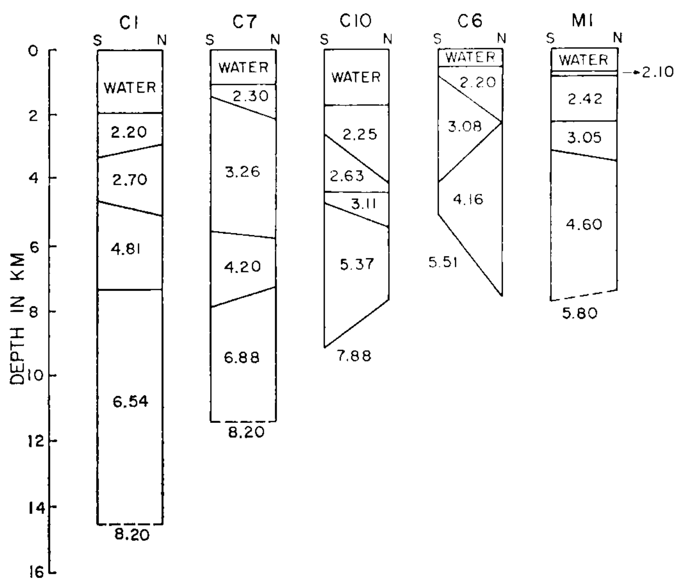

FIG. 9. Crustal sections for some of the reversed profiles. The dashed lines indicate the minimum depth to the *M* discontinuity. The numbers are velocities in kilometres per second. The letters denote the orientation of the lines in the north–south direction.

and 17, also within the graben, show similar velocities but variable layer thicknesses (Table 2). Expendable line 18 was located to the east of the graben in an area of rough bottom topography and a velocity of $5 \cdot 8$ km s^{-1} was measured. This is tentatively correlated with pre-Cambrian basement because of exposures of pre-Cambrian rocks on nearby islands. There is no sediment cover above this layer.

Several other expendable sonobuoy experiments were carried out on the continental shelf surrounding the basin. It is more appropriate to discuss these in a later paper in relation to the seismic reflection and other geophysical results. They are, however, listed in Table 2.

5. Discussion

Several facts emerge from the results of the refraction experiments which must be accounted for in any reconstruction of the evolution of Baffin Bay. First, the extent of oceanic crust in the central basin, outlined by C. Keen *et al.* (1972), is confirmed by the recent refraction results. On the northern continental slope, which is characterized by a relatively gentle gradient, oceanic crust is found beneath areas of relatively shallow water depths, about 1100 m, buried by up to 7 km of sediment. It would appear that high rates of sedimentation are extending the continental margin beyond the true transition from continental to oceanic crust as defined by gravity and seismic measurements (see also C. Keen *et al.* 1972). Mayhew, Drake & Nafe (1970) have suggested that this extension also occurs on the Labrador Sea continental margin. Similar thicknesses of sediment are measured on the true continental shelves; 7 km in the Smith Sound area and at least 7 km in the Melville Bay graben. These areas may be of economic importance in terms of petroleum deposits.

The crustal structure measured on all refraction lines in the central basin differ in velocity layering and layer thickness from the more typical oceanic crustal structure found in the major ocean basins (Ludwig *et al.* 1970). One might expect that the basin would be underlain by typical oceanic crust, depressed by several kilometres of sediment, as observed in other small, enclosed oceanic areas, the Gulf of Mexico for example. Instead, the main oceanic crustal layers (omitting the sedimentary layers) are significantly thinner, 4 km instead of 6 to 7 km, and the mantle is shallow, 10 km. The Labrador Sea is also characterized by a thin crust in its central ocean basin (Drake *et al.* 1963; Mayhew 1969). The shallow mantle is consistent with the positive free-air gravity values, about 20 mgal, measured in the central basin (C. Keen *et al.* 1972). On many lines the velocity of layer 3 is somewhat below the oceanic mean and layer 2 was not detected. However, the mantle velocities do fall within the range associated with normal mantle rocks with a mean of $8 \cdot 0$ km s^{-1}.

The thin crust, found in both of these ocean basins, is the most predominant feature of the area. It may perhaps be related to the age of the crustal material. Near active mid-ocean ridges the crust appears to thicken with distance away from the ridge axis and hence perhaps with age (Ludwig *et al.* 1970; Le Pichon *et al.* 1965). The Labrador Sea and Baffin Bay regions ceased spreading about 50 My ago, the youngest oceanic crust therefore being about 50 My old, and the oldest about 80 My (Le Pichon, Hyndman & Pautot 1971; Laughton 1971). Observations near the northern mid-Atlantic ridge suggest that the crustal layers are thinner, 4 to 5 km, than the average for the deeper basin. This is also true for the East Pacific rise.

Measurements of total crustal thickness of layers 2 and 3 over the northern mid-Atlantic ridge have been plotted against water depths and a least squares straight line fitted to the data (Le Pichon *et al.* 1965). If the age-water depth relationship of Sclater, Anderson & Bell (1971) is used to convert the depths of this straight line fit to age, an age of 80 My corresponds to a crustal thickness of about 6 km and 50 My, to a crustal thickness of about 5 km.

We therefore suggest that there may be a direct relationship between crustal thickness and age although there is considerable scatter in the measurements of crustal thickness. This relationship measured near the mid-Atlantic ridge is consistent with the Baffin Bay refraction results and the postulated age of the crust there. This implies that the correlation may be universal and not restricted to presently active ocean ridge systems. It may be significant that the thickness of layer 3 appears to increase towards the margin to the east, in a region of older crust although this is not observed on the western side of the basin (Fig. 7). The results also support the hypothesis that at least the lower part of layer 3 is formed by an alteration of the mantle rocks, probably by serpentinization. This would, in time, lower the *M* discontinuity to a depth more consistant with results in other ocean basins (Vine & Hess 1970).

Refraction and reflection results in southern Baffin Bay and over the Davis Strait sill show that the latter is probably underlain by a thick pile of oceanic crustal material, perhaps similar to that beneath Iceland (Palmason 1970). The $6 \cdot 2 \text{ km s}^{-1}$ velocity of the main crustal layer falls within the range of velocities measured for the same layer beneath Iceland. The crustal thicknesses are comparable although the Davis Strait area appears to have a somewhat thicker crustal section—8 to 16 km beneath Iceland, 22 km beneath Davis Strait. The crust under Iceland is underlain by an anomalously low velocity mantle, $7 \cdot 2 \text{ km s}^{-1}$, whereas the unreversed measurement in Davis Strait indicates a normal mantle velocity. As the latter was measured over 75 km it seems unlikely that the true velocity is significantly less. The difference in mantle velocity and in crustal thickness is more probably related to the situation of Iceland over an active mid-ocean ridge system where the high temperatures beneath have substantially reduced the velocity. The results support the hypothesis that the Davis Strait was once the site of a hot spot or upwelling convection plume (Morgan 1971).

Another possibility, although more unlikely, is that the sill is a continental fragment. The seismic refraction results are in some respects similar to those obtained for Rockall plateau (Scrutton 1972) and the Seychelles Bank (Davies & Francis 1964). Both these areas appear to be underlain by continental crust 22 to 31 km thick and the mantle exhibits normal velocities of about $8 \cdot 1 \text{ km s}^{-1}$. There are three main reasons for rejecting a continental origin for Davis Strait. First, unlike these continental fragments, layer 2 can be traced from the Baffin Bay basin to the top of the sill which is convincing evidence that both the sill and the basin are floored by the same rock type and hence may have a similar origin. Second, both microcontinents mentioned above exhibit two high velocity layers, with velocities of about $6 \cdot 3$ and $6 \cdot 9 \text{ km s}^{-1}$. Scrutton (1972) has suggested that this may be typical of continental fragments. There is no evidence for two main crustal layers beneath the Davis Strait. Finally, if the Labrador Sea and Baffin Bay were formed by sea-floor spreading and Greenland and North America were once joined, it is difficult to conceive of a mechanism for retaining such a thick continental crustal block between the two main continental masses. Although these arguments do not provide absolute proof, they favour a sea-floor spreading origin for Davis Strait.

An important result of the refraction measurements is that there is no detectable central ridge in Baffin Bay. Systematic errors in the measurements across the traverse from west to east of about $0 \cdot 5$ km may occur so that a ridge with an elevation equal to or less than this value may be present. Also the sediment thickness remains constant across the traverse. In the Labrador Sea, on the other hand, a ridge is detected with topographic relief of about 1 km (Le Pichon *et al.* 1971) and the sediment thickness varies from 3 to 4 km near the margin to 1 to 2 km over the central ridge.

Sclater *et al.* (1971) have shown that depth to the top of oceanic layer 2 is directly related to the age of the sea floor and that the relationship is compatible with a cooling, contracting plate as it moves away from the centre of spreading. The same model predicts that 50 My after spreading has ceased the excess elevation of the ridge crest will have decreased to about 0·5 km. This is compatible with the postulated spreading history of the Baffin Bay area and explains the lack of a detectable median ridge. The differences between Baffin Bay and the Labrador Sea in terms of their basement topography may be explained by a suggestion of Le Pichon *et al.* (1971) that the Labrador Sea is still active and opening at a very slow rate.

The depth-age curve of Sclater *et al.* (1971) cannot be applied to the Baffin Bay basin to give an estimate of its absolute age. Although we corrected the depth for the isostatic effect of the thick sediments, the predicted age for the central part of the basin, 15 My, is much less than other geophysical evidence would allow. The depth-age relationship is invalid in other areas, such as the Reykjanes Ridge which is situated near a hot spot. There the elevation of the ridge crest is greater than of the mid-Atlantic ridge further from the postulated hot spot. The excess elevation appears to be largely due to a more intense thermal anomaly in the mantle below (Talwani, Windisch & Langseth 1971). If the Davis Strait was once the site of a hot spot there may be a remanent thermal anomaly in the mantle below causing a somewhat greater elevation of the ocean basement rocks and hence the unrealistic age predicted by the depth-age relationship of Sclater *et al.* (1971).

The constant thickness of sediment across the Bay is in conflict with the idea of sea-floor spreading. Furthermore, if we assume that there was once a ridge in the centre that has now subsided it should be possible to observe dipping reflectors on the reflection records. These results, however, show about 2 km of flat lying sediment across the basin, underlain by a disturbed zone in which no clear reflectors can be seen (C. Keen *et al.* 1972; M. Keen *et al.* 1972). This suggests that at least half of the sediments were deposited after most of the ridge subsidence had occurred. The distribution of sediments during the period of active spreading is difficult to determine from the refraction measurements. Possibly about 2 km of the sediment has been derived by denudation of the surrounding continents by glaciation from late Tertiary to the present time. Even allowing for the possibility that much of the deposition may be recent, rates of sedimentation must have been quite high when sea-floor spreading was active—about 10 cm/k.y. allowing for compaction and assuming that 1·5 km of the sediment presently observed was deposited between 80 and 50 My ago. Possibly the temperature within layer 2 was sufficiently high because of its rapid blanketing by sediment that the basalts lost much of their initial magnetization through low grade metamorphism. This might explain the lack of well-defined magnetic lineations in the central basin.

6. Conclusions

Seismic refraction results in Baffin Bay show that the central basin is floored by oceanic crust covered by 3 to 7 km of sediment; the thicker sequences being formed at the foot of the northern continental slope. The northern continental margin appears to be extending seaward over oceanic crustal material and the bathymetric expression of the shelf break is not a reliable indication of the true transition from oceanic to continental crust. The Melville Bay graben, a major feature on the west Greenland shelf is filled by at least 7 km of sediment and similar sediment thicknesses are measured in the vicinity of Smith Sound on the northern continental shelf.

The Davis Strait sill to the south of Baffin Bay exhibits a seismic structure similar to that beneath Iceland; the main difference being that a normal mantle velocity is found beneath the former. The refraction results, supported by a seismic reflection line in the area, suggest that the sill was formed, like Iceland, by an excessive outpouring of oceanic basalts and hence may once have been the site of a hot spot.

The oceanic crust in the central basin is abnormally thin, about 4 km excluding the sediments, and is underlain by mantle material with velocities ranging from

7·7 to 8·5 km s^{-1}. These values are within the range associated with 'normal' mantle rocks elsewhere in the oceans. No evidence for a central ridge was found. This is consistent with the degree of subsidence calculated for a ridge system which has been inactive for about 50 My. The most characteristic feature of the basin is the thin oceanic crustal layers. A direct relationship between the age of the sea floor and crustal thickness is suggested as a possible explanation. Thicknesses in Baffin Bay compare well with those measured near other active ridge systems where the age of the crust is similar. Further crustal thickening may be accommodated by alteration of the mantle rocks and therefore at least the lower part of Layer 3 in the deep ocean basins may be composed of serpentinized peridotite.

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