EARTH SCIENCE SYMPOSIUM ON OFFSHORE EASTERN CANADA

Ottawa, February 1971

Edited by
Peter J. Hood

NATIONAL ADVISORY COMMITTEE ON RESEARCH IN THE GEOLOGICAL SCIENCES

ASSOCIATE COMMITTEE ON GEODESY AND GEOPHYSICS OF THE NATIONAL RESEARCH COUNCIL

Published by the Geological Survey of Canada as Paper 71-23

EARTH SCIENCE SYMPOSIUM
ON OFFSHORE EASTERN CANADA

Cover design based on
Lakes, Rivers and Glaciers,
National Atlas of Canada
Special Interim Series, 1967.

Production editing
and Layout
Leona R. Mahoney

GEOLOGICAL SURVEY OF CANADA

PAPER 71-23

EARTH SCIENCE SYMPOSIUM ON OFFSHORE EASTERN CANADA

Proceedings of a symposium held in Ottawa February 22nd-24th, 1971 under the sponsorship of the National Advisory Committee on Research in the Geological Sciences and the Associate Committee on Geodesy and Geophysics of the National Research Council

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DEPARTMENT OF ENERGY, MINES AND RESOURCES

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Price: $7.50 Catalogue No. M44-71-23

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Information Canada
Ottawa
1973

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FOREWORD

The Earth Science Symposium on Offshore Eastern Canada was held in Camsell Hall at the Department of Energy, Mines and Resources in Ottawa from February 22nd to 24th, 1971. Because of its multi-disciplinary nature, the symposium was sponsored both by the National Advisory Committee on Research in the Geological Sciences and the Associate Committee on Geodesy and Geophysics of the National Research Council. The two and a half day symposium was opened by the Deputy Minister of Energy, Mines and Resources, Mr. J. Austin, and was closed by Mr. Arne Nielsen, President of Mobil Oil (Canada) Ltd. whose company was the first to obtain offshore permits on the east coast. A total of 373 persons attended the symposium: of these, 215 were from industry and the remainder were government and university scientists. The regional representation included 130 from Alberta, 63 from the U. S. A. and 14 from overseas. The main objective of the symposium was to summarize the available geological knowledge of the continental shelves and slopes of eastern Canada and to make available jointly to exploration personnel in industry and government and university scientists a shared insight into the problems and opportunities of developing commercial oil and gas reserves in that region. Each of the sessions was a multi-disciplinary study of one or more areas, and the papers for a given area were arranged so that those bearing on surficial geology were first and the papers which followed were arranged in order of increasing depth penetration. Thus in a given session the geological edifice for a specific area was studied from the top downward. As there is only one geological edifice in a given area, the results presented in the various papers should have been mutually compatible. To this end authors were given a three-month-period from the date of the symposium to revise their papers for the Proceedings Volume in order to take account of other earth scientist's results presented at the symposium. The Organizing Committee for the symposium consisted of E. R. Deutsch, A. K. Goodacre (Secretary), G. D. Hobson, P. J. Hood (Chairman), M. J. Keen, N. J. McMillan, B. R. Pelletier, W. H. Poole and D. I. Ross. It is our hope that this endeavour might serve as a model for other multi-disciplinary assaults on areas of the continental crust in Canada and elsewhere.

We are especially grateful to Dr. Y. O. Fortier, Director of the Geological Survey, for his support and encouragement as well as for authorizing publication of the proceedings as a Survey report. We wish to thank the following persons, who together with various members of the Organizing Committee, acted as session chairman during the symposium: B. D. Loncarevic, P. J. Savage and P. H. Serson. Thanks are also extended to Leona R. Mahoney and Martha J. McLean for layout and production editing, to Janet M. Crump and Sharon J. Parnham for the final typescript and to the many reviewers who made valuable suggestions for improvement of the manuscripts.

Peter J. Hood
Editor and Chairman,
Organizing Committee.

Ottawa,
March, 1973.

(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 1)

1. PALEOMAGNETISM IN EASTERN CANADA: A KEY FOR RECONSTRUCTING THE ATLANTIC

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Abstract

Paleomagnetic results meeting minimum stability criteria are available from about 25 major rock units of known age (200 to 2,000 m.y.) in the Canadian Appalachian region and Labrador. By their relative proximity to the continental margin such rocks can provide key evidence on plate movements predating the evidence obtainable from present oceanic crust. Three types of paleomagnetic comparison are examined: (1) between eastern Canada and other North American areas; (2) across the North Atlantic; (3) between parts of eastern Canada. The Triassic, Permian, Carboniferous and some Precambrian data are adequate for type (1) comparisons. Generally, even widely separated contemporaneous North American sites yield similar pole positions, suggesting (a) that the ambient Earth's field was mainly dipolar, and (b) that North America has not been unduly deformed since the late Paleozoic. This justifies using combined North American data for type (2) comparisons with Europe which, for post-Carboniferous times, are simultaneously compatible with (i) a Wegener-type separation between these continents and (ii) their joint northward movement representing either continental drift or polar wandering. The North American and European data also support an earlier proposal by the author for a three-stage initial opening of the North Atlantic, where (I) a late Paleozoic single-plate assemblage of North America and Europe, similar to the fit of Bullard et al. (1965), is (II) ruptured by a northward movement of North America relative to Europe, probably accompanied by some longitudinal separation between the two continents; (III) this is followed by the major separation of these continents, initially about a "Triassic" pivot located near the Alaskan pivot favoured by Wegener. Tests of "pre-Wegenerian" reconstructions, e.g. Wilson's "proto-Atlantic", involve all three types [(1), (2), (3)] of comparison and require pre-Carboniferous evidence which is so far inadequate. Systematic paleomagnetic sampling in the future should lead to conclusive tests, but not only more data but a critical choice of localities is required. For example, unambiguous evidence on the rotation of Newfoundland is unobtainable from paleomagnetic comparisons with New Brunswick, but should be sought by comparing contemporaneous sites in western Newfoundland and in North America outside the Appalachian belt.

REFERENCE

Bullard, E.C., J.E. Everett, and A.G. Smith
1965: The fit of the continents around the Atlantic, in Symposium on Continental Drift; Phil. Trans. Roy. Soc. London, Series A, v. 258, no. 1088, p. 41-51.

(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 3-22)

2. LATITUDE MAPS OF THE EASTERN NORTH AMERICAN-WESTERN EUROPEAN PALEOBLOCK

J. L. Roy, Division of Geomagnetism, Earth Physics Branch, Ottawa

Abstract

Paleolatitude maps obtained from eastern North American and western European paleomagnetic data indicate that, in the Late Paleozoic, these regions and their adjacent continental shelves were in low latitudes and were parts of a single block. The Permo-Carboniferous paleoequator transects the Maritime area of eastern Canada and therefore, if the hypothesis that oil and gas fields are more likely to be formed in low latitudes is valid, the probability for finding hydrocarbons in the Carboniferous and Permian sedimentary rocks of the Maritimes and the Gulf of St. Lawrence is much enhanced. A northward movement (with respect to a stationary paleopole) of the block during the Carboniferous can be detected. The Mesozoic paleolatitude maps indicate that the rupture of the block and the opening of the North Atlantic Ocean did not follow each other closely in time. A latitudinal displacement of North America with respect to Europe places the time of rupture during the Lower or Middle Triassic. It is suggested that, during the Triassic, eastern North America and western Europe slipped one against the other in a north-south direction (by about 20° in latitude), with little separation in longitude occurring. The data do not provide any evidence for an appreciable opening of the North Atlantic until after the Triassic.

INTRODUCTION

Paleomagnetic data have often been used to test the continental drift theory and to reconstruct ancient land masses or continents. In 1956, it was shown (e.g. Irving, 1956; Runcorn, 1956) that the polar wandering curves (constructed on the assumption that the field was dipolar) of North America and Europe did not coincide, suggesting that the two continents were not in the same relative positions as they are today. However, the polar paths could not be determined very accurately because most of the results available at that time were based on observations of NRM, that is, the total remanent magnetization as found in a rock. Since a rock may acquire different magnetizations at different times in its history, the NRM does not necessarily accurately represent the field at the time of formation although in favourable cases it may do so approximately.

With the analytical techniques now available (e.g. alternating field and thermal cleanings, and chemical leaching), it can be determined if one or more magnetizations are present and it is often possible, with the aid of a fold, tilt or contact test to determine within given limits the age of magnetization of one of the constituents. With extensive sampling of a rock formation, the mean direction of the ancient field and consequently the pole relative to the sampling location can be determined to an accuracy of 5° or so. The simple presentation of a pole result is often difficult to visualize and it is certainly difficult to compare it with geological and other geophysical data.

An alternative method is to work back from the paleopole and calculate the paleolatitudes of localities of interest. These can then be tested against reconstruction of continents obtained by other means. This more readily understood procedure is used in this paper.

The North American-European Paleoblock

The problem is to determine 1) the relative positions of the North American and European continents for Late Paleozoic time, 2) the time at which the block ruptured, and 3) the time at which the North Atlantic Ocean began to open; 3) and 2) may coincide but this is not necessarily so. The two continents have been fitted in a number of ways and one reconstruction which has recently drawn much attention is that of Bullard et al. (1965). They closed the Atlantic Ocean by fitting the continental slopes at the 500-fathom bathymetric contour. The fit can be tested paleomagnetically and a few tests using polar wandering curves (Wells and Verhoogen, 1967; Hospers and Van Andel, 1968; Larson and La Fountain, 1970; Deutsch, 1973) have been carried out. The selection of data used in each test has varied depending on the literature available to the author and on the criterion of acceptability of each author. As shown in the preceding paper (Deutsch, 1973), the exercise then consists in rotating the polar path curves about the axis of rotation (or pivot point) used by Bullard et al. (1965). Coincidence of the two curves upon rotation of 38° which is the angle used in the topographic fit would confirm the goodness of fit. Furthermore, in principle, the time of the start of the opening of Atlantic would be shown by the point in time where the two curves would separate. When comparing these works it is found that as the standard of acceptability is increased, the confirmation of the topographic fit becomes more evident. Little information on the time(s) of rupture and drifting apart of the two continents was obtained except that it occurred between Upper Permian and Late Mesozoic time, an uncertainty of perhaps 175×10^6 years. In particular, no information was given on the pattern of rupture. It has been suggested (Larson and La Fountain, 1970) that the North Atlantic began to open as early as the Triassic. However, this is based on the assumption that North America and Europe started drifting apart at the time of rupture. As pointed out by Hospers and Van Andel (1968), the relative movement of the two continents is not necessarily a simple pulling apart but may have followed a more complex course. For example, the pattern of rupture may have been such that the North Atlantic Ocean did not start to open immediately.

Data Selection

Selection and geological time scale of data (Table 1) are made so as to provide as accurate a set of maps as possible. Many new North American and several U.S.S.R. results (translations of E.R. Hope) which have never been used previously in this type of work are included. The standard of acceptability is as follows: 1) the stability of magnetization had to be tested by magnetic and/or thermal cleaning, 2) the sampling had to be extensive enough so that the mean direction was judged fairly representative of the field at the time of magnetization, 3) the age of magnetization had to be determined within reasonable limits. In some cases, results which might possibly be reliable were rejected because of lack of stability-testing details in the original published paper. Most of the U.S.S.R. results were obtained from the Synoptic Tables of U.S.S.R. paleomagnetic data (Khramov and Sholpo, 1967, translated by Hope); the results with a stability measure greater than 0.8 were accepted.

Since the purpose of this study is to obtain information about the North Atlantic Ocean, the sampling area was confined to the adjacent regions. The restriction was made in order to avoid introducing data which might not relate directly to the opening of the Atlantic. Indeed, the continents as they are known today may not have moved as a whole block and may be constituted of different land masses or paleoblocks. Therefore, it is possible that paleomagnetic data obtained from an area far removed from the Atlantic Ocean has no bearing on the opening of the Atlantic. The sampling area was then limited to eastern North America and part of Europe which is sometimes called stable Europe, that is Europe west of the Urals and north of the Alpine front (line of the Carpathians, the Alps and the Pyrenees).

Paleolatitude Maps

The procedure was to obtain a mean European paleopole from European sampling sites; the latitudes of the European block were then drawn and transposed onto the reconstruction map of Bullard et al. (1965). The same procedure was applied to North American results. Paleolatitude maps have been derived for four geological periods. The Upper Devonian and Lower Carboniferous results yield the same paleopole approximately and constitute the basis of the first map. No significant difference could be found between poles from Upper Carboniferous and Permian formations of the same continent and the results have been combined in the second paleolatitude map. Because of the scarcity of available Upper Mesozoic results, the Jurassic and Cretaceous results have been combined to produce the fourth map. The reliability of each set of latitudes has been graded. These reliability indices have been assigned not only on the basis of the number of individual poles but also on the extent of coverage, determination of the time span of magnetization and the assessment of the author of the reliability of his results. Because of the multiplicity of variables, objective grading is difficult and these indices should therefore be regarded as a relative indicator of the reliability of each map based partly on the subjective judgement of the author.

Upper Devonian-Lower Carboniferous Paleolatitude Results

During Upper Devonian-Lower Carboniferous time, the correspondence between the European and North American paleolatitudes is good and both land masses were near the equator (Fig. 1). Thus the paleomagnetic results support the topographic fit obtained by Bullard et al. (1965) indicating that the North Atlantic Ocean was closed and that western Europe and eastern North America were parts of a single block during the period considered.

Upper Carboniferous-Permian Paleolatitude Results

The latitude agreement for Upper Carboniferous-Permian time (Fig. 2) is even better than that obtained in Figure 1 since the paleolatitudes are

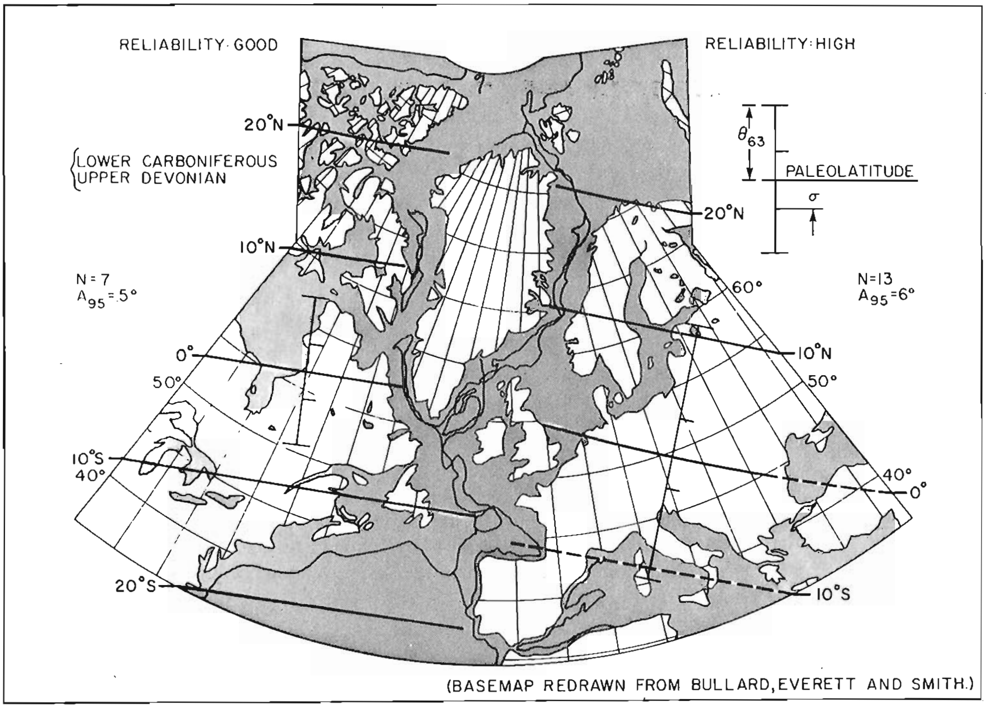

Figure 1. The North American and European latitudes are nearly aligned. This indicates that the paleomagnetic data support the hypothesis that the two land masses were parts of a single block. N is the number of individual poles (see Table 1). A_{95} is the semi-axis of the circle of confidence at the 5% level. The standard error (σ) and circular standard deviation (θ_{63}) are indicated separately for the European and North American results.

parallel; this is probably due to the greater number of observations (N) available for this time period. For these paleolatitudes to be almost continuous, western Europe and eastern North America had to be parts of a single block. Although there is only one Upper Permian result from North America, there are many from Europe and those results do not differ from the mean. The paleolatitude map is probably valid for most if not all of Permian time indicating that the block was still intact in the Late Permian.

The fact that the block is on the paleoequator is also of some interest. Irving and Gaskell (1962) found that a large number of oil fields were located in low paleolatitudes (<20°). The location of the Permo-Triassic oil and gas fields given in Donavan (Kent, 1968) are indicated on Figure 2, and it can be seen that these fields lie near the paleoequator. In view of the fact that the Permo-Carboniferous paleoequator transects the Maritime area of eastern Canada and if the hypothesis that oil and gas fields are more likely to be formed in low latitudes is valid, then the probability for finding hydrocarbons in the Carboniferous and Permian sedimentary rocks of the Maritimes and the Gulf of St. Lawrence is much enhanced.

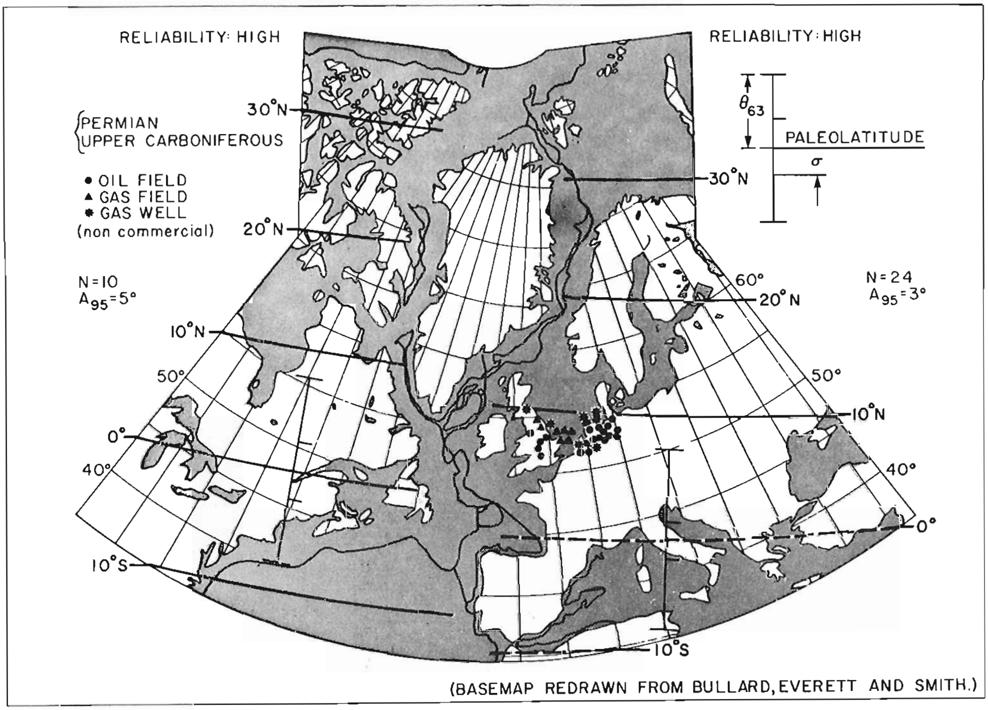

Figure 2. The agreement between the latitudes of the two land masses indicates that the block was still intact. N, A_{95}, σ and θ_{63} as per Figure 1.

Assuming a stationary pole, comparison of Figures 1 and 2 shows that a northward displacement of about 10° of the block is indicated between the two periods. This result agrees with the suggestion often made (e.g. Irving 1966; Storetvedt, 1968; Roy, 1969) than an apparent polar movement occurred during Carboniferous time. It also indicates that the Upper Devonian and Lower Carboniferous rocks have not been remagnetized during the Permian as it is often feared since no displacement would be found if this had happened.

Triassic Paleolatitude Results

The difference between paleolatitudes of the European and North American blocks is about 20° (Fig. 3) indicating that the unit had broken into two parts by the Triassic. Eight out of 10 North American results are from Upper Triassic formations and therefore the paleolatitudes obtained do not necessarily apply to the Middle or Lower Triassic formations. Therefore the rupture must have occurred during Middle or Lower Triassic time or possibly at the end of the Permian. Lower Triassic results are required to determine a more precise time of rupture.

It is interesting to compare the slanting of the paleolatitude lines with respect to the paleolatitude lines of Figure 2. The slanting itself may be explained by a westward displacement of the pole during the Triassic.

TABLE 1

Selected data for map compilation

Age	Paleoblock	Location	Lat., Long. °N, °E	Particulars	Test[+]	Pole °N, °E	Ref.[1]
Cretaceous							
Jurassic							
	North America	U.S.A.	38,282	Mesozoic Dykes	a	66,146	13
		U.S.A.	43,287	Mount Ascutney Gabro	a	64,187	33
		U.S.A.	44,288	White Mountains Volcanics	a	85,126	33
		CANADA	45,287	Mount Megantic Intrusives	a	68,171	24
		CANADA	46,287	Monteregian Hills	a	69,187	27
		CANADA	79,256	Isachsen Diabase	a	69,180	28
	Mean					71,172	
	Europe	U.S.S.R.	55,039	Altian Eleurolites	0.6*	60,167	76
		NORWAY	78,015	Diabase Dykes, Spitsbergen	a	58,178	23
	Mean					59,173	
Triassic							
	North America	U.S.A.	40,283	Pennsylvania Diabase	a	62,105	2
		U.S.A.	40,285	Neward Group (N.J.)	a	63,108	15, 32
		U.S.A.	42,287	Massachusetts Lavas	a	55,088	19
		U.S.A.	42,287	Connecticut Valley Igneous rocks	a	65,087	14
		CANADA	44,294	Shelburne dyke (N.S.)	a	69,098	30
		CANADA	45,293	Grand Manan Island Lavas	a	80,100	6
		CANADA	45,295	North Mountain Basalt (N.S.)	a	66,113	26

Age	Paleoblock	Location	Lat., Long. °N, °E	Particulars	Test[+]	Pole °N, °E	Ref.[1]
		CANADA	45,295	North Mountain Basalt	a	73,104	6
		CANADA	52,292	Manicouagan Dacite	t	57,089	36
		CANADA	52,292	Manicouagan Structure	a	60,088	29
	Mean					65,097	
	Europe	GERMANY	46,012	Ladinian Volvanics	a	46,155	7
		U.S.S.R.	49,038	Donbass, Sediments	a	51,146	20
		ENGLAND	51,357	Keuper Marls (Sidmouth)	a	44,134	11, 12
		ENGLAND	53,358	Keuper Marls	a,t	43,131	8, 9
	Mean					46,141	
Permian							
Upper							
Carboniferous							
	North America	U.S.A.	40,279	Dunkard Series (W. Virginia)	a	44,122	18
		U.S.A.	40,283	Mauch Chunk formation	t	43,127	22
		CANADA	46,294	Hurley Creek formation	t	39,125	44
		CANADA	46,295	Cumberland formation	t	36,125	41
		CANADA	46,296	Pictou Group	t	41,132	40
		CANADA	46,296	Prince Edward Island sediments	t	40,126	5
		CANADA	46,297	Prince Edward Island redbeds	t	42,133	40
		CANADA	47,296	Prince Edward Island Basic Intrusive	a	52,113	25
		CANADA	48,294	Pre Pictou	a	24,133	5
		CANADA	48,295	Bonaventure formation	t	38,133	40
	Mean					40,127	

TABLE 1 (cont'd.)

Age	Paleoblock	Location	Lat., Long. °N, °E	Particulars	Test[+]	Pole °N, °E	Ref.[1]
	Europe	U.S.S.R.	38,038	Asselian Stage Cupriferous/ sandstone	a, t, 1.0	40,152	143
		FRANCE	44,007	Estérel, Igneous and sedim. rocks	a, t	49,145	1,3,5, 39,47
		U.S.S.R.	48,038	Red Clays of Asselian stage	1.0	46,166	142
		U.S.S.R.	48,038	Red Clays of Asselian stage (Lower part)	1.0	43,160	144
		U.S.S.R.	48,038	Araucarite Suite, Donbass	1.0	38,170	152
		U.S.S.R.	48,038	Clays and Aleuro-lites Kasimovian stage	0.8	40,182	154
		U.S.S.R.	49,038	Clays and Aleuro-lites, Avilov suite	0.8	40,184	153
		GERMANY	50,008	NAHE Igneous rocks	a	46,167	31
		POLAND	50,020	Volcanics, Krakow District	a	43,165	3
		U.K.	50,356	Devon Sediments (3 sites)	a, t	46,165	10
		POLAND-CZECH.	51,016	Sedimentary and Igneous rocks	a, t	39,177	4
		U.K.	51,356	Exeter Lavas	a	50,149	47
		U.K.	51,356	Exeter Lavas	a, t	46,165	11
		FRANCE	54,006	Volcanics, Nideck-Donon	t	47,169	39

Age	Paleoblock	Location	Lat., Long. °N, °E	Particulars	Test+	Pole °N, °E	Ref.[1]
		U.S.S.R.	54,052	Clays of Upper Tartian Substage	0.8	49,173	123
		U.S.S.R.	55,038	Clays of Gzhelian stage	a, 0.8	46,177	149
		U.S.S.R.	55,053	Upper Tatarian Substage Transvolga Region	0.8	45,173	126
		SWEDEN	56,014	Skane Dolorite dykes	a	37,174	35
		U.S.S.R.	56,014	Clays of Czhelian stage (Moscow region)	1.0	42,167	148
		U.S.S.R.	56,039	Red clays, Kasimovian stage	a	42,169	155
		U.S.S.R.	56,055	Red Clays, Ulfimian stage	0.8	40,168	134
		U.S.S.R.	57,055	Red Clays, Kazanian stage	0.8	40,167	133
		SWEDEN	59,013	Mt. Billinger and Mt. Hunneberg Sills	a	35,175	35
		NORWAY	60,010	Igneous Complex of Oslo	a	47,157	16, 43
	<u>Mean</u>					<u>44,168</u>	
Lower Carboniferous							
Upper Devonian							
	North America	U.S.A.	45,292	Perry Lavas (Maine)	a	24,128	34
		CANADA	45,293	Perry Fm. Volcanics	a	26,109	5
		CANADA	45,293	Perry Fm. Sediments	a	35,121	5

TABLE 1 (cont'd.)

Age	Paleoblock	Location	Lat., Long. °N, °E	Particulars	Test[+]	Pole °N, °E	Ref.[1]
		CANADA	45,293	Perry Formation	t	32,118	37
		CANADA	46,296	Maringouin Formation	t	34,117	43
		CANADA	46,296	Shepody and Enrage Formations	t	34,119	42
		CANADA	48,301	Codroy Group (Newfoundland)	a	30,127	5
	Mean					31,120	
	Europe	U.S.S.R.	52,059	Porphyrites, Tournasian	a, t	20,163	194
		U.S.S.R.	55,036	Red Clays, Moscovian	a, 1.0	32,166	164
		U.S.S.R.	55,039	Moscovian Stage Ozery City	a, 1.0	34,170	163
		U.K.	56,357	Scottish Lavas	t	18,161	17,44
		U.S.S.R.	57,031	Redrocks, Upper Frasnian substage	a, 0.8	29,164	206
		U.S.S.R.	57,057	Bauxites and Hydro-hematites	1.0	35,181	209
		U.S.S.R.	58,033	Red rocks, Famen-mian stage	a, 0.8	34,158	205
		U.S.S.R.	59,034	Red clays, Visean stage	1.0	45,156	188
		U.S.S.R.	59,034	Oka Strata of Visean stage	a, 1.0	49,164	189
		U.S.S.R.	59,034	Tula Horizon, Visean stage	1.0	41,158	190
		U.S.S.R.	59,034	Red rocks, Famennian stage	a, 1.0	32,159	204
		U.S.S.R.	60,033	Red Loams, Lower Frasnian substage	a, 0.8	28,151	207

Age	Paleoblock	Location	Lat., Long. °N, °E	Particulars	Test[+]	Pole °N, °E	Ref.[1]
		U.S.S.R.	61,037	Red Beds, Visean	a,1.0	46,152	191
	Mean					36,162	

[+] The test column is to be read as follows: a = alternating field cleaning; t = thermal cleaning; the number is the stability measure as given in Ref. 21 of Appendix I; stability increases as number increases to unity.

[1] See Appendix for references used. References 76 to 209 refer to the number given in Reference 21.

[*] Although the stability measure is not very high, a reversal and a tilt test render their result acceptable with reservations.

[θ] a recent and acceptable (a and t cleaning) Upper Carboniferous result (pole: 44°, 159°) from the U.K. (Storetvedt and Gidskehaug, 1969) changes the mean by less than 0.4°.

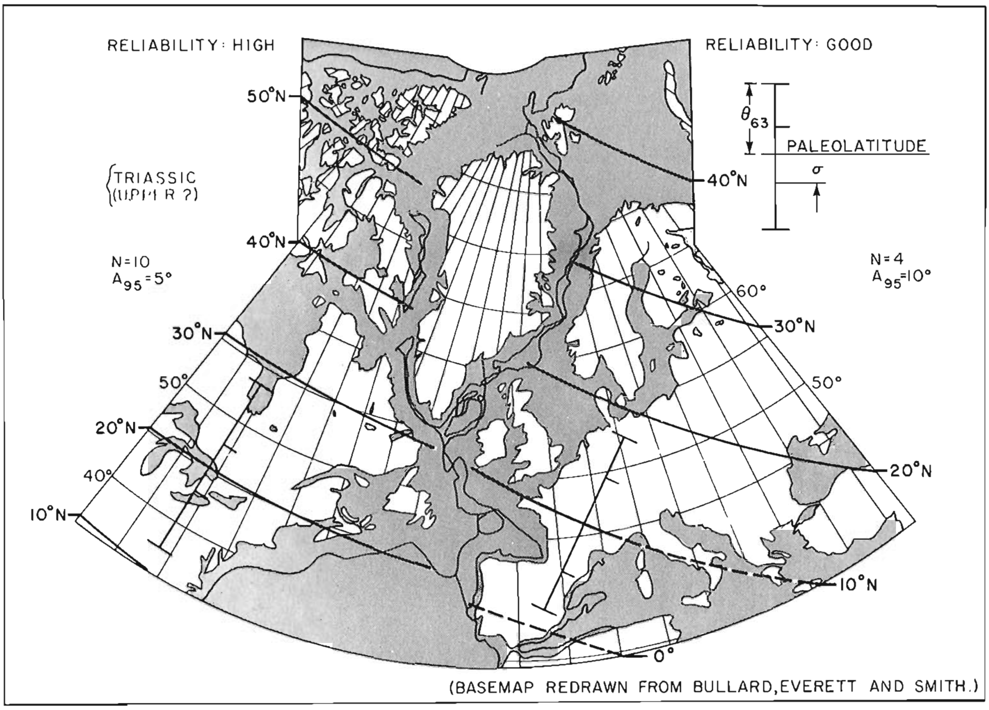

Figure 3. The difference in latitudes indicates rupture of the block. From this pattern it is suggested that the rupture occurred by strike slip motion. N, A_{95}, σ and θ_{63} as per Figure 1.

The agreement of the slanting angle between the two blocks may be quite significant. A simple latitudinal displacement of about 20° of the two parts with respect to each other is sufficient to obtain continuity of the paleolatitude lines. A longitudinal displacement is not required; in fact, if such a displacement is introduced, the latitudinal displacement must be increased. Furthermore, from the polar wandering curves, presented later, there is no evidence of a longitudinal displacement in the Triassic. The difference in longitude between poles of the two blocks was about 40° in Permian time; the difference in Triassic was still about 40°. This would indicate that the two parts initially underwent a sinistral strike slip without appreciable separation. Thus the paleolatitude results presented in Figure 3 indicate that the North Atlantic Ocean did not exist until after the Triassic, that is, not earlier than 180 million years ago. From the paleolatitude difference between Figures 2 and 3, it is likely that most of the displacement was acccomplished by a northward movement of about 20° of the North American block.

Jurassic-Cretaceous Paleolatitude Results

Although there are not many results for the Jurassic-Cretaceous period, the reliability of the North American results is good and a tentative paleolatitude map has been drawn (Fig. 4). Again, a northward movement of North America of about 15° is indicated. Because the paleopole is rather close to the present pole, the separation of the two blocks or the width of the North Atlantic cannot be established with any accuracy specially since the reliability of the European latitudes is only fair. However, the placing of the European block approximately at about its present location is not inconsistent with these data. From the North American results, the pole was at 172°E and 71°N. This position is about 180° from the present European longitude. So, if Cretaceous Europe is placed in the present European location or thereabouts, the Cretaceous latitudes would be parallel to and smaller by 19° than the present latitudes. The map indicates that they are parallel and smaller by about 30°. Then although the data are insufficient to determine the width of the North Atlantic in Cretaceous time, they are more consistent with the idea that the ocean was at least partly open. It may be tentatively suggested that the discrepancy of 11° (19° ~ 30° mentioned) latitude is an indication that the European block had not yet reached its present location relative to the pole.

Polar Wandering Paths

Pole displacements with respect to western Europe and eastern North America have been plotted from the same set of data in Figure 5. The circles of confidence at the 5% level are shown. For the European Jurassic-Cretaceous (pole 4), N=2 and its circle of confidence should be regarded as a rough indicator only.

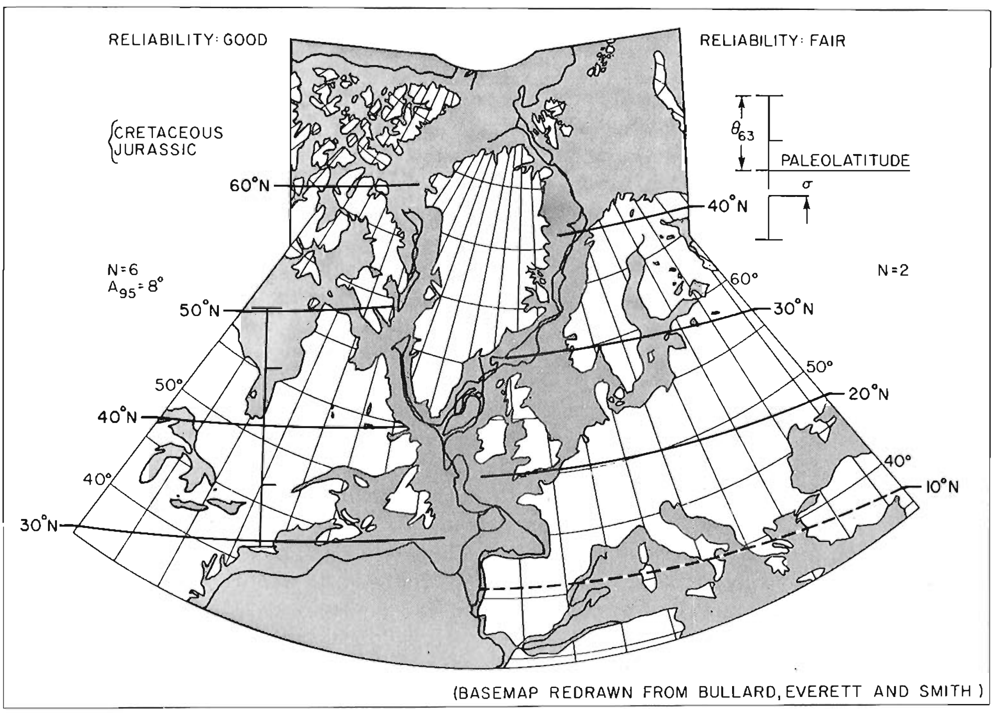

Figure 4. It is suggested (see text) that the Atlantic was then partly open. N, A_{95}, σ and θ_{63} as per Figure 1.

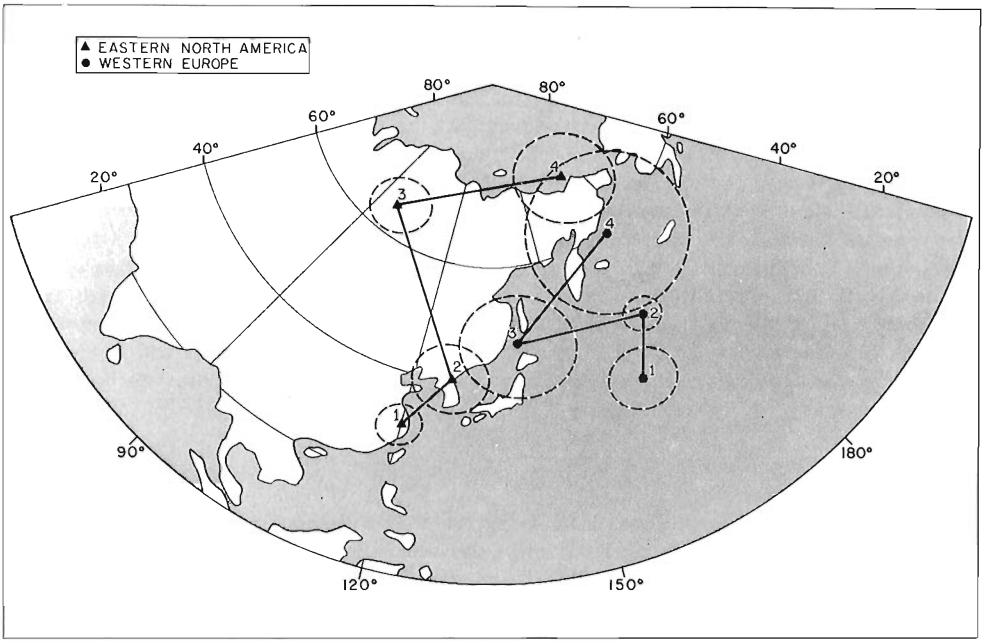

Figure 5. Polar wandering paths of the two land masses. The numbers refer to the four previous figures. Circles of confidence at the 5% level are shown.

There is some similarity between the two polar wandering paths and the westward movement of the pole during Triassic time is clearly shown. The distance between the European and North American curves is approximately constant, up to the Upper Triassic being about 40°. The separation of the Cretaceous poles is about 13° i.e. the separation has been reduced to one-third of the maximum, indicating that with respect to the pole the two blocks have travelled two-thirds of the distance that they had to travel from Triassic time to the present. This does not necessarily mean that the North Atlantic Ocean then opened to two-thirds of its present size since the displacement might not have been linear but much more complex.

To summarize the Mesozoic results, it seems that the rupture of the two blocks was by strike slip movement and occurred prior to the Upper Triassic epoch, and that the opening of the North Atlantic did not really start before Middle or late Mesozoic time. This opening time is roughly in agreement with the conclusion reached from extrapolation of sea-floor spreading results. For example, a latitudinal rupture followed much later by the start of the opening of the North Atlantic is sufficient to explain a uniform spreading rate of 1.4 cm/year (Emery et al. 1970; Fig. 35) without necessitating a reduction (0.8 cm/year) of spreading rate for the period 110-270 million years ago.

The gaps to be filled in order to establish a more definite picture are the obtaining of more paleomagnetic results from Upper Permian (especially North American) and Lower Triassic rocks in order to establish the time of rupture, and results from Jurassic and Cretaceous rocks in Europe in order to ascertain the time of opening of the North Atlantic Ocean.

ACKNOWLEDGMENTS

I wish to thank E. Irving for his suggestion to use paleolatitude maps in this work and for placing his library on paleomagnetism at my disposition. The help of T. Whillans with the compilation of the results was greatly appreciated. I am indebted to E. Irving and P.H. Serson of the Earth Physics Branch, and to P.J. Hood and E.J. Schwarz of the Geological Survey of Canada for critical reading of the manuscript.

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(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 23-42)

3. SUBSIDENCE AND FRACTURING ON THE CONTINENTAL MARGIN OF EASTERN CANADA

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Abstract

Many of the world's continental margins developed through rifting, as zones of subduction, and as zones of transform faulting. The seaboard of eastern Canada developed predominantly by rifting; however, it developed also in part by fracturing along transform faults. We show that there was substantial subsidence associated with rifting on the eastern seaboard, and that the changes in rates of subsidence are characteristic of its history. The rate increases in the early stages of opening, a phenomenon which could perhaps be used to date the beginning of rifting, and slows in the later stages, associated either with increase in distance from the spreading axis, or with cessation of spreading. The nature of the fracturing along lines of weakness may be similar to the deformation found at margins elsewhere now occupied by active transform faults. These characteristically possess a graben-and-horst structure, similar to those developed in eastern Canada, and this suggests that similar structural features should underlie the southern margin of the Grand Banks.

INTRODUCTION

Many of the world's present continental margins developed in one or more of three ways. Some evolved through "rifting" - development of an oceanic ridge or zone of accretion beneath a continental region. Others are now occupied by zones of subduction. A third type is now or was formerly associated with a transform fault. It is convenient for brevity to call these three "rifted", "trench" and "transform fault" margins. The margin of the eastern seaboard of Canada developed in part through rifting, and associated with this was substantial uplift and subsidence ascribed by Sleep (1971) to thermal expansion and contraction. The margins were associated also with transform faults. The axis of rifting is offset along lines of weakness in the continent being split, and these lines subsequently control the location of transform faults in the ocean basin (Wilson, 1965). Consequently, in the early development of the ocean basin two parts of the splitting continent could be slipping by each other. Subsidence and slipping may have been responsible for many of the features now seen on our continental margins, and we attempt to examine them in this paper.

It is clear that there has been substantial subsidence on the margin of the eastern seaboard of North America. The shallow water sediments found in deep wells on the Atlantic coast is evidence of this. The evidence for fracturing associated with "slipping" is less clear, because so much of any evidence that may exist will have been covered by sediment. Consequently we have chosen to examine a margin known to be associated with a transform fault which is active now, the west coast of Canada. The features seen here allow comparison with the results of studies already completed on the east coast and may be a guide to experiments which should be done in the future.

Rifted Margins

Rifted margins develop with associated uplift in the early phases, and subsidence in the later phases. The uplift and subsidence may be caused by thermal expansion and contraction. A scheme describing their evolution has been proposed by Schneider (1969), Vogt (1970) and Vogt and Ostenso (1967). The earlier phases of rifting are perhaps represented now by the Eastern African rift valleys, the Red Sea and the Gulf of Aden. The development in the later phases has to be inferred from the history of margins such as those bordering the Atlantic Ocean. This can be gathered from geomorphological studies (King, 1971), and from studies of sediment in wells drilled near the continental margin.

The subsidence of the margin led to the succession of thick sediments found along the Atlantic coast of eastern North America. The subsidence can be substantiated by examining the change in depth of samples deposited near sea level at different times, and by comparing it with the subsidence of the ocean basins (Fox et al., 1970; Menard, 1969). It seemed likely that it would be informative to investigate data from more restricted localities than those investigated by Fox et al. A variety of well logs from the Atlantic coast of North America have been examined, from Florida to the Grand Banks of Newfoundland, and the data from these compared with the data for subsidence in the Atlantic Ocean.

Subsidence on the Eastern Seaboard of North America

Data

We have used data from individual wells from a variety of areas, the Florida shelf, Georgia, North Carolina, Maryland, the Grand Banks, and the continental margin off northeastern Newfoundland. The sources of these data are given in Table 1, and the locations are shown in Figure 1.

The two wells from the Florida shelf are the Joides holes J-1 and J-2 (Joides, 1965). These were chosen (in preference to J-3 and J-6) because the sediments in them appeared to be demonstrably of shallow water origin. The wells drilled by Amoco Canada-Imperial Oil Limited on the Grand Banks seem suitable for the same reason (Bartlett and Smith, 1971). Of the two wells (Grand Falls and Tors Cove) for which information is available only one (Grand Falls) has been used, the region of the other having been demonstrably affected by salt diapirism. The data for Tors Cove is also more limited. Orphan Knoll was drilled by Joides in 1970, and preliminary information was taken from Laughton (1970), supplemented by A.S. Ruffman (personal communication). Orphan Knoll is interesting because it may be a "continental" fragment similar to comparable fragments off Portugal for example (Black et al., 1964). It is also interesting because the Knoll lies at the margin of a small ocean basin, the Labrador Sea, which was opening once, but unlike the areas of the other wells at the margin of the Atlantic, it is not opening now (Hood and Bower, 1971; Le Pichon et al., 1971). Only the data for samples older than 60 m.y. have been used in the case of Orphan Knoll, these samples being of shallow or fairly shallow water origin. Stratigraphically higher samples may have originated in deep water, and the depths in which they formed will be uncertain; the uppermost certainly originated in deep water because the present depth of the Knoll is 1800 m.

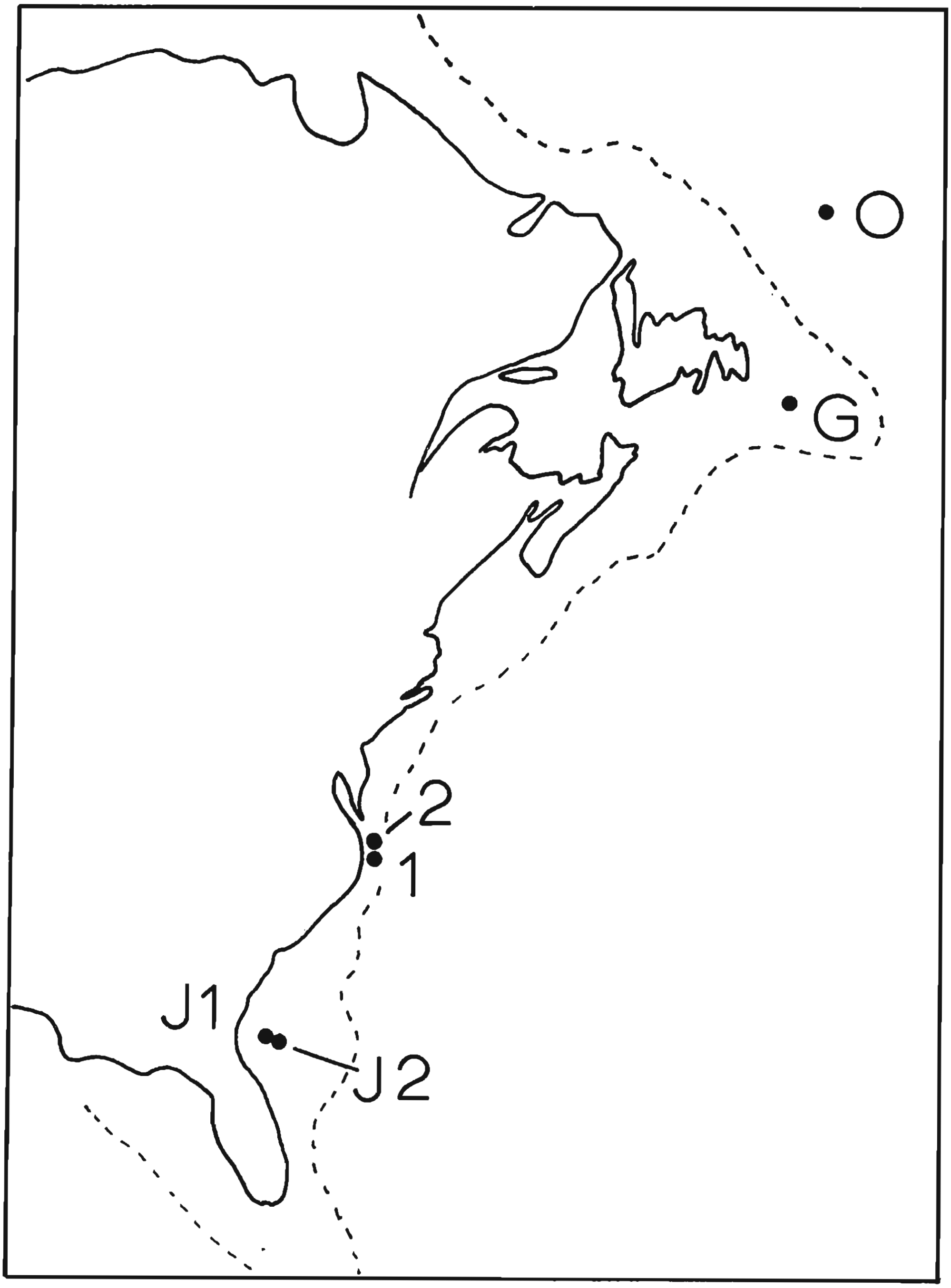


Figure 1. Location of wells on the emerged or submerged Atlantic coastal plain of eastern North America used in this paper. O-Orphan Knoll; G-Grand Falls; 1-Esso Hatteras Light Well No. 1; 2-Esso North Carolina Well No. 2; J-1, J-2, - Joides, off Florida.

Table 1

Sources of Data

1.	Orphan Knoll	Laughton, 1970
2.	Grand Falls	Bartlett and Smith, 1971
3.	Esso Standard Oil Co., Hatteras Light Well No. 1	Swain, 1951; Maher, 1965
4.	Esso Standard Oil Co., North Carolina Esso Well No. 2	Swain, 1951; Maher, 1965
5.	Joides wells J-1, J-2	Joides, 1965
6.	Wells in Georgia	Herrick, 1961
7.	Joides Leg 2	Peterson, 1969; Emery *et al.*, 1970
8.	Joides Leg 3	Maxwell, 1970
9.	Joides Leg 14	Unpublished Joides manuscript report

The two wells from North Carolina, Esso Hatteras Light Well No. 1 and North Carolina No. 2 are valuable because they are both deep, and have been studied in some detail.

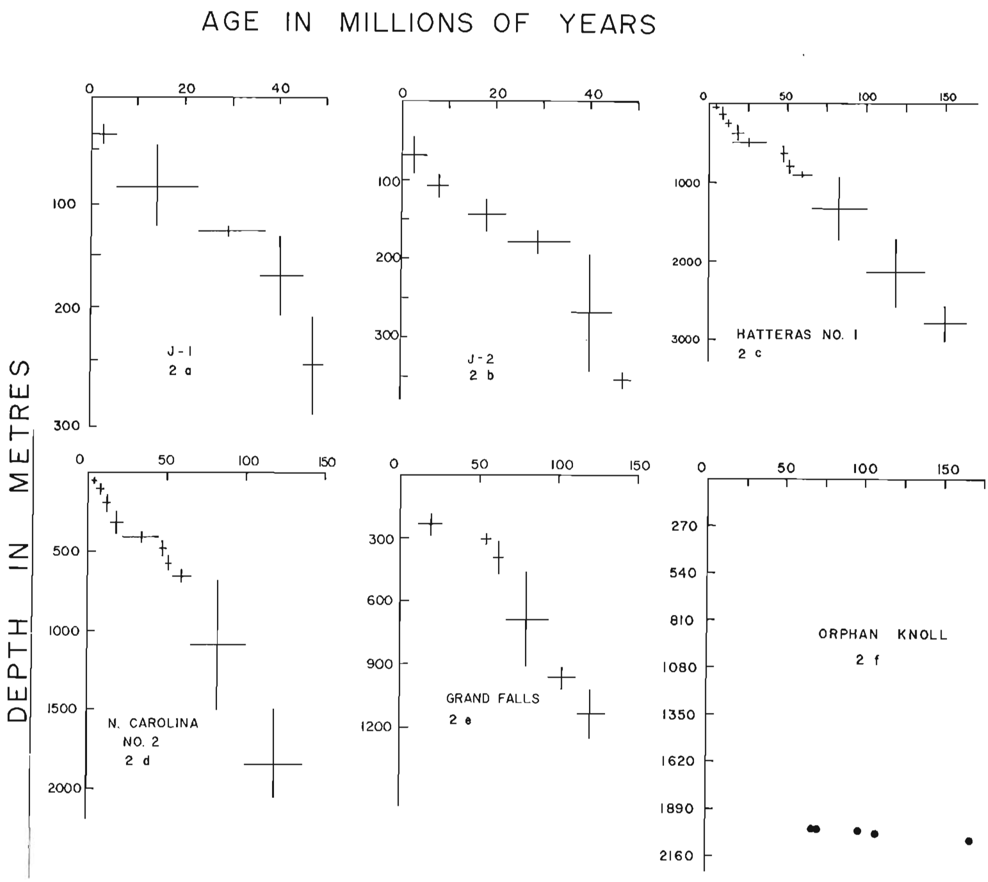
Subsidence has been estimated by plotting depth below the sea floor against age, the implicit assumption, perhaps crude, being that an horizon of age t at depth h has subsided h in the last t years. Various errors may arise.

Errors due to uncertainty in age are of three kinds. The absolute time-scale may be incorrect; an horizon's age may be known only within rather broad limits; an horizon may be incorrectly dated or correlated. We have used the time scales of Berggren (1969) for the Tertiary, and the Geological Society of London (1964) for the Mesozoic. The correlation between North American and European terminology is the one proposed by Maher (1965).

We have no way of estimating the uncertainty due to error in the absolute values of the time-scale. The effect of the uncertainty due to the possible range of an horizon's age is illustrated in Figure 2. In this figure, if the age "Campanian to Maestrichtian" is reported the range in ages would be 76 to 45 m.y. A possibly more serious source of error may lie in the actual age assigned to an horizon.

An example of this can be seen in Figure 3. The depths to the Woodbine, Washita and Frederiksburg in the Hatteras Light Well No. 1 reported by Swain (1951) and Maher (1965) are different. This alters substantially the form of the depth-age curve, and consequently if the relationship is exponential, the time constant of subsidence. This could be an isolated case, and there is no way of checking if only one person has documented a well. However Sleep (1971) has shown that the rates of subsidence below the Woodbine are apparently very rapid; this would result if the correlations reported by Maher (1965) for the Atlantic Coast were all systematically in error in the same sense as may be the case in the Hatteras Light Well No. 1.

Uncertainty concerning the depth of water in which samples originated is also, of course, a source of error whose magnitude is hard to estimate. The authors of the Joides (1965) report used in one instance the criteria of good sorting and the broken nature of calcareous skeletal debris to reach their conclusion that the limestones in question originated in shallow water. In the



(a) Joides well J-1 off Florida;
(b) Joides well J-2 off Florida;
(c) Esso Standard Oil Co., Hatteras Light well No. 1, Dare County, N.C.,
(d) Esso Standard Oil Co., North Carolina Esso well No. 2, Dare County, N.C.,
(e) Amoco-Imperial Grand Falls well;
(f) Joides well site III, Orphan Knoll.

Note that the ranges in ages and depths reported have been plotted except in (2f) for which the ranges are too small to reproduce

Figure 2. Subsidence estimated from wells on the margin of eastern North America. Depth in metres is plotted against age in millions of years. Locations are shown in Figure 1. Sources of data are given in Table 1.

case of the Joides holes J-1 and J-2 errors of only several tens of metres could alter the apparent significance of the data. The apparent subsidence seen in the Grand Falls well is greater, so that errors would have to be in the range of 100 m or more, and be systematic, to affect the significance of the curves plotted. If there has been uplift at any stage we might see gaps in the stratigraphic section. Such an effect would be difficult to distinguish from simple erosion unless sub-aerial sediments are present and have been recognized.

Apparent subsidence due to compaction cannot be estimated unless the thickness of the whole sedimentary section above "basement" is known, together with the lithology of the whole section. The effect can be described qualitatively however. A sample from a depth below present sea-bottom of say 100 m is anomalously deep because the extra load of 100 m of sediment above will cause the section beneath to consolidate. If subsidence is measured with the present surface as zero the apparent subsidence measured will be larger than the true subsidence. At basement itself there is no correction for consolidation because there is nothing beneath to consolidate. At the present surface there is also no consolidation correction because there is no additional load above. The correction between basement and surface can be estimated using the tabulations given by Hamilton (1959). We did this using the Esso Standard Oil Hatteras Light Well No. 1 (Swain, 1951), approximately 2800 m deep, which penetrates "basement" beneath Upper Jurassic sediments.

The corrections were estimated assuming the whole section was clay and shale (which it is not), and this should show the maximum likely effect. The estimates showed that consolidation can be ignored at the present stage of these studies. The qualitative conclusions reached later would be reinforced if consolidation were taken into account.

Results

The data from six wells illustrate the general nature of subsidence with time along the eastern seaboard. Depths against ages are plotted in Figure 2. The curves for Grand Falls, Hatteras Light Well No. 1, North Carolina Well No. 2 and Florida Joides Wells J-1 and J-2 show that the subsidence is roughly exponential with time. Subsidence through the later part of the Jurassic, the Cretaceous and the early part of the Tertiary was relatively rapid, the rates being in the range of 1 to 2.5 centimetres per thousand years (cm/ky), but was slower in the later part of the Tertiary. The Grand Falls and Hatteras Light wells show the whole phenomenon, the Joides wells only the subsidence for the last few ten of millions of years, the wells being shallower.

Orphan Knoll illustrates a different phenomenon. There was apparently little subsidence until a time later than approximately 75 million years ago, after which the rate of subsidence must have increased. The time at which this occurred coincides approximately with the time of opening of the Labrador Sea (Le Pichon et al., 1971). Consequently we suggest what follows as a working hypothesis. Before the main phase of opening of an ocean basin, subsidence is slow; during the main phase of opening, subsidence is rapid, but the rate decreases with time. The decrease comes about in association with the increase in distance of the margin from the axis of the spreading centre (the cases of Grand Falls, Hatteras Light, North Carolina No. 2, J-1 and J-2), or with decrease in the rate of spreading (perhaps for the case of Orphan Knoll).

The subsidence of margins has been treated quantitatively by Sleep (1971). He has shown that the subsidence caused by thermal contraction after initial uplift can be expressed as

$$S = k\left[\exp(at) - 1\right]$$

where S is the sediment accumulated since time t years ago at a given distance from the continental margin, and k is a constant, for the given distance. The value of the parameter a is the reciprocal of the time constant. Sleep assumed

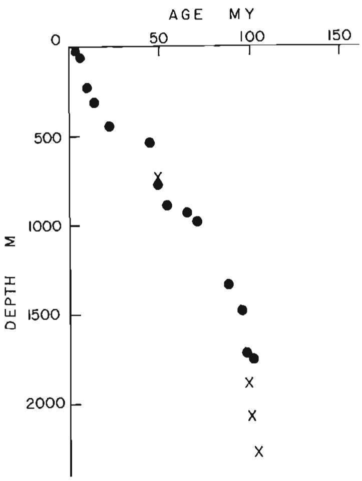


Figure 3.

A comparison between ages in millions of years and depths in metres for Hatteras Light well No. 1 following correlations assigned by Swain (1951), and where different, Maher (1965). The data for tops of horizons are plotted.

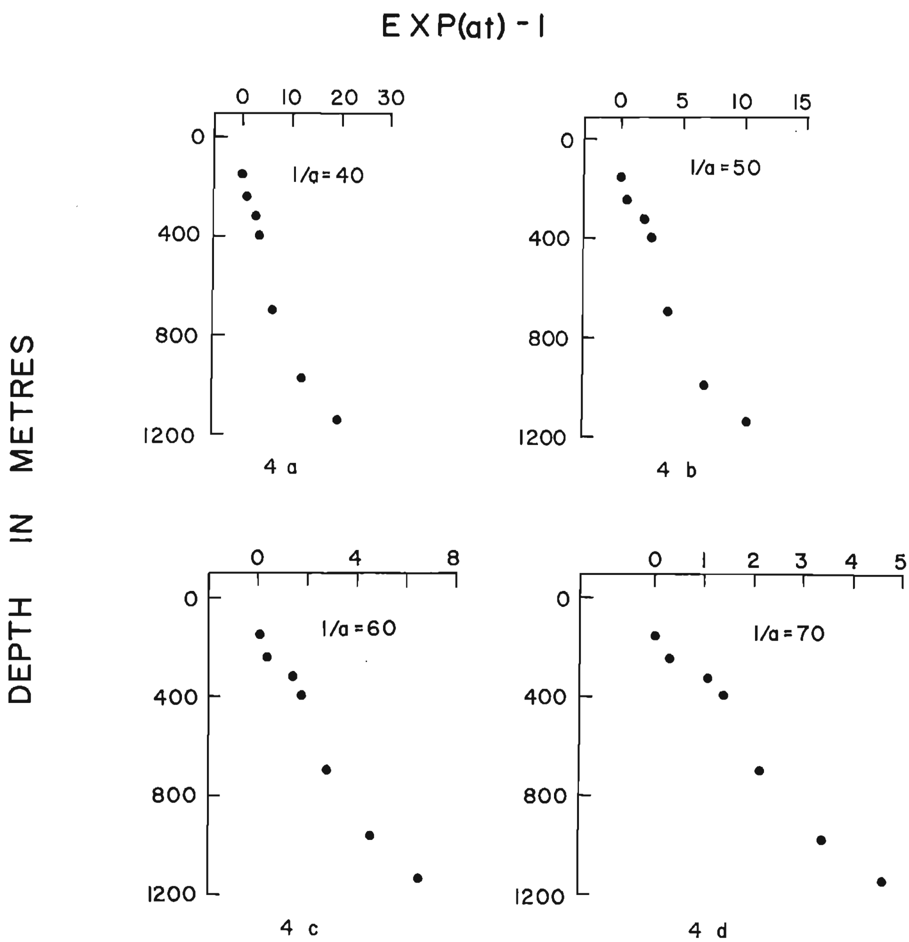


Figure 4. Depth of horizons in the Grand Falls well plotted against values of (exp(at)-1) with values of (1/a) of (a) 40, (b) 50, (c) 60 and 70 m.y. See text.

that (1/a) has the value of 50 million years, using studies of mid-ocean ridges as a guide. This deduction can be tested by plotting values of S against [exp(at) – 1 . If the correct value of the time constant (1/a) is assumed a straight line should result which passes through the origin.

Figure 4 shows the relationship for the Grand Falls well. The appropriate value of (1/a) may be greater than 70 million years. We claim no particular virtue for this number, it is critically dependent upon good data. However, the data so plotted are approximately linear and if good data become available it could be worth-while to investigate systematically the value of this constant at different margins.

It is intuitively obvious that subsidence should vary across a continental margin. If thermal expansion and contraction control uplift and subsidence then as new ocean floor is generated at an axis of spreading different part of the margin are heated and cooled at different times, and by different amounts. We may also expect to observe any effect which will arise if isostatic compensation is not local, but is regional. We examined these geographical aspects using data from wells in the State of Georgia.

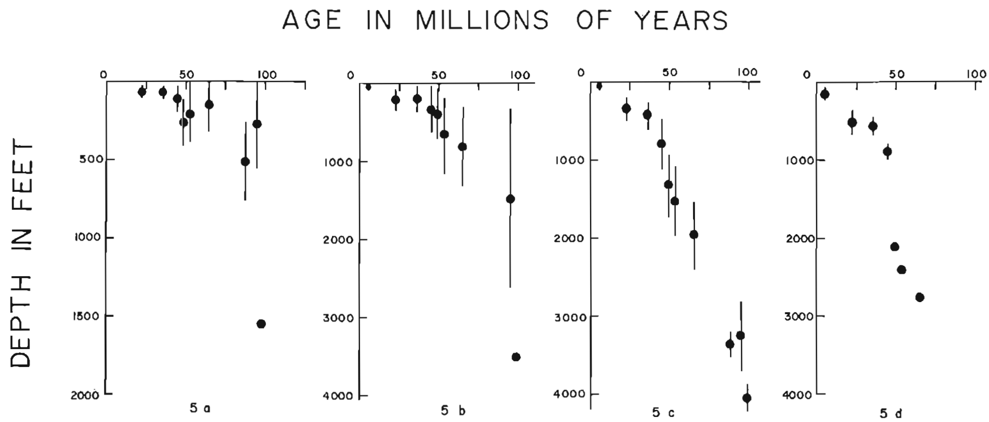
These wells were chosen because the data was extensive, and had been compiled by one person (Herrick, 1961). One well from each county was selected (usually the deepest well). The wells were then grouped according to distance from the fall line, the boundary between the Coastal Plain and the Piedmont, 0 to 225 km, 225 to 450 km, 450 to 660 km and over 660 km. The mean depth and standard deviation for each horizon within each zone was then calculated, to give in effect "average" subsidence curves for the four zones. These are shown in Figure 5. It is obvious that there is geographical variation in subsidence. The rates increase with distance from the fall line.

Subsidence in the Atlantic Ocean

The subsidence at the margins appears to be comparable to subsidence of mid-ocean ridges. Menard (1969) has suggested that in the first 10 million years a ridge subsides at approximately 10 cm/ky, and 3 cm/ky thereafter. The data of Joides for depth and age of basement found in Legs 2 and 3, and on part of Leg 14, in the North and South Atlantic (Fig. 6) are plotted in Figure 7, using information drawn from Peterson (1969), Maxwell (1970) and Emery et al. (1970). Figure 7 shows the depth below sea level and age of the igneous rock recovered at the bottoms of the holes (dated paleontologically from fauna in the overlying sediment). It seems unlikely that the ages are in doubt by more than a few million years. Maxwell et al. (1970) show for the South Atlantic sites that ages deduced from magnetic field anomaly patterns are in remarkable agreement with the ages assigned for paleontogical reasons.

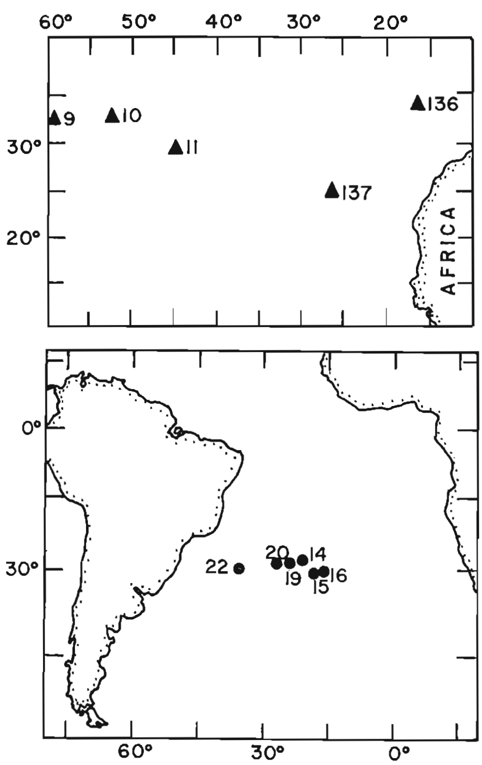
We assume, perhaps erroneously, that the crest of the mid-ocean ridge was at the same depth with respect to sea level at whatever time it formed. If this is so, Figure 7 shows the subsidence of the ridge. The rate between the interval 20 to 120 million years is 2 cm/ky. This is comparable to rates of subsidence at the margins during the main phase of opening. Presumably if the rates of subsidence of the ocean basin adjacent to the margin on the one hand, the continent adjacent to the margin on the other were not similar, the lithosphere would break. If it does break, a zone of subduction could arise, as Dewey (1969) has hinted.

Only two wells from Leg 14 from the eastern North Atlantic have been plotted on Figure 6 because the information released so far is only



(a) Wells 0-225 km from the fall line
(b) 225 to 450 km
(c) 450 to 660 km
(d) over 660

Figure 5. Average subsidence for wells in Georgia. The means and standard deviations of depths are plotted against age. Points with no bars indicate that only single values are available.



(a) North Atlantic

(b) South Atlantic

Figure 6.

Location of Joides sites used in the compilation of Figure 7.

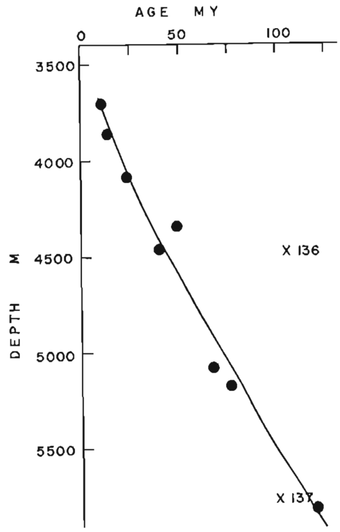


Figure 7.

Depth to basement in metres against age in millions of years found on Joides Legs 2, 3 and part of 14 in the North and South Atlantic.

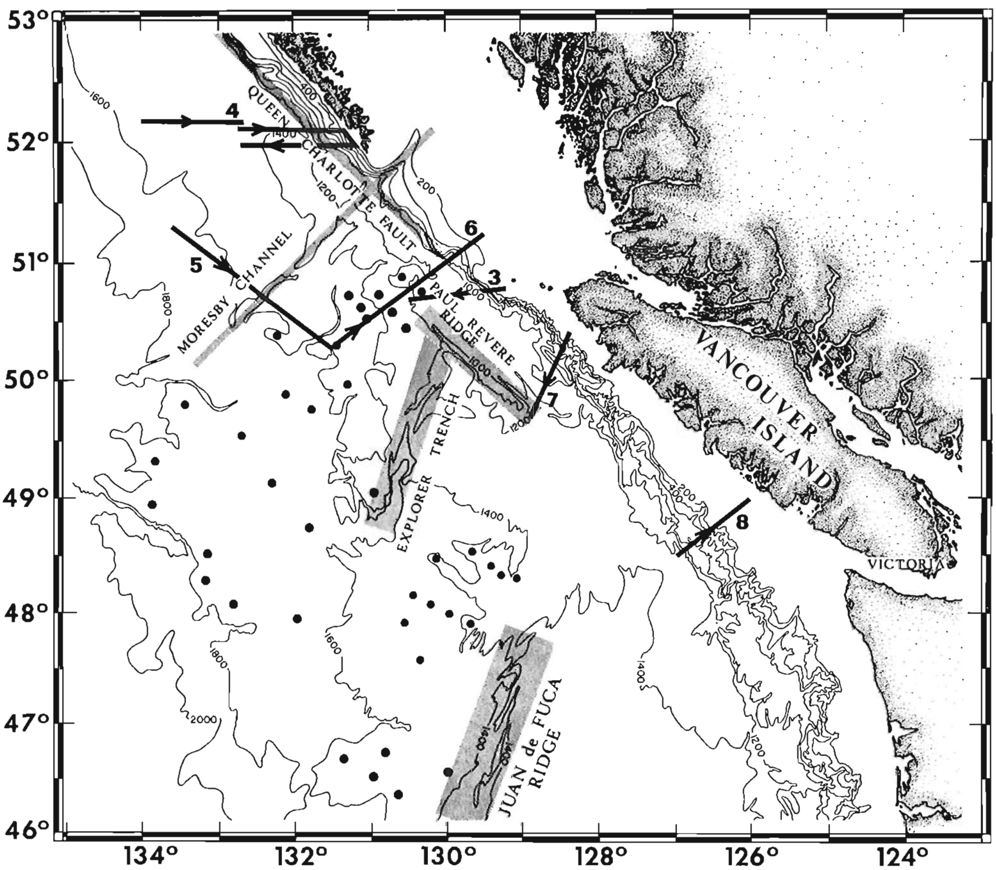


Figure 8. Major features of the northeastern Pacific Ocean west of British Columbia, and location of reflection profiles referred to in the text. (The dots are dredge stations, not referred to).

preliminary. It is worth noticing that the results from Site 136, southwest of Gibraltar, were reported as showing that igneous rock recovered from the hole is anomalously young with respect to the boundary of the magnetic quiet zone, that boundary considered as an isochron. It is also anomalously shallow (4477 m in depth, 110 m.y. old), and it will be interesting in future studies to see if this is due to the location of the site, relatively close to the Azores-Gibraltar "ridge".

One further point should be made. It is sometimes difficult to date the time of break up of the continents with certainty. Observations from deep wells may help. For example, the data for the Labrador Sea suggest that the region started spreading approximately 70 m.y. ago, in agreement with the estimates of others, as we have shown.

Transform Type Margins: Fracturing

We pointed out in the INTRODUCTION that some of the features of the eastern seaboard may have developed through fracturing along lines of weakness in the continent as it was split. Structures along offsets in the margin, for example the southern boundary of the Grand Banks, could have been caused by slippage along the opposing continental margin (see for example Wilson, 1965). It is perhaps curious that the structures found on mainland and offshore Nova Scotia which could have been associated with this type of fracturing (because they trend in an easterly direction) are basins bounded by normal faults. A good example is the structure associated with the Orpheus gravity anomaly beneath the Scotian Shelf (Loncarevic and Ewing, 1967; King and MacLean, 1970).

The evidence from the southern margin of the Grand Banks and its extension to the southeast (called by Le Pichon and Fox, 1971, the Newfoundland Fracture Zone) is still obscure. There is a thick Mesozoic and Tertiary sediment cover, and there has been so far inadequate study. Consequently we have sought evidence for deformation associated with transform faults at continental margins from another region, the west coast of Canada, where the Queen Charlotte Islands' fault is found (Fig. 8). Examination of the region may allow us to plan experiments on the eastern margin with intelligence. The examination should also bear upon the nature of fracturing in areas such as the coast of west Africa between Freetown and Lagos, and the north coast of Brazil west of Cap de Sao Roque. Le Pichon and Hayes (1971) have suggested that features such as the North Brazilian Ridge are associated with fracture zones between South America and west Africa.

We should point out however that the Queen Charlotte Islands' fracture zone arose in a different way from the fracture zones of the Atlantic. It is an example of a transform fault which has developed in association with a triple junction, and moreover bounds a wholly oceanic plate – the Pacific plate – and a plate bearing a continent – the American plate (Atwater, 1970; McKenzie and Morgan, 1969). The triple junction off British Columbia is a ridge – trench – transform fault junction (RTF). The Queen Charlotte Islands' fault is the transform fault, the Explorer "Trench" the ridge – a continuation of the Juan de Fuca Ridge along an unnamed fracture zone. The trench of the triple junction is the region of compression west and southwest of Vancouver Island. The evidence from this region is important because plates are interacting actively now, and because the margin is undergoing compression in one part, but strike-slip motion in an adjacent part.

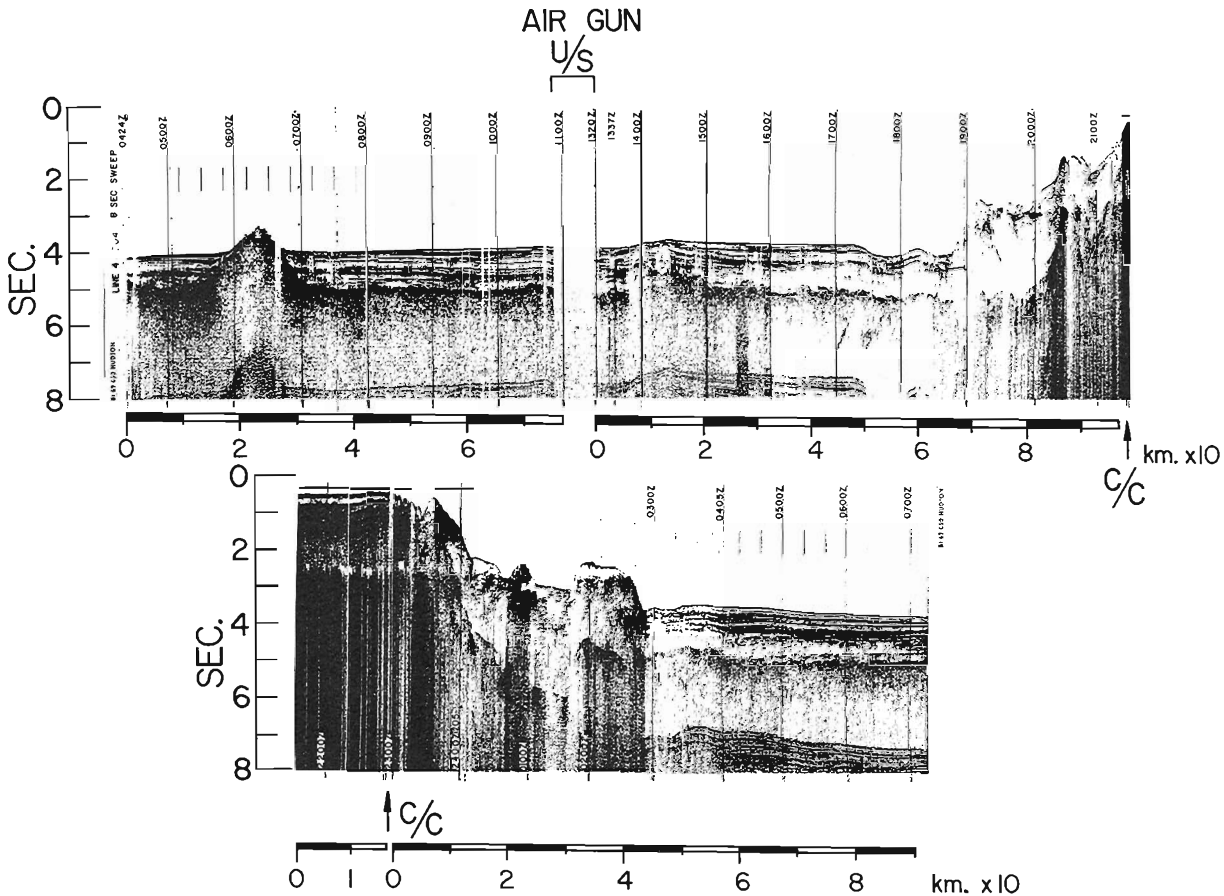


Figure 9. Air gun seismic record for Line 4N (northerly section of line 4, upper) and Line 4S (southerly section of line 4, lower).

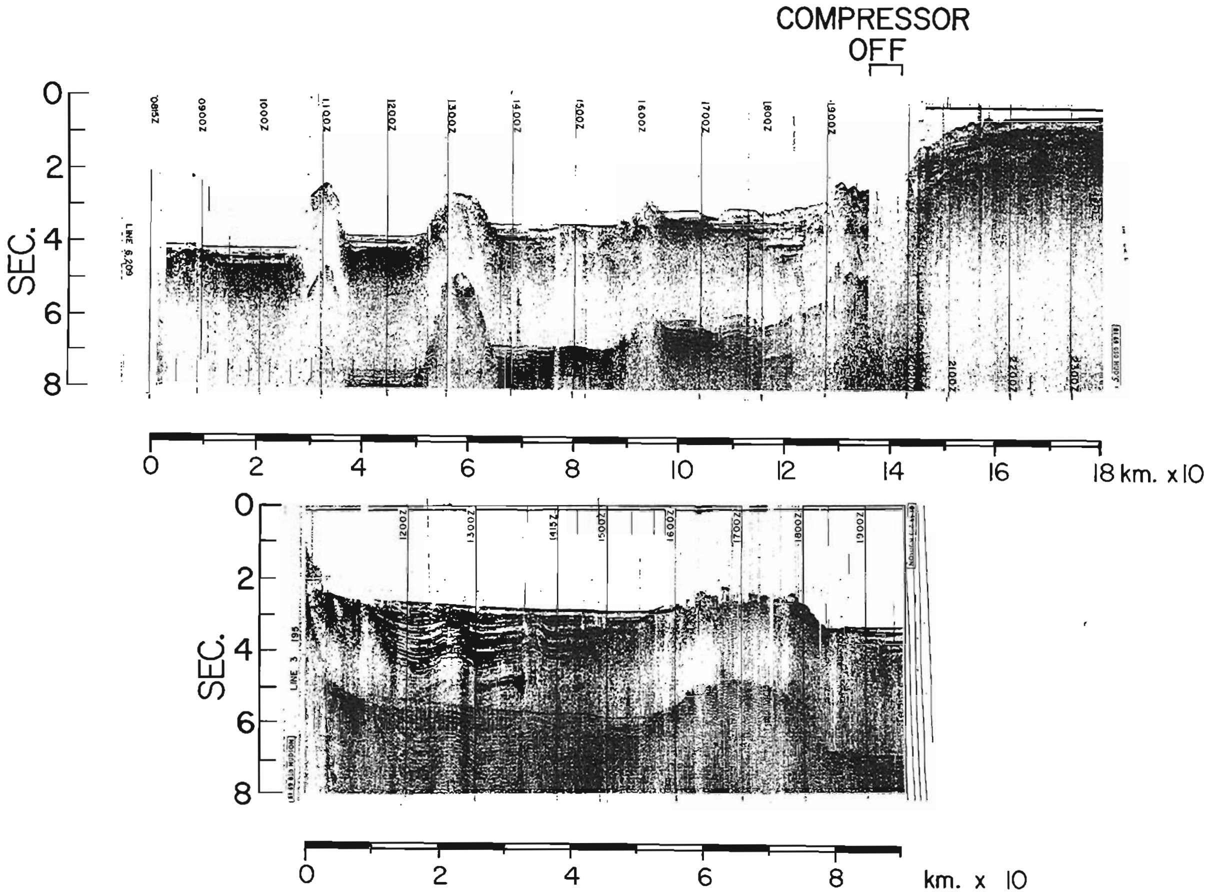


Figure 10. Air gun seismic profiler record for Lines 6 (upper) and 3 (lower). Note that the continental slope is to the left on line 3.

Six reflection profiles (Fig. 8) were obtained by C.S.S. Hudson in 1970 across the margin of British Columbia off the Queen Charlotte Islands and Vancouver Island (Srivastava et al., 1971). Lines 4N and 4S cross the Queen Charlotte Islands' Fault. Lines 6 and 3 cross the margin just north of the junction between the Explorer Trench (a spreading centre), the Fault and the Paul Revere Ridge (not a spreading centre?). Lines 7 and 8 cross the margin east of the Explorer Trench – they lie across the Juan de Fuca plate and the American plate boundary, whereas the northerly profiles cross the boundary between the Pacific Plate and the American Plate. We might expect to see a transition between "slippage" tectonics in the north to compressional tectonics on the margin in the south, reflecting motion between the Pacific Plate and the Juan de Fuca Plate, with the American Plate.

The Reflection Profiles

A description of the main features shown on these profiles has been published elsewhere (Srivastave et al., 1971) and only a brief account, relevant to the present discussion, is given below (Figs. 9, 10, 11).

The northerly lines: 4N and 4S (Fig. 9)

The top of layer 2 is relatively flat beneath the ocean basins, and is overlain by flat-lying sediment, deformed only slightly. Layer 2 rises to form seamounts with steep slopes (20°). The sediment beneath the ocean basin, over 1 km thick, is not deformed by the seamounts, and must have been deposited after the formation of Layer 2 and the seamounts. However, some deformation of the oceanic sediment is associated with rises in the top of Layer 2. Faulting is confined wholly to the continental slope, which is steep and narrow (about 3 km wide). A narrow zone of horsts and grabens may change the elevation of the top of Layer 2. At 0200Z on Line 4S a flat topped seamount appears to have originated as a horst. Layer 2, with oceanic sediment overlying it, and later transparent sediment on top of the oceanic sediment may have been faulted up. If this is true the "normal" faulting is later than the sedimentation in the adjacent ocean basin. The evidence is not good, however. Folded sediments are confined to the horst and graben region of the continental slope.

The central lines: 6 and 3 (Fig. 10)

Line 6 lies to the north of the triple junction, and line 3 to the south. The continental margin is much broader in this central region than it is to the north – it is approximately 100 km wide, depending upon the definition of the seaward limit. The zone over which the top of Layer 2 is deformed is also much broader. The more northerly profile, Line 6, shows a series of steps in Layer 2, the ocean floor stepping up as Layer 2 steps upwards towards the continent. A broad rise in Layer 2 can be seen on both lines (Line 6, 1700Z; Line 3, 1700Z). The depression between the rise and the continental slope is occupied by a thick sedimentary basin. The sediments of the basin seen in the southern line (Line 3) are broadly folded, with crests of anticlines more tightly folded than the intervening "synclinal" regions. This phenomenon could, perhaps, be caused by diapirism. The sediments of the basin are at least 3 km thick.

The southerly lines: 7 and 8 (Fig. 11)

The margin here is also broad, again about 100 km. Layer 2 beneath the ocean basin, seen on the easterly extremity of Line 7 is rather flat-lying. The sediment beneath the ocean basin is only about .25 km thick. The margin is intensely deformed. The broad rise in Layer 2 and the basin of line 3 appears again. Flat-lying sediment appears to occur in deep troughs, up to 3 km thick, as at 1030Z Line 7. This may however represent a continuation of the phenomenon seen in Line 3 - crests of anticlines could have broken through, leaving apparently undeformed sediment between. We see, as it were, a "ridge and trough" province formed by folding (and breaking) of sediments. The sea floor above this deformed region is not flat, except in occasional ponds. This suggests that in this region folding is active now. Alternatively, we could be seeing the effect of diapirism.

The reflection profiles and transform faulting

The main characteristics of the margin between the Queen Charlotte Islands and Vancouver Island are given in summary in Table 2. North of the Explorer trench the continental margin is narrow. Sediment folding is confined to the inner, continental side of the margin, and this folding may be associated with deformation at an earlier phase than that occurring now. Layer 2 is uplifted, associated with the development of horsts and grabens, a development which may be a recent phenomenon. This could be a mechanism for incorporating Layers 2 and 3 into the continental margin, and hence into the continent. Such a region could be fruitful for dredging. There is little deformation of the oceanic sediment adjacent to the transform fault, except in association with horsts and grabens.

By contrast, the region to the south where there may now be compression is associated with a wider, more deformed margin. Sediment folding is intense and conspicuous. The sediment which has been and is now being folded could be continental rise or oceanic sediment, and we see here a mechanism for incorporating such material together with oceanic crust into a continent without the phenomenon of a topographic trench or an obvious sinking plate.

The margin between profiles 6 and 4N possesses a "stepped" nature, it is a region of horsts and grabens. The margin to the south does not. Thus it appears as though this is a characteristic feature of the margin occupied now by the transform fault. The zone of grabens generally lies seaward of the zone of compression, and may be younger but this evidence is poor. The compressional structures, to the south of the Juan de Fuca-Pacific plate boundary, are not overlain by undeformed sediment and it appears that deformation is occurring at the present. This suggests that the Juan de Fuca plate is not coupled to the American plate but that the two are interacting.

It is not our purpose to discuss in any detail the present plate motions in this complicated area. Tiffin et al., (in press) have discussed some aspects of these. However, the results show that compression has occurred west of the Queen Charlotte Islands and this might be associated with the past occurrence of a zone of subduction in this area. It is possible that the triple junction between the Farallon, American and Pacific plates once extended at least as far north as the Queen Charlotte Islands. The mostion of this RTF junction to the southeast along the American plate boundary is consistent with plate motions some 10 m. y. ago according to MacKenzie and Morgan (1969).

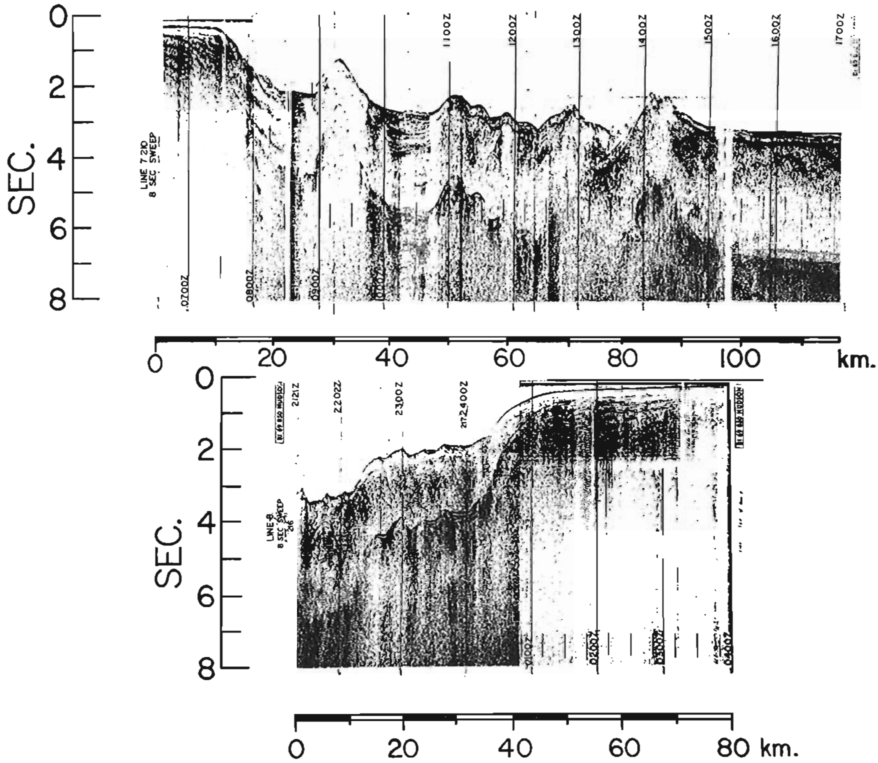


Figure 11. Air gun seismic profiler record for Lines 7 (upper) and 8 (lower).

Comparison with the eastern margin

Such studies that have been done suggest that many of the features observed over the margin off the west coast are to be seen south and south-east of the Grand Banks. Keen, Loncarevic and Ewing (1971) pointed out that the pattern of positive and negative gravity anomalies over the Grand Banks suggested that large blocks of continental crust underlay the Banks, the blocks separated by troughs and cracks. Auzende et al. (1970) made a study of the southern margin of the Banks and the southeast Newfoundland Ridge. They concluded that normal faulting could be observed on the continental block of the Banks, and that southeast of the Banks a "Newfoundland fracture zone" was marked by a series of horsts, and intrusions of igneous rocks.

It is obvious that similarities exist in the structures associated with a transform fault off the west coast and with a possible fracture zone on the east coast. However, we have not established criteria which would let us identify in a unique way a fracture zone in a situation such as may exist off the Grand Banks.

Table 2

Phenomena Observed on the Reflection Profiles

North	Line	Horsts, grabens and steps	Compression and extent
	4N	Yes	Yes , 22km
	4S	Yes	Yes , 22km
Profiles cross	6	Yes (over 80km)	Yes , 33km
triple junction	3	Not obvious	Yes , 45km
	7	No	Yes , 56km
South	8	No	Yes, at least 40km (record incomplete)

CONCLUSIONS

We have shown that subsidence has played an important role in the development of the margin of the eastern seaboard of Canada. The history of subsidence reflects the history of rifting and ocean-floor spreading. If accurate subsidence curves could be obtained they might aid in finding the time at which the main phase of opening started.

The fracturing associated with an active transform fault on the west coast of Canada appears to lead to zones of horsts and grabens. The history of a margin such as the west coast of North America is complex. The nature of triple junctions changes as ridges and their associated transform faults successively meet the margin. Consequently the margin will experience oscillations of trenches and transform faults along its length corresponding to successive RTF and FFT junctions.

Comparable deformation exists on the southern margin of the Grand Banks, and its southeasterly extension. Study in detail of this area could resolve two problems of general importance in the history of opening of the North Atlantic. One problem, mentioned in the previous paragraph is the time of opening – is it Triassic or Jurassic? The other is the line of opening – was it along the continental margin, or along the boundary of the quiet magnetic zone?

ACKNOWLEDGMENTS

The seismic reflection profiles referred to were obtained as a part of the HUDSON – 70 cruise. We thank L.H. King, J. Woodside, D.I. Ross, A.S. Ruffman, S. Srivastava, D.L. Barrett and L.H. King for help at sea and in the preparation of this paper.

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4. THE CANADIAN ATLANTIC CONTINENTAL MARGIN: PALEOGEOGRAPHY, PALEOCLIMATOLOGY AND SEAFLOOR SPREADING

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Abstract

Microorganisms, foraminifera in particular, are excellent indicators of environmental conditions and are essential for intercontinental biostratigraphic correlations and paleoecological interpretations of continental margin sediments. Planktonic/benthonic ratios, calcareous/arenaceous ratios, coiling directions, ultramicrostructure of test walls, microfaunal/macrofaunal ratios, oxygen isotope data, paleomagnetic data, and sediment associations have been utilized in the synthesis of Mesozoic to Recent sediments on the Canadian Atlantic continental margin.

A wide variety of depositional environments (coastal dunes, offshore banks, deltas, back reefs, warm shallow bays, and inner shelves) were present during its formation. Tectonism in the form of broad upwarping was intermittent. Salt dome tectonism, so characteristic of the margin area, locally alters this pattern but does not change the regional character.

An understanding of the relative position of the landmasses adjoining the Atlantic basin, and their origin, is essential for a proper interpretation of paleoecology, paleoclimatology and paleooceanography. This information is utilized to interpret watermass characteristics and circulation patterns in the north Atlantic from the early Mesozoic to the present.

INTRODUCTION

The biostratigraphy, sediment characteristics and depositional environments of the Canadian Atlantic continental margin have been described by Marlowe (1965); Marlowe and Bartlett (1967); Bartlett (1968, 1969); King _et al._ (1970); and Bartlett and Smith (1971). Bartlett (1967) suggested evidence for the presence of a proto-Gulf Stream off Nova Scotia at least as early as mid-Miocene time. He further emphasized that sediments underlying the continental shelves in the Atlantic region were a continuation of the Gulf Coast and Atlantic coastal plain physiographic province.

This report is an attempt to utilize the characteristics of present ocean circulation and the associated microfaunas, foraminifera in particular (Plates 1-3), as a guide to the interpretation of oceanic circulation in the past. In this respect it must be pointed out that planktonic foraminifera will be carried passively in surface currents whereas benthonic species will reflect the environment at the sediment water interface. The inferred position of the continents during the Mesozoic and Cenozoic periods and the similarities (cosmopolitanism) and dissimilarities (provincialism) of faunas and floras in widely separated geographic areas are paramount to this discussion. It was

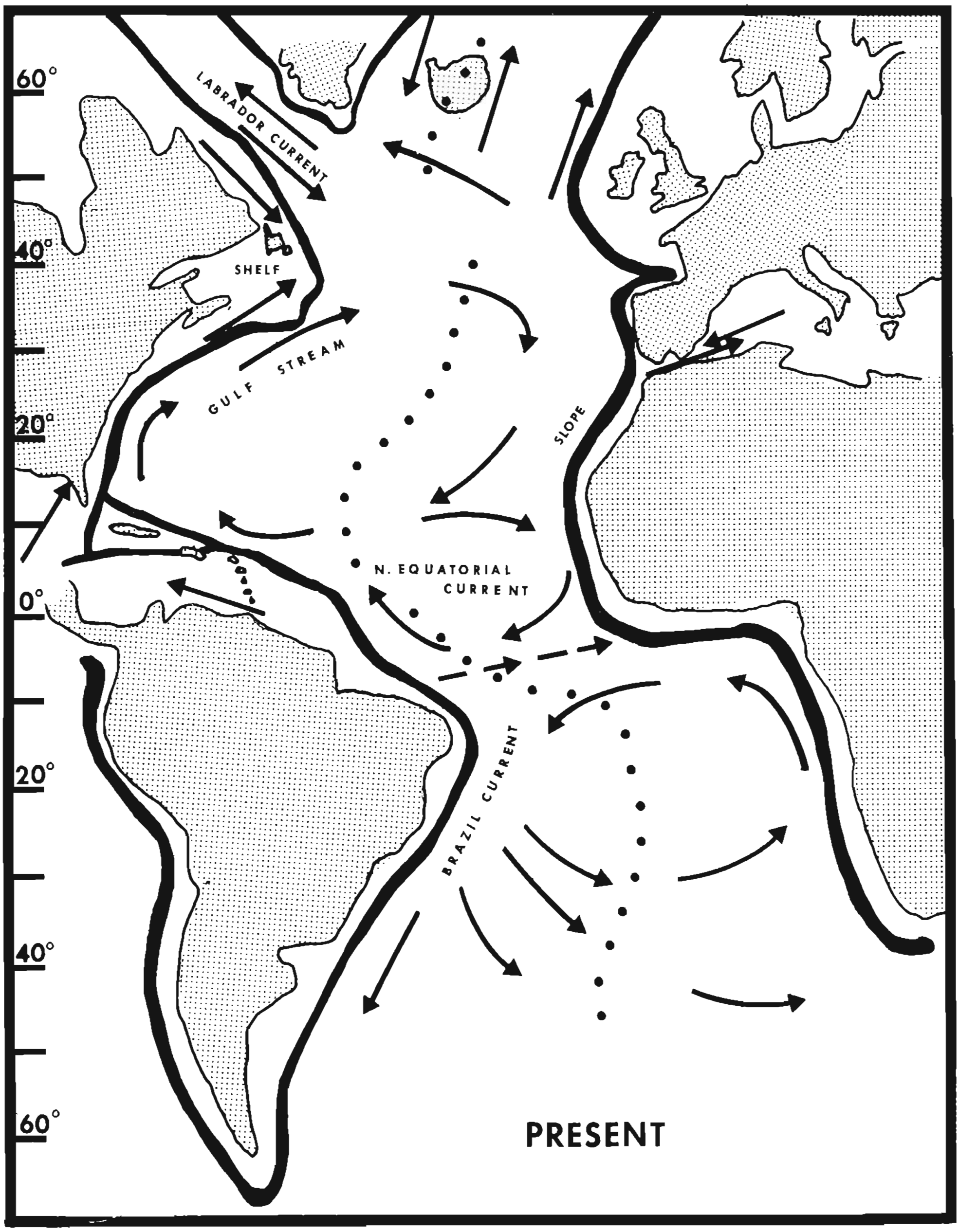

Figure 1. Surface circulation in the North Atlantic at present showing some of the principal currents affecting the Atlantic continental margin. Broken arrows at equator represent equatorial counter current. Heavy dark line approximates the 1,000-metre bathymetric contour.

suggested by Bartlett (1969), that watermass distribution during the Mesozoic and Cenozoic was quite different from that prevalent today. This was also suggested by Bandy (1960, 1964, 1967) when he noted that keeled globorotalids, globotruncanids and rotaliporids were limited to minimum sea temperatures of 17°C. The change in position of continental landmasses, more than one ocean basin and delay in opening or spreading of more northern areas were suggested as plausible reasons for these differences by Bartlett (1969).

Faunal provincialism in the North Atlantic since the late Miocene could be attributed to increasing isolation because of plate movement and the gradual closure of the Strait of Gibraltar that restricted Atlantic-Mediterranean circulation. Information published in Geotimes, v. 15, no. 10, p. 12-15, December 1970, from Leg 12 and Leg 13 of the GLOMAR CHALLENGER confirmed this hypothesis. Therefore, microfaunas from the continental margins are essential for developing concepts and interpreting the paleogeography, paleoclimatology and significance of plate tectonics in the Atlantic region.

The utilization of microfaunas in an interpretation of oceanic basins and continental margins requires an understanding of surface circulation (Fig. 1), wind-induced "Ekman Spiral" and density (thermohaline) circulation. The Ekman Layer is from 100 to 150 metres thick and generally moves at right angles to the prevailing wind direction. This surface layer provides an environment for reproduction, growth and transportation of planktonic foraminifera. Consequently, the distribution of planktonic organisms in ancient sediments will provide information on surface currents, winds and climates.

In the absence of continents the oceans would be dominated by east-west motion. The Gulf Stream, a body of water that is paramount to the discussion, is a strong current on the westward side of the Atlantic Ocean. It is a result of the return flow of water that was squeezed south by the wind-driven convergence of surface waters in the entire central North Atlantic. However, a piling up in the Gulf of Mexico, a "rubbing" against the continental margin and Coriolis force are essential for its formation.

Thermohaline circulation is induced by density differences in various watermasses as a result of salinity and temperature differences. Unrestricted circulation with both the Arctic and Antarctic Oceans is essential for the development of efficient thermohaline circulation, the main driving force of the present oceans. Mixing of various watermasses occurs between the surface and deep water layers through an upward movement, convergence and divergence, near the equator, of the deep cold bottom waters. Consequently the deep and bottom water may reach the surface and be governed by wind-induced motion until it again cools, becomes more dense, and sinks, to complete the cycle. The present deep and bottom waters are relatively uniform over a wide geographic area, and although they may have been significantly warmer during the Mesozoic and Cenozoic, they are assumed to have been consistent in both physical and chemical characteristics during those periods. The development of provincialism in benthonic faunas must therefore, be attributed to progressive geographic isolation because of continental drift, as a result of the movement of major tectonic plates, development of deep and wide ocean basins, and changes in surface circulation (major transporter of larvae) in the North Atlantic.

Planktonic foraminifera (Plates 1-3) are distributed in broad latitudinal belts and narrow vertical zones in the present oceanic watermasses. Recent distributions as noted by Bé (1959); Bé and Hamlin (1967); Bé et al. (1971); Bartlett et al. (1968); Bradshaw (1959); and Parker (1967) can be

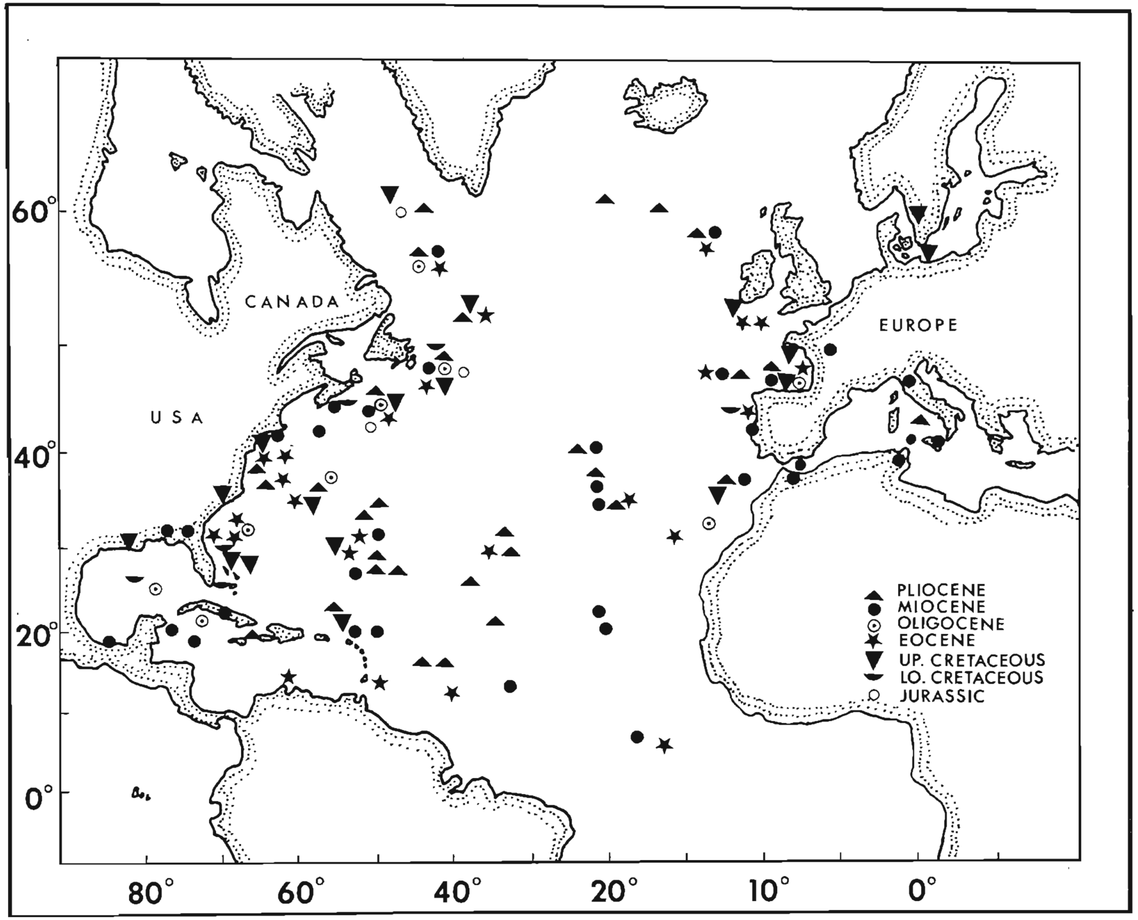

Figure 2. Distribution of Mesozoic and Cenozoic sediments in the North Atlantic. Faunal evidence from these localities was used in constructing oceanic circulation patterns since the Permian.

utilized in interpreting the characteristics of ancient watermasses. Nutrients, temperature, salinity and depth are the principal controlling factors in limiting their distribution. Furthermore, planktonic/benthonic ratios, calcareous/arenaceous ratios, coiling directions, pore area to surface area, the ultramicrostructure of test walls, oxygen isotopes and paleomagnetic data have been utilized for this synthesis.

The distribution of foraminifera in Mesozoic to Recent sediments in the North Atlantic region (Fig. 2) has been studied in detail for this analysis by the author. Comparable faunas have been recorded in widely separated parts of the world by Bandy (1960, 1963, 1966); Banner and Blow (1965); Bartlett (1967, 1968, 1969, 1971); Berggren (1960, 1962, 1964, 1968); Bolli (1957a, b, c, 1959, 1964, 1966); Bronnimann and Brown (1953, 1956, 1958); Cann and Funnell (1967); Ciffelli (1965); Cushman (1936, 1939); Ericson et al. (1961, 1963); Funnell (1964, 1971); Hamilton (1953, 1956); Heezen and Sheridan (1966); Herman (1963); Jenkins (1964); Kennett (1966, 1968, 1970); Loeblich and Tappan (1957, 1961); McTavish (1966); Menard and Hamilton (1964); Nakkady and Osman (1954); Olausson (1961); Olsson (1960, 1964); Parker (1964, 1965a, b, 1967); Reid (1962); Riedel and Funnell (1964); Smith (1955); Stainforth (1948); Stephenson (1936); Subbotina (1953); and Todd and Low (1964) to name only a few.

Depositional Environments

Sedimentological, stratigraphic and structural interpretations of the area have appeared in numerous scientific journals, e.g. Bartlett and Smith (1971). Consequently, the diverse nature of the sediments, biostratigraphy and tectonic framework of the Canadian Atlantic continental margin is quite well known. The faunas and sediments indicate that a wide variety of depositional environments existed on the Atlantic continental margin during the Mesozoic and Cenozoic periods. These environments would be constantly under the influence of the surface waters existing at that time and, consequently, should provide information on watermass distribution and current configuration.

Many sequences began with the subaerial, mainly dune environments of the quartz and quartz/limestone arenites at the bases of the Neocomian, Upper Cretaceous, and Miocene. Also subaerial were the very low-lying land areas of stream channels, flood-plains, and swamps of the Neocomian and portions of the mid-Eocene. These environments dominated the Middle Cretaceous as well, when a succession like that of the classic cyclothems was deposited. In terms of sediment volume, the most important marine environments were the shallow, warm, nearly flat bottoms of estuaries, lagoons and the open shelf, which received abundant, fine, terrigenous sediment during the upper portion of the Late Cretaceous, the Paleocene and earliest Eocene, the mid-Eocene and portions of the Middle and Late Miocene. Offshore sand banks were scattered in Middle Miocene time. An open shelf slightly deeper than present existed during the first appearance of cool subarctic waters in the Late Miocene. Lesser dispersals of terrigenous sediment allowed the accumulation from warm marine waters of the argillaceous, fine limestones and marls of the early Late Cretaceous, Middle Cretaceous, Mid-Eocene, and Mid-Miocene. Cessation of this sediment influx, along with a very slight restriction of the waters, allowed for the deposition of the massive limestones of the Upper Jurassic.

Tectonism

The foraminifera and associated sediments indicate that tectonism on the Atlantic continental margin was generally one of low level activity. Salt dome intrusions and diapiric structures periodically alter this pattern, but do not change the regional expression.

The primary rifts that initiated the formation of the Atlantic basin were probably comparable to those existing in the Red Sea at present. The limited size of this proto-Atlantic basin and the associated heat emanations dictate that the oceanic or sea waters of the Permo-Triassic Atlantic were in the form of concentrated brine. This would be an extremely favourable period for the deposition of evaporite sequences. The vertical movements and accompanying transgressions and regressions on the continental margin must have been coincident with periods of quiescence and both lateral and vertical movement in the ocean basins. Each of these movements should be reflected in changes in ocean circulation, watermass distribution and faunal associations. This is substantiated by the fact that the margins were exposed to denudation and nonpreservation of sediments at least eight times during their development.

The reconstructions and positions of the continents represented by Figures 3 to 7, are derived from information by Hess (1962); Bullard et al. (1965); Bullard (1969); Vine (1966); Heirtzler and Hays (1967); Funnell and Smith (1968); Emery et al. (1969); Kay (1969); Berggren and Phillips (1970); Dewey and Horsfield (1970); Le Pichon (1968); Wilson (1966, 1968); and Mitchell and Reading (1969). The fit of the continents around which I have placed the surface circulation patterns is essentially that of Bullard et al. (1965), at the 1,000-metre bathymetric contour. There are already some objections to this fit, especially in the Bahama Bank-Caribbean Sea. However, it provides an excellent framework on which to transpose the Mesozoic-Cenozoic ocean circulation patterns proposed here.

The initial Atlantic was probably sluggish and affected by monsoons, comparable to, or weaker than, those of the present Indian Ocean. Similarly, the present water loss of $6X10^6 m^3/sec.$ (Neumann and Pierson, 1966; Sverdrup et al., 1942), from the South to the North Atlantic and the piling up of north equatorial waters in the Gulf of Mexico as the mechanism for initiating the Gulf Stream, is extremely important in recognizing the displacement of the thermal equator and various watermasses and their circulation patterns in the past.

Permo-Triassic Period

The pre-drift construction, (Fig. 3) is essentially that of Bullard et al. (1965) and indicates that the Atlantic was probably closed along the crest of the present mid-Atlantic Ridge and the initial configuration was probably comparable to the present Red Sea. Consequently, this period was extremely favourable for the formation of evaporite facies. Sluggish surface and bottom currents, high evaporation rates, high temperature, and highly saline dense waters are envisaged and the author believes that the climate was necessarily extremely warm and arid. Planktonic microorganisms would be extremely sparse. It is also assumed that there was a uniform benthonic fauna in the proto-Atlantic which was not differentiated into distinctive faunal provinces until Late Eocene-Early Miocene time. This provincialism has continued to

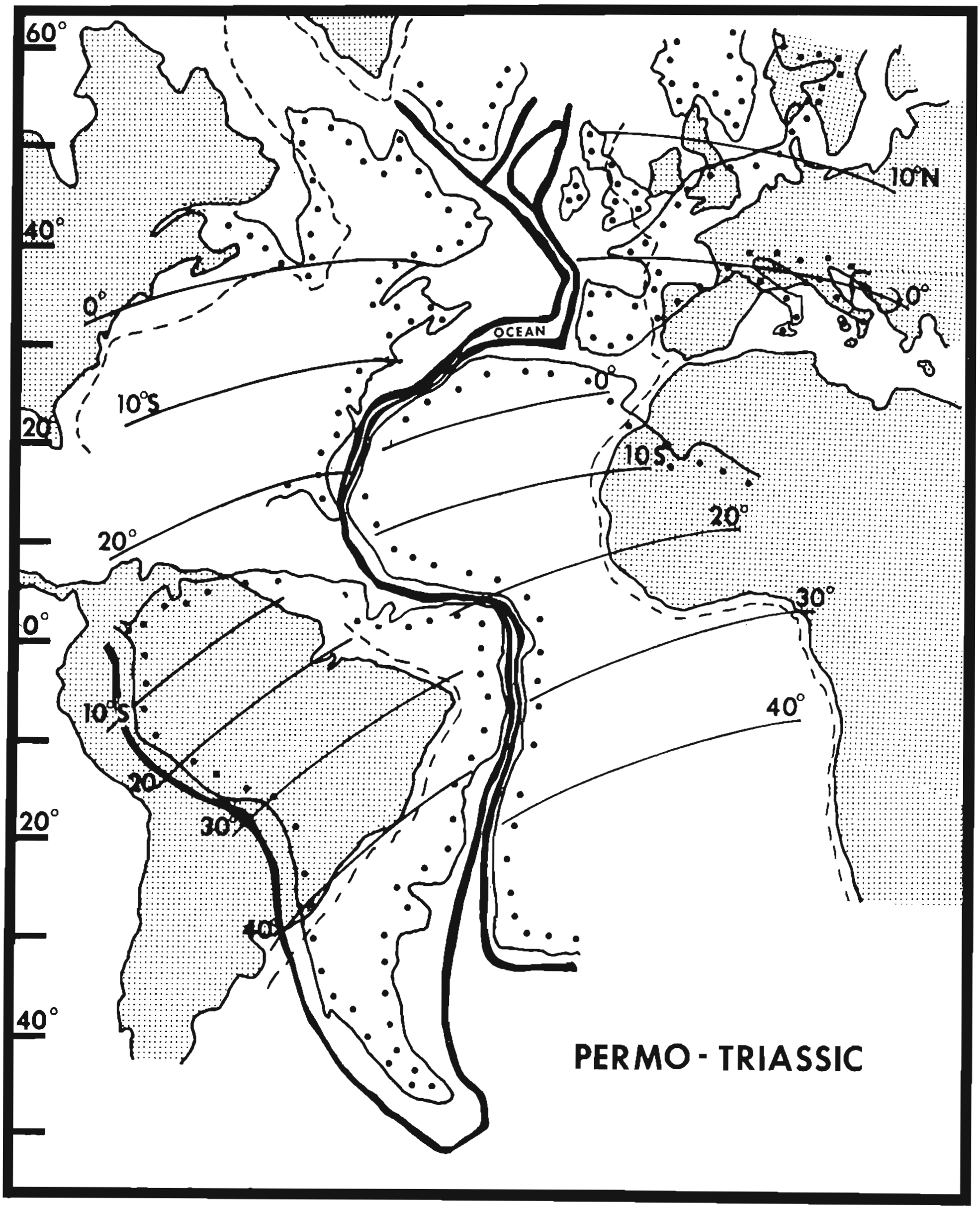

Figure 3. Pre-drift construction modified after Bullard et al. (1965) showing the proto-Atlantic. Different equatorial positions are superimposed because of continental "best fits" proposed by various authors.

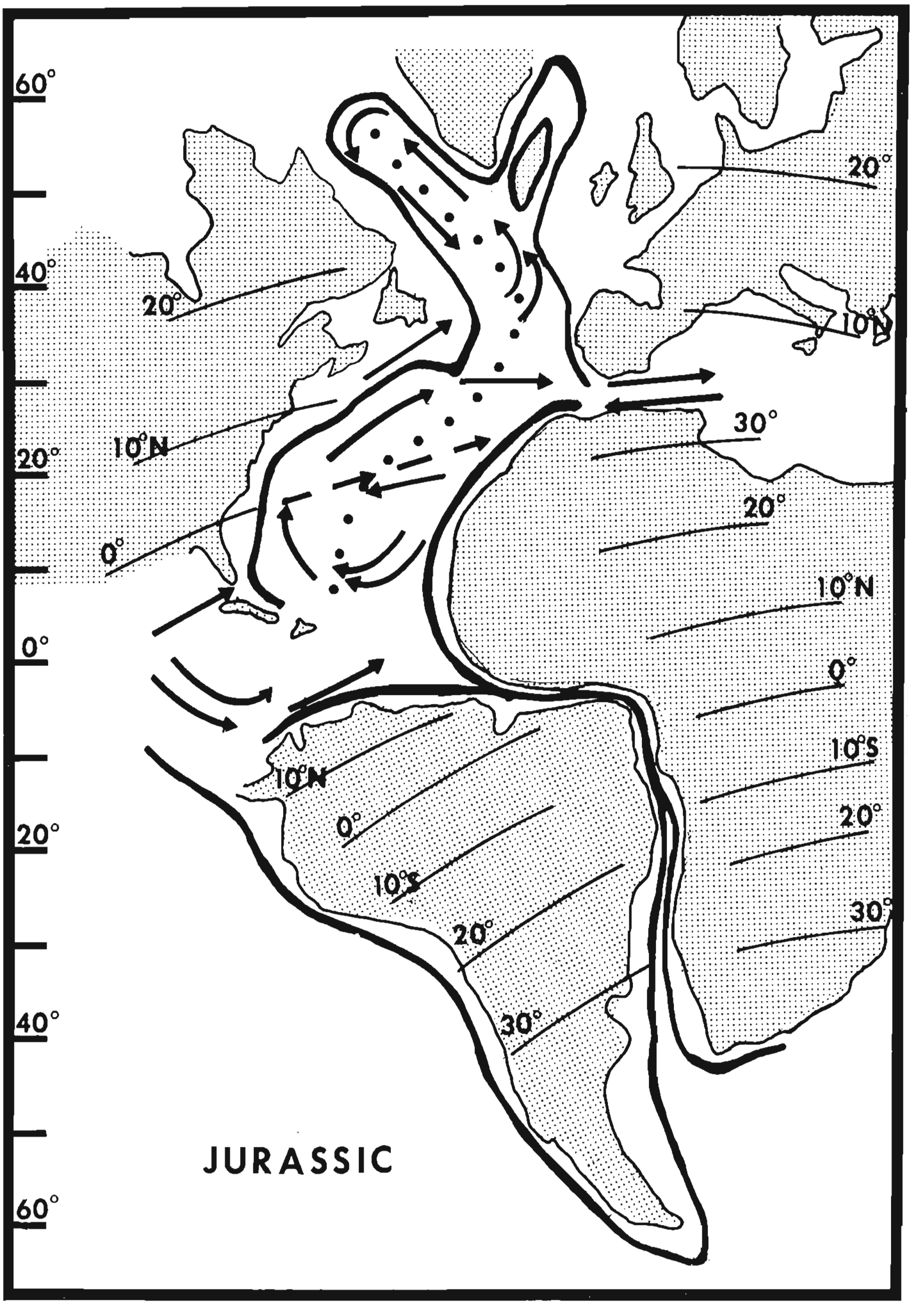

Figure 4. The Atlantic ocean during Jurassic time, showing a North Atlantic with connections to the Mediterranean (Tethys) and Pacific. The narrow, land-enclosed, sinuous Atlantic probably resembled a Mediterranean Sea with sluggish currents resulting from density differentiation through evaporation.

the present time because of geographic isolation and distinct variations, in temperature, salinity, density and related physical and chemical parameters and, consequently, these watermass characteristics provide effective barriers to faunal migration.

Jurassic Period

The central Atlantic (Fig. 4) was opening at approximately 0.8 cm/yr (Funnell and Smith, 1968) and could have opened as much as 1,120 kilometres during the Jurassic Period. There is no concrete faunal evidence on the Atlantic continental margin to imply that the North Atlantic-Greenland-Europe area was open. However, Berggren and Phillips (1970) suggest that the Labrador Sea reached its full width 130 m.y. ago. Jurassic limestones with shallow water, inner-shelf benthonic faunas have been recovered (1969) by the present writer from the Labrador Sea. Apparently, the South Atlantic remained closed during this period, although there may have been circulation between the Pacific and Tethys (Mediterranean) through the opening between North and South America.

There were doubtlessly at least two factors that prevented the formation of dense cold bottom waters and permitted the Atlantic to have the characteristics of a Mediterranean Sea during the Jurassic period.

1. The thermal equator (Fig. 4) appears to have been significantly north of its present position, or the continents were south of their present position, and one would expect subtropical waters to extend at least as far north as Labrador. Surface circulation would be wind-induced, whereas deep-water circulation would be dependent on density differentiation brought about through evaporation.
2. The limited, if any, access to cold, deep bottom water from either polar region would prevent thermohaline circulation, the main driving force of the present oceans.

It is concluded from floral, faunal and sedimentological evidence that the climate was extremely warm and humid. Faunal and floral evidence for this period is extremely sparse in sediments accumulated on the Atlantic continental margin. The planktonic foraminifera are characterized by primitive globigerinids whereas the benthonic assemblage is typical of inner shelves. Both faunas are cosmopolitan in their distribution and have been reported on both sides of the Atlantic. The thick limestones of Jurrasic age attest to environmental conditions comparable to those of the present day low-energy Caribbean areas. There is no evidence for the existence of extensive organic reefs. However, the presence of high concentrations of carbonaceous material in some sediments substantiates proximity to heavily forested land areas or widespread swamps.

Cretaceous Period

The Central America-Africa area (Fig. 5) had opened another 920 kilometres and attained 73 per cent of its present width. Also, it is assumed that some spreading was probably occurring between the Grand Banks and Spain. Computer plots (Berggren, Woods Hole Oceanographic Institution, personal communication) suggest that a warm current flowed into the Labrador Sea and along the west coast of Greenland by mid-Cretaceous time. However,

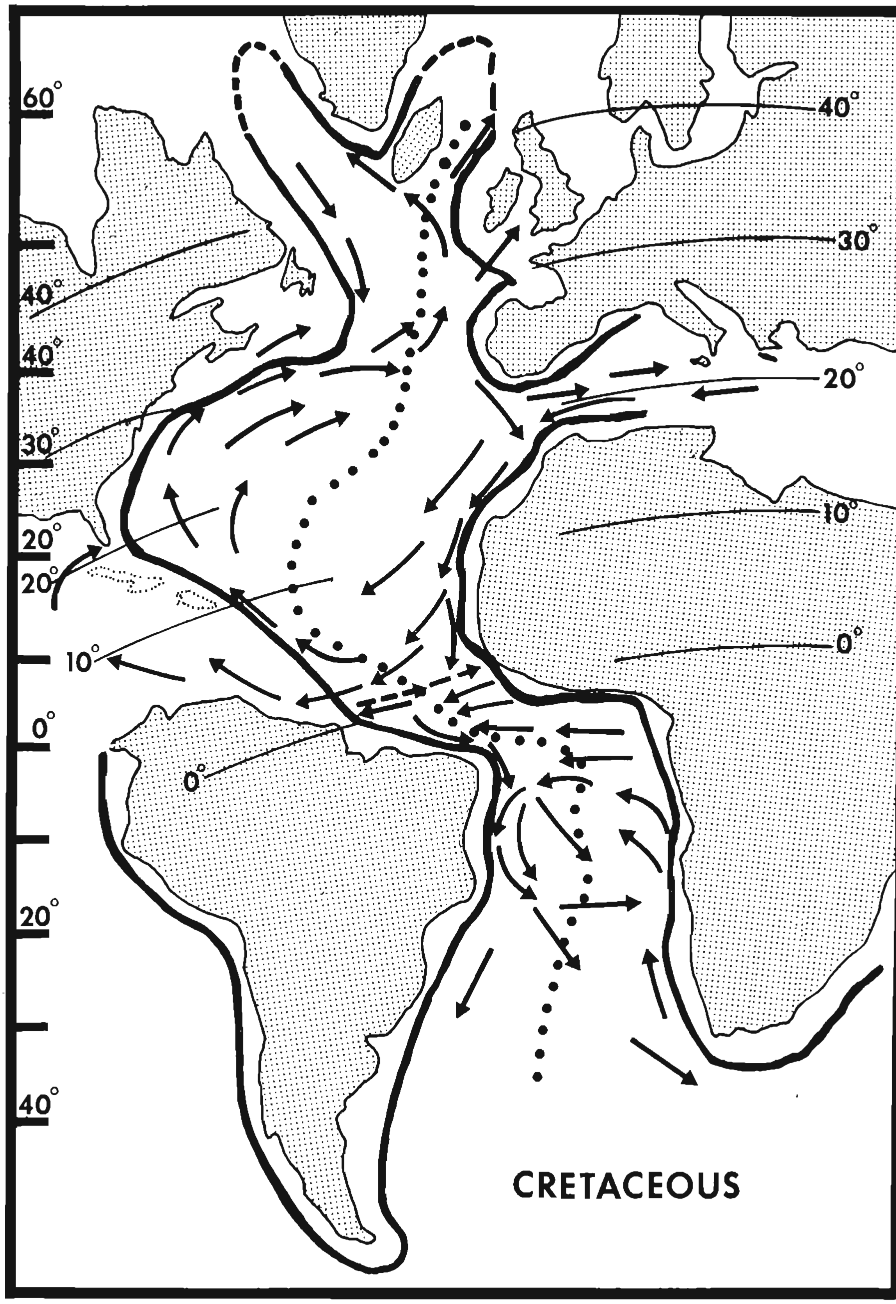

Figure 5. The Atlantic Ocean during Cretaceous time, showing an enlarged North Atlantic and an opening South Atlantic. The acutely sinuous shape near the equator would restrict free exchange between watermasses in this area. Convergence and divergence should occur significantly north of the equator. The absence of cold deep bottom waters and the close proximity of landmasses would cause drag and eddy diffusion. Rotation of the earth and Coriolis force should maintain sluggish circulation without the distinct thermal characteristics prevalent today.

concrete faunal evidence for the presence and penetration of distinctly boreal or subarctic water into the North Atlantic is lacking.

During the Lower to Mid-Cretaceous the thermal equator was somewhat to the south of its present position. Generally, the microfaunal evidence suggests that waters of the North Atlantic were slightly colder and/or deeper than those in the Upper Jurassic and Upper Cretaceous periods.

Open circulation between the Caribbean, western North Atlantic and Mediterranean permitted the dispersal and development of comparable planktonic microfaunas. Benthonic assemblages in these widely separated geographic areas show a high degree of similarity. Planktonic species with more pore space than surface area, and well-developed keels imply the presence of warm waters (at least 17°C) during the late Middle and Upper Cretaceous. Similarly, paleontologic evidence in both the Middle and Upper Cretaceous with species of Rotalipora and Globotruncana supports the contention that warm waters existed in the mid to high-latitudes of the North Atlantic. Paucity of planktonic faunas at some intervals attests to sediment influx and the generally brackish nature or unfavourability of the continental margin environment. The similarity in the benthonic faunas is a reflection of the lack of variation in the physical properties of the bottom waters of the Atlantic basin, the close proximity of South America and Africa and the land areas of the extreme North Atlantic (Fig. 5). This substantiates the author's contention that the Arctic polar seas had not yet invaded the North Atlantic.

During the Neocomian, the preserved record indicates that subaerial and beach deposits covered the eroded surface of Jurassic limestones. Also, estuaries and stream channels were periodically flooded with warm marine waters. Coal cyclothems and intermittent marine deposition are most typical of the Middle Cretaceous period. Quartz arenites, dune deposits and soft white chalk are most characteristic of the Upper Cretaceous deposits. Biostromal limestones represent accumulations in environments not unlike that of the "Oyster Reefs" of the present Gulf of Mexico (Bartlett, 1968; Bartlett and Smith, 1971).

From the evidence of the microfauna the following history of the Atlantic area can be deduced for Upper Cretaceous time. During the Upper Cretaceous planktonic foraminifers with subtropic characteristics, strong Tethyan affinities and world-wide distribution, indicating widespread uniform surface currents, typify the Atlantic continental margin sediments.

Lower Tertiary Period

By Lower Tertiary time, the Central Atlantic (Fig. 6) had opened to 85 per cent of its present width and the North Atlantic had opened to 96 per cent of its present width. Also, the South Atlantic had opened to 50 per cent of its present width, permitting limited but restricted circulation between the two watermasses. The initiation of spreading on the Reykjanes Ridge probably opened the Atlantic to the Arctic seas (Phillips in Berggren and Phillips, 1970). This may represent the first major incursion of cold Arctic surface water into the Atlantic Ocean basin. However, faunal evidence on the Atlantic continental margin, Caribbean, and the Atlantic basin (Fig. 2) indicates that this event is more likely to have occurred in the Miocene. On the other hand, it is believed by Berggren and Phillips (1970) that such circulation may have initiated the depositional environments for Eocene cherts which formed so extensively in the North Atlantic and other oceanic watermasses. However,

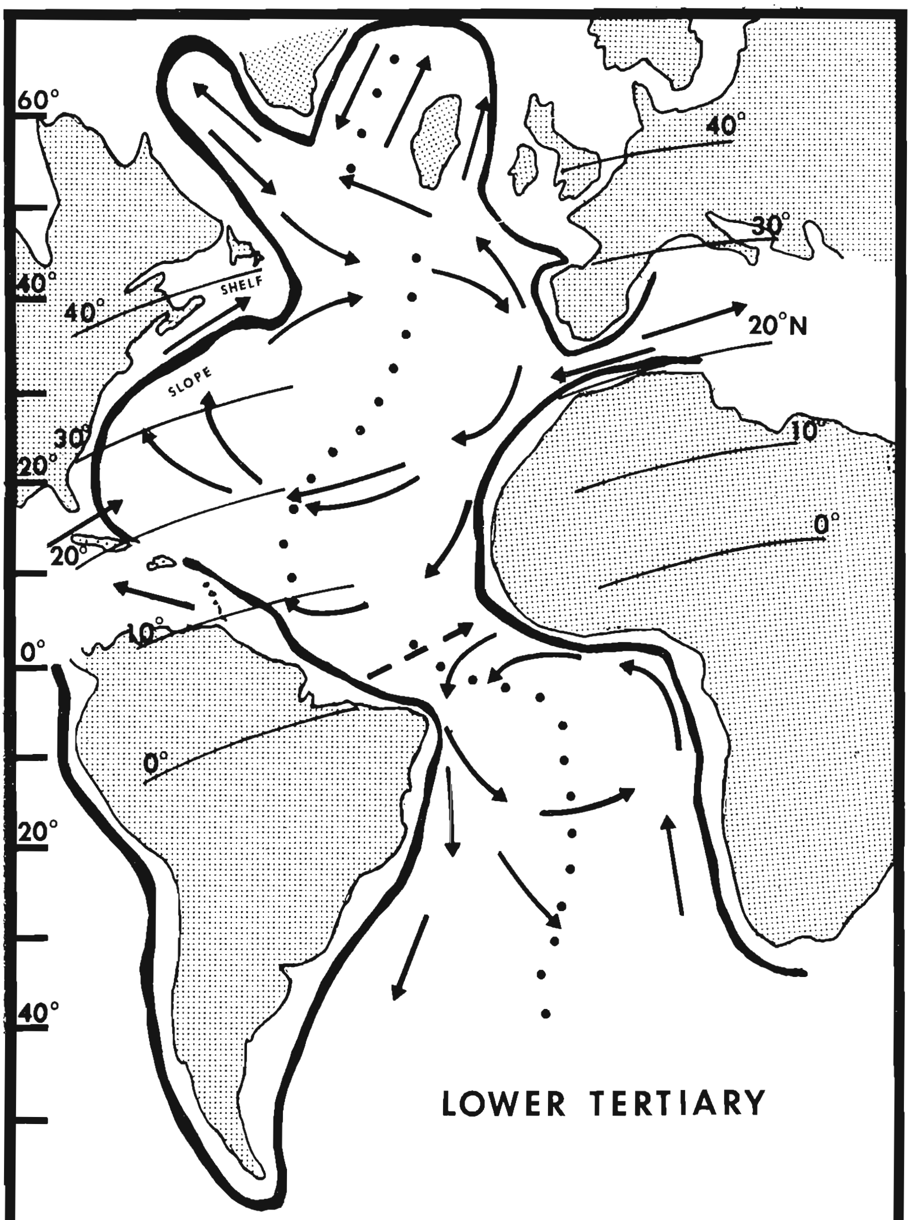

Figure 6. Lower Tertiary showing the Mediterranean (Tethys) - Pacific connection, a possible Arctic surface connection and an expanding South Atlantic. A gyre may have existed at 40° N. Lat. Similarly, the northward-flowing warm water would be the result of deflections from the Tethys-Pacific Current. The distribution of landmass and displacement of the thermal equator would be significant in the equatorial region.

Gibson and Towe (1971) equate these Eocene cherts with volcanism in the North Atlantic region. They believe that a volcanic origin provides a consistent explanation for the presence of siliceous deposits in a wide variety of marine environments. This view is more consistent with the present author's proposed oceanic circulation patterns.

On the Atlantic continental margin the Lower Tertiary period was characterized by the presence of warm shallow seas in which lime-muds alternated with land-derived sediments and highly siliceous biogenous deposits. The relative positions of the North American and European plates indicate that the equator could be approximately 15 to 20 degrees south of its present position. A general cooling in the North Atlantic and/or the influx of polar surface waters may also have lowered the average sea temperatures in the northern hemisphere a few degrees. It is significant to note that Globorotalia, a genus that is today more characteristic of tropical and subtropical waters, is minimal in Lower Tertiary sediments on the Canadian Atlantic continental margin, but is widespread in Caribbean and European sections.

During the Paleocene-Lowest Eocene (Fig. 5) the environment was typically low-lying and coastal-nearshore and bathed by waters deflected from the Tethyan-Pacific exchange. Sediment deposition in the Middle Eocene occurred in warm, shallow marine environments with slight current activity seaward of any deltaic influence; consequently, sudden influxes of terrigenous material and the presence of highly siliceous material were intermittent.

The published literature (Bartlett and Smith, 1971; Berggren and Phillips, 1970; LeRoy, 1953; Cushman, 1951; and Cushman and Renz, 1946, 1947) shows a close similarity between Paleocene benthonic faunas in North Africa, the Gulf Coast and Caribbean areas. Lower to Middle Eocene faunas have a lesser degree of similarity with Upper Eocene assemblages indicating a further degree of provincialism.

During the Oligocene the deposits in the western North Atlantic were consistently deep water and were characterized by abundant radiolaria and planktonic foraminifera. Similarly, Bandy (1970) reported an abyssal sequence in the Middle Oligocene-Lower Miocene in Eastern Panama and Northern Colombia. According to Berggren (in Berggren and Phillips, 1970) planktonic microfaunal assemblages indicate that there was a significant cooling trend in the latest Eocene-Early Oligocene, but this may equally be attributed to faunal differences because of facies change. Because of inadequate faunal data the history of this epoch is obscure. However, lateral spreading (magnetic anomaly 5 to magnetic anomaly 13), during this period was extremely slow (Dewey and Horsfield, 1970) and could account for significant vertical movement in both the ocean basins and continental margins. This would imply increasing vertical movement with decreased lateral spreading.

Benthonic faunas of Oligocene-Miocene age show a marked tendency towards provincialism, a further indication of geographic and/or environmental isolation. Planktonic faunas also show, for the first time, the influence of well-defined vertical watermass zonation and distinct surface current patterns in the North Atlantic.

Upper Tertiary Period

The Central and North Atlantic (Fig. 7) had opened to 96 per cent of its present width (Berggren and Phillips, 1970) by Miocene time; spreading between the Grand Banks and Spain had also ceased. The influx of cold water

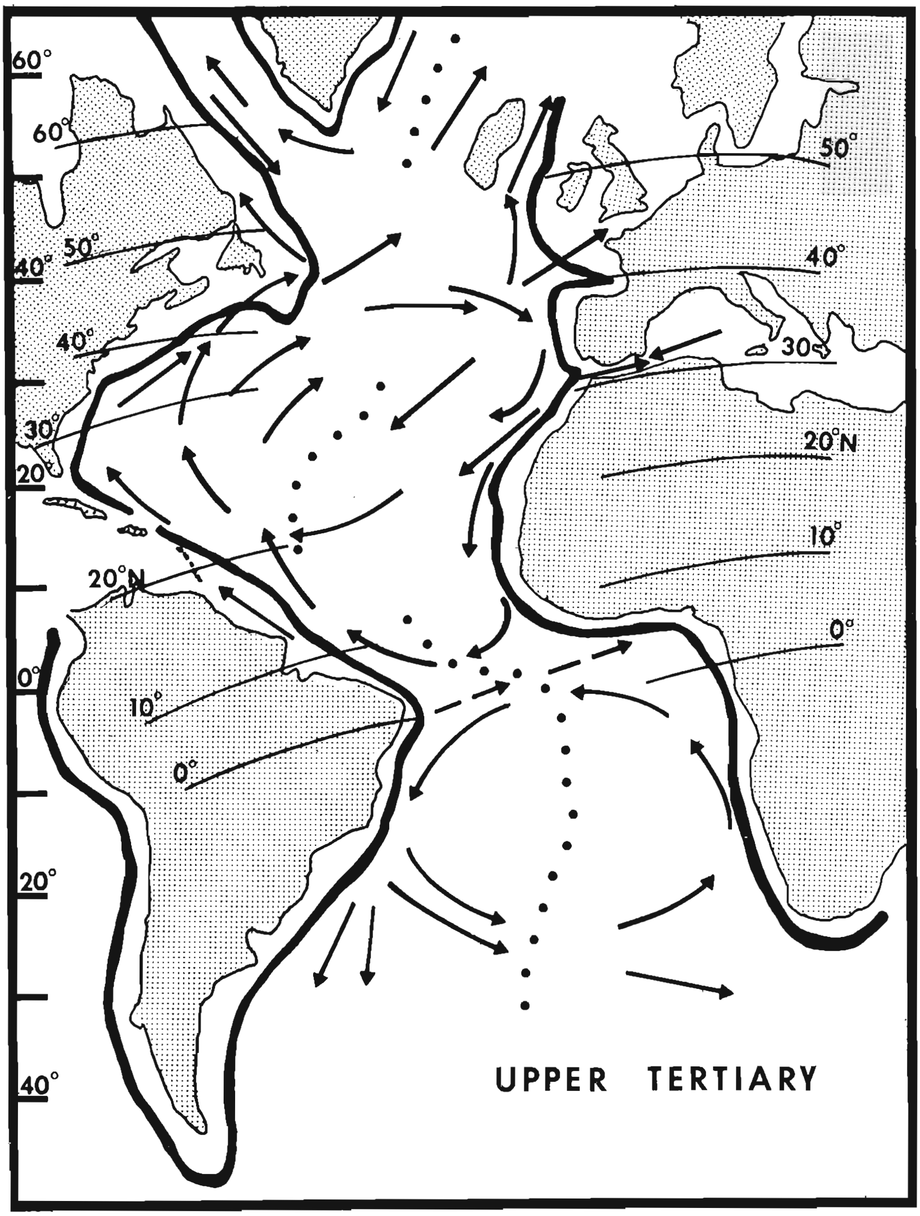

Figure 7. The Atlantic Ocean during the Upper Tertiary, showing almost unrestricted circulation with the Arctic Sea. Gibraltar and Panama are assumed to be closed by Mid-Miocene time. The distribution of landmass, closure of Panama, opening of Gibraltar, Arctic watermass incursion, and the initiation of the thermohaline circulation would displace the thermal equator to its present position and thus initiate present-day oceanic circulation patterns.

species resembling Globigerina pachyderma and G. bulloides immediately above the warm water Globorotalia fohsi fohsi faunas (mid-Miocene) may indicate a major unconformity, but also suggests a definite cooling trend that persisted into the Pleistocene. Ocean circulation patterns as we know them today were actually being initiated. As mentioned earlier, polar surface waters may have entered the Atlantic basin in the Eocene, but truly Arctic or boreal planktonic species did not appear on the Atlantic continental margin until mid-Miocene time with the incursion of Arctic deep and bottom water indicating a definite opening to the Arctic sea. Reports from Leg 12 of the GLOMAR CHALLENGER (Geotimes, v. 15, no. 9, p. 10-14) suggest that the first significant cooling preceding the Pleistocene did not occur until approximately 5 m.y. ago.

Deep water circulation between the Atlantic and Mediterranean probably also ceased during the Mid-Miocene. The information obtained on Leg 13 of the GLOMAR CHALLENGER (Geotimes, v. 15, no. 10, p. 12-15) was interpreted as indicating compression between Africa and Europe, and that the deep basin of the Mediterranean was completely isolated from the Atlantic during the Late Miocene. The N8/N9 Orbulina datum has not been recorded on the Atlantic continental margin. It is assumed that the influence of subtropic waters of the Gulf Stream declined after major development during the time represented by the N3/N4 Globigerinoides datum near the base of the Aquitanian.

It has been suggested by El-Naggar (1967) that provincialism, as far as certain planktonic assemblages are concerned, for example Globigerina compressa/Globigerina daubjergensis, may have occurred in Europe during the Paleocene.

It was also noted by Berggren and Phillips (1970) that noticeable changes i.e. dissimilarities occurred in the benthonic foraminiferal faunas throughout the Mediterranean-Caribbean areas about Mid-Miocene time. This further substantiates the contention of distinct watermass development and geographic isolation of these areas.

At present large volumes of North Atlantic Drift water flow into, and lesser volumes leave the Mediterranean daily. Consequently, an effective blocking of this water during the Miocene would enable large volumes of Atlantic "intermediate water" to flow southward and permit larger volumes of Gulf Stream water with its subtropic species to penetrate farther north. Moreover more rapid cooling of these warm saline waters because of the development of an Arctic and perhaps an Antarctic connection, would permit larger volumes of water to sink and move southward, and would therefore, effectively increase the rate of circulation in the North Atlantic. This process would have at least a two-fold effect; 1) first, a general overall warming of the oceanic surface waters, and 2) later, a general cooling throughout the entire watermass. The dense water from the Mediterranean that contributes so significantly to the Atlantic deep and bottom water today would have to be derived from another source, presumably the Arctic deep and bottom water and/or the Atlantic surface water. Seasonal variations would be important. This mechanism may have been instrumental in the development of conditions that resulted in the initiation of Pleistocene glaciation.

If cold dense currents did not flow southward from the Arctic then circulation throughout the world oceans (the Atlantic in particular) would remain sluggish. The Gulf Stream, which owes its origin to the piling up of North Equatorial water in the Gulf of Mexico, would be negligible until the development of effective barriers (Caribbean Islands) or complete closure of the area between North and South America (Panama). Again, the author does not believe the Gulf Stream operated as a major force on the Atlantic continental margin until Miocene time (Bartlett, 1968). There is obviously a relationship between the gradual opening of the North Atlantic basin and the influx of cold polar bottom waters and their subsequent effect on all ocean watermasses, currents, climates and faunal assemblages.

It has been suggested by Bartlett and Smith (1971) that during the Middle and Upper Miocene, a basal quartz arenite was deposited on the eroded surface of the top of the Eocene. Deposition of Mid-Miocene sands occurred in aeolian environments, whereas the clayey silt deposits were typical of foreset and bottomset deltaic beds. Marine sands and silts are also evident; these muddy sediments represent short-term overloading because of tectonic oscillations in the source area. Miocene seas were presumably shallow and warm along the entire continental margin. Tectonically, large volumes of clastics were being contributed from a western source area. This isostatic rebound may be the result of adjustment following the consistent sinking of the continental margin during the Oligocene.

10 Million Years Ago - Present

According to Berggren and Phillips (1970) the Atlantic-Mediterranean opening was reactivated approximately 5 m.y. ago.

Information from the GLOMAR CHALLENGER Leg 12 cruise places the onset of glaciation in the mid-Pliocene (5 m.y. ago). Equally significant for paleogeographic and paleoclimatic reconstructions of the Late Cenozoic history of the North Atlantic was the first appearance of Arctic planktonic species on the Atlantic continental margin in the Mid-Miocene. This provides the first concrete evidence of the incursion of persistent Arctic watermasses, the opening of the Atlantic basin to the north and the presence of a distinct Gulf Stream in the North Atlantic.

The Mid-Miocene occurrence suggests that cold Arctic waters were in contact with warm Gulf Stream waters at least 8 m.y. to 15 m.y. ago and that an effective barrier or connection between North and South America brought about the formation of Gulf Stream waters before Pliocene time (6 m.y. - 8 m.y. ago). A polar cooling and the movement of warm waters significantly north of their present position would provide the increased precipitation essential for the formation of continental ice sheets.

During the Lower Tertiary and early part of the Upper Tertiary (up to the boundary between the Lower and Mid-Miocene, Burdigalian-Langhian, 20 m.y. ago), benthonic foraminiferal faunas of the Mediterranean region exhibit a strong degree of similarity with those in the Gulf Coast region, the Caribbean-Antilles region, Central America and Northern South America (Colombia, Ecuador, Venezuela) and the Atlantic continental margin off eastern Canada. By Middle to Upper Miocene time, Langhian-Tortonian, 10 m.y. ago, distinct differences occurred in the benthonic microfaunas in these widely separated geographic localities.

Benthonics also became extremely provincial in character during this period. Mediterranean forms became more distinctly related to those of European areas and were in turn dissimilar to Nova Scotian and Caribbean species. Planktonic species (Plates 1-3) developed as subarctic, transitional, subtropical and tropical assemblages that very closely resemble present-day distributional patterns.

All of these faunal variations (species, ultrastructure, geographic distribution) can be attributed to distinctly different oceanic watermasses and circulation patterns because of changes in both landmass and seafloor configuration.

Since the Middle Miocene there has been a consistent gradual decline, with minor reversals, in the earth's temperature. As previously mentioned this was the time during which faunal provincialism developed, and significant changes from strictly surface circulation to increased thermohaline circulation occurred in the North Atlantic.

Pliocene sediments containing planktonic foraminifera have not been definitely identified on the Atlantic continental margin. Sediment cores on the Tail of the Banks and the Central Atlantic contain good Pliocene planktonic assemblages (Bartlett *et al.*, 1968). The early Pliocene is marked by significant warming, followed by a gradual cooling at approximately 3 m.y. to 2 m.y. ago at the onset of continental glaciation.

All of the Pleistocene sediments on the Atlantic continental margin are characterized by brackish water benthonic foraminifera and the Arctic planktonic species *Globigerina pachyderma* and *Globigerina quinqueloba* and the Subarctic or Transitional species *Globigerina bulloides* intermixed with Subtropic species such as *Globigerinoides ruber* (pink). The Pleistocene was a period of rapid change on the Atlantic continental margin. The exposure of banks and shelves and the restricting or opening of straits in the western Atlantic again greatly altered circulation throughout the North Atlantic during the Pleistocene. Labrador Current waters were diverted far to the east (Mid-Atlantic Ridge) and the Gulf Stream consistently flowed into the Bay of Fundy and Gulf of St. Lawrence.

SUMMARY

Microorganisms, foraminifera in particular, are excellent biostratigraphic and environmental indicators. A thorough understanding of the paleontology and biostratigraphy of an area is essential for a proper interpretation of plate tectonics. Similarly a knowledge of the relative position of the continents through time is a prerequisite for paleogeographic, paleoecologic and paleooceanographic reconstructions. The stratigraphic, taxonomic and ecologic data from microfaunal investigations have been utilized to interpret both the oceanic and climatic conditions that existed in the North Atlantic during deposition of the Mesozoic and Cenozoic sediments on the Atlantic continental margin. A thorough knowledge of the relative position of the landmasses adjoining the Atlantic Basin is also essential for a proper interpretation of paleoecology, paleoclimatology and paleooceanography. Environmental analysis and the synthesis of formative processes also provide data essential for tectonic synthesis.

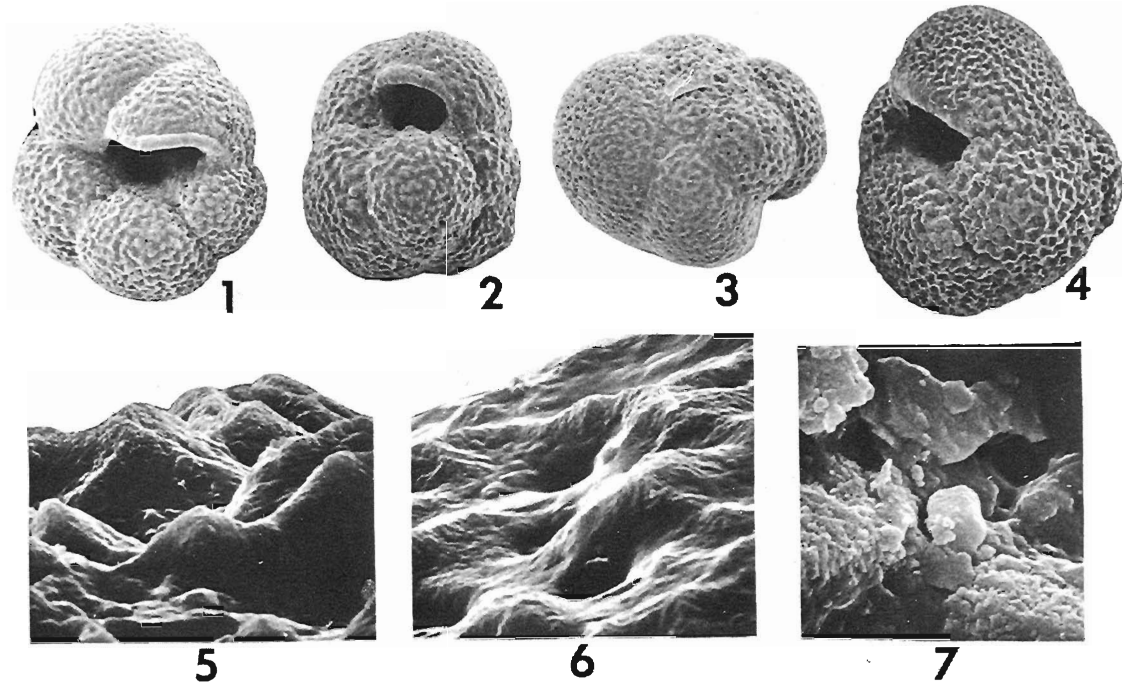

Plate 1. Arctic-Subarctic, Figs. 1-7: Globigerina pachyderma (Ehrenberg)

1, 2. Apertural view showing five chambered and four chambered forms with thick apertural rim, x250;
3. Dorsal view with five chambers visible, x250;
4. Ventral view showing thick test with prismatic calcite crystals, x250;
5. Microstructure of penultimate and ultimate chambers showing test thickening through growth of calcite rhombs, x9000;
6. Microstructure of earlier formed chamber showing pores, x9000;
7. Microstructure showing laminar layers of calcite in apertural region, x9000.

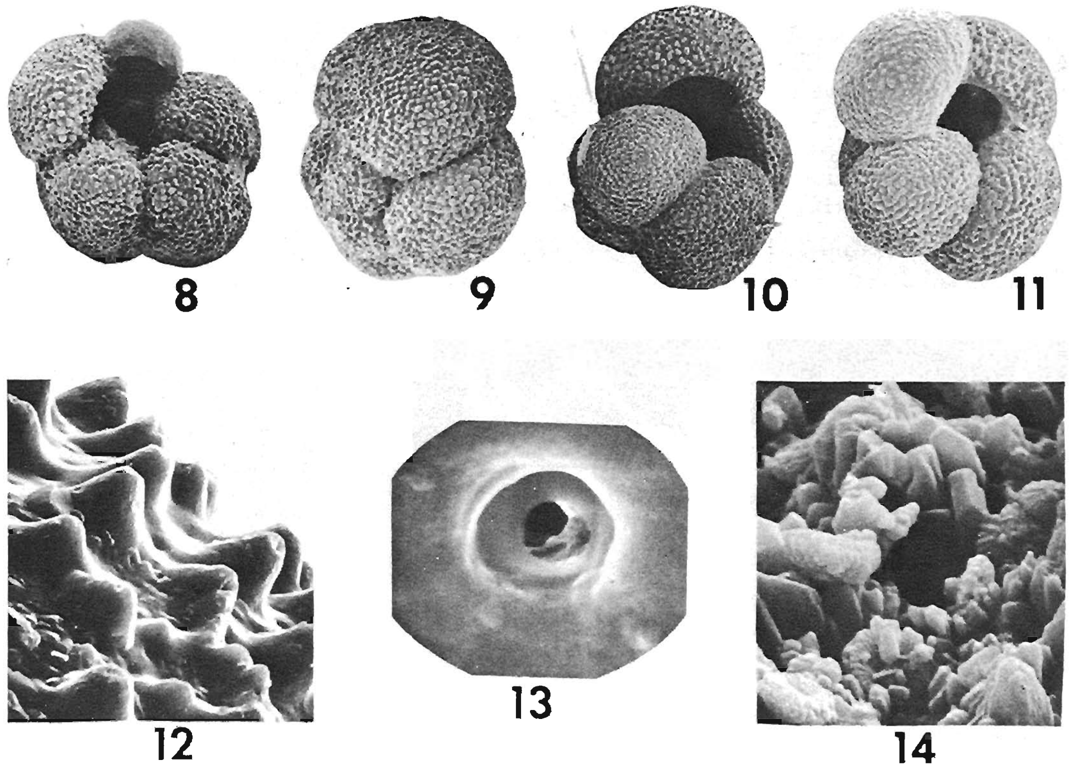

Plate 1 (continued) Figs. 8-14: <u>Globigerina</u> <u>bulloides</u> d'Orbigny

8, 10, 11. Apertural views showing large umbilical aperture and irregular nodose tests surface; note development of reduced non-ornamented accessory chamber in figure 8, x150;

9. Dorsal view showing coiling characteristics, x150;

12. Microstructure of test surface showing knobby spinal bases, x2700;

13. Detail view of individual pore, narrowing towards interior, x2700;

14. Microstructure of apertural region showing concentration of coccolith plates, x9000.

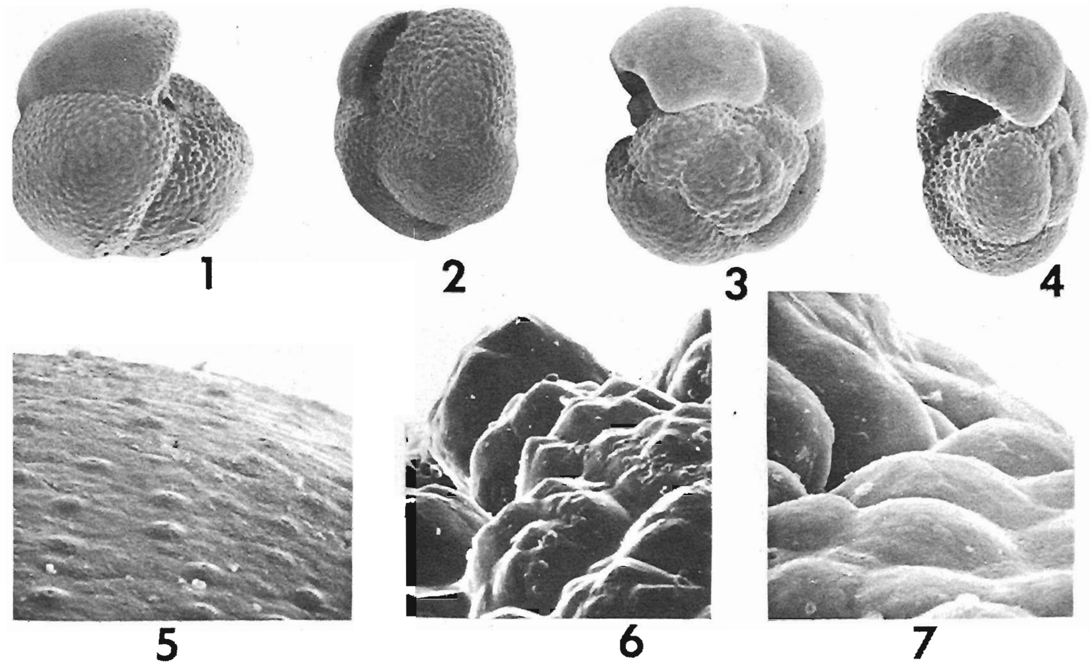

Plate 2. Transitional-Subtropic, Figs. 1-7: <u>Globorotalia</u> <u>inflata</u> (d'Orbigny)

1-4. Ventral, apertural and dorsal views of test showing inflated chambers, extraumbilical aperture, rounded periphery, smooth ultimate chamber and irregular knobby earlier chambers, x130;

5. Microstructure of ultimate chamber showing relatively smooth surface with slightly elevated mounds, each with a single pore, x2700;

6. Microstructure of apertural region, showing thick calcite deposition and development of thick nodes and spine like bases, shell microstructure suggests an affinity to <u>Globigerina</u>, x2700;

7. Microstructure of dorsal surface of test wall, note regularity of undulations, x2700.

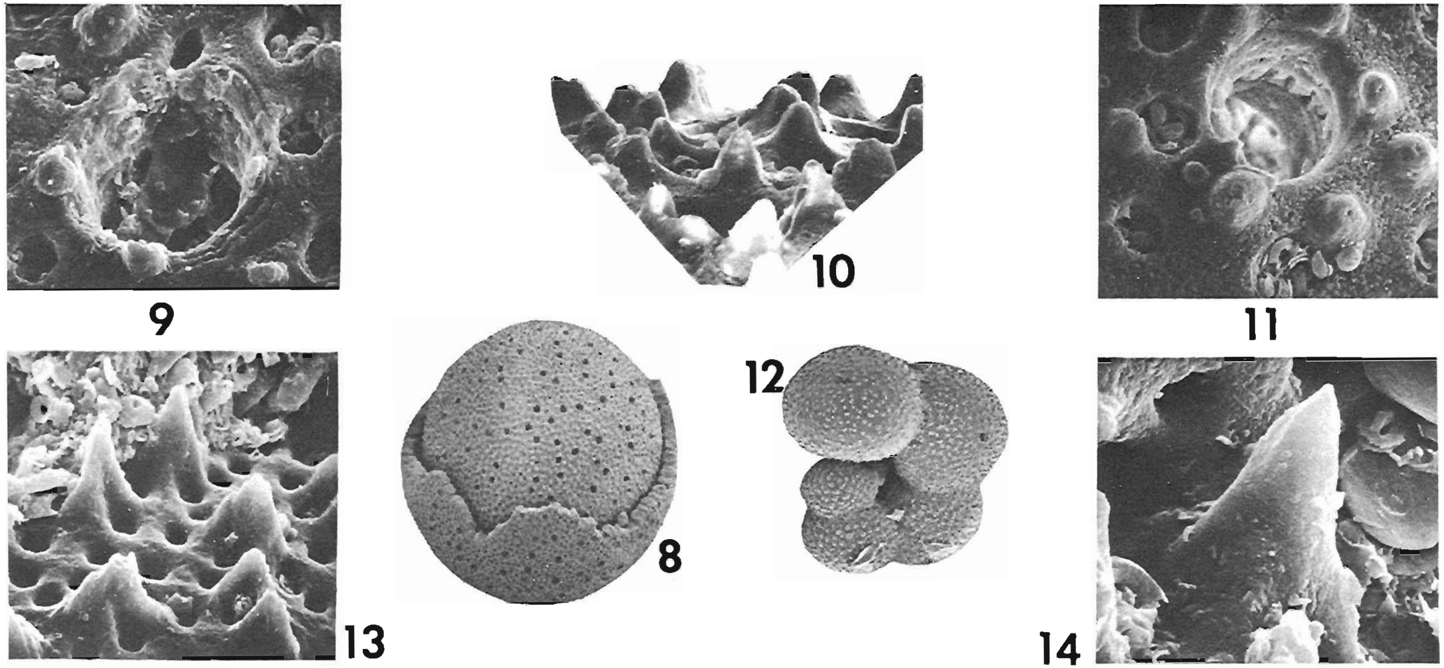

Plate 2 (continued), Figs. 8-11: Orbulina universa (d'Orbigny); Figs. 12-14: Globigerinatella aequilateralis (Brady)

8. View of test showing two distinct spherical walls, two distinct pore sizes, and knobby spinal bases, x130;
9. Microstructure of outer wall showing detail of large pore, coccolith filled smaller pores, and broken spinal bases, x2700;
10. Microstructure of outer wall emphasizing spinal bases, these have affinities to certain Globigerinoides, x2700;
11. Microstructure of inner shell wall, note similarity to outer wall, x2700;
12. View showing trochospiral test, x130;
13. Microstructure showing surface pores and spinal arrangement, note that spines partially encompass pores and that all are grooved for passage of living protoplasm, x2700;
14. Microstructure of spine, strands of protoplasm apparently pass through these as well as the major pores, x9000.

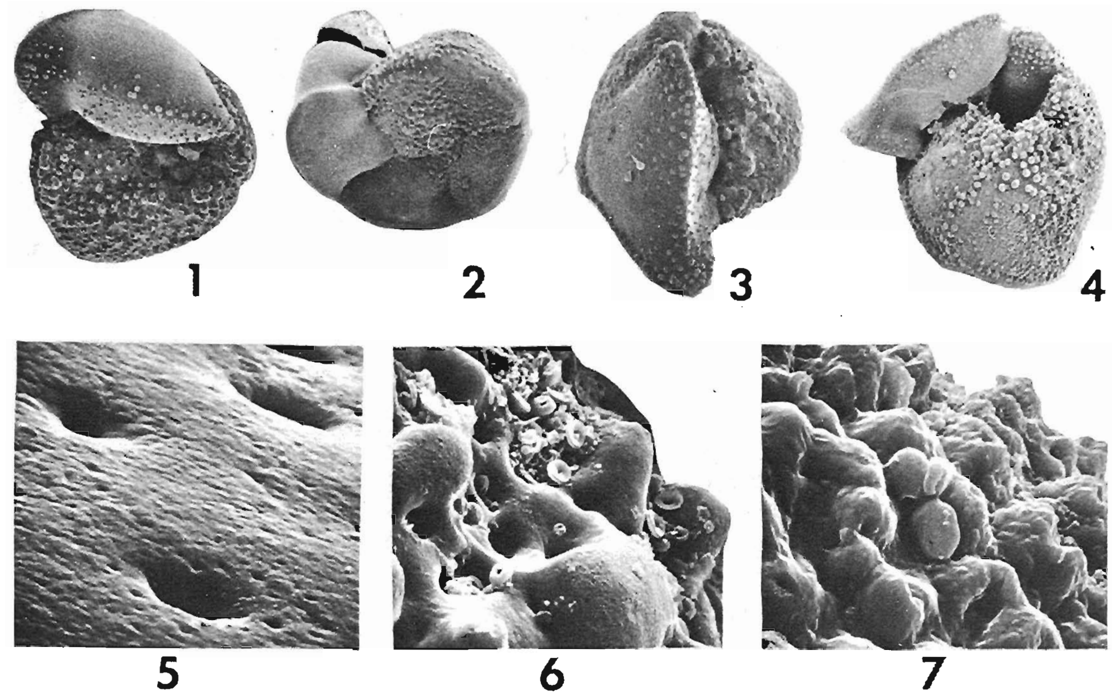

Plate 3. Subtropic-Tropic, Figs. 1-7: <u>Globorotalia</u> <u>truncatulinoides</u> (d'Orbigny)

1-4. Ventral, dorsal and apertural views showing characteristics of test morphology, distinct keel, concave dorsal surface, sinuous sutures, and irregular surface of chambers on ventral side, also characteristic open umbilicus and slit like aperture, x130;

5. Microstructure of ventral surface of ultimate chamber, note granular surface and large distinct pores surrounded by small pores as in <u>Orbulina</u>, x12,500;

6. Microstructure of early formed chambers, dorsal view, pores remain distinct, are filled with coccoliths and surrounded by secondary shell growth, x2700;

7. Ventral view of irregular secondary calcite overgrowths in umbilical region, x2700.

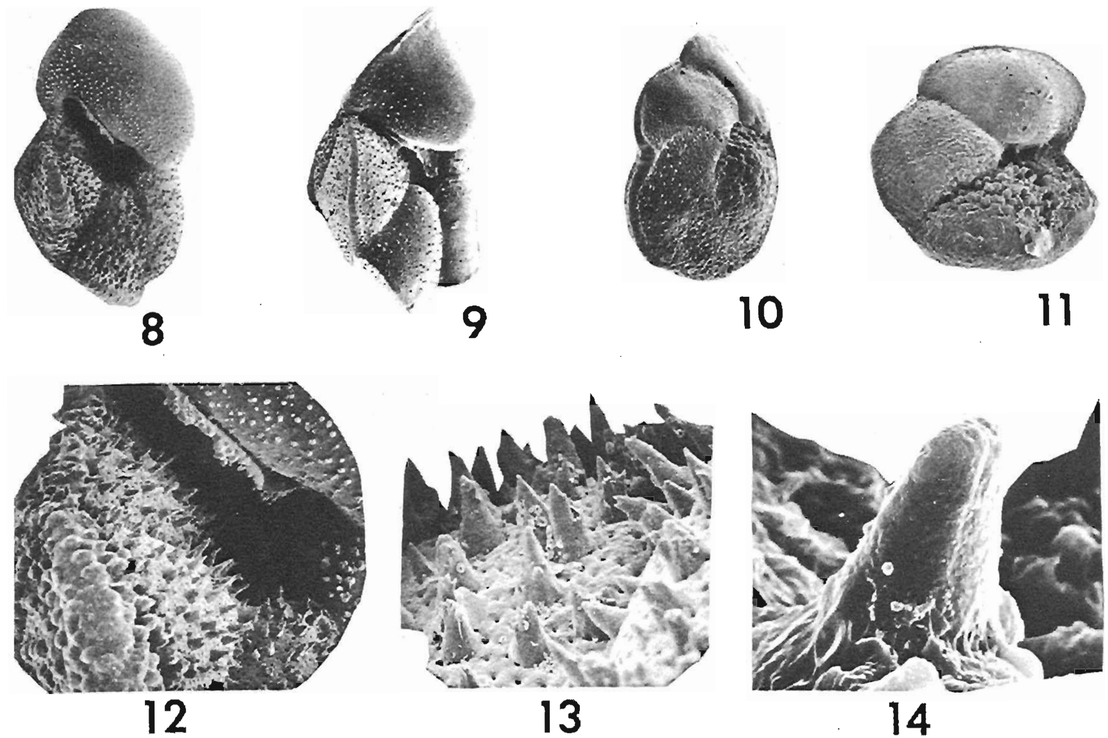

Plate 3 (continued); Figs. 8-14: <u>Globorotalia</u> <u>hirsuta</u> (d'Orbigny)

8-11. Apertural, dorsal and ventral views showing test morphology, note relatively smooth ultimate chamber, keel, and pustule development in apertural region, x100;

12. Detail view of apertural opening showing thick keel, well developed spines and lip, x300;

13. Detail of spinal arrangement and pores in apertural area, x1000;

14. Microstructure of individual spine, x3000.

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(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 73-107)

5. REGIONAL GEOLOGY OF OFFSHORE EASTERN CANADA

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Abstract

The continental shelf of eastern Canada extends from the Gulf of Maine to the head of Baffin Bay in the Northwest Territories (Lat. 42° to Lat. 77° N), a distance of approximately 3,500 miles (5,600 km). On the mainland, the Maritime Provinces, exclusive of Labrador, occupy the northern part of the Appalachian System. This belt of folded and unfolded rocks, was involved in the Middle to Late Ordovician (Taconic) and Middle Devonian (Acadian) orogenies. Middle and Late Paleozoic faulting and folding along the narrow Fundy Geosyncline resulted in the deposition of Carboniferous continental and marine sediments in intermontane troughs, followed by flat-lying Late Pennsylvanian and Permian sediments. Triassic sediments occupy an area of renewed faulting in the Bay of Fundy and Chedabucto Bay. The region has been positive since that time. On the continental shelf and slope, total thickness of sediments overlying a granitic or metamorphic basement, as determined from seismic refraction, magnetic depth-to-basement calculations and estimates from the total depths of 17 wells drilled to date, are indicated to be greater than 5 km (16,000 feet) on the Scotian Shelf and greater than 6 km (20,000 feet) on the Grand Banks and the northern Newfoundland banks. Results of drilling show typical coastal plain sediments, Jurassic to Tertiary in age, covered by a layer of Quaternary sediments that includes glacial drift. Triassic and Late Paleozoic sediments may be represented in troughs on the inner part of the Shelf. In the northern part of the Gulf of St. Lawrence and on Anticosti Island, nearly flat-lying Early Paleozoic platforms rocks dip gently to the south and are at least 3.7 km (12,000 feet) thick. In the central and southern part of the Gulf, sediments reach a maximum thickness of 7.6 km (25,000 feet); velocity data and extrapolation from onshore geology suggests the presence of mainly Permo-Carboniferous and possible Triassic sediments. No information has been released to date on three wells drilled in the Gulf of St. Lawrence. No published refraction data is available for the Bay of Fundy, but Triassic and Carboniferous sediments are predicted from onshore geology and from regional tectonics. Seismic refraction profiles display up to five velocity layers which may be roughly correlated throughout the area: Layer 1, 5,900-7,200 ft/sec (1.8-2.2 km/sec); Layer 2, 7,300-11,200 ft/sec (2.2-3.4 km/sec); Layer 3, 10,500-14,800 ft/sec (3.2-4.5 km/sec); Layer 4, 14,200-18,400 ft/sec (4.3-5.6 km/sec); Layer 5, 15,500-20,800 ft/sec (4.7-6.3 km/sec). An approximate correlation of the layers is made with Tertiary, Upper Cretaceous, Lower Cretaceous-Triassic, Permo-Carboniferous and basement. The distribution of the deepest refraction layers suggests a fragmentation of the underlying basement in post-Carboniferous time with initial deposition in marginal troughts, followed by the onlap of coastal plain sediments.

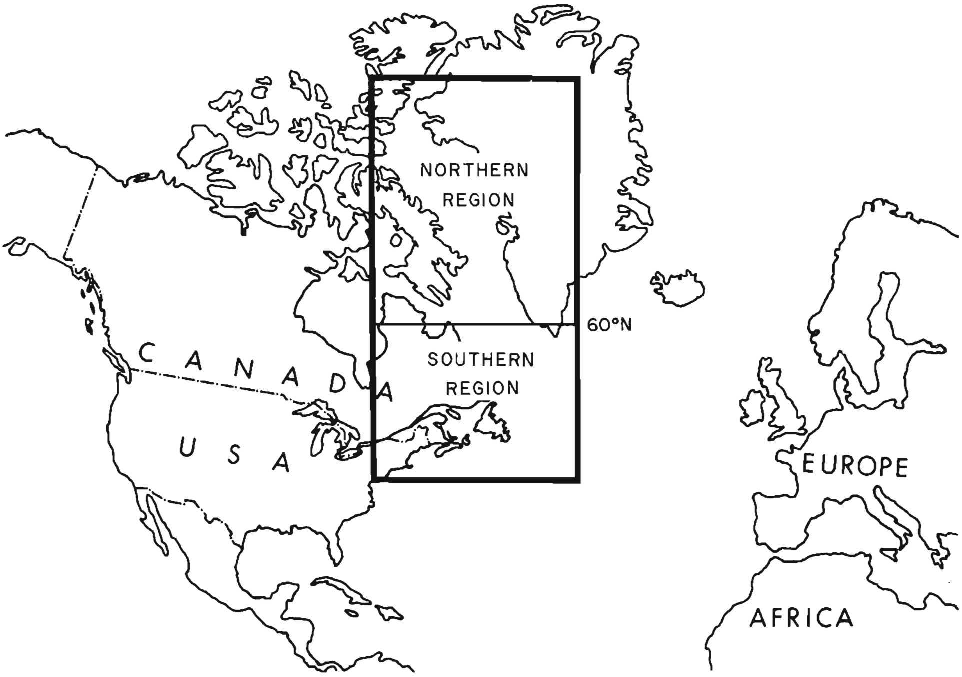

Figure 1. Index map to eastern Canada offshore.

North of the Appalachian System and at right angles to it are the narrow Labrador and Baffin continental shelves, presumably underlain by Precambrian basement rocks ranging in age from Archean to Proterozoic. The total thickness of sediments on the Labrador Shelf, based on depth estimates from magnetics and refraction data, is in excess of 6 km (20,000 feet) with compressional velocities consistent with the coastal plain sediments to the south. On the Baffin Shelf, a total thickness of sediments in excess of 9 km (30,000 feet) has been inferred from magnetic survey data. Onshore geology suggests the shelf may be underlain mainly by, Cretaceous and Tertiary sediments. Ordovician rocks are present in down-faulted structures along part of the coastline.

Although oil shows and seepages have been reported in each of the Maritime Provinces, the only commercial oil and gas production is from the Mississippian Horton Group at the Stony Creek field in New Brunswick. Many of the offshore wells drilled (as of February, 1971) have shows of oil and gas, but no commercial production has been reported by the operators.

Petroleum exploration on the continental shelves of offshore eastern Canada has stimulated research in the theory of continental drift, which could have a direct bearing on the nature and distribution of sediments, and hence petroleum along the continental margin.

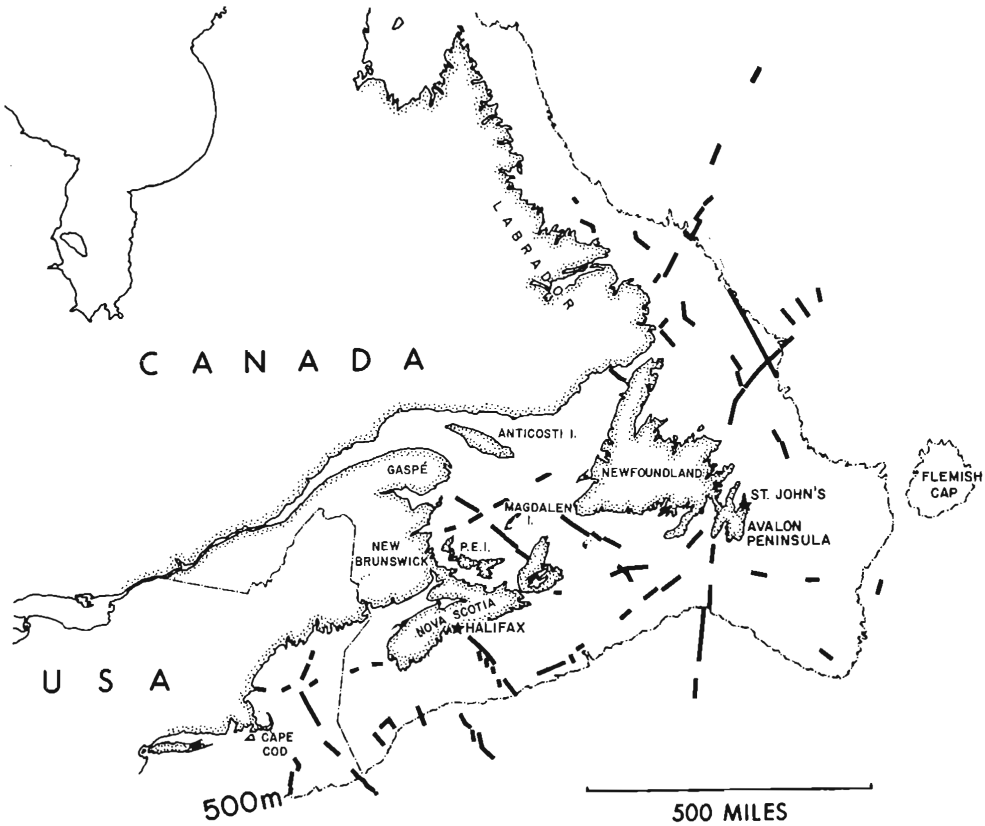

Figure 2. Seismic refraction profiles. Variously Woods Hole, Lamont Geological Observatory, Dalhousie University, Geological Survey of Canada.

INTRODUCTION

The continental shelf of eastern Canada extends from the Gulf of Maine off Nova Scotia to the northern end of Baffin Bay in the Arctic Islands (Lat. 42° N to Lat. 77° N), a distance of approximately 3,500 miles (5,600 km) (Fig. 1). The shelf varies in width from a maximum of 300 miles (480 km) on the Grand Banks to a minimum of 5 miles (8 km) at the northern tip of Baffin Island, and encompasses an area greater than 500,000 square miles (1.45 million sq km). Included in this region are the Scotian Shelf, Grand Banks, Newfoundland and Labrador Banks, Baffin Shelf and the inland waters of the Gulf of St. Lawrence and the Bay of Fundy.

Because of the very large areal extent, geographic position, and geological variations, the area has been divided into a southern and northern region with the dividing line at 60° N Lat. (Fig. 1).

SOUTHERN REGION OF OFFSHORE EASTERN CANADA

History of Exploration of Southern Region

Since the middle 1930's, a number of seismic refraction surveys have been carried out by government agencies and by universities in a continuing study of the continental shelf off Nova Scotia, Newfoundland and Labrador. Notable among these are Woods Hole Oceanographic Institution, Lamont Geological Observatory, Bedford Institute, Geological Survey of Canada and Dalhousie University, and Figure 2 shows the location of the seismic refraction profiles shot to date.

In recent years more detailed surveys in the form of continuous seismic profiler, gravimeter, magnetometer, bottom coring and dredging have been carried out mainly by Canadian government agencies. The most significant regional compilations were published by Drake et al.. (1965), Sheridan and Drake (1968) and Mayhew et al., (1970) to whom the authors are indebted for much of the basic data in this paper.

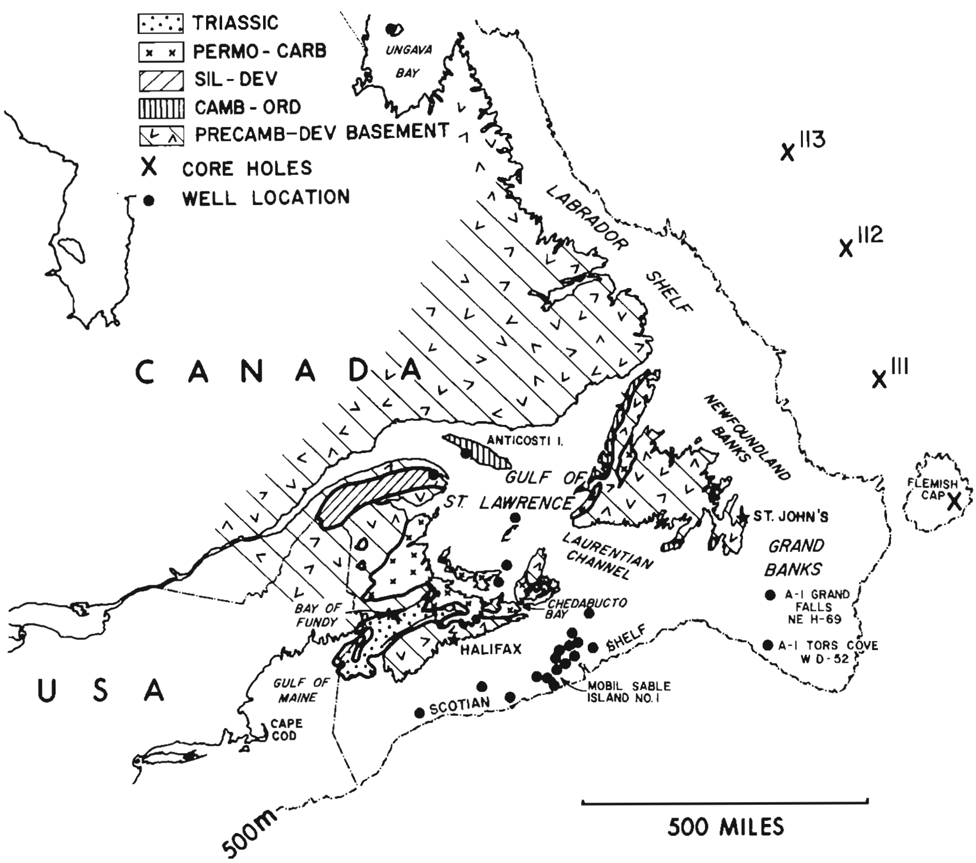

Figure 3. Surface geological map eastern Canada (southern region). Well locations as of February, 1971.

The first active interest in offshore petroleum exploration was in 1959, when Mobil Oil Canada Ltd. obtained 1.2 million acres of exploration permits on the Scotian Shelf on and around Sable Island. By the end of 1970 almost the entire offshore region to the edge of the shelf and in parts of the continental slope and rise were under lease.

The first two wells in offshore eastern Canada, Pan Am IOE A-1 Tors Cove W, D52 and Pan Am IOE A-1 Grand Falls NE H-09, were drilled in 1966 from a surface barge on the Grand Banks to depths of 4,834 feet (1,450 metres) and 5,250 feet (1,575 metres) respectively (Fig. 3). This was followed in 1968 by Mobil Sable Island No. 1, the first deep well on the Scotian Shelf, which was abandoned at 15,106 feet (4,532 metres). Information released on these wells indicate a typical sequence of Atlantic coastal plain sediments from Jurassic to Tertiary in age. Gas shows were reported from the Sable Island[1] and Tors Cove wells. In addition, gas and/or oil shows have been reported in trade journals from a number of other wells presently on tight-hole status.

As of February 22, 1971, thirteen additional wells had been drilled by Shell Oil on the Scotian Shelf, and two wells had been drilled by the Hudson Bay-Fina-Getty-Skelly consortium and one by Texaco in the Gulf of St. Lawrence.

Regional Geology of Southern Region

The simplified regional geology of the mainland portion of the area is shown in Figure 3. Major geological subdivisions include Triassic, Permo-Carboniferous, Silurian-Devonian, Cambro-Silurian tectonic cover and basement. The term basement includes crystalline rocks, volcanics, intrusives and metamorphic sediments, mainly of Precambrian to Devonian age.

Triassic rocks comprising fluviatile sediments and basic lava flows and some Carboniferous rocks are exposed on the shores of the Bay of Fundy. Continuous seismic profiler surveys by Swift and Lyall (1967) and others indicate the Bay to be a half graben floored by Triassic rocks with a downfaulted side to the north. On Figure 3, the seaward termination of the Triassic into the Gulf of Maine has been taken from Uchupi (1966), who concludes that the Gulf is floored by Paleozoic metamorphics and granites and Jurassic intrusives, covered by a veneer of Pleistocene and Recent sediments.

Permo-Carboniferous continental and some marine sediments occupy intermontane fault troughs in Nova Scotia, New Brunswick and Newfoundland. These rocks have been traced into the Gulf of St. Lawrence by seismic refraction surveys (Sheridan and Drake, 1968), and probably underlie Triassic rocks in the Bay of Fundy.

The Silurian and Devonian, marine and continental facies of Gaspé occupy a greatly deformed northeast-trending basin that escaped the intense Acadian folding. The nearly flat-lying sedimentary rocks of Ordovician-Silurian age which occur on Anticosti Island and the Ordovician-Cambrian sequence of western Newfoundland are cover rocks on the southern flank of the Grenville Orogen (Fig. 4).

As may be surmised, the onshore geology gives an incomplete picture of the sedimentary sequence present offshore. Refraction seismic data, together with wells drilled, grab samples and cores, show the continental

[1] Sable Island E-48 well, a 10 mile stepout from Sable Island no. 1 well, is an indicated discovery from 17 zones in the Cretaceous. Additional wells are now being drilled directionally from a platform to evaluate this discovery.

shelf to be comprised of coastal plain sediments similar to those present further south along the eastern seaboard of the United States. Sediments of Jurassic, Cretaceous and Tertiary age consist of sandstones, shales, siltstones, evaporites and some carbonates. Older Permo-Carboniferous and Triassic rocks exposed onshore can with reasonable certainty be traced by seismic surveys onto the inner shelf, and may also be present along portions of the outer shelf and slope.

On the Flemish Cap, a recent drill-hole by the Bedford Institute recovered granodiorite with a Precambrian or early Cambrian age (592 m.y.) indicating the cap is a detached portion of the continental shelf (Pelletier, 1971). Cores from Sites 111, 112 and 113 (Fig. 3) in the deep ocean basin were recovered by the JOIDES Deep Sea Drilling Project to determine the age of oceanic basement (Laughton et al., 1970). From an examination of the core recovered at Site 111 on Orphan Knoll, it was concluded that Orphan Knoll is a detached piece of the continent that foundered to oceanic depths between mid-Jurassic and Eocene time. The hole drilled at Site 112 encountered oceanic basalt, and the age for the overlying sediments was estimated to be 70 million years. This compares favourably with the extrapolated age from sea-floor spreading of 67 million years. Site 113 was abandoned before basement was reached.

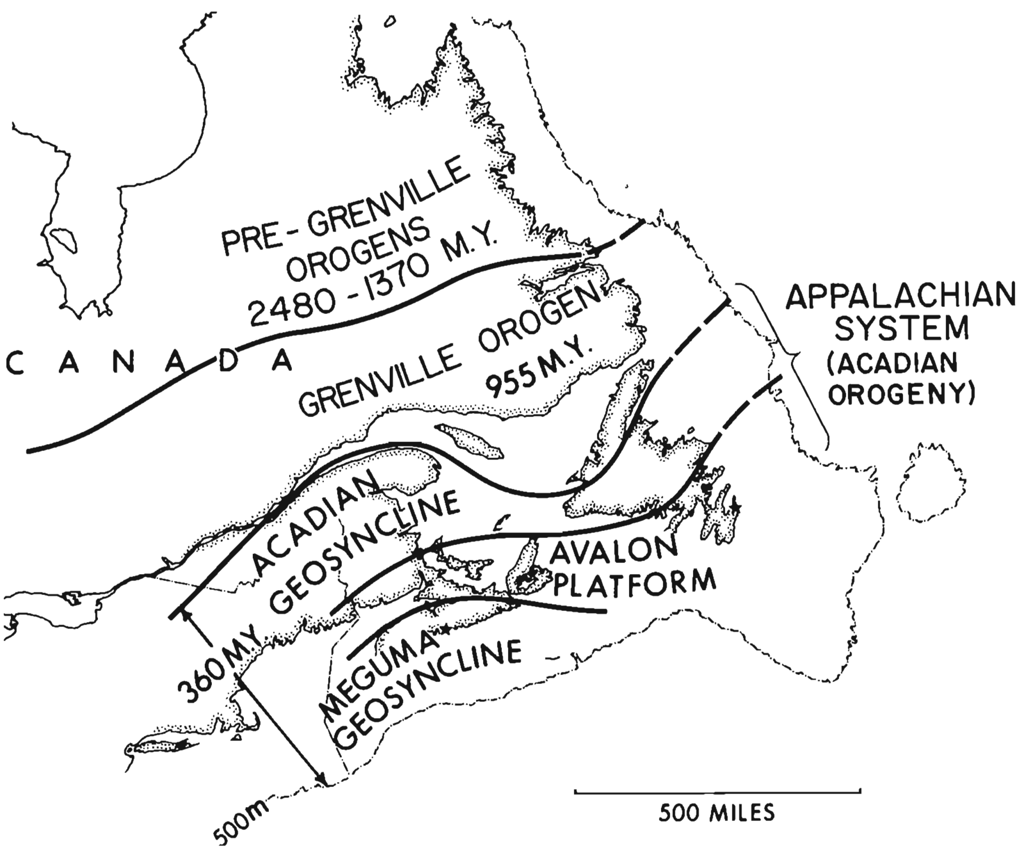

Figure 4. Mid-Paleozoic tectonic map showing area subjected to Acadian Orogeny (Modified from Poole, 1967).

Tectonic History of Southern Region

The simplified tectonic map (Fig. 4) shows the tectonic relationships during the middle Paleozoic. The Acadian Geosyncline, Avalon Platform and the Meguma Geosyncline are all part of the much larger Appalachian geosynclinal system. The Acadian Geosyncline, deformed during the Taconic Orogeny (440 m.y.) was refolded during the Acadian Orogeny. Together with the Avalon Platform and Meguma Geosyncline, it became part of the rising Acadian Orogen.

It is to be noted that crystalline and metamorphic rocks constituting basement became progressively stabilized from north to south in younger rocks with the major orogenic belts crossing the shelf at right angles to the continental margin east of Labrador and Newfoundland. Where these tectonic belts cross the shelf, significant structural changes are likely to be present affecting deposition offshore.

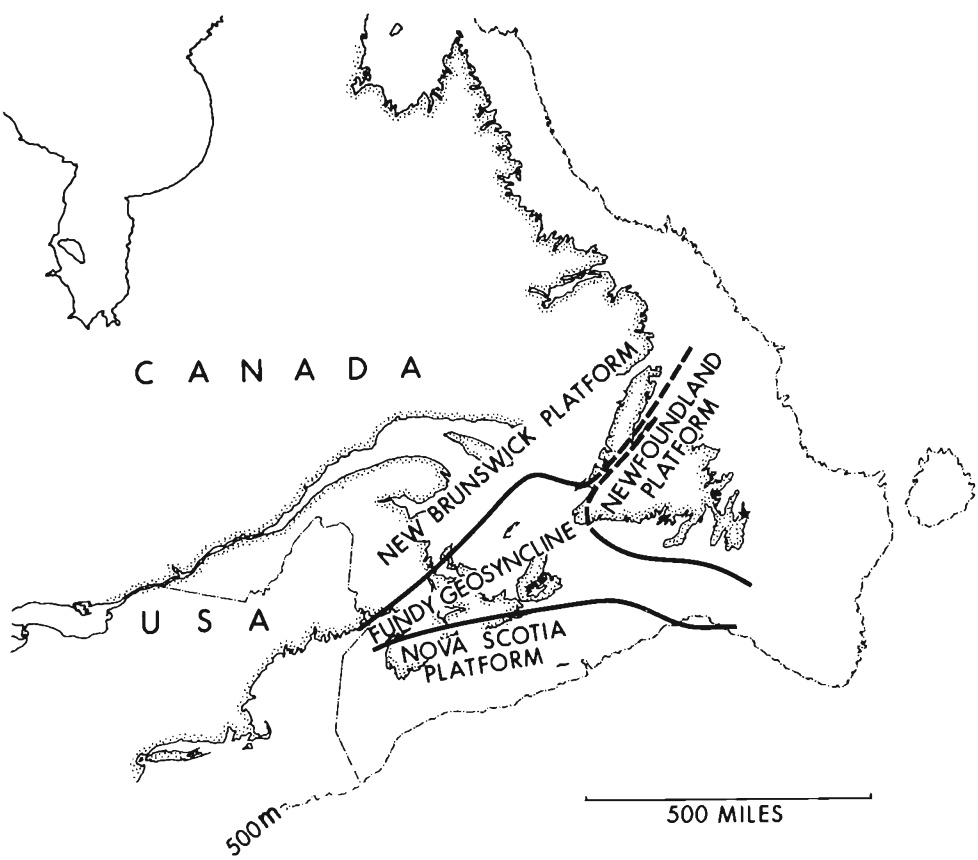

Figure 5. Late Paleozoic tectonic map showing area of Fundy Geosyncline subjected to Maritime disturbance (Modified from after Poole, 1967).

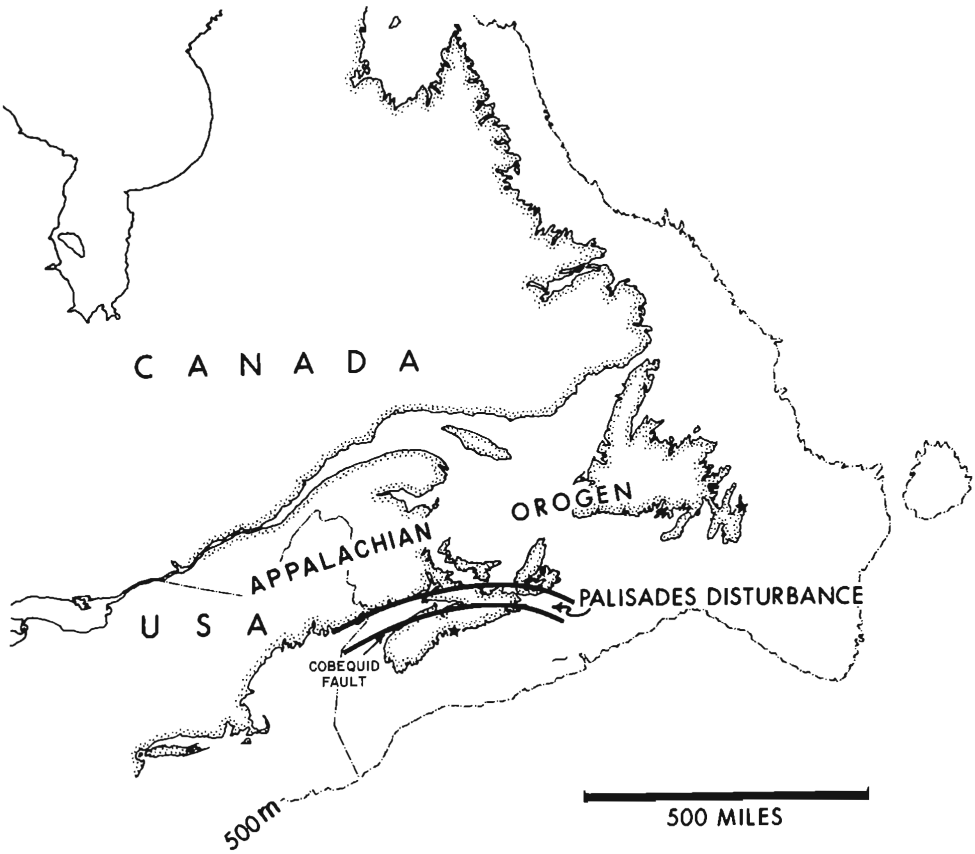

Figure 6. Triassic tectonic map showing local Palisades Disturbance (Modified from Poole, 1967).

Through the late Paleozoic, the Acadian Orogen was deformed mainly by faults, warps and gentle basinal subsidence, that gradually decreased in intensity (Poole, 1967). Folding and faulting in the Fundy Geosyncline, may have extended across the seaward edge of the southern Grand Banks (Fig. 5). Towards the end of the Carboniferous, nearly flat-lying Pennsylvanian and Permain sediments provided post-tectonic cover for most of the area. A late-stage adjustment, the Triassic Palisades Disturbance reactivated movement along the Cobequid-Chedabucto fault zone (Fig. 6).

From Triassic time through to the present, the Maritime Provinces have remained a positive area with little recorded tectonic activity on land, the only sedimentary occurrences being local deposits of Cretaceous age in the Shubenacadie area of Nova Scotia (Stevenson, 1959). Epierogenic uplift of the Appalachian System during late Cretaceous and early Tertiary was accompanied by intrusion of stocks and plugs in the provinces of Quebec and Newfoundland, and general subsidence and faulting along the continental margin.

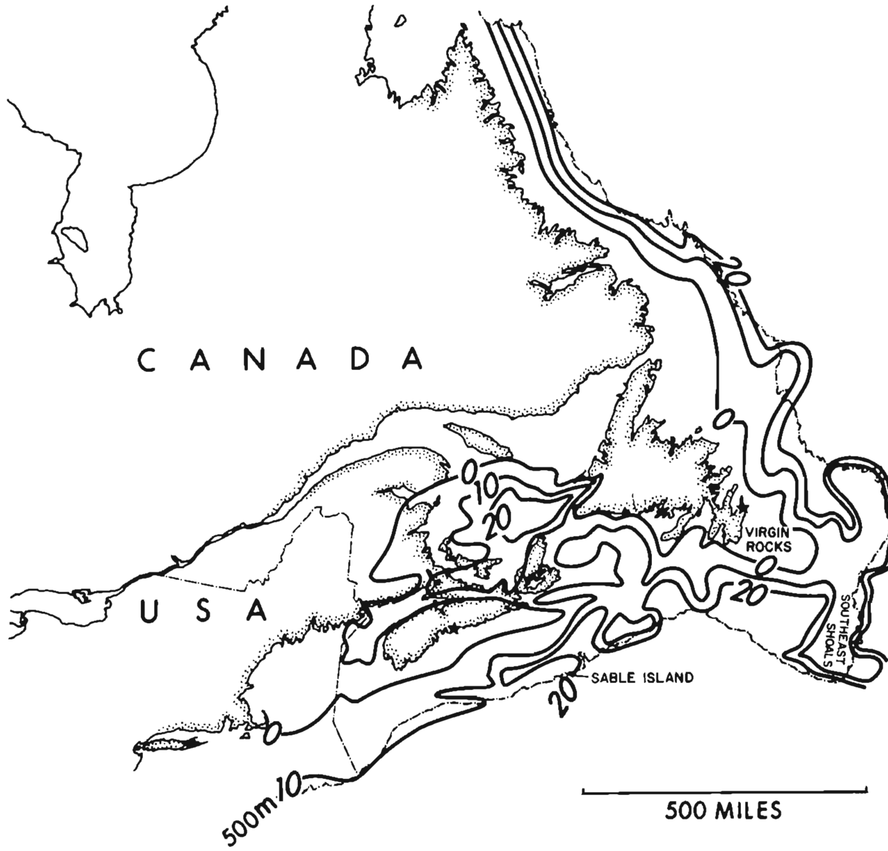

Figure 7. Total thickness sediments (post-Acadian)based on velocity cut-off of 5.0 km/sec (16,400 ft/sec) or greater. Isopach interval thousands of feet.

Offshore Southern Region

Seismic refraction surveys, magnetic interpretations and information derived from wells drilled to date indicate great thicknesses of sediments are present on the continental shelves, slopes and rises of the southern region. Figure 7 shows the total thickness of post-Acadian sediments calculated from all available data. Utilizing refraction data, petroleum basement has been assigned compressional velocities of around 5.0 km/sec (16,400 ft/sec) and greater, depending on the area. This velocity is consistent with velocities of 4.9-6.1 km/sec (16,000-20,000 ft/sec) obtained by Drake et al. (1959) from typical basement wells on the Atlantic coastal plain of the United States. In some areas the deepest refractors have velocities as low as 4.5-4.8 km/sec. Lacking other data these velocities are assumed to be basement, and may represent low-grade metamorphic rocks (Table 1).

As some of the Carboniferous evaporites and carbonates on the inshore portion of the Fundy Geosyncline have velocities in excess of 5.0 km/sec and since there is no way of being certain that all velocities in excess of this figure are basement, we must conclude that sedimentary thicknesses shown are minimum figures.

TABLE 1

SUMMARY CORRELATION REFRACTION LAYERS

Location of Refraction Profiles	Refraction Layers (Velocity km/sec)					
	Layer 1 1.8-2.2	2 2.2-3.4	3 3.0-4.5	4 4.3-5.6	5 5.0-6.3	6 6.0-6.7
Gulf of Maine	1.5-1.8		3.7			
Manticus		2.1			5.9-6.0	
Northeast Georges Bank	1.7-1.8	2.6-2.8		4.7-5.1		
Browns Bank	1.7-1.9	2.0-2.8			5.4	
Georges Bank Centre	1.5	2.9			5.6	
Georges Bank South	1.7-2.1	2.6-3.4		4.8		
Georges Bank Southwest	1.7		3.9			6.6
Portland-Maine	1.7-1.8	2.2-2.8	3.0-4.5		5.0-5.9	
Halifax-Sable South	1.6-1.8	2.3-2.8	3.4-3.9		5.1-5.9	
Banquereau-Artimon Bank	1.8	3.2		4.6		
St. Pierre Bank	1.7-1.8	2.8-2.9	3.8-4.1	5.0-5.3		6.1-6.3
Chedabucto Bay		2.0-2.3	3.5	4.3-4.6 5.4-5.5		
Laurentian Channel South	1.8	2.2-2.4	3.3-4.0	5.3-5.6	6.0-6.3	
Laurentian Channel North		2.1-2.6	3.9-4.2	4.4-5.4	5.4-6.3	
Gulf of St. Lawrence		2.0-2.6	3.2-3.8 4.0-4.4	4.8-5.4	5.5-6.2	
South Newfoundland	1.8-2.0	2.3-2.7	3.7-4.3		5.4-6.0	6.0-6.7
Grand Banks	1.8-1.9		3.7-4.1	4.9-5.3	5.8-6.2	
South Avalon Peninsula	1.6		4.3		5.5	
Grand Banks Southeast	1.7	2.7	4.3		5.7	
Grand Banks East	1.7				5.5	
Grand Banks Northeast[1]	1.8-2.2	2.2-3.3			5.0-6.1	
Grand Banks Northeast[2]	1.7-2.0	2.6	3.3-3.5	4.5-4.7		
Strait of Belle Isle			4.0		5.7-5.9	6.6-6.9
Strait of Belle Isle South				4.9-5.2	6.2	
Labrador	1.7		3.6-4.4			
Placentia Bay		2.7	4.0-4.5		5.2-6.0	6.4-6.6

Velocities to the right of heavy line are taken as basement. In the cases of Northeast Georges Bank, Georges Bank South, Banquereau-Artimon Bank and Grand Banks Northeast velocities shown for basement are low but could represent low-grade metamorphic rocks.

Refraction profiles shown on Figure 2 are mostly oriented at 90 degrees to the general depositional strike, which helps to outline the broad, regional depositional pattern on the shelf and slope. Depth-to-magnetic basement calculations by Margaret Bower (1961, 1962) and others give additional depth control in certain areas. Although data is not available on which wells penetrated basement, the total well depths have been used as a minimum depth-to-basement. Total thickness of onshore sedimentary strata was taken from Howie and Hill (1970).

The gravity compilation map (Fig. 8) by the Bedford Institute, which covers only part of the map-area, has been simplified to show negative and positive gravity anomalies. On Newfoundland and in the Gulf of St. Lawrence the gravity results are presented as Bouguer Anomaly maps and in the offshore region, as Free Air Anomaly maps. This grvaity anomaly map has been utilized in conjunction with total thickness of sediments to interpret the configuration of the basement. The deep evaporitic trough identified by

Loncarevic and Ewing (1966) extending seaward from Chedabucto Bay produces an elongated gravity minimum, the so-called Orpheus anomaly. Positive gravity anomalies on the edge of the Scotian Shelf confirm the interpretation by Officer and Ewing (1954) of a basement ridge at the shelf break, separating an inner sedimentary trough from an outer trough on the continental slope and rise. Other gravity minima near the Magdalen Islands and Prince Edward Island (Fig. 2) are also interpreted as thick sedimentary sequences containing evaporites.

On the northeastern Grand Banks, a large negative gravity anomaly is interpreted as a major sedimentary basin opening to the north. Separating this area from the basin to the south is a topographically high basement area (Fig. 7) with Precambrian strata outcropping at the Virgin Rocks which are similar to Precambrian outcrops on the Avalon Peninsula (eastern Newfoundland). Further to the east and south, Southeast Shoals may represent an underwater topographic extension of the Precambrian rocks of the Avalon Peninsula (Williams and Lilly, 1965). The Flemish Cap appears as a positive gravity anomaly separated from the main continental shelf by a deep sedimentary trough. This trough continues along the east and south side of Grand Banks.

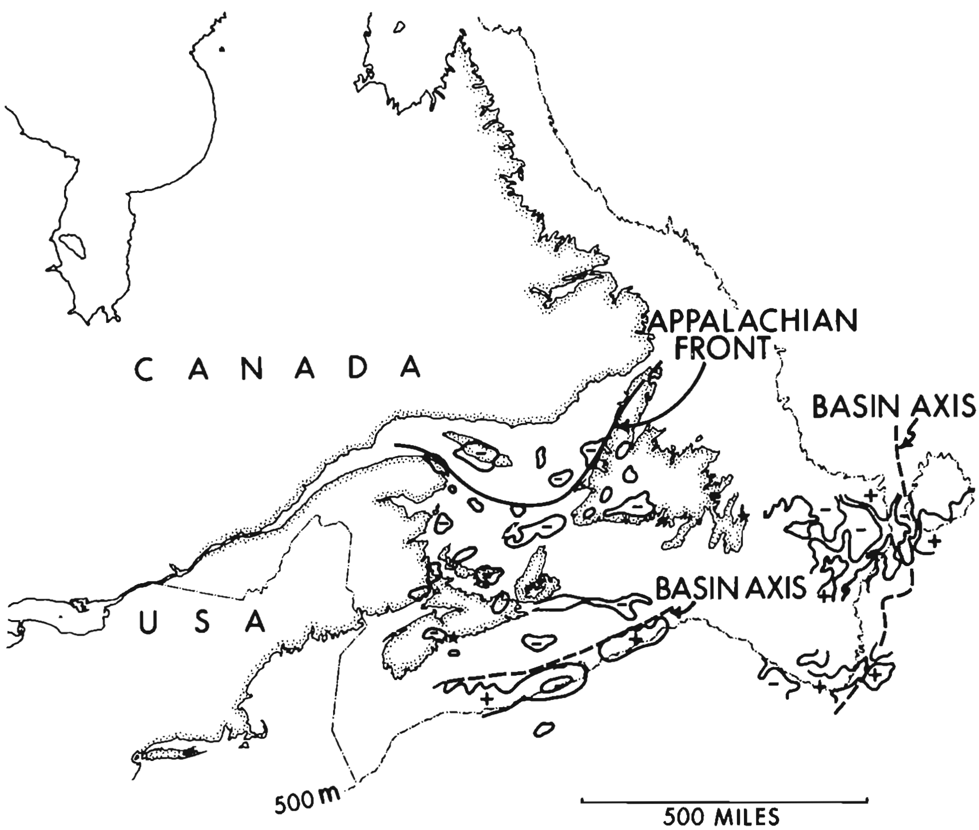

Figure 8. Bouguer anomalies - Gulf of St. Lawrence; free air anomalies - Grand Banks and Scotian Shelf (Modified from Bedford Institute compilation).

In addition to the gross features obtained from seismic refraction and gravity, continuous seismic profiles have been obtained on the Scotian Shelf (King, 1970), in the Gulf of Maine (Uchupi, 1966), in the Bay of Fundy (Swift and Lyall, 1967) and on the Grand Banks and Labrador Shelf (Grant, 1966, 1968). King's work on the Scotian Shelf indicates the presence of folded Cretaceous rocks in the area of the Orpheus gravity anomaly trending out of Chedabucto Bay, and the presence of at least two unconformities, at the base and near the top of the Tertiary. King and MacLean (1970a) report the location of a diapiric structure believed to be evaporitic on the Scotian Shelf close to the two Shell Oil wildcat abandonments, Iroquois J-17 and Abenaki L-57. Other diapiric structures have been variously reported on the Grand Banks, the Laurentian Channel and at the foot of the continental slope near the tail of the Grand Banks (Schneider, 1970).

Maximum thickness of sediments as deduced from all available data is in excess of 20,000 feet (6 km) over much of the outer parts of the continental shelves. No refraction or published depth to magnetic basement data is available for the Bay of Fundy and consequently no sediment thicknesses are given for that area.

On the Labrador Shelf (Fig. 7), the western edge of a marginal topographic trough appears to represent the boundary between Precambrian rocks and coastal plain sediments (Grant, 1966). This is confirmed by the change in character of aeromagnetic profiles (Hood et al., 1967) which indicates a marked increase in the depth to the crystalline basement near the western edge of the marginal trough. Sediments thicken rapidly eastward with indications of perhaps 30,000 feet or more on the Labrador continental slope and rise.

Seismic Velocity Intervals

In addition to the total thickness of sediments deduced from seismic refraction data, an approximate subdivision of the sedimentary section may be obtained by using the interval velocities of the different refracting layers. Much of the information for this compilation was taken from a report on refraction survey data by McConnell and McTaggart-Cowan (1963) and from Drake et al., (1959). Compressional velocities from seismic refraction profiles have indicated up to five different refracting layers (see Tables 1 and 2). Most of the layers can be correlated in a general sense from profile to profile and from area to area. An attempt has been made to correlate these seismic refractors with the known gross offshore geological section, bearing in mind that velocities reflect the lithification of the sediments and that no direct correlation can be made with stratigraphy (Drake et al., 1959).

Velocity Layer 1

The uppermost layer, Velocity Layer 1 (Fig. 9), has compressional velocities ranging from 5,900-7,200 ft/sec (1.8-2.2 km/sec). This is the unconsolidated layer of Officer and Ewing (1954). The unit is present on the outer portions of the continental shelf and slope, reaching a maximum thickness is excess of 6,000 feet (1.8 km) and is correlated approximately with the Tertiary section. This velocity layer is not recognized in the Gulf of St. Lawrence, but may be present in minor thicknesses in the Laurentian Channel and portions of the Gulf of St. Lawrence. It is not known to be present in the Bay of Fundy.

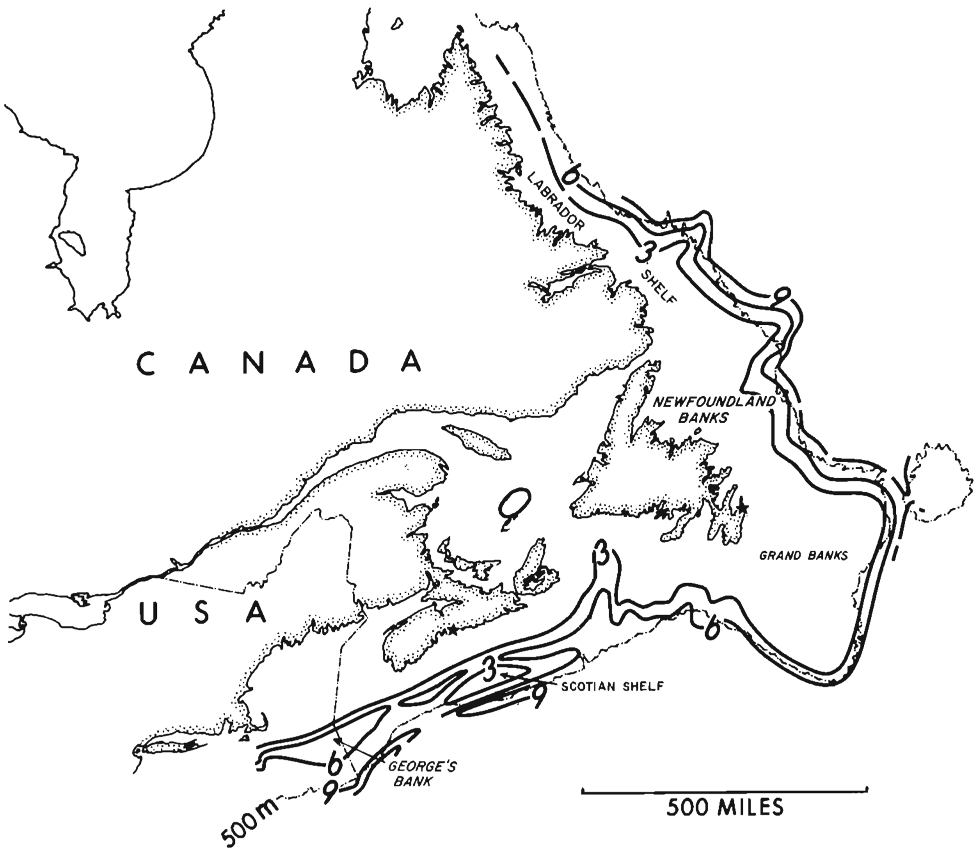

Figure 9. Velocity layer 1, Compressional Velocities 5900-7200 ft/sec (1.8-2.2 km/sec). Isopach Interval in thousands of feet.

Velocity Layer 2

Velocities in the range of 7,300-11,200 ft/sec (2.2-3.4 km/sec) are recorded from this layer (Fig. 10). This is the semi-consolidated layer of Officer and Ewing (1954). Strata represented are present everywhere on the coastal plain, reaching a maximum in excess of approximately 9,000 feet (2.7 km), but only a thin layer is present in the Gulf of St. Lawrence. The layer is identified approximately with Upper Cretaceous sediments. On the southern side of the Grand Banks, the few seismic profiles available (Fig. 2) show Velocity Layer 2 to be thin or absent, suggesting non-deposition on the southern Grand Banks or that subsequent uplift and erosion has occurred. On the northern Grand Banks, seismic data is not available but the large negative gravity anomaly (Fig. 8) suggests that Velocity Layer 2 may be present. On the Scotian Shelf and Georges Bank, where more refraction profiles have been shot than for any other part of the region deposition appears to have been controlled by a series of troughs and ridges paralleling the continental shelf edge.

Velocity Layer 3

Sediments with velocities in the 10,500-14,800 ft/sec (3.2-4.5 km/sec) range are assigned to Layer 3 (Fig. 11). The layer appears, within the limits of the seismic control (Fig. 2), to be absent from Georges Bank, Banquereau Bank and the banks south of Newfoundalnd. It may be present on the northern portion of the Grand Banks, but no refraction coverage has been obtained in that area. It would seem from the velocity range and regional distribution of the velocities that these represent older sediments possibly Pennsylvanian clastics in the Gulf of St. Lawrence and Triassic, Jurassic and Lower Cretaceous in faulted troughs on the outer edge of the continental shelf. Alternatively these velocities may represent areas of carbonate facies within a dominantly clastic depositional province.

On the Scotian Shelf, the trough shown has been assigned a possible Triassic age by Officer and Ewing (1954). This velocity interval is also present beneath Sable Island (Berger et al., 1965) and should have been penetrated by the Mobil Sable Island well which bottomed in Lower Cretaceous or Jurassic sedimentary rocks. The absence of Triassic sediments in this trough does not, however, rule out the possibility of the presence of Triassic in other troughs along the outer part of the Shelf.

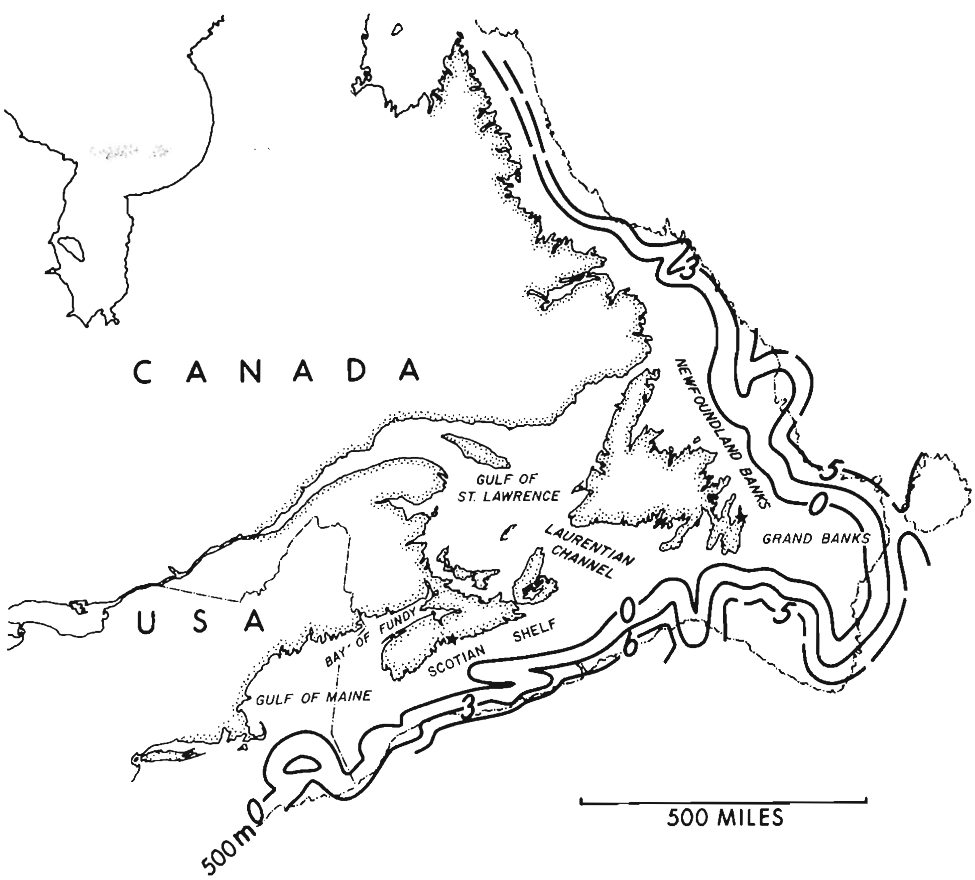

Figure 10. Velocity layer 2, Compressional Velocities 7,300-11,200 ft/sec (2.2-3.4 km/sec). Isopach Interval in thousands of feet.

Velocity Layer 3 probably represents the initial fragmentation of the offshore portion of the Acadian Orogen into separate basins which, along with the Labrador Shelf, was eventually blanketed by the Atlantic coastal plain sediments.

Velocity Layer 4

Sediments with the highest velocities 14,100-18,400 ft/sec (4.3-5.6 km/sec) have been assigned to Layer 4 (Fig. 12). These sediments are present mainly in the Gulf of St. Lawrence and have been correlated by Sheridan and Drake (1968) with Carboniferous rocks on the mainland[1]. Correlation can also be made with Permo-Carboniferous sediments in the Chedabucto trough (Fig. 3), Laurentian Channel and possibly along the south side of the Grand Banks. No refraction data is available for the Bay of Fundy,

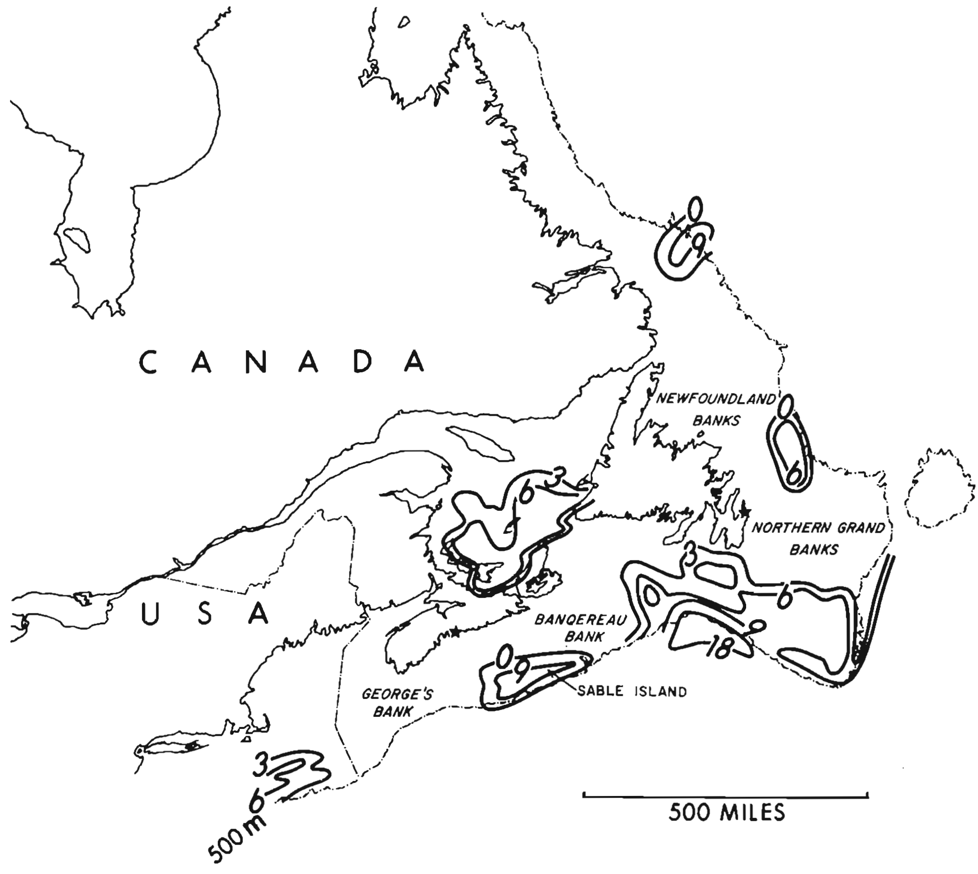

Figure 11. Velocity Layer 3, Compressional Velocities 10,500-14,800 ft/sec (3.2-4.5 km/sec). Isopach Interval in thousands of feet.

[1] Stratigraphic data on three wells in the Gulf of St. Lawrence released from confidental status in 1972, confirms sediments from Pennsylvanian to Mississippian age overlying a metasedimentary basement.

Figure 12. Velocity Layer 4, Compressional Velocities 14,100-18,400 ft/sec (4.3-5.6 km/sec). Isopach Interval in thousands of feet.

but it is not unreasonable to suppose that Permo-Carboniferous sediments which outcrop on the north side of the Bay of Fundy also underlie this Triassic basin.

Velocities in the range 15,500-16,500 ft/sec (4.7-5 km/sec) on parts of the Scotian Shelf have been assigned to basement in line with reported velocities for low-grade metamorphics and granites on the New England coast and to the south (Drake et al., 1959; Hood, 1967a). It is possible that some of these values represent high-velocity late Paleozoic or early Mesozoic carbonates. It might be conjectured from the isopach pattern that the dominant structural movement during the late Paleozoic Maritime Disturbance may have been through the Laurentian Channel and along the south side of the Grand Banks. Sheridan and Drake (1968) and King (1970), from a consideration of the refraction, seismic profiler and gravity results, deduced no large-scale displacements in an east-west direction, but concluded that the structural grain crosses the Laurentian Channel at an angle which parallels the Appalachian System.

Cross-Sections Across the Continental Shelves of the Southern Region

The index map of cross-sections (Fig. 13) gives the location of pertinent cross-sections to be described. These are seismic refraction sections, with the exception of A-A' which is a structural stratigraphic section and P-U is an aeromagnetic profile.

Section M-M' Across the Scotian Shelf

Section M-M' (Fig. 14) which is known as the Halifax section (Officer and Ewing, 1954), shows the top three refraction layers overlying an interpreted crystalline basement with velocities generally in excess of 18,400 ft/sec (5.6 km/sec). At the continental shelf break and slope, basement velocities decrease to 16,800 and 14,800 ft/sec (5.0 and 4.4 km/sec). This velocity decrease has been explained by Sheridan and Drake (1968) as due to a dying-out of the effects of metamorphism of the Acadian Orogeny. Another interpretation is that because these velocities are in general low for basement rocks, and because they seem to be associated with the structure of the continental margin, they may be unmetamorphosed, late Paleozoic (post-Acadian) or early

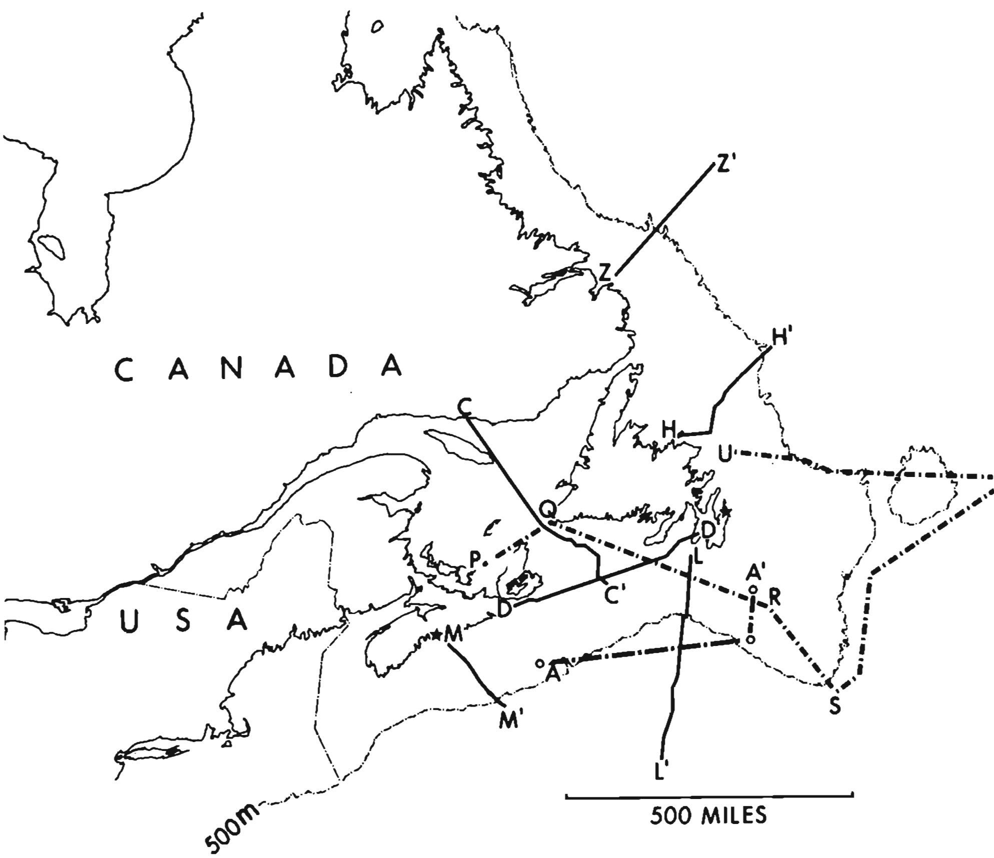

Figure 13. Index map of cross-sections.

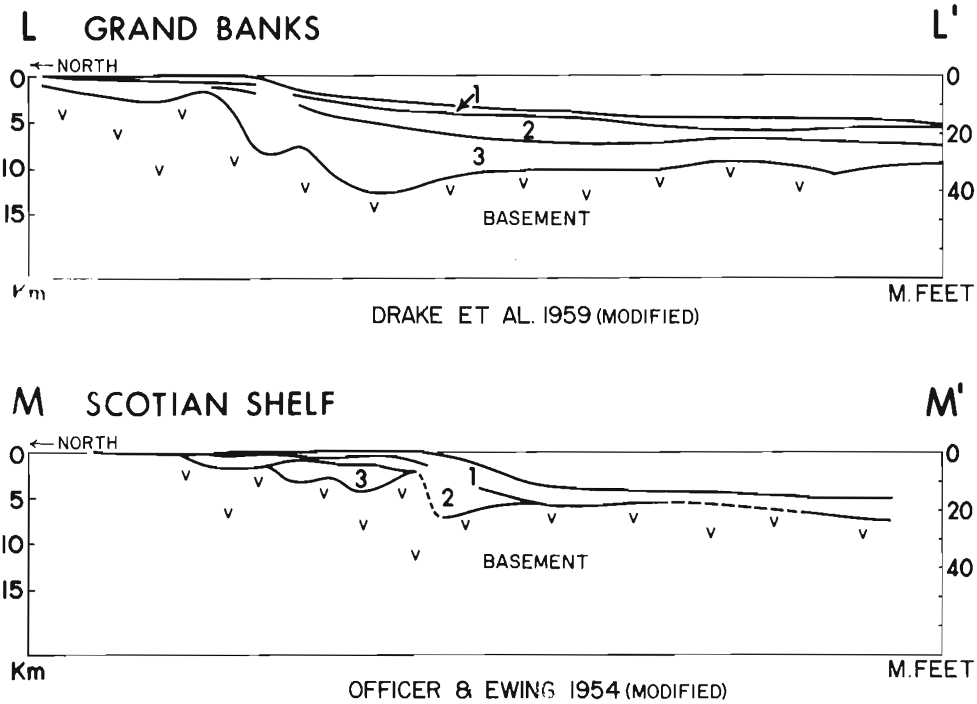

Figure 14. Seismic refraction sections, Scotian Shelf (M-M') and Grand Banks (L-L'). Velocity Layers 1, 2, and 3 overlying basement.

Mesozoic sediments deposited in marginal troughs on the outer edge of the continental shelf. The section shows the outer basement ridge separating the inner shelf deposits from the much larger sedimentary trough on the slope and rise which may contain older sediments. Berger et al., (1964) came to a similar conclusion based on the results of a refraction survey on Sable Island. High basement velocities under the island and on strike with the island decreased toward the continental margin indicating the presence of possible late Paleozoic or early Mesozoic sediments.

Section L-L' Across the Grand Banks

The Grand Banks section L-L' (Fig. 14) contains the three upper velocity layers. As in the Halifax section there is a prominent basement ridge with a thick prism of sediments at the foot of the continental slope and rise. The forty-thousand foot (12 km) sedimentary section may be partly accounted for by submarine slides. Basement velocities in this section are similar to those on the Halifax section.

Section A-A' from Sable Island to the Grand Banks

The structural-stratigraphic section (Fig. 15) from the Sable Island well to the Grand Falls well shows these two wells have penetrated mainly

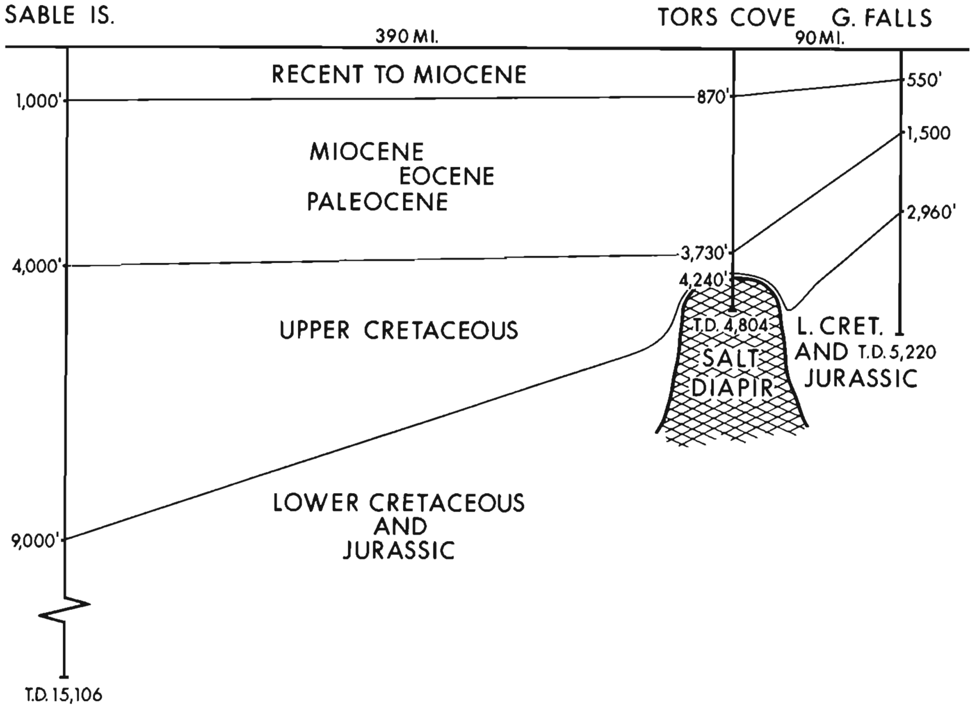

Figure 15. Structural-stratigraphic section Sable Island - Grand Banks.

shales, sandstones and siltstone with some thin beds of limestone. The Tertiary section at the Tors Cove and Sable Island wells are about equal in thickness. The Grand Falls well which is topographically higher on the shelf has a thinner Tertiary section. The evaporitic diapir encountered at the Pan Am IOE A-1 Tors Cove W, D-52 well makes it impossible to compare the lower stratigraphic section of this well with Mobil Sable Island No. 1. Sidewall cores from the caprock above the evaporitic diapir at Pan Am IOE A-1 Tors Cove W, D-52 is reported to contain indigenous palymorphs which have a range from Triassic to Lower Cretaceous Albian, but are most common in the Jurassic (Bartlett, and Smith 1971).

It is not possible to make more than an approximate correlation of the refraction velocities with the sedimentary sections encountered. On the basis of a comparison of the sedimentary thickness at Sable Island with the velocity data given by Berger et al., (1965), it is suggested that Zone 1 is probably Tertiary and Zones 2 and 3 represent the combined Upper Cretaceous, Lower Cretaceous and Jurassic.

Section C-C' Across the Gulf of St. Lawrence

Section C-C' (Fig. 16) (after Sheridan and Drake, 1968) crosses the Gulf of St. Lawrence at right angles to the strike of the Appalachian System. The section shows flat-lying Cambro-Silurian cover rocks separated by the Logan Fault from folded and metamorphosed rocks of the Appalachian System (Acadian Orogen). Layer 4 has been correlated with Carboniferous strata of the Maritime Disturbance. It is overlain by Layers 3 and 2 which in this area probably represent flat-lying Permo-Pennsylvanian cover rocks. Total thicknesses of sediments within these basins is in excess of 25,000 feet (7.5 km). A pre-Carboniferous structural ridge appears to extend across the Cabot Strait from Newfoundland towards Cape Breton Island.

Section D-D' Across the Laurentian Channel

Section D-D' (Fig. 16) extends from Chedabucto Bay to Avalon Peninsula. Here again troughs filled with Carboniferous or younger sediments (Layer 4) are overlain by Layers 3 and 2, similar to Section C-C'. The presence of basement ridges and troughs is substantiated by gravity data and seismic profiler (King and MacLean, 1970c).

Section P-S'

Section P-S (Fig. 17) is an interpretation of part of the airborne magnetometer profile of Hood and Godby (1965) interpreted by J. Britton (unpublished). The section contrasts older late Paleozoic sediments underlying the Gulf of St. Lawrence with the onlapping coastal plain sediments of the Grand Banks.

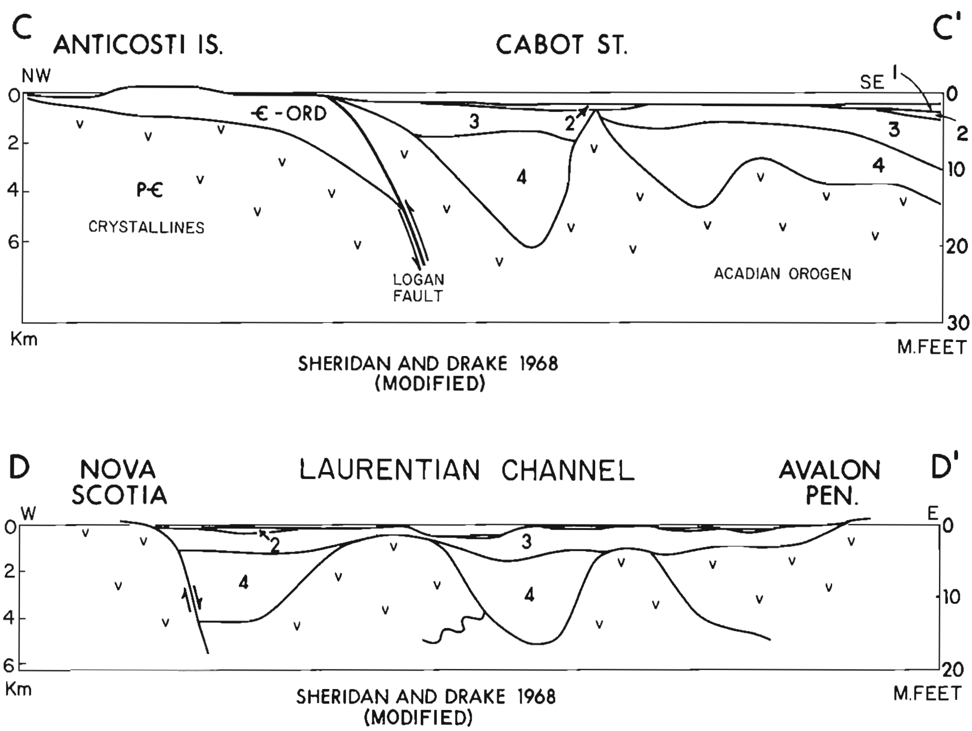

Figure 16. Seismic refraction sections, Gulf of St. Lawrence (C-C') and Laurentian Channel (D-D'). Velocity layers 1, 2, 3 and 4 overlying Basement.

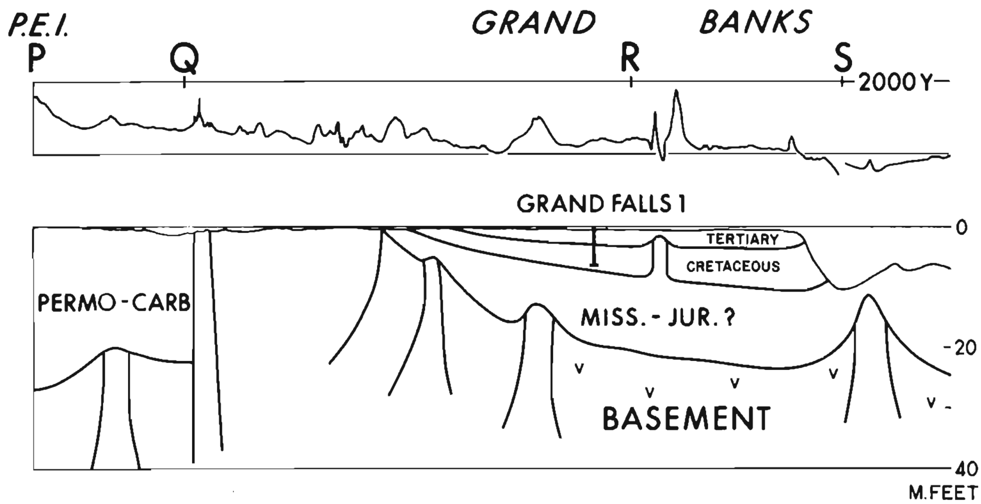

Figure 17. Airborne magnetometer profile (Hood and Godby, 1965). Interpreted by J. Britton (Unpublished).

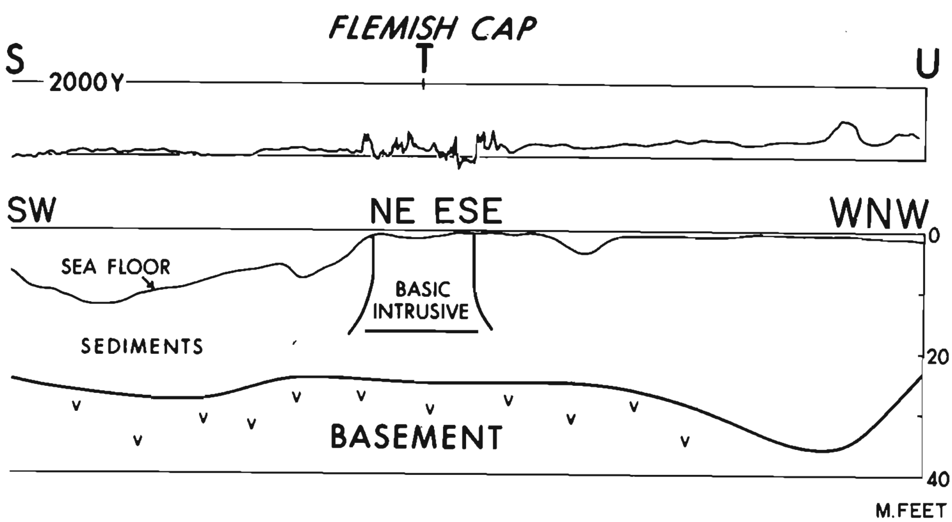

Figure 18. Unfolded airborne magnetometer profile across the Flemish Cap (Hood and Godby, 1965). Interpreted by J. Britton (Unpublished).

Section S-U' Across the Flemish Cap

Section S-U' (Fig. 18) is the remainder of the Hood and Godby (1965) airborne magnetometer profile which crosses the Flemish Cap twice. A core taken by the Bedford Institute on the Cap in 490 feet (147 m) of water recovered six inches of granodiorite having an age of 592 m.y. The interpretation is that the Cap is floored by Precambrian intrusive rocks and is a detached part of the continental shelf. A magnetic depth to basement calculation by Hood and Godby (1965) indicated a 25,000-foot (7.5 km) sedimentary section between the Flemish Cap and the continental shelf.

Sections H-H' and Z-Z'

Sections H-H' and Z-Z' (Fig. 19) on the Newfoundland and Labrador Shelves contain coastal plain sediments represented by velocity Layers 1, 2 and 3 onlapping a crystalline or metamorphic terrain.

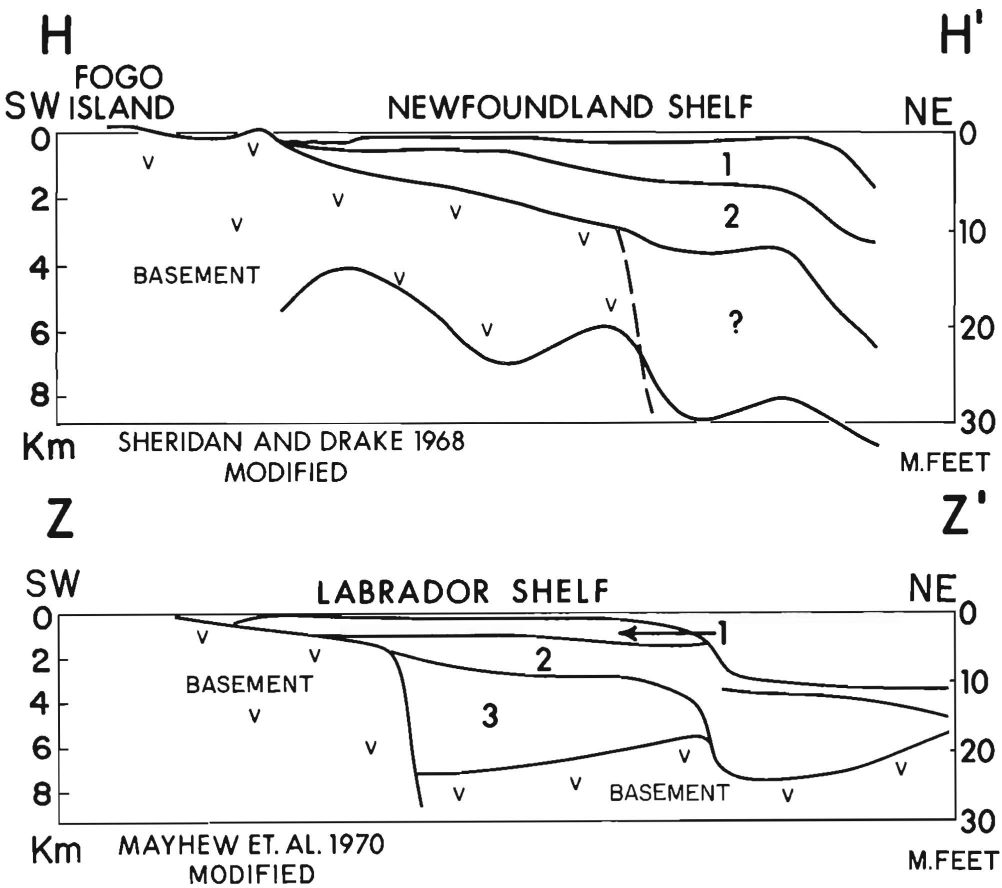

Figure 19. Seismic refraction sections showing coastal plain sediments. Velocity Layers 1, 2 and 3 over Basement.

On the inner part of the continental shelf on section H-H' the velocities of the basement rocks are between 5.7-6.3 km/sec (18,700 - 20,600 ft/sec). On the outer part of the continental shelf and slope, velocities decrease to 16,800 and 17,200 ft/sec (5.14-5.25 km/sec). As discussed on the Halifax section M-M' (Fig. 14), these latter velocities are well within the range of unmetamorphosed sediments and may be late Paleozoic or early Mesozoic rather than basement rocks. The underlying basement velocities may be sufficiently close to these velocities that their presence cannot be detected.

In the section Z-Z', the coastal plain sediments (Velocity Layers 1 2 and 3) are lying on a much older basement – the Grenville Orogen. On the inner portion of the shelf there is a significant inferred structural change in the attitude of the underlying basement, considered to be a half graben with the oldest sediments (Velocity Layer 3) probably being early Mesozoic. There is an interesting relationship between the two sections in that an apparent structural change appears to be present in each section about the same distance shoreward from the edge of the shelf, which indicates the probability of the presence of a deep sedimentary trough paralleling the outer part of the shelf.

It is worthy of note that Sheridan and Drake (1968) mention that the basement velocities identified with the Acadian Orogen do not persist to the edge of the continental shelf, and so inferred that the Acadian orogenic belt was not continuous with eastern Europe. If, however, the continuation of the Acadian orogenic belt to the edge of the continental slope has been masked by overlying high-velocity Paleozoic or Mesozoic rocks, valuable support is afforded to the theory of continental drift.

Correlation of Velocities with Stratigraphy

The correlation chart (Table 2) is an attempt to correlate the various velocity layers with the presence of rocks identified in the geological column by drilling or from rock outcrops. As discussed previously, it is not possible to correlate compressional velocities with stratigraphy, and this is merely an attempt at gross correlation. It is extremely hazardous to speculate on the lithology of rocks forming Veolocity Layer 4, because the velocities could indicate either granites, metamorphic rocks or sediments. Some velocities which have been included as basement in Layer 5 may actually be sediments with high compressional velocities.

NORTHERN REGION OF OFFSHORE EASTERN CANADA

The northern region covers the Arctic portion of offshore eastern Canada from the northern tip of Labrador to the head of Baffin Bay. The 500-metre bathymetric contour approximately delineates the shelf break (Fig. 20).

The Baffin Shelf is narrow in comparison with the continental shelf to the south, it varies in width from approximately five miles (8km) opposite Bylot Island to 110 miles (176 km) off Cumberland Sound. At the shelf edge, the bottom descends rapidly to oceanic depths (2,600 m). Opposite Cape Dyer in the Davis Strait, the depth of water is 800 metres. Rock outcrops on Baffin Island and on the west coast of Greenland are for the most part of Early Proterozoic age and form part of the Canadian-Greenland Shield. The few sedimentary exposures include Cambro-Ordovician cover rocks, on the west coast of Baffin Island, and Cretaceous and Tertiary marine and continental clastics on the East Coast. Thick Tertiary basaltic lavas up to 25,000 feet (7.5 km) thick overlie clastics in the Disko Island area of West Greenland. At Cape Dyer, basalts 1,400 feet (420 m) in thickness overlie thin clastics of probable Paleocene age.

Near the southern tip of Greenland, dyke swarms have been dated as Jurassic (Watt, 1969). Watt concluded that the distribution and age of the dykes suggests that initial rifting between Greenland and Labrador took place about the middle of the Mesozoic era.

Shipborne magnetometer profiles across the Bay interpreted by J.W. Murray, W.G. Libby and R.L. Chase (as reported in Oilweek May 11, 1970) indicate sediments in excess of 30,000 feet (9 km) along the continental shelf of Baffin Island, with an apparent basement ridge near the shelf break (Hood and Bower, 1970). Since published seismic refraction results are not available, sediment depth determinations are entirely based on magnetic data, and the computed sedimentary section could include Cambro-Ordovician cover rocks such as are believed to be present in fault grabens in Cumberland Sound, Frobisher Bay and Hudson Strait. The only well in the area was drilled by Premium Homestead on Akpatok Island in Ungava Bay; the hole was completed at 1,217 feet (365.1 m) in Precambrian rocks and encountered only early Paleozoic sediments.

Estimates of total thicknesses of sediments from airborne data on the west coast of Greenland are between 16,000 and 20,000 feet (4.8 and 6 km).

TABLE 2

APPROXIMATE CORRELATION
OF VELOCITY LAYERS WITH STRATIGRAPHY

Basement defined as crystallines, volcanics and metamorphics with velocities in excess of 16,400 ft/sec (5.0 km/sec)

VELOCITY LAYER	SCOTIAN SHELF	GRAND BANKS	GULF OF ST. LAW.	NFLD.& LAB. BANKS
1 5900 - 7200 Ft/SEC	TERT.	TERT.	TERT.-CRET.	TERT.
2 7300 - 11,200 Ft/SEC	U.CRET.	U.CRET.	U.CRET.	U.CRET.
3 10.500 - 14,800 Ft/SEC	L.CRET. - TR.	L.CRET. - JUR.	L.CRET.-PENN.	L.CRET. - JUR.
4 14.100 - 18,400 Ft/SEC	BASEMENT or PALEOZ.-MESOZ.	CARB.	CARB.	BASEMENT or PALEOZ.-MESOZ.
5 15.500 - 20,800 Ft/SEC	BASEMENT	BASEMENT	BASEMENT	BASEMENT

Labrador Sea Ridge

The Labrador Sea Ridge was predicted by Wilson in 1963, who stated that its presence had not been noted because of insufficient sounding data. Previous work by Lamont Geological Observatory had shown that the ocean bottom in the central Labrador Sea was not typical of ocean basins, but more closely resembled that found over the Mid-Atlantic Ridge. Later seismic profiler work by Drake et al., (1963) and by Johnson et al., (1969) confirmed the presence of a basement ridge (5.5 km/sec) near the axis of the Labrador Sea, and Godby et al., (1966) presented aeromagnetic data which showed that bands of magnetic anomalies were located on either side of this central axis

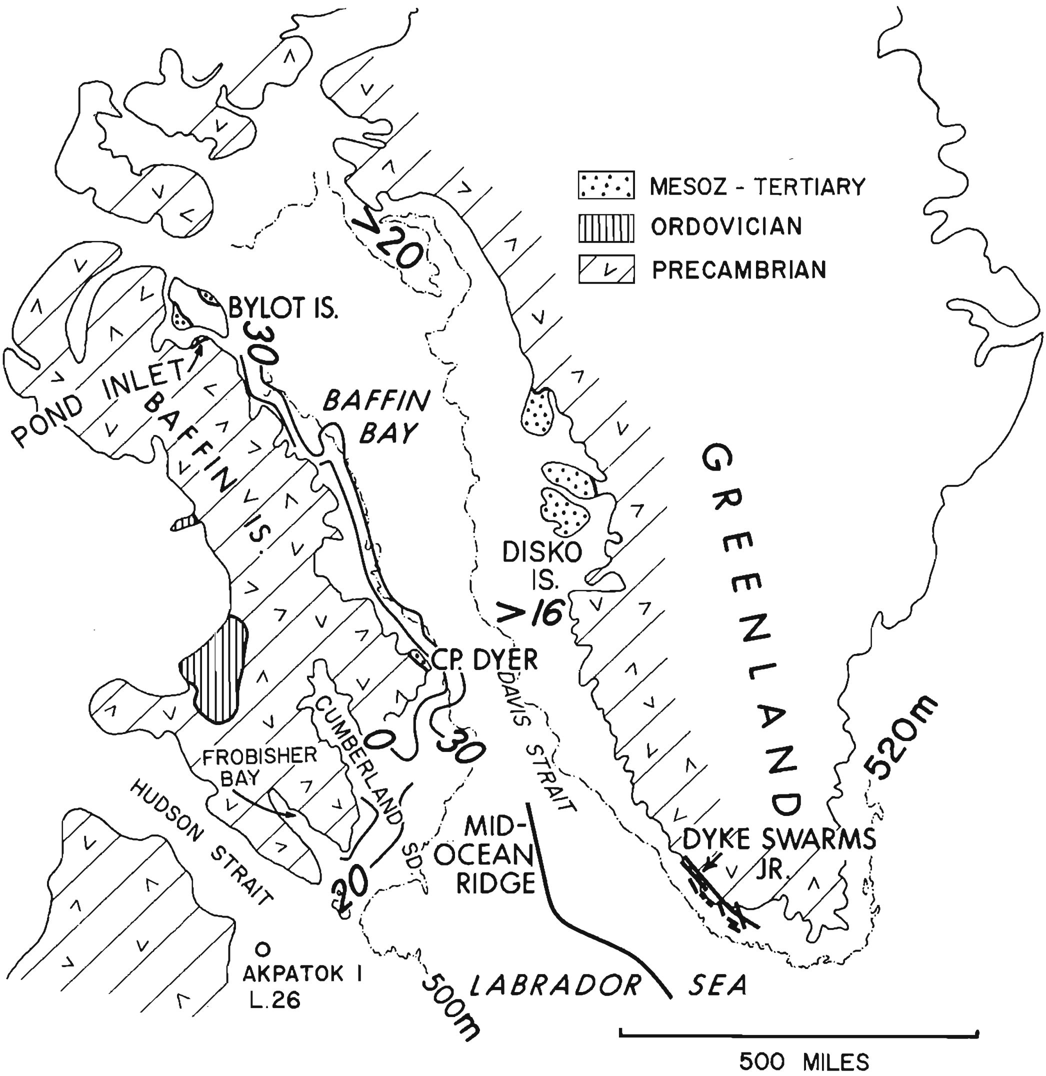

Figure 20. Surface geological map eastern Canada (northern region), showing estimated total thickness of sediments from aeromagnetic data in thousands of feet.

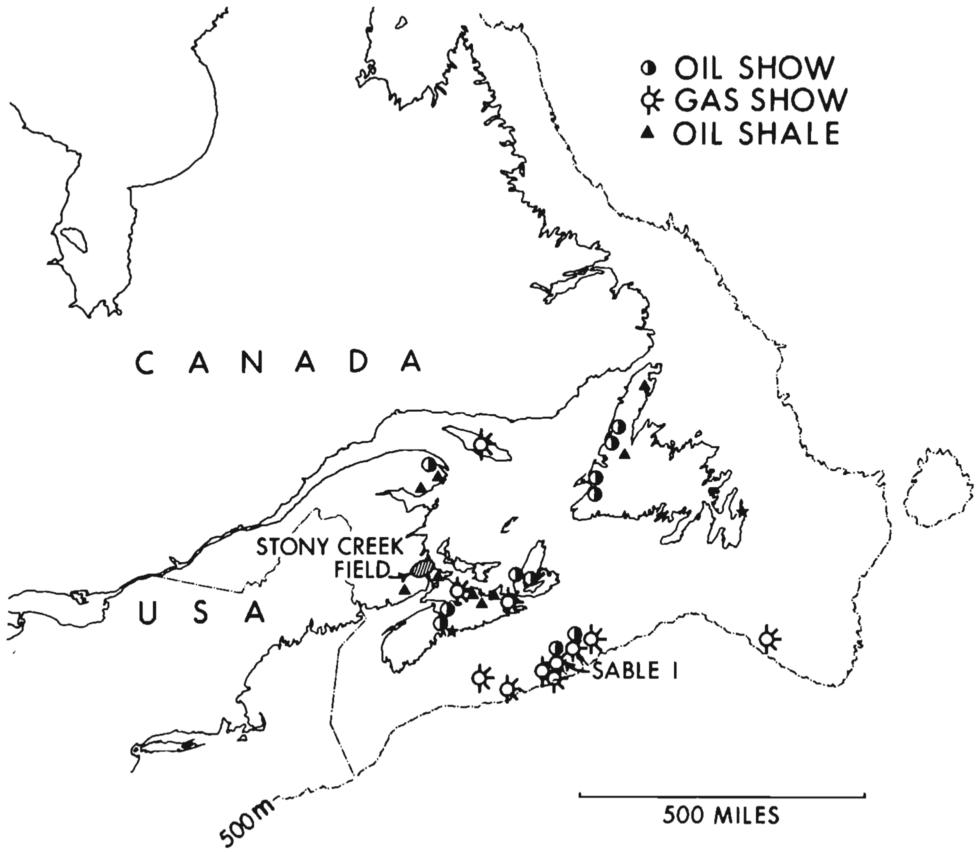

Figure 21. Oil and gas occurrences in eastern Canada - Maritime Provinces as of February, 1971 - Howie – modified.

indicating that oceanic crust underlies the sediment in the Labrador Sea. The ridge can be traced as far as the Davis Strait where it is buried beneath about 1 km of sediment. Because of the lack of seismicity over the ridge, and the absence of characteristic magnetic anomalies, Drake postulated a relic mid-ocean ridge, developed in earlier time than the mid-Atlantic Ridge.

Hydrocarbon Occurrences

Figure 21[1] is a simplified reproduction from Howie (1970) giving hydrocarbon occurrences in the Maritime Provinces. On land, oil and gas shows have been reported in sediments of Ordovician, Devonian and Carboniferous age. In addition, there are numerous occurrences of oil shale and albertite mainly in Carboniferous rocks.

[1] Release of the Shell Onondage E-84 well (16 miles southwest of Sable Island) in November, 1971 showed this well to have approximately 100 feet of gas and oil pay with no water in an overall interval of 400 feet. Mobil Tetco Sable Island E-48 on which information was also released in 1971 is an indicated gas and oil discovery in 17 separate zones in the Cretaceous.

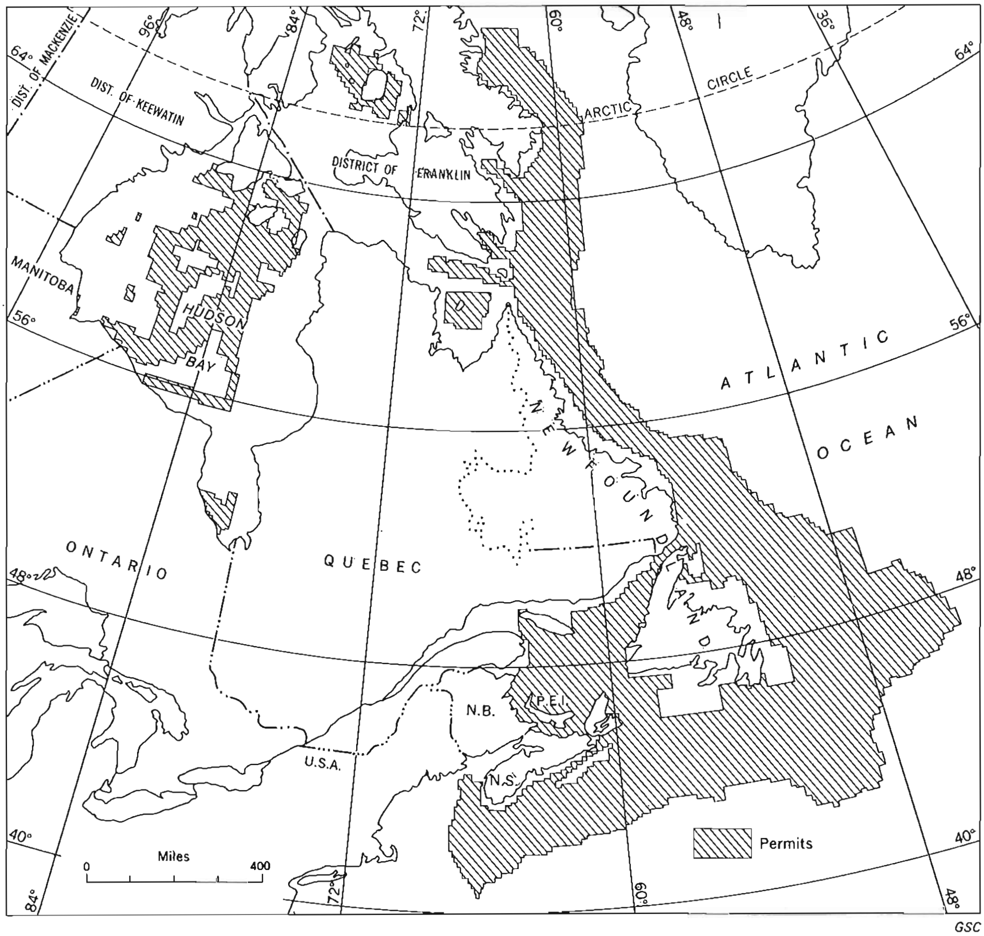

Figure 22. Approximate acreage held by industry under petroleum and natural gas permit, February, 1971, (289 Million Acres).

Although little information is available on the offshore wells, of the twelve wells on which official announcements have been made as of February, 1971 only two have not reported oil and/or gas shows. One well, Shell Mic Mac J-77, on the Scotian Shelf, was reported to have obtained encouraging oil shows which were considered non-commercial by the company at the time of the press release.

No oil and gas occurrences are known to be present on the Labrador Shelf or on the Baffin Shelf to the north. This is understandable because of the almost total absence of drilling as of the date of this paper as well as the very few sedimentary outcrops in this region.

Industrial Land Holdings

In the early part of 1971, total acreage held by industry in the form of offshore permits was approximately 289 million acres. This includes a recent filing by Imperial Oil on 22 million acres in deep water off the Newfoundland and Labrador Shelves (Fig. 22). By the middle of 1971 additional extensive filing on acreage in water depths up to 2,500 feet (750 m) have been made by oil companies on the Grand Banks, the Labrador Sea and Baffin Bay.

Continental Drift

A discussion of offshore areas would not be complete without brief references to the Continental Drift Theory and the implications concerning the nature, distribution and age of sediments on the continental shelf and slope.

Figure 23 is a modification of a diagram by Drake and Nafe (1968) in which the pre-drift alignment of the continents was made so that there is continuity of the Caledonian and Hercynian tectonic belts, and the quiet zones of Heirtzler and Hayes (1967) are in juxtaposition. Drake and Nafe propose that the quiet magnetic zones are underlain by original oceanic crust, and that the proto-ocean so formed together with marginal troughs was the site of deposition of Paleozoic or early Mesozoic sediment-filled troughs on the outer edge of the continental shelves and slopes, as deduced from changes in the velocity of assumed basement rocks. This hypothesis gives credence to the view that the thickness of sediments on the continental shelves may be greatly in excess of that estimated from preliminary geophysical surveys. It also indicates that restricted enivronments and marginal structures favourable for the generation and accumulation of hydrocarbons, have been in existence for a very long time.

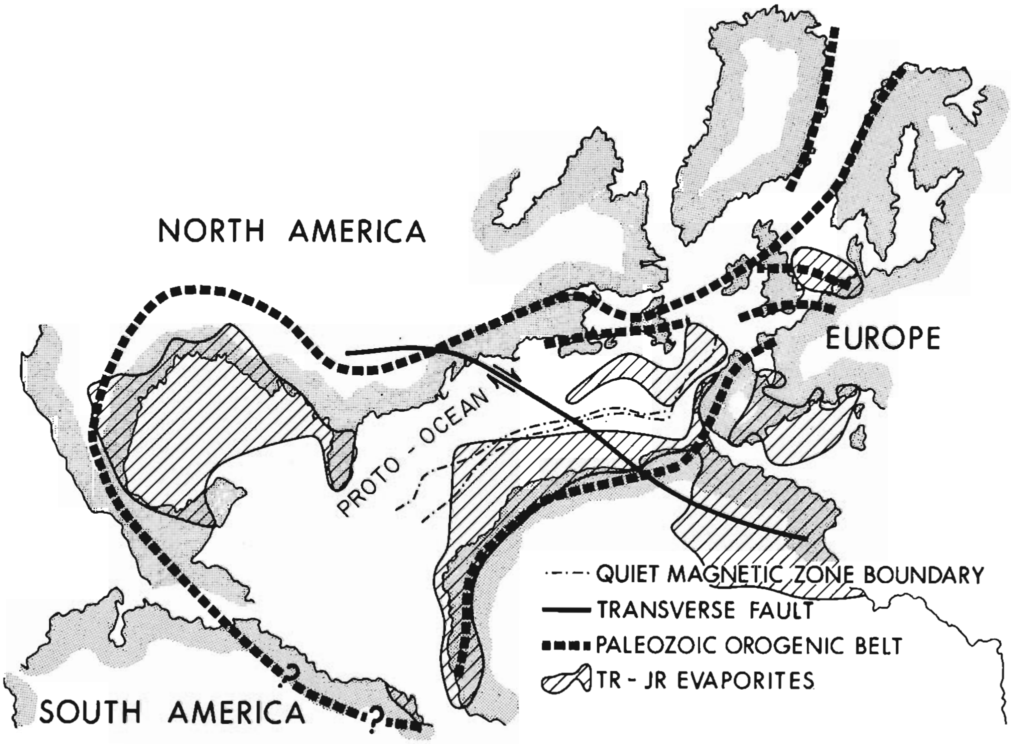

Figure 23. Pre-continental drift relationships modified from Drake and Nafe (1968).

ACKNOWLEDGMENTS

The authors are indebted to Teledyne Exploration Ltd., Calgary, Johnston Testers, Calgary and to the Geological Survey of Canada for assistance in the preparation of the figures.

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(Earth Science Symposium on Offshore Eastern Canada
Geol. Surv. Can., Paper 71-23, 1973, p. 109)

6. OFFSHORE EXPLORATION DRILLING AND PRODUCTION

W. W. Greentree, Shell Canada Ltd., Dartmouth, N. S.

Many companies have taken out exploration permits off the East Coast of Canada. Seismic surveys have been conducted on many permits and the second phase of activity, exploration drilling, has commenced. A variety of drilling vessels is available to drill offshore, however, few are suitable to conduct year-round drilling operations in this hostile environment. A significant discovery would promote the third phase, offshore production, and the subsequent development of resources. This talk outlined the equipment required to drill and produce offshore wells with emphasis on Shell Canada's current offshore program.

(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 111)

7. GEODYNAMICS PROJECT

C. L. Drake

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No abstract available

(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 113-151)

8. QUATERNARY SEDIMENTATION IN THE BAY OF FUNDY

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Abstract

The bulk of Fundy's sediment cover was emplaced under sub-aerial conditions during the Pleistocene low stands of the sea. Sub-bottom profiles reveal 0-30 m of outwash and till, and up to 100 m of finer sediment.

The suspended sediment dispersal system has the following components: 1) an oscillating turbid water mass, with an average turbidity of 6.6 mg/l; 2) bay-floor and margin provinces undergoing winnowing, which serve as fine sediment sources; 3) bay-floor and margin provinces undergoing fine sediment accretion, which serve as fine sediment reservoirs; 4) minor fresh, turbid water input; and 5) minor salt, turbid water output into the Gulf of Maine. Three different rates of sediment transfer may be detected. Exchange between the mud provinces and the overlying water masses occurs during each semi-diurnal tidal cycle. Some sediment undergoes long-term storage of fine sediment in reservoir provinces, with residence times on the order of several millenia. Ultimately fine sediment either escapes at the mouth of the bay (on the order of magnitude of 1.6×10^6 metric tons per year) or undergoes permanent burial on the bay floor. It appears possible to resolve the system into two sub-systems: the main system of bay floor provinces and associated turbid water masses, and a marginal system of intertidal mud flats and associated water mass. The latter is volumetrically less important, but its rates of sediment transfer are much higher.

The modern, high-energy tidal regime is reworking the Pleistocene deposits and is generating additional sediment by means of coastal erosion. These materials are being redistributed as the following textural provinces: 1) gravel provinces that consist of thin lags which cover most of the central and southeastern bay; 2) sand provinces which are tide-maintained, plano-convex lenses occurring mainly in the central and eastern portions of the bay; and 3) mud provinces which occur along the northwestern side. This distribution partly reflects the pattern of tidal currents, with stronger currents on the southeastern side, and a counterclockwise residual circulation.

A lithofacies analysis and an examination of sorting indicate transverse patterns of sediment distribution in the Bay, which may be due to a long-term effect of a standing wave. A ternary plot of the gravel-sand-mud constituents illustrates the relationship of sedimentary texture to hydrodynamic vigour as seen in the two dominant facies: (1) the traction load which is

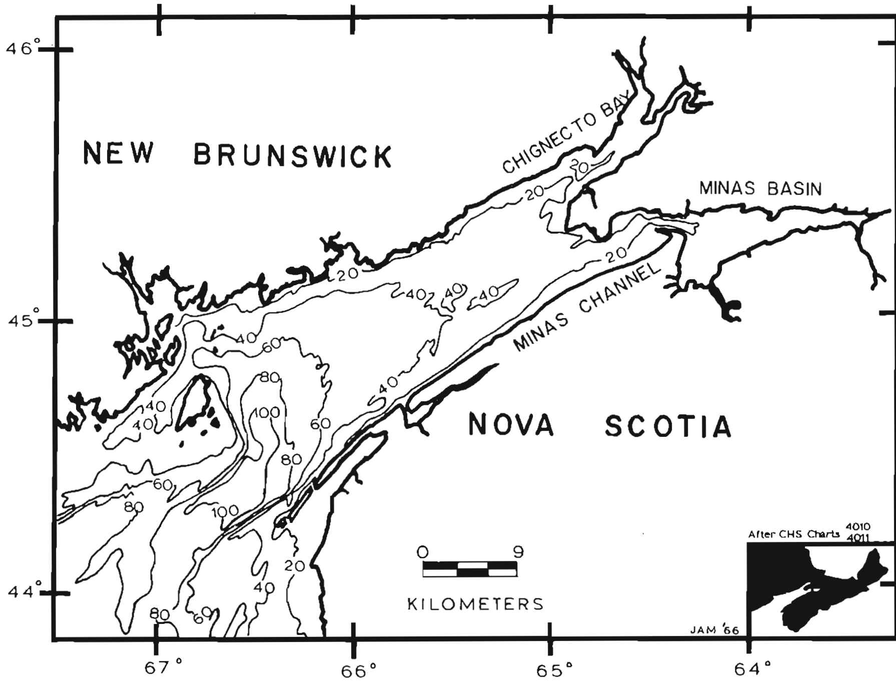


Figure 1. Location map of Bay of Fundy showing bathymetry in metres (modified after J.A. Miller, 1966).

associated with extreme tidal energy, and (2) the traction load associated with lesser tidal energy, and which contains an admixture of sediments settled from suspension.

INTRODUCTION

The Bay of Fundy (Fig. 1) is a funnel-shaped body of water lying between Nova Scotia and the Canadian mainland. The Bay of Fundy proper is 144 km long and 100 km wide at the base. The northeast end bifurcates into northeast-trending Chignecto Bay, and the east-trending Minas Basin. The bay has been incised into red continental mudstones and sandstones, and tholeiitic basalts of a Triassic half-graben (Swift and Lyall, 1968a, 1968b). Fundy was named Rio Fondo by 16th century Portugese navigators who were impressed by its enormous tides. Fundy has tidal currents of 0.75 to 2 m/sec. in a water column averaging 75 metres deep.

The bulk of Fundy's sediment cover was emplaced under sub-aerial conditions during the repeated glacial episodes of the Pleistocene, and these relict materials are evolving into sediments adjusted to the modern hydraulic regime. Where this process has run to completion, the resulting deposits bear the distinctive impress of a high energy tidal regime, a facies which has heretofore received little attention in the literature.

GEOLOGICAL AND PHYSIOGRAPHIC SETTING

Geology

The physiographic basin known as the Bay of Fundy is nearly coincident with the Acadian Structural Basin, a half-graben of probable Newark Age which contains friable red sedimentary clastics and basalt. It was first described in detail by Powers (1916). There is, however, an offset between the two basins, such that the border fault system lies 2 to 10 km offshore from the northwest shore, but on the southeast shore, a strip of Triassic rock 10 to 20 km wide is sub-aerially exposed. Southeast of the border fault, the Triassic rocks have subsided into a broad syncline, the Fundy Syncline. In addition to the border fault and syncline, the underlying rocks are deformed by a series of second-order faults and synclines.

The Triassic Acadian Basin is superimposed on the earlier Fundy Basin of Pennsylvanian Age (Bell, 1944; Belt, 1965), and Pennsylvanian clastic rocks outcrop beneath the northeastern sector of the bay. The bifurcation of the bay and of the Triassic Basin is inherited from the earlier basin. Uplands of lower Paleozoic metamorphic and granitic rocks flank the Fundy Basin on the northeast and southwest, and lie between the two upper bays. The bedrock geology beneath the bay is described in detail by Swift and Lyall, (1968a).

Physiography

A pocket of lower Cretaceous sediment filling a tributary valley indicates that the physiography of the Fundy region has been inherited from the Mesozoic (W. Take, Nova Scotia Museum, pers. comm.). However, mapping of the floor of the bay by sub-bottom profiling techniques indicates that the original erosional surface has been modified beyond recognition by Pleistocene glaciation (Swift and Lyall, 1968b). This glacial pavement has a gradient of 2 m/km and less. A periglacial river system is incised into its surface. Bedrock troughs at the mouth of the bay, and at the mouths of the two upper bays are also of glacial origin.

Shoreline

The shoreline consists primarily of an intertidal, wave-cut, bedrock terrace up to 300 metres in width (Klein, 1963). The terrace is veneered with less than a metre of coarse, variable, poorly sorted detritus of mainly local origin (frequently a muddy pebbly sand), and it is fronted by a sea cliff. These cliffs are receding at rates probably measured in centimetres or decimetres per year. Where composed of weak basal Triassic sandstone, the rate of retreat of promontories may be 2 to 3 metres per year (Churchill, 1924). Bedrock lows in the shoreline are occupied by masses of glacial drift of which outwash gravel is an important component (Borns and Swift, 1966; Swift and Borns, 1967a, 1967b; Swift, 1968). Here sand and gravel bluffs replace the bedrock sea cliff, and the supply rate of coarse detritus is sufficiently high to produce a complex series of inertidal shingle spits and bars. Locally, the cliff line is cut by rivers that incised valleys 60 metres below present sea level. These valleys were filled with gravel deltas during the Wisconsin ice retreat, and modern tidal currents now scour the upper surfaces of these relict deltas (Swift and Borns, 1967a; Ali, 1964; Ali and Laming, 1966).

In protected coves and estuaries, more extensive constructional intertidal deposits occur, including tidal mud flats backed by tidal marshes (Swift, 1968). Where the rate of supply of sand is sufficiently high, and hydraulic conditions are suitable, extensive intertidal sand bodies have developed (McMullen and Swift, 1966; Swift et al., 1966; Swift et al., 1967; Swift and McMullen, 1968).

The changing shape of the Bay of Fundy shoreline is partly related to submergence. Grant (1970) reports submergence of approximately 30 cm/century for at least the last 4,000 years, most of which is anomalous although some (6 cm/century) is eustatic. He states from geological evidence that a rise in high tide, or an increase in tidal range is partly responsible for the anomalous submergence, together with the effects of the additional water load and epeiric downwarping. Physiography, bedrock and soil erosion, and sedimentation along the shore must be affected by such large-scale action. Ultimately the shoreline must change.

PHYSICAL OCEANOGRAPHY

Tides

The tides of Fundy are semi-diurnal and anomalistic (markedly higher during the lunar perigee than during the lunar apogee), and are famous for their enormous range. The entering tide at the mouth has a mean range of 6.4 metres (Grand Passage, Nova Scotia) and the range increases to 13 metres at the head of the main Bay (Halls Harbour). The enormous tidal range is a consequence of the dimensions of the Bay, the latter of which averages 75 m in depth and 300 km in length. From the formula by King (1963, p. 74) the critical length for a standing oscillation with a node at the mouth of the Bay and maximum amplitude at the head is 296 km, which closely agrees with the measured length of 300 km; also, the natural period of the bay is 6.29 hours which is almost one-half of the semi-diurnal period, 6.21 hours. Therefore, the conditions for resonance are approximated, and the entering tidal wave is amplified to the point where constructive interference is balanced by tidal friction. The tidal wave in Fundy is therefore a standing wave rather than the progressive wave found in some other embayments. Consequently, high water is attained nearly simultaneously throughout the Bay, being only 24 minutes later at the bifurcation than at the mouth of the Bay. After analyzing the tide in Fundy, as a damped cooscillating system, Harleman (1966) notes that it is "not highly resonant" in that the amplification of the tide from the ocean entrance to the head of the bay is only 2 1/2 times. He points out that the entrance tide at Cutler, Maine is 4.3 metres, and "has already undergone a primary amplification due to the shoaling effect of the continental shelf at the seaward edge of the Gulf of Maine". Shoaling is a relatively minor contributor to amplification of the tidal wave within the Bay, adding .3 to .6 inches to the diurnal spring tides. The main reflecting areas of the tidal wave appear to be Cape Hopewell in Chignecto Bay, and Scotts Bay, in the Minas Channel.

Fundy's tidal wave moves 96 km^3 of water through its mouth twice a day (Bowden, 1962). The resulting tidal currents are parallel to the bay's axis (Fig. 2). At half flood and mid-depth they average 103 cm/sec (2 knots) on the south shore and 77 cm/sec (1.5 knots) on the north shore. The difference

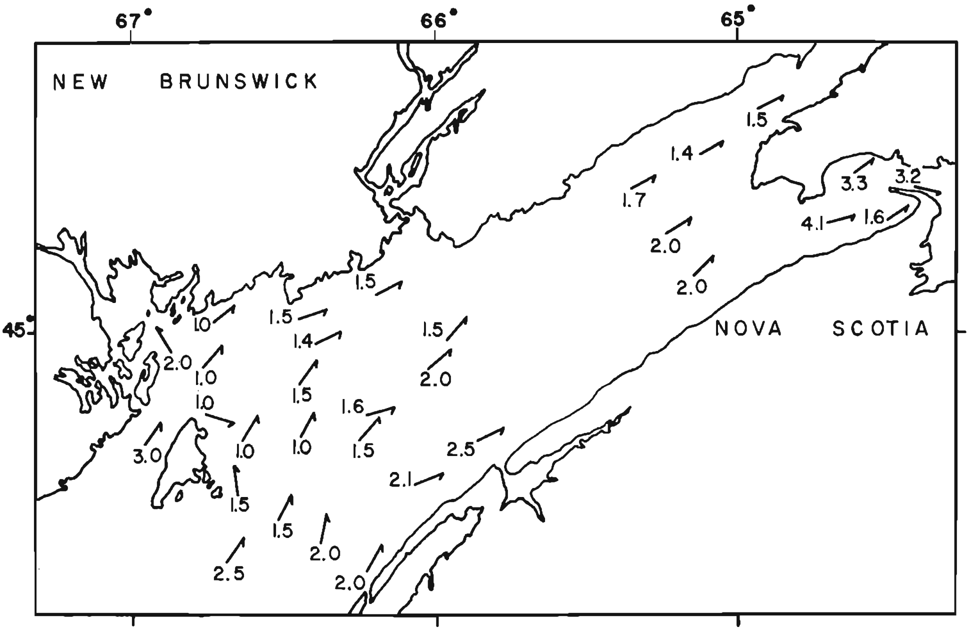


Figure 2. Tidal current velocities (in knots) at mid-depth and half flood tide (from Anderson, 1928; Forrester, 1958; and Bedford Institute of Oceanography Data Series 66-2-D).

in velocities is due to the Coriolis effect on the tidal currents, which tend to bank up on the south shore when flooding, and on the north shore when ebbing. Towards the head of the bay, shoaling and narrowing retard the tidal wave, and current velocities increase to 206 cm/sec (4.0 knots) in the Minas Channel and up to 556 cm/sec (11.0 knots) in the Minas Passage (Cameron, 1961). During spring tides, currents run 30 to 50 per cent faster (Farquharson, A. O. L., pers. comm.). The currents maintain an essentially constant speed down to the viscous boundary layer, and at most places the tide turns at the surface and below at practically the same time (Dawson, 1908).

Waves and Ice

Winds in the Bay of Fundy blow mainly from the south and southwest, and during the rest of the year, from the west and northwest. Therefore, Fundy's coasts are generally exposed. Winter waves are over 0.3 m high 90 per cent of the time, over 1 m 50 per cent of the time, and over 4 m 10 per cent of the time. Waves of all heights are 5 to 10 per cent less frequent in the summer.

Drifting ice in the spring plays an active role in the transportation of sediments in the Bay of Fundy. The ice collects its load by direct incorporation of muddy water, and by grounding on the tide flats where it acquires sand and gravel additionally. Hind (1875) collected and melted ice samples in Windsor Bay in order to determine their sediment load. He concluded that ice with a load acquired by grounding carried an average 4.54 g/m^3 and ice with a load acquired by direct accretion carried 2.38 gm/m^3.

Run off

The St. John River (Table I), draining 55.655 km^2 of New Brunswick contributes 70 per cent of the water entering the Bay of Fundy.

TABLE I

Discharge rates of rivers tributary to the Bay of Fundy (Watson, 1936)

Rivers	Area	Yearly mean discharge rate
St. Croix	3,885 km^2	65.9 m^3/sec
Magaguadavic	2,719	59.1
St. John	55,685	936.7
North Shore Rivers	2,486	64.9
Chignecto Bay Rivers	5,594	135.7
Minas Basin Rivers	8,547	238.9
Annapolis	2,549	66.9
Bay of Fundy	15,540	421.2
Total	97,011	1,989.0

Discharge is bimodal with a maximum of 1218 m^3/sec during the November rains, and a second maximum of 2856 m^3/sec during the May meltwater floods. During periods of minimal discharge, the rate is approximately 532 m^3/sec.

Water Structure and Circulation

Bailey et al. (1953) have summed 29 years of observations on the western corner of Fundy, near Passamaquoddy Bay. They note that during this period, mean surface temperatures vary from 1.2°C in March to 11.5°C in September, and surface salinity varies from 31.1‰ in May to 32.9‰ in October. Both saline and thermal stratification develop throughout the spring and summer. During the May floods the vertical salinity gradient reaches a maximum of 1.0‰ over 90 m of water. During this period the brackish effluent of the St. John River can easily be traced westerly from St. John past Bailey's station (Fig. 3). By September the stratification in this corner of the bay is mainly thermal with a vertical gradient of 1°C over 90 m.

Our survey of the bay in May of 1965 showed that for each depth, isohaline and isothermal lines are nearly parallel to the long axis of the bay, with the warmest, saltiest water occurring along the southeast shore. Here salinities are 32.0 to 33.8‰. Surface temperatures were variable, but below 20 m, this southwestern water mass was 5.4°C. The coldest, freshest water appeared in the northeast sector of the bay with values as low as 22.0‰ and 2.8°C respectively. Stratification gradients up to 4‰ and 3°C over 40 m occurred along the north shore (Fig. 4). Most of the upper bay was stratified to the extent of a single temperature and salinity unit or less over this depth, while the lower bay was homogeneous.

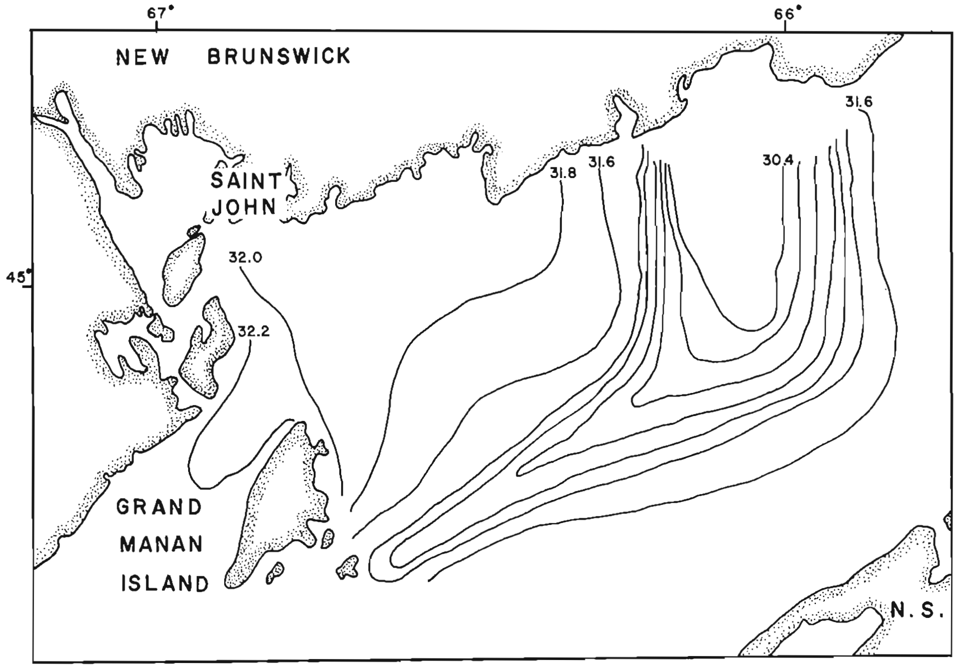
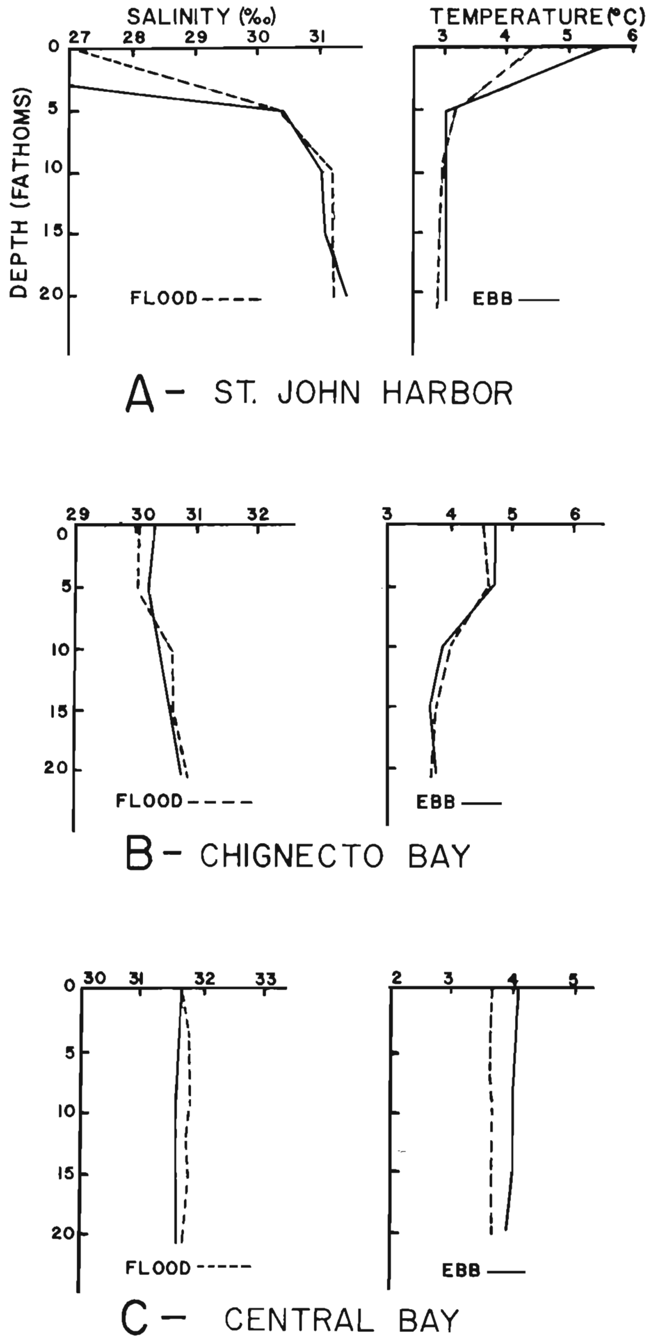


Figure 3. Horizontal distribution of salinity at the surface during the spring freshet of April 17 – May 5, 1930 (from Hachey and Bailey, 1952). Isohalines are in parts per thousand.

Density profiles taken by Watson in June, 1936 (Fig. 5) yield further detail on the structure of Fundy's water. The indication of tidal mixing in a turbulent boundary layer is particularly interesting in this diagram. The density isopleths, which are normally horizontal, turn downwards as they occur within 20 to 50 m of a shoaling bottom. Where the bottom rises steeply, the area of mixing is small, and there is much less effect on the isopleths. Toward the head of the bay, the bottom has shoaled to an extent that the boundary layer emerges on the surface, and the isopleths are vertical. Aerial photographs in this sector reveal turbulence clearly outlined as vertically rising "boils" of more turbid water separated by partitions of less turbid water (see Fig. 14).

The density structure of Fundy's water reflects a counterclockwise pattern of residual currents (residual after subtraction of diurnal and semidiurnal tidal components). This circulation pattern (Fig. 6), was determined from drift-bottle studies (Mavor, 1922a; Bumpus, 1959; Bumpus and Lauzier, 1965), and from current metres (Bedford Institute of Oceanography Data Series 66-2-0). This latter reference cites residual current velocities of 5.2 to 10.3 cms/sec (0.1 to 0.2 knots) at most points. The greater flood tide on the south shore and greater ebb on the north shore result in a counterclockwise (cyclonic) residual current gyre.

Watson (1936) stresses thermohaline gradients as a factor in oceanic circulation. Since the St. John River effluent is diverted southward, the salt water that it consumes by turbulent mixing is partially supplied by a saline



A. southwest of St. John Harbour,

B. Chignecto and

C. Central Bay area. A and B are affected by the spring runoff of the St. John and Petitcodiac rivers, respectively.

Figure 4. Temperature and salinity profiles.

underflow at the mouth of the bay (Fig. 5). On the south shore, mixing is a result of tidal turbulence in which the mixed water moves across the bay under the impetus of the density gradient caused by the salt-water consumption of the St. John effluent (see Watson, 1936). Thus, the thermohaline circulation of Fundy consists of an influx of salty water at both sides of the mouth at depths over 100 m, surface outflow on the north, surface inflow on the south, and a lateral (northwest to southeast) transfer at mid depth. This thermohaline circulation interacts with and reinforces the coriolis circulation.

The effect of wind on the circulation pattern is less well known. Hachey (1934a; see also Hachey, 1934b and 1934c) suggested that the southwest winds of summer will tend to keep surface water from leaving the Bay, and hence increase its stratification. Lauzier (Fisheries Research Board, St. Andrews, N.B., pers. comm.) suggested that the predominant westerly

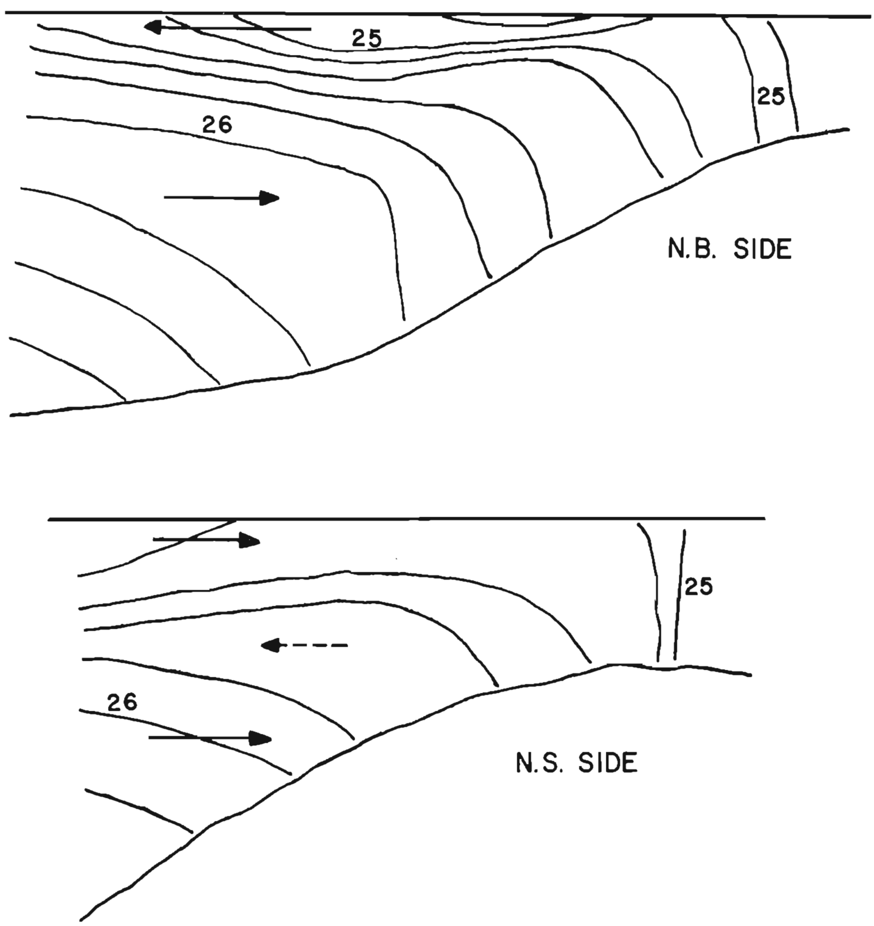


Figure 5. Longitudinal density sections, Bay of Fundy, June 15 to 18, 1932 (from Watson, 1936). Arrows represent hypothesized water movements as explained in text.

winds cause upwelling to occur along the New Brunswick coast, with movement of the deeper water from south to north across the bay. Watson (1936) concluded that wind effects may predominate at the head of the bay, and perhaps in the Grand Manan Channel, but that the main circulation of the bay is due to Coriolis forces.

SUSPENDED SEDIMENT

Methods

During the spring and summer of 1965, a detailed study of the suspended sediment system was undertaken by Miller (1966), and water samples were collected from 43 stations (Fig. 7). Each station was sampled at half-flood and half-ebb, at the bottom, 1 m from the bottom, 10 m from the bottom, and at the surface. The samples were collected in 6-litre Van Dorn bottles and stored in plastic bottles to await filtration. After resuspension, one litre of water from each sample was filtered through a pre-weighed, 0.22-micron Millipore filter, in which was dried and weighed, to determine the concentration of suspended material.

The amount of particulate organic carbon in the bottom water was determined by processing the material filtered and refrigerated while at sea. The determinations were made with an infrared carbon dioxide analyser

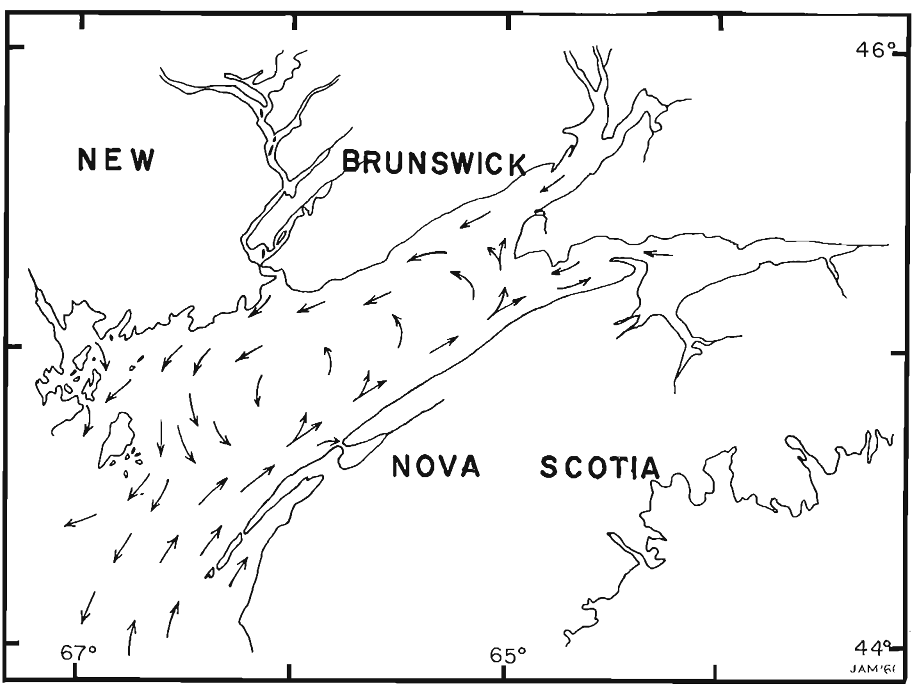


Figure 6. Residual circulation pattern, Bay of Fundy, June 15 to 18, 1932 (<u>see</u> text for sources of data).

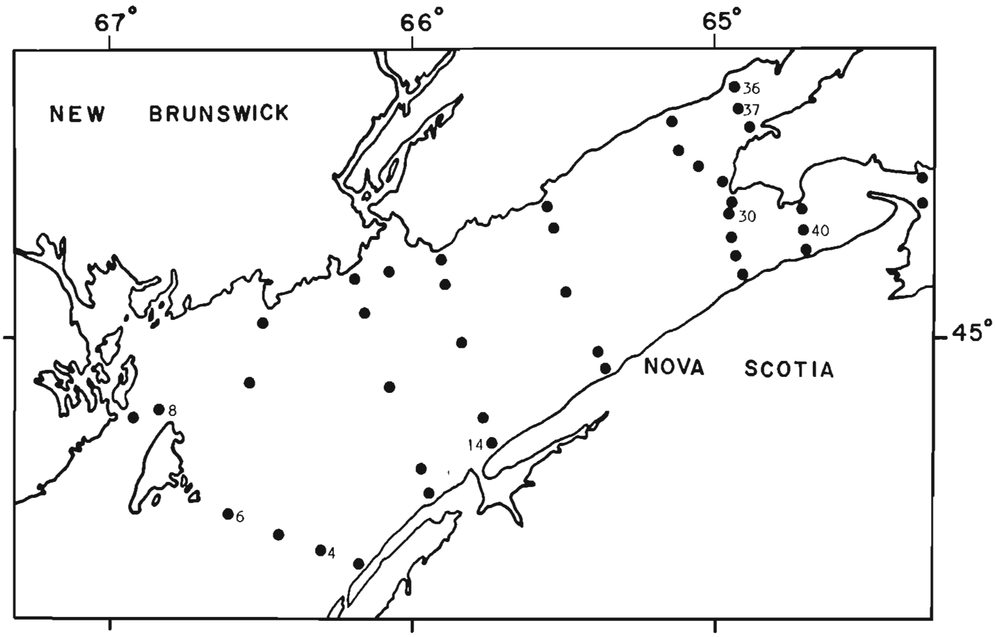


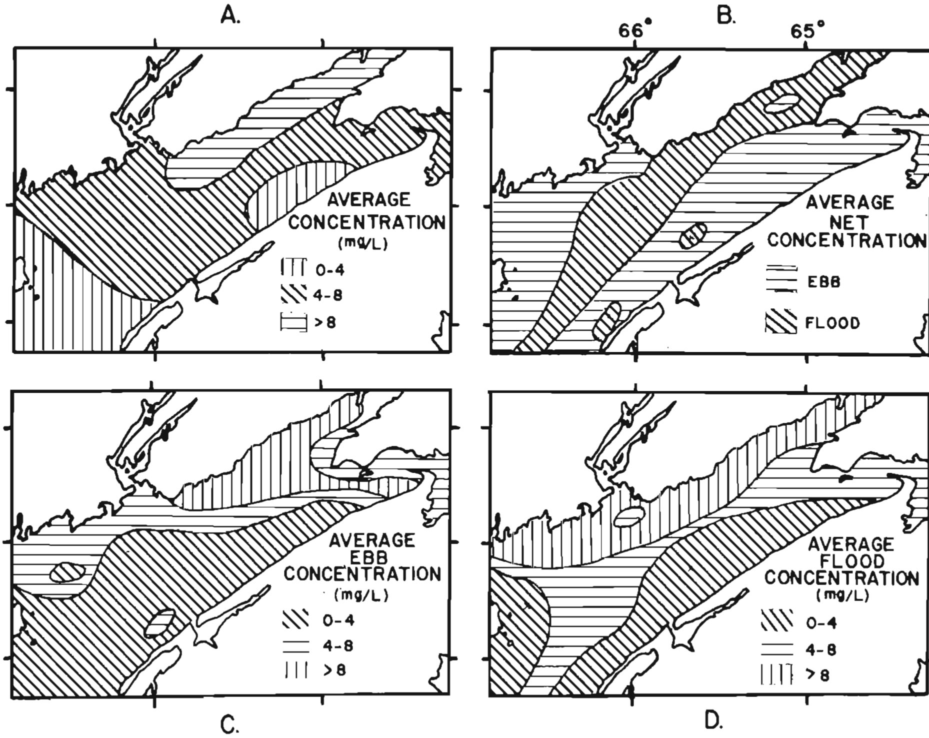
Figure 7. Portion of sample not for suspended sediment study.

(Miller, 1966, p. 25). X-ray diffraction was used to determine the major mineral suites present in the suspended materials, and to assess their relative abundance. Methods of identification are described by Miller (1966, p. 27). After weighing, to determine suspended sediment concentration, the Millipore filters were examined under a binocular microscope in order to estimate grain size, and identify organic materials.

Suspended Sediment Concentration

Measured suspended sediment concentrations varied from 0.2 to 30.4 mg/l, with an average value of 6.6 mg/l for 263 samples. In Figure 8a, the average suspended sediment concentration for the entire water column throughout the tidal cycle is presented. In this illustration, turbidities in excess of 8 mg/l occur in a strip along the New Brunswick coast north of St. John. Figures 8c and 8d contrast half-ebb and half-flood turbidities averaged through the water column. Here the flood tide generates a zone of high turbidity along the entire New Brunswick coast; at half-ebb this zone has contracted to the position shown in the generalized map of Figure 8c.

This pattern of turbidity through time and space has the following probable causes: 1) The turbidity is generally higher on the New Brunswick coast because it is primarily the result of the resuspension of bottom materials, and the floor of this sector, due to the residual current pattern, is a depot for fine sediment generated throughout the bay; and 2) The turbidity on the New Brunswick side is higher at half-flood than at half-ebb, despite weaker currents then, because: (a) the sector has just passed through a half

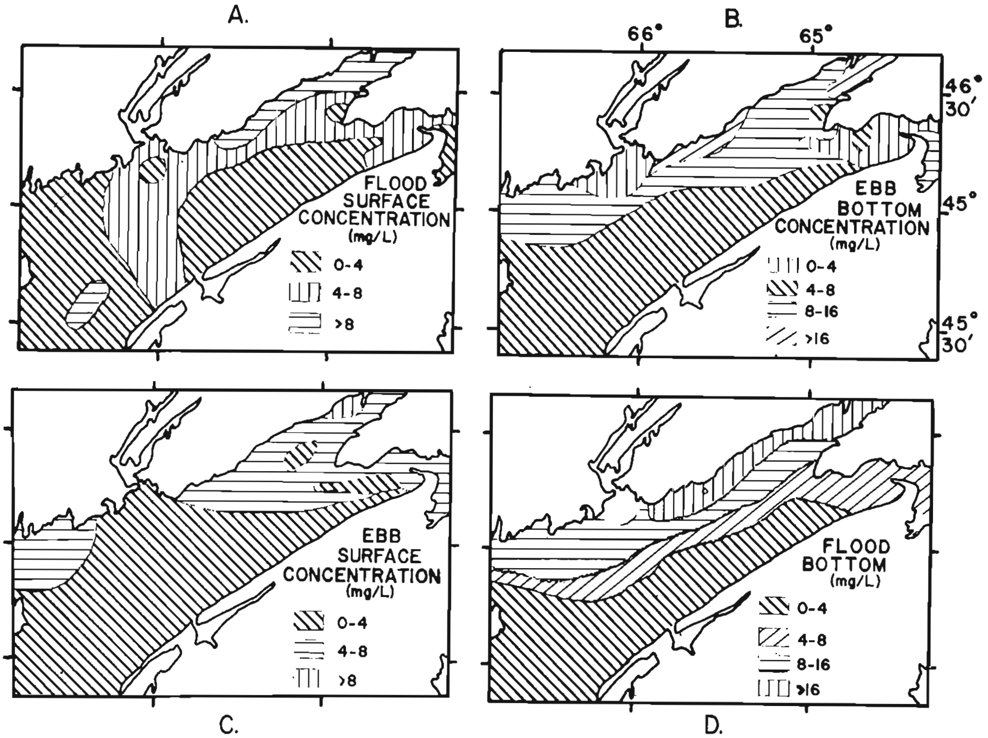


A. Suspended sediment concentration for entire water column.
B. Net suspended sediment concentration for whole water column.
C. Half ebb suspended concentration for whole water column.
D. Half flood suspended concentration for whole water column.

Figure 8. Suspended sediment concentration in Bay of Fundy.

tidal cycle of extremely shoal water, when oscillatory wave surge can multiply the erosive effect of the tidal stream; (b) as the tide returns, waning wave surge is compensated by the intensifying tidal stream. The time-velocity asymmetry of the tidal wave near shore results in strongest flood currents during the first half of the flood; (c) during ebb tide, the nearshore zone receives the drainage of the intertidal zone. The velocity of this drainage is uniformly high through ebb tide, since it is maintained by its own hydraulic head, and the water is highly turbid. This turbidity becomes entrained by the returning flood stream.

Surface and bottom turbidities at half-ebb and half-flood are compared in Figure 9. Generally bottom turbidities are higher than surface turbidities, which indicates that most of the sediment is locally resuspended. Figure 10 illustrates vertical turbidity profiles at half-ebb and half-flood. Profiles from the upper bay show marked gradients. Profiles from the lower bay (stations 4, 6 and 8) show negligible gradients. This suggests that turbidity is far travelled, and that the water column has either been homogenized, or has lost its coarser particles prior to arrival at the station or both.



A. Surface concentration of suspended sediment, half flood.
B. Bottom concentration of suspended sediment, half ebb.
C. Surface concentration of suspended sediment, half ebb.
D. Bottom concentration of suspended sediment, half flood.

Figure 9. Surface and bottom concentration of suspended sediments, Bay of Fundy.

At station 20 it seems a far-travelled turbid water mass appears to occur in the upper portion of the water column, sufficiently extensive to cause a reversed gradient throughout the water column.

Figure 8 shows net concentration of turbidity, and indicates which phase of the tide (flood or ebb) is more important in the suspension of particular matter. The distribution is obtained by comparing the weight-per-volume values of the flood and ebb tide samples, and plotting the higher concentration in terms of flood or ebb tide. As previously indicated, the half-flood tide stations are the most turbid on the northwest side of the bay, except for the sheltered sector south of St. Johns. Southwest of St. John Harbour, and on the southeast side of the bay, the ebb tide is the most turbid, probably because during this period each station receives from the north and the adjacent shoreline more turbid water than can be generated locally.

Composition of Suspended Sediment

Detrital materials comprise more than 75% by volume (visual estimation) of suspended sediments in the Bay of Fundy. Silt-sized and finer particles were the most abundant sizes, but sand grains occurred in 60% of the samples. The sand was mainly fine to very fine with the coarser particles found in the near-bottom samples. Suspended sand was most abundant on the southeast side of the bay (Fig. 11a), reflecting the character of the bottom. The mineral species characteristic of sand classes were also most abundant on this side, including feldspar and calcite (Fig. 11b, Table II).

TABLE II

Average Mineralogical Composition of Suspended Sediments

	Corrected Values (Scattering Factor of Johns, Grim and Bradley, 1954)	Uncorrected Values
Quartz	3.2%	11.7%
Feldspar	1.2%	4.8%
Calcite	0.5%	2.9%
Illite	71.1%	49.7%
Chlorite	2.1%	17.0%
Kaolinite	3.6%	7.5%
Halloysite	18.0%	6.3%

Clay minerals present include illite, halloysite, kaolinite, quartz, and chlorite, in order of abundance.

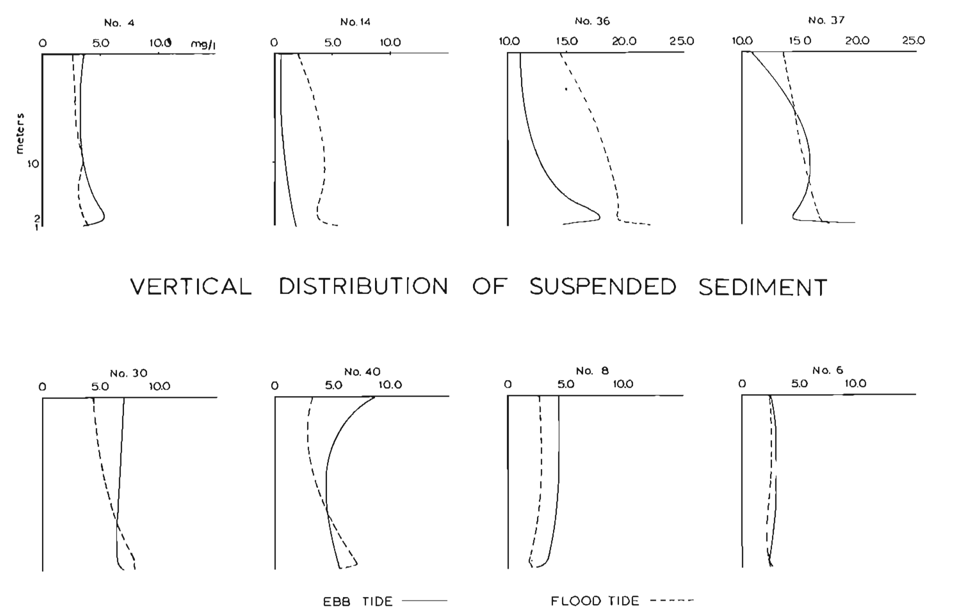
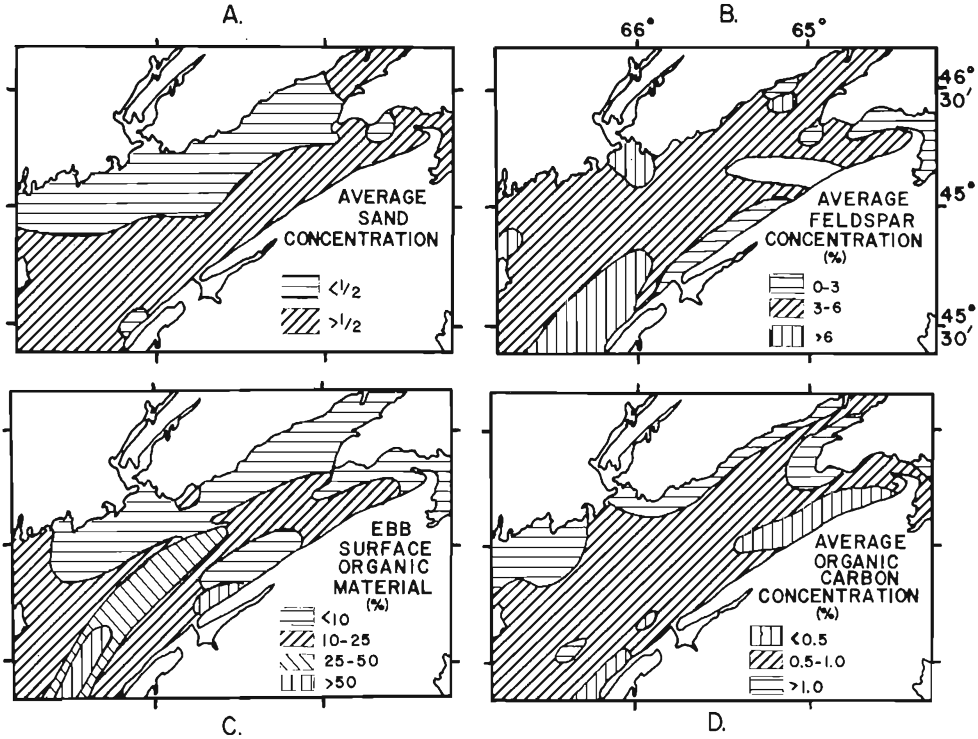


Figure 10. Turbidity profiles from selected stations in the Bay of Fundy.



A. Concentration of suspended sand averaged through the tidal cycle and through the water column.
B. Average concentration of suspended feldspar.
C. Per cent suspended organic matter, surface, half ebb.
D. Organic carbon, per cent of total sediment.

Figure 11. Nature and concentration of various suspended sediments in Bay of Fundy.

The volume per cent of organic detritus was determined by visual estimate. This category includes diatoms, foraminifers, large algae, molluscs, and material that appeared on filters as "organic film", a fine, yellow-green material that could not be resolved under the microscope at X50. The material is more abundant in the upper part of the water column (average surface value of 18% versus average bottom value of 9%). The highest bottom concentrations (up to 35 per cent) are found along the southeast side; the highest surface concentrations (up to 80%) are at the mouth of the bay (Fig. 11c).

Organic carbon analysis revealed particulate organic carbon values of 0.05 - 0.74 mg/l which made up 0.26 - 2.65 per cent of the total material in suspension. The distribution (Fig. 11d) is the inverse of that determined by visual estimation by volume per cent, with highest concentrations along the northwest side of the bay. The discrepancy is due to the masking effect of the large amounts of inorganic suspended material on the northwest side, and to the low density of the material. Per cent of whole foraminifers and diatoms by visual estimation does coincide with the distribution of organic carbon.

THE SUSPENDED SEDIMENT SYSTEM

The suspended sediment system in the Bay of Fundy is an open system. It consists of four components, 1) an oscillating body of turbid water, 2) a substrate that exchanges water with the overlying water mass, 3) a minor turbid fresh-water input, and 4) a minor turbid salt-water output into the Gulf of Maine.

Sediment transfer through this system is occurring at three different rates. The mud provinces of the northwest side exchange fine sediment with the overlying water mass during every tidal cycle, and appear to be in a state of dynamic equilibrium with it. Grab sampling at slack, low and high water within these provinces has locally revealed the presence of a layer of highly fluid mud a few centimetres thick, that appears to have settled out at that time. This material is resuspended as the tide begins to flow.

On the southeast side of Fundy the slack-tide mud layer has not been observed; the bottom is probably sufficiently rough for the vestigal rotary current to generate a weak turbulence during this period. There is, however, a long-term transfer of fine sediment winnowed from the Quaternary drift and incorporated into the suspended load. Probably much of this material is slowly restored to the bay floor, on the aggrading mud facies of the northwest side. However, the area in which such aggradation is possible may be shrinking before the advance of the "transition" sand provinces.

There is evidence to indicate that a portion of the suspended sediment load is escaping into the Gulf of Maine. Forgeron (1962) plotted the distribution of bottom sediment colour for most of the Bay of Fundy and Gulf of Maine. His map shows that the colour 10YR4/2 may be traced from the northwest side of the Bay of Fundy into the central portion of the Gulf of Maine, suggesting a pathway of suspended sediment movement. Residual current measurements (Forrester, 1958) indicated the presence of an outward movement of 0.005 - 0.01 knots around Grand Manan Island. Drift bottle data gathered by Chevrier (1959) show that there is a net movement of surface water out of the northwest corner of the bay and into the Gulf of Maine. Lauzier (pers. comm.) carried out circulation studies in this region using bottom drifters, which indicated a residual movement out of the bay that ranged between 0.1 and 1.0 nautical miles per day. If it is assumed that residual outward transport occurs throughout the Grand Manan Channel and west of the island as far as the 100 metre contour, that the average rate of transport is .037 km/hr, and that the average suspended sediment concentration is 2.7 mg/l, then the yearly suspended sediment discharge into the Gulf of Maine is on the order of 1.6×10^6 metric tons.

During the course of our investigation, we were able to detect, but were not able to resolve clearly, a nearshore subsystem of suspended sediment transport, consisting of fine sediment exchange between intertidal mud flats and their associated water masses. Where fully developed, the intertidal mud flats are prisms of silty clay and clayey silt with flat upper surfaces (salt marsh) at mean high tide, gently sloping forward surfaces of bare mudflat. The prism rests on an inclined till surface (see Fig. 12 and 13). In deep coves these sediment masses may be quite voluminous, with surface areas up to 25 sq. km, and thicknesses in their centres in excess of 15 m, as indicated by sub-bottom profiles. In open coasts they are absent, or consist of ephemeral lenses 60 cm or less in thickness, just below the high-tide sand or shingle beach, resting on the sand and gravel mantle of the main tide

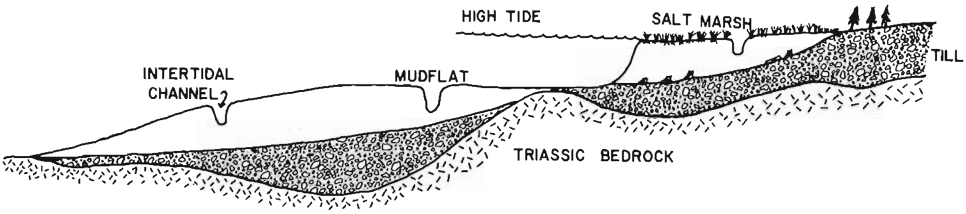


Figure 12. Schematic section through a Bay of Fundy sheltered intertidal zone.

Figure 13a. Salt marsh transgressing (spruce) forest, Grand Pre, Nova Scotia on south side of Minas Basin.

Figure 13b. Rapidly retreating marsh face, same locality.

Figure 13. Illustrations of sheltered intertidal zones.

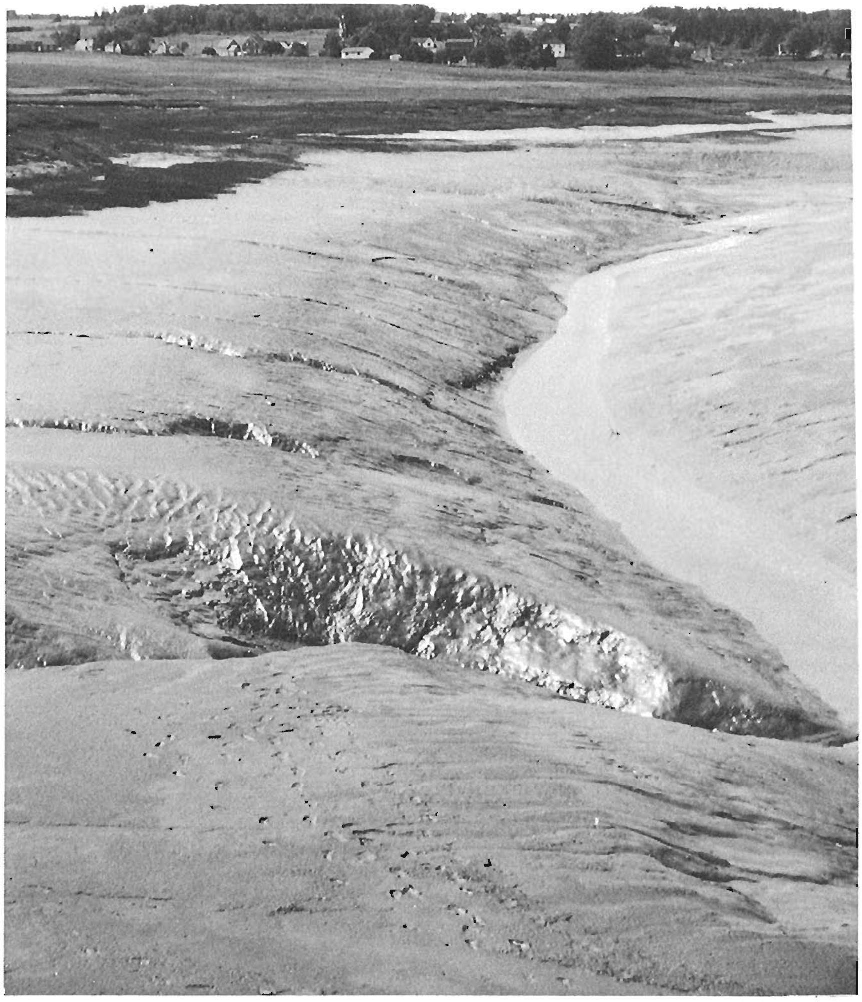


Figure 13c. Tidal creek in mudflat, Noel Bay, Nova Scotia, on south side of Minas Basin. Here banks are failing by plastic flow.

flat. The total area of such intertidal mud flats is probably on the order of 500 sq. km, or about one tenth of the total area of the submarine mud provinces. The rate of exchange of sediment with the overlying water masses however, is higher than for the bay floor provinces. The intertidal mud flats are subject to intense wave activity through half of the tidal cycle. More importantly, they are incised by meandering tidal channels up to 45 metres deep with rates of lateral migration of tens of hundreds of centimetres per week. In addition to short-term exchange by these mechanisms the intertidal mudflats are subjected to a long-term exchange of sediment with the bay water due to rising sea level. As the marsh surface rises and advances over

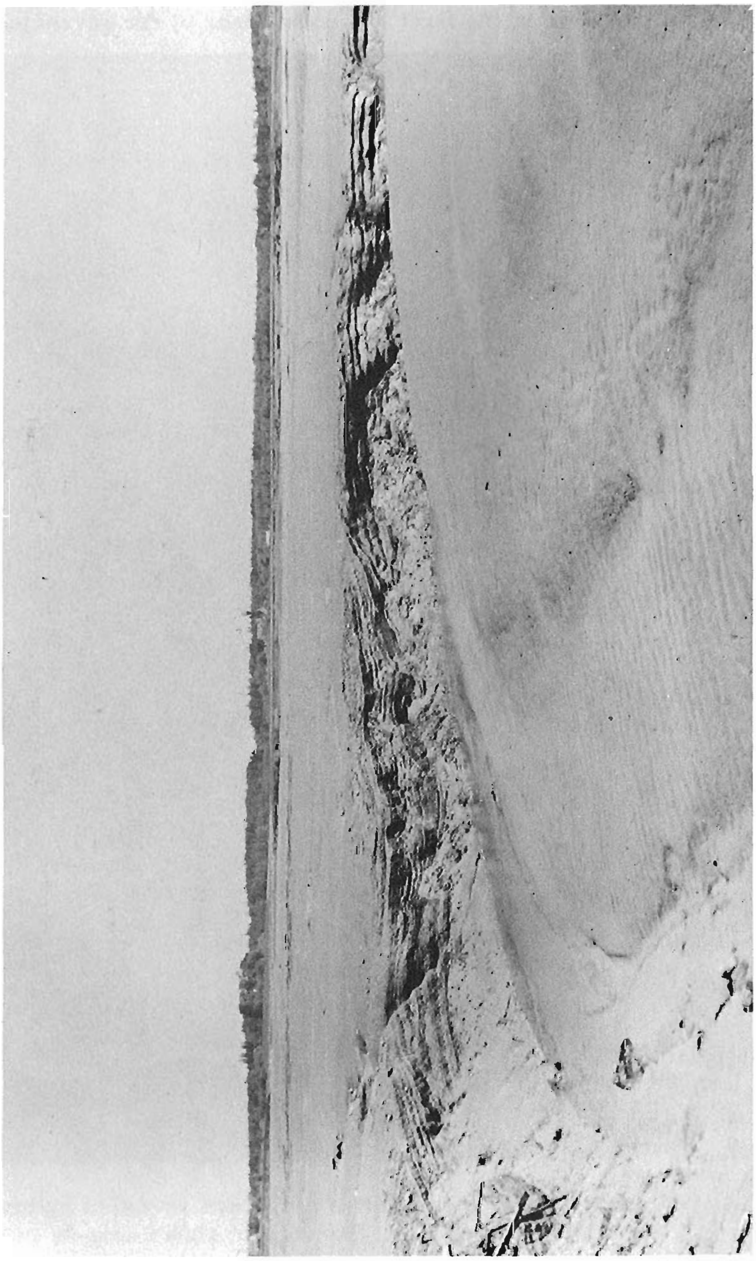


Figure 13d. Meandering tidal creek, Noel Bay.

the subaerial surface, the marsh front and the mud flat below it retreats, often re-exposing the stumps of spruce forests drowned several millenia before (Harrison and Lyon, 1963). Thus the maximum residence time of a clay particle in such a mud flat can be no more than a few millenia.

Nearshore turbidity tends to rise exponentially to nearshore values as high as a gram per litre in the first 15 centimetres of the advancing flood

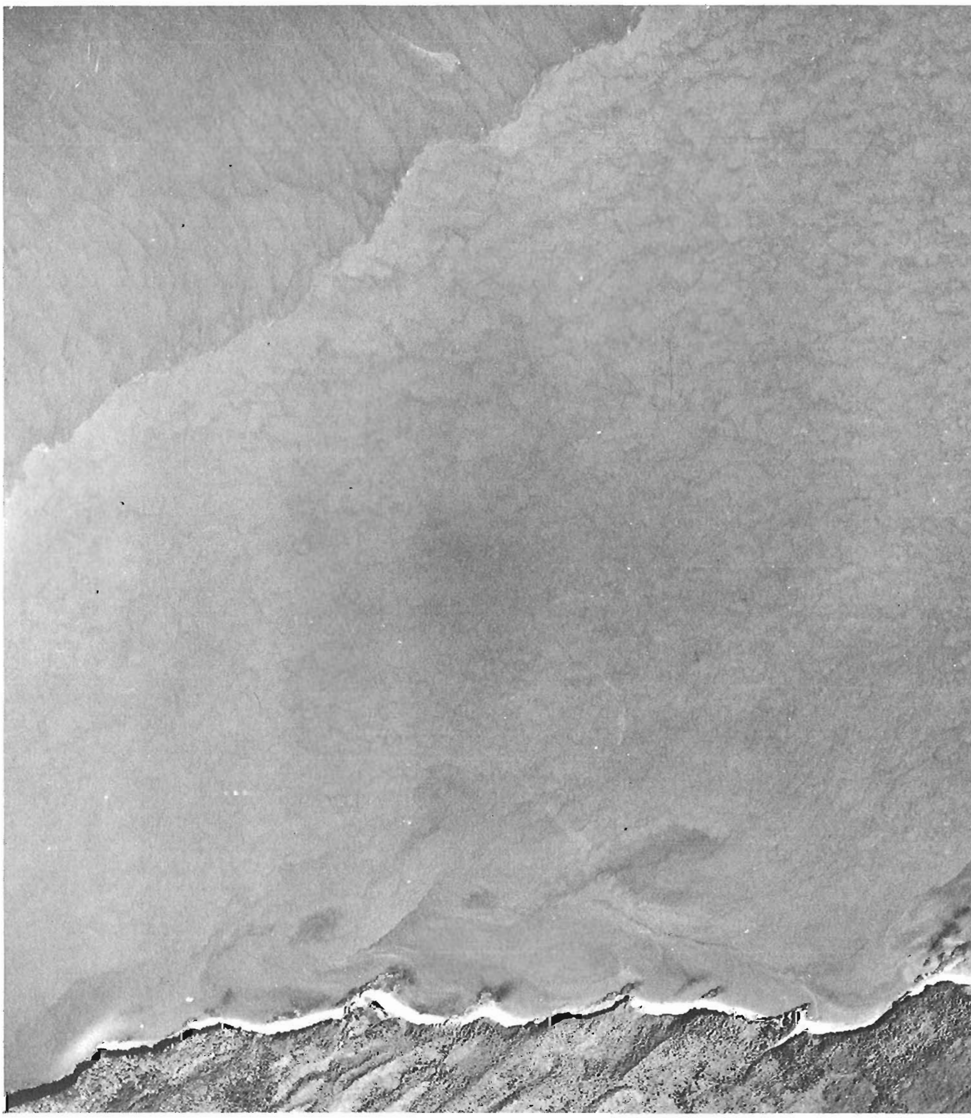


Figure 14. Aerial photograph showing water structure revealed by turbidity, south side of Chignecto Bay. Prominent shear zone is generated by separation of boundary, layer as tidal stream passes Cape Chignecto. Top to bottom overturn in both water masses outlined by turbidity. Note countercurrents near shore.

tide. The nearshore water mass that exchanges sediments with the tide flat is commonly a distinct entity for some portion of the tidal cycle which is isolated from the main tidal stream by the separation of a vertical boundary layer generated by a nearby cape (Fig. 14). Elsewhere, eddies which develop between nearshore countercurrents and the main offshore tidal stream (Fig. 15) cause significant amounts of the nearshore turbid water to be entrained into the offshore water mass. While the volume of sediment within

Figure 15. Aerial photograph showing nearshore countercurrents causing entrainment of nearshore, highly turbid water by main tidal stream, south shore, Chignecto Bay.

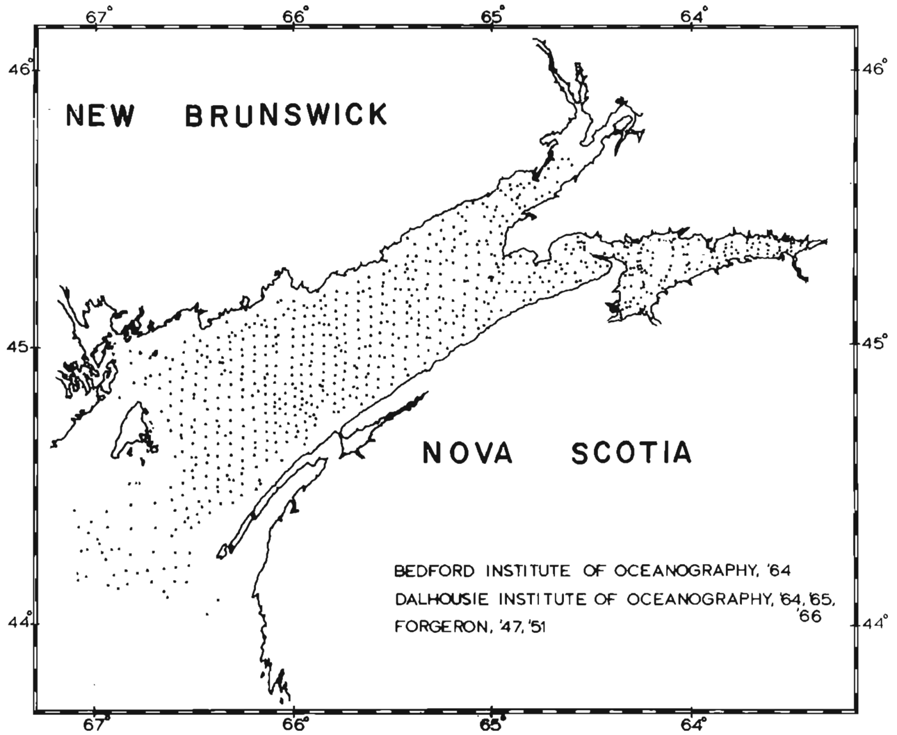


Figure 16. Grab-sample net, Bay of Fundy.

this nearshore subsystem of fine sediment transport is at any one time a small portion of the total system, its rate of sediment exchange is probably sufficiently high to render it a significant part of the total system.

BOTTOM SEDIMENT

General

Bottom sampling of the Bay of Fundy was carried out by the Bedford Institute of Oceanography and the Dalhousie Institute of Oceanography during the period 1961 to 1966 (Fig. 16). Samples were collected on a 2-km sample grid within the Bay of Fundy proper, using a .2m^3 Van Veen grab sampler. The sand and silt-clay fractions were subjected to sieving and pipette analysis, respectively.

Sub-bottom profiling was carried out as a joint operation of Bedford and Dalhousie Institutes of Oceanography during the spring of 1966 (Fig. 17), and was the primary responsibility of the third author. The profiles were made with a Huntec Mark 2A Hydrosonde profiling system using a 165 Joule sparker. The system was programmed to provide four sparks per second with a sweep time of 250 milliseconds. Filters were set at 152 to 2329 cps, a bandpass which avoided interference with the ship's electrical system. The spark electrode was towed at about 5 knots behind the CNAV SACKVILLE, a Canadian naval auxiliary corvette.

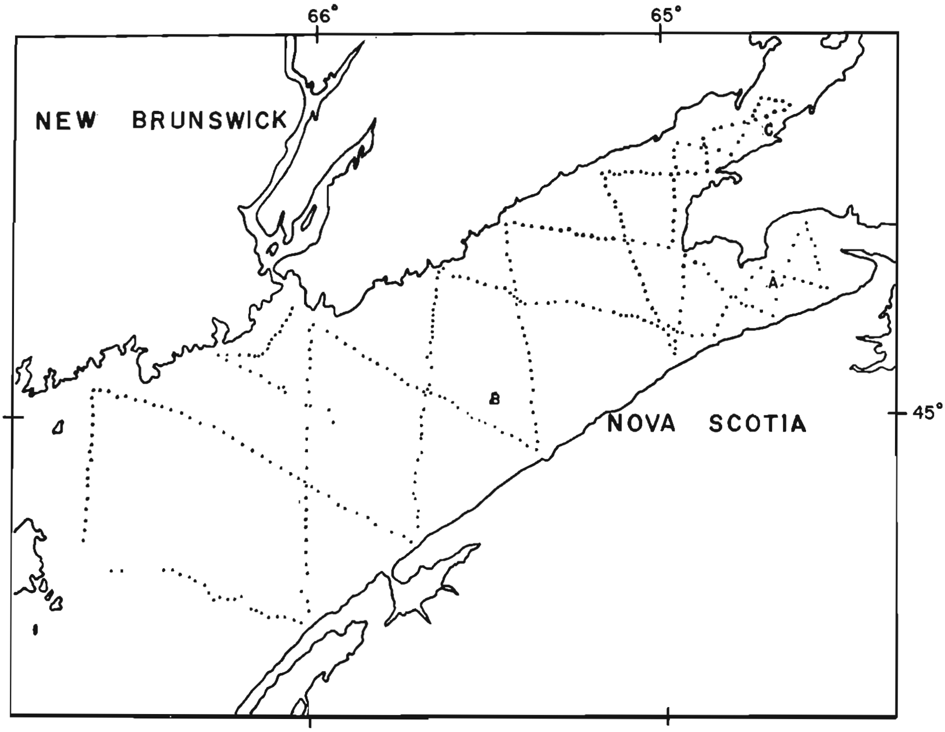


Figure 17. Sub-bottom profiling net, Bay of Fundy. Dots along net are fiducials marked on the Hydrosonde seismic records. Notations A, B, and C refer to seismic reflection profiles discussed in Figure 21.

Textural facies of bottom sediments in the Bay of Fundy are shown in Figures 18 and 19. Figure 19 presents the per cent probability of finding the namesake sediment type within each textural province. The probability is lowest for the coarser sediments, and highest for the fine sediments.

Textural provinces in the Bay of Fundy fall into three main groups consisting of gravels, sands and muds. Gravels floor most of the bay. A large area of sand occurs in the centre of the bay, and sand provinces also form transition zones between the mud and gravel provinces. Mud provinces extend along the northwestern margin of the bay. An isopach map (Fig. 20) of the Quaternary section indicates that thicknesses in excess of 30 m are widespread only beneath the mud provinces. The thinner section beneath the sand and gravel provinces exhibit patterns on the sub-bottom profiles which we interpret as till (massive, "sharkskin" pattern, due to numerous intersecting hyperbolas generated by point sources; see Fig. 21); or as glaciofluvial outwash (stratified pattern). These interpretations have locally been verified by tracing them into the intertidal zone where they can be examined in outcrop. Thus, the Holocene sand and gravel provinces appear to be skins of

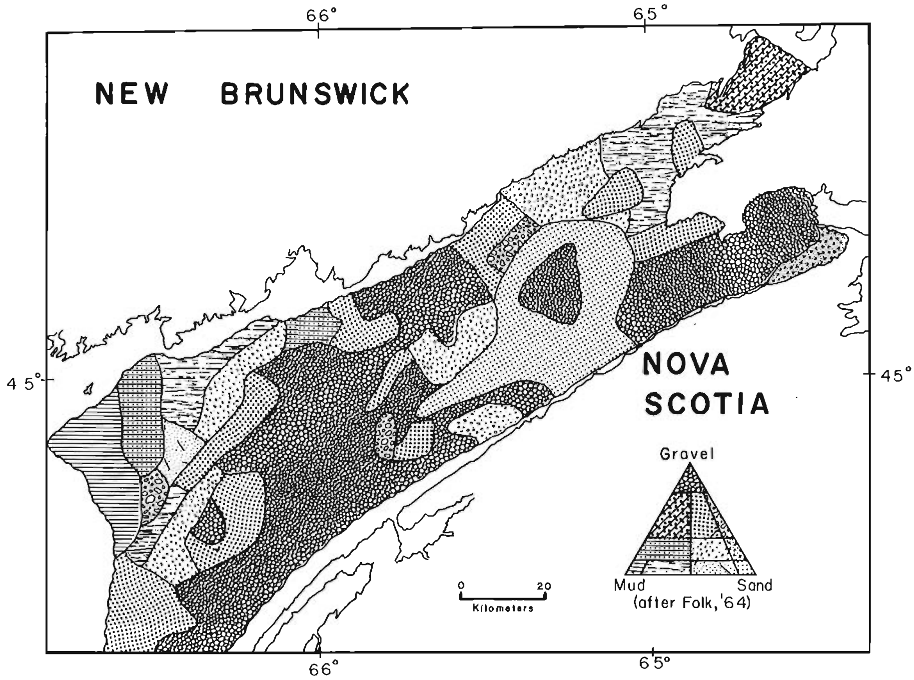


Figure 18. Textural facies in the Bay of Fundy.

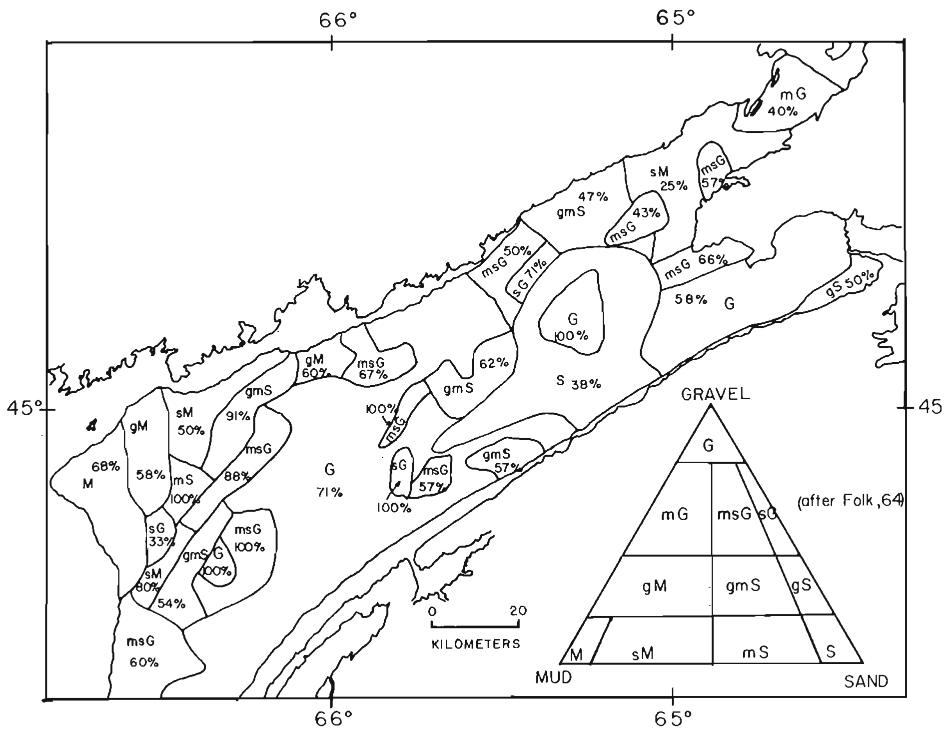


Figure 19. Percent probability of finding namesake texture for textural provinces, Bay of Fundy.

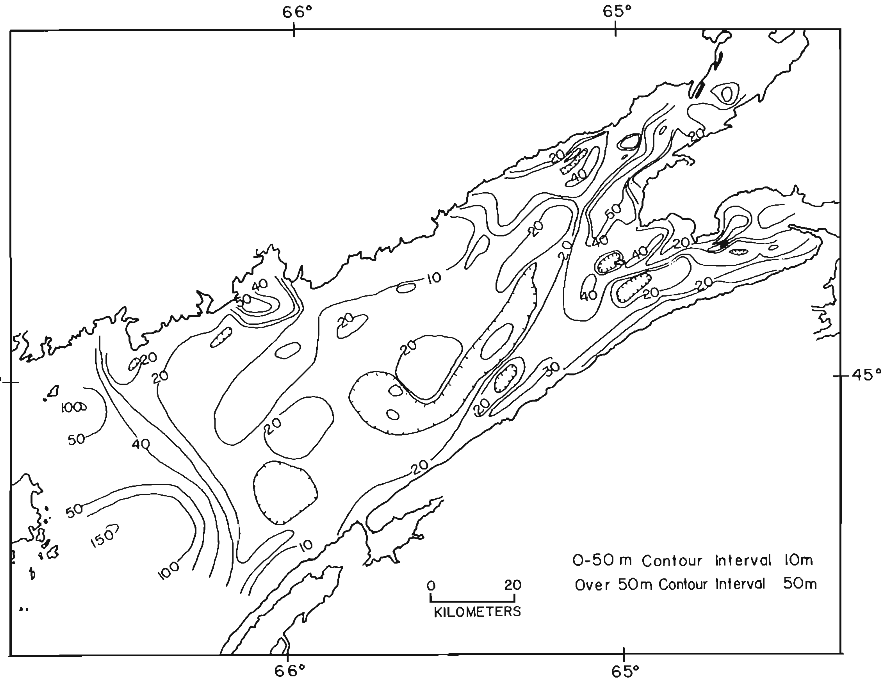


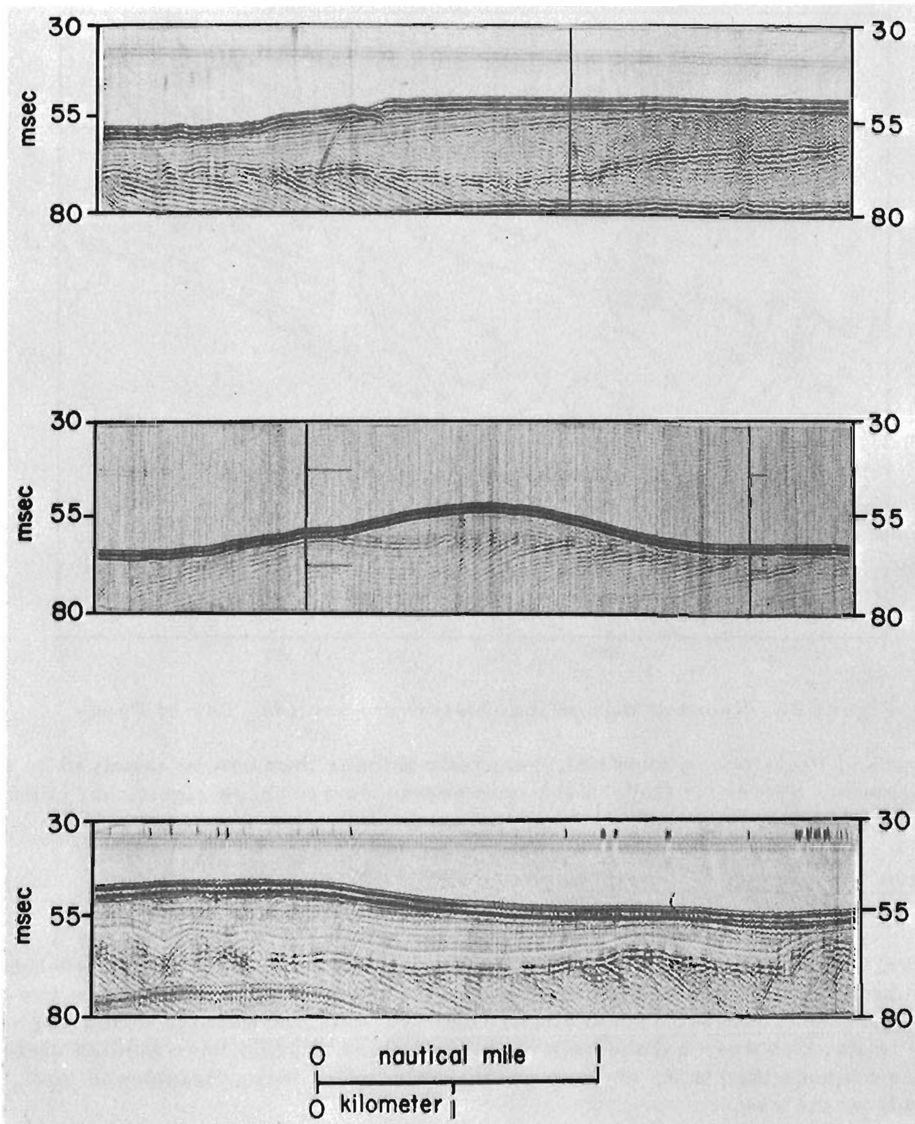
Figure 20. Isopach map of the Quaternary section, Bay of Fundy.

reworked Pleistocene material, generally thinner than can be resolved by the sub-bottom profiler. Only in the mud provinces are there significant build-ups of Holocene materials.

Gravel Provinces

Medium to coarse pebble gravels, sandy gravels, and muddy sandy gravels occupy 22% of Fundy's floor. They occur mainly beyond the 40-metre contour (Fig. 18) where they are immune to wave action. The one- to two-knot (0.5 to 1.0 m/sec approximate) tidal currents are capable of moving very fine to fine pebbles, but the bulk of the bay-floor gravels have median diameters coarser than this, and are presumably relict from Pleistocene low stands of the sea.

This hypothesis is strengthened by the relationship of modal diameter (determined by intercept measurement) to roundness of the modal size class (determined by visual estimation) for Fundy gravels. Figure 22 presents this data as a scatter plot, for 10 submarine samples, and for 14 samples of known provenance from Fundy's margin. The submarine samples (diagonally ruled fields) fall into two groups. Five bay-floor gravel samples are sub-angular, and the individual stones are in many cases encrusted with calcareous or chitinous organisms (Fig. 23). Fishermen report vast areas floored by such encrusted pebbles which they refer to as "copper bottoms". Clean, sub-rounded to well-rounded gravels occur only in such local zones of high tidal velocity as the Minas Passage, and Chignecto Bay, off Cape Enrage. The stippled field represents marginal gravels of potential source



A. Stratified drift overlying till on Triassic strata, central Fundy.
B. Solitary sand waves localized by Triassic bedrock highs, central Fundy.
C. Stratified muds, lower Chignecto Bay. Bedrock is Triassic. Second echo echo is visible at lower left.
Note: Top, middle and lower panels are A, B, and C respectively and correspond to locations in Figure 17.

Figure 21. Sub-bottom profiles of the Bay of Fundy.

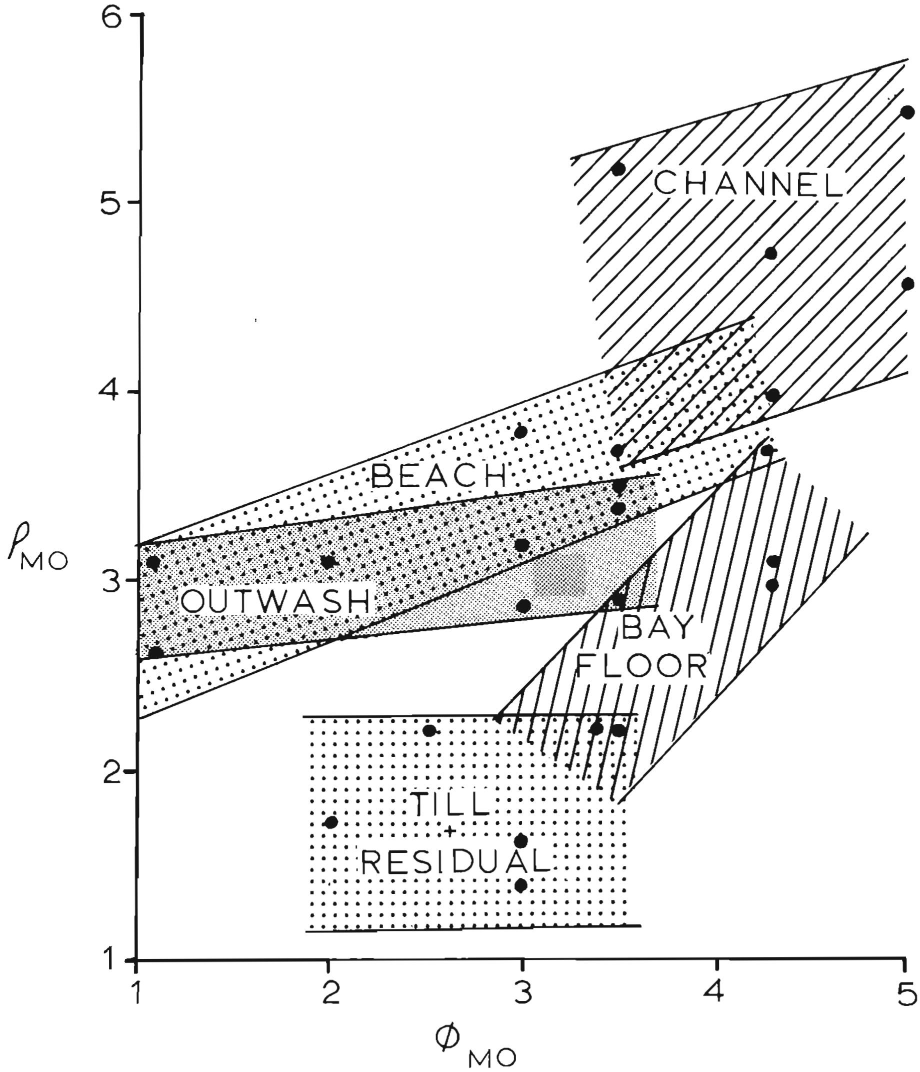


Figure 22. Modal diameter vs. roundness of modal size class for Bay of Fundy gravels.

environment. Residual gravels, till stones, and glaciofluvial outwash are sufficiently angular to serve as sources for bay-floor gravel. Thus, the vast sectors of encrusted pebbles appear to be relict from the Pleistocene. Only in the high velocity straits are gravel bottoms active.

The major petrographic components of Fundy gravels in order of abundance are granitic and gneissic rocks, red sandstone and shale, drab sandstone and shale, and basalt. The first group is of pre-Carboniferous origin, the second group is primarily of Carboniferous origin, and the third and fourth groups are primarily of Triassic origin. The composition of

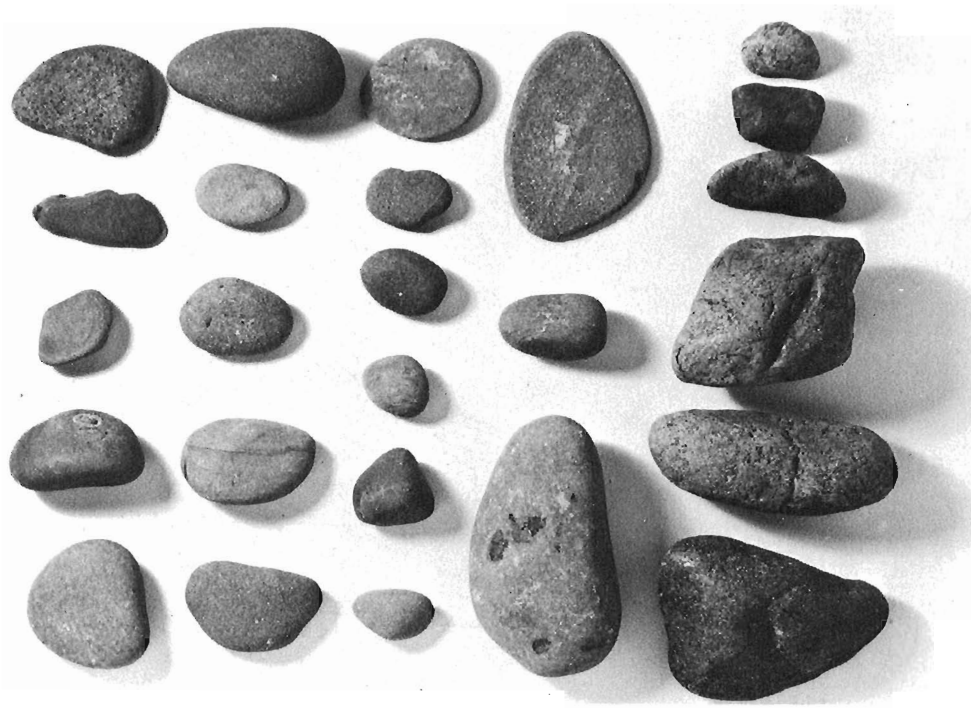


Figure 23a. Bay floor gravels from central Fundy.

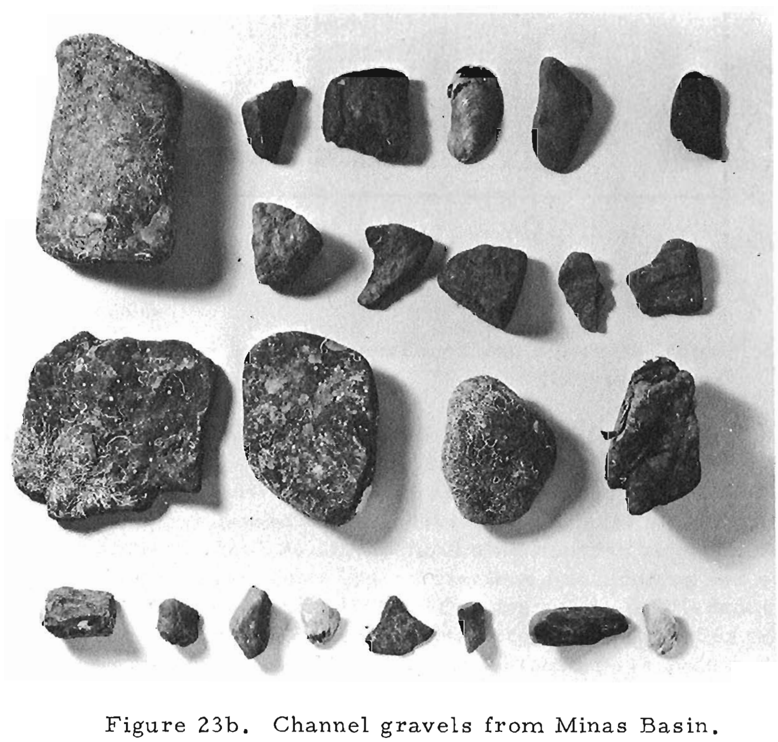


Figure 23b. Channel gravels from Minas Basin.

gravel samples generally reflects the composition of the underlying bedrock, and also its resistance to abrasion. Gravel samples from pre-Carboniferous and Carboniferous terranes contain 20% or less of far-travelled material. But for gravel samples taken from friable oven-red Triassic mudstones, the percentages are often reversed, unless there is abundant sea-floor outcrop.

Sand Provinces

Sands, gravelly sands, and muddy sands occupy 22% of the bay floor (Fig. 18). Median grain size is medium to very coarse, except in the vicinity of mud deposits where it is fine to very fine. Two major sand deposits are present, one in a transverse north-south band towards the head of the bay, and a second transverse band across the eastern corner of the bay. Where they border mud deposits, these sand provinces may be in part hydraulically maintained transition zones.

Forgeron (1962) has collected 28 gravity cores in the Bay of Fundy, up to 2 m in length. His data indicate that the contact between the mud provinces and the fine sand provinces dips gently southeast; thus the fine sands appear to be transgressing the muds. Forgeron suggests that the transitional fine sands are in fact lags generated during the retreat of the zone of mud accumulation through the late Holocene, as Fundy's currents intensified (see Swift and Borns, 1967a, and 1967b for a discussion of the ontogeny of the tidal regime). Forgeron reports sand-filled borings within the buried mud, and interfingering of thin beds of mud and sand (Forgeron, 1962, p. 109).

The main mass of sand toward the head of Fundy may be an outwash delta generated by a late Pleistocene periglacial river, whose channels have been traced by sub-bottom profiles (Swift and Lyall, 1968b), or generated by meltwater from a late Pleistocene local ice cap centered on southern Nova Scotia (Hickox, 1962), or by both. Sub-bottom profiles from this area reveal irregular solitary sand waves with amplitudes up to 2 m, localized by bedrock or till highs (Fig. 21). They are mainly transverse forms with flood asymmetry (Fig. 24).

In the Minas Channel and Minas Basin, the Friable Wolfville sandstone (basal Triassic) forms the bedrock, and the overlying tills are very sandy. Sand released from the till and exposed bedrock has been swept by the tide into plano-convex sand with upper surfaces may reach into the intertidal zone. These surfaces are deformed into longitudinal sand bars that extend for several kilometres and bear extensive sand wave fields (see Swift et al., 1966; and Swift and McMullen, 1968).

Mud Provinces

Muds and muddy sediments occupy 20% of the floor of the Bay of Fundy (Fig. 18). They occur along the northeast side of the Bay of Fundy from Chignecto Bay to Grand Manan Island. On sub-bottom profiles they are well stratified and acoustically transparent (Fig. 21), and up to 100 m thick. It is possible that the thickest sections may be in part Pleistocene freshwater lake deposits that predate the Holocene transgression. Cores and grab samples indicate abundant ice-rafted cobbles and pebbles in the upper 2 m of the mud deposits. The mud deposits are localized by the weaker tidal currents of the northeast side of the bay, and by the step-wise increase in

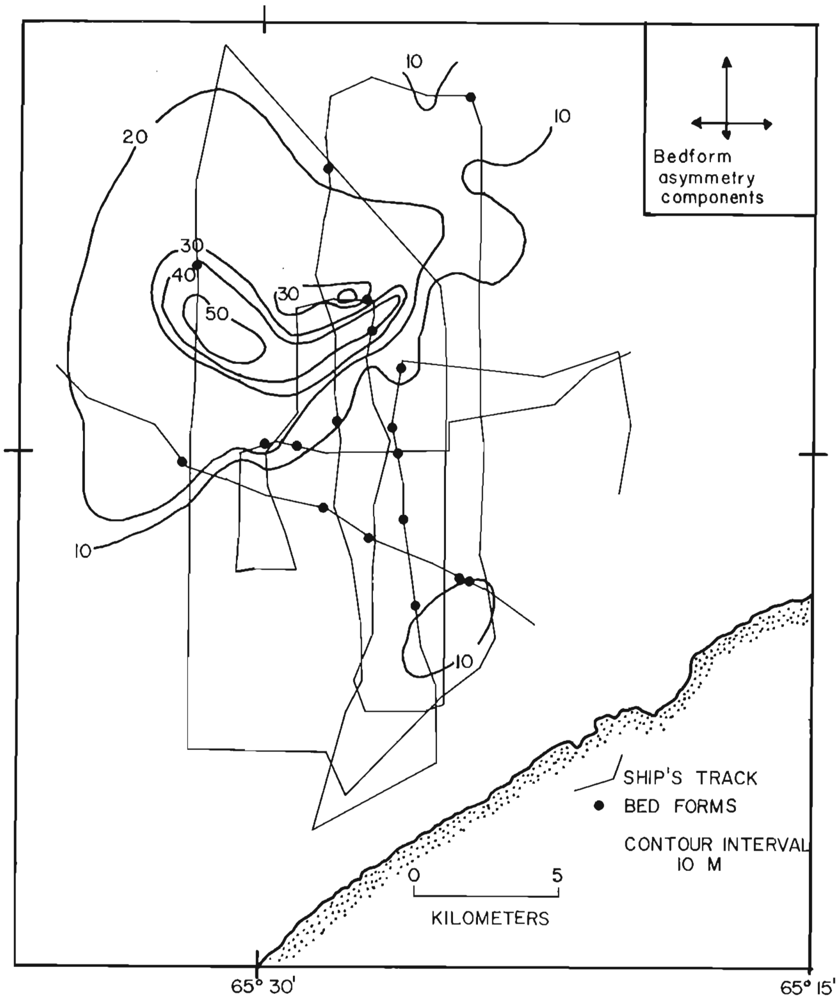


Figure 24. Locations of solitary sand waves and their asymmetry in plane of profile.

suspended sediment concentration as the counterclockwise residual tidal currents receive the successive discharge of the Minas Basin, the Chignecto Bay, and the St. John River.

LITHOFACIES RATIOS AND TRENDS OF SORTING VALUES

From the gross petrologic plot of the samples shown earlier (Fig. 18), it was decided to carry out a lithofacies analyses based on a restricted binary classification, in order to introduce some refinement in the interpretation of the facies trends. To carry out these analyses the lithofacies ratios were adopted from the classification of McMullen, (see Sen Gupta and McMullen, 1969, p. 477) which is given in Table III.

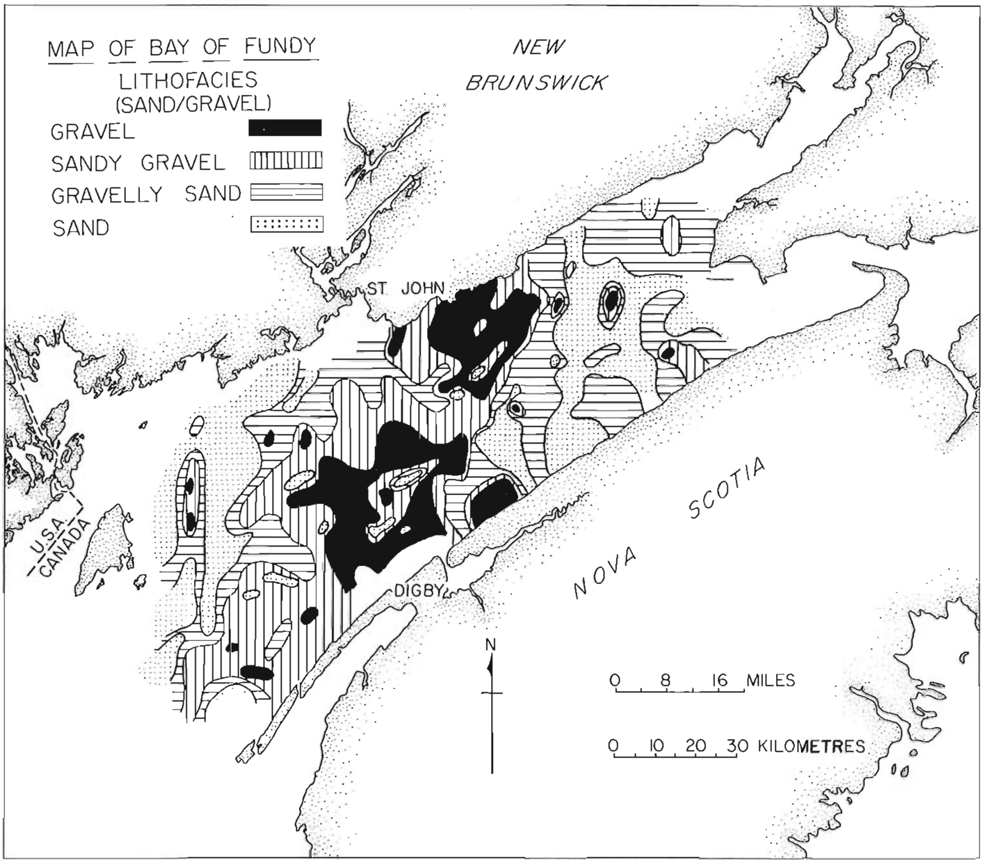


Figure 25. Lithofacies map of sand and gravel occurrences in the Bay of Fundy. Note similarity of pattern to those in Figures 26 and 28. (Figs. 25 to 29 after Pelletier in Pelletier and McMullen, 1971).

TABLE III

Lithofacies Classification (after McMullen)

Lithofacies	Criterion
gravel	$\geq$80% is > 2.0 mm
sand gravel	50-80% is > 2.0 mm
gravelly sand	10-50% is > 2.0 mm
sand	$\geq$90% is > 0.063-2.0 mm
muddy sand	10-50% is > 0.063 mm
sandy mud	50-75% is > 0.063 mm
mud	$\geq$ 75% is > 0.063 mm

The lithofacies analyses are given in two parts. In the first part the lithofacies ratio of sand to gravel was plotted (Fig. 25), and it became

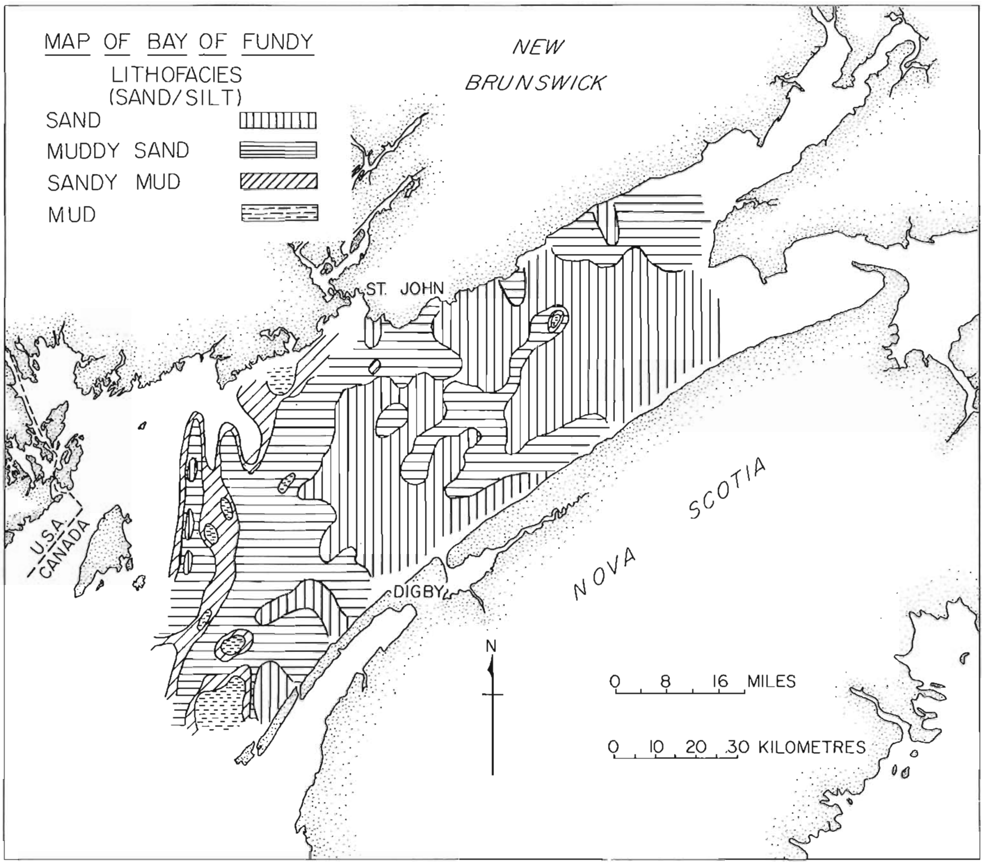


Figure 26. Lithofacies map of sand and mud occurrences in the Bay of Fundy.

apparent immediately that the sediments occurred in more-or-less alternating areas occupied by respectively coarse and fine material. These trends occur oblique to the south shore in a north-south direction, but are somewhat more perpendicular to the north shore. The distance between the axes of these trends is approximately equidistant along the length of the Bay. In the second part of the facies analysis the lithofacies ratio of sand to mud was plotted (Fig. 26), and this reflected a similar trend although generally the coarser material is in the upper part of the Bay.

As an adjunct to the facies studies moment measures were calculated, and a plot of the second moment (standard deviation) (Fig. 27) was made. This presentation of the relative degree of sorting, showed a good correlation with the trends of the lithologic ratios. Areas of better sorted sediments lie in the upper part of the Bay, with one at the lower part, and are separated by areas of more poorly sorted sediments. These areas of good and poor sorting coincide respectively with the coarser and finer sediment provinces (Fig. 28).

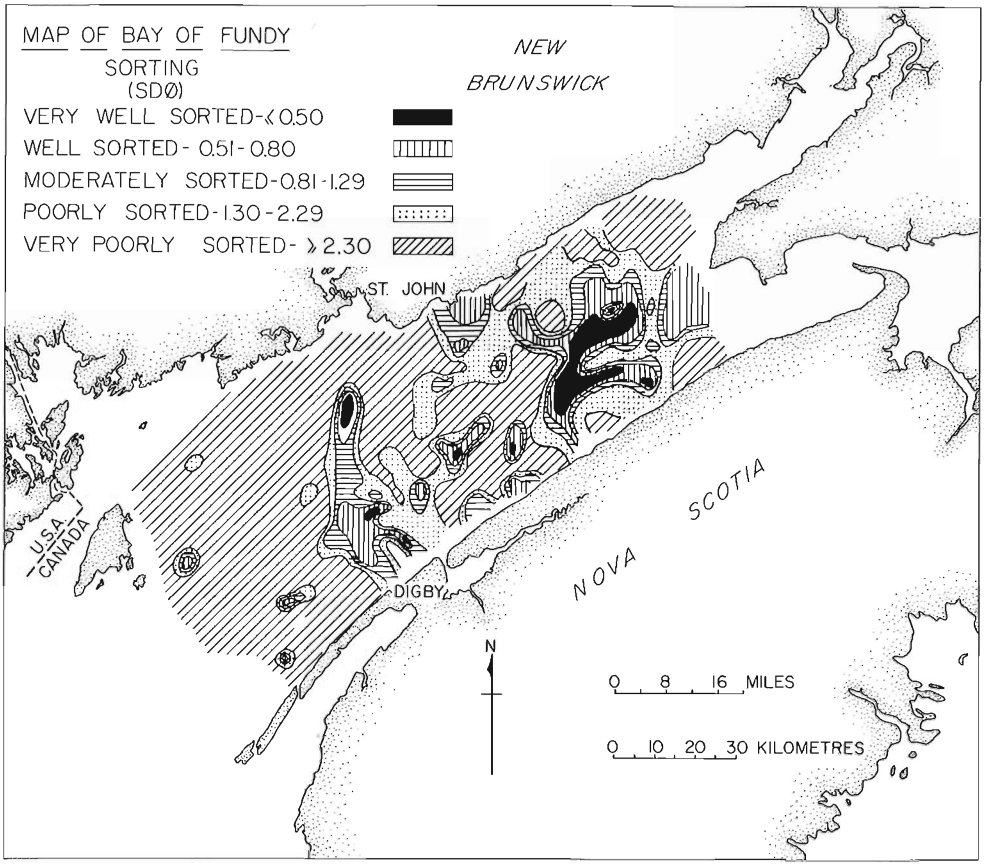


Figure 27. Sediment sorting based on standard deviation (SD ϕ) in Bay of Fundy.

The pronounced, almost equally spaced trends suggest that a combination of factors is responsible for such an occurrence. This phenomenon may be due to erosion of the relict Pleistocene sediments under conditions of a standing wave, and one that may have migrated in time as the spacing between the trends is almost one-quarter of the length of the Bay.

Hydrodynamic vigour and sedimentary texture

Although quantitative data are not available for bottom currents over each sampling site, a qualitative and relative assessment can be made on the relationship of hydrodynamic vigour and sedimentary texture. This is shown by means of a ternary plot (Fig. 9) in which the major lithologic components of gravel, sand and mud are represented as 100 per cent at each apex respectively. On the premise that decreasing texture corresponds to decreasing hydrodynamic vigour, the regime of greatest hydraulic energy is placed at the gravel apex, and that of the least energy is placed at the mud apex with the intermediate regime placed at the intermediate sand apex. Thus sediments depositing from tractional transport and in equilibrium with their

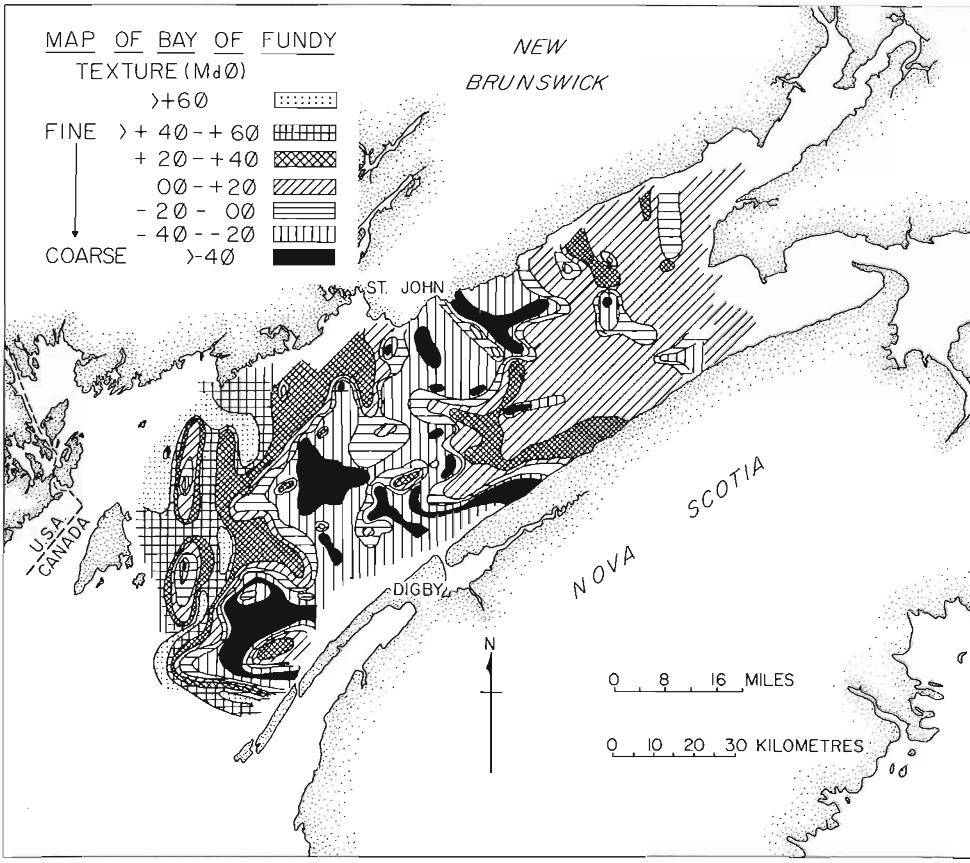


Figure 28. Distribution of median diameters of individual samples.

corresponding hydraulic regime should have the position of their lithologic ratio plot along the boundary between these apices. Any departure from this position must represent deposition from suspension such as in the extreme case of a sediment with a lithologic ratio that plots along the gravel-mud boundary. Sediments containing a partial load from suspension will have the position of their lithologic ratio plot in an intermediate position between the suspension boundary and the traction boundary i.e. somewhere within the field of the diagram.

In the Bay of Fundy it appears that two major sub-systems of mechanical sedimentation are in operation. One is a strong tidal system in which sediments appear to be in equilibrium with their hydrodynamic regime as shown by the plot of their lithologic ratios. In this case 56 per cent of the 300 samples selected from the Bedford Institute surveys plot along the gravel-sand-mud boundary, a fact which is consistent with the observation of a decrease in texture corresponding to a decrease in hydrodynamic vigour. The second system of mechanical sedimentation appears to involve settling from suspension. This is understandable considering the volume of sediments of the finer sizes contributed by rivers and from erosion of the shoreline,

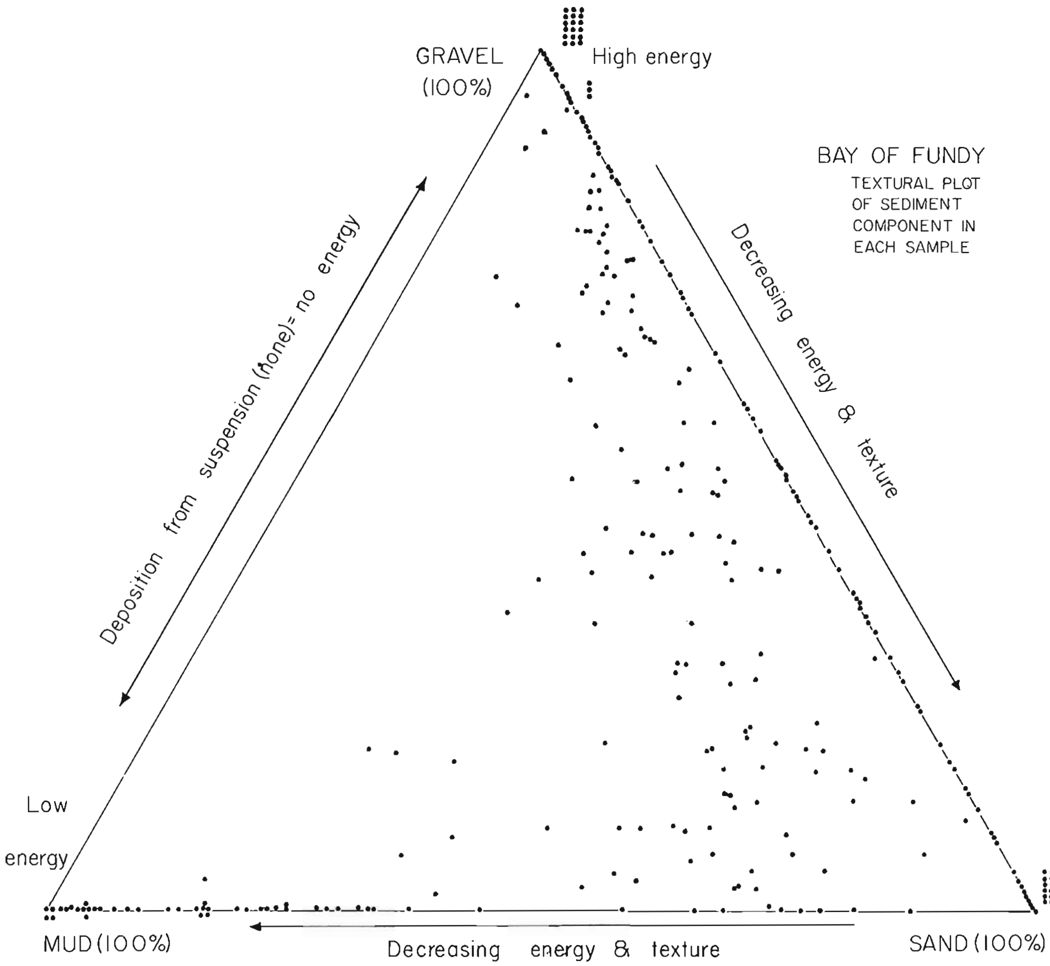


Figure 29. Ternary diagram showing relationship of textural composition for individual samples shown by dots and the relative energy available to erode at the sampling site in Bay of Fundy. Note: dots outside of ternary diagram refer to a plotted position at the adjacent apex.

together with the bottom load in parts of the Bay that is thrown into suspension by tidal action. Overall, the sediments in the Bay of Fundy lie between two extremes of relationships with hydrodynamic vigour. On the one hand they are related to a system of close hydrodynamic equilibrium such as in the Minas Basin system (Pelletier and McMullen, 1972), and on the other hand, they are related to a system of epeiric-sea deposition involving some deposition involving some deposition from suspension such as that in the Hudson Bay system (Pelletier, 1969).

ACKNOWLEDGMENTS

The work presented in this paper was carried out under the auspices of the Atlantic Oceanographic Laboratory and the Atlantic Geoscience Centre Bedford Institute of Oceanography, Dartmouth, Nova Scotia; Dalhousie University Institute of Oceanography and Department of Geology, Halifax, Nova Scotia. The work of the first author was funded by the Nova Scotia Research Foundation, and also by grants A1945 and A2686 of the National Research Council of Canada. We thank R.M. McMullen, W.I. Farquharson, R. Cormier and C. Langford, all formerly of the Atlantic Oceanographic Laboratory, P.J. Wangerksy and M.J. Keen of Dalhousie Institute of Oceanography, J.E. Blanchard of the Nova Scotia Research Foundation, F. Medioli, P. Schenk and C.G.I. Friedlaender of the Dalhousie Geology Department, D.J. Stanley of the Smithsonian Institution, A.E. Cok of Adelphi University, D. Laming of the University of Swansea, and T.T. Davies of the University of South Carolina. These people discussed various aspects of the work with the authors and provided many useful suggestions. W.D. Forrester of the Atlantic Oceanographic Laboratory, and G. Drapeau of the Atlantic Geoscience Centre offered many critical comments on the final manuscript.

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(Earth Science Symposium on offshore Eastern Canada
Geol. Surv. Can., Paper 71-23, 1973, p. 153-166)

9. CRUSTAL STRUCTURE BENEATH THE BAY OF FUNDY FROM OCEAN TIDE LOADING

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Abstract

The M_2 tide in the Bay of Fundy has a large amplitude, 5 metres, and a well-known spatial distribution. It acts as a large sinusoidal load, period 12.42 hours, on the earth's surface. The loading effects of the tides on gravity or tilt have been measured at six stations in Nova Scotia. The M_2 tilt amplitude from an eight-month series of tilt observations at Rawdon, (45.07°N, 63.80°W) 25 km. south of the Bay of Fundy, is compared with theoretical model tilts. The layered theoretical spherical earth models have been solved using the finite element method.

The observed tilts show good agreement with the seismic refraction models from the Atlantic coast of Nova Scotia. The models suggest that the Nova Scotian crust is 35 km. thick, consists of three major layers, and is underlain by normal mantle.

INTRODUCTION

Our knowledge of the elastic properties of the earth's crust and upper mantle is provided almost exclusively by seismology. An independent determination of these elastic properties is provided by tidal-loading observations because the deformation of the earth under a surface load is directly related to the elastic properties of the earth beneath the load.

Measurement of elastic properties by tidal loading differs in two important aspects from the seismic technique. Firstly, the strain-inducing mechanism, the ocean tides, cover large areas of the earth's surface. We therefore expect that tidal-loading observations will reflect earth structure averaged over a large area. Secondly, the tidal-loading input frequencies are five orders of magnitude lower than seismic frequencies and there is some evidence that the earth's response is significantly anelastic at these lower frequencies (Zadro, 1964).

The luni-solar gravitational attraction on the earth can be expressed as a sum of sinusoidally-varying component waves, each with its own characteristic amplitude and frequency. The amplitudes and frequencies are determined by the masses and relative motions of the sun-earth-moon system. A description of the luni-solar potential is given by Melchior (1966). When describing tidal phenomena it is normal to discuss the behaviour of each of the component waves separately. The amplitude and phase of each component is found by spectral analysis of suitable length of tidal record. In tidal loading studies the most important component wave is the principal lunar semidiurnal (M_2, period 12.42 hours) because it has the best known spatial

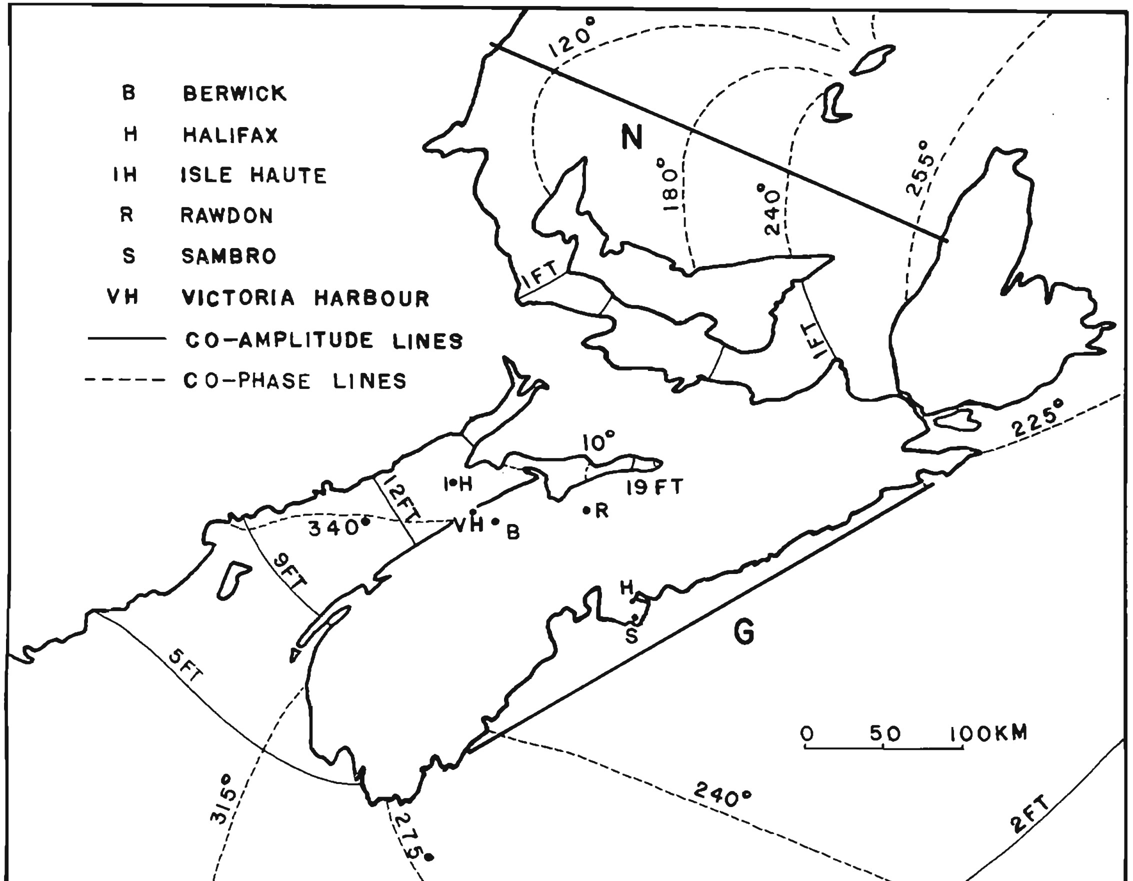

Figure 1. Simplified diagram of the semi-diurnal tide distribution around Nova Scotia. The amplitudes are given in feet and phase lags in degrees with respect to the equilibrium tide at meridian 60° W. The diagram also shows the sites of tilt and gravity stations and the seismic profiles - G; Port Herbert - Long Lake - Cole Harbour, and N; Tracadie - Cheticamp.

distribution and greatest average amplitude. The earth's rotation causes an apparent passage of the moon around the earth every 24.84 hours. The period of motion is not 24 hours because the moon is also moving in its orbit around the earth. The M_2 tide is produced by this apparent motion of the moon. Its period is half that of the moon's apparent motion because the tidal forces raise two tidal bulges on opposite sides of the earth. The response of the ocean water masses to the gravitational attraction of the sun and moon is modified by the shape of the oceans, resulting in spatial variations of the amplitudes and phases of the ocean tidal constituents.

During the last five years Dalhousie University has undertaken a study of crustal structure beneath the Bay of Fundy and the Nova Scotian continental shelf using ocean-tide loading. It is hoped that the results will give a general picture of the crust and upper mantle, that can be compared with results from seismic refraction profiles. The tides in the Bay of Fundy provide an ideal input load. The M_2 component has a well-known spatial distribution, and an amplitude in excess of five metres in the Minas Basin.

In addition to providing direct evidence on earth structure, tidal-loading observations indicate the influence of the ocean tides on the earth tide. The earth tide is the deformation of the earth resulting from the direct luni-solar gravitational attraction on the solid earth. Both ocean tides and the earth tide result from the luni-solar gravitational attraction, therefore, the measurements of surface tilt and gravity have contributions from both earth tide and tidal loading. The relative magnitude of the two effects depends upon latitude and distance from an ocean. In gravity measurements the earth tide dominates, the maximum tidal-loading effect is ten percent of the total amplitude. Tilt measurements around the Bay of Fundy, however, show a tidal-loading amplitude many times larger than the earth tide amplitude. At inland stations tidal loading causes small perturbations on measurements of the earth tide. In fact, as a result of recent tidal gravity studies Kuo (1970) and Farrell (1970) have suggested that the major spatial perturbation of the earth tide results from the indirect effect of the ocean tides. Until the response of the earth to tidal loads is well understood, it is impossible to determine the influence of crustal and upper mantle inhomogeneity on the earth tide. Does the earth behave like a layered elastic sphere under tidal loads or do crustal blocks tilt independently as suggested by Tomaschek (1953)?

Tilt and Gravity Measurements in Nova Scotia

The north-south and east-west components of surface tilt at Rawdon (45.07°N, 63.80°W) were measured by Lambert using Verbaandert-Melchior horizontal pendulums (Lambert, 1970; Melchior, 1966). Lambert has also measured tidal gravity on Isle Haute in the Bay of Fundy (45.25°N, 65.00°W), at Victoria Harbour (45.11°N, 64.88°W), and Berwick (45.03°N, 64.75°W), using a LaCoste-Romberg tidal-recording gravimeter. Because the gravity measurements are relatively insensitive to earth structure, only the more sensitive Rawdon tilt measurements are discussed in this paper. Further measurements have been made on the Atlantic coast of Nova Scotia; tilts at Sambro (44.47°N, 63.62°W) were measured using horizontal pendulums (Beaumont et al., 1970), and gravity at Halifax (44.63°N, 63.75°W) was measured using a TRG 1 tidal-recording gravimeter. The Sambro tilt measurements show tidal loading effects of the same order of magnitude as those from Rawdon. Because Sambro is close to the Atlantic Ocean, these tilt measurements are most sensitive to near surface structure. The gravity measurements at Halifax were made in conjunction with Dr. J. T. Kuo of Colombia University and have not been analyzed so far. The locations of the tilt and gravity stations and the distribution of the M_2 tide around Nova Scotia are shown in Figure 1.

Tilt Measurements at Rawdon

(i) Analysis of data

Spectral analysis of the tilt records from Rawdon gave the following amplitudes (in milliseconds of arc) and phase lags (in degrees) for the north-south and east-west components of the M_2 constituent (Lambert, 1970).

North-south 26.4 ± 0.5 msecs., $188.2 \pm 1.0°$
East-west 27.4 ± 0.6 msecs., $274.1 \pm 0.8°$

The phase lags are given with respect to the phase of the M_2 constituent of the luni-solar potential at Rawdon.

(ii) Correction of the observed tilt for the earth tide and Newtonian attraction

As previously explained, part of the measured amplitude is due to the earth tide (the body tilt). The tilt response of the solid earth to the tide-generating forces is characterized by the diminishing factor, D_2, which is the ratio of the tilt measured, or theoretically predicted from models, to the tilt experienced on a perfectly rigid earth. Alsop and Kuo (1964) have shown theoretically that for reasonable earth models D_2 should not differ from 0.680 by more than two percent. Melchior (1966) on the other hand, believes that $D_2 =$ 0.706. This result represents the weighted average of a large number of observations. I have assumed a diminishing factor of 0.700. The discrepancy between the theoretical, observed, and assumed diminishing factors has little influence on the conclusions reached because the tidal-load amplitude is much larger than the earth-tide amplitude. The relative magnitude of the

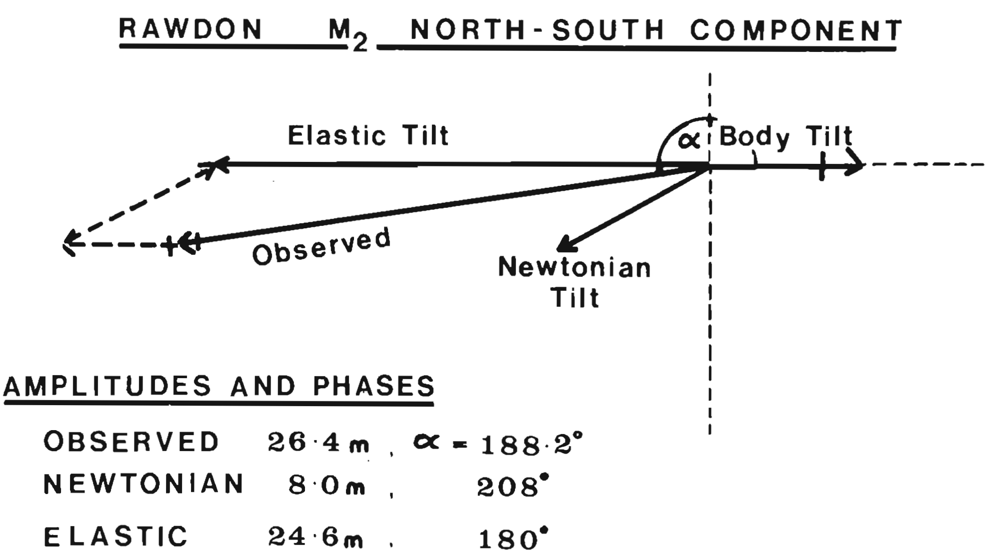

Figure 2. Vector plot for the north-south component of tilt at Rawdon.

earth tide and the tidal load is shown in the vector plots (Figs. 2 and 3) for the north-south and east-west components of tilt. The theoretical tilt on a rigid earth was calculated using the method given by Longman (1959). In the vector plots, the amplitudes and phases of the observed, earth tide, Newtonian, and elastic tilts are represented as vectors with magnitude equal to the amplitude of the tilt and angular rotation (α) equal to the phase of the tilt with respect to the earth-tide tilt at Rawdon. The total tidal loading amplitude is the vector sum of the elastic and Newtonian tilts.

A tiltmeter measures the angle between the normal to the earth's surface and the direction of the earth's gravitational field. The tidal water masses, in addition to deforming the earth's surface, change the direction of the gravitational field by adding a horizontal component. This Newtonian attraction is measured by the tiltmeter but it is independent of the elastic response of the earth and is therefore treated separately. The M_2 tidal charts given by Yuen (1967), Dohler (1964), and Dietrich (1944), for the Bay of Fundy, Gulf of St. Lawrence, and North Atlantic Ocean were used to calculate the Newtonian tilt using the method outlined by Lennon (1961).

The elastic tilt is given by subtracting the earth-tide tilt (body tilt) and Newtonian tilt from the observed tilt. The elastic tilt is the tilt that results from the elastic yield of the earth under the surface load and is the tilt compared with the theoretical tilts from a series of earth models.

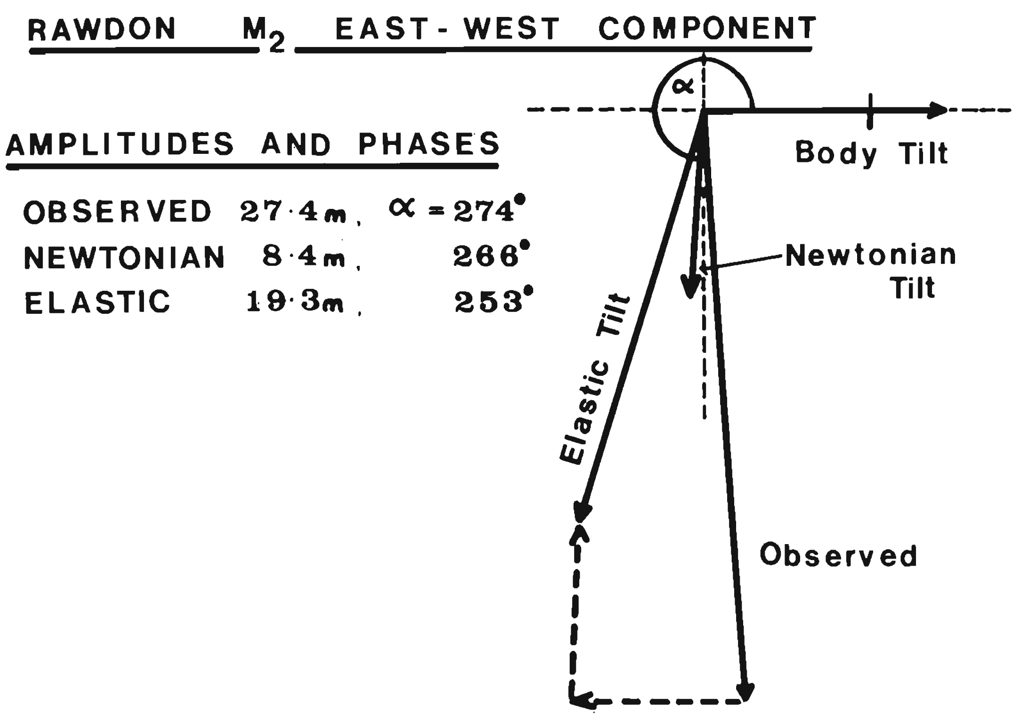

Figure 3. Vector plot for the east-west component tilt at Rawdon. The Body tilt is the tilt due to the earth tide. The Newtonian tilt is the tilt due to the direct attraction of the water masses. The Elastic tilt is the tilt due to the elastic yield of the earth. The phase lags (α) are with respect to the Body tilt, and the amplitudes in milliseconds of arc.

To be precise, there is a small term due to the redistribution of mass in the earth that has been omitted. An estimate of the magnitude of this term (Beaumont and Lambert, in preparation), using the load Love numbers given by Farrell (1970), shows that the omission is justified. The amplitude of the redistribution of mass term is less than 1.5 per cent of the elastic tilt. The errors in the model tilts are probably larger than this redistribution term because the tidal charts may not accurately represent the tidal distribution. When more accurate tidal-loading observations are available and the ocean tide distribution is known in greater detail the redistribution of mass effect must be included.

The corrected north-south and east-west elastic tilt amplitudes are:

North-south 24.6±0.5 msecs., 180.0±1.0°

East-west 19.3±0.6 msecs., 253.0±0.8°

The phase lags, as before, are expressed with respect to the M_2 constituent of the luni-solar potential at Rawdon. The errors assigned to the elastic tilts are the same as the observed tilt errors. The use of the same errors implicitly assumes that the tidal maps are exact. Until many more tidal loading observations are available a better estimate of the tidal distribution is not possible.

Finite Element Model

The observed elastic tilt must now be compared with the elastic tilt of theoretical earth models. The deformation of simple elastic earth models under surface loads has been discussed by Slichter and Caputo (1960), and Longman (1962, 1963) and more recently by Kuo (1969), and Farrell (1970). Slichter and Caputo's model and Longman's model are whole earth models that do not include detailed crustal and upper mantle structure. Kuo's model and Farrell's model are flat earth approximations that only give an accurate representation of the earth's response within 1000 km. of the load. Farrell's original model which is based on Gilbert's normal mode model 8734 (Backus and Gilbert, 1970, p. 165) has a structure that is considerably different from that believed to exist beneath the Bay of Fundy. Farrell has recently extended his results to include a spherical gravitating earth model (Farrell, pers. comm.).

In this paper I consider a layered spherical earth model under a surface point load. The deformation of the model is found using the finite element method (Zienkiewicz and Cheung, 1967). The finite element method has been used extensively in civil engineering, soil mechanics, and rock mechanics to study the deformation of complicated structures under arbitrary loads. It is an approximate method of solution and relies on the division of a continuum into blocks, called elements, that are interconnected only at nodal points. Forces are applied at the nodal points and the structure is solved for the equilibrium nodal point displacements by equating the external work done in displacing the nodal points to the internal work done on the materials within the elements. The accuracy of the nodal point displacements depends on the configuration of the elements. A large number of small elements that closely represent the structure modelled will generally give accurate nodal point displacements.

The axisymmetric finite element earth model used in this study has 902 elements, 41 along the surface by 22 arranged in the radial direction. A simplified diagram of the model is shown in Figure 4. The elements are

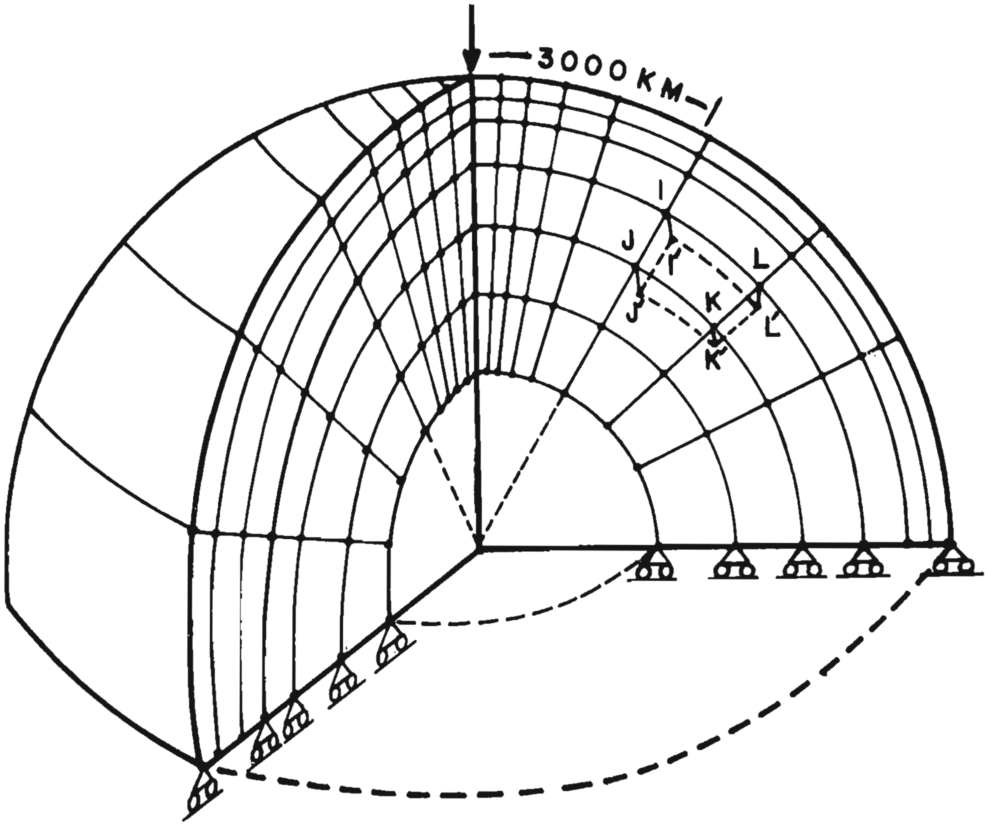

Figure 4. Simplified diagram of a section through the finite element earth model. Under the point load (P) the nodal points I, J, K, and L are displaced to I^1, J^1, K^1, and L^1.

small near the origin, where the displacements must be accurate, and increase in size in proportion to the distance from the origin. A single nodal point load is applied at the surface on the axis. The solution gives the nodal point displacements throughout the model.

The radial displacements of the surface nodal points give the Green's function for radial displacement. The Green's function is the response of the model to a delta-function load. Cubic spline interpolation (Ahlberg et al., 1967) of the nodal point displacements produces a smooth continuous Green's function. The surface tilt Green's function was obtained by differentiating the cubic spline interpolating function.

The model has been shown to give solutions that are accurate to one per cent (Beaumont and Lambert, in preparation), partly by comparison with Kuo's solution for a seven-layer earth model. Comparison with Farrell's model 8734 also gave good agreement, although in this instance exact comparison was not possible because model 8734 has a structure which is too complicated to be represented exactly by the finite element model.

At distances greater than 3000 km. from the point load the solution deteriorates because the finite element model is non-gravitating; it does not

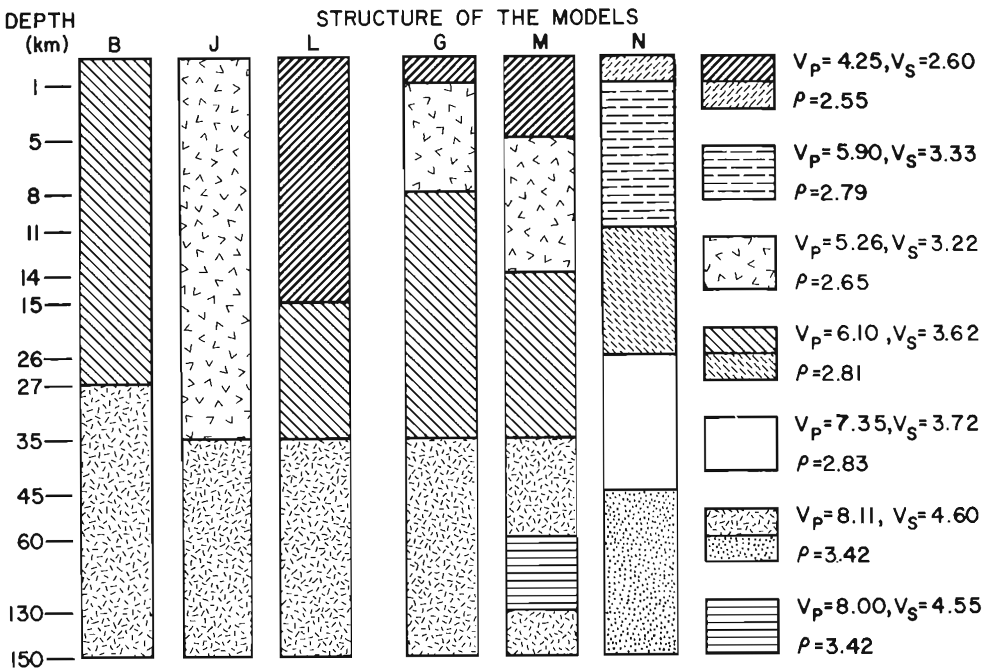

Figure 5. Models B, J, and L are unrealistic. Models G, M, and N are considered realistic. At depths greater than 150 km. all models have similar structures. The properties of the layers are expressed in terms of compressional velocity, Vp, shear velocity, Vs, and density ρ.

model the earth's gravitational field. The inaccuracy in the Green's function beyond 3000 km. is unimportant because the magnitude of the Green's function in this region is small. Therefore, the contribution to the total tilt at Rawdon from the tides beyond 3000 km. is also small. The elastic tilt at Rawdon for each model was calculated by summing the effects of small areas of the ocean tide distribution weighted by the Green's function; that is, convolving the tilt Green's function with the tidal distribution.

RESULTS

The tilt Green's functions for 14 simple models were calculated using the finite element model. The crust beneath the Bay of Fundy was assumed to be continental type and consequently the depth to the Moho was only varied between 27 km. and 45 km. An oceanic model gave tilts that are far too small. The structure of six of the models is shown in Figure 5. At depths greater than 150 km. the models have similar structures, based on the Bullen B seismic model. Other models had different properties below 150 km. depth, but the Rawdon tilt measurements appear insensitive to earth structure below 100 km.

The tilt Green's functions corresponding to the models are shown in Figures 6 and 7. The Green's functions have been normalized by the Newtonian

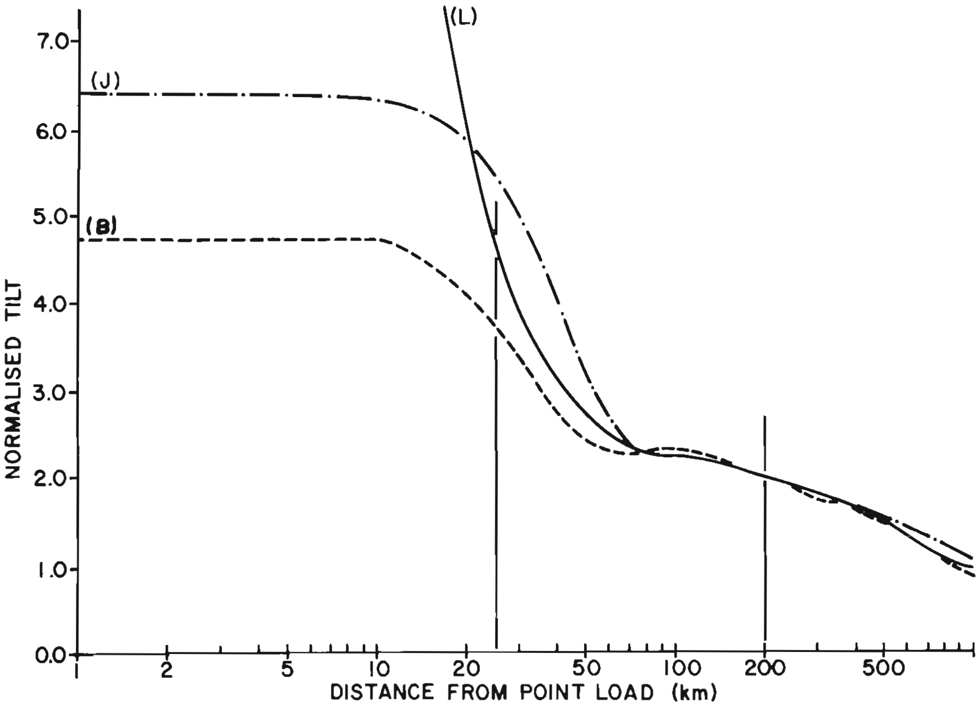

Figure 6. Normalized tilt Green's Functions for models B, J, and L.

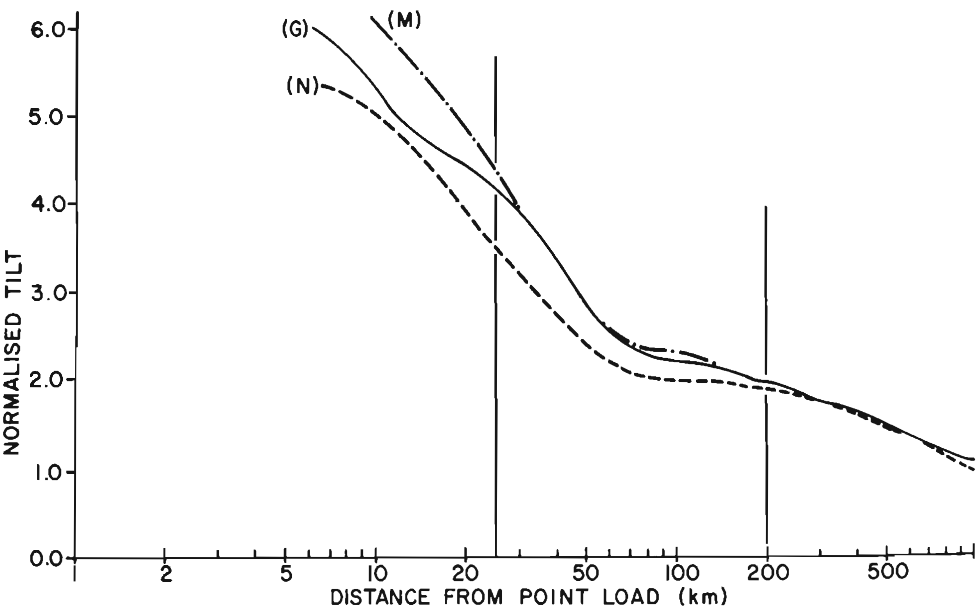

Figure 7. Normalized tilt Green's Functions for models G, M, and N.

attraction of the point load. The interpretation of the diagrams is best illustrated by an example. A point load 25 km. from the tiltmeter on model G will produce an elastic tilt that is 4.2 times larger than its Newtonian tilt, whereas a point load 200 km. from the tiltmeter produces an elastic tilt that is only 2.0 times larger than its Newtonian tilt. The Green's functions for homogeneous earth models would appear as horizontal lines.

Models B, J, and L are unrealistic because they have unrealistic material properties and very simple structures. The results are included to illustrate bounds for the acceptable models.

Models G, M, and N are based on the results of seismic refraction and gravity profiles for Nova Scotia. It is unfortunate that no refraction line has been shot in the Bay of Fundy, therefore, no direct comparison of the tidal loading results with the refraction results is possible. The structure beneath the Bay of Fundy is probably intermediate between models G and N, as the two refraction lines on which the models are based were shot on opposite sides of Nova Scotia (see Figure 1).

Model G is an average of the model from Barrett's Atlantic coast, Port Herbert-Long Lake-Cole Harbour seismic profile (profile G, Figure 1) (Barrett et al., 1964). A 1 km. thick layer of sediment has been added at the surface to agree with the sediment isopach map for the Minas Basin. The presence of the sediment layer is unimportant because measurements at Rawdon are too far from the Minas Basin to sense earth structure at depths less than five km.

Model M is similar to model G but has thicker sedimentary and crystalline basement layers. It also has a slightly low velocity region between 60 km. and 130 km.

Model N is an average of the model from Ewing's Cheticamp to Tracadie profile (profile N, Figure 1) (Ewing et al., 1966). The crustal structure of this model is significantly different from model G. The mantle, at a depth of 45 km., has an anomalously high Pn velocity of 8.50 km./sec., and is overlain by a 19 km. thick intermediate layer with a velocity of Vp= 7.35 km./sec.

A comparison of the elastic tilts, predicted by the models with the observed north-south and east-west components of elastic tilt is shown in Table 1. For the north-south direction models B, J, and L can be rejected. Model M gives very good agreement with observations but model G also falls within the error limits. The Gulf of St. Lawrence model, model N, gives a tilt that is too small and can therefore be rejected. In the east-west direction, models G and M give tilts that are too large, but model N agrees closely with the observation.

CONCLUSIONS

The elastic earth model that agrees with the north-south component of tidal loading tilt at Rawdon is also in good agreement with the seismic refraction model for the Atlantic coast of Nova Scotia. This model suggests a three layer crust with a Moho at 35 km. The model that agrees with the east-west component of tilt is in agreement with the seismic refraction model from the Gulf of St. Lawrence. This model suggests a thickened continental type crust with a Moho at 45 km.

The results are preliminary. They consider the tilt observations from one site. Better control can be obtained from a profile of observations. The tilt Green's functions show that the tilt at Rawdon is most sensitive to earth structure near the Moho. A station between Rawdon and the Minas Basin would provide more information on the near-surface structure, and a further measurement 100 km. from the Bay of Fundy would delimit upper mantle structure.

Additional observations will also demonstrate whether laterally homogeneous models are realistic. Lateral inhomogeneity may account for the apparent difference between the north-south and east-west models. The effects of lateral inhomogeneities can be investigated using the finite element method. The most obvious lateral inhomogeneity in earth structure is the continental margin. Preliminary calculations suggest that changes in crustal structure at the continental margin could not be detected at Rawdon. However, a lateral change in elastic properties of the upper mantle would have a slight effect. The tilt Green's functions show that beyond 100 km. from the load, tilts are insensitive to crustal structure. Consequently, the models represent a weighted average of the crustal structure over a 100 km. radius area surrounding Rawdon.

A more reasonable explanation for the differences between north-south and east-west models lies in our poor knowledge of the Atlantic Ocean M_2 tides, which contribute 30 per cent of the tilt at Rawdon. The tilt and gravity observations from the Atlantic coast stations will place constraints on the tidal distribution.

MODEL	T_{N-S}	T_{E-W}	STRUCTURE
OBS.	24·6	19·3	
B	21·7	20·5	RIGID CRUST
J	28·9	24·9	GRANITE CRUST
L	25·5	23·0	SOFT CRUST
G	24·3	21·9	ATLANTIC COAST A
M	24·7	22·6	ATLANTIC COAST B
N	20·8	19·6	GULF OF ST. LAWRENCE

Table 1. Comparison of observed and model tilts for Rawdon. T_{N-S} and T_{E-W} are the north-south and east-west tilts in milliseconds of arc.

The results are significant for two reasons. Firstly, tidal-loading tilt observations have been made with sufficient precision that realistic earth models are required to explain them. The results owe much to the large amplitude tides in the Bay of Fundy and it is not claimed that measurements could be repeated with sufficiently high precision in other areas of the world at present. However, given better instrumentation tidal-loading measurements could complement seismic refraction and surface wave studies. Secondly, the response of realistic elastic earth models to surface loads can now be calculated. Farrell's results are applicable to laterally-homogeneous whole-earth models, whilst the finite-element method can simulate complicated structures in the near field. A self-gravitating finite-element earth model could simulate loading on an arbitrarily complicated earth.

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(Earth Science Symposium on Offshore Eastern Canada, Geol. Surv. Can., Paper 71-23, 1973, p. 167-194)

10. RECENT SAND SPILL-OVER OFF SABLE ISLAND BANK, SCOTIAN SHELF[1]

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Abstract

Sediment movement off Sable Island Bank to deeper water on the outer Scotian Shelf during Holocene to present time is proposed on the basis of bottom sampling and photographic investigations in this region of the Canadian Atlantic margin. The transfer of sediment from the outer shelf to the slope, rise and abyssal plain beyond is called spill-over. Two events that most affected off-shelf sedimentation on this continental margin during late Quaternary time are glaciation and eustatic sea level oscillations. As sea level attained its near-present position and the actual configuration of bottom currents was established on the drowned bank surface, texturally modified relict (or palimpsest) sands began the present pattern of movement on Sable Island Bank. This bottom current activity has resulted in spill-over of lag sands off the bank, and deposition of thin discontinuous layers (including some turbidites) on the slope and rise south of the bank, and in the Gully Trough and The Gully north and east of the bank respectively. Emplacement of spill-over sands in historic time is suggested, and bottom photographs show that off-bank movement is continuing at present. Cores collected seaward of Sable Island Bank indicate that terrigenous sand supplied by this submerged platform not only accumulates on the slope, but has, in some instances, reached the rise and even the Sohm Abyssal Plain hundreds of kilometres to the south.

[1] Editor's Note

The editor has been obliged to change various geographical names in this article to conform with those approved by the Canadian Permanent Committee on Geographical Names on the recommendation of its Subcommittee on Undersea Features. The geographical names used by Dr. Stanley and his co-authors in this and earlier articles were, with the official Canadian name in parentheses, as follows:- Banquereau Bank (Banquero Bank), Kapuskasing Canyon (Verrill Canyon), Nova Scotian Shelf (Scotian Shelf), Sable Island Canyon (Logan Canyon), Sackville Canyon (Dawson Canyon), and The Gully Canyon (The Gully).

INTRODUCTION

Geologists have long recognized the importance of off-shelf sediment transport that occurred during Pleistocene low stands of sea level when rivers and wind, and ice in higher latitudes, were able to move sediment across emerged platforms directly onto the uppermost slope. The role of present day submerged continental shelves as sources for coarse sediment transported across the shelf-break to the slope and beyond is considerably less well documented. There is ample evidence that both fine and coarse sediments, at least locally, are now bypassing the shelf-break of narrow and generally tectonically-active platforms such as those off California (Moore, 1969) and the circum-Mediterranean area (Heezen and Ewing, 1955; Stanley et al., 1970). However, much less information is available on modern sedimentation dispersal patterns on the outermost sectors of broad shelf regions located far from landmasses. An example in point is the outer shelf Atlantic margin off northeastern North America.

This study, an outgrowth of a broader survey of late Quaternary history and progradation of the outer continental margin off Nova Scotia (Stanley et al., 1971), is a discussion of modern sand transport off Sable Island Bank. This off-shelf transfer of sand and silt is called spill-over. The purpose of this paper is threefold: to provide both direct and indirect evidence of the lateral movement of coarse graded material from this sector of the Scotian Shelf to the slope, rise and Sohm Abyssal Plain to the south; to describe petrological attributes of this sandy spill-over facies; and to interpret these observations. Modification of the surface and margin of Sable Island Bank by marked glacial events and by erosion and deposition associated with late Pleistocene to Holocene shoreline fluctuations (James and Stanley, 1968; King, 1970) has directly influenced the spill-over pattern presently observed. A review of these late Quaternary events is essential, therefore, in a discussion of spill-over.

Geographic Setting of the Study Area

Sable Island Bank is an elliptical sand-covered bank covering an area of about 17,000 sq. km on the outer Nova Scotian Shelf (Fig. 1). The bank is approximately 250 km in length and 115 km in maximum width (when using the 90 metre or 50 fathom isobath as bank boundary). Sable Island, located on the east-central sector of the Bank, lies about 185 km south-southeast of Cape Canso and 334 km southeast of Halifax on the Nova Scotian mainland. Gully Trough, an elongate depression with depths of up to 200 m, is oriented roughly east-west. It separates Sable Island Bank from Middle Bank to the north. Gully Trough, displaying a complex and irregular topography, deepens abruptly to form The Gully east of Sable Island Bank. This impressive canyon, the largest such feature on the outer Scotian Shelf, separates Sable Island and Banquero Banks (Marlowe, 1967, 1969; Stanley, 1967).

The seaward (southeastern) margin of Sable Island Bank forms the ENE-WSW trending shelf-break and slope; the transition from shelf to continental slope occurs within a distance of 2 to 4 km (Fig. 2). The slope bordering the seaward edge of the Bank is approximately 240 km in length and extends from the southwest margin of the Bank (at about 62°00'W, 43°00'N) to The Gully (at about 59°07'W, 43°45'N).

The break between shelf and slope (gradient > 1:40) occurs at depths ranging from 110 to 146 m, but is found most frequently between 119 (65 fm)

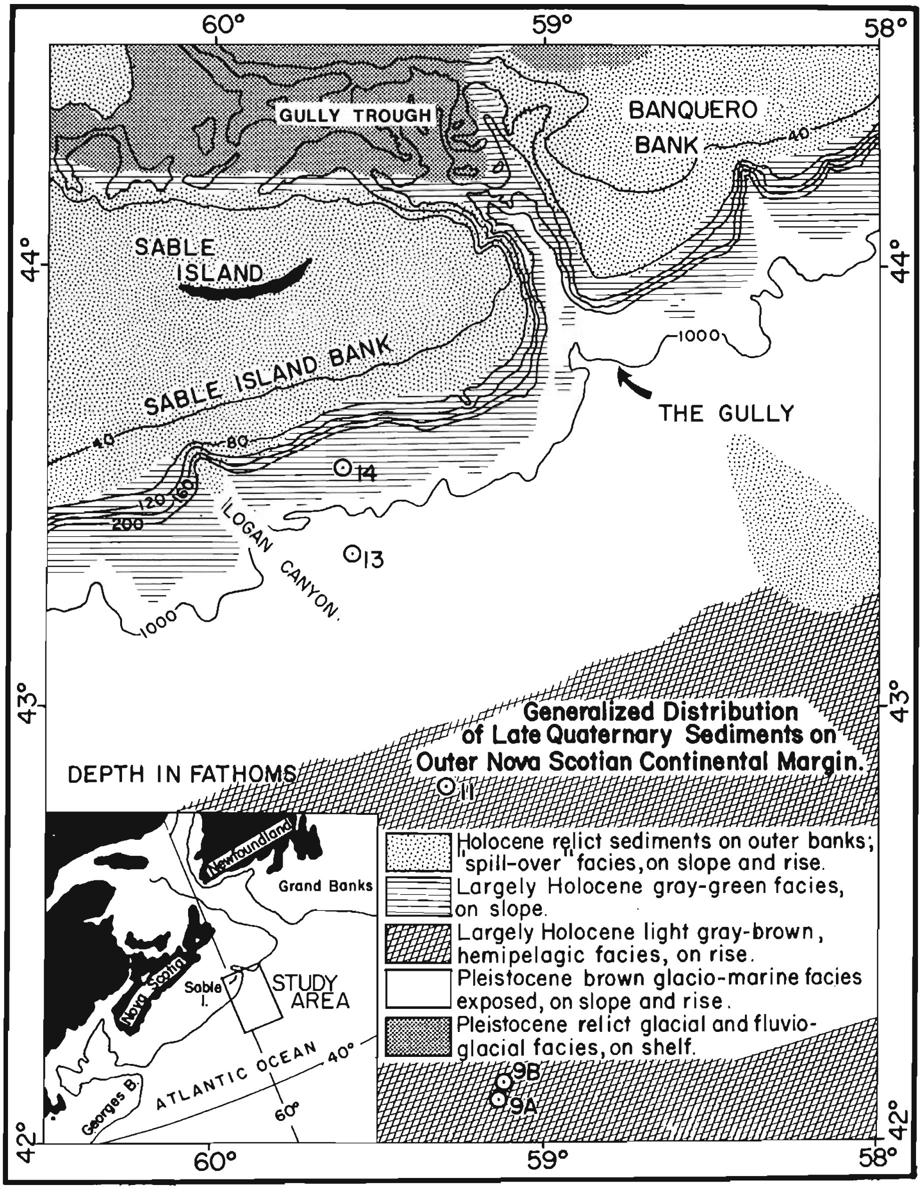

Figure 1. Map showing physiographic features and generalized distribution of late Quaternary sediments on the outer continental margin off Nova Scotia discussed in text (Stanley and Silverberg, 1969). Depths in fathoms.

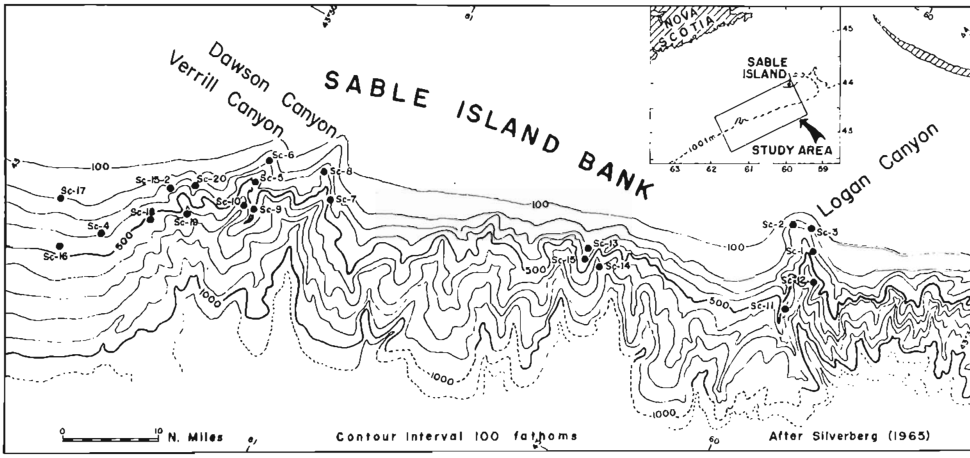

Figure 2. Dissected Nova Scotian continental slope southwest of Sable Island (modified after chart by Silverberg, 1965). Note smooth, non-scalloped slope west of the Verrill Canyon, in contrast with areas between Dawson and Logan Canyons. Depths in fathoms.

and 137 m (75 fm). The shelf-break occurs at similar depths along most of the outer margin between Newfoundland and the area south of New England (Uchupi, 1968), and these depths coincide closely with the maximum drop in sea level during Wisconsinan time (Milliman and Emery, 1968; Stanley et al., 1968). Gradients from 1:10 to 1:25 characterize the upper slope (Fig. 2). Only three large depressions actually head on the outer shelf margin at depths of less than 200 m in the region west of The Gully: (a) Logan Canyon, about 5 km wide at approximately 60°03'W, 43°35'N (almost due south of Sable Island); (b) Dawson Canyon, about 4 km wide at 61°09'W, 43°19'N; and (c) Verrill Canyon, about 3 km wide at 61°17'W, 43°16'N.

The slope becomes considerably dissected below 500 m: more than twenty large NNW-SSE trending depressions 1 to 8 or more kilometres across are noted on the charts between 59° and 61°W Long. Cross-sectional profiles oriented parallel to the slope indicate that most valleys tend to be U- rather than V-shaped (Stanley and Silverberg, 1969). Gradients along the top of intervalley ridges are approximately 1:20; that of valley axes are commonly about 1:12. Submarine valleys and gullies are straight to slightly sinuous, and most extend to the base of the slope. Some valleys appear to lack tributaries but this may be due to an insufficiently-tight sounding net. Relief of several larger depressions exceeds 500 to 700 m on the upper continental rise, but none of the valleys appear to extend as far as the Sohm Abyssal Plain to the south (Pratt, 1967). Certain valleys such as Verrill Canyon, heading near the shelf-break, appear to die out near the base of the slope. It is noteworthy that most valleys head at depths greater than 400 to 700 m, and in some cases below 900 m. Large broad mounds, some of them covering an area exceeding 25 sq. km, occur beyond the distal termination of gullies at base-of-slope depths.

It is difficult to determine the exact boundary between slope and rise since the decrease in gradient from 2° to less than 1° is gradual. The upper limit of the rise is somewhat steeper and deeper (1,800 to 2,700 m) than that

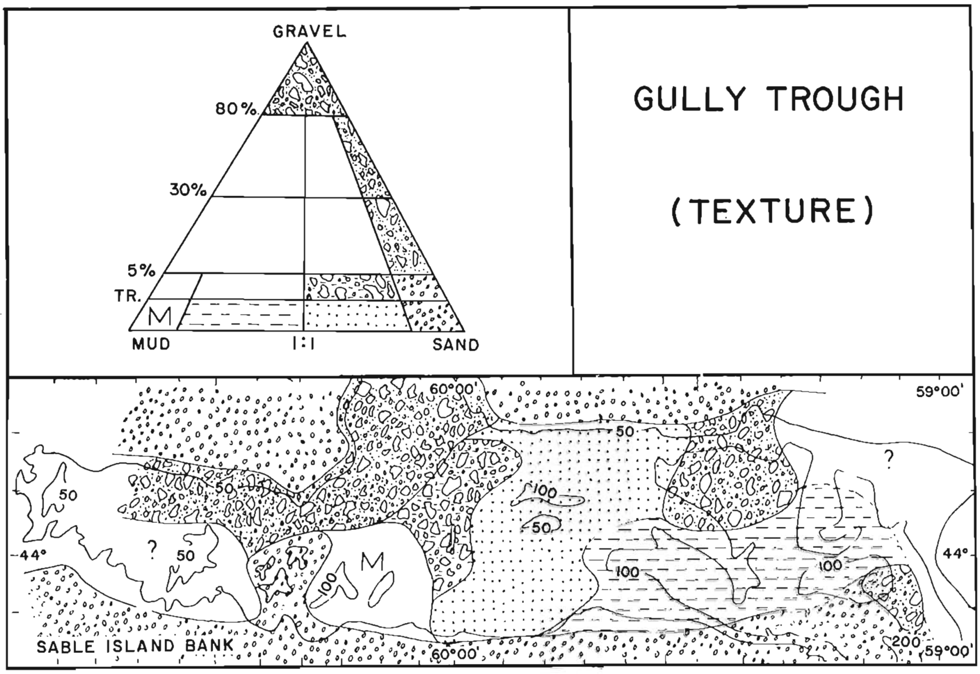

Figure 3. Textural distribution of surficial sediment in Gully Trough north of Sable Island Bank (classification after Folk, 1954). Note tongues of sandy sediment draping from bank into this depression. Depths in fathoms.

off some areas off the eastern United States (Heezen et al., 1959; Pratt, 1967). The lower rise merges with the Sohm Abyssal Plain. The combined width of the slope and rise southeast of Sable Island Bank is approximately 180 nautical miles (334 km).

Sediment Facies on the Outer Margin

General Distribution

Sand and sand-gravel admixtures are dominant textural types on Sable Island Bank as determined by grab sample and camera surveys of the bank and adjacent area (James and Stanley, 1968, Fig. 5; King, 1970). The sand-mud textural limit (commonly referred to as the "mud line") does not coincide with the shelf-break nor does it parallel isobaths on the upper slope. Sand, in fact, appears to drape irregularly from the bank surface into the Gully Trough to the north (Fig. 3) and The Gully to the east (Fig. 4), and onto the slope to the south.

Recognition of sedimentary facies on the continental slope and rise south of the bank is based on a petrological study of four sets of cores detailed in Stanley et al. (1972). The cores examined include:

(a) 21 piston cores (referred to as Sc cores; position shown in Fig. 2) collected on the slope adjoining the southern margin of Sable Island Bank and described by Silverberg (1965);

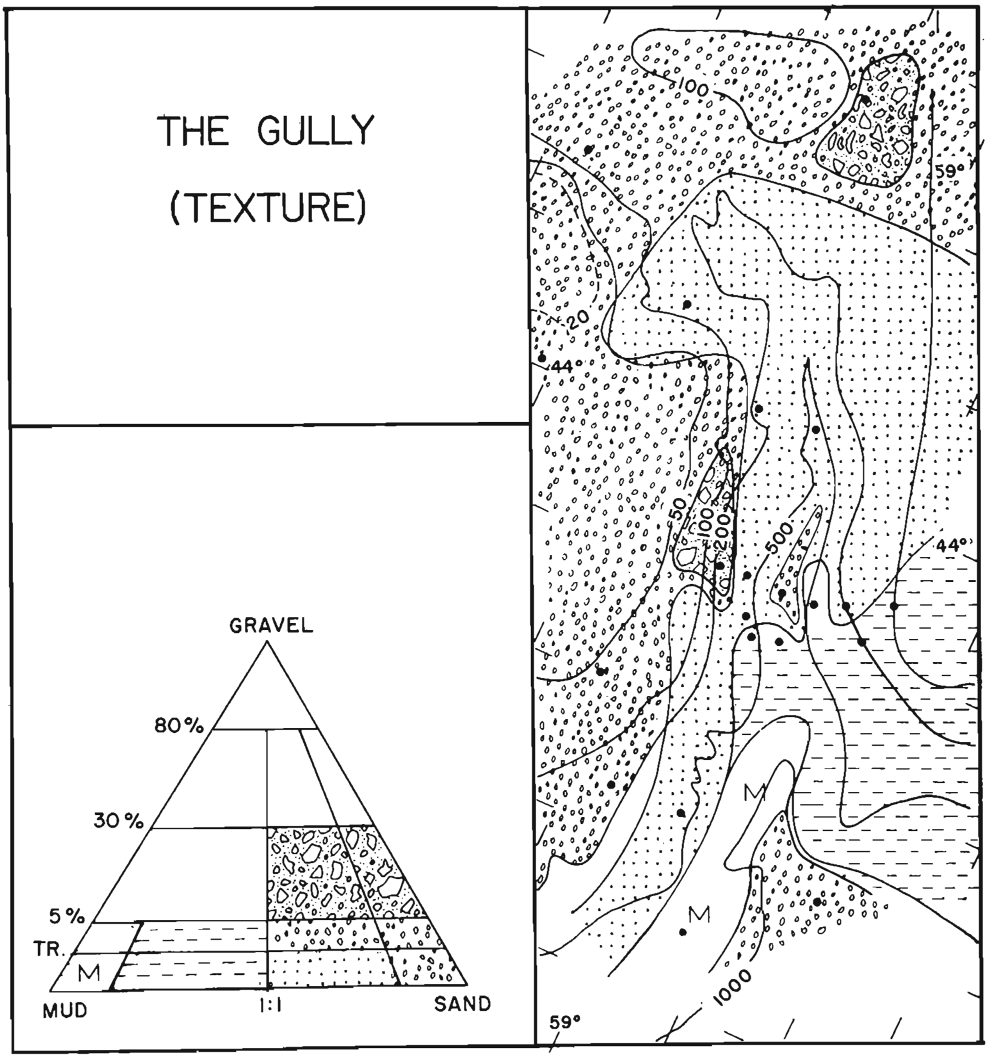

Figure 4. Textural distribution of surficial sediment in The Gully east of Sable Island Bank (classification after Folk, 1954). Note sand draping from the bank onto the eastern wall of The Gully. Depths in fathoms.

(b) 5 piston cores (referred to as HUD-30 cores; position shown in Fig. 1) collected by Stanley and Swift on the CSS HUDSON in 1964 along a north-south transect from the slope due south of Sable Island to the lower rise;

(c) cores collected by the Bedford Institute of Oceanography in the vicinity of The Gully and adjacent slope and detailed in Marlowe (1964), James (1966), Stanley (1967) and Stanley and Silverberg (1969, Fig. 3);

(d) 30 cores collected by the Lamont-Doherty Geological Observatory on the slope and rise off Nova Scotia, and examined by Sutton (1964; in Stanley et al., 1972, Fig. 17). Selected Lamont-Doherty cores collected in this area have also been described by Ericson et al. (1961), Hubert (1964), Conolly et al. (1967) and Hubert and Neal (1967).

Analysis of grab samples collected in Gully Trough shows that the grain size distribution of surficial sediment is, at least locally, depth dependent [data is plotted on a Folk (1954) textural triangle]. At depths of less than 80 metres, sediment is dominantly sand; samples below 180 metres in the trough contain a predominant mud (silt and clay) fraction. Locally, sand drapes from Sable Island Bank well into this topographic low (Fig. 3). The textural type is generally more varied between 80 and 180 metres: sand and gravel occur between 80 and 140 metres; sand, gravel and mud between 120 and 160 metres; and sand and mud between 160 and 180 metres. There is some overlap of these textural zones. Small bank tops and isolated hummocks within Gully Trough are covered with sand. Intermediate depths contain sand and gravel with or without a mud admixture, and below 160 metres, as well as in isolated basins, mud is prevalent. Sediment on the western section of this region, between Banquero and Sable Island Banks, comprises gravel to a depth of 175 metres and clean sand in water deeper than 260 metres.

An examination of grab samples and cores collected in The Gully indicates that sediment consists mainly of sand and mud with minor amounts of gravel and cobble-sized particles (Marlowe, 1964; Stanley, 1967). Grain size data, plotted using a Folk (1954) textural triangle (Fig. 4), show marked local variations superimposed on the general trend of decreasing grain size and increasing silt and clay with increasing depth (James, 1966). Sable Island Bank above 200 metres and the upper reaches of The Gully are composed primarily of sand or sandy gravel. Most of the northern portion and the west wall of the canyon are covered with muddy sand. Muddy sand also predominates at a depth of 900 m on the east wall of the canyon and again at a depth of 2,860 m in the axis. A zone of sandy mud drapes from the top of Banquero Bank, down the east wall, and across the mouth of the canyon. Clean sand is found at 1,400 m in the axis of the canyon. Mud is predominant west of the axis near the mouth of the canyon.

The predominant surficial textural type on the slope and rise south of the bank is mud (i.e., silt and clay, largely of terrigenous origin) with varying, but minor, amounts of gravel, sand and coarse silt. Muddy sediments generally display either an olive-grey or brown to reddish brown coloration. Thin layers of clean sand, and occasionally sandy gravel and gravel, are also noted in cores on the slope. Sand strata, most of them laminated but some of them graded turbidites, are also noted in cores collected on the rise and Sohm Abyssal Plain.

Stratigraphy of Late Quaternary Facies

Determination of the late Quaternary stratigraphic sequence in the study area (cf. Stanley et al., 1972) is based on examination of the four sets of cores referred to in the previous section as well as to earlier studies by other workers (Ericson et al., 1961; Heezen and Drake, 1964; Conolly et al., 1967). Olive-grey mud of late Pleistocene to Holocene age generally overlies the brown facies of Pleistocene age sequence (this sequence was also recorded by Heezen and Drake, 1964; Conolly et al., 1967) but exceptions to this are noted. To contact between the two sediment types can be sharp or gradational. Interbedded olive-grey and brown sediments are noted in some cores. The olive-grey mud is generally less than 2 to 3 m thick. This facies began to accumulate 15,000 to 20,000 years ago (late Pleistocene) and continued through much of Holocene time, and covers an older reddish brown Pleistocene unit on much of the upper- and mid-slope region. Occasional thin partings of sand occur within the olive-grey facies, but more commonly within the brown facies. The third and stratigraphically youngest sediment type, noted in some of the cores collected on the slope, is a relatively clean sand unit that occurs in thin laminae (2 to 15 cm) at the uppermost surface. Somewhat thicker layers of sand have been cored in The Gully (Marlowe, 1964; Stanley, 1967). It is this sand facies of recent age that will be emphasized in the present study. A fourth facies, observed on the lower rise (Fig. 1, cores HUD 30-9A, B), is a pale yellowish brown mud of Holocene age that covers the Pleistocene reddish brown facies. The Holocene as well as Pleistocene units on the lower rise contain sand layers including laminated and graded strata.

A simplified sediment distribution chart of the study area based on the coring program is shown in Figures 1 and 16. The olive-grey mud and older brown mud-sand-pebble facies appear to be irregularly distributed on the slope surface. The sand facies, whose distribution is even more irregular, most often appears in tongues that locally drape on the margins of Sable Island Bank. The patchy late Quaternary facies distribution is, in part, a function of topography (Stanley and Silverberg, 1969, Figs. 5 and 7).

Petrological Characteristics of Modern Spill-Over Sands

The uppermost thin sand facies mapped seaward of Sable Island Bank is easily distinguished from older olive-grey muddy sediment and brown mud-sand-gravel facies that lie beneath it in cores. Radiographs show that these sands are either laminated or contain bands of closely-spaced fine mottles (probably burrows in most cases) with scattered distinct coarse mottles (probably granules, shell hash and small pebbles). Textural analyses of the surface sand facies show that it occupies the sand and silty sand sectors on textural triangle diagrams (Silverberg, 1965). This is in contact with samples of the red-brown Pleistocene facies that occupy mainly sand-silt-clay, sandy silt, and silty sand positions on the triangle and those of the olive-grey facies that generally comprise silt and clayey or sandy silt. A log showing the textural variation within an upper slope core (Sc-9) that penetrates the spill-over sand facies above the brown Pleistocene section is shown in Figure 5.

Tan and yellowish brown Holocene sediment on the lower rise generally consists of well-sorted very fine, silty clay, clayey silt and silt and is fairly uniform in texture. Some of the sand strata in this base-of-slope environment are recognized as turbidite layers (one such layer, for instance, was noted in Core 30-9A).

Textural analysis of samples from 30 Lamont-Doherty cores collected on the slope, rise and Sohm Abyssal Plain south of Nova Scotia shows that these outer margin core samples are generally finer grained and less well-sorted than adjacent shelf deposits. Sandy sedimentation units on the slope and upper rise are generally coarser grained than those on the lower rise and Sohm Abyssal Plain. However, there are no obvious differences that are strictly related to water depth (Stanley et al., 1971, Table 1). Sorting

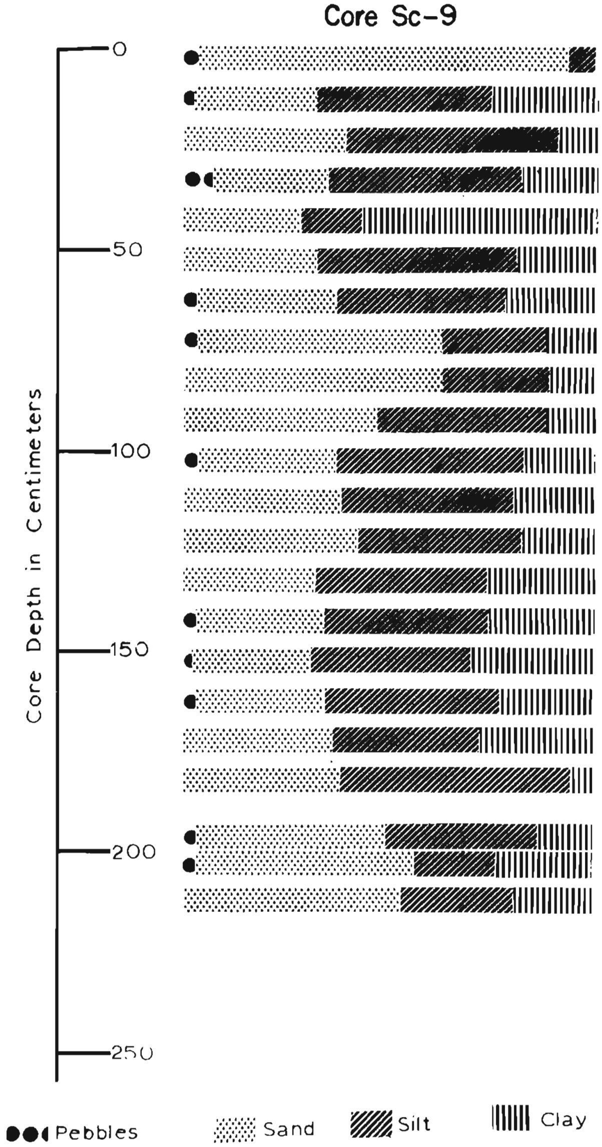

Figure 5.

Textural variation within an upper slope core (Sc-9, see location in Fig. 2). The thin surficial spill-over layer is markedly more sandy than the much older reddish brown section above which it lies.

and kurtosis values of sandy slumped (contorted) sediments and of turbidites (graded) are different from those moved and modified by marine currents (laminated). Variation in grain roundness, on the other hand, does not appear to bear any relation to the environmental province or mode of origin, but probably reflects the character of the original sediment sources (in this case, Sable Island Bank).

Cumulative frequency curves, plotted on a probability scale (Fig. 6), show the size class spread and indicate a generally poorer sorting of the

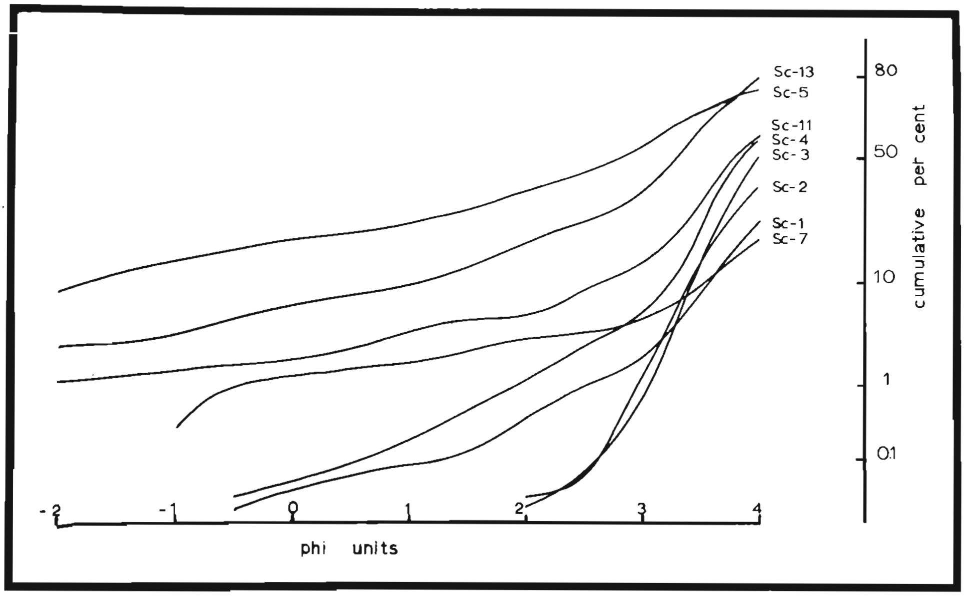

Figure 6. Cumulative coarse fraction size curves of surficial sediment collected at core locations on the upper and middle continental slope south of Sable Island Bank (see Fig. 2 for Sc core locations). Spill-over sands (Sc-5 and -13) are markedly coarser than sandy stringers in older grey and brown facies.

sandy surficial slope sediment than that of the adjacent bank sands. On the upper slope, relatively clean surface sand layer samples (Sc-5 and Sc-13) are relatively coarse grained and show a small secondary mode in the coarse sand to fine pebble range; sorting is poor. Coarse layers within the older grey facies generally comprise predominantly very fine sand with only traces of coarse sand. Samples Sc-2 and Sc-3 are somewhat better sorted but contain less sand (the modes lie in the silt range). In general, coarser grained sediments are more poorly sorted than finer grained ones.

The light mineralogical composition of the surficial sand spill-over facies is shown in Figure 7. Quartz, the dominant component of the sand-size fraction, occurs in several forms. Many of the grains are iron-stained and the surfaces may be smooth, pitted, or frosted; they are similar to those observed on the bank surface. An almost complete gradation from well-rounded to very angular grains is found in most samples. The dominant accessory component consists of terrigenous grains, particularly rock fragments. Other common accessory components include feldspar, mica, shell material, and foraminifera. Glauconite (generally dark green, somewhat knobby grains with a smooth or finely pitted surface), dark terrigenous grains, and carbonate "hash" are additional components which may also be present. Mica, generally in low amounts, occurs mostly as flakes of muscovite; small amounts of biotite and phlogopite are noted. Fragments of red, green and grey siltstone, red and grey sandstone, quartzite, mica schist, quartz-biotite gneiss, basalt, and granite are the principal lithologies comprising the lithic fraction.

Shell material consists of fragments in various stages of destruction. The bulk of shell fragments are relatively fresh although some of the larger smooth and rounded grains show evidence of abrasion. Foraminiferal tests make up much of the shell component; many broken individuals are noted in grain counts. Carbonate hash is a category including aggregates of a pale,

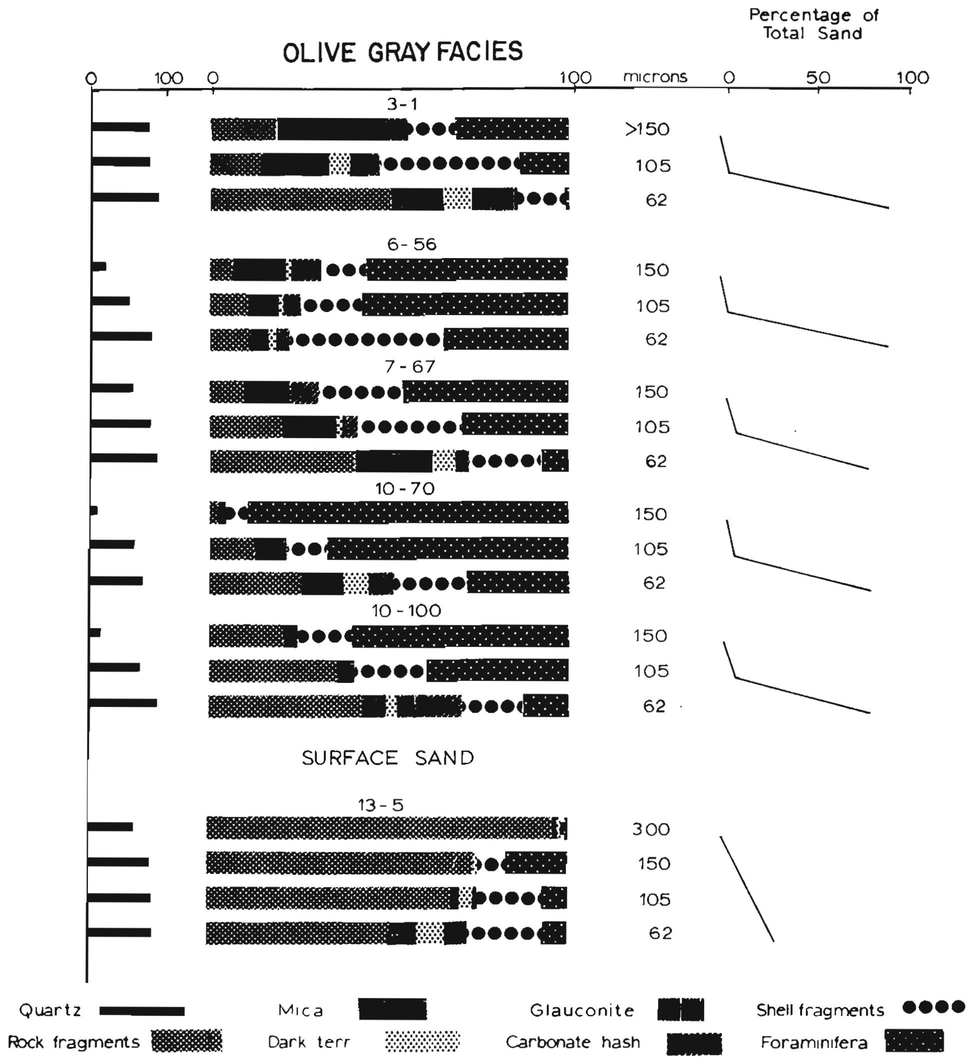

Figure 7. Light mineral composition of surface (spill-over) sand facies and that of the coarse fraction of older olive-grey facies on slope off Sable Island Bank. Sample numbers refer to Sc cores shown in Figure 2. Quartz is dominant mineral in both facies, but composition of accessory minerals (percentage excluding quartz) shows that surficial sands contain higher percentages of lithic fragments.

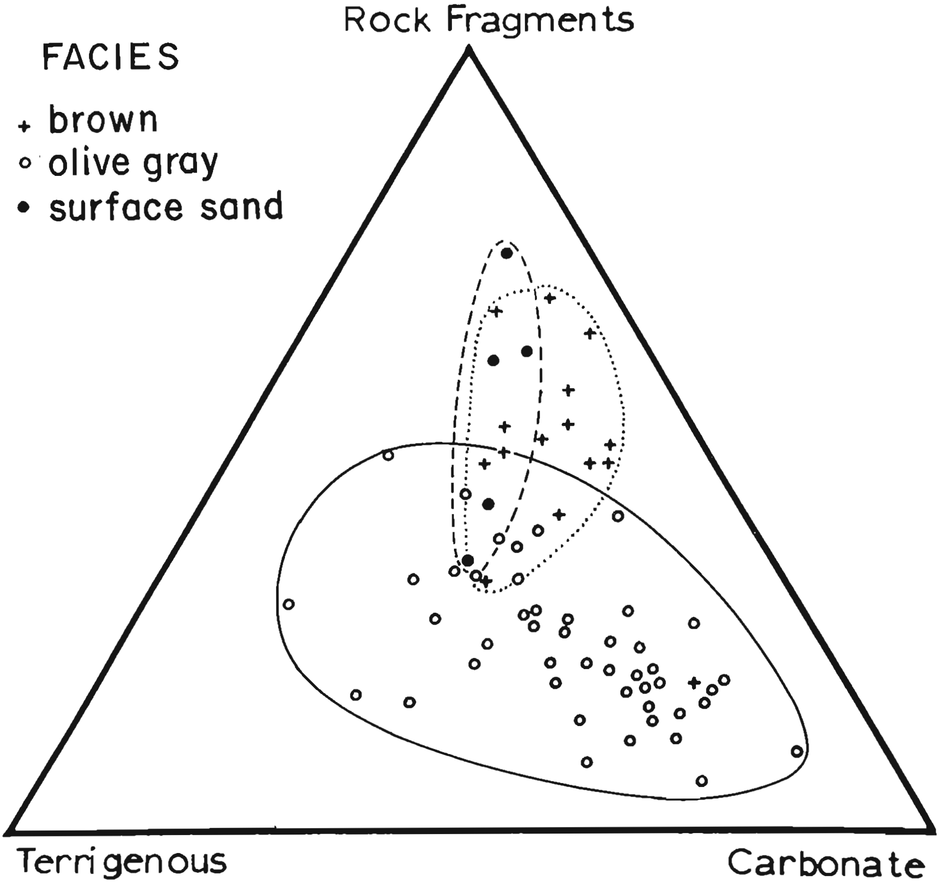

Figure 8. Cross mineral composition of the surface (spill-over) sand facies and coarse fractions of older grey and brown coarse facies on the continental slope. Samples from Sc cores shown in Figure 2.

greenish flaky material, fine shell fragments, and occasionally scattered quartz grains, all loosely cemented by carbonate material. The flaky material is probably an altered clay product.

Similarity between the accessory light mineral composition of the upper sand facies on the slope and that of the sand fraction within the much older (Pleistocene) brown facies below it (Fig. 8) is noteworthy. The composition of both of these facies, in turn, is similar to that of Sable Island Bank as described in James and Stanley (1968) and Cok (1970).

The heavy mineral assemblages of the surficial sand layer on the slope and rise are also similar to those mapped on Sable Island Bank although the proportion of minerals differs. On the bank the relative per cent of garnet generally exceeds that of hornblende (James and Stanley, 1968; Cok, 1970), while on the slope the relative per cent of hornblende is invariably higher than garnet (Fig. 9). This difference can probably be attributed to size sorting (e.g., generally finer sands on the slope than on the shelf would contain a lower amount of garnet, an inherently large mineral).

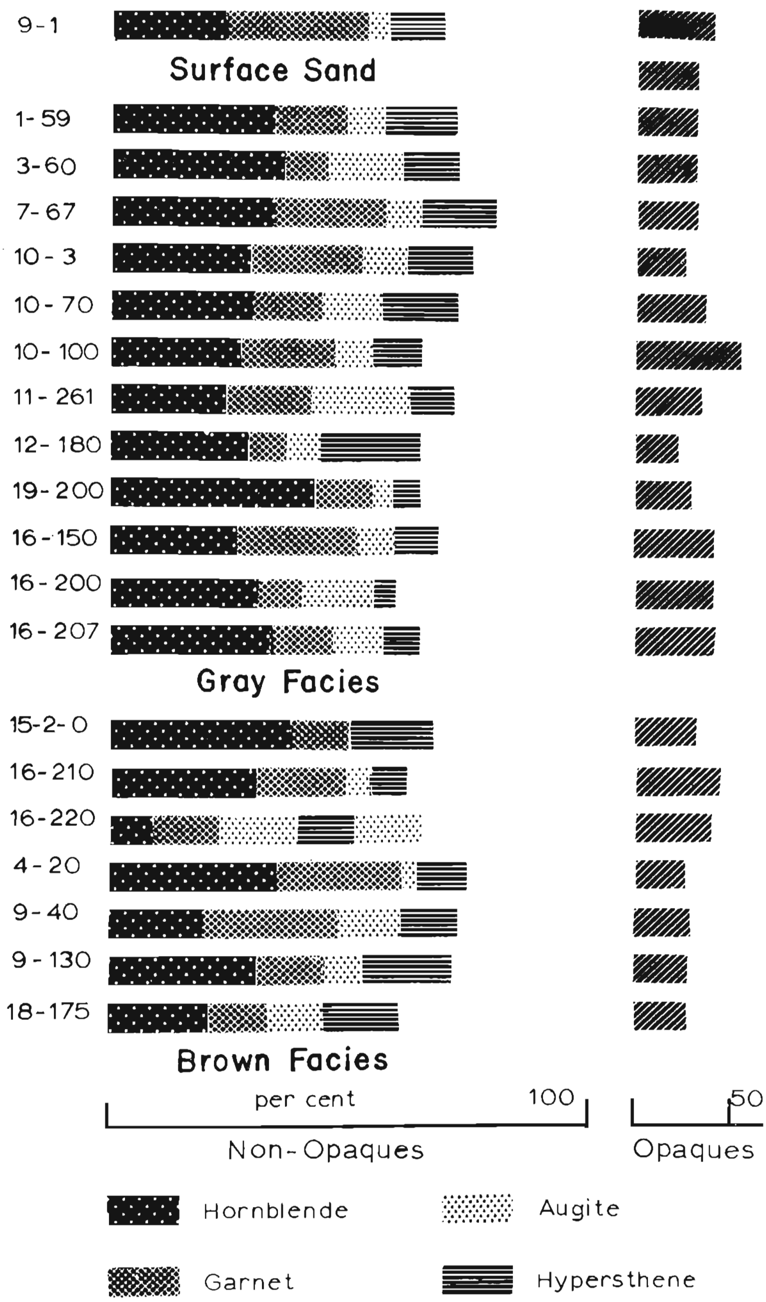

Figure 9. Principal heavy minerals suite in surface spill-over sand, and older grey and brown sediment facies. Sample numbers refer to Sc cores collected on the Scotian continental slope shown in Figure 2.

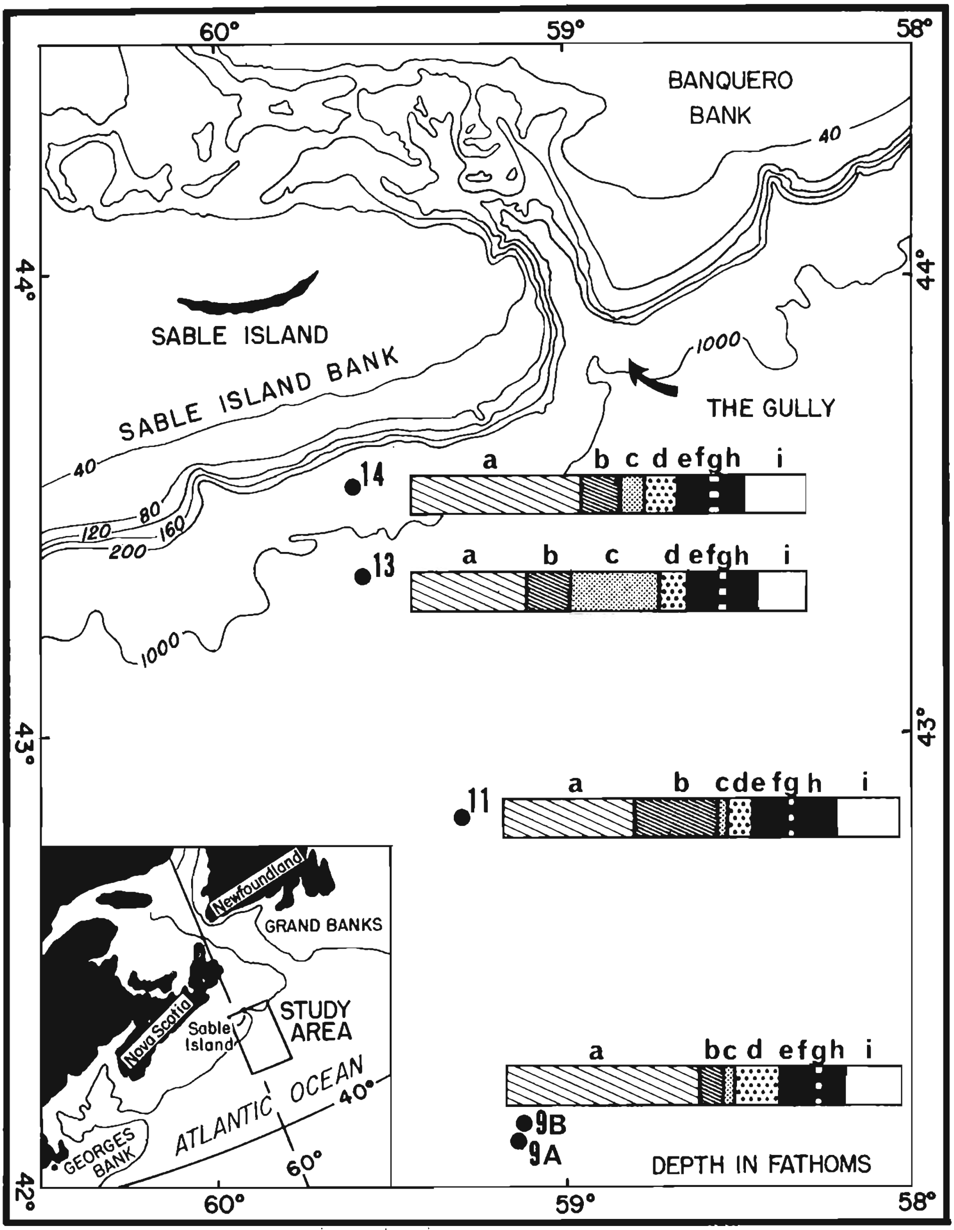

Figure 10. Relative percentages of transparent heavy minerals observed in core (HUD-30) samples south of Sable Island Bank. a, hornblende; b, garnet; c, alterite; d, epidote; e, zircon + tourmaline + rutile; f, augite; g, hypersthene; h, metamorphic suite (staurolite, kyanite, etc.); i, other minerals. Depths in fathoms.

Sand on the lower rise (Fig. 10) and Sohm Abyssal Plain (Hubert and Neal, 1967; Stanley et al., 1972, Table 3) south of Sable Island Bank also shows a similar suite of light and heavy minerals. In most of these samples, varieties of garnet and amphibole account for approximately half of the heavy mineral suite. In several cases, the amount of garnet exceeds that of hornblende. The proportion is generally related to grain size, i.e., higher amounts of garnet are generally found in coarser grained samples. This similarity between the light and heavy minerals of Sable Island Bank and outermost margin sediments shows the important role played by the shelf as a continuing source of downslope transported terrigenous deposits. Mineralogy and texture of rise and Sohm Abyssal Plain sands indicates that these coarse materials have occasionally bypassed the slope, and can be dispersed hundreds of kilometres to the south of the Bank. Similar conclusions were made by earlier workers such as Ericson et al. (1952) and Hubert and Neal (1967) in their more general regional surveys of Western Atlantic Basin dispersal patterns.

Late Pleistocene to Modern Events that have Resulted in Spill-Over

Quaternary Progradation

The broad shallow banks sited along the outer Scotian Shelf have played a unique role in the sediment dispersal system of the Nova Scotian continental margin. They have served as reservoirs for sediment received under one set of conditions (glacial, glacio-fluvial and glacio-aeolian) during the Pleistocene and subsequently released sediment under a second set of conditions (marine littoral conditions, and deeper marine wave- and tide-agitated environments) during the Holocene (James and Stanley, 1968). This shift in sediment dispersal processes and patterns on the outer banks of the Scotian Shelf through late Quaternary time has resulted in a corresponding sequence of depositional events on the slope and rise. The consanguinous nature of these two sedimentary provinces is clearly indicated by the similarity of mineral suites on Sable Island Bank and in cores on the slope and rise seaward. Seismic surveys show that the late Quaternary history of this outer margin has been one of progradation: a net seaward extension of the slope and rise by sediment accretion (Uchupi and Emery, 1967; Emery et al., 1970). The late Quaternary to recent events that have resulted in spill-over of sand off Sable Island Bank are discussed in the following two sections.

Sedimentation on Sable Island Bank during Subaerial Exposure

During the maximum low eustatic stand of the sea in Wisconsinan time, sea level is believed to have stood at a depth of approximately 110 to 120 metres below present sea level [approximately 17,500 to 20,000 years B.P. according to Curray (1965), and closer to 15,000 years B.P. according to Milliman and Emery (1968)]. During this stage the shoreline coincided closely with the shelf-break, as determined by a study of terraces (Stanley et al., 1968).

Individual tongues of Pleistocene ice approached Banquero and Sable Island Banks but apparently did not over-ride them (Stanley and Cok, 1968). Hence the bank surface received enormous volumes of outwash, much of it sand, during periods of maximum glacial advance. Petrologic investigations indicate that most of the brown to reddish brown stained glacial and fluvio-glacial material preserved on the outer banks, including Sable Island Bank, originated in Carboniferous and Triassic terranes lying several hundred kilometres to the northwest (Bay of Fundy, Prince Edward Island and New Brunswick regions). Some of this material accumulated on the Bank as a thickening outwash plain which locally underwent aeolian modification (James and Stanley, 1968; Medioli et al., 1967). Fluvial processes resulted in deposition of mud to pebble-size material on bank surfaces, and local pockets of gravel encountered on Sable Island Bank are probably relict from this phase. However, much of the brown material must have bypassed the shoreline (the exposed bank margin) thus becoming available for deposition on the slope and beyond.

The coastal margin was dissected with deep bays formed by the heads of the large canyons (Logan Canyon and The Gully, for example) which indent the seaward bank margin. These embayments undoubtedly served as outlets for meltwater. In some cases they may have contained heavily-stratified water bodies with well-developed estuarine circulation, capable of serving as hydraulic traps for accumulating fine sediment (Postma, 1967). During peaks of outwash aggradation, the resulting estuarine clay deposits would be loaded by rapidly growing intra-estuary deltas of coarse sediment (Swift and Borns, 1967). Transfer of all grades from the bank tops onto the steep bank margins was probably accelerated during those phases when ice tongues migrated closest to the outer banks. This is confirmed by an examination of brown mud-sand-gravel sections cored on the slope and rise south of Sable Island Bank (Stanley et al., 1972).

The seaward edge of Sable Island Bank serving as the coast received the brunt of erosion by surface waves, and if meltwater had sufficiently stratified coastal waters, then possibly by breaking internal waves. The relatively steep slope (5° or more) of the outer bank margin meant that storm waves were not damped by a shallow shelf and that the coast was under intensive attack. Slumping of the heterogeneous coastal material would be expected and would provide a large volume of sediment for transfer downslope (Stanley and Silverberg, 1969, Fig. 9).

Sedimentation During and Subsequent to Inundation

In late Pleistocene time sea level began to rise first slowly and then more rapidly (Emery, 1968, Fig. 15). The transgressing surf stripped sediment from the retreating shore face and released this material on the adjacent shelf floor for resedimentation by marine currents (Swift, 1968). This resedimentation process has resulted in a second set of petrographic attributes being printed over the original textures. Most obviously, the finer fractions (silt and clay) have been winnowed out leaving a clean sand lag on the bank surface whose pebbly admixture reflects the original periglacial origin (James and Stanley, 1968, Fig. 5). Such hybrid sediments (reworked relict sediments) have been designated palimpsest (Swift et al., 1971). Processes resulting in the modification of texture have been more effective on the outer banks of the Scotian Shelf than in the deeper central shelf physiographic province.

After water deepened over Sable Island Bank, the blanket of palimpsest sand continued to undergo textural modification and transport according to a new and well-defined system of sediment dispersal. This system has

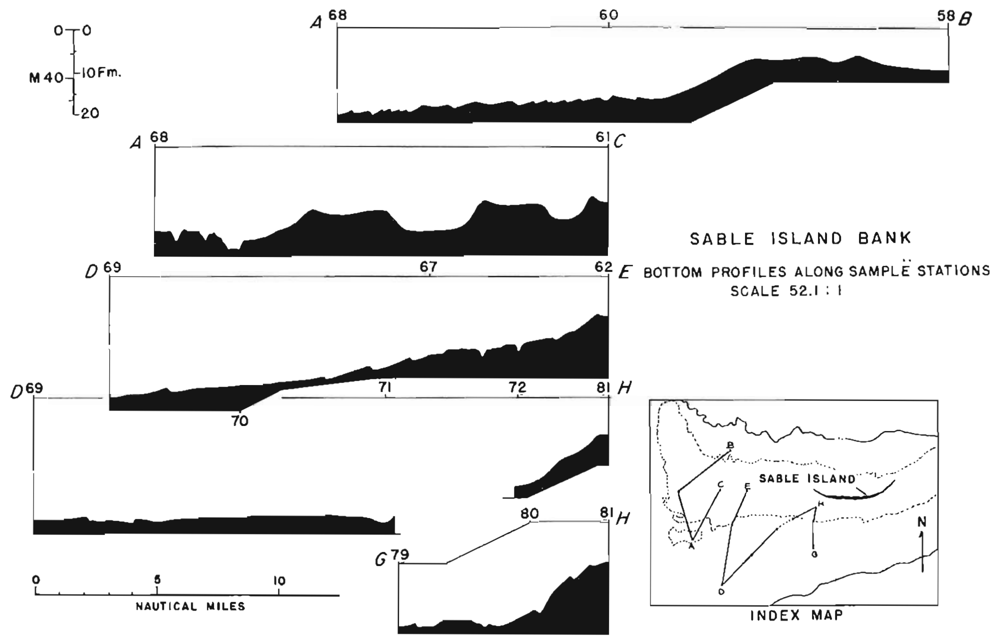

Figure 11. Bathymetric profiles of the surface of Sable Island Bank plotted from PDR and RCA depth sounder records obtained in May 1965. Sand waves and associated bedforms related to intense current activity that molds the bank surface.

been best studied on Sable Island Bank where its elements include Sable Island, the emerged crest of the bank, and the surrounding submarine bank surface.

Sable Island consists of two parallel beach-dune ridges which, with the rise of sea level, have converged towards the crest of the bank until they are contiguous. This sediment reservoir exchanges material with the adjacent shallow sea floor according to a seasonal cycle whereby sediment is accreted to the island in summer by marine processes and is later deflated by strong winter winds and waves (James and Stanley, 1967).

The island was initiated by convergence of the dune ridges in late Pleistocene - early Holocene time (Medioli et al., 1967), but it is presently maintained as a dynamic system of water and sediment circulation. The island and surrounding bank would, in fact, appear to be a circulating sand cell of the type described by Van Veen (1936) from the floor of the southern bight of the North Sea. The distribution of large-scale sand waves indicates a clockwise circulation of residual tidal and wave-driven currents (James and Stanley, 1968). Large sand waves are present between 10 to 70 m depth, both north and south of Sable Island and on the western part of the bank (Fig. 11). Smaller sand waves, averaging 6 to 7 m in height and having wave lengths from 300 to 1,000 m are most abundant south of Sable Island. Their asymmetry indicates a predominant wave migration from east to west, while north of the island, sand-wave asymmetry indicates migration from west to east. A sequence of lines trending north-south across the submerged bar west of the island shows that sand waves in that sector are migrating

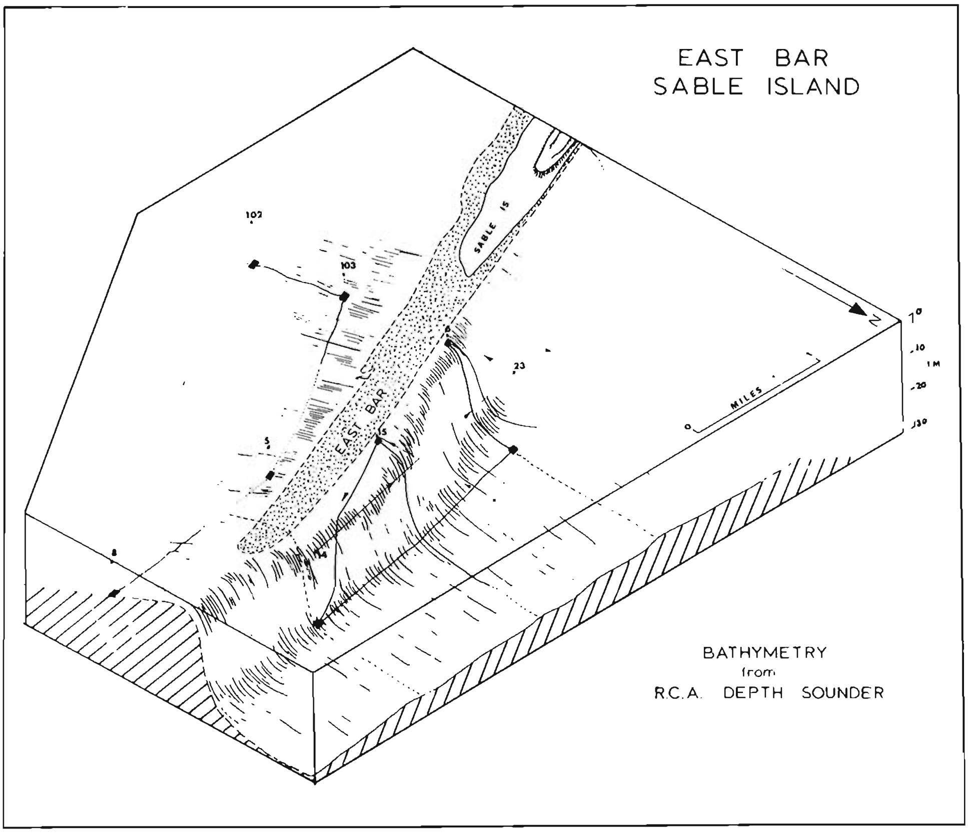

Figure 12. Diagram showing submarine topography in the vicinity of the east terminal bar of Sable Island showing intense reworking of the east-central sector of the bank surface by current activity. The northern face is steeper than the more gentle southern slope (based on RCA depth sounder data obtained in May, 1965).

northward. The east bar of Sable Island (Fig. 12) has been likened by an asymmetric sand wave with a steeper northern face suggesting predominant movement of sand toward the north just east of the island.

Smaller scale structures, such as asymmetric ripple-marks observed on most bottom photographs of the bank surface (Figs. 13 and 14) show vectorial properties that are much more variable than those of sand waves. The regional pattern of these ripples conforms, in a general way, to the sand-wave circulation pattern: ripples off the west end of the island indicate predominantly northward and eastward transport, and those off the east end show southward and westward flow; those on the northwest part of the bank migrate west and north. Station 78 at the southern margin of the bank near the head of Logan Canyon shows a predominant southerly (or off-bank) transport; Stations 8 and 9 indicate off-bank transfer into The Gully.

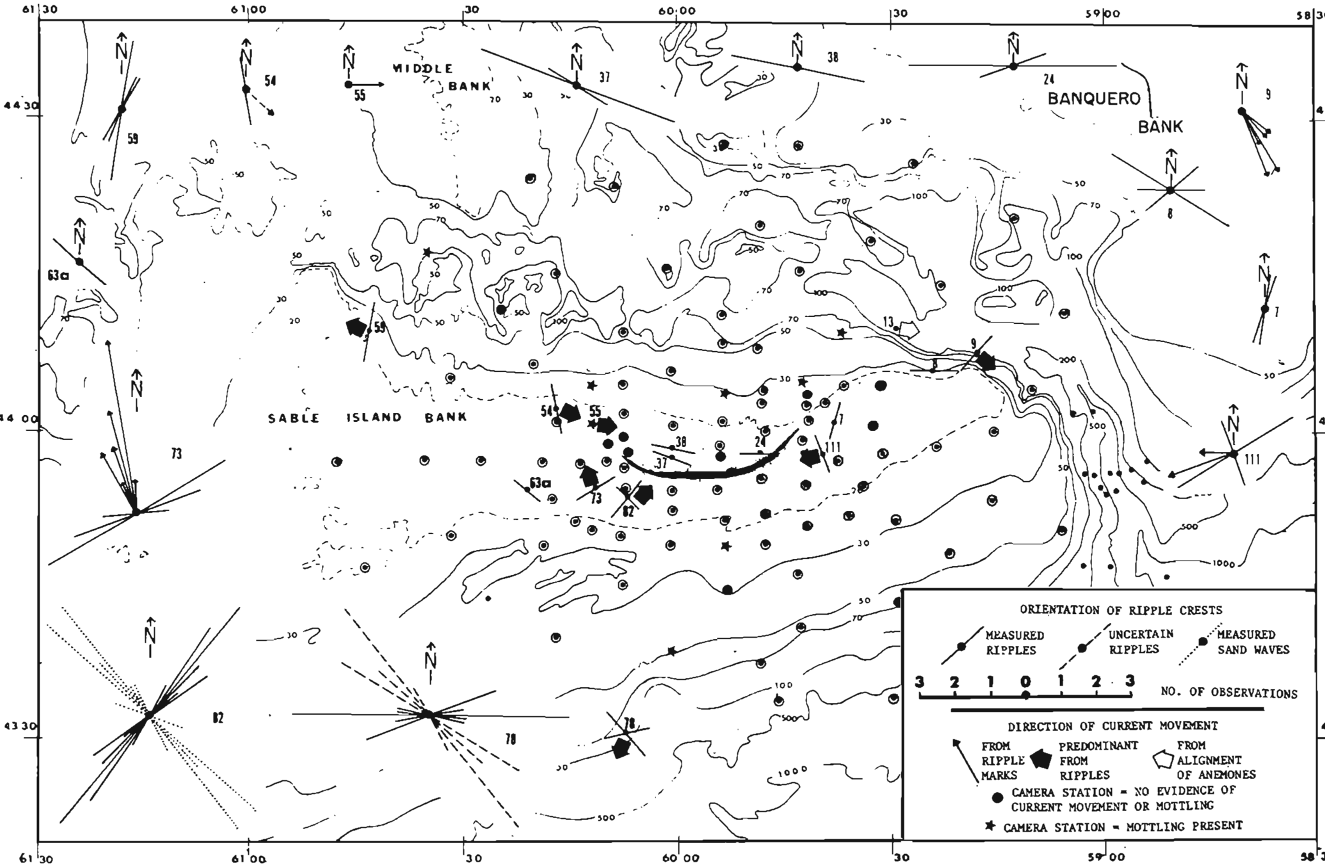

Figure 13. Direction of current dispersal on Sable Island Bank based on features such as ripple-marks observed on bottom photographs collected in May, 1965. Depths in fathoms.

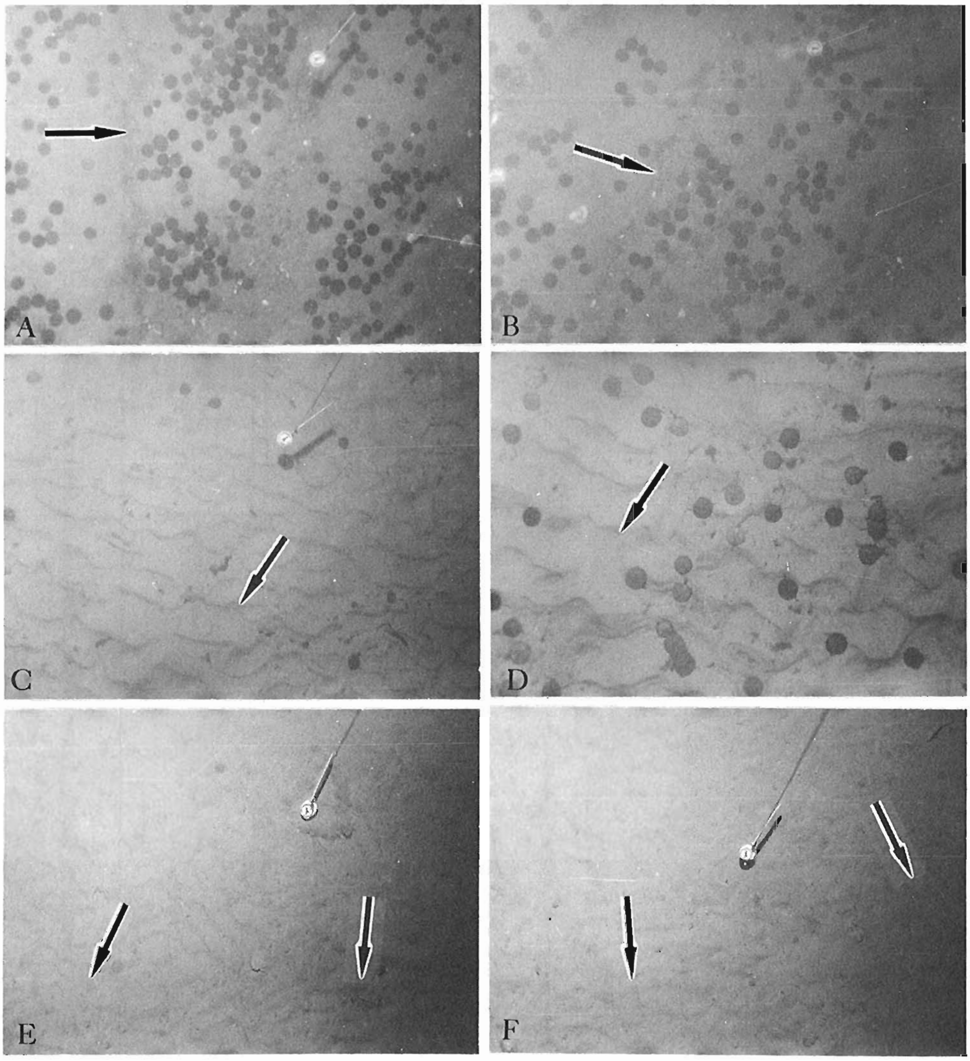

Figure 14. Selected bottom photographs obtained in May, 1965 showing off-bank transport of sand by bottom activity. A and B, (Station 8 on northeast margin of bank; see Fig. 13) sand waves at a depth of 52 m showing dominant current direction toward northeast, or into The Gully. Common sand dollar, Echinarachnius parma (Lamarck) and compass and vane (33 cm) give scale. C and D, (Station 9 on the northeast margin of Sable Island Bank and west wall of The Gully Canyon; see Fig. 13), asymmetric ripple-marks at a depth of 256 m show southeast movement of sand. E and F, (Station 78 at the shelf-break on the southern margin of bank near the head of Logan Canyon; see Fig. 13), rounded (older) irregular asymmetric ripple-marks at a depth of 175 m showing predominant southerly movement of sand.

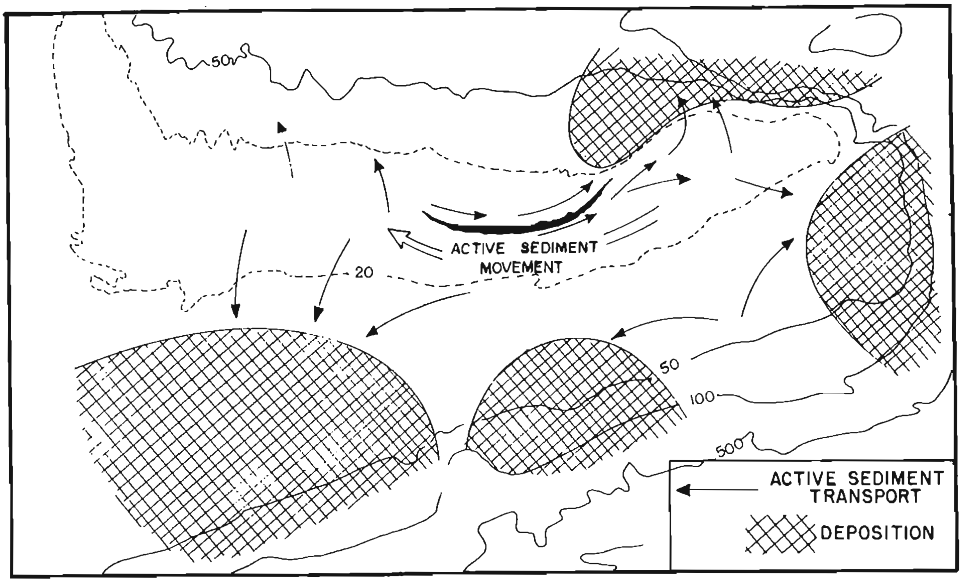

Figure 15. Interpretation of bottom circulation pattern on Sable Island Bank showing dominant sediment dispersal pattern based on bedform and petrological interpretations (after James and Stanley, 1968). Sand moves off the bank surface onto adjacent slope in areas of net deposition. Depths in fathoms.

Radial Dispersal and Modern Sediment Spill-Over

While bedforms indicate primarily a clockwise system of sediment transport, textural and mineralogical patterns on the bank surface (James and Stanley, 1968) reflect a net long-term process of radial dispersal off the bank into the adjacent lows. Transport patterns are clearly marked as orthogonals to isopleths of mineralogy, mean grain size, sorting and skewness. In general, sediment on Sable Island Bank is being transported from central areas of relatively coarse, coarse-skewed sand to marginal areas of relatively fine, fine-skewed sand. These marginal areas tend to be less well sorted than zones where unmixing and transport is most intensive. Sediment-yielding lag deposits are richer in the denser and coarser heavy minerals (opaques, garnet); sediment-receiving areas are richer in the lighter heavy minerals (hornblende, kyanite, and tourmaline). The intensity of iron-staining of relict quartz is assumed to be inversely proportioned to the intensity of abrasion during modern transport; areas of stain-free quartz are areas indicated by other criteria to be areas of active sediment unmixing and transport. A pattern of erosion, transport and deposition on Sable Island Bank emerges upon integration of all data (Fig. 15).

Thus, the present cyclical movement results in a net loss of sediment from the bank surface by transfer of sand-sized material off Sable Island Bank. Bottom photographs showing actual or recent movement of silt and sand by bottom currents on the outermost shelf, shelf-break and upper slope and beyond is direct evidence of spill-over. Retrieval of this sand in bottom grabs and cores is further evidence of off-shelf transport. Additional indirect evidence is available from a study of bathymetric profiles made across the shelf-break (James, 1966). Many of the profiles oriented normal to the shelf-break display a smooth, continuous and very regular convex-up profile extending between the flat surface of the Bank and the upper slope, in some cases to depths of 300 m or more. This smoothing has resulted in part from off-bank transfer of sand that has buried irregularities such as the prominent 65 (119 m) and 80 (146 m) fathom terraces often found along the outer margin (Stanley et al., 1968).

There appear to be several large areas of sand spill-over: south and southwest of Sable Island, from the outer shelf onto the upper slope; north of Sable Island, from the bank margin into Gully Trough; and east of the bank, into The Gully. Evidence for sediment spill-over into The Gully have been presented elsewhere (Stanley, 1967). Relatively clean palimpsest sands are draped onto both east and west canyon walls (Stanley, 1967, Fig. 3B), presumably from the adjacent bank surfaces. Further evidence of spill-over in the shelf-break area south of the bank was cited earlier: sand drapes from the outer shelf onto the upper slope, and cores in the area penetrate a surficial sand layer above older muds.

DISCUSSION

Slumping, turbidity current transport, and ice-rafting were dominant processes on the outer margin off Nova Scotia during the Pleistocene (Brundage et al., 1967; Pratt, 1967; Stanley and Silverberg, 1969; Uchupi, 1969; Emery et al., 1970). The importance of these mechanisms in the downslope transfer of sediment became progressively less important as the ice retreated and rising sea level drowned the continental shelf, thus reducing considerably the supply of sediment transported onto the upper continental slope. However, core and photographic data cited in this report show that Sable Island Bank, even now, is serving as a source of terrigenous material, mostly sand, that accumulates on the slope and perhaps occasionally even on the rise and Sohm Abyssal Plain hundreds of kilometres to the south. Interpretation of sedimentary sections in cores indicates that as Sable Island Bank continued to submerge during Holocene time, the development of its lag blanket of winnowed sand and sandy gravel has approached completion. These reworked and modified glacial and fluvio-glacial materials on the bank surface are recognized as a palimpsest deposit. Progressively, less and less fines (of glacial origin) were generated by littoral and sea floor erosion of the bank surface. Rates of fine sediment deposition on the adjacent slope thus decreased during Holocene to present time while spill-over sands, emplaced as a result of storms that frequently modify the bank surface, became a progressively more prominent part of the upper- to mid-slope section south of the bank (Fig. 16). This is confirmed by an examination of slope cores which reveals, in many cases, a thin cap of relatively clean sand. It is more than likely that many of these have been emplaced in historic time.

As early as 1952, Ericson and co-workers called attention to sands of shallow water origin on the ocean floor of the Western North Atlantic and attributed their presence to turbidity current activity. More recent studies of textures of Atlantic deep-sea sands (Hubert, 1964; Sutton, 1964) and of deep-water mass movement (Heezen et al., 1966; Schneider et al., 1967) have

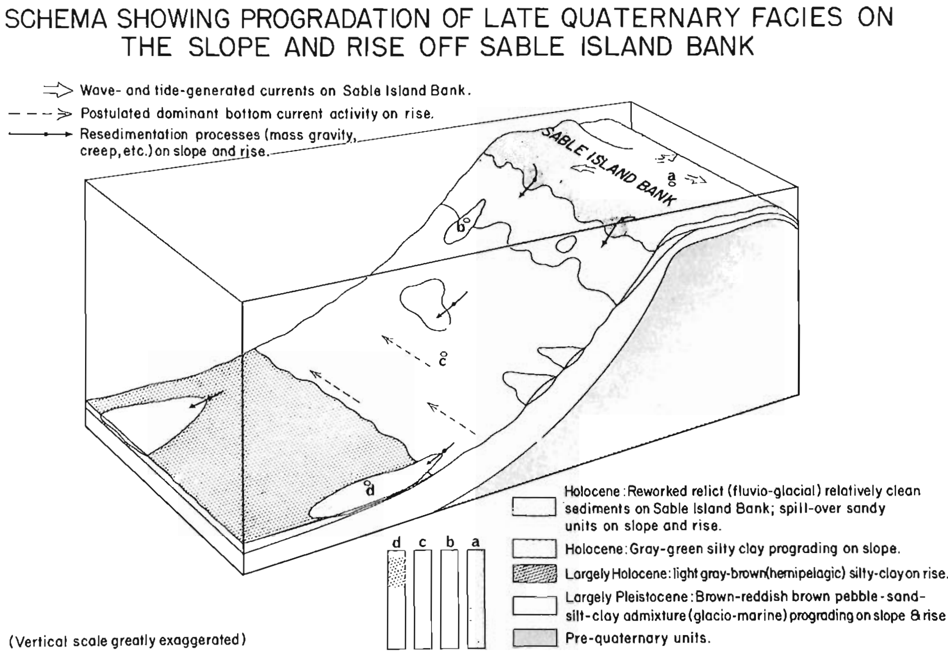

Figure 16. Interpretation of sedimentary facies distribution on the outer Scotian margin in the vicinity of Sable Island Bank. This schema shows late Quaternary progradation on the slope and rise, based on stratigraphy reported in Stanley et al. (1971). Postulated modern dispersal pattern shows spill-over of sands from a sector of Sable Island Bank onto the upper slope. Sand, occasionally by-passing slope and upper rise, accumulates as thin sheets on the lower rise (see core d; this core is based on actual core HUD-30-9A in Fig. 1).

suggested that ocean-bottom currents also may be of primary importance in the transport of deep-sea sands. Although many of these deep-sea sand layers in cores south of Sable Island Bank are, in fact, current laminated and display a low clay content, it is probable that most sands in deep marine environments were nonetheless originally emplaced by sporadic turbidity-current transport (Kuenen, 1967) and subsequently received their secondary structure and texture through reworking by bottom currents (Stanley, 1970, Fig. 12; Stanley et al., 1971). Unfortunately, there is virtually no documentation of the process, or processes, that produce off-shelf movement and downslope transport of sand in this region. It is postulated that off-shelf transfer, as indicated by draping of shelf sand over gullies and depressions of the upper slope, is activated by storm waves (particularly frequent and intense during winter months) and internal waves breaking on the upper slope and seaward bank surface exposed to open ocean conditions. How long the sand remains immobilized on the upper slope (slopes in excess of 5° are not uncommon) before it is transferred downslope is unknown. Channelized flow in submarine valleys appears to be an important method of bypassing the slope as attested by the presence of sand in The Gully axis (Stanley, 1967) and levees bordering some submarine valleys on the rise south of the Scotian margin (Hurley, 1964; Pratt, 1968). That both coarse and fine grades can, in fact, bypass the slope is indicated by the presence of sand horizons in the upper portion of rise and Sohm Abyssal Plain cores (see core logs in Stanley et al., 1971, and hypothetical core d, Fig. 16).

This study of the Sable Island Bank region supports the contention of others (Hubert and Neal, 1967, and others) that mineralogical dispersal patterns formed during the late Quaternary progradation of the western North Atlantic continental margin reflect predominantly a downslope movement of sediment normal to the continental margin. We would go farther, however, and conclude that the off-bank dispersal pattern is continuing at present.

ACKNOWLEDGMENTS

We are indebted to four organizations for their generous support of different phases of this investigation: Bedford Institute of Oceanography, Dalhousie University (Institute of Oceanography), Lamont-Doherty Geological Observatory of Columbia University and the Smithsonian Institution. Sampling was conducted in 1964 to 1966 on the CSS HUDSON, CSS KAPUSKASING and CNAV SACKVILLE: the Captains, officers and men of these ships are thanked for support so efficiently given in the work at sea. Financial support for this project has been provided to one of us (D.J.S.) by the Smithsonian Institution Research Foundation grants No. 235320 (FY 1970) and 436330 (FY 1971).

We acknowledge with gratitude our colleagues who have made special contributions to this study: Dr. A.E. Cok, Adelphi University, and Mr. H. Sheng, Smithsonian Institution, for helping us to identify heavy minerals from slope cores; Dr. J.I. Marlowe (formerly with Bedford Institute of Oceanography), Miami Dade Junior College, for generously sharing his Gully Canyon cores with us; Mrs. A.E. Cok and Mr. H. Sheng for processing core and rise samples; and Mr. L. Isham for drafting and modifying figures. Dr. T.-C. Huang, University of Rhode Island and Mr. D. Magnusson, Mobil Oil Canada Ltd., critically read the manuscript.

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(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 195-196)

11. GEOLOGY OF THE SCOTIAN SHELF AND ADJACENT AREAS

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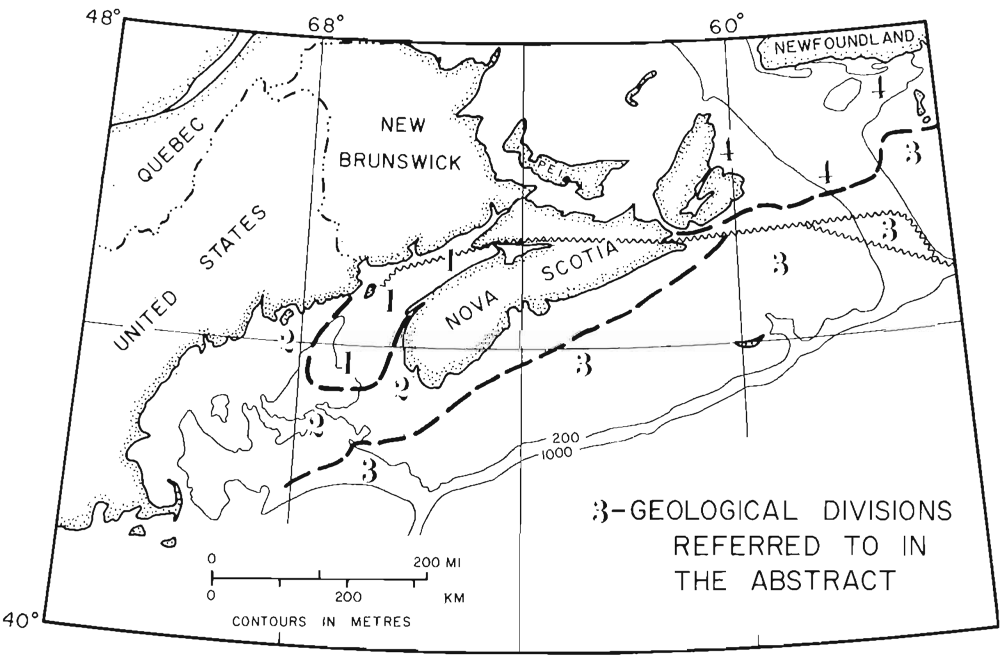
Abstract

The geology of the Scotian Shelf and adjacent areas is interpreted on the basis of structural and stratigraphic relationships and acoustical reflectivity as revealed by a study of continuous seismic-reflection profiles, bedrock control from adjacent shore geology and a few dredged samples, and gravity, magnetic, and seismic refraction data. The geology of the offshore falls into four major divisions: (1) An area of Triassic Basins in the Bay of Fundy and northern Gulf of Maine. The Bay of Fundy section essentially is a half-graben with the Scots Bay Formation overlying the North Mountain Basalt and older Triassic formations. The northeastern Gulf of Maine is a synclinal structure containing Triassic sedimentary and volcanic rocks. The northwestern Gulf of Maine essentially consists of fault-bounded basins containing a thick Triassic sequence; (2) An area of acoustical basement and sedimentary outliers which underlie the eastern central part of the Gulf of Maine. The Meguma Group and White Rock Formation extend southwest of Yarmouth and the acoustical basement farther west consists of undifferentiated pre-Pennsylvanian rocks; (3) A Coastal Plain Province of Tertiary, Cretaceous,



and Jurassic strata which underlies Georges Bank, the Scotian Shelf, and the outer part of the Laurentian Channel. South of Nova Scotia these rocks have a gentle dip seaward and lap on the Meguma Group and Devonian intrusives of the Appalachian Province approximately 50 km offshore. At the northeast end of the Scotian Shelf mildly-folded Early Tertiary, Cretaceous, and undifferentiated Pennsylvanian-Jurassic rocks occur north of the sub-surface seaward extension of the Cobequid-Chedabucto fault system; (4) An area of Pennsylvanian and older rocks south and east of Cape Breton Island.

(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 197-218)

12. SALT STRUCTURES EAST OF NOVA SCOTIA

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Abstract

Air-gun seismic reflection profiles across the Laurentian Channel and the Scotian Shelf west of it have been obtained to further clarify the fold and fault structures of the area. Structures in the Sydney-St. Pierre Basin of Carboniferous rocks are poorly defined by the seismic profiles, but folds and a postulated diapir are indicated in the southern portion. The folds may trend northerly, or otherwise perhaps trend in an east-west direction. The Scatari Ridge basement block dips rapidly southward beneath the Orpheus Basin, and the large prominent folds within the basin strike nearly east-west beneath the Laurentian Channel but apparently strike in a more southwesterly direction beneath the banks west of the Channel. Salt was penetrated in a well southwest of Sable Island; at least two diapirs are recognized north of Sable Island, and an apparent graben and folds indicate that salt structures are common, possibly as part of a continuous area of salt structures from Sable Island to the Sydney-St. Pierre Basin. Shallow domes or folds and other disturbances, including the Tors Cove salt structure, have been mapped by seismic traverses over the western Grand Banks of Newfoundland and the deep ocean to the southwest. The frequency of disturbances on the seismic records indicates that one or more major salt basins underlie the shelves, slopes, and rise east of Nova Scotia. The shelf-edge basement ridge generally mapped as extending northeastward to Sable Island is deeper than previously realized.

INTRODUCTION

Air-gun seismic reflection profiling was carried out in the shelf and slope region east of Nova Scotia in 1967, 1969, and 1970 (using R/V TRIDENT) as part or all of Cruises 41, 73, and 84 of R/V TRIDENT respectively. Professors Dale C. Krause and Jean-Guy Schilling of the University of Rhode Island conducted that portion of Cruise 41 which traversed the Scotian shelf, and the author was Chief Scientist on the two later cruises. Cruises 73 and 84 were conducted with the primary purpose of mapping subbottom structure, with particular reference to the distribution and nature of salt structures in offshore eastern Canada.

Diapirs including two known salt structures have been reported at widely scattered localities offshore. Watson and Johnson (1970) describe probable diapirs at approximately 39° N, 50° W, southeast of the Grand Banks. Keen (1970) reports one in the outer Laurentian Channel, and King and MacLean (1970a) report a diapir about 25 miles (46 km) north of Sable Island. Another is shown on an unpublished reflection line shot across the Laurentian Channel by R/V GLOMAR CHALLENGER (A. Ruffman, personal communications, 1971).

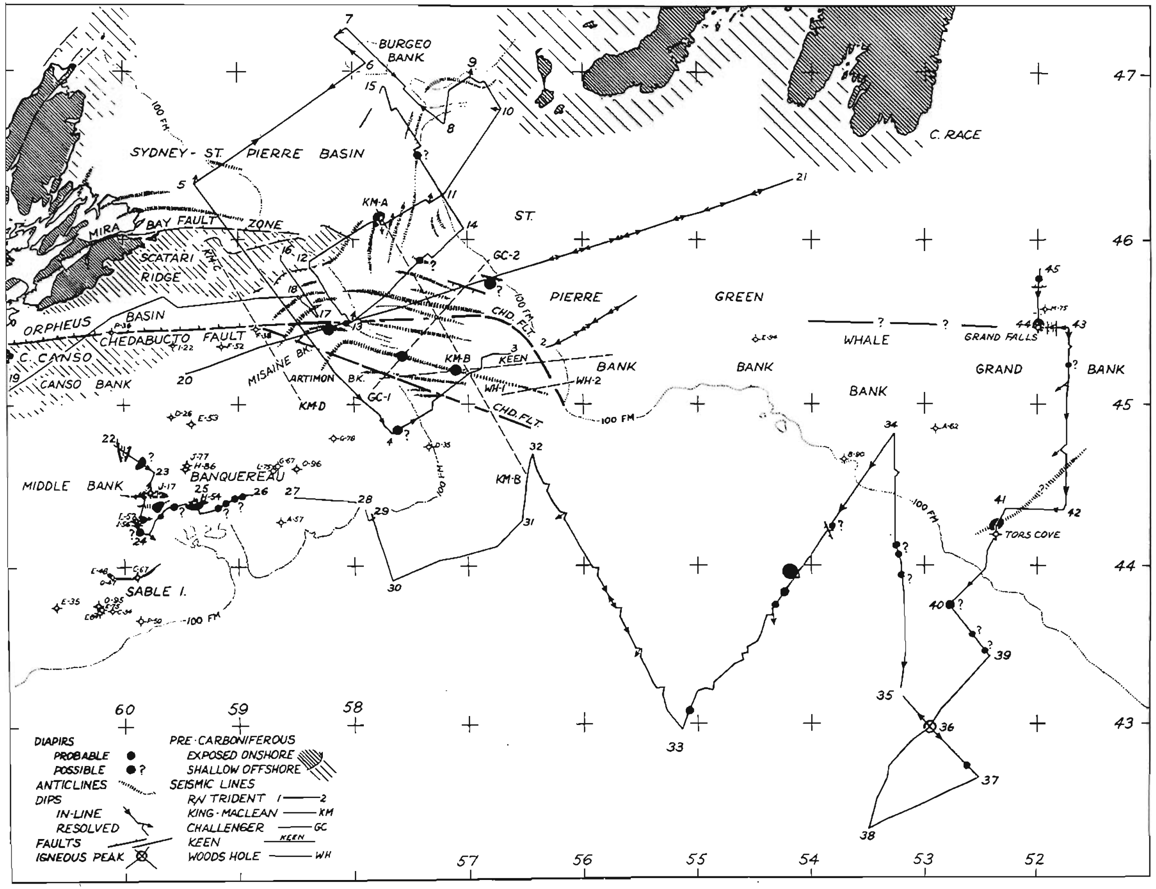

Figure 1. Index map of study area. TRIDENT seismic lines: 1-19, Cruise 73, 1969; 20-21, Cruise 41, 1967; 22-45, Cruise 84, 1970. Other lines as labelled. Frequency of structures and exploratory wells evident near Sable Island. Prominent folds and diapirs shown in Misaine-Artimon Banks and Laurentian Channel west of St. Pierre Bank. Northerly trending interpretation chosen for structures on Line 11-12, also south-westerly trends in Misaine Bank area. Apparently a cluster of structures occurs offshelf to the south and west of Tors Cove uplift. Well locations as of January, 1972.

In addition, the Pan American Tors Cove well, near the southern edge of the Grand Banks, penetrated into a salt dome shortly after reaching Mesozoic formations. The TRIDENT 41 line across the Laurentian Channel shows well-developed folding in the continental shelf section, and prompted the programming of Cruises 73 and 84. More recently, Pamenter (1971) reports that the first of the offshore Shell wells, Onondaga E-84, drilled into salt at a location about 18 miles (33.3 km) southwest of the Mobil C-67 well on Sable Island, thus proving the occurrence of salt far to the west of the Tors Cove D-52 well. The salt was encountered at 12,990 feet (3,962 m) below about 5,000 feet (1,525 m) (drill depth) of Mesozoic formations. Whether or not the salt is diapiric is not stated.

The seismic profiling done with R/V TRIDENT utilized Bolt airguns of up to 10-cubic-inch volume and Raytheon and Alpine recorders. Navigation was with Loran C, supplemented by celestial fixes and dead reckoning. Minor adjustments were made to adjust to bathymetry as shown on Canadian Fisheries Chart 4041 (Banquereau and Misaine Bank) and Bathymetric Charts 801 and 802 (Bay of Fundy to Gulf of St. Laurence, 1969, and Newfoundland Shelf, 1970).

Regional Geology of the Eastern Scotian Shelf

The regional geology of the eastern Scotian Shelf has been described by Alcock (1949), Geological Association of Canada (1950), Williams (1967), Sheridan and Drake (1968), Uchupi and Emery (1968), Webb (1969), Emery *et al.*, (1970), King and MacLean (1970b) and others. King and MacLean's (1970b) paper is particularly useful in that it defines and describes a number of major tectonic elements, particularly the Sydney-St. Pierre Basin extending northeasterly and easterly from the Sydney Basin of Cape Breton Island (Fig. 1), the Scatari Ridge of shallow pre-Carboniferous basement extending eastward from Scatari Island and Scatari Bank, (separated from the basin by the Mira Bay Fault), and the eastern extension of both the Orpheus gravity anomaly (and the Carboniferous basin it represents) and the Cobequid-Chedabucto Fault system which bounds the Orpheus Basin on the south. They and Hood (1967) have shown further that the Meguma basement block extends eastward for some distance from Cape Canso at a shallow depth on the south side of the Chedabucto Fault. Its upper surface dips southward, gently at first and then more steeply beneath the Cretaceous-Tertiary shelf cover, which is in excess of 15,000 feet (4,650 m) thick at Sable Island (Howie, 1970). King and MacLean (1970b) interpreted from their numerous continuous-seismic profiles where folding had occurred and concluded that the folds are essentially limited to the Orpheus Basin. This basin is believed to contain a thick salt-bearing Carboniferous section beneath the Tertiary and Cretaceous sediments (Loncarevic and Ewing, 1967), and King and MacLean suggest that the salt is involved to at least some degree in the folding seen in the overlying shelf formations. They further show that the folding affects the Cretaceous formations in the Laurentian Channel area but that the folded Cretaceous is separated from the overlying unfolded Tertiary by an angular unconformity. Presumably the folding is the result of renewed movement on the Carboniferous Chedabucto Fault (King and MacLean, 1970b).

Mississippian Windsor salts are known to occur widely in the Maritime Provinces and the Gulf of St. Lawrence (Howie and Cumming, 1963), and thus the salt that underlies the shelf and the deep ocean east of Nova Scotia may also be Windsor evaporites. However, a Mesozoic age for the salt in the

Tors Cove well has been suggested (Howie, 1970). If the salt at Tors Cove is Mesozoic, the salt near Sable Island may also be that age, where the Mesozoic-Tertiary section likewise is thick and marginal to the continent. Presumably the Mesozoic salt deposits were deposited early in the present cycle of Atlantic Ocean opening and hence would be found along the continental edge and perhaps also occur beneath the continental slope and rise. Whether the outer edge of the Windsor salt, perhaps a rift cut-off edge, is sufficiently far seaward to be overlapped by a postulated inner margin of a Mesozoic salt formation, hence to cause a double occurrence of salts in the section, is not at all clear, but it remains a distinct possibility for the outer shelf region.

Interpretation of Continuous Seismic Records

Because of the limited penetration achievable with the seismic profiling equipment used, especially in waters directly underlain by Carboniferous beds or by other strongly reflecting sedimentary formations and also because of the multiple reflections developed in shallow-water shooting it is recognized that only very shallow salt bodies might be outlined directly. Accordingly, the various ways in which salt structures can be detected on the records might usefully be listed and anticipated as follows:

1. Shallow diapirs showing both steeply upturned beds and an unstratified core; these features are identifiable by the fact that reflections in the section are not present for a short distance - the so-called reflection cutout.
2. Shallow diapirs or possible diapirs, without sharp upturning of beds and also identifiable by reflection cutout.
3. Deep-seated swells or anticlines, not necessarily diapiric, identifiable only by the shallow upfolded beds; definite structures but interpreted as salt structures only by association with known or postulated salt beneath them.
4. Normal fault structures, particularly graben, without or without discernable upfolding of beds; also definite structures but not necessarily attributable to salt structures at depth.
5. Miscellaneous disturbances, including faults and folds, of uncertain geometry and origin, likewise only tentatively associated with postulated salt movement at depth.

Features of types 3, 4, and 5 do not necessarily define salt occurrences, whereas types 1 and 2 do define either salt or shale uplifts. Type 5 features inevitably include occurrences of any kind which are too poorly defined to assign otherwise, hence may include artifacts as well as real geological structures which are too complex to be revealed by the seismic survey.

The lines shot during Cruises 73 and 74 include a number of short "dog-leg" segments which were made in order to obtain true dip-strike control along otherwise straight traverses. While some of them were in areas of very local structure and others suffered from poor record quality, in general these corners have proven helpful in this study and offer improved data towards a regional structure study.

Results of Continental Shelf Study

Figure 1 shows the locations of seismic lines interpreted in this report, with location numbers at appropriate points. The northern part of Line 4-5, the entire Laurentian Channel crossing Line 5-6 and Lines 6-11, 14-15, and part of Line 11-12, all are north of the Mira Bay Fault and hence in the Sydney-St. Pierre Basin. The Carboniferous section is thick in that area (Sheridan and Drake, 1968) and only shallow, gentle fold structures are detectable. Several folds are seen near Location 5 and are thought to be continuous with folds seen onshore in the Sydney area; a north-northeasterly dip is resolved at Location 5. Line 5-6 shows a gentle syncline across the Laurentian Channel rising to a slight structural high under Burgeo Bank. The Carboniferous beds likewise dip to the northwest and to the southeast off Burgeo Bank, indicating that it is a structural high. Likewise, the western tip of St. Pierre Bank appears as a high, with beds dipping northwest or north from the corner at Location 8, dipping north at Location 9 but again dipping more westerly at Location 10. Dips are gentle. Mild folds may be wrapped around the western tip of the Bank.

A system of north-northeast trending folds may be inferred as crossing Line 11-12, if the finger-like pattern of Tertiary and Pennsylvanian outcrops shown by King and MacLean (1970b, Fig. 11) is in fact representative of structure. The Tertiary section there is a series of fillings of erosional valleys, and therefore might have no structural connotations, but as Line 11-12 crosses several appropriately placed but rather poorly defined folds or possible folds (type 5, above), it may also be that the Tertiary remnants are filling eroded anticlinal valleys. If, however, the Tertiary trends are independent of structural trends, then the fold trends would best be interpreted as running approximately east-west parallel to the Mira Bay Fault trend in the same manner as the folds to the south lie more or less parallel to the Chedabucto Fault. This trend is in fact suggested by the northerly dip resolved at the corner about 6 miles (9.6 km) west of Location 11.

About midway between Location 11 and Location 12 the line crosses Line KM- A-B of King and MacLean (1970b) and then jogs towards Location A before again turning towards Location 12. An uplift is interpreted on the north corner of the jog, and is adjacent to indications of an uplift near Location A on Line KM- A-B. It is interpreted that a diapir is crossed there, although whether a domal one or a linear feature is uncertain. The magnetometer records a 20-gamma low over the feature, and it is therefore interpreted as a salt diapir rather than an igneous intrusion.

Line 11-12 crosses the Mira Bay Fault Zone near 46° N, 58° W, and the seismic profiles indicate the presence of an anticline immediately adjacent to the eastern portion of the Scatari Ridge (Fig. 1). Location 12 appears to be on the Scatari Ridge of shallow pre-Carboniferous basement, or else over a small sedimentary re-entrant on its south side. Line 12-13 passes over a fold or flexure 3 miles (5.6 km) south of Location 12 and into the deeper Orpheus Basin of thick Carboniferous rocks overlain by Cretaceous sediments. Lines 16-17, 17-18, 4-5 (Webb) and Lines 20-21-22 and KM- C-D of King and MacLean (1970b) also define the same south-dipping limb of the Scatari Ridge block, trending perhaps 260 degrees or nearly west. The seismic profiles show a rapidly thickening Cretaceous section to the south across the Orpheus Basin, which contains a number of folds and possible faults. These structures underlie the northeastern portion of a large zone of complex bathymetry lying

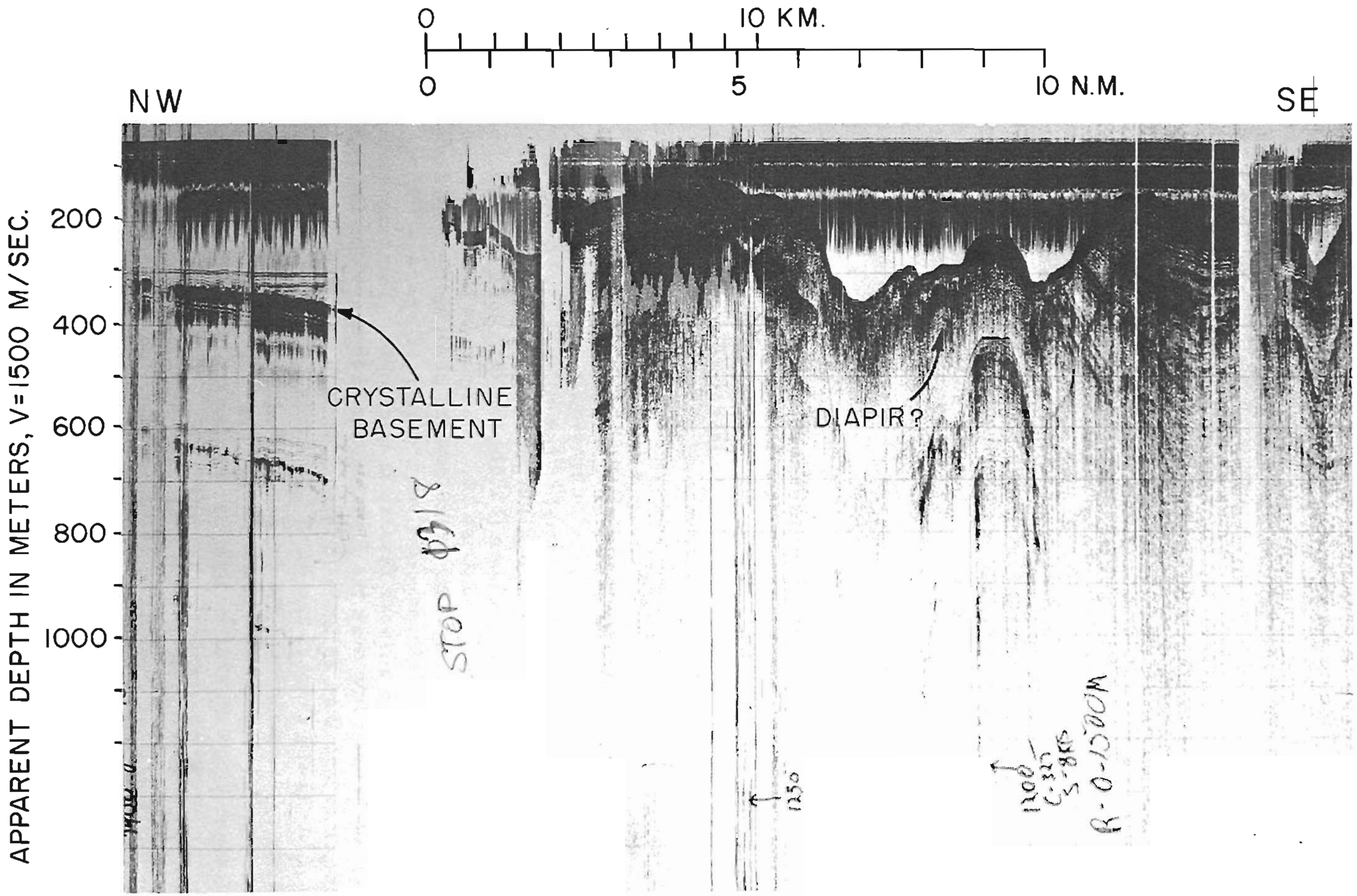

Figure 2. Centre portion of TRIDENT Line 4-5 across Orpheus Basin to Scatari Ridge. Prominent fold (diapiric?) evident at 1200 time mark, near Location 18, in Mesozoic beds, and Scatari Ridge crystalline basement visible beneath shallow cover at northwest end.

between Scatari and Banquereau Banks. Although the zone has some flat shoals, e.g. Canso, Misaine, and Artimon Banks, it is characterized by many small closed depressions and isolated hills (Canadian Hydrographic Service Charts 4041, 801) representing a modified subaerial erosion surface (King and MacLean, 1970b). Another such zone leads eastward into The Gully near Sable Island.

The bathymetric trends between Artimon and Scatari Banks are interpreted by this writer as being at least in part structurally controlled in the area closely traversed by Lines 12-13, 16-17, 17-18, and part of Line 4-5, (see Fig. 2). If so, the structures there strike approximately northeasterly parallel to the prominent bathymetric grain. A second alternative is that they strike more easterly, directly into the folds in the Laurentian Channel between 45° and 46° N; possibly the area is structurally complex with both trends involved. The easterly strike, or the occurrence of both trends, appears more likely to become dominant to the south and east of Location 13.

One prominent structure or en echelon structural zone, variously appearing as an anticline or as a south-dipping flexure, appears to trend northeasterly across the lines about 4 miles (7.4 km) north of Location 17 (see Figs. 1, 2), thence to curve more easterly into a large anticline which crosses Line 20-21 and Line KM-A-B, a few miles west and south of their common intersection in the Laurentian Channel (Fig. 3). A second such anticlinal structure, possibly including diapirs on Line 20-21 at the east end of Misaine Bank, extends to Line GC-1-2 in the Laurentian Channel, and a third (or a branch of the second) seems to trend about 105 degrees (east-southeast) across Line KM-21-22 of King and MacLean (1970b) (not shown on Fig. 3), through a diapir on the CHALLENGER Line GC-1-2, and through the diapir defined by Keen (1970). It probably continues into the strongly disturbed, seemingly folded, mostly east-dipping zone seen on Line WH-1-2 (Emery and Uchupi, in preparation), and strikes essentially parallel to and north of the extended southern branch of the Chedabucto Fault (King and MacLean, 1970b). Southward from Artimon Bank the folds either die out or are too deeply buried to be recognized, the visible section thickening to the southeast or east by progradation.

In the Laurentian Channel proper, the largest features are the almost east-west trending folds and diapirs centered within 45°-46° N, 57°-58° W. As noted above, their strike is determined with considerable confidence by careful comparison of Lines 20-21, obtained on TRIDENT Cruise 41 (Fig. 3), and Line KM-A-B (KM-4-5) of King and MacLean (1970b). The three largest folds referred to in the discussion of the Scatari to Misaine Bank area, above, are accompanied by several smaller anticlines to the northeast. However, the eastern portions of both show a flat or gently southwestwardly dipping structure, probably separated by a fault from the prominent folds beneath the central channel. Whether this undisturbed structure extends far beneath the St. Pierre Bank eastward is unclear, although some folds and faults of rather uncertain quality are picked on Lines 20-21 and 1-2. Woods Hole's Line WH-1-2 certainly shows that such structures do occur to the edge of the Bank if not beneath it.

TRIDENT Lines 22-26 and 27-28 were laid out to examine parts of the bathymetrically complex areas north and east of Sable Island because one diapir and several wildcat well locations were already known to be located there. Line 22-23 initially crosses a 4-mile (7.4 km) wide basin with a narrow central ridge separating two slightly upbowed deep floors, with relief of about 100 metres. Each deep overlies a V-shaped discontinuity, either a graben or a partly filled valley (Fig. 4). Two of the subsurface planes have true

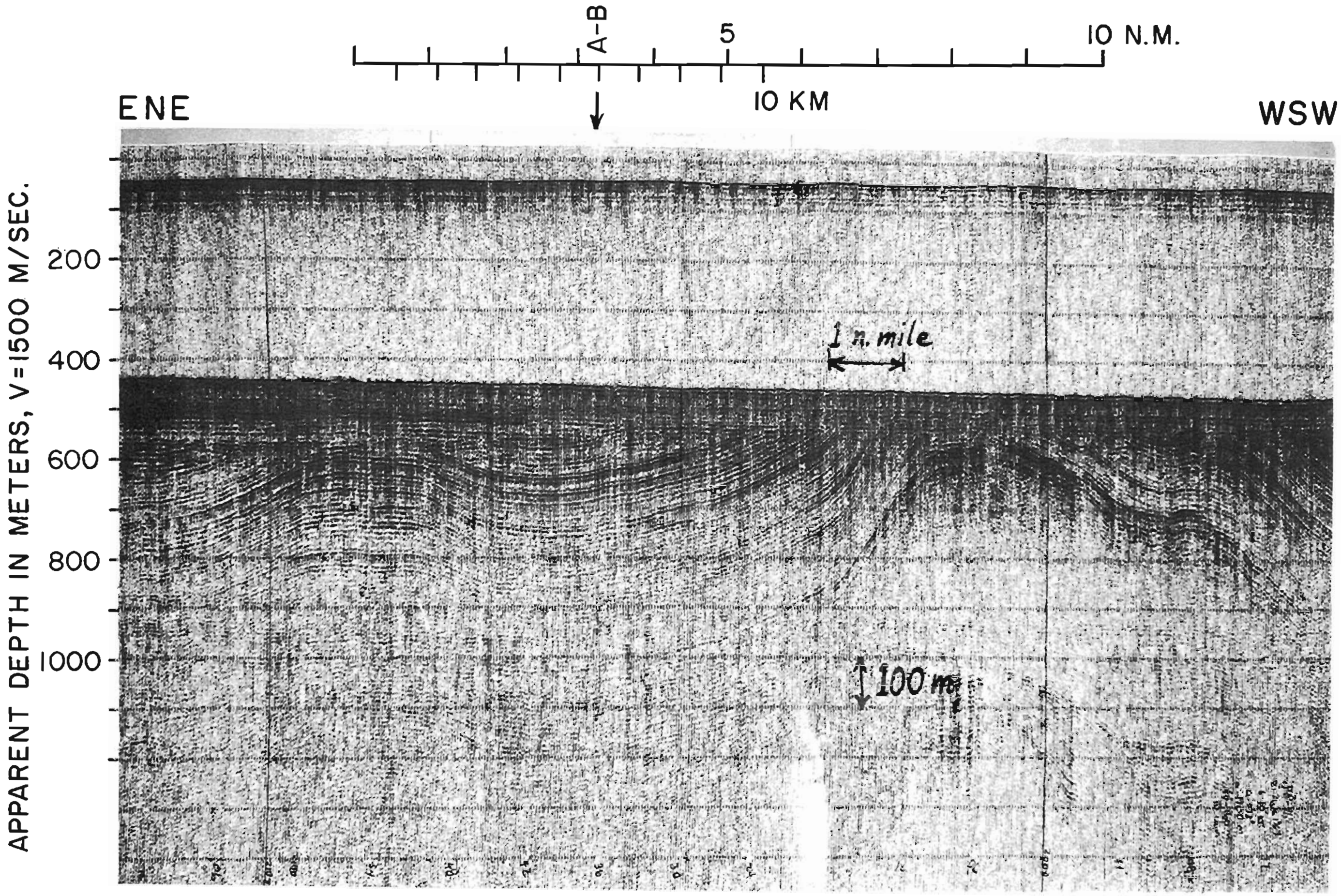

Figure 3. Four anticlines on TRIDENT Line 20-21 in Laurentian Channel. Crosses King and MacLean Line KM-A-B in center of photograph, permitting interpretation of fold trends. Folded beds are Cretaceous, flat beds above unconformity are Tertiary and Quaternary.

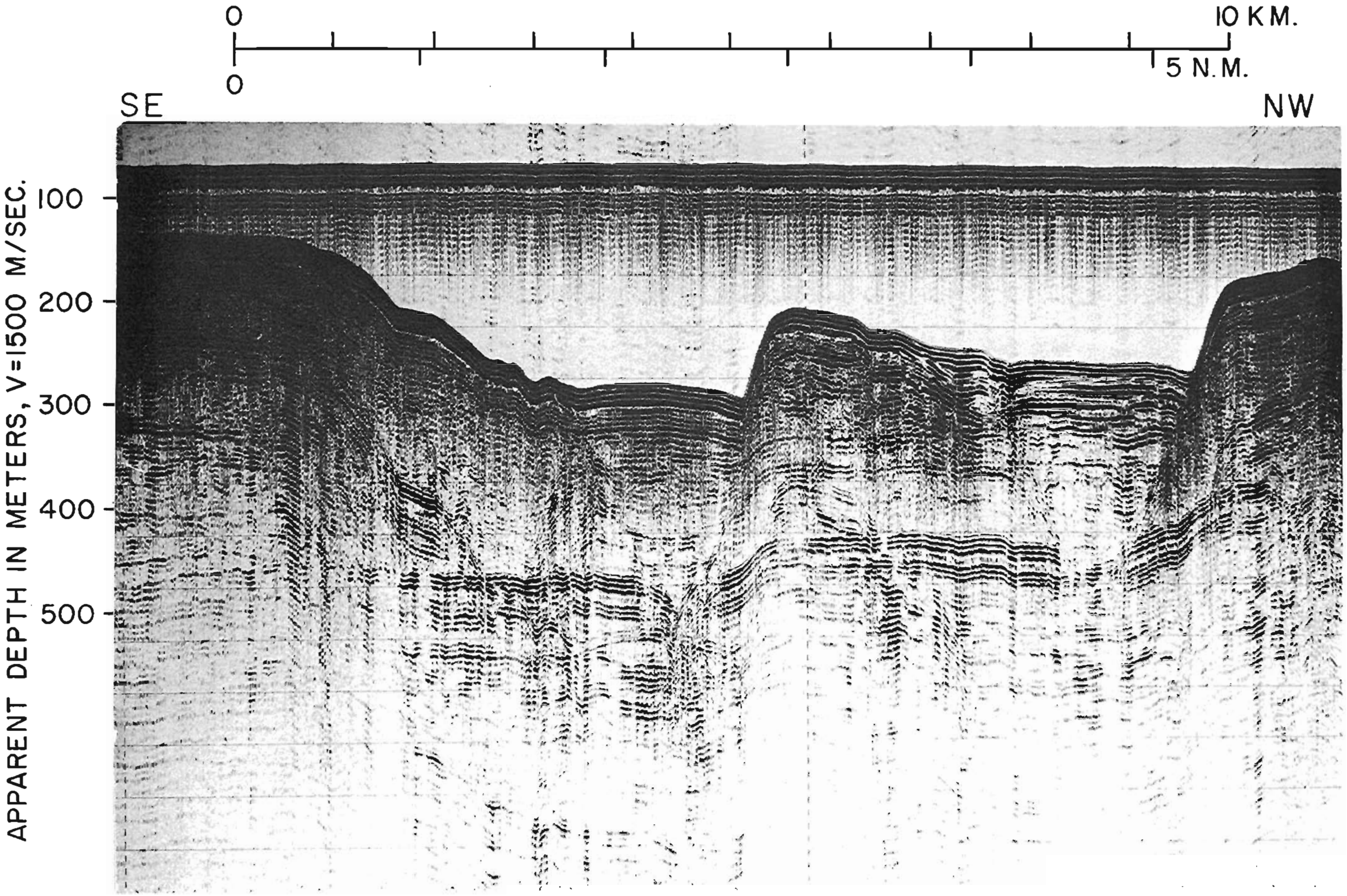

Figure 4. Two valleys believed to overlie graben over deep-seated salt structures, near Location 22 on western portion of TRIDENT Line 22-23. Edge of Banquereau shoal on left.

dips of 31 and 35 degrees, assuming a sediment velocity of 2,250 metres per second and a degree of obliquity of track to strike as determined by bathymetric trends. Subsurface slopes dipping more than 30 degrees may possibly be buried valley walls, but their regular planar nature and relative steepness indicate that they are in fact fault planes. While these dips are less than the usual 60 degrees for a normal fault, perusal of salt dome cross-sections included in Murray (1961) show several faults dipping about 30 degrees, specifically in shallow positions above salt domes where the fault crosses the domal crest. This author favours the fault interpretation for the discontinuities seen on Line 22-23 and on Line 23-24 at the south edge of Banquereau Bank, and further infers that they overlie deep-seated salt pillows or domes (type 4 structure, above). The western portion of Banquereau Bank appears to be underlain by disturbed north-dipping beds and appears to be terminated southward by a south-dipping normal fualt of perhaps 38 degree dip. Diapiric structures may underlie the Bank. The Shell Iroquois J-17 well was drilled near Line 23-24 a short distance north of the fault.

South of Banquereau Bank, Line 23-24 crossed a small uplift lying immediately west of the high diapir described by King and MacLean (1970a), and then crossed another distinct diapir about 8 miles (14.8/km) to its southwest (Figs. 5, 6). Shell wildcat Abenaki L-57 and J-56 wells were drilled to the south or southwest of the crest of this latter structure. Location 24 may overlie another diapir. Line 24-25 skirted King and MacLean's diapir on its south and east, and Line 25-26 passed southward (across a syncline or graben?) and over the Shell Missisauga H-54 well dome or half dome, thence back onto Banquereau Bank eastward towards Line 27-28. Structure beneath Banquereau Bank is indeterminate because of penetration and multiple reflection problems, but what can be seen appears to be considerably disturbed as if the entire section is faulted and perhaps folded. It might well overlie a broad area of salt swells and pillows, but the field evidence is inconclusive.

Fig. 6 is a detailed map of the area around Sable Island, showing bathymetry, wells, the Cruise 84 track, and interpreted subsurface structure. The diapir of King and MacLean (1970a) at 44° 22'N, 59° 42'W and the Abenaki L-57 well diapir (TRIDENT 84) are "type 1" structures as seen on the records, with sharply upturned beds and reflection cutouts. The remainder of the diapir structures shown are considered as possible rather than probable and if they are, in fact, diapirs they are most likely deeper structures than the two described above. Specifically, structures traversed by Line 22-23 and the northern part of Line 24-25 are interpreted as type 4, shallow normal faults over salt at depth, and the group of diapirs (?) on Lines 24-25 and 25-26 east of about 59° 35'W are of types 2 and 5, reflection cutouts and miscellaneous other disturbances. These diapirs (?) are shown in Figure 6 as being centered on the TRIDENT 84 track, but in fact most of their crests would lie on one side or the other of the track. They may very well represent portions of one or two broad salt massifs rather than individual domes.

Fault orientations are based largely upon bathymetric trends, on the assumption that these trends are structurally controlled to some extent, and the structure contour trends are inferred similarly in part. The two distinct diapirs may well be parts of a single northeasterly trending salt ridge, possibly extending past the Mobil-Tetco E-48 discovery well on the west end of Sable Island and the Shell Abenaki L-57 and J-56 wells, to the Shell MicMac H-86 and J-77 wells, or beyond. In that case the Shell Iroquois J-17 and Missisauga H-54 wells are on separate or branching structures. The Halifax-Burgeo

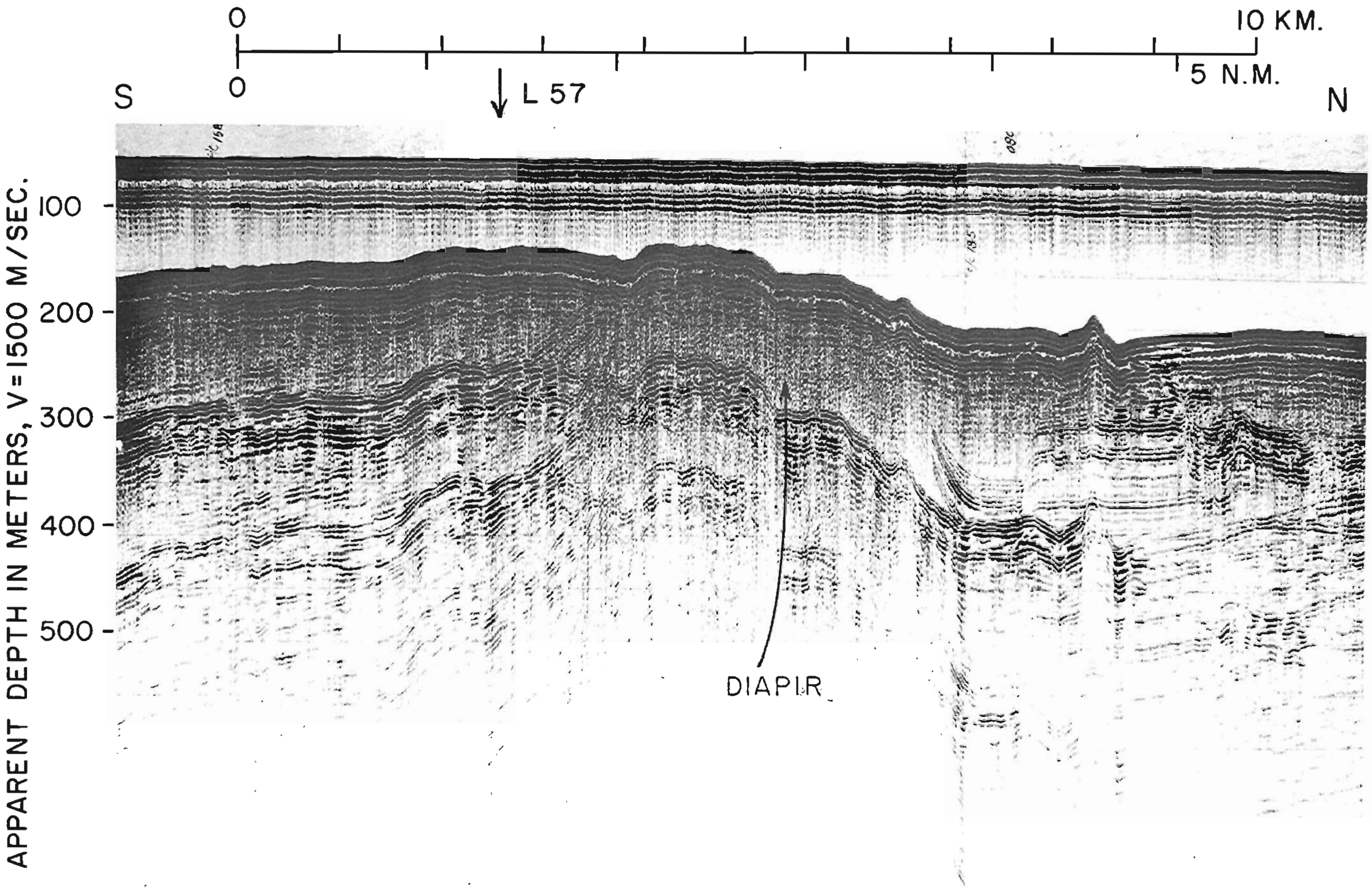

Figure 5. Diapir structure on TRIDENT Line 23-24. Shell Oil L-57 well being drilled 0.5 mile (0.8 km) to the west of line at time of survey in 1970. Upturned beds visible between sea bed return and first multiple in Cretaceous or Tertiary beds.

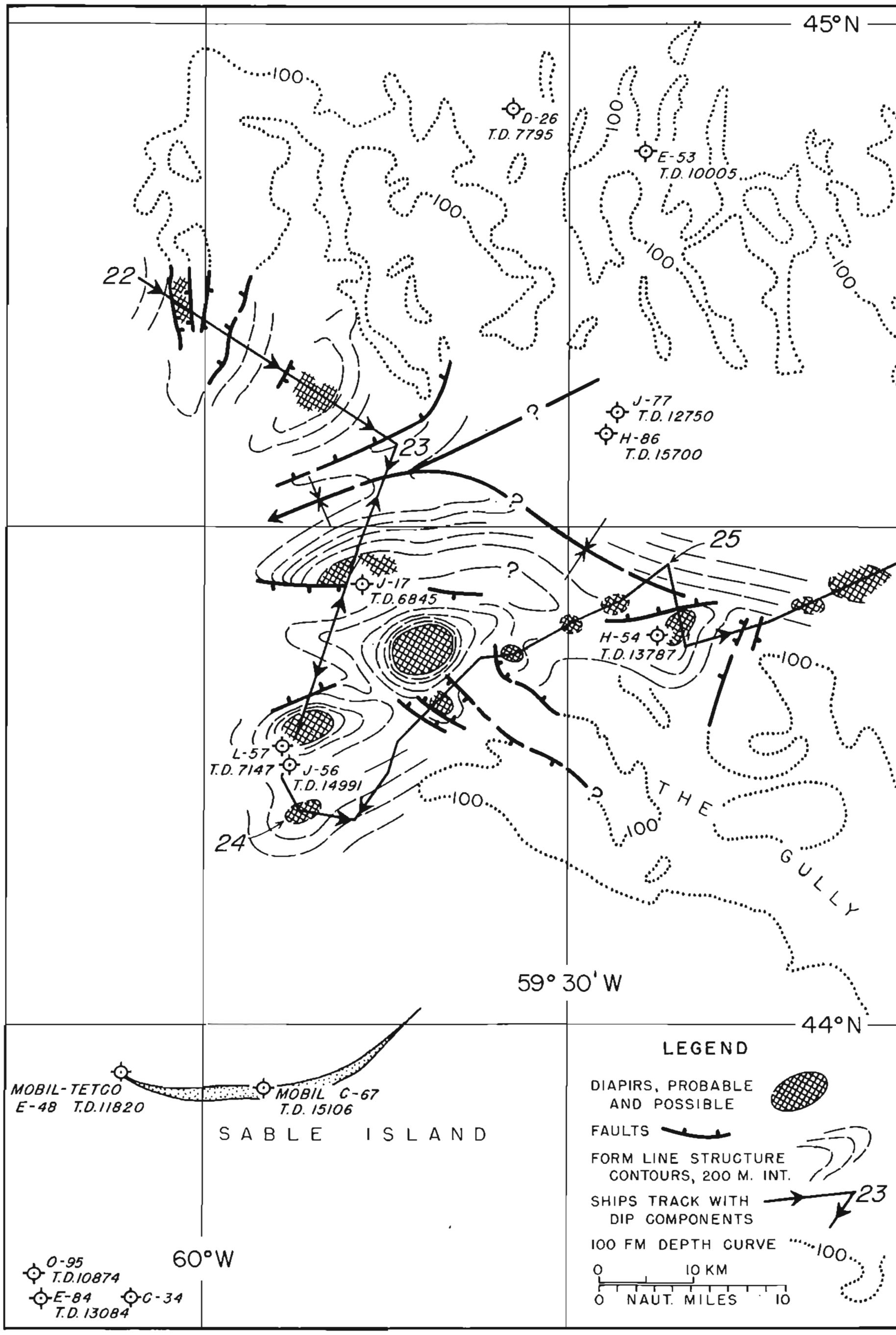

Figure 6. Detailed map of area around and north of Sable Island. Structure shown by 200-metre interval form line contours on shallow horizons, with the addition of interpreted faults and diapirs. Structural interpretations based on reflection seismic records, supplemented by bathymetry and gravity. Two alternative extrapolations shown for the syncline north of the J-17 well. Well locations as of January 1972.

Bouguer Anomaly Map (Stephens et al., 1971) shows a northeasterly trending negative anomaly passing through the diapir area and, although at the very edge of the contoured data, supports this continuous ridge hypothesis. On the other hand the probable east-west orientation of the Iroquois J-17 faulted half dome (?) suggests a similar easterly trend for the syncline north of it, near Location 23, in which case it might well cut across the gravity trend and into the structural low which is interpreted to occur on the north side of the Mississauga H-54 dome. Figure 6 shows both of these alternative trends for the syncline. The gradients indicated by the structure contours, which have a 200-metre interval, are those seen on the TRIDENT 84 records and as shown by King and MacLean (1970a). They are not numbered for particular depths, however, even for a phantom datum, because of the impossibility of making correlations across faults and other disturbed portions of the records. More and better data would be needed in order to construct an accurate structural map.

No structure is shown under or near Sable Island on Figure 6 because of inadequate data. The island has been interpreted as being on or near the crest of a basement ridge mapped as extending from latitude 38° N to Sable Island beneath the outer edge of the continental shelf (Maher and Applin, 1971; Berger et al., 1965). However, the Shell Oneida well bottomed at 13,484 feet (4,112 m) total depth, fully 2,500 feet (763 m) deeper than Maher and Applin's contoured basement surface, at a location about 90 miles (150 km) west-southwest of Sable Island. Likewise, Mobil Sable Island C-67 well bottomed at 15,106 feet (4,604 m) total depth (Howie, 1970) without reaching either the 5,000-foot (1,525 m) (?) thick Jurassic formations or the salt which were encountered nearby in the Shell Onondaga E-84 (Pamenter, 1971), while Berger et al., (1965) indicated a thickness of only about 14,750 feet (4,499 m) of sediments beneath the island as calculated from seismic refraction data. Clearly the depths to basement are consistently too shallow when obtained from refraction work beneath the outer Scotian Shelf, thus casting the very existence of the shelf-edge basement ridge in doubt. Perhaps the evaporites there have velocities sufficiently high to be confused with basement, suggesting that the "basement ridge" as mapped may in fact be a relatively persistent trend of salt structures.

The short traverses on Whale Bank reveal little more than seaward dip and one or two possible faults. Lines 40-45 reveal rather more. Immediately north of the Tors Cove D-52 well a sharp anticline is evident despite the difficulty produced by multiples (Fig. 7). A less well-defined feature, possibly diapiric, occurs on Line 41-42 and another possible disturbance occurs on Line 42-43, 15 miles (27.8 km) beyond Location 42 in line with the Tors Cove well and the Line 41-42 feature. These three disturbances may define a northeasterly strike for the Tors Cove structure rather than its being a strictly circular dome. If so, it would be like the Laurentian Channel structures, elongated folds, almost certainly salt-cored, with one or several diapirs along the crest.

Line 42-43 otherwise exhibits little structure south of Lat. 45° N, but disturbed zones (faults?) and swells are relatively frequent from there to the end of the line at Location 45. Folds are crossed at the course jog at 45° 08'N, and graben (?) and possible diapirs are again frequent past the Grand Falls H-09 well and along Line 44-45.

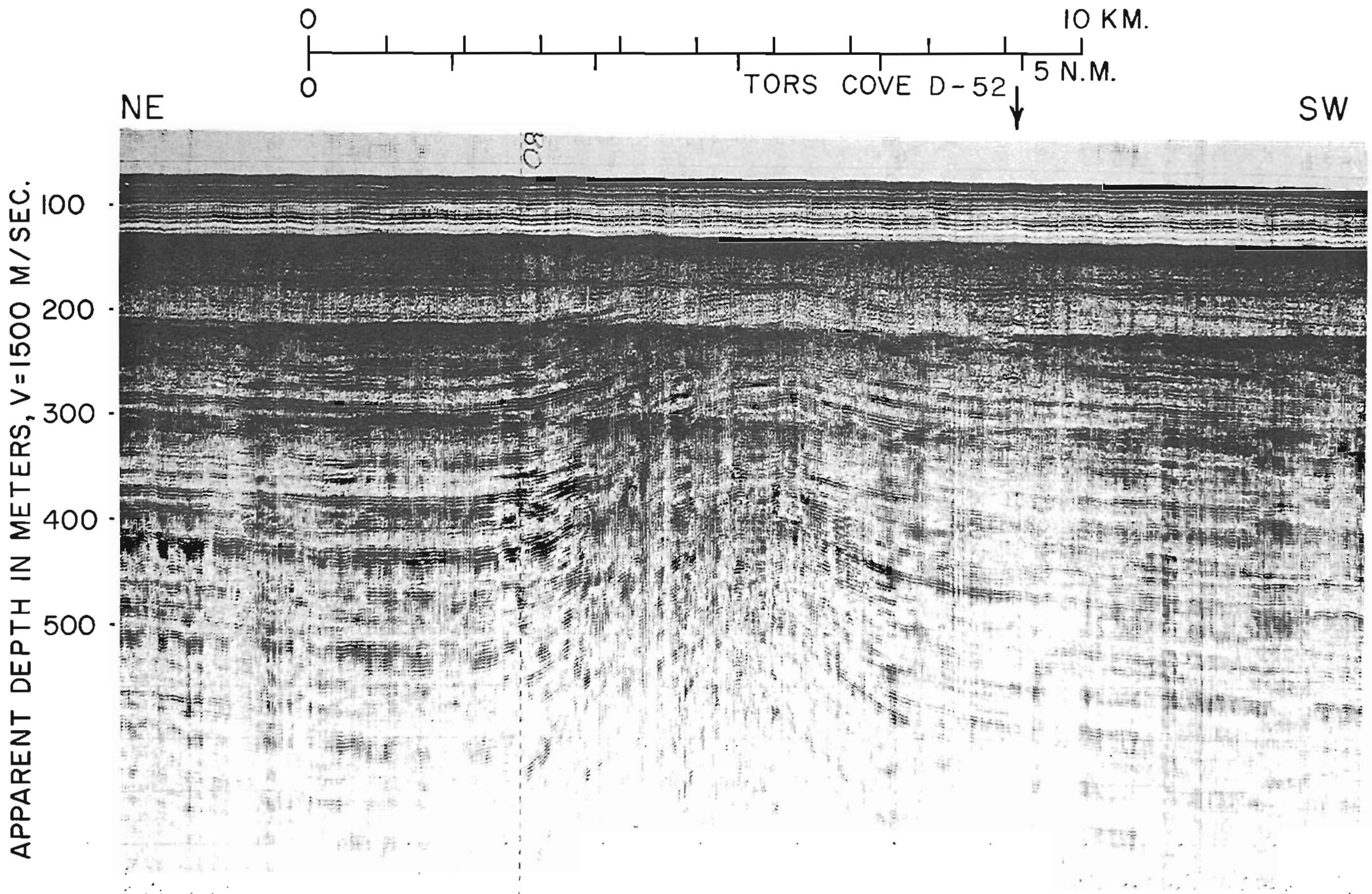

Figure 7. Anticline (diapiric?) at Tors Cove D-52 well location, TRIDENT Line 40-41 on Grand Banks. Pan American well was drilled to south of crest and encountered salt at depth.

Results of Continental Slope and Rise Study

Cruise 84 made one short excursion off the shelf, Lines 29-32, and two longer ones, Lines 32-33-34 and 34-36-37-40 and beyond, in order to evaluate the possibility of diapirs (Mesozoic salt?) beneath the slope and rise. The seismic cross-sections obtained on Lines 29-32 were affected by rough topography and show little, and parts of the other lines were likewise unproductive. Line 32-33 traversed an extensively slumped area to within about 20 miles (37 km) of Location 33, then passed into an apparently undisturbed sedimentary terrane exhibiting a number of good reflectors (turbidites?). The large slump area is that presumed to have moved in 1929 (Heezen and Drake, 1964).

Line 33-34 shows a cluster of apparent diapirs from 43° 40' to 44° 20'N. One, just south of 44° N (Fig. 8) was encountered at two course-change corners, going into and out of a 90 degree jog intended to reveal a true dip-strike in that part of the slope. The seismic record obtained at the 90-degree corners show that the diapir must lie to the northwest of the line. Line 34-35 indicates several possible diapirs, the section being confused by slump features. Another possible diapir occurs at Location 40, perhaps part of a large group of diapirs extending across all three lines. Such a group might be concentrated in a band parallel to the southeasterly striking shelf edge, but it might equally well trend more easterly and even northeasterly through the Tors Cove structure. Most of these diapirs are interpreted from occurrences of reflection cutout or upturned beds.

The articles by Sheridan and Drake (1968) and Drake et al. (1968) include structural contour maps of the area contoured on pre-Upper Pennsylvanian and on crystalline basement respectively. Both maps show the Chedabucto Fault trend extending essentially eastward from Cape Canso across the Laurentian Channel and well across the Grand Banks, at perhaps 45° 30'N. A southwesterly plunging structural basin is mapped to the south of the extended Chedabucto trend, separating it from a second structural high beneath the southeastern part of the Grand Banks. This basin would enclose and may serve to limit the cluster of salt diapirs believed to extend across the three above-mentioned slope lines and to the east and northeast through the Tors Cove well. The structures seen in the vicinity of and north of the Grand Falls well might thus belong to a second cluster of salt uplifts occurring to the north of the eastwardly extended Chedabucto Fault, essentially in the position of the Orpheus Basin in the vicinity of Nova Scotia. That the Chedabucto Fault might extend entirely across the Grand Banks, perhaps bounding a ridge separating two sedimentary and salt basins does not seem improbable to this writer. The Cobequid-Chedabucto Fault is one of the largest faults of the Maritime Provinces with possibly more than 100 miles (185 km) of dextral strike-slip during Carboniferous time (Webb, 1969), in which case it would be expected to continue many miles offshore. Indeed, Schenk showed that it could be continuous with the South Atlas fault of Morocco (Schenk, 1972).

The diapir-like structures seen in the lower slope-upper rise, as on Line 33-34, are part of the irregular ridge mapped by Emery et al. (1970) and Maher and Applin (1971) as extending from the Blake Plateau to the Grand Banks. They consider it to be either a basement ridge or a diapiric complex but favour the former. This author is impressed by the frequency of diapir and diapir-like structures in the area and prefers to interpret the slope-rise features for the most part as diapirs, presumably of salt.

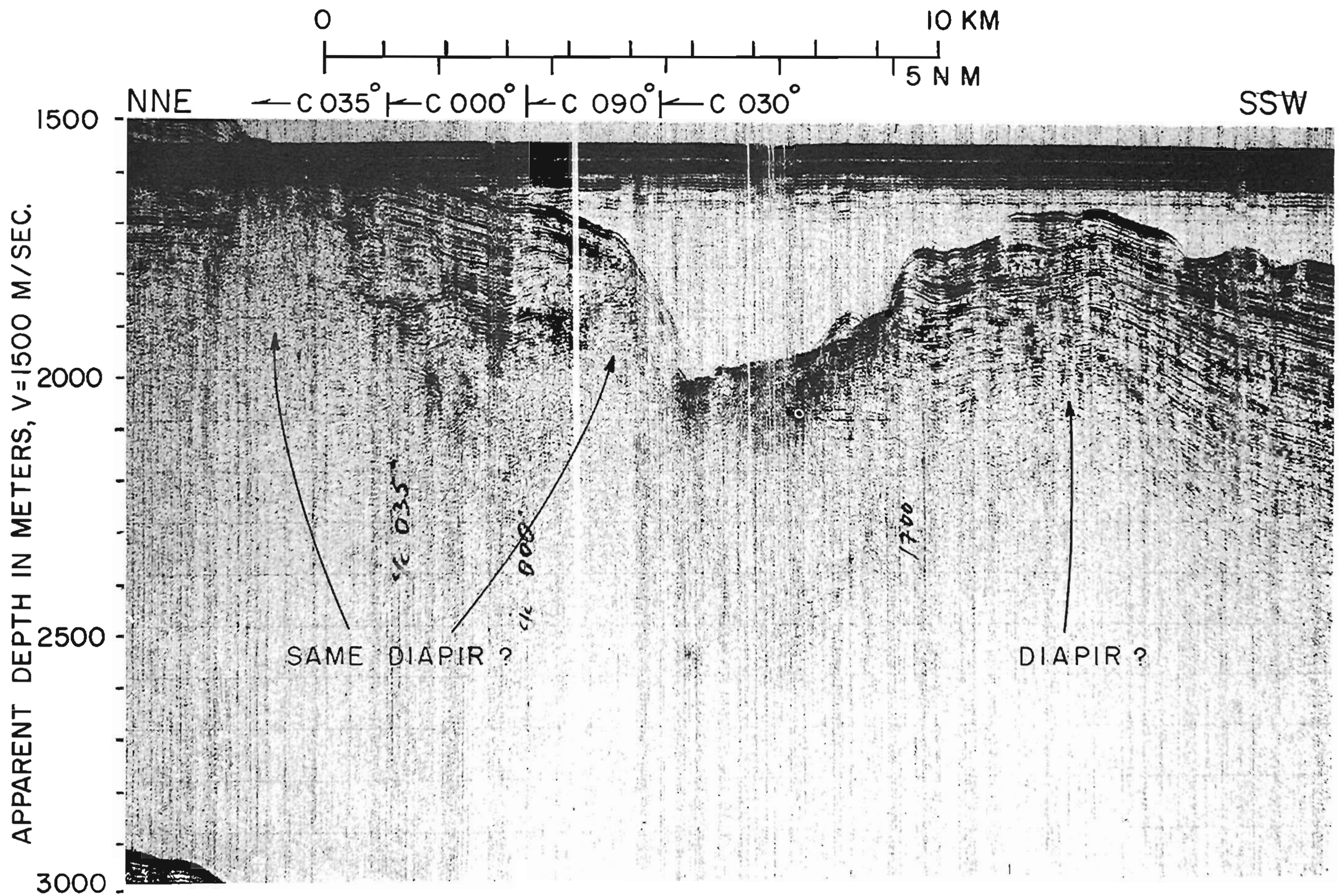

Figure 8. Diapirs on TRIDENT Line 33-34, centred at approximately 43° 50'N, 54° 15'W. At least one diapiric (?) uplift interpreted in southwest portion (to the right of the 1700 time mark), and two crossings of what may be a single diapir in shallower portion (to left), on either side of heavily marked 000 and 035 degree course change marks.

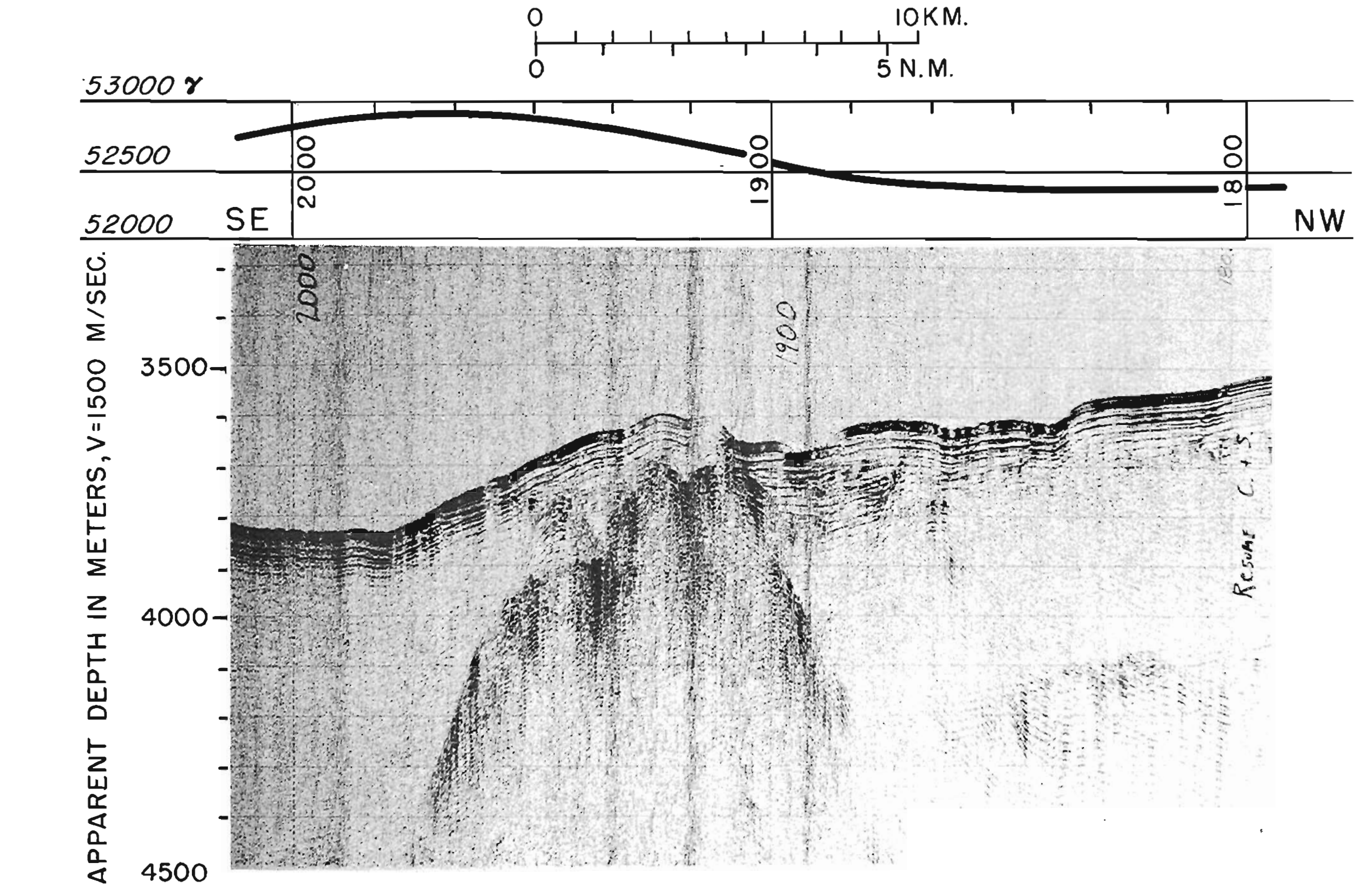

Figure 9. Probable igneous body at Location 36 on TRIDENT Line 35-37. A 500-gamma magnetic anomaly occurs over "dark" mass on this and subsequent crossing, indicating that it is igneous, either intrusive or a buried extrusive peak. Strong reflections from it are in contrast to reflection cutouts seen in Figure 8, believed to be diapirs of salt or mud. Disturbed portion of record from 1820 to 1900 hours could be salt or mud diapirism on the flank of the igneous body.

A probable exception occurs at Location 36, crossed on Line 35-37 and recrossed by Line 38-39 (Fig. 9). The bedding there is interrupted by an intrusion or buried peak, but the core exhibits higher seismic reflectivity than do most of the other diapir cores, and the magnetometer record peaked some 500 to 600 gammas on each of the two crossings. The feature is interpreted as a buried volcanic seamount, or perhaps an igneous intrusion, rather than a salt diapir.

Frequency of Occurrence of Salt Diapirs

For those lines which were essentially randomly oriented, such as Line 33-34, it is worth considering the probability of a diapir being discovered and the resultant implications concerning the number of salt structures likely to occur in the area. Consideration of salt-dome distribution in southern Louisiana and in the Isthmus of Tehuantepec, Mexico (Murray, 1961), reveal one dome per 380 km^2 and one per 185 km^2 respectively; one per 280 km^2 might be average, meaning the distance between domes (on a rectangular grid pattern) would average about 17 km, or every 9 to 10 nautical miles. If a typical shallow dome may be considered as being 2 miles (3.2 km) wide for recognition purposes (i.e., at least an arching on the record), then a random traverse would have one chance in five of seeing a dome when crossing a line of domes, and even with the domes not in a grid pattern the traverse should see about one-fifth of the domes within a 10 mile (16 km) wide swath. Thus one dome in as many as 50 miles (80 km) might be typical for a random traverse across such a salt dome province.

If the salt structures are in the form of anticlines a random line will intersect them more often. Trusheim's Figure 3, (1960) which is a map of salt structures in northwest Germany, reveals one ridge structure per 20 km across strike, one per 35 km along the average strike (the structures curve somewhat), or perhaps one per 25 km over all (about one per 13 nautical miles). For comparison, the group of east-west folds mapped here in the Laurentian Channel average 4 nautical miles (7.4 km) apart for the small ones, 10-12 nautical miles (18.5-22.2 km) for the large ones. The two known diapirs north of Sable Island are 8 nautical miles (14.8 km) apart, and the several supposed diapirs on Lines 33-34 and 34-35, off the Grand Banks shelf, average 8 to 10 nautical miles (14.8 to 18.5 km) apart.

It is concluded that the structures encountered in the Laurentian Channel, north of Sable Island, and off-shelf west of Tors Cove are distributed in a manner compatible with a salt dome province interpretation, and that if anything their spacing on the records is more suggestive of salt ridges than of isolated circular salt domes.

SUMMARY AND CONCLUSIONS

Seismic traverses, in part randomly located, reveal easterly trending anticlines and diapirs in the Laurentian Channel north of Lat. 45° N, possibly changing to southwesterly trends beneath the banks west of the Channel. Salt was penetrated in one well southwest of Sable Island. Several domes and probably broad areas of salt swells underlie the banks immediately north and northeast of Sable Island, and the detailed bathymetry of Chart 4041 shows several circular or oval patterns as if broad swells underlie the entire Banquereau

Bank northeast of Lines 22-26. It is interpreted that one salt structure province does underlie the entire Sable Island to St. Pierre Bank area, the salt of necessity being Windsor salt at least as far south as about 45° N; younger salt may occur nearer the shelf edge. This salt province is bisected in its western portion by the basement ridge off Cape Canso south of the Chedabucto Fault. The salt structure area probably extends northward around the plunging eastern end of Scatari Ridge into the Sydney-St. Pierre Basin, past the cluster of poorly defined structures seen near the Mira Bay Fault at 46° N, 57° -58° W. It is not known how far these structures continue eastward beneath St. Pierre Bank. The basement ridge hitherto mapped as extending northeastward to Sable Island beneath the outer shelf edge is much deeper than previously reported, as required by new drill hole results. Salt could have been interpreted as basement based on high seismic velocities, in which case the "ridge" might in fact be a complex of salt structures rather than a basement uplift.

A number of structures presumed to involve salt movement are mapped in the deeper water west and southwest of the Tors Cove D-52 well which is known to have penetrated a salt dome or ridge, and similar structures are seen northward from the Grand Falls H-09 well. There may be two clusters of such structures beneath the western Grand Banks, separated by a continuation of the Chedabucto Fault of Nova Scotia.

Although an angular unconformity occurs between Cretaceous and Tertiary rocks in the Misaine Bank area (King and MacLean, 1970a), and Bartlett and Smith (1970) date the latest movement of salt at Tors Cove as Early Eocene, the recognition of the Tors Cove and neighbouring structures on shallow seismic records indicates that some diastrophism has occurred in late Tertiary to Quaternary time.

ACKNOWLEDGMENTS

The author wishes to express his deep appreciation to Captain Barnes Collinson and the crew of R/V TRIDENT for their co-operation, to the University of Rhode Island Graduate School of Oceanography and its funding agencies (National Science Foundation, Office of Naval Research) which made possible the use of the vessel, to Dale C. Krause and Jean-Guy Schilling for their valuable advice, and National Science Foundation which financed part of the writer's own activities through Grant GA-518. Lewis King and Brian MacLean kindly devoted several days to a comparison of records during the writer's visit to Bedford Institute in March of 1971, and Alan Ruffman and Elazar Uchupi provided copies of records in the Laurentian Channel area. Thomas Casadevall, Martin Fisk, Frederick Frodyma, Frederick Haug, Michael Page, Frank Raffaldi, John Richmond, and James Wessel, students, and William Hahn, Joel Knee, and Frank Rose, oceanographic technicians, served ably in science parties at sea. Carl Hobbs worked on Cruise 73 records and Richard Skryness did a Senior Honors thesis on the Cruise 84 data at the University of Massachusetts (Skryness, 1971). Errors and omissions are the author's own.

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Geol. Surv. Can., Paper 71-23, 1973, p. 219-222)

13. NOVA SCOTIA, MOROCCO AND CONTINENTAL DRIFT

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The theory of plate tectonics has revolutionized the study of mountain belts on the earth's continents. This concept grew from geophysical, sedimentological, and petrological studies mainly of the North Atlantic basin. On the western flank of this basin, the classical Appalachian orogen is a test-case for the extrapolation of the theory back into geological time.

Previous work has concentrated on the application of plate tectonics to the northwestern Northern Appalachians. Contributors at the Gander Conference (Kay, 1969) extended the orogen northeastward to northern Ireland, northern England and Scotland. Bird and Dewey (1970) expanded the following ideas which were originally advanced by Wilson (1966). Thus Late Precambrian rupture of a megacontinent gave birth to a proto-Atlantic. This Atlantic widened throughout the Cambrian and Early Ordovician by production of an ocean-floor plate along a median, accreting margin similar to the present Mid-Atlantic Ridge. During the Middle Paleozoic, plate consumption along both sides of the proto-Atlantic narrowed this seaway. The two major orogenies in the Northern Appalachians record the collision of North America first with island arcs (the Ordovician Taconian) and finally with Africa (the Devonian Acadian). In the Middle Mesozoic, rifting again of the mega-continent initiated the present North Atlantic, which is still widening. Such trans-oceanic correlations are most visible parallel to tectonic/sedimentologic/metamorphic facies belts. Correlations across such belts, as proposed in the model presented, would be very hazardous without the medium of the Meguma massif of Nova Scotia and the Anti-Atlas area of Morocco.

In preview, southeastern Atlantic Canada may have been built by several cycles of the following stages: marine sedimentation in a previous North Atlantic ocean; continental collision between Africa and North America; subaerial redbed sedimentation and volcanism; and continental rifting initiating a younger North Atlantic ocean. The subsequent rifting did not occur exactly along the earliest margin. Rifting was very ragged, so that parts of one continent were transposed to the other, and sialic fragments became offshore micro-continents embedded in oceanic plate.

The oldest-known rocks of the Maritimes, the George River of Nova Scotia and Green Head of southern New Brunswick, can be correlated with the Paleohelikian Precambrian of Morocco. This miosynclinal assemblage built the western continental shelf of Africa, and probably extended northwestward across the Grenville Province. Late Precambrian (Grenvillian) continental collision between Africa and North America welded the two continents together. The Fourchu and Morrison River of Cape Breton, the Coldbrook of New Brunswick, and the Harbour Main/Love Cove of Newfoundland can be correlated in part with the early Hadrynian Ouarzazate System of Morocco. This pyroclastic/redbed sequence records mainly subaerial, intermontane deposition in rift zones. The fragmentary break-up of the mega-continent initiated

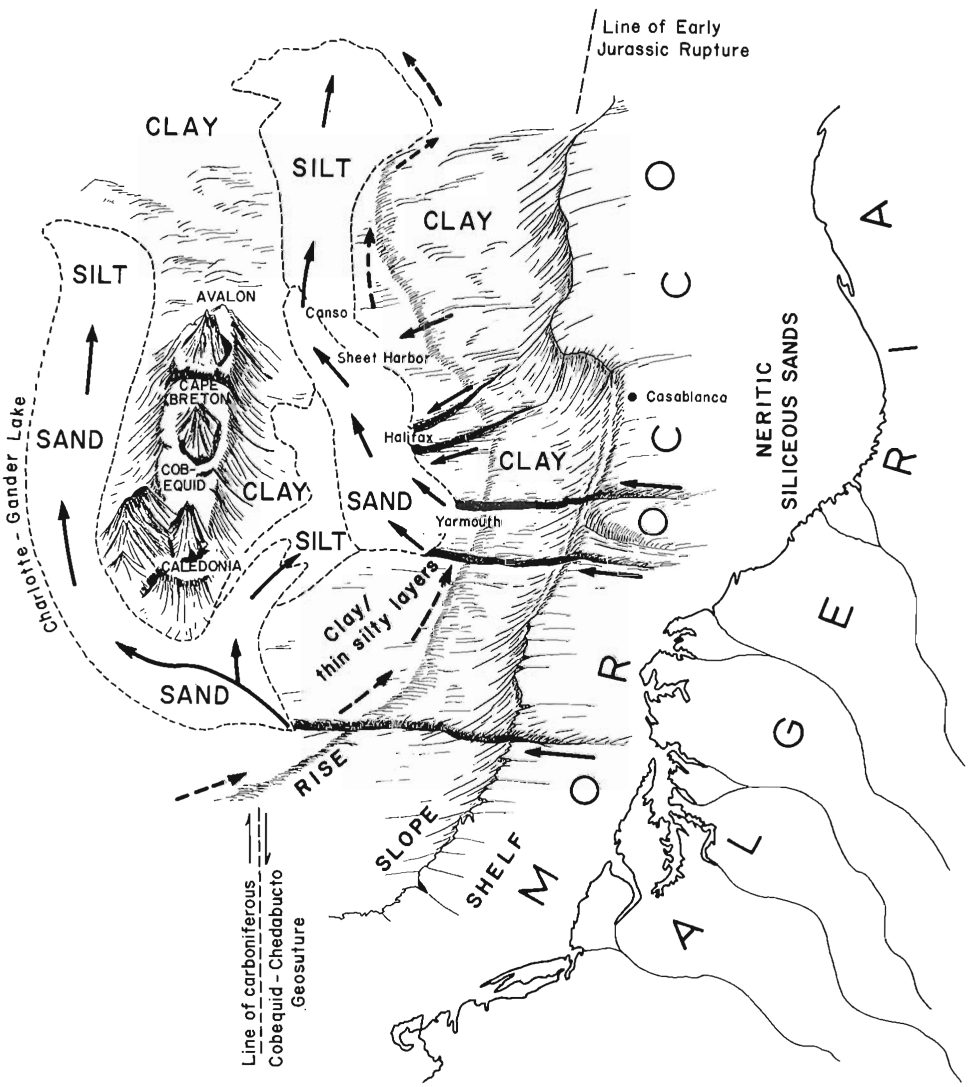

Figure 1. Speculative setting along the "eastern" Paleozoic North Atlantic during deposition of the Meguma Group. Delta complex from Algeria; continental shelf in Morocco; upper Meguma perhaps a lower continental rise contourite; lower Meguma perhaps an abyssal fan. Rotation of diagram 180 degrees restores setting to Recent sedimentation along the western North Atlantic with the sialic block as the Bermuda Rise (after Horn et al. 1971).

the proto-Atlantic and created sialic fragments which became offshore micro-continents. The broken block of Cape Breton and the Cobequid Mountains of Nova Scotia, and the Caledonian Mountains of New Brunswick are fragments of this micro-continent. Present analogies are possible with the Danakil Alps of the Red Sea area, the Canary Islands off Morocco, and the Flemish Cap-Orphan Knoll off Newfoundland. Late Precambrian continental glaciation of Africa is recorded both there and in the Conception Group of southeastern Newfoundland (Bruckner, in press). However just as later in the Pleistocene, both land masses bordering the Late Precambrian Atlantic were glaciated. From the Cryptozoic into the Phanerozoic, carbonate sedimentation was continuous on the African shelf, the western rise-complex was starved, and the offshore micro-continent was slowly flooded as it moved eastward with Africa away from the up-arching accreting margin. North America was carried west.

During the Paleozoic, rise-complexes derived from the African Precambrian Shield, spread as aprons around the micro-continent, Africa collided again with North America in the Middle Devonian, and continental redbeds and volcanics accumulated over the sutured region. The Meguma Group of Nova Scotia is part of a Cambro-Ordovician rise complex, the other part remaining in Africa as part of the Atlas area and perhaps as the base of the eastern Canary Islands. Other parts of this complex which smothered the micro-continent are the Charlotte Group of New Brunswick and Maine, and the Gander Lake/Baie d'Espoir units of southeastern Newfoundland (Fig. 1). Late Ordovician continental glaciation is unique in northwestern Africa. Possible glacial erratics in the White Rock Formation may suggest not only the ages of this unit and that of the underlying Meguma Group, but also correlation of the formation with the Dunn Point Volcanics of northern Nova Scotia as well as the idea that the province is African in origin. During the Ordovician and Silurian, intense volcanism occurred on the flanks of the micro-continent. Continental collision with North America sandwiched the micro-continent and accompanying volcanic/sedimentary rise-complexes between America and Africa. Initial collision occurred in the Middle Devonian but adjustments continued into the Late Carboniferous. The Windsorian marine units may record vestiges of the proto-Atlantic. During the Carboniferous and Early Mesozoic, sub-aerial redbeds and volcanics accumulated over the suture of the mega-continent. Beginning in the Carboniferous, strike-slip movement along the Cobequid/Chedabucto and the South Atlas faults moved the Meguma massif and the Atlas area toward their present positions. Middle Mesozoic rifting initiated the present North Atlantic. However, the tensional stress-field was complicated by the proto-Atlantic micro-continent, so that separation was again very ragged. The old rise-complex of the Mauritanides in the Anti-Atlas area was plucked as a block off Africa. Part remained clinging to the micro-continent in North America as Nova Scotia and so moved westward away from the accreting Mid-Atlantic Ridge, part straggled eastward with Africa as the eastern Canary Islands. Rise-complexes derived from the African craton are now forming off Africa and are flanking its offshore volcanic micro-continent – "The present is the key to the past".

In conclusion, at least three North Atlantic oceans may have existed – during the Middle Proterozoic, the Early Paleozoic, and from the Late Mesozoic-Recent. The Early Paleozoic Atlantic was closed at least along the latitude of Nova Scotia-Morocco. Deposits of continental glaciation in the Late Precambrian and the Ordovician may establish inter-regional and inter-continental chronostratigraphic units, and may also discriminate between drift models to indicate that Nova Scotia is a remnant of northwestern Africa. The cycle of marine sedimentation, collision, continental sedimentation, and rifting is repetitive, so that stages of earlier cycles may be compared both to later phases, and to the Recent. The model is a crude, over-simplified test-case which will be tested by work presently being done in both Morocco and Nova Scotia (Schenk, 1971, 1972).

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(Earth Science Symposium on Offshore Eastern Canada,
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14. MARINE SEISMIC OPERATIONS: OFFSHORE EASTERN CANADA

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Abstract

Catalina Exploration and Development Ltd. of Calgary were the first Canadian company to carry out marine seismic operations as a contract service in the search for hydrocarbons on the east coast. A fishing vessel which was suitable for the purpose was acquired, refitted and rechristened the M/V CALGARY CATALINA. Seismic surveys have been carried out by the CALGARY CATALINA on the Grand Banks of Newfoundland, the Scotian Shelf and in the Gulf of St. Lawrence. The resultant seismic cross-sections show a variety of interesting geological features such as diapirs, faults, graben, and some structures of unknown origin and significance in the possible entrapment of hydrocarbons.

INTRODUCTION

Having first established that there was an adequate Canadian market to support continuing marine seismic operations, the next step for Catalina Exploration and Development Ltd. was to acquire a suitable vessel. After specifying the optimum characteristics for a marine seismic ship, a suitable ship was acquired whose characteristics corresponded closely to the ideal and it was subsequently christened the M/V CALGARY CATALINA (Fig. 1). This motor vessel had been originally built as a tuna seiner in the shipyards at Luzon, Quebec. It has a length of 167 feet, a beam width of 36 feet, and a gross tonnage of 975.26 long tons. The M/V CALGARY CATALINA has a cruising speed of 12 knots, and there is adequate room for personnel accommodation and equipment.

The ship was sailed from Halifax to Picton, Nova Scotia, where it was refitted, equipped and brought to Canadian Shipping Inspection (CSI) standards, the latter at considerable extra cost in time and money. The refit required clearing the stern and central part of ship of all fishing gear. The aft end of the ship had to be redesigned and the vessel shortened a couple of feet to permit efficient handling of the marine geophone cable (streamer). The major items of equipment installed in the ship were an energy source and compressor, geophone streamer and seismic instrumentation. The energy source chosen for the system was a tuned array of 24 Bolt air guns. These were mounted in the stern of the ship on sliding beams that could be moved aft over the stern so that the guns could be lowered into the water. The guns varied in volume from 10 to 40 cubic inches and were charged by a Norwalk five-stage compressor which was capable of delivering 250 cfm at 2000 psi pressure. The Norwalk compressor was mounted amidships between the decks. The seismic recording equipment consisted of a 24-trace DFS III 21-track binary-grain amplifier system with read-after-write transport and SIE camera. This

*Now of Caravel Exploration Ltd., Calgary.

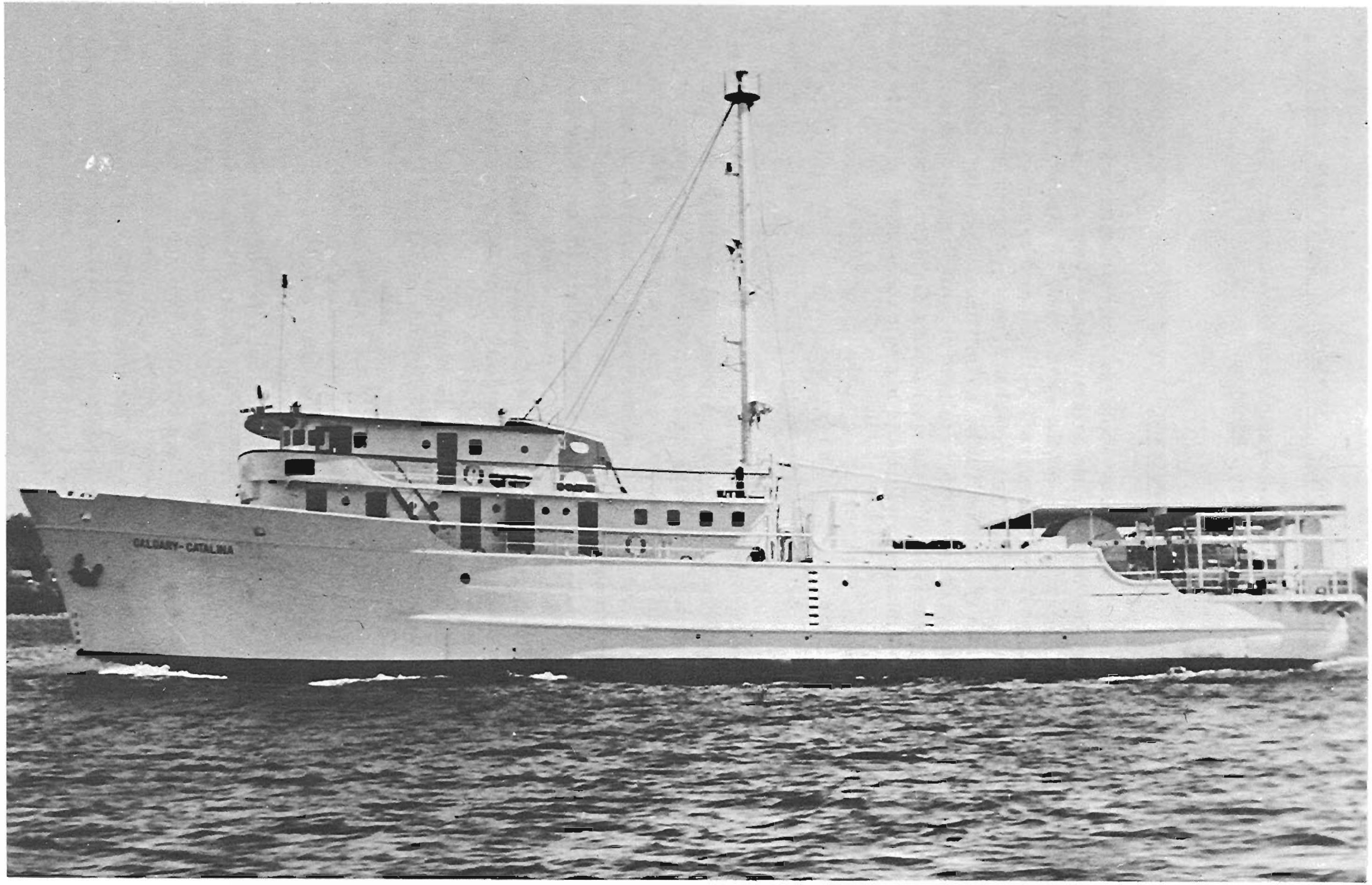

Figure 1: The M/V CALGARY CATALINA of Catalina Exploration and Development Ltd., Calgary.

highly sophisticated recording system was mounted in a dog house (instrument cabin) on the main deck adjacent to the work area aft (see Fig. 1). A 7200-foot marine seismic streamer was housed aft on a 12-foot diameter reel beneath a steel canopy (see Fig. 1) and was controlled both electrically and hydraulically.

Thus the CALGARY CATALINA was equipped to gather large volumes of seismic data quickly using techniques which provided very high percentages of multiple coverage. The data thus acquired was subsequently processed using computer techniques which have developed rapidly over the past several years.

PROCESSING OF MARINE SEISMIC DATA

Large reconnaissance surveys, must have a processing sequence designed to fit a broad area, and at the same time cope with special problem areas. Three of these special problem areas are:

1. Geologic - Where the best combination of field and data centre parameters is not readily apparent to solve the problem, as an example, a shallow high velocity section.
2. Reverberation - Reverberations of such strength that deconvolution and stacking may not provide the attenuation necessary, will require special processing techniques.
3. Velocity Control - Gathering of proper velocity information is dependent upon data quality and frequency of sampling.

The processing sequence can be divided into four operations which are illustrated in Figure 2.

1. Input and preliminary processing
2. Parameter determination
3. Parameter application
4. Final stack and display

The actual processing sequence is:

1. Quality control of field records
2. Gain recovery
3. Vertical summing
4. Predictive deconvolution
5. Time variant deconvolution
6. Normal moveout
7. Digital filter
8. Stack
9. Section modulation, if necessary
10. Film display

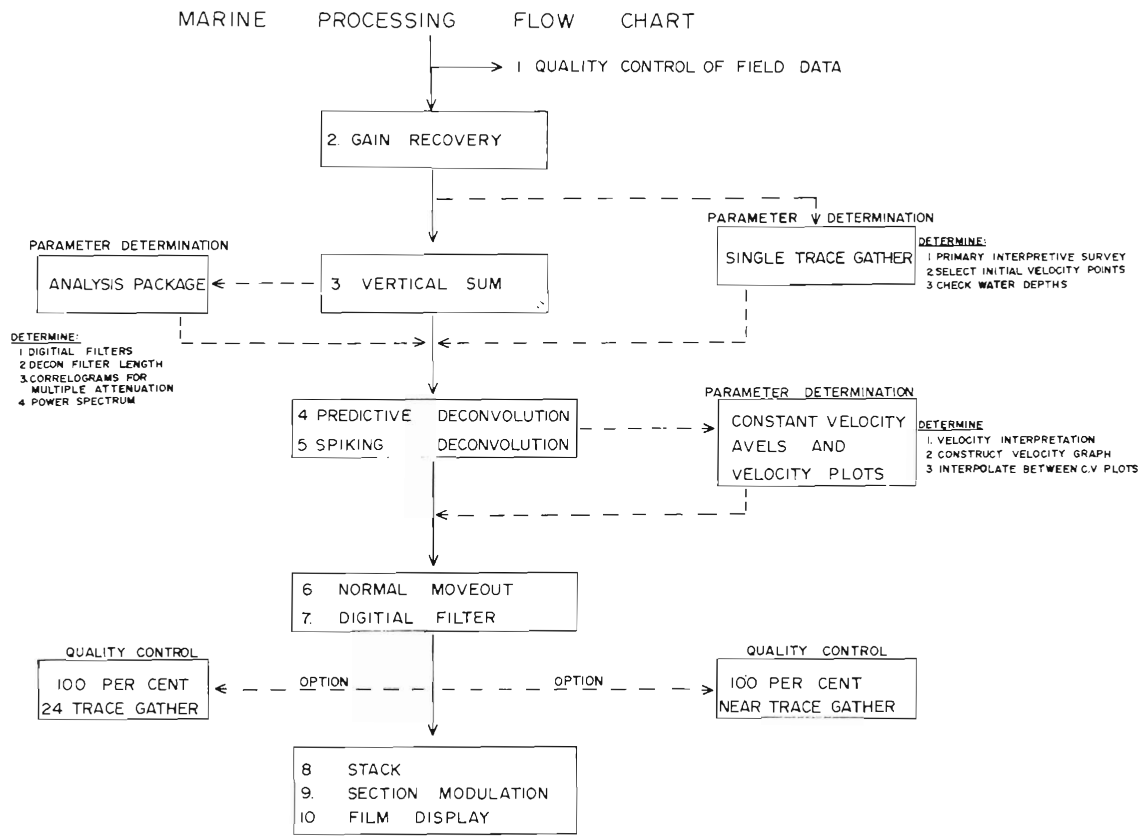

Figure 2. Marine processing flow chart.

Parameter Determination

1. A Single Trace Gather

A single trace gather is taken out between steps two and three, is scaled and filtered, and provides:

a. A preliminary interpretive look at the section being processed,

b. a check on the accuracy of the water depth,

c. the location of velocity control points which are influenced by geology and water bottom contour. This section can be processed through deconvolution, normal moveout and digital filtering and is presented as a final end product.

2. Analysis Package

An analysis package is run at step three and provides:

a. From digital filter sweeps and power spectrum the hi-cut and lo-cut filter points and if necessary, the start times and overlap zones if the filters are to be time variant.

b. Deconvolution analysis, which includes length of filter decided, gate start and end times and overlap zones. From the autocorrellogram the effective multiple attenuation of the predictive deconvolution filter is ascertained. The gap in the operator is from the water bottom depth or nth zero crossing from autocorrelation. Autocorrelation displays are prepared before and after deconvolution.

3. Velocity Analysis

Velocity analysis is performed after step five and can take a number of forms. One effective method is by constant velocity stacking.

Constant Velocity Stacking

The records needed to make up one stacked 24-trace record are gathered and filtered. A constant velocity is used from T_o to T_{max} to apply moveout. The records are then stacked and the output displayed. On the Scotian Shelf for example, the starting velocity is 5000 ft/sec, the velocity is increased by an increment of 500 ft/sec and individual stacks made for each velocity up to 8000 ft/sec. The velocity increment is then increased by 1000 ft/sec and sections run for velocities up to 12,000 ft/sec. This results in 11 stacked records and provides first, an interpretive look at the section in terms of velocities and second, the output can show geologic changes more easily than the automatic velocity plot described below. Another procedure for velocity determination is using AVELS and velocity power plots.

AVELS and Velocity Power Plots

In this method, 12 records are gathered and vertically summed to provide an increased signal-to-noise ratio. The resulting record is muted to remove noise trains, deconvolved and filtered. A starting velocity or delta T is specified and a velocity or delta T increment, to increase the starting velocity. The velocity or delta T is applied to the record and it is stacked to provide one output trace. The increment is applied to the velocity or delta T and the record stacked again and continued to a maximum output of 96 traces. The resulting output traces can be outputted to a plotter for visual display or to a line printer for a printout. Power plots are then prepared. The output records after AVEL are divided up into gates and the velocity spectral power is computed along a trace for a given velocity, the gate powers are then averaged horizontally and vertically providing intermediate velocity control.

4. Quality Control

Quality control consists of a 100% near gather and a 100% 24 trace gather taken out just before stacking and while optional is extremely useful in section interpretation providing:

1. Final analysis of parameter selection before stack.
2. Additional interpretive information to client.

Figure 3 demonstrates the effects of the use of inadequate velocity control in processing marine data. It also suggests that the geophysicist must be very prudent in deciding whether data of poor quality is the result of faulty processing techniques or the result of inappropriate data acquisition procedures and parameters.

Offshore Seismic Operations by Catalina

To demonstrate the value of marine seismic operations in throwing light on the geology of the area surveyed, a number of seismic cross-sections taken from Catalina's data library will be discussed. These sections are located in the circled areas on Figure 4 which also shows the location of Catalina's existing and proposed control.

Grand Banks of Newfoundland

Figure 5 indicates the change of regional dip of the sediments to the east of the Avalon Peninsula on the Grand Banks. There is a hinge line near the right side of the figure where the rate of dip changes abruptly from approximately 300 ft/mile (57 m/km) to twice that rate.

Figure 6 shows the geological section located on the east side of the Flemish Pass. Note the change in dip between the shallower and deeper events resulting in an eastward termination of the sedimentary section.

Figure 7 shows a continuation of this profile westward to the western side of the Flemish Pass. The sedimentary section on this side of the Flemish pass wedges out toward the west. There is a small basement anomaly on the west edge of the pass. The water depth varies from approximately 960 feet (290 m) to 4000 feet (1200 m) across the Flemish pass.

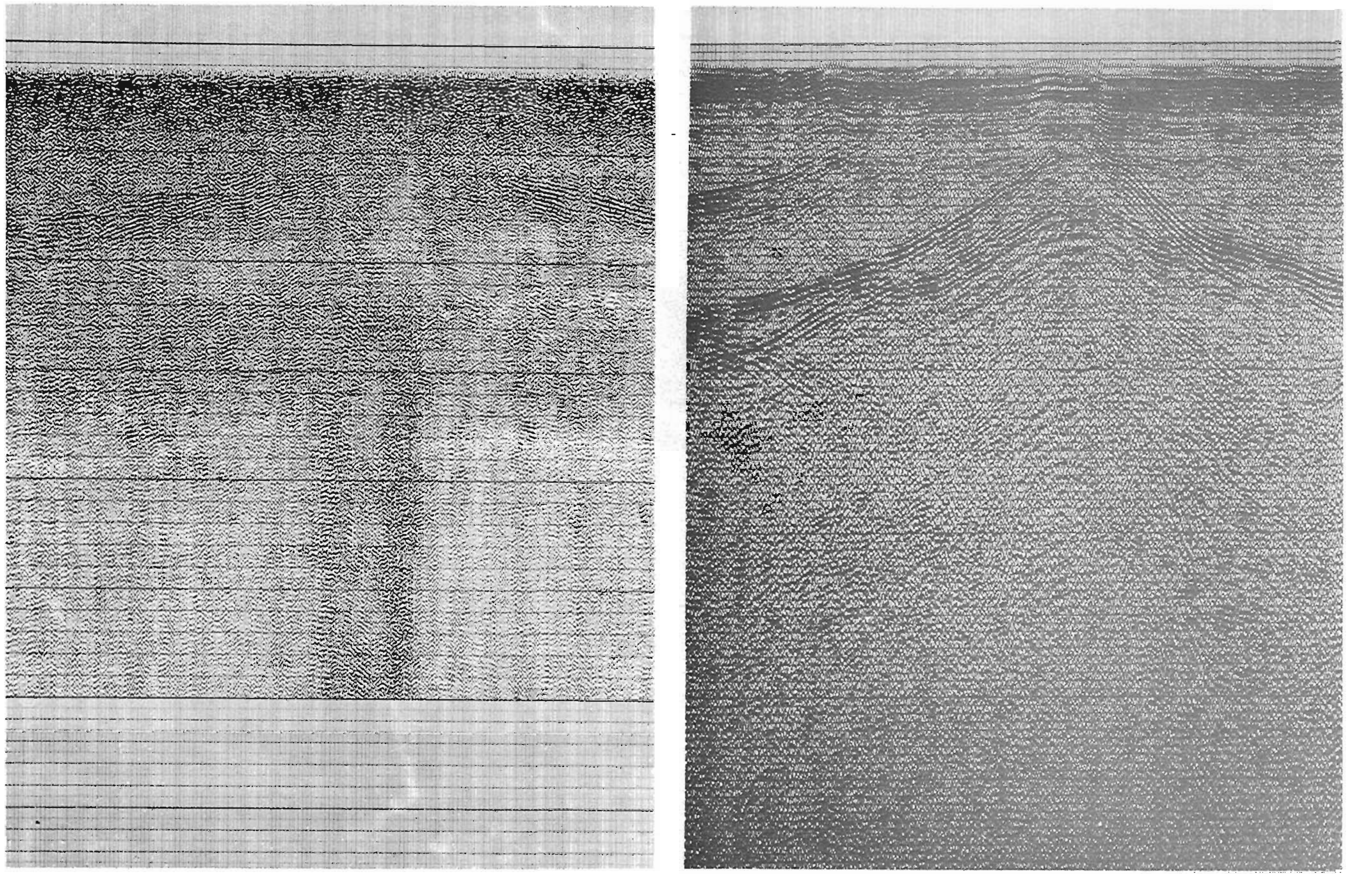

Figure 3. The same seismic cross-section processed in two different ways. The cross-section on the right was processed using inadequate velocity control.

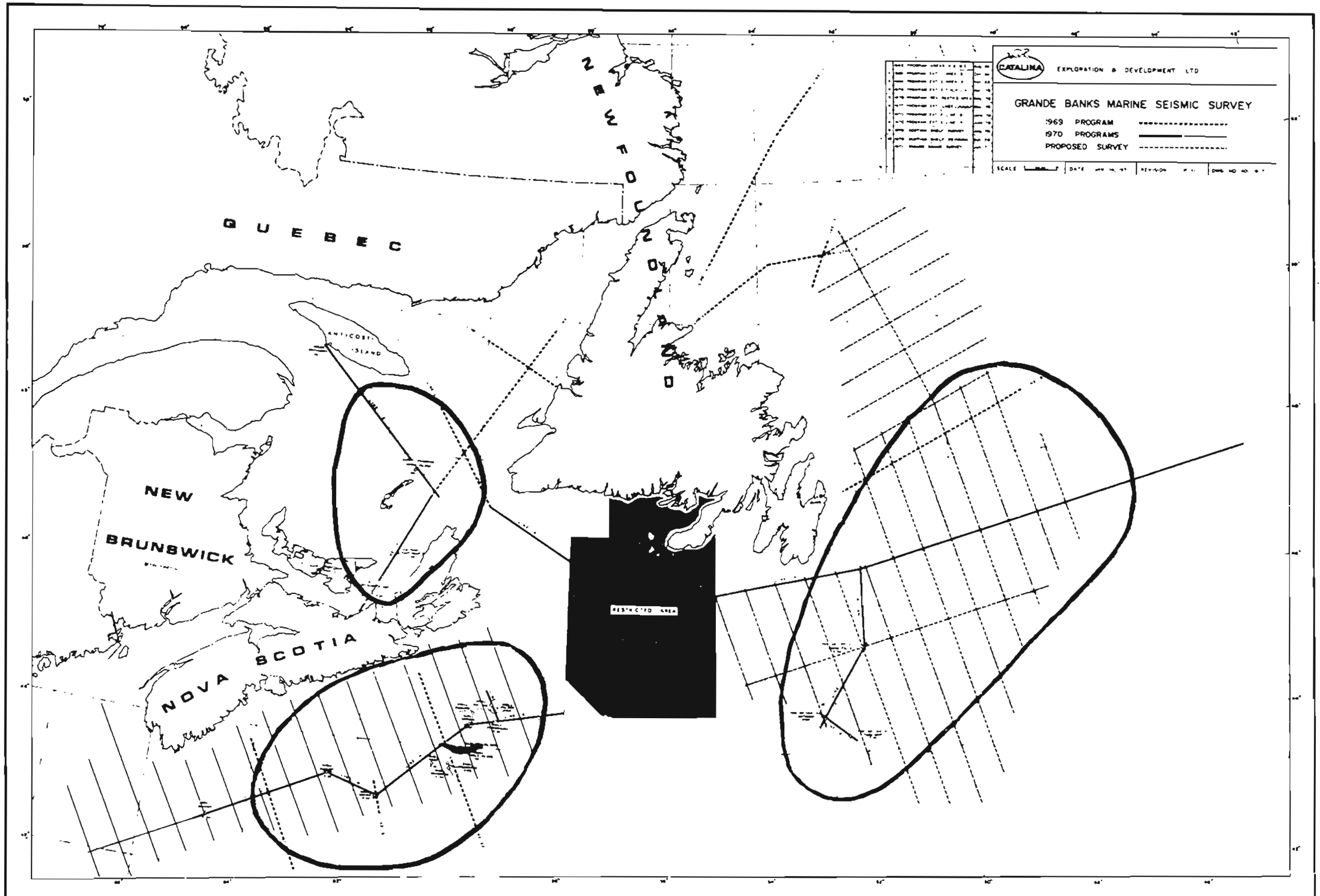

Figure 4. Location map showing seismic profiles acquired by Catalina.

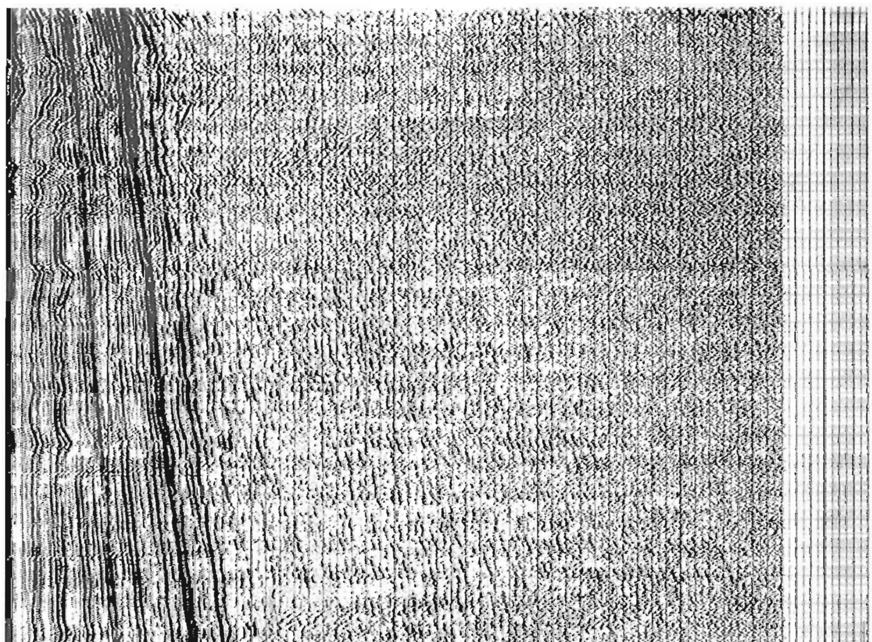

Figure 5. Seismic profile located to the east of the Avalon Penninsula, Newfoundland.

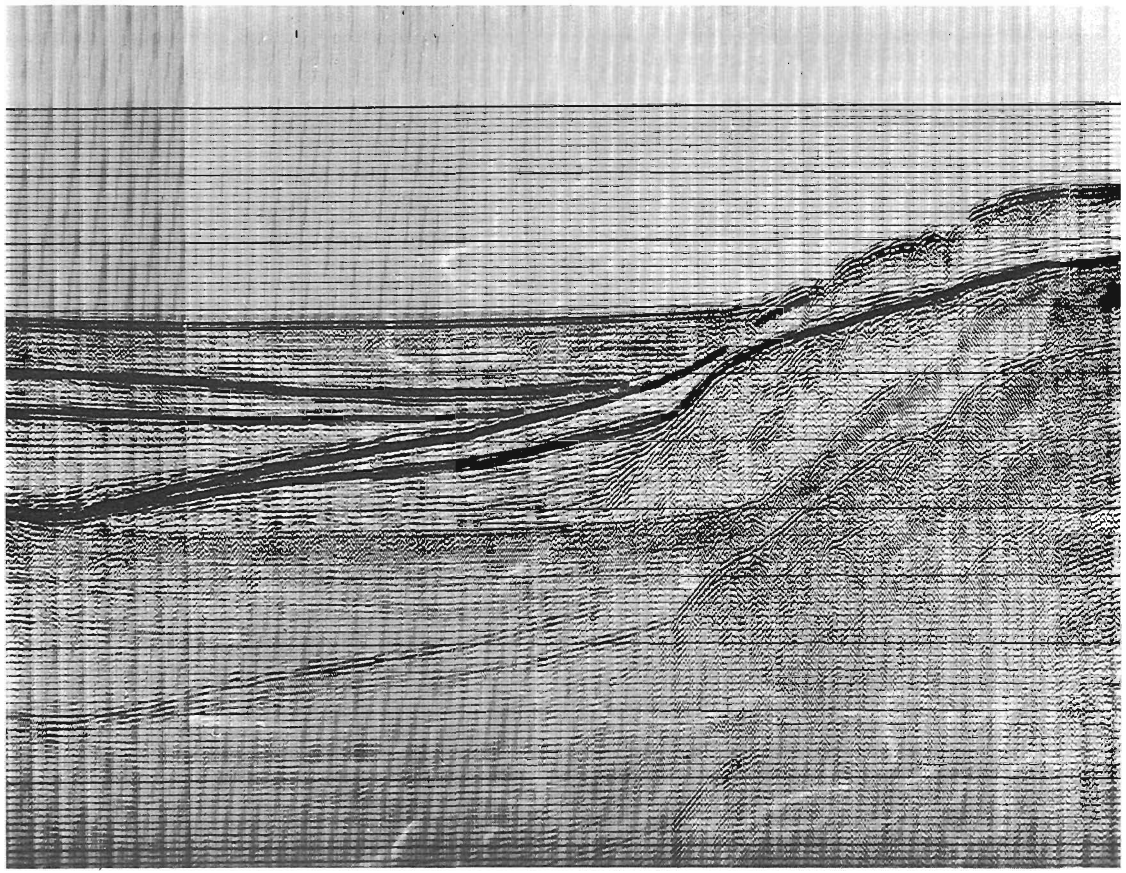

Figure 6. East-west seismic profile located on the east side of Flemish Pass and includes part of the Flemish Cap.

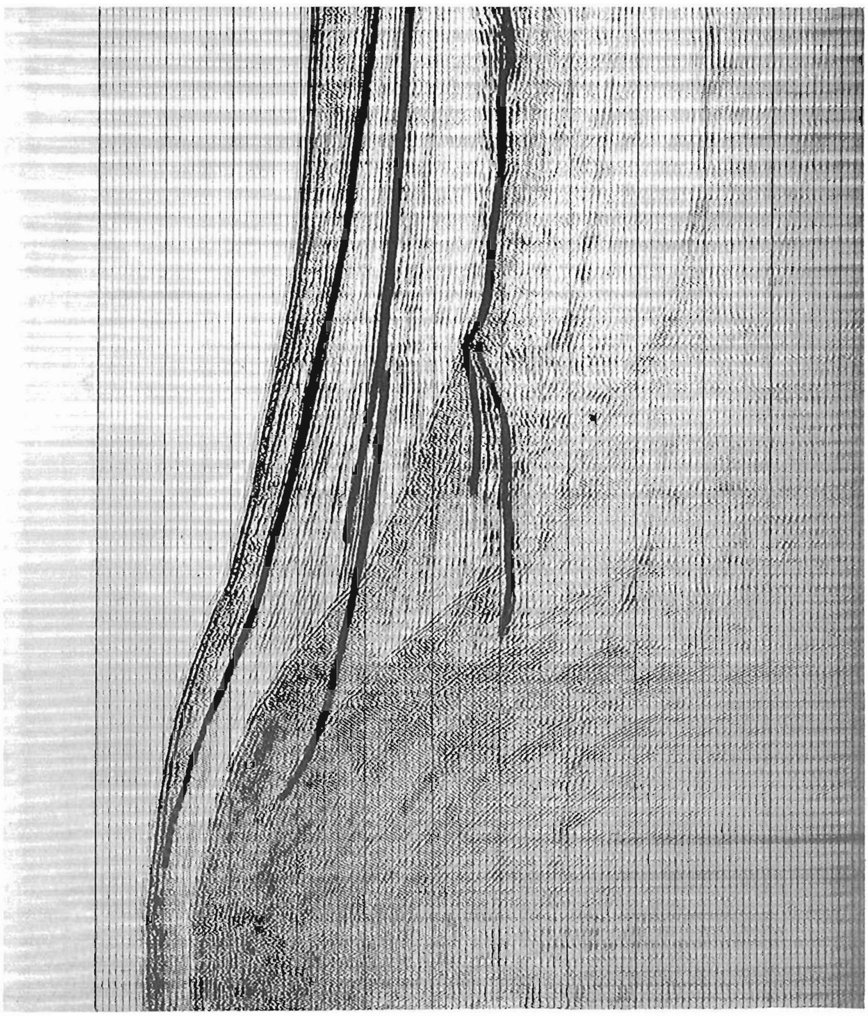

Figure 7. East-west seismic profile located on the west side of Flemish Pass.

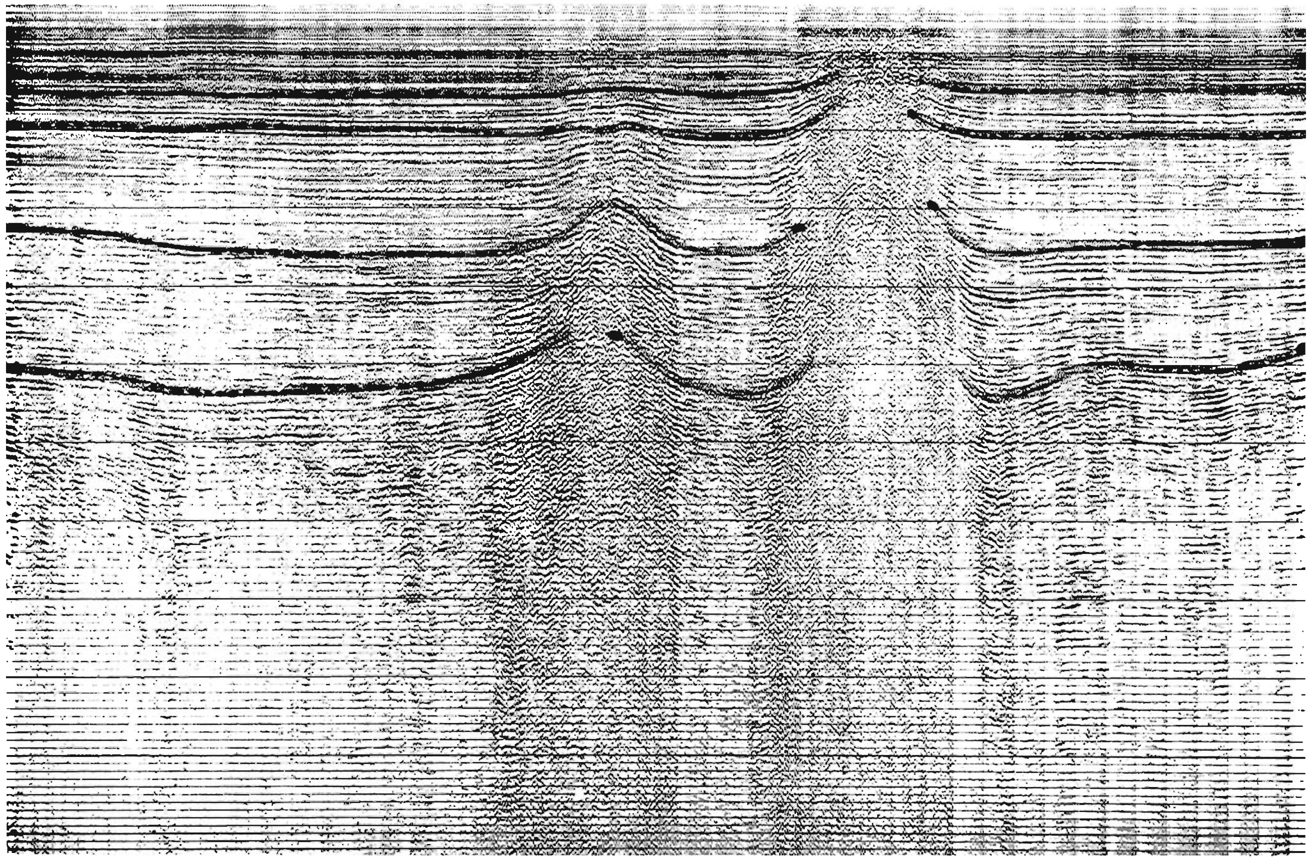

Figure 8. Northwest-southeast seismic profile located south of the Avalon Peninsula, Newfoundland Two bodies, possibly salt diapirs, intrude the sedimentary section.

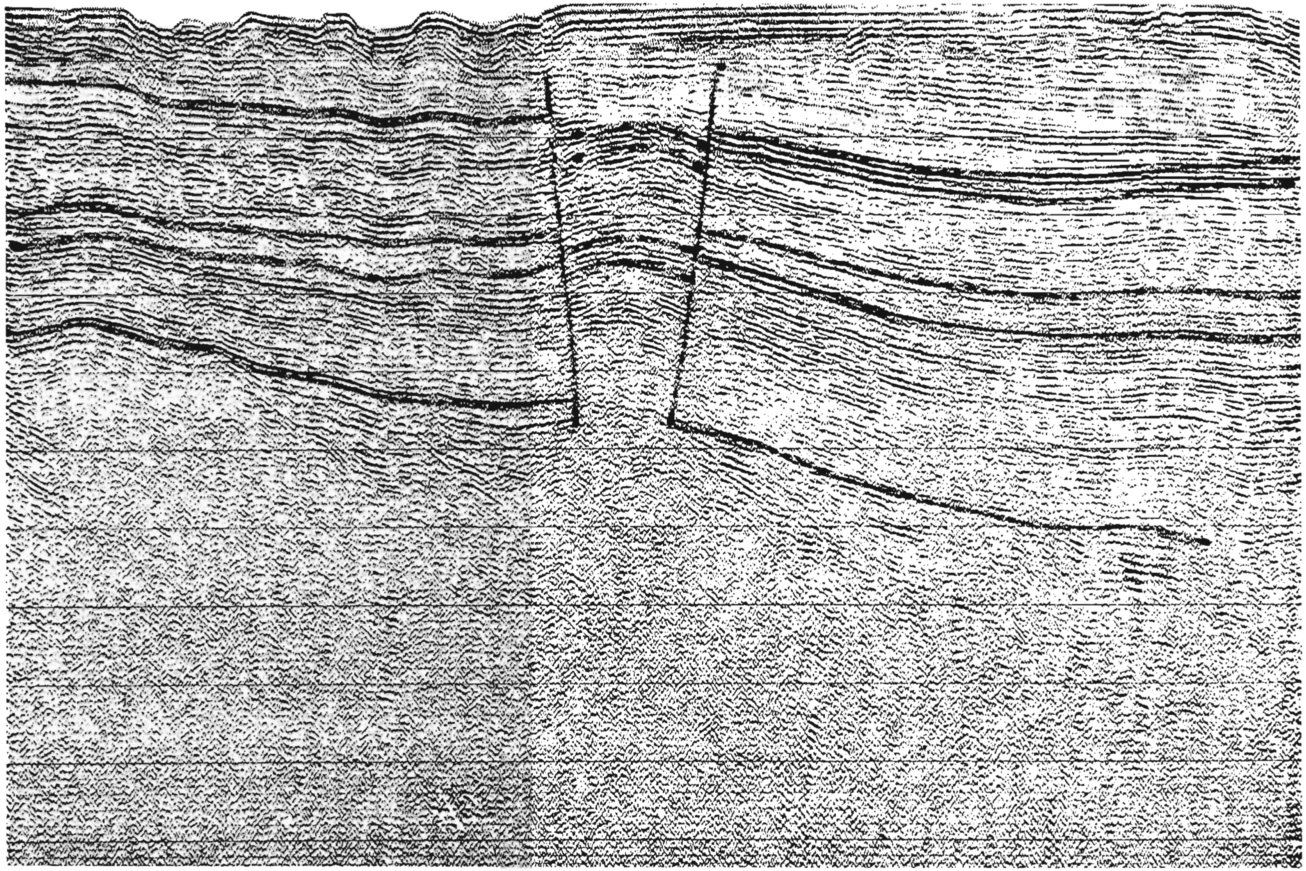

Figure 9. Northwest-southeast seismic profile located near the north end of the Scotian Shelf. Note the graben in the centre of the profile.

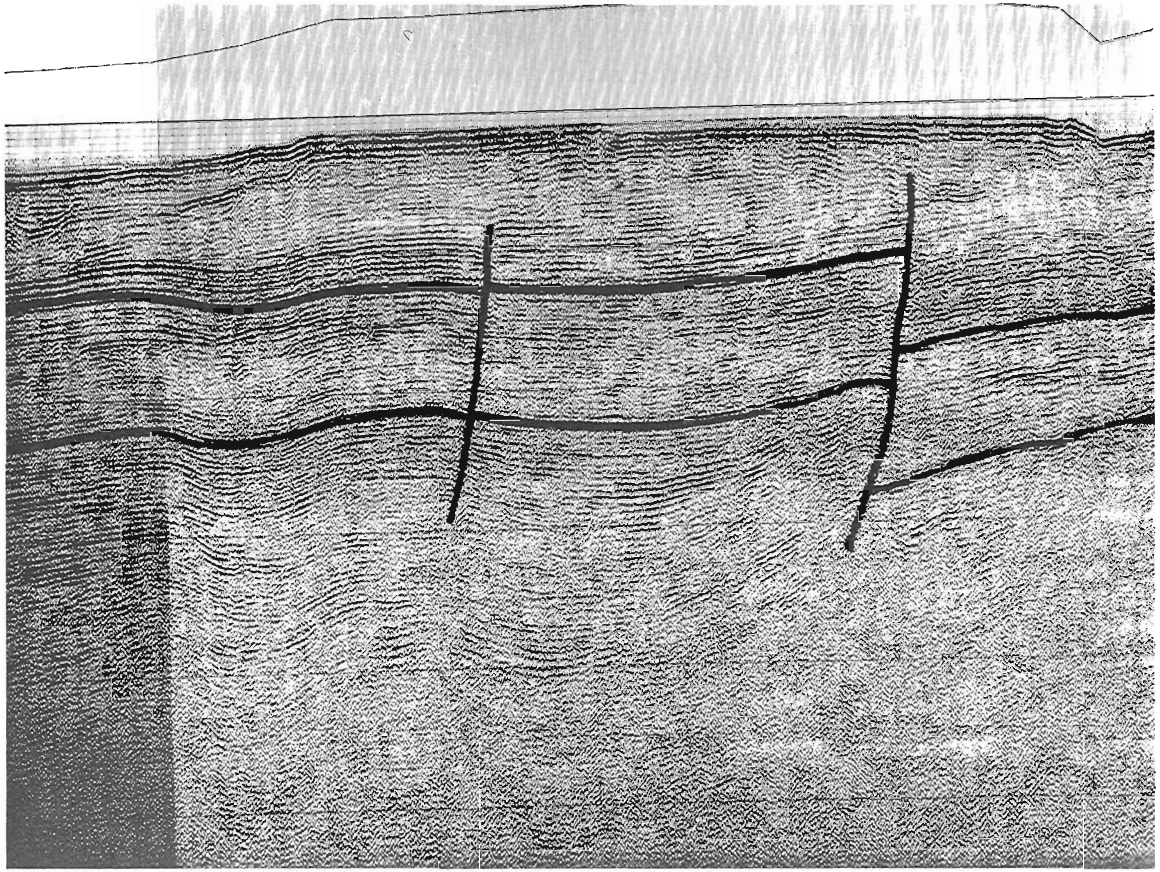

Figure 10. Northwest-southeast seismic profile located on the Scotian Shelf on which a reverse fault is apparent.

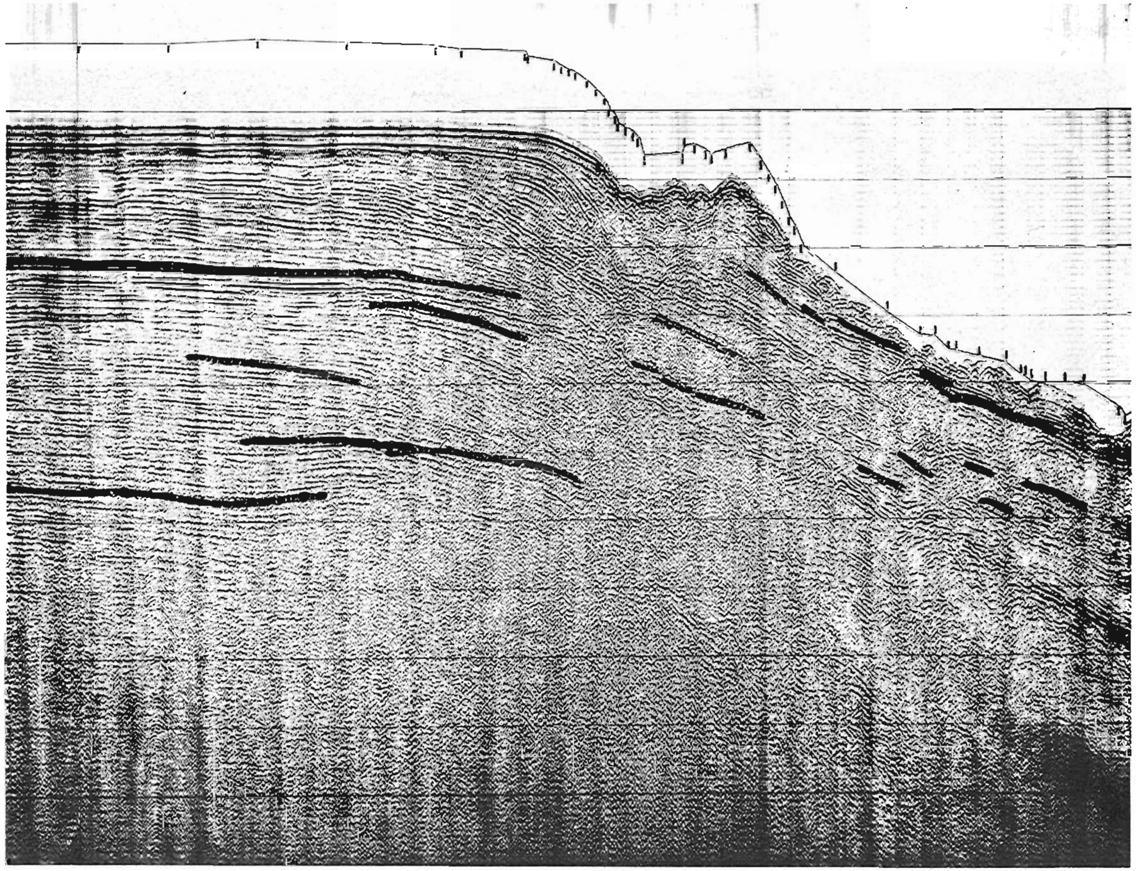

Figure 11. Seismic profile across the continental shelf edge and slope on the Scotian Shelf which indicates that some slumping has taken place.

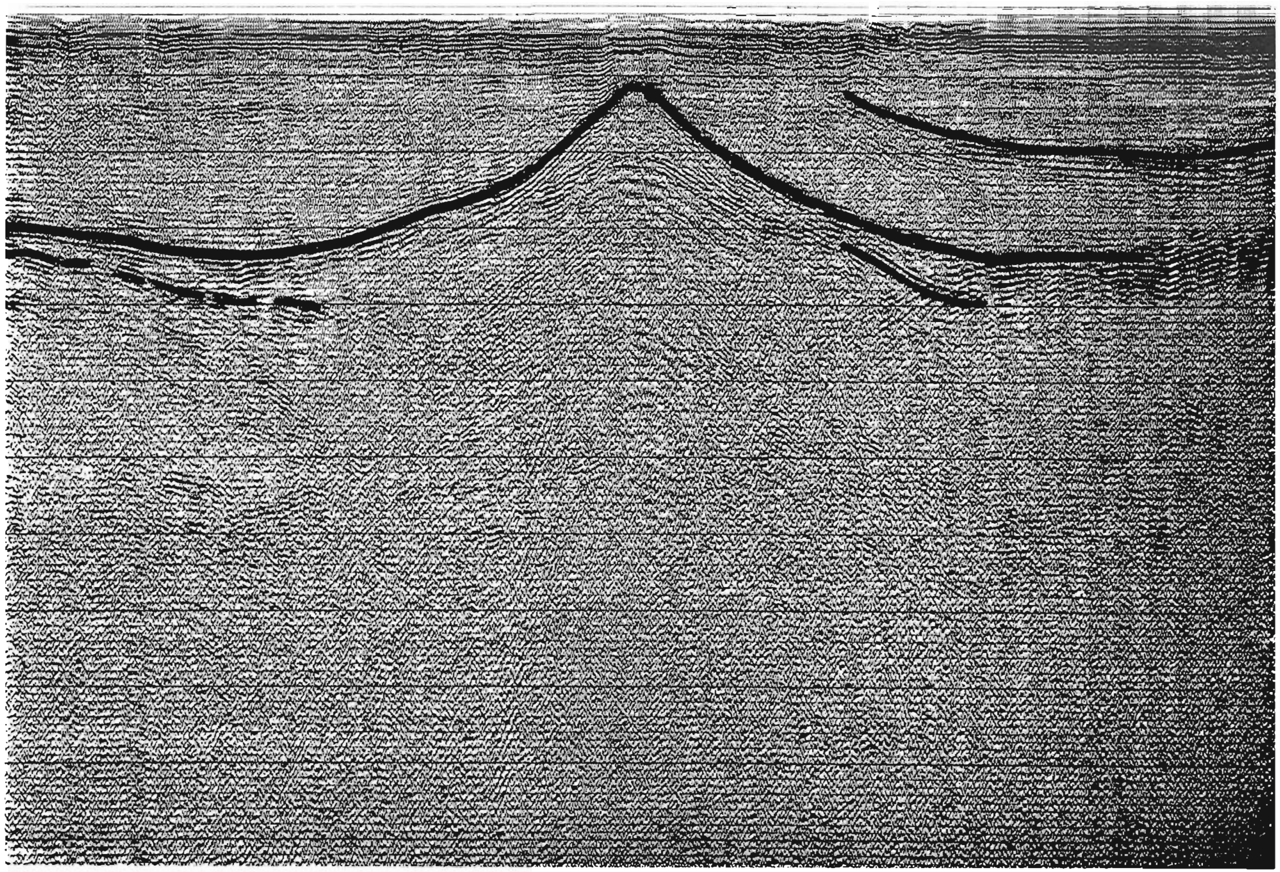

Figure 12. Seismic profile in the Gulf of St. Lawrence showing an anticlinal feature of possible significance in the entrapment of hydrocarbons.

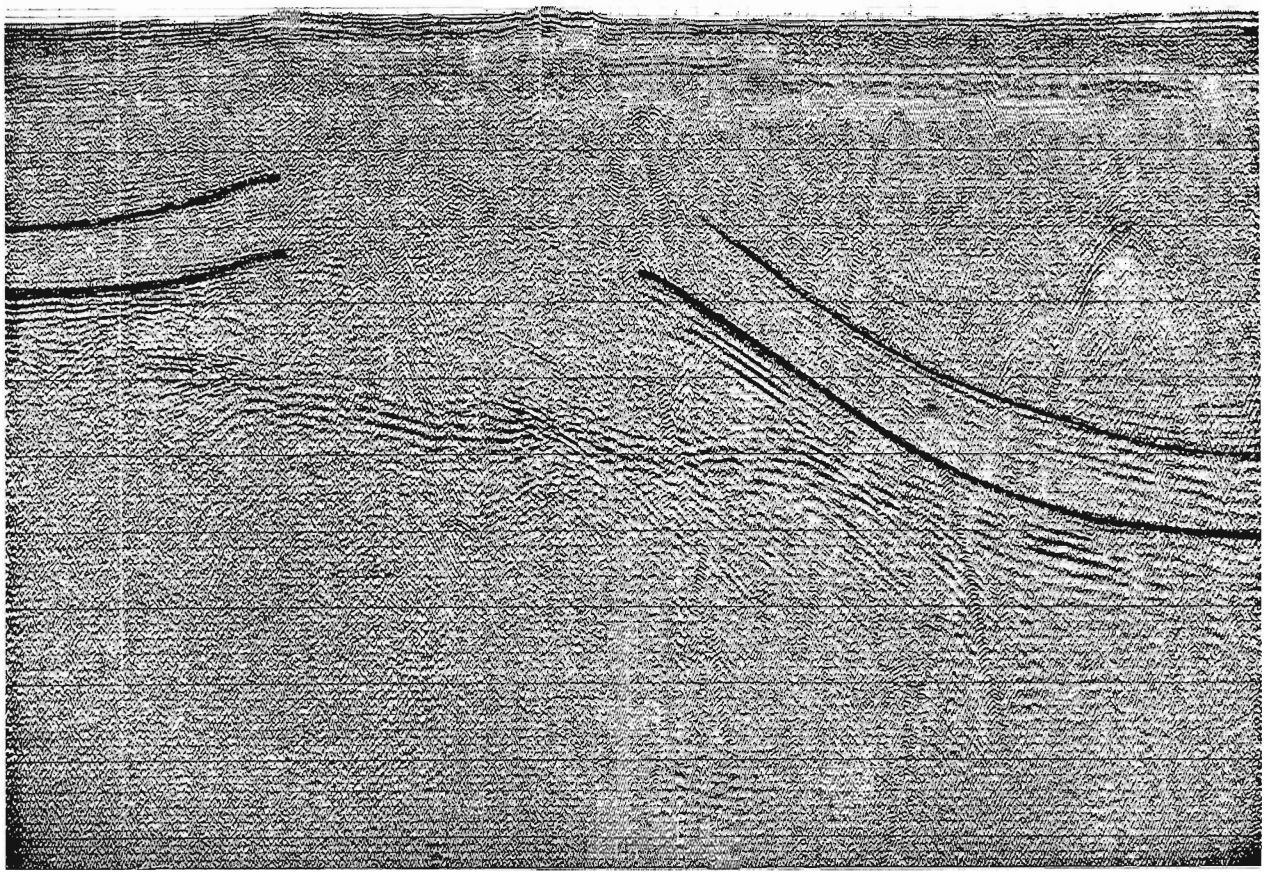

Figure 13. Southeast-northwest seismic profile in the Gulf of St. Lawrence showing a large reef-like structure which has little sea-floor expression.

The section on Figure 8 is located almost due south of Avalon Peninsula on southeast Newfoundland and has a northwest-southeast orientation. It is readily apparent that the sedimentary section has been penetrated by two intrusives, possibly salt diapirs. Water depth is constant at about 480 feet (146 m) over the length of this profile.

Scotian Shelf

Figure 9 is a northwest-southeast oriented profile located near the north end of the Scotian Shelf. Regional thickening toward the edge of the continental shelf is demonstrated. A graben is apparent in the middle of the seismic cross-section which has 100 to 200 milliseconds of throw or approximately 900 feet (274 m).

Figure 10 is also a profile taken on the Scotian Shelf and has a northwest-southeast orientation. The central feature on this profile is a reverse fault of considerable magnitude possibly as much as 2500 feet (762 m) of throw. To the southwest about ten miles, a small normal fault is also apparent which is possibly associated with the larger fault in the section.

Figure 11 shows the continental shelf edge. The lack of continuity of the acoustic events and the irregularities of the water bottom at the edge of the shelf are evident in this section indicating that some slumping has taken place. The water increases in depth from 250 feet (76 m) to 5300 feet (1615 m) across the profile.

Gulf of St. Lawrence

The data on Figure 12 was recorded in the Gulf of St. Lawrence. The characteristics of the prominent anomaly here are unlike those exhibited on earlier sections on the Grand Banks. Present available information in this area is inadequate to identify the physical nature of this anticlinal feature and its significance in the possible entrapment of hydrocarbons.

Figure 13 is taken from another line in the Gulf of St. Lawrence. The profile runs southeast-northwest, and a large reef-like structure is apparent. Note the discontinuity of the shallower markers. The ocean floor reflects to a much lesser degree the structural configuration of the subsurface.

The above cross-sections contain considerably more supporting detail than could be interpreted directly onto the figures. The main objective of this article has been to illustrate some major geological features encountered in various areas of the continental shelf in the Canadian Maritimes, which is a fascinating one in which to work, and one that holds great promise of major offshore hydrocarbon accumulations.

ACKNOWLEDGMENTS

We wish to thank P. J. Hood of the Geological Survey of Canada for help in the preparation of this paper. We also wish to thank the Department of Energy, Mines and Resources for the opportunity of presenting and publishing this paper and hope that it will stimulate interest in the further use of continuous seismic profiling techniques.

(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 241-252)

15. A GRAVITY SURVEY OF THE SCOTIAN SHELF

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Abstract

During the summer of 1970, 692 underwater gravity stations were established on the Atlantic continental shelf of Canada. The areas surveyed include the Laurentian Channel, Cabot Strait, and parts of St. Pierre Bank and the Scotian Shelf. Stations were located on a 15-km grid and detailed profiles run across the Orpheus anomaly and across a diapiric structure northeast of Sable Island. During all three cruises of CNAV SACKVILLE, standard marine Decca and radar were used as primary navigation aids but tests were also made to evaluate the accuracy of differential Omega navigation. The gravity data are presented in the form of a Bouguer anomaly map. The anomalies generally strike in an easterly direction across the continental shelf and the dominant feature is the linear Orpheus anomaly which extends 250 km eastwards from Chedabucto Bay and is flanked to both north and south by positive anomalies. The northern positive anomaly appears to be related to Proterozoic metavolcanic rocks with interspersed Devonian basic intrusions. A broad negative anomaly south of the Orpheus anomaly is probably underlain by a Devonian granite batholith. Between Cape Breton Island and the Miquelon Islands a broad positive anomaly area appears to delineate the underwater extent of the Avalon Platform and in this area density variations within the pre-Carboniferous basement are probably the primary source of variations in the gravity field. Several gravity anomalies are distorted or terminate at one or both margins of the Laurentian Channel; this observation suggests that deep-seated structure may have been a factor in the formation of the Channel.

INTRODUCTION

The regional underwater gravity measurements made on the north-eastern part of the Nova Scotia continental shelf in 1970 by the Gravity Division of the Earth Physics Branch of the Department of Energy, Mines and Resources are part of a program to systematically measure the earth's gravitational field over Canada's lakes, inland seas, continental shelves and margins. These measurements are part of a larger program to provide gravity data over all of Canada for (i) geodetic studies, (ii) studies of structural geology as an aid to the petroleum and mineral exploration industries and (iii) fundamental investigations of the earth's crust and upper mantle. Figure 1 shows the area which was covered by the 1970 underwater gravity surveys. Within the survey area, which stretches about 600 km from Halifax to the Newfoundland coast, there are 692 underwater gravity observations

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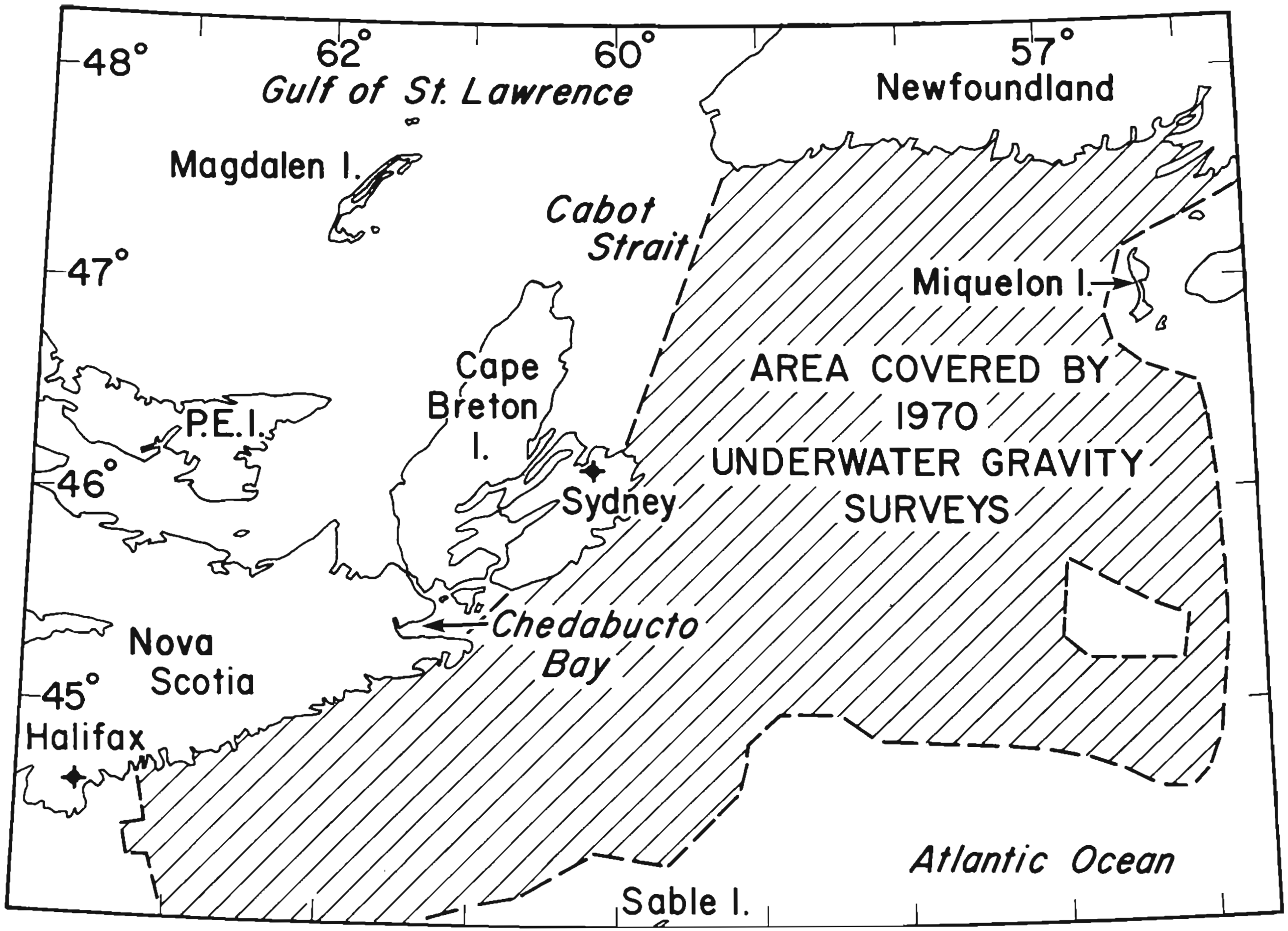

Figure 1. Location map showing area covered by underwater gravity surveys in 1970.

spaced about 15 km apart on a square grid. Marine Decca chains 2, 6 and 7 were used for navigation and the party was at sea a total of 42 days. A small portion of the area was not surveyed due to a combination of bad weather and poor navigation conditions but this gap should be filled up in 1971.

Stephens et al. (1971) have already described the equipment used for the survey and have described the reduction and accuracy of the gravity data, so this information will not be repeated here. This report briefly describes a shipboard evaluation of the Omega navigation system and presents a slightly expanded version of the preliminary interpretation of the Bouguer anomaly field presented by Stephens et al. (1971).

Preliminary Evaluation of Omega VLF Navigation

Using the existing Decca navigation facilities, the Gravity Division will be able to continue to survey the Nova Scotia continental shelf for two more seasons. However, there are extensive areas northeast of Newfoundland and along the Labrador Coast where standard Decca is not available so an important phase of the 1970 work was to evaluate the usefulness of Omega VLF navigation in positioning underwater gravity measurements. The reader is referred to a paper by Kirby (1970) for a description of how the Omega system works. A Tracor 599R Omega receiver was set up at Sydney airport in Nova Scotia to monitor fluctuations in the Omega pattern and the monitor corrections were subsequently applied to the Omega readings obtained on board ship. Omega transmitting stations at Aldra, Norway; Trinidad, W.I.; and Forrestport, N.Y. provided useable signals on a frequency of 13.6 kc/s throughout the tests; no attempt was made to receive signals on 10.2 kc/s although these should also be useful. Transmissions from the Omega station at Hawaii could be detected but they were generally too weak to use.

A comparison of 11 sets of Omega fixes and geographical coordinates obtained from a Magnavox 702 CA Satellite receiver indicates that discrepancies of 1 to 2 km between the Omega and satellite navigation fixes occur under good daytime conditions. These discrepancies consist of a systematic westward displacement of about 0.5 km of the Omega positions relative to the satellite fixes and a random scatter of about 1 km. The systematic shift may occur because the Omega and satellite fixes are referred to different datums. Kirby (1970) reports systematic offsets in his Omega positions but the offset varies in direction and magnitude from day to day. The random scatter is caused mainly by randomness in the Omega data as the error in the satellite data should be only 0.1 km because the satellite data were carefully selected on the basis of previous operational experience at known locations. A comparison of 272 sets of Omega and Decca fixes (Fig. 2) indicates that most of the results agree to within 1 or 2 km although a few disagree by more than 4 km. The larger discrepancies usually occur at night when the Omega and Decca signals are least stable and skywave signals may cause interference. Most of the discrepancy between the Omega and Decca fixes arises in the Omega data as a comparison of 27 sets of Decca and satellite fixes indicates that the random error in the Decca fixes is only about 0.8 km.

A combination of Omega fixes and dead-reckoning, supported by satellite navigation, should give positions that are adequate for reconnaissance and regional underwater gravity surveys in areas along the east coast of Canada where standard Marine Decca is not available but for very detailed underwater work, one cannot rely on Omega navigation, because an error in N-S position of 4 km represents an error of 3 mgal at these latitudes.

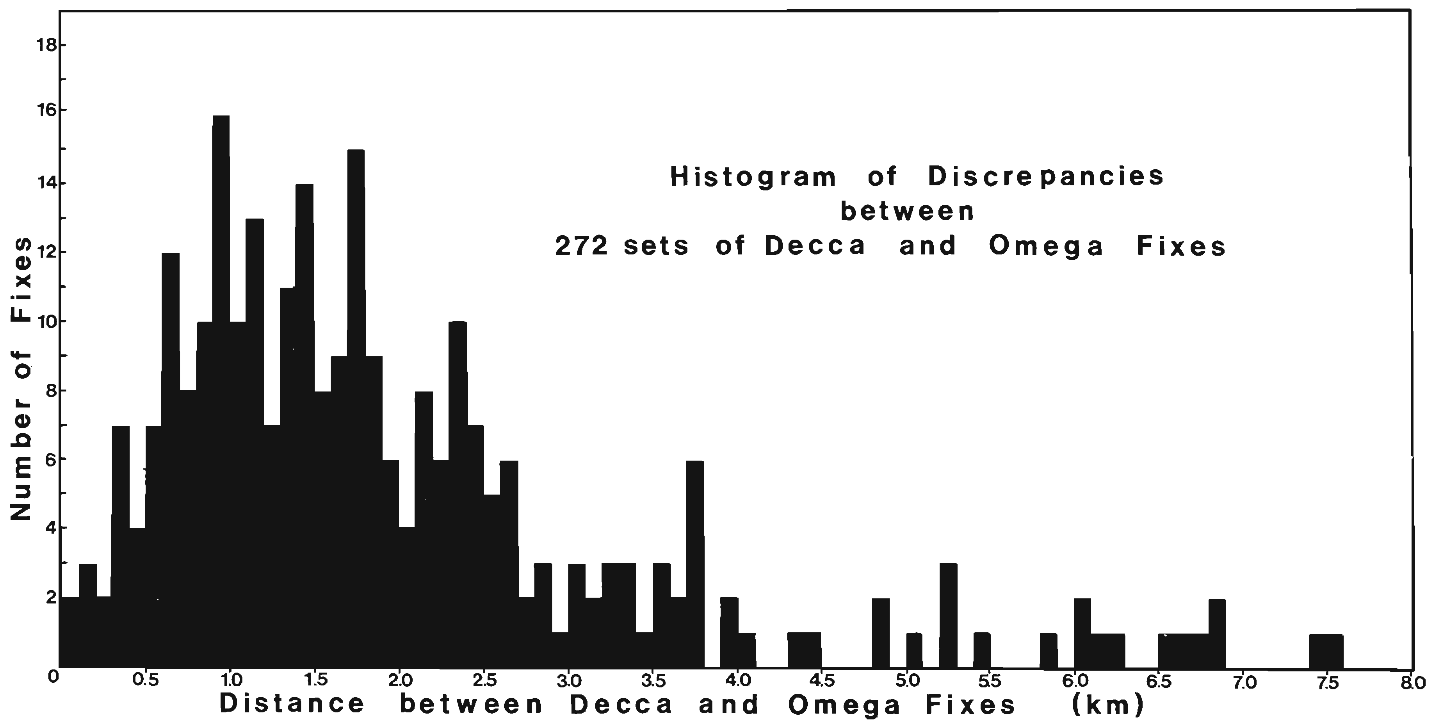

Figure 2. Histogram of distances between Decca and Omega fixes.

Preliminary Interpretation of the Bouguer Anomaly Field

Figure 3 is a simplified version of the 1:1,000,000 Bouguer anomaly map in the report by Stephens et al. (1971). The Bouguer gravity anomalies in Nova Scotia, Newfoundland and the Gulf of St. Lawrence west of Cabot Strait have been described and interpreted by Garland (1964), Weaver (1967) and Goodacre et al. (1969) respectively. In the area east of Cabot Strait and off the Nova Scotia coast, the gravity anomalies generally trend in an easterly to northeasterly direction, although in the northwestern quadrant of the survey area they tend to converge towards Cape Breton Island. The Bouguer anomalies become more positive seawards, as is characteristically found on the continental shelves. In the survey area, the anomalies vary in magnitude from -50 mgal for the Orpheus anomaly to 91 mgal at the continental margin in the southeastern corner of the area (Fig. 3).

The two primary geological influences on the gravity field are density variations within the pre-Carboniferous basement and the density contrast between overlying Carboniferous sedimentary basins and the basement. In some areas the gravity anomalies extend across the shelf from known structures on land and may be used to infer the seaward extension of the onshore geology. Thus, the Middle Bank Low (Figs. 3 and 4) is attributed to an intrusion of Devonian granite as it displays the same broad characteristics and the same trend as the negative gravity anomalies west of Halifax where Devonian granite is the predominant rock type. King and Maclean (1970) have also delineated the northern margin of this intrusion by using acoustic and surface gravity data and their interpretation is in agreement with the present gravity results.

The most impressive gravity feature in the survey area is the linear, east-west trending Orpheus negative anomaly (Fig. 3). This anomaly, which reaches minimum values of -50 mgal and is accentuated by flanking positive anomalies of about 40 mgal (Fig. 4), was discovered and mapped by members of the Bedford Institute using a shipborne gravity meter. Loncarevic and Ewing (1967) suggest that the Orpheus gravity anomaly is caused by low density Carboniferous, and possibly younger, sedimentary rocks lying within a trough in dense pre-Carboniferous basement. Studies, using models which have simple shapes, indicate that a sedimentary basin, which is about 6 km deep and which offers a density contrast of 0.4 g/cm^3 with respect to the surrounding rocks, provides a good fit to the underwater data (Fig. 5). As the flanking rocks probably have a density of about 2.8 g/cm^3, the mean density of the sedimentary basin is 2.4 g/cm^3. Seismic data (Ewing and Hobson, 1966) from another part of the Orpheus gravity anomaly indicate the compressional wave velocity increases with depth in the sedimentary rocks, therefore, there is probably a corresponding increase of density with depth from about 2.2 g/cm^3 near the top to about 2.6 g/cm^3 at the bottom.

On either side of the Orpheus low, the high anomalies appear to reflect the presence of relatively shallow, dense pre-Carboniferous rocks. These pre-Carboniferous basement rocks form buried ridges which are covered in places by less than 1 km of Carboniferous and Cretaceous sediments (Sheridan and Drake, 1968). The northern high anomaly extends west to Cape Breton Island and could be due to the seaward extension of the metavolcanic rocks of the Proterozoic Forchu Group which outcrop on the mainland (Weeks, 1960). Intruding the Forchu Group at Cape Breton are several

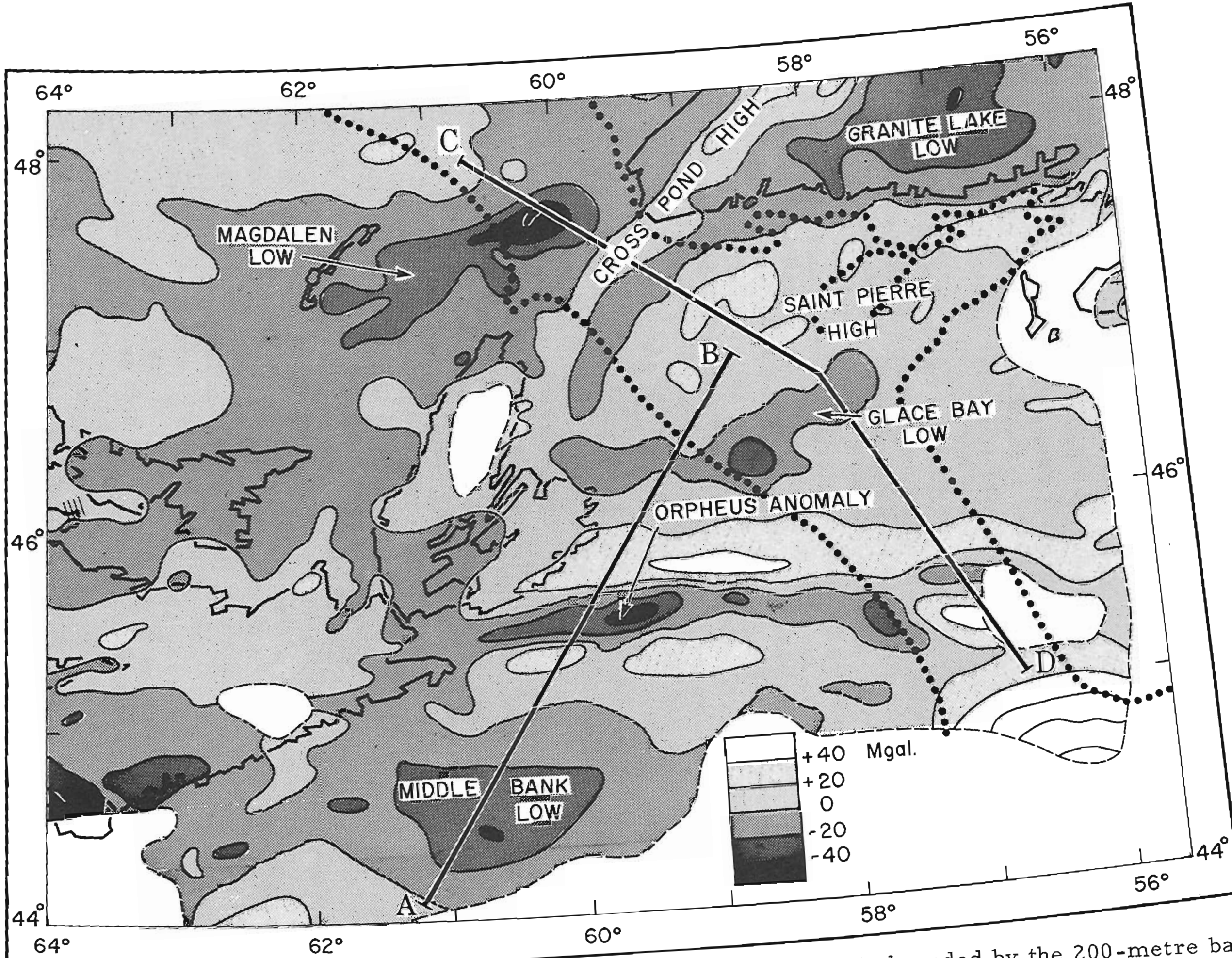

Figure 3. Simplified Bouguer anomaly map. The Laurentian Channel is bounded by the 200-metre bathymetric contours shown as dotted lines.

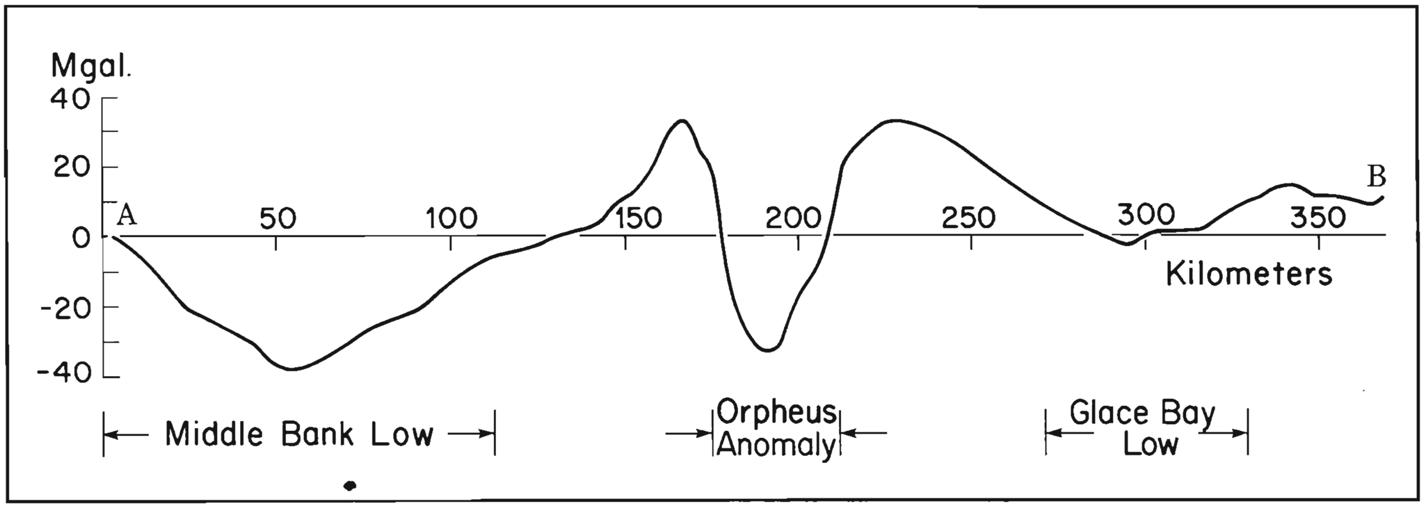

Figure 4. Bouguer anomaly profile A-B across the Middle Bank Low, the Orpheus Anomaly and the Glace Bay Low.

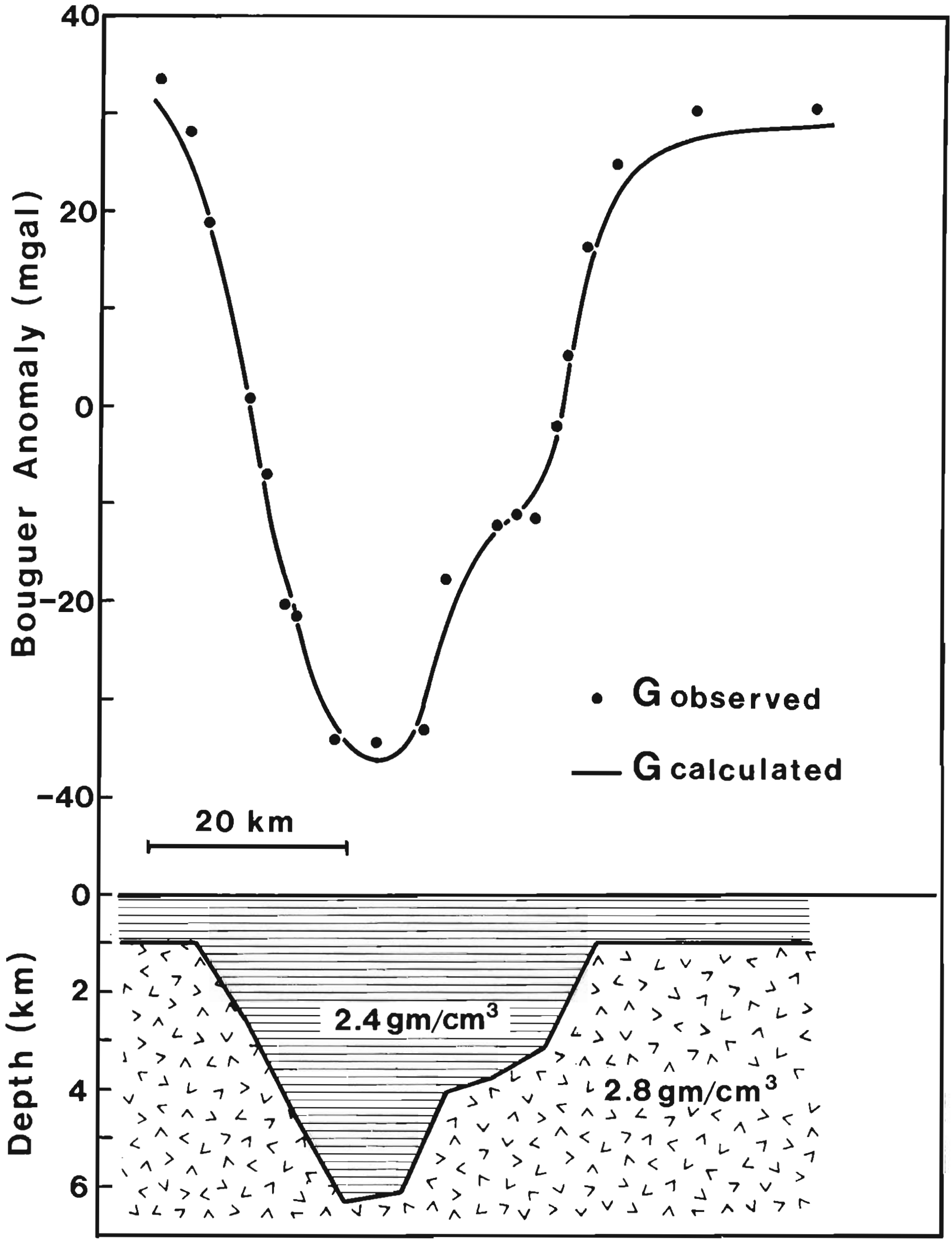

Figure 5. Underwater gravity profile across Orpheus Anomaly, calculated Bouguer anomaly and simple structural model (data from profile AB).

Devonian basin intrusions. Similar rocks may contribute to the high magnetic anomalies (Bower, 1962) and to the high gravity anomalies east of Cape Breton.

The broad positive anomaly region in the northern part of the survey area seems to correlate in a general way with the underwater extension of the Avalon Platform (Poole, 1967); therefore, the St. Pierre gravity High and the adjacent positive region to the southeast probably reflect the presence of Precambrian sedimentary and volcanic rocks similar to those in southeastern Newfoundland.

The Glace Bay gravity Low (Fig. 3) extends east from Glace Bay, N.S., and abruptly swings northeast as it crosses the southwestern edge of the Laurentian Channel. At Glace Bay, Carboniferous sediments are exposed and their gradual thickening to the northeast is suggested by seismic observations (Sheridan and Drake, 1968) which indicate a crescent-shaped (concave southward) sedimentary basin lying between Cape Breton Island and the Miquelon Islands. The configuration of the seismically-determined basin bears little resemblance to the gravity pattern as the deepest part of the basin coincides with the Saint Pierre gravity High. Therefore, throughout much of this area the gravity variations are probably caused primarily by density variations within the pre-Carboniferous basement although variations in the thickness of the sedimentary column will also contribute to the gravity field. As Figure 6 shows, there is no simple relation between the St. Pierre High or the Glace Bay Low and magnetic or seismic data. However, the high compressional wave velocities of 7 km/s or greater at depths of 6 to 10 km apparently reflect heavy basic igneous or metamorphic rocks that should produce a broad gravity high.

The Granite Lake gravity Low and the Cross Pond gravity High (Fig. 3) have been described by Weaver (1967) who correlated them with Devonian granite and Devonian gabbroic intrusions respectively. In addition to the gabbro, high-grade metamorphic rocks may also contribute significantly to the Cross Pond High. Offshore seismic profiles and short wavelength magnetic anomalies (Fig. 6) indicate a pre-Carboniferous ridge which rises to within 0.5 km of the surface and which corresponds with the southwestern extension of the Cross Pond High.

The origin of the Laurentian Channel is an interesting and controversial subject and it seems that the present day form of the trough is caused by glaciation. However, it should be noted that some of the gravity anomalies on the shelf are distorted or terminate at one or both margins of the Laurentian Channel. As Loncarevic and Ewing (1967) point out, the east-west trend of the Orpheus negative anomaly is interrupted (and possibly terminated) at the southwestern margin of the Channel. The Glace Bay Low swings abruptly northeast as it crosses the southwestern margin and it terminates near the northeastern margin. In the Gulf of St. Lawrence, the trend of the Gaspé High is disturbed at the Channel as this positive anomaly sweeps around from Gaspé towards the Newfoundland coast. These observations suggest that deep-seated structure may have been a factor in the development of the Laurentian Channel but this suggestion is only speculative and other geophyscal studies including deep seismic reflection and refraction work should be done not only where the gravity anomaly field seems disturbed by the Laurentian Channel but also in areas where the field is not disturbed as in the vicinity of the Cross Pond High.

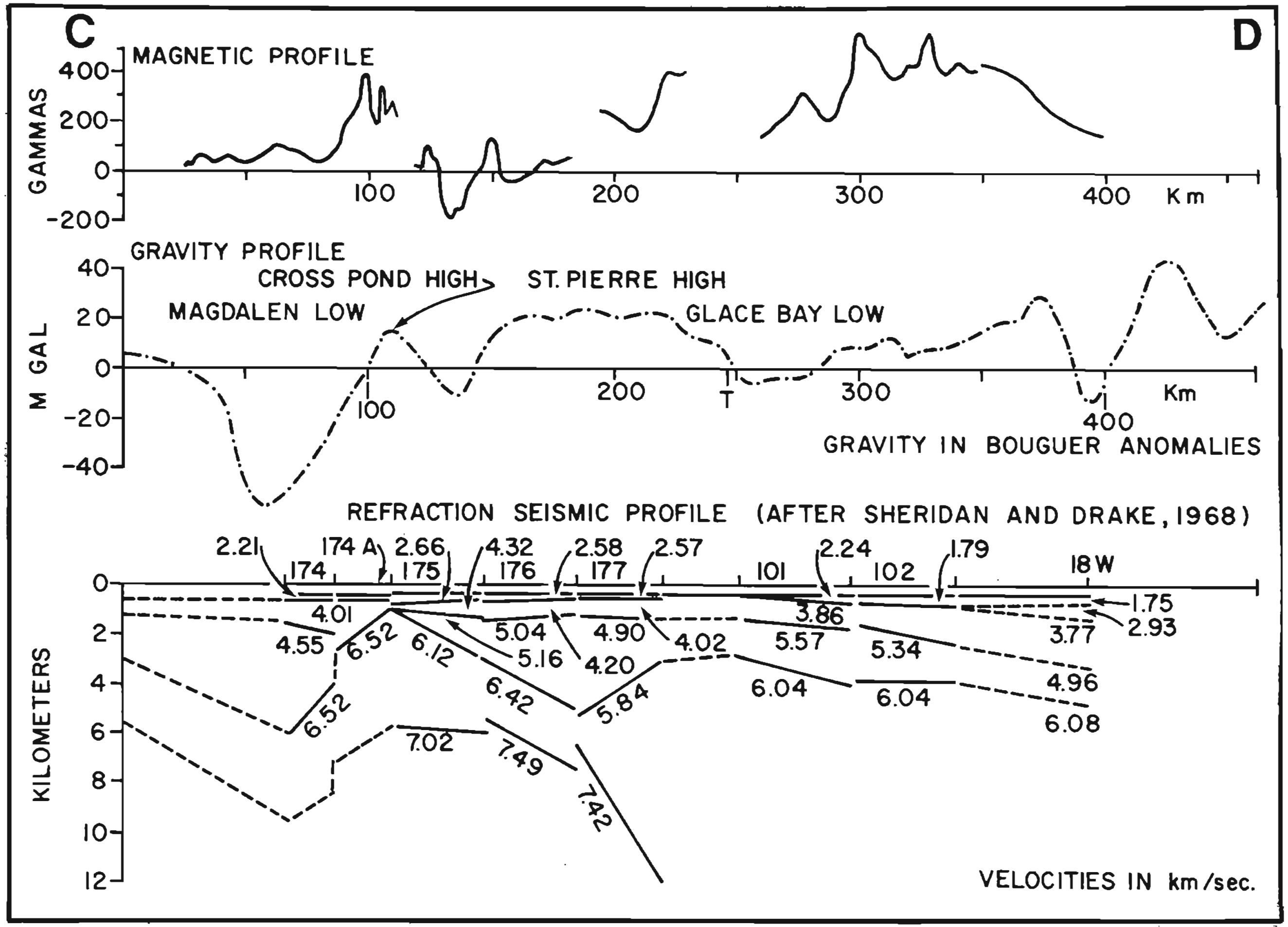

Figure 6. Magnetic, gravity and seismic profiles (C-D) along Laurentian Channel.

At the mouth of the Laurentian Channel, the high gravity anomaly of 91 mgal may be related to the thinning of the sialic crust at the continental margin. However, a relatively high residual magnetic anomaly at this location suggests that the high gravity anomaly might be caused partly by a basic intrusion or partly by dense pre-Carboniferous rocks. Another positive gravity anomaly 100 km southeast of Halifax appears to be restricted to the shelf and it may possibly be produced by high-density basement rocks consisting of either Devonian intermediate or basic intrusions or pre-Carboniferous metamorphic rocks belonging, perhaps, to the Cambro-Ordovician Meguma Group.

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16. THE SABLE ISLAND DEEP TEST OF THE SCOTIAN SHELF

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Abstract

Mobil Oil Sable Island No. 1 (C-67) drilled to a total depth of 15,106 feet on the Scotian Shelf of the Canadian Atlantic offshore, constituted the first deep test in the region. The well drilled 190 miles (306 km) east of Halifax, Nova Scotia on the outer shelf utilizing Sable Island as a drilling platform, penetrated a Tertiary-Cretaceous section.

The well section consists predominantly of clastic strata composed of 4,050 feet of Tertiary and Quaternary, and 11,056 feet of Cretaceous strata of which upwards of 6,000 feet can tentatively be assigned to the Lower Cretaceous. These sequences can be subdivided into a total of 11 units based on gross sandstone abundance, paleontological data, as well as additional lithologic criteria. The succession of these units indicates the presence in this part of the Scotian Shelf of fluctuating, but mainly marine conditions of Cretaceous and Tertiary deposition in very shallow to bathyal water depths.

Encouraging gas shows were tested in several zones as well as a trace of oil on a test near the bottom of the hole. Geochemical and lithologic criteria suggest the former presence of environmental conditions conducive to the generation of oil and gas. Porous sandstones were encountered throughout much of the succession. Unpublished seismic and other data suggest the possibility of structural features in the region related to salt tectonism and contemporaneous faulting. Stratigraphic entrapment conditions for oil and gas may also be present.

INTRODUCTION

Mobil Oil Sable Island No. 1 (C-67) was drilled to a total depth of 15,106 feet on the Scotian Shelf of the Canadian Atlantic offshore. The well was intended primarily as a stratigraphic test of the region.

It is the purpose of this paper to outline the nature of the sedimentary succession encountered; its geologic relationship to the Scotian Shelf and neighbouring areas; as well as the economic significance to this region of offshore Eastern Canada.

The Scotian Shelf as defined by Berger *et al*. (1966), consists of an inner shelf, an area of depressions and isolated banks in the central part of the shelf, and an outer banks region. Most of the unconsolidated sediment cover on Sable Island and other banks consists of sand and gravel whereas the depressions contain loosely compacted clay and silt (King, 1970). The sand and gravel cover of the banks is relict Pleistocene glacial outwash (James and Stanley, 1968). Considerable reworking of these deposits by strong bottom currents on Sable and other outer banks has continued to the present time (Stanley and Cok, 1968).

The well was drilled on Sable Island which marks the only emergence on the outer shelf off of Nova Scotia. The island is a narrow, arcuate, dune-covered sand bar about 21 miles (34 km) long, that is known as the "Graveyard of the Atlantic" as a result of upwards of 500 vessels (Grosvenor, 1965) having been wrecked about the treacherous, shifting shoals. In the present instance, the island is of particular interest in that it provided a drilling platform for the first exploratory test of strata on the Scotian Shelf. Sable Island is situated approximately 190 miles (306 km) east of Halifax. The cyclical movement of sediment by current action around this island may be one of the leading factors owing to its preservation in this region of high seas and winds (James and Stanley, 1968). The well was located near the centre of the island at Lat. 43°56'05"N, Long. 59°51'01"W. Figure 1 highlights the location of the well with particular reference to the Scotian Shelf. Information on wells drilled subsequently in this area are still held as confidential.

The well was spudded on June 7, 1967 and abandoned on January 2, 1968. The logistics and actual drilling operations of this remote exploratory well have been described in some detail by Brusset (1969).

CRETACEOUS AND TERTIARY REGIONAL GEOLOGY

Isolated pockets of stratified clay, silt, sand and lignite containing a Lower Cretaceous microflora have been found on mainland Nova Scotia about 32 miles (51 km) northeast of Halifax (Stevenson, 1969). Cross-stratification and pebble imbrication as reported in Swift and Lyall (1968, p. 333) indicate a westerly drainage direction. These are the only known occurrences of Cretaceous surface exposures in the Atlantic Provinces (King et al., 1970, p. 145).

Speculations dating back many years suggested that Cretaceous and Tertiary strata as exposed on the Atlantic Coastal Plain of the United States extended northeastward beneath the continental shelves south of Nova Scotia and Newfoundland. These early speculations were founded on rock fragments containing Tertiary or Cretaceous fossils obtained from fish trawls at scattered localities on the shelves.

Seismic studies on the Scotian Shelf and adjacent areas provided additional bases. Officer and Ewing (1954) reported on a series of refraction profiles which included the Browns Bank section and the well-known Halifax section. Their profiles indicate the presence of three velocity layers of sediments. From accumulated data on both the submerged and emerged Atlantic Coastal Plain of the United States they identified: an unconsolidated layer of Upper Cretaceous, Tertiary and Quaternary age; a semi-consolidated layer primarily Lower Cretaceous and possibly Jurassic; and a consolidated sediment layer interpreted as Triassic. Drake et al. (1959) concluded from a compilation of a great mass of geophysical data that two sedimentary troughs are present and filled with Mesozoic and Tertiary sediments. These two troughs occur under the shelf and continental slope and are separated by a basement ridge. This young sedimentary system was extended north of Cape Hatteras to the Scotian Shelf and adjacent areas. It was compared, furthermore, to the Appalachian system as restored for Early Paleozoic time (p. 176). The basement ridge complex is indicated by Emery et al. (1970) to have served as dams to trap land-derived sediments during the Mesozoic Era.

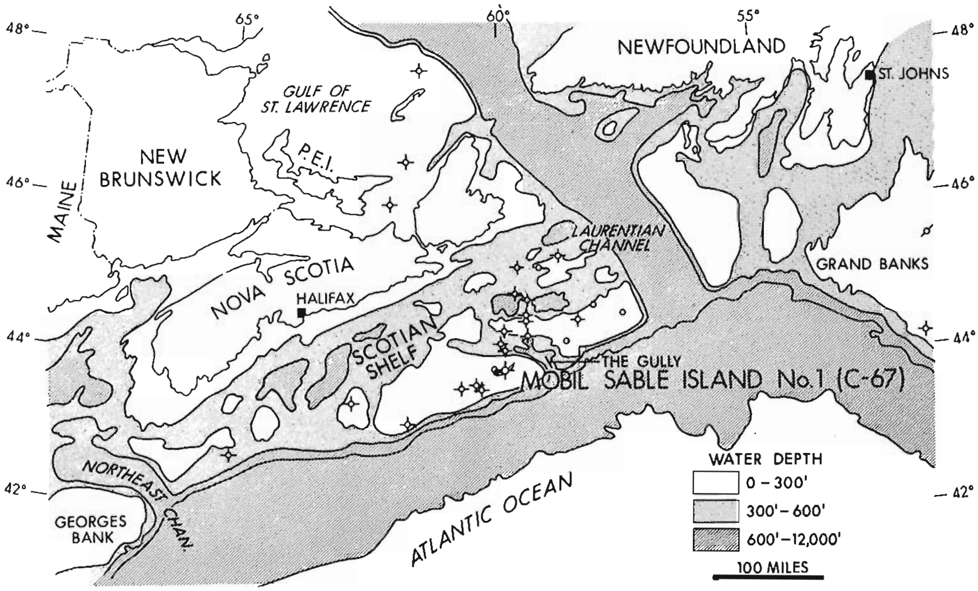

Figure 1. Scotian Shelf and adjacent areas with location of Mobil Oil Sable Island No. 1 (C-67).

Seismic refraction profiles on and in the vicinity of Sable Island by Berger et al. (1965, p. 959) indicated a total thickness of 14,750 feet for all sediments comprised of 3,250 feet with a velocity of 5,900 ft/sec., 3,950 feet with a velocity of 10,000 ft/sec. and 7,550 feet with a compressional velocity of 12,500 ft/sec.

Morphologic studies by Berger et al. (1966) suggest the presence of a submerged fall line in water depths of 510 feet (186 m) adjacent to the southeastern coast of Nova Scotia. Studies primarily concerned with unconsolidated sediment based in part on echograms and reflection seismic on the Scotian Shelf south of Halifax by King (1967, 1970) indicate underlying strong reflecting horizons interpreted as Cretaceous or Tertiary bedrock dipping southward at less than one degree. These beds appear to be truncated in the updip direction about 28 miles (45 km) from shore.

Cretaceous (Cenomanian) submarine bedrock samples were dredged and described by King et al. (1970) from a locality 62 miles (100 km) north-northwest of Sable Island. The locality was also observed and photographed from a submersible (King and MacLean, 1970a). The integration of these data with reflection seismic results conducted on the Scotian Shelf and Laurentian Channel areas has enabled King and MacLean (1970c, 1970d) to map an angular unconformity approximately between the Cretaceous and Tertiary; hence, differentiate the two, over broad areas of the Scotian Shelf. They interpret the Tertiary as essentially flat over regularly seaward-dipping Cretaceous, or flat over folded Cretaceous as in the Orpheus Anomaly area.

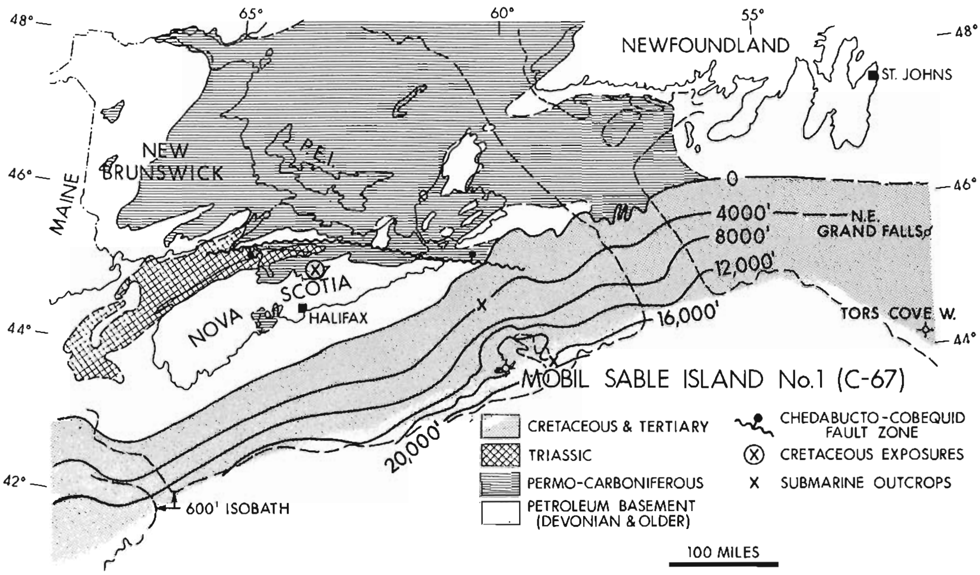

Figure 2. Regional geology of Scotian Shelf and adjacent areas showing generalized Cretaceous-Tertiary thickness.

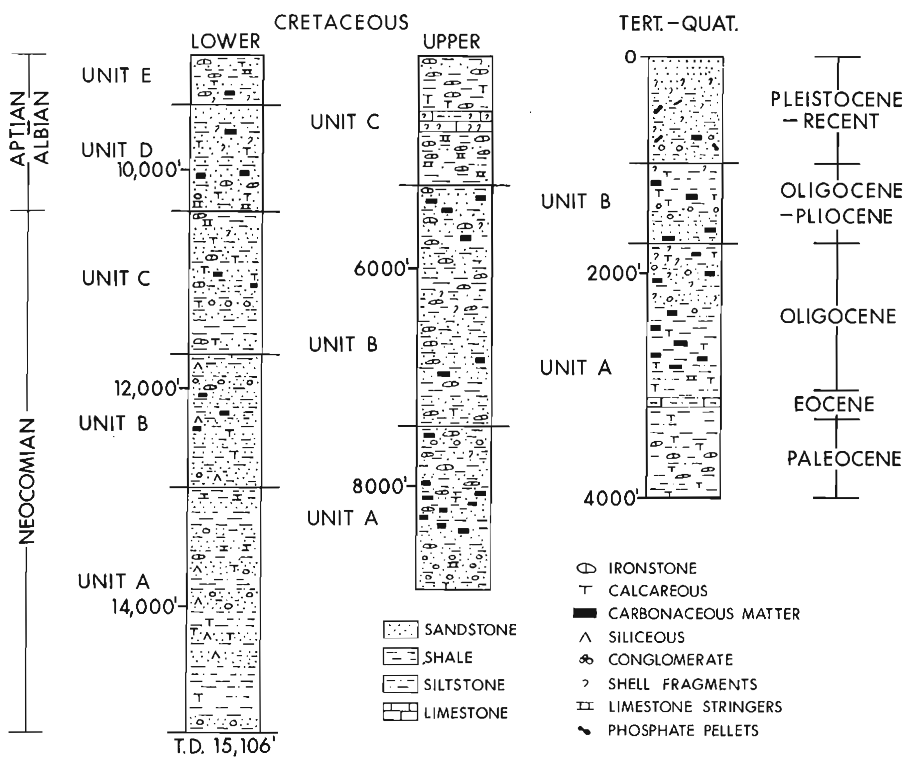

Figure 3. Generalized lithologic section of Mobil Oil Sable Island No. 1 (C-67).

The northernmost known submarine exposures of Tertiary rocks along the eastern coast of North America was the section exposed on Georges Bank (Marlowe, 1969, p. 1091), prior to the Bedford Institute's project in The Gully. Here dredging indicated as much as 4,593 feet (1,400 m) of Oligocene and Miocene (Tertiary) beds crop out in the walls of The Gully.

On the Grand Banks east of the Scotian Shelf the presence of Cretaceous and Tertiary strata is documented by Pan Am IOE Al W Tors Cove D-52 (Lat. 44°11'14"N, Long. 52°23'42"W) and by Pan Am IOE Al NE Grand Falls H-09 (Lat. 45°28'19"N, and Long. 52°00'03"W). The stratigraphy of these wells has been considered by Bartlett (1969), Howie (1970) and more recently by Bartlett and Smith (1971).

Mobil Oil Sable Island No. 1 drilled an entirely sedimentary succession to 15,106 feet. Shale and sandstone comprise the bulk of the well section. Preliminary study of the fossil content indicates this well section can be subdivided into 4,050 feet of Tertiary and Quaternary, 4,900 feet of Upper Cretaceous and 6,156 feet of Lower Cretaceous strata. One of the most significant findings provided by the well was documentation for upwards of 11,000 feet of Cretaceous strata in this portion of the Scotian Shelf.

On the basis of the Sable well and other data, Figure 2 contains generalized isopachs of the total Cretaceous-Tertiary sequence on the Scotian Shelf. This interpretation suggests the presence of a thick basin infilled with these beds in the general vicinity of Sable Island.

STRATIGRAPHY

No formal stratigraphic breakdown of the thick sandstone-shale sequence in the Sable Island well has been attempted; moreover, available paleontological data is preliminary. Instead, the well section has been divided into informal units based on gross sandstone or shale abundance and other lithological attributes. In effect, the succession was broken into groups of sandstone units with the age as an additional factor in the subdivision. The lithology of these 11 units is summarily illustrated on Figure 3.

Lower Cretaceous Section

Unit A (15,106 ft. - 12,905 ft.)

This unit consists in overall aspect of about 75 per cent shale and 25 per cent sandstone. The shales are dark grey-brown to brown in colour, generally silty. Pyrite is a common accessory mineral. The sandstones in this unit are characterized by the coarseness of the angular to subangular grains, poor sorting and relatively high argillaceous content. The sandstones are siliceous in part. Black bituminous matter and carbonaceous flecks are common in some zones. Unit A, in the lower part of the well, is interpreted to be a transgressive neritic shale and sandstone deposit probably representing the maximum Lower Cretaceous transgression on this part of the Scotian Shelf.

Unit B (12,905 ft. - 11,685 ft.)

This unit contains 56 per cent shale and 44 per cent sandstone in gross aspect. The shales are generally varying shades of grey, and, although silty in part, are much purer than in Unit A. Thin maroon shale interbeds occur scattered in the unit. Pyrite and black carbonaceous specks are fairly common with thin beds and partings of bituminous shale.

The sandstone beds of unit B are coarse to medium grained. They are generally less shaly than the previous unit, particularly in the thicker zones. Some beds are locally quite calcareous and a few are siliceous.

Unit B contains neritic microfossils. The relatively high sandstone percentage with the tendency towards more mature grain rounding and sorting, along with interbeds of purer shales suggests moderate sedimentation rates in the inner neritic to paralic realm.

Unit C (11,685 ft. - 10,377 ft.)

Shale and siltstone account for 71 per cent of this unit and sandstone 29 per cent. It contains the marine fossils of undifferentiated neritic water depths. The shales are generally some shade of grey. A red to maroon zone is present near the top of the unit.

The sandstone beds range generally from very fine to medium grained. The grains are moderately to well-sorted although poorly sorted toward the base. The argillaceous content is variable but quite constant within individual sandstone beds. The overall sandstone percentage and vertical distribution of individual sandy zones would suggest that unit C represents an inner neritic succession of offshore bar complexes.

Unit D (10,377 ft. - 9,410 ft.)

The unit contains neritic fossils in a succession dominated by sandstones with several discrete beds exceeding 80 feet in thickness. In overall aspect, the unit contains 58 per cent sandstone. The sandstones contain generally coarse to medium subangular to subrounded grains. Most of these beds have at least 90 per cent sand-size clastic particles. Some of the sandstones are calcareous. The interbedded shales of this unit are mainly grey with some red and maroon interbeds.

Unit D appears to represent a succession of offlapping beach and associated marginal marine deposits. The high sandstone percentage would support this concept. The comparatively high grain-to-matrix ratio of the sandstones is interpreted as indicative of rather slow deposition in a stable high energy environment.

Unit E (9,410 ft. - 8,950 ft.)

Unit E is dominated by fine clastics, that is, shale and siltstone which make up 92 per cent of the interval. The shales are predominantly greyish brown, although maroon and black shale zones are present toward the base. Unit E has various aspects to support the concept that it may in part represent a tidal flat deposit. A diamond-core cut between 9,280 feet and 9,309 feet indicates the presence of lenticular lamination, channeling, and intraformational shale-pebble conglomerate.

Upper Cretaceous Section

Unit A (8, 950 ft. - 7, 468 ft.)

This Upper Cretaceous interval is composed of 53 per cent sandstone and 47 per cent shale. Neritic zone fossils have been obtained from the unit. Coarse sandstones dominate particularly toward the top and base of the unit. Black carbonaceous material occurs occasionally scattered throughout the succession but is especially abundant in the interval 7, 960 feet to 8, 430 feet. Shales in this unit are generally silty, dark greyish brown to black.

The relatively high percentage of coarse sandstone with abundant fossils and glauconite indicates a shallow water marine origin for Unit A probably at inner neritic to paralic water depths of deposition. The carbonaceous zone could indicate proximity to lagoons or some other source of vegetal material.

Unit B (7, 468 ft. - 5, 250 ft.)

Shales comprise 63 per cent of this unit which contains undifferentiated neritic microfossils. Dark shades of greyish green to greyish brown silty and sandy shales are most common.

The brownish sandstone zones in this unit commonly occur as 25-foot beds or much less in thickness. Grain-sorting is moderate to rather poor with variable amounts of shaly matrix. Carbonaceous flecks and pyritized coal fragments are common in the upper 500 feet.

Upper Cretaceous Unit B is indicated to have been deposited rather rapidly in the inner neritic environment. This is supported by the presence of finer, yet poorly sorted sands, with abundant interbedded shales all deposited under reducing conditions of deposition.

Unit C (5, 250 ft. - 4, 050 ft.)

This 1, 200 foot unit completes the Upper Cretaceous succession in the Sable Island No. 1 well. It is composed of 75 per cent shale and about 4 per cent sandstone. The remainder of the unit consists of 200 feet of impure carbonate. This is the most significant thickness of carbonate in the entire well section. The unit contains a middle to outer neritic microfauna.

The shales of this unit are dark grey, greenish grey with minor amounts of brownish grey and black shale. Interbedded stringers of limestone occur between 4, 670 feet and 5, 250 feet. The limestone unit between 4, 470 feet and 4, 670 feet is white to cream with about 40 per cent chalky argillaceous matrix binding calcareous prisms of disaggregated pelecypod fragments. The sandstones in this unit are confined to three thin zones.

Upper Cretaceous Unit C is a middle to outer neritic succession. The carbonate zone has the appearance of a shell hash which was deposited either by deep outer shelf currents or the accumulation may indicate a slight regressive pulse with the buildup of an outer shelf bar of partially disintegrated shell fragments.

A notable change in structural attitude of the strata is evident on the computed dipmeter near the Cretaceous-Tertiary boundary. The upper Cretaceous beds dip easterly at from 2 degrees to 4 degrees whereas the

Tertiary beds have a consistent southerly dip of the same order. It may indicate subsidence of the Cretaceous or early Tertiary outer shelf at this part of the continental margin followed eventually by southerly directed Tertiary progradational buildup over the slope which resulted from the subsidence.

Tertiary Section

Unit A (4, 050 ft. - 1, 725 ft.)

The lower Tertiary unit is composed of 83 per cent shales and 14 per cent sandstone. The remainder of the unit comprises a 70-foot limestone. A prolific microfauna indicates the presence of Paleocene, Eocene and Oligocene in the interval. Outer neritic to bathyal depths of moderately warm waters of deposition are indicated.

Various shades of brown dominate the shales of this deep-water succession. The shales are sandy and silty throughout. The limestone is a dark brown, cryptocrystalline, argillaceous carbonate.

The sandstones occur as thick massive beds with gradational contacts within the upper 600 feet of the unit. They consist of poorly consolidated, coarse, conglomeratic, poorly sorted grains. The grain-to-matrix ratios are high indicating fairly clean sandstones.

The upper sandstones of this unit may be turbidite deposits from a nearby high-energy shelf where the initial winnowing of the fines may have occurred before their influx onto the deep water slope which was gradually up-building.

Unit B (1, 725 ft. - 980 ft.)

Unit B contains 58 per cent sandstone. Scarse microfaunal and microfloral data indicate a mainly Miocene to Pliocene age. The poorly consolidated sandstones are mostly brownish grey and coarse grained. Lignite and carbonaceous flecks are present throughout much of this unit. A sandy coquina zone composed of pelecypod and gastropod fragments occurs near the top of the unit. Brown, silty, partly calcareous clay comprises the fine clastic portion of the unit.

The common presence of lignitic material and the relative absence of microfauna suggest that much of this unit may be nonmarine in origin. The ill-sorted, dirty nature of the abundant sands supports this possibility.

Unit C is considered to be a regressive paralic unit which was deposited on a low alluvial coastal plain close to a source of vegetal material. The coquina zone with an associated littoral to inner neritic fauna near the top of the unit may represent a brief marine invasion.

Quaternary Section

The Pleistocene to Recent succession of unconsolidated clastics is 980 feet thick of which about 63 per cent is sand. It consists of a lower division dominated by brown, sandy clay grading to clayey sand. The upper part of this unit is composed of unconsolidated poorly to moderately sorted subrounded quartz sand. In the top 300 feet all of the silt and clay has been winnowed out leaving a grain framework. The pure sands of the upper part of this section are relict glacial material which has been reworked by marine agents during the Holocene rise in sea level.

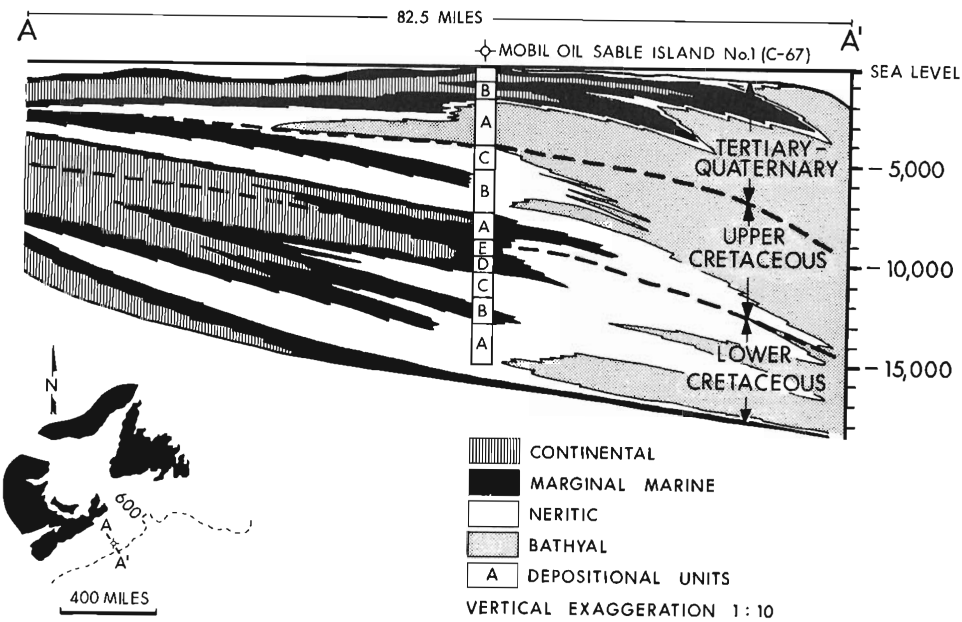

Figure 4. Diagrammatic cross section through Mobil Oil Sable Island No. 1 (C-67) illustrating major depositional environments.

DISCUSSION

Figure 4 vertically stacks the sequence of environments deduced for the eleven units just discussed. The diagrammatic section runs essentially at right angles to the trend of the Scotian Shelf and alternatively, at right angles to depositional strike. The vertical compilation of these gross environments provides an insight into major transgressions and regressions hence provides an insight into the total depositional sequence (Visher, 1965).

This diagram (Fig. 4) indicates a number of major transgressions and regressions across this part of the Scotian Shelf during Cretaceous and Tertiary times. The fundamental significance to the region of these shifting strand lines is the implied tectonic instability at the continental margin in this area. A rather high degree of shelf-flexing must have occurred in this region because of the comparatively thick deposits originating in subparalic depths of deposition as interpreted for much of the Cretaceous and Tertiary strata. Deitz and Holden (1966) emphasize prograded paralic wedges in their model of the Atlantic miogeocline.

Petrographic examination of sidewall cores and other data indicates that the majority of the sandstones can be classified as subfeldspathic. Locally sublithic sandstones are present particularly below 11,000 feet.

Studies of rock fragments and grain characteristics as well as heavy mineral suites indicate a varied provenance. Volcanic and metamorphic rock fragments such as phyllite are common below 12,000 feet. Large, rounded quartz grains with abraded overgrowths point to a sedimentary source as well. This source is dominant particularly near 10,000 feet. Other indicators include rounded stable heavy minerals as well as sedimentary rock fragments. In addition, reworked Permo-Pennsylvanian pollen has been found near 9,300 feet (J.F. Stone, personal communication). Sheared composite quartz grains, microcline, sodic plagioclase and orthoclase are more abundant higher in the well section indicating the unroofing of a sheared granite or gneissic source. Schistose metamorphic rock fragments are also commonly present in the upper part of the well section.

The data suggest a terrane composed of locally exposed Meguma-like basement, with broad areas of Permo-Carboniferous strata with the gradual unroofing of the Meguma basement complex with time. An additional source is provided by more locally derived Triassic sedimentary and volcanic rocks.

Dipmeter information indicates a south to southeastward direction of transport for the most part. Hence, the great bulk of Cretaceous and Tertiary clastic strata on this part of the Scotian Shelf were derived from a distributive province situated in the same general area as mainland Nova Scotia.

A much thicker Cretaceous sequence is evident in the Sable Island No. 1 well as opposed to the Grand Banks wells such as the Grand Falls H-09 well which was drilled relatively shelfward. It is difficult to compare with the Tors Cove West D-52 well since it apparently drilled into a salt piercement.

The bathyal to supra-littoral Oligocene-Miocene succession in the Sable Island well section correlates toward the shelf edge with thick outcrops in The Gully of similar-age fine clastic strata (Marlowe, 1969). Finally, the thick sedimentary succession penetrated by the Sable Island well tends to substantiate earlier interpretations of a thick sedimentary succession based on seismic refraction such as reported by Officer and Ewing (1954) and others; but, contrasts markedly with the apparent local veneer of Lower Cretaceous continental beds locally preserved on mainland Nova Scotia.

ECONOMIC ASPECTS

There were 13 occurrences of gas reported during the drilling and testing of the Sable Island well. The most positive and numerous indications of gas were obtained in the Lower Cretaceous Section. A recovery of 0.4 cubic feet of formation gas was made on a wireline test at 13,465 feet. A conventional bottom-hole test of the lowest 514 feet in the well reverse circulated gas to surface. In addition, the gas detector indicated the presence of gas in nine zones below 12,500 feet through analysis of the drilling mud. Two gas shows were detected from beds of Upper Cretaceous age.

A bottom hole sampler contained 50 cc of 39° API crude oil from a test bottoming at 15,106 feet. The composition of the oil was found to be consistent with an origin in the deeper sediments encountered in the well.

Geochemical source rock analyses were conducted on many sidewall and drill-cuttings samples. The work yielded evidence of fairly high total organic content in the samples as well as other aspects suggesting a close approach to the characteristics of petroleum source rocks.

Various sedimentational and other inferrences can be drawn to suggest favourable petroleum source rocks in the area of the well. These include: the common development of zones containing finely divided carbonaceous specks of vegetal origin; the presence of warm water faunas which might be conducive to high organic productivity; as well as the position of the well section adjacent to the continental margin yet often close to ancient shore lines (Hedberg, 1964). The organic material for source rocks was largely preserved by a reducing environment of which there are a number of indicators such as the common occurrence of pyrite and siderite. There is some evidence to infer restriction under marine conditions by both stratigraphic and structural barriers. For example, bar and other deposits have been interpreted in the well.

The well stratigraphy indicates that sandstones provide the bulk of the effective Cretaceous and Tertiary reservoir beds in this portion of the Scotian Shelf. About one third of the Cretaceous strata are comprised of sandstone that averages 12 per cent porosity in the Lower Cretaceous, and about 22 per cent in the rest of the well section. Such porosities may be prevalent over a broad area of the Scotian Shelf. The development of locally improved reservoir facies is possible in response to the growth of contemporaneous structures of various kinds.

Continuous seismic reflection and other data indicate several types of structures on the Scotian Shelf which could trap large quantities of oil and gas. A dome-shaped structure which may contain an evaporitic core has been reported by King and MacLean (1970b) to occur 28 miles (45 km) north-northwest of Sable Island. A possible salt diapir has also been reported in Laurentian Channel by Keen (1970). Structures resulting from salt tectonism besides domes include possible inter-diapiric residual highs and deep seated salt core anticlines. Salt domes offer areas of interest on which to focus exploration because of their size and great number of possible entrapment situations. Halbouty (1967) has indicated nine major types of traps associated with salt domes in the Gulf Coast of the United States.

Other entrapment possibilities on the Scotian Shelf include a wide range of structures related to faulting including those which are a result of contemporaneous faulting. Finally, the Sable Island well section itself suggests the possibility of various dominantly stratigraphically controlled entrapment situations.

CONCLUSIONS

The section penetrated by the Sable Island No. 1 (C-67) well further documents the extension of the submerged Atlantic Coastal Plain on the Scotian Shelf; with thick deposits of Cretaceous, as well as Tertiary, sandstone and shale. Analyses of the well section reveal fluctuating largely marine conditions of clastic deposition suggesting the instability of this part of the continental margin. Finally, the well section offers encouragement in the search for economic accumulations of oil and gas on the Scotian Shelf.

ACKNOWLEDGMENTS

Sincere thanks are extended to Mobil Oil Canada, Ltd. for permission to publish this paper. The writer acknowledges the assistance of many colleagues at Mobil. In particular, W.H. Monro prepared the original lithologic data basic to this paper, and J.E. Cooper contributed a number of ideas especially relative to Upper Cretaceous and Tertiary stratigraphy. Research scientists at the Field Research Laboratory of Mobil Oil Corporation in Dallas supplied a considerable amount of data relating to the age and environmental significance of microfaunas and floras, lithologic analyses, source rock and other analyses. C.E. Carlson, W.E. Dawson and W.H. Monro of Mobil Oil Canada, Ltd., Calgary, critically read the manuscript and offered numerous helpful suggestions. M.G. Wanner of Mobil assisted in and supervised the drafting of the illustrations.

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(Earth Science Symposium on Offshore Eastern Canada
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17. LATE MESOZOIC AND CENOZOIC OF THE SABLE ISLAND BANK AND GRAND BANKS

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Abstract

The Amoco wells in the Grand Banks and the Mobil well below Sable Island in the eastern Scotian Shelf gave the first continuous stratigraphic sections for the shelf region of offshore Eastern Canada.

Integrated analysis and stratigraphic synthesis of lithic and faunal data showed a minimum of seven unconformity-bounded sequences present. These are Pleistocene, Middle and Late Miocene, Intra-Eocene, Paleocene and earliest Eocene, Late Cretaceous, Middle Cretaceous, and Neocomian in age. Additional units of Middle and Late Jurassic age and probable Late Triassic (Keuper) age are also present.

The rocks include halite and dolomite of probable Late Triassic age, limestones of the Upper Jurassic, Middle Cretaceous, lower Upper Cretaceous, mid-Eocene and mid-Miocene, and sandstones which dominate the Neocomian, Upper Eocene and Middle Miocene successions. Variable proportions of siltstones, silty mudstones and claystones occur almost throughout the section.

Depositional environments ranged from aeolian, for quartz arenites, through very low land, for deltaic stream and swamp deposits, to lagoonal, bank and open-shelf warm-marine environments, for fine sand to clay-size terrigenous sediments or skeletal carbonates and lime muds.

Periodic interregional tectonic oscillations produced the alternately depositional and erosional episodes of the major baselevel transit cycles. The result is the present sedimentary wedge, thickening due to differential preservation toward the continent's edge. This wedge represents one-half or less of all late Mesozoic and Cenozoic time.

Several diapirs, presumably all of evaporites, occur in the region. One penetrated into the Tors Cove well section, another affected the Grand Falls well section, a third caused faulting of the Sable Island section.

A sedimentologic and related tectonic analysis of Orphan Knoll shows that this fragment began to separate from North America in the earliest Middle Cretaceous, but did not sink until the early Tertiary.

INTRODUCTION

The Pan American (now Amoco) Grand Falls and Tors Cove wells in 1966 were the first to penetrate stratigraphic sections of Mesozoic and Cenozoic age in the shelf regions of offshore eastern Canada. The next year Mobil Oil drilled a well below Sable Island on the Scotian Shelf, which also penetrated a Cenozoic and Upper Mesozoic section. In 1970 the JOIDES Deep Sea Drilling Project drilled two wells in Orphan Knoll (Laughton et al., 1970), a continental plate remnant in deep water northeast of Newfoundland. Again, a Cenozoic and Upper Mesozoic section was penetrated. Recent discoveries of Cretaceous and Miocene rocks on the Scotian Shelf (Bartlett, 1968; Marlowe, 1969; King et al., 1970) came entirely from piston cores and dredged samples, and their correct stratigraphic positions are unknown.

The first modern, integrated, stratigraphic analyses of well sections in the Atlantic Margin regions of North America was published by Bartlett and Smith (1971) using the information from the Amoco wells. It is the purpose of this paper to extend this work to include the Sable Island and Orphan Knoll wells.

For this analysis the author has used all available data from the Pan Am 10E A 1 Grand Falls well, at Lat. 45°28'19"N and Long. 52°00'03"W, the Pan Am 10E A 1 Tors Cove well, at Lat. 44°11'14"N and Long. 52°23'42"W, the Mobil Sable Island No. 1 well, at Lat. 43°56'05"N and Long. 59°55'01"W, and the JOIDES Deep Sea Drilling Project Orphan Knoll (Site 111), at Lat. 50°26'N and Long. 46°22'W. These data came from a detailed study of the well cuttings, the sidewall cores, and the various mechanical logs of the three shelf wells, and a detailed examination of the microfauna. All data concerning the Orphan Knoll well are from Laughton et al. (1970) and Ruffman and van Hinte (1971).

It should be emphasized that the overall method of analysis involved independent analysis of the sediments and the mechanical logs by the author, a separate analysis of the faunas by G. A. Bartlett, and then a blending by the author of the two sets of data into a complete time-stratigraphic synthesis. For this reason a thorough knowledge of the faunas and their location in the sections was critical to the synthesis, even though this paper emphasizes the lithic aspects. Through such integration the author has been able to continue the work, begun for the paper by Bartlett and Smith (1971), of delineating the space-time history of the Grand Banks and surrounding regions, in both the purely stratigraphic sense and the tectonic sense.

The drilling of these four wells continues the author's unique opportunity to test and use the sequence hypothesis of Professor H. E. Wheeler in relation to this large and previously unknown region. The philosophy relevant to this hypothesis and the terminology related to it have already been discussed (Blackwelder, 1909; Wheeler, 1958, 1964; Bartlett and Smith, 1971). A procedure for analysis pertaining to unconformity - bounded sequences was given in detail in Bartlett and Smith (1971, p. 66-67).

Unconformity-bounded Sequences present in Offshore Wells

By following this procedure, seven sequences were delineated in the Grand Banks region (Fig. 1) and the upper six of these were also found in the Sable Island Bank well (Fig. 2). These unconformity-bounded units span the following time intervals: Pleistocene to Recent; lower mid-Miocene to latest Miocene; Intra-Eocene; Paleocene to earliest Eocene; Late Cretaceous (Turonian to mid-Maestrichtian); Middle Cretaceous (Aptian to Cenomanian); and Early Cretaceous (Neocomian). A Middle and Upper Jurassic section is also present in the Grand Falls well and a Middle Jurassic section is present in the Orphan Knoll well (Laughton et al., 1970; Ruffman and van Hinte, 1971). In neither was the entire Jurassic sequence penetrated to its basal unconformity. None of the wells reached basement.

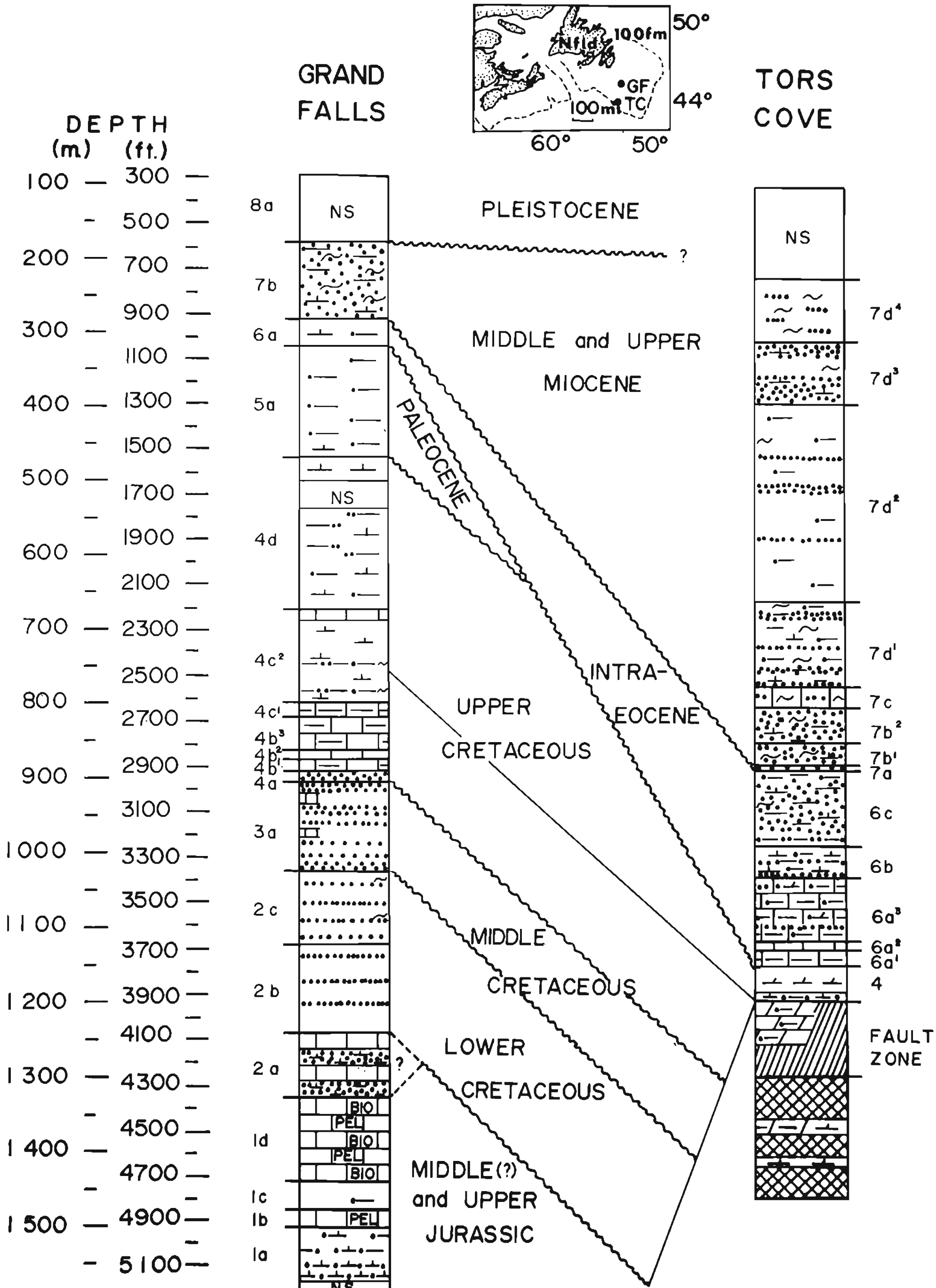

Figure 1. Stratigraphic section of the Grand Banks of Newfoundland (from Bartlett and Smith, 1971).

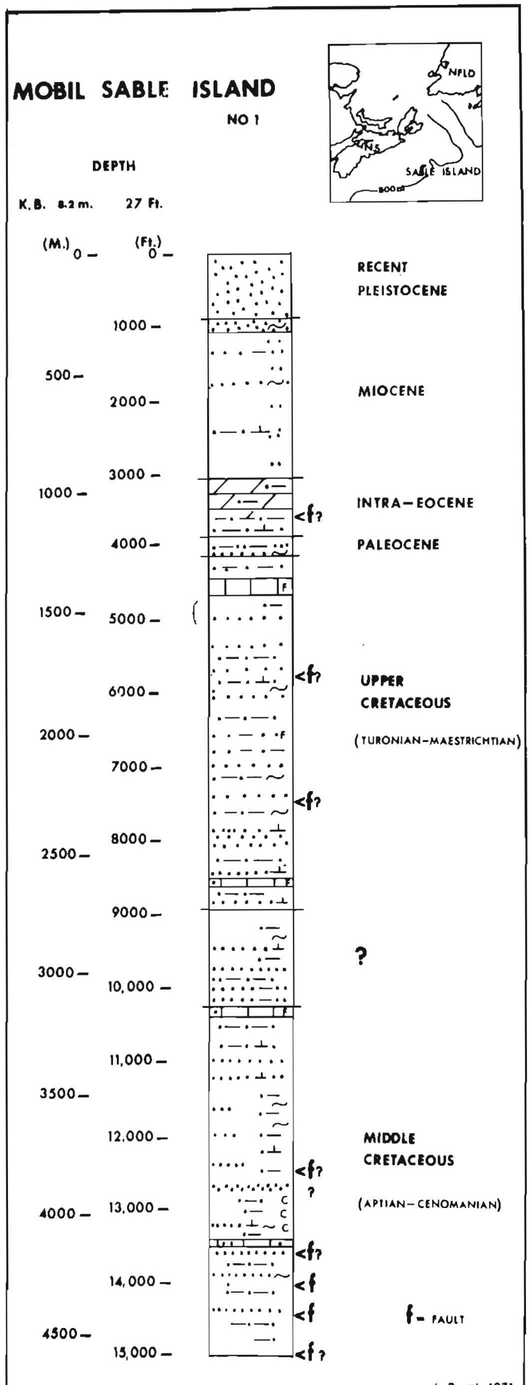

Figure 2. Stratigraphic well section in the Sable Island Bank, Scotian Shelf.

Description of the Sequences

The following are condensed descriptions of the various sequences as they are found in the Tors Cove well, the Grand Falls well and the Sable Island No. 1 well. Another description of the lithology and stratigraphic succession in the Sable Island well is given elsewhere in this volume (Magnusson, 1971).

Middle and Upper Jurassic Sequence

The discovery at Orphan Knoll of terrestrial sandstones with a Bajocian microflora shows that the lowest unit penetrated in that well was only a relatively short distance above the sub-Bajocian, Absaroka-Zuni discontinuity of Sloss (1963, p. 104). As the lowest terrigenous units in the Grand Falls well are calcareous and no coarser than silt size it is assumed that they were deposited somewhat later than these Bajocian sandstones. Thus, the entire Middle and Upper Jurassic sequence cannot yet be delineated from any of these wells.

The 802-foot (245 m.) Jurassic section in the Grand Falls well had earlier presented two other difficulties. The very sharp contacts between the lower siltstone-mudstone units and the two upper limestone units had already raised the question of undiscovered discontinuities in this section (Bartlett and Smith, 1971, p. 69). The author had thought that these discontinuities were tectonic rather than stratigraphic and should not have dismissed the idea, for seismic reflection data (Webb, 1971) has shown that the location of the Grand Falls well is close to a diapiric structure. This diapir appears to have caused faulting in the lower part of that section.

The other, and as it proved, related difficulty with respect to this Jurassic section concerned Formation 2a, the lower unit of the Neocomian sequence which overlies the Jurassic in the Grand Falls well. This 295-foot (90 m.) unit contains an interbedded mixture of coarse-sand-size lithoclasts of biopelmicrite and biopelsparite from the uppermost formation of the Jurassic combined with quartz sand in a micrite matrix. The limestones were derived from Formation 1d, but this formation contains no quartz sand, as it was deposited in a shallow, warm-water, low to medium energy environment, in a tectonically very stable region. This extreme and sharp change in environment from that which produced the 361-foot (110 m.) thick limestone unit to that in which the mixed unit had formed remained an enigma until the discovery of a diapiric structure just north of this well supplied an explanation for this latter stratigraphic unit.

For a diapiric intrusion into the rocks nearby, beginning during the erosional phase between deposition of the Upper Jurassic limestones and deposition of the basal Neocomian unit, would have caused a sudden and local uplifting of the limestones to an elevation significantly above baselevel. As a result they would have been rapidly eroded; but the energy level in the surroundings at the onset of Neocomian deposition would have been low enough to allow deposition of the carbonate lithoclasts near the erosional site. The energy level during the erosional hiatus would have been sufficient to cause introduction of quartz sand, as the surface of the bank was almost certainly subaerial at that time. Subaerial erosion also would easily explain the fragmentary nature of the limestone material in Formation 2a. Thus, a mixed deposit of quartz sand and carbonate lithoclasts would have formed near the diapiric uplift in earliest Neocomian time.

As the next formation of the Neocomian sequence (2b) is composed almost entirely of terrigenous fines it is obvious that any diapiric movement had ended and its related geomorphic effects had been eroded or covered before deposition of this formation began.

Neocomian Sequence

This sequence, unlike any other discussed in this paper, is present only in the Grand Falls well. The upper two formations of the unit, totalling 718 feet (219 m.) in thickness, consist of varicoloured shales and mudstones which are slightly carbonaceous to coaly. Several thin, very fine grained sandstones are scattered in the section. The upper formation (2c) is calcareous in part and is slightly glauconitic, showing a progression from essentially non-marine deposition in Formation 2b to a mixture of marine and non-marine deposition in Formation 2c. The non-marine environments are those of a very low-lying flood-plain and delta swamp. The marine waters were warm and very shallow.

Middle Cretaceous Sequence

This sequence occurs in the Grand Falls and the Orphan Knoll wells and is represented in the Mobil Sable Island No. 1 well. In the Orphan Knoll section this sequence is 190 feet (58 m.) thick and consists of a variety of glauconitic calcarenites and biocalcarenites (Laughton et al., 1970; Ruffman and van Hinte, 1971). These represent very shallow, warm-water depositional environments, completely cut off from any source of terrigenous fines. The Middle Cretaceous unit in the Grand Falls well is 400 feet (122 m.) of mudstones, sands, and minor limestones. The mudstones and sands are carbonaceous and unconsolidated, and the whole unit strongly suggests the classic coal cyclothem deposited in a deltaic, stream-channel and swamp environment of exceedingly low relief. The few limestones and the exclusively planktonic foraminifera in the formation represent brief incursions of warm-marine waters.

The Middle Cretaceous sequence cannot be closely defined in the Mobil Sable Island No. 1 well, although it is undoubtedly present. Its upper boundary appears to occur at either 8,950 feet or at 10,270 feet, and at Total Depth of 15,106 feet the well was still in rocks of this age. Thus the unit ranges from 6,156 + feet (1,875 + m.) to 4,736 + feet (1,443 + m.) in thickness. Between 10,270 and 10,430 feet, and again between 13,450 and 13,510 feet, there are several thin biosparite units. Several coal beds are present between 12,720 and 13,330 feet. The remainder of the unit consists of an interbedded mixture of quartz arenites and sub-feldspathic quartz arenites with silty mudstones. The arenites are variously calcareous and glauconitic. These sediments are closely similar to those of the same age in the Grand Falls well and, at least in part, also represent coal cyclothems.

The sideritic, glauconitic quartz arenites and sideritic siltstones of this sequence that were dredged from the Scotian Shelf NNW of Sable Island (King et al., 1970) are like several of the sandstones found in the upper half of the Middle Cretaceous section in the Sable Island well.

It had been planned to use the Cretaceous interval in the Sable Island well in a direct stratigraphic manner. Unfortunately, it was found that both the Upper and Middle Cretaceous sections are cut by numerous faults. These are especially noticeable from 3, 600 to 3, 650 feet, from 7, 500 to 8, 800 feet and from 13, 600 feet to Total Depth. The faults were recognized by one or a combination of the following criteria: repetition of faunas (pers. comm.: G. A. Bartlett), excessively steep dips in cores, crushed quartz grains, siliceous sandstone cements, slickensides, lost circulation zones, and high pressures near the bottom of the hole.

Having worked with the Tors Cove section, the author quickly suspected the presence of a nearby diapiric structure to be the cause of this faulting. Magnusson (1971) suggested the possibility of structures in the region due to salt tectonism and contemporaneous faulting. Webb (1971) afforded the critical seismic information, having found several domes and diapiric structures within the Shelf just north and east of Sable Island.

Upper Cretaceous Sequence

This sequence is the oldest one present, at least as a remnant, in all four wells. In the Orphan Knoll section it is represented only by an exceedingly thin (33 ft. -10 m.) section of chalk (Laughton et al., 1970; Ruffman and van Hinte, 1971). The Tors Cove Upper Cretaceous section was involved in salt diapirism and became truncated and dolomitized. The more complete section in the Grand Falls well, which is 1,434 feet (437 m.) thick, begins with a basal, thin, pure quartz arenite, typical of the arenites which are so widespread above interregional discontinuities. This unit is overlain sharply by the interregionally characteristic and geologically singular Upper Cretaceous chalks. These variously argillaceous carbonates, 239 feet (73 m.) in thickness, are overlain by 1, 154 feet (352 m.) of variously calcareous mudstones, which exhibit that distinctive North American Upper Cretaceous element of an influx of terrigenous sediment in sufficient quantity to mask carbonate deposition. Thin limestones in the middle of this mudstone section are biostromes of non-argillaceous shell calcarenites and calcrudites, like the "oyster-reef" accumulations of the present Gulf of Mexico.

The depositional environments of the chalks, mudstones and limestone biostromes all appear to have been in very shallow, warm, marine waters. In the case of the mudstones and the biostromes these were near of very low relief and a fluctuating coastline position where terrigenous influx was able in general to overcome the carbonate-producing abilities of the marine environment.

The Upper Cretaceous section in the Sable Island well contains the mudstone and shell biostrome elements of the Grand Falls section. It also contains fine-grained glauconitic sandstones. However, although the pelecypod calcrudites are thicker in the Sable Island well, there is no hint of the interregionally characteristic Upper Cretaceous chalks.

Although it is possible that the influx of terrigenous sediment was sufficient here, throughout the entire Turonian to Maestrichtian interval, to suppress or mask chalk accumulations, it is also possible, in view of the preceding discussion, that the chalks earlier present in this section have been faulted out. Thus, this unit, which may range from 4, 790 feet (1, 460 m.) to

6,110 feet (1,861 m.) in thickness, due to the uncertain position of the Upper - Middle Cretaceous boundary, also cannot be used as a normal stratigraphic section.

Paleocene and Lowest Eocene Sequence

This sequence is present in the Grand Falls and Sable Island wells, and is possibly present in the Orphan Knoll section. In the Grand Falls section it consists of 496 feet (151 m.) of silty, pyritic mudstone. In the Sable Island well it consists of 260 feet (80 m.) of slightly glauconitic mudstones and minor, fine-grained, calcareous, glauconitic sandstones. In both these wells deposition was again closely related to a very low-lying coastal region, with marine incursions being more persistent in the region of the Sable Island Bank. The minor variations in lithology between this sequence and those above and below it point to relatively small variations in the region's depositional environments and terrigenous sediment sources from one depositional phase to another, during the time from high in the Upper Cretaceous until the mid-Eocene, even though these phases were separated by erosional episodes.

Intra-Eocene Sequence

This sequence is represented in the Grand Falls well only as a thin remnant. In the Sable Island well it is an 800-foot (244 m.) unit of microcrystalline, silty dolomite and calcareous siltstones. In the Tors Cove well it is an 878-foot (268 m.) unit of variously argillaceous and silty, micritic limestone and dolomitized limestone overlain by silty mudstone and fine-grained, calcareous and very argillaceous sands. An equivalent of this upper portion of the Tors Cove section has not been preserved in the Sable Island well due to the early Miocene erosional episodes.

The sediments of this sequence were deposited in a warm, shallow-marine environment of slight current activity, situated seaward of a very low-lying deltaic region. The influence of terrigenous sediment dispersal from this deltaic complex became progressively greater, as indicated by the change upward from variously argillaceous limestones and dolomitic limestones to sands. The non-calcareous character of some of the sediments suggests that sporadic subaerial deposition may have occurred.

Middle and Upper Miocene Sequence

This sequence, like the Intra-Eocene sequence below it, is found in all three shelf wells. Again, a thinner section is preserved in the Grand Falls well under the upper, erosional discontinuity. This thinner unit, however, is still 344 feet (105 m.) thick, and is composed of glauconitic and calcareous sands. The section in the Tors Cove well, although 2,183 feet (665 m.) in thickness, begins at the base with essentially the same sedimentary unit. Above this is a sandy, glauconitic, micritic limestone, 88 feet (27 m.) thick. The upper five-sixth of the Tors Cove section consists of mudstones, claystones and variously calcareous, glauconitic and argillaceous sands, the latter concentrated in the lower and upper portions.

The basal part of this succession was deposited in an offshore bank area of moderate current energy. The presence of sandy limestone indicates that the waters were warm and probably quite shallow. The waters deepened during deposition of the mudstones and claystones, while organic mixing destroyed much of the depositional interlayering. The resultant depositional surface was exceedingly flat and featureless over very large areas.

The Sable Island section, with a closely comparable thickness of 2,202 feet (671 m.), contains the sandy mudstones and argillaceous, variously glauconitic sands of the other section, but here these sands are concentrated in the upper 200 feet (60 m.). There are no sands or sandy limestones in the lower part of the section, suggesting either that terrigenous sands were not supplied to this region or that the transporting energies of the currents were not strong enough to move sand-size particles. The lack of carbonate sediment is due to an abundance of terrigenous fines.

A composite stratigraphic section for the Miocene Period has been constructed (Bartlett, 1968; Marlowe, 1969) using data from samples dredged off the slopes of The Gully, a submarine canyon located at the east edge of the Sable Island Bank. Microfaunal work by Bartlett (pers. comm.) has shown that most of the Miocene is represented in this section, including some of the Lower Miocene not found in the wells. The microfauna is characteristic of warm marine waters, as are the lime muds present in the section. The thickness given by Marlowe for this section is twice that for either the Sable Island or Tors Cove Miocene sections. The order of lithologies in this column, however, is basically similar to that in the well sections, especially the Tors Cove section, with greater quantities of sand in the lowermost and uppermost portions.

Pleistocene and Recent Sequence

This sequence of sands is found in all three shelf wells. In the Tors Cove well the unit is 412 feet (126 m.) in maximum thickness and the the Grand Falls well its maximum thickness is 321 feet (98 m.). The Sable Island well section, however, is 898 feet (274 m.) thick. The depositional environment for the Grand Banks section was marine, as it was for most of the Sable Island section. The upper 170 feet (52 m.) of this latter section, however, is devoid of organic remains and was deposited in an island environment like that present today.

Salt Dome Unit (probable Upper Triassic)

This additional stratigraphic unit, present in the lower portion of the Tors Cove well but in a salt-dome tectonic relationship with surrounding units, was earlier thought to be a Permian depositional entity (Bartlett and Smith, 1971). As diapiric structures, presumably consisting of evaporites, have since been found at numerous other localities in the eastern Canadian continental margin (Keen, 1970; King and MacLean, 1970a; Webb, 1971) an accurate depositional age for these evaporites is critical to an understanding of the stratigraphic history of the region.

The author has been studying the interregional occurrence of Mississippian to Jurassic evaporites in the North Atlantic region, in terms of their contained and associated lithologies, and the related faunas, in order to find any characteristics that would uniquely identify the stratigraphic succession in which each major evaporite unit occurs. The salt and the variegated and grey, dolomitic shales and argillaceous, very fine dolomites interbedded with it, which occur in the offshore eastern Canada region, are so closely comparable to those in the known Upper Triassic (Keuper) evaporite succession of the North Atlantic region it is very probable that they also were deposited during that time.

Relation between Seismic and Stratigraphic Sections

Numerous seismic refraction surveys have been done over the Scotian Shelf and the Grand Banks regions, and data from some of these have been interpreted in terms of travel times through the sediment in order to construct generalized sections. Comparison with the well sections has brought out the following correlations. In the Sable Island Bank region, the top of the Upper Cretaceous biostromal limestone unit is the surface used by Officer and Ewing (1954) as the top of their "Consolidated" unit. Berger et al. (1965) place the top of their "Semi-consolidated" unit at the position of the upper surface of the Intra-Eocene carbonate unit. In the Grand Banks region the top of the Upper Cretaceous biostromal limestones was used by Press and Beckmann (1954), in a section very near the Grand Falls well, as the top of their "Consolidated" unit. The top of the Upper Jurassic limestone unit has the same position as the top surface of their "Basement" unit. The top of the mid-Miocene sandy limestone is in the same position as the top surface of the "Semi-consolidated" unit in the section of Bentley and Worzel (1956) located slightly west of the Tors Cove well. As salt diapirism has affected the lower part of this well section it is only possible to state that either the top of the Upper Cretaceous biostromal limestones or the top surface of the Upper Cretaceous chalks can be correlated with the top of their "Consolidated" unit.

It is clear from these correlations that the upper surfaces of limestone units serve as the best reflecting surfaces within the stratigraphic section of the Eastern Canadian Continental Margin.

Seismic and Geomorphic Delineation of Regional Unconformities

As the essential defining characteristic of a sequence is its bounding unconformities, the sequence analysis of this paper, using well sections, is closely related to the delineation of erosional surfaces by seismic-reflection profiling or through geomorphic analysis. The earliest example of the latter is Spencer's paper (1890), in which he concludes that the basic form of the Laurentian Channel, between the Scotian Shelf and the Grand Banks, was formed by fluvial erosion "probably" in the later Tertiary period. This would require the surrounding higher area to have been subaerially eroded at the same time. From seismic data both Uchupi (1969) and King and MacLean (1970b) conclude that the basic geomorphic forms of the Scotian Shelf are the result of fluvial erosion in immediately pre-glacial time, and Keen, Loncarevic and Ewing (1970) state that the upper bedrock surface of these forms is an erosional surface. King and MacLean (1970b) also note an irregular, fluvial-erosion surface overlain by Tertiary sediments, and they postulate a pre-Cretaceous subaerial erosion surface in the Laurentian Channel area. Thus, seismic and geomorphic investigations led to the finding of three of the eight regional erosional discontinuities delineated by sequence analysis of the well sections of these regions.

Mesozoic and Cenozoic Tectonism of Sable Island Bank and the Grand Banks

The nature of the baselevel transit cycles that occurred in the Grand Banks and Sable Island Bank region in response to interregional epeirogenic oscillations, together with the study of the sediments and faunas contained in the resultant unconformity-bounded sequences, have shown that only low-level tectonic activity occurred in that region during late Mesozoic and Cenozoic time. Scattered diapiric tectonism locally altered this situation but did not change its interregional expression.

The author has portrayed the area-time history of the Grand Banks region in Figure 3, using a generalized, two-dimensional form which shows the baselevel transit patterns. The time-stratigraphic units used in Figure 3

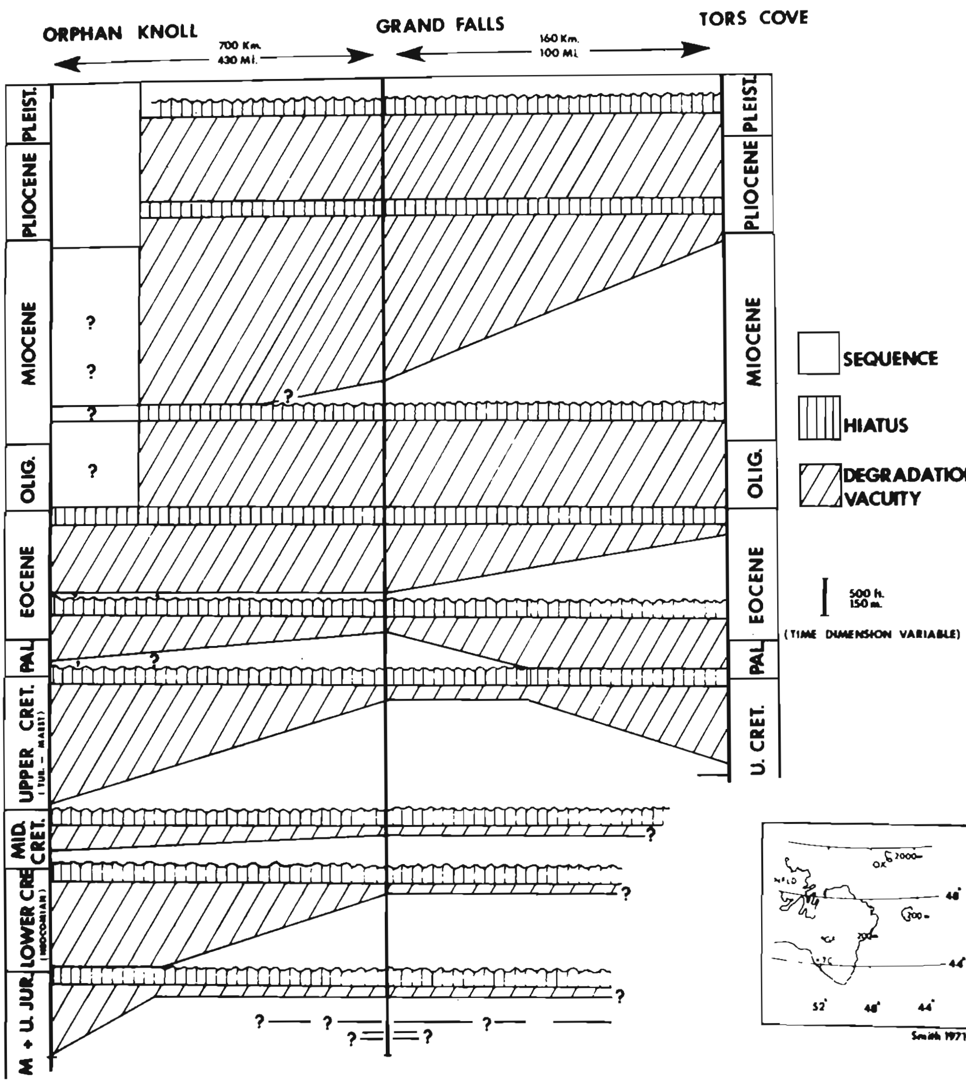

Figure 3. Area-time cross-section of the Grand Banks region since mid-Jurassic.

are as defined and used in Wheeler (1963, 1964). The sum of each sequence-degradation vacuity pair shows the total succession produced during the depositional phase of each baselevel transit cycle. The degradation vacuity is that part of the total unit removed during the succeeding erosional phase (the hiatus) of that cycle, or of later cycles. The hiatus itself records the time necessary to perform the erosional removal. The sequence is, therefore, simply that portion of the original depositional unit that has been preserved to the present.

The vertical thickness scale in Figure 3 is constant within the sequences but the vertical time dimension is not constant and is distinctly variable with respect to absolute time. Figure 3 shows, however, the minimum holostromal thickness that was deposited in the region during the sub-baselevel episode of each baselevel transit cycle since mid-Jurassic time. As stated in Bartlett and Smith (1971, p. 79) "the vertical thickness of each hiatus can only be considered an approximation but the faunal gaps present in the sections suggest that it is valid as a minimum expression of the time span of the various suprabaselevel episodes that affected the region".

A detailed discussion of the tectonic relationships in the Grand Banks region is given in Bartlett and Smith (1971, p. 79-80). It is concluded that the periodic tectonic oscillations which affected this region produced a succession of unconformity-bounded sequences, the regional form of each sequence being that of a wedge thickening toward the continental edge. A study of the lithic and faunal data shows that this thickening is due to differential preservation of holostromal units and not to differential depositional thicknesses.

A comparison of the thickest section for each sequence present in either the Grand Falls or the Tors Cove wells with the section of the same sequence in the Sable Island No. 1 well shows that the Sable Island section throughout is characteristic of a continental margin-edge section. Almost every sequence present is at least as thick as its equivalent in the Grand Banks wells, that is, the section exhibits considerable preservation of each of the sequences present.

Also notable in Figure 3 is the reversal of the thickness pattern for the Paleocene-lowest Eocene and the Upper Cretaceous sequences in the lower part of the section. This reversal is entirely due to mid-Early Eocene salt-dome tectonism. It is now known, as discussed earlier, that the Jurassic units in the Grand Falls section were also affected by diapiric movement, but the last movement of this diapir to affect that section occurred between Late Jurassic and Neocomian times. It is possible, in view of the rock type present in the lowest Neocomian formation, that the diapiric movement did not entirely end until earliest Neocomian time. It is now evident that the Sable Island No. 1 well also shows considerable faulting resulting from nearby salt dome diapiric movement. Accordingly, all three wells are to some extent affected by this second characteristic type of tectonic movement for the Eastern Canadian Continental Margin region, and no section affected only by inter-regional tectonism is yet available.

Stratigraphic and Tectonic Relations of Orphan Knoll to the Grand Banks

A comparison, using Figure 3, of the time-stratigraphic and tectonic histories of the Orphan Knoll area and the southern Grand Banks region gives significant results concerning the time of separation of this continental plate fragment. All information concerning the Orphan Knoll well section, its lithologies and ages is taken from Laughton et al. (1970) and Ruffmann and van Hinte (1971).

In the Jurassic, Orphan Knoll was still attached to North America, as the Jurassic sediments show active subaerial sedimentation of sand-size sediments, and, thus, continuity with a larger continental mass. The thinness of preservation of this sequence also shows that the Orphan Knoll area was tectonically more positive during the following hiatuses than was the Grand Banks region itself, as only the lower part of the sequence is preserved on the Knoll.

Although this tectonic activity produced a subaerial erosional surface at the base of the Middle Cretaceous and although the glauconitic calcarenites deposited during the Middle Cretaceous demonstrate that only shallow, marine water covered the area at this time, the fact that none of the great amount of terrigenous sediment being supplied to the Grand Banks region reached the Knoll, during limestone deposition, suggests that the Knoll had already sufficiently separated from the continental plate to allow bypassing of excess terrigenous sediment into the Atlantic Basin. For otherwise, these limestones are similar to those of the Middle Cretaceous sequence on the Grand Banks.

Although separated, Orphan Knoll remained tectonically more positive during the erosional episode between the Middle and Upper Cretaceous depositional episodes, reducing the Middle Cretaceous holostrome to a very thin remnant.

An exceedingly thin Upper Cretaceous chalk sequence is found on the Knoll, and has been assigned a Maestrichtian age (Laughton et al., 1970; Ruffmann and van Hinte, 1971). Although interregionally the Upper Cretaceous chalk units begin with the Turonian, the separation of the Knoll would have allowed the large quantities of terrigenous sediment that were introduced into the Grand Banks region during the Maestrichtian to have been either deposited on the Banks or bypassed into the Atlantic Basin, without any reaching the depositional surface atop the Knoll regardless of its depth.

But separation with respect to sediment dispersal is not the only aspect to be considered. There is also the question of tectonic separation with respect to the interregional tectonic oscillations which affected eastern North America throughout the late Mesozoic and the Cenozoic. If the chalks in the Orphan Knoll section are Maestrichtian in age then Orphan Knoll was not only sedimentationally separated from the Grand Banks but was also acting as a tectonically independent unit compared with the rest of eastern North America. For if there are no Turonian or Senonian units preserved on Orphan Knoll, a direct contrast exists with the other wells being considered, in which Turonian and Senonian units are present under the Maestrichtian. Thus, if the Orphan Knoll chalks are Maestrichtian in age, and do not have any Turonian affinities, the Grand Banks region was below baselevel during Turonian and Senonian time while Orphan Knoll would have had to be above baselevel for at least the latter portion of that time interval.

The thin Eocene clay unit in the Knoll section may be divided by a major discontinuity, as shown in Figure 3, for the Paleocene and lowest Eocene sequence on the Banks is composed exclusively of terrigenous fines and could be an equivalent unit to at least part of the Orphan Knoll "Eocene". However, if there is any unit equivalent in age to the Intra-Eocene sequence of the Banks, then the upper surface of the Knoll must have been in fairly deep water as there is no lithic similarity between any section of that age on the Knoll and those on the Banks. Whether the Knoll was tectonically independent of the North American continental mass at this time cannot be ascertained until detailed faunal data are available.

In the upper portion in Figure 3 a separation is shown between the section for the Grand Banks region and that for Orphan Knoll, as the published information suggests a continuous Knoll section from the latest Miocene to the Recent. Whether tectonically attached to the continental mass or not it appears that during this period the Knoll was in sufficiently deep water and was not swept by strong currents so that sedimentation could be continuous.

CONCLUSIONS

1. The Pan American (Amoco) Grand Falls and Tors Cove wells in the Grand Banks of Newfoundland and the Mobil Sable Island No. 1 well were the first to penetrate Mesozoic and Cenozoic sections in the offshore eastern Canada region, and encountered the only known sections of Jurassic and probable Upper Triassic (Keuper) sediments in the Atlantic continental margin of North America.

2. An integrated analysis of all available lithic and faunal data allowed delineation of the space-time history of the Grand Banks and Sable Island Bank during late Mesozoic and Cenozoic time; initially in terms of the comprised stratigraphic entities, the sequence and the lacunas, and, lastly, in terms of the tectonic features which produced these stratigraphic results, or later affected them.

3. This analysis, following Wheeler (1958, 1964) delineated seven sequences of the following ages: Pleistocene, mid-Miocene to Upper Miocene, Intra-Eocene, Paleocene to lowest Eocene (upper Danian to middle Thanetian), Upper Cretaceous (Turonian to mid-Maestrichtian), Middle Cretaceous (Aptian to Cenomanian), and Lower Cretaceous (Neocomian).

4. The following rock types are present: (1) limestones, dominating the Upper Jurassic, lower Upper Cretaceous and Middle Eocene, and significant in the Middle Miocene; (2) sandstones, dominating the lowest Cretaceous, Middle Miocene, portions of the Upper Miocene and the Pleistocene, and significant in the Neocomian, Middle Cretaceous and Upper Eocene; (3) mudstones, dominating the Middle Jurassic, Neocomian, upper Upper Cretaceous, Paleocene and lowest Eocene, and portions of the Upper Miocene, and significant in the Middle Cretaceous and (4) halite and dolomite in the Tors Cove well section, of probable Keuper age.

5. A wide variety of depositional environments were present during the subbaselevel episodes that affected the region. These include the subaerial, mainly dune environments of the quartz and quartz/limestone arenites at the base of the Neocomian, Upper Cretaceous and Miocene sequences. Also subaerial were the very low-lying land areas of deltaic stream channels, floodplains and swamps during the Neocomian and parts of the Middle Cretaceous and mid-Eocene. In terms of sediment volume, the most important marine environments were the shallow, nearly-flat bottoms of near-shore open bays and the open shelf, which received abundant, fine, terrigenous sediment during Middle Cretaceous, the upper Upper Cretaceous, Paleocene and lowest Eocene, mid-Eocene and parts of the Middle and Upper Miocene. An offshore sand bank was present in Middle Miocene. Lesser dispersals of terrigenous sediment allowed accumulation from these warm marine waters of the argillaceous, fine limestones and marls of the lower Upper Cretaceous, mid-Eocene

and mid-Miocene. Virtual cessation of this influx, along with a very slight restriction of the waters, allowed production and deposition of the massive limestones of the Upper Jurassic and the biostromes of the Middle and Upper Cretaceous.

6. An essential contrast exists between the present, irregular, erosional surface of this suprabaselevel episode and the past, regular, depositional surfaces of the subbaselevel episodes, which have been preserved. In addition, the sedimentary environments associated with the present surface have been affected drastically by eustatic sea level changes, whereas the past environments were epeirogenically controlled.

7. There are notable similarities in geological history between the Grand Banks and the Sable Island Bank regions. The stratigraphic positions of the interregional discontinuities, the ages of the sequences present, especially their lower portions, their thicknesses, and the close lithic resemblance of their Middle Cretaceous, upper Upper Cretaceous, Paleocene, Intra-Eocene carbonate, and Upper Miocene sections show that these regions have been acted upon as part of an interregional geologic enity.

8. The sequence analysis afforded an intrinsic framework with which to analyze the two principal tectonic features of the Mesozoic and Cenozoic of the eastern Canadian continental margin, i.e. periodic, interregional oscillations, which expressed themselves in the alternately depositional and erosional phases of the baselevel transit cycles (Fig. 3), and salt dome intrusions, whose stratigraphic effects can be seen in Figures 1, 2, and 3.

9. The subbaselevel depositional episodes of the baselevel transit cycles appear to have been times of essentially equal downsinking over the whole region, while the suprabaselevel, erosional phases were times of distinctly differential uplift.

10. In areas that have had no salt-dome activity, the known stratigraphic result of the geologic history of the region is a pile of unconformity-bounded sequences of Mesozoic and Cenozoic sedimentary rocks, which, due to preservation, thicken toward the continental edge.

11. Only about one-half of late Mesozoic and Cenozoic time is represented by the preserved sedimentary sequences at the continental edge of the Sable Island and Grand Banks.

12. It appears that the last movement of the salt diapir in the Tors Cove well occurred in mid-Lower Eocene and that of the diapir north of the Grand Falls well occurred in the very earliest Neocomian.

13. Varied lines of evidence demonstrate the presence of several faults in the Middle Cretaceous and lower Upper Cretaceous section of the Sable Island well, which are probably due to nearby diapiric movement.

14. The upper surfaces of the various limestone units are the best seismic reflectors in the region.

15. The Orphan Knoll fragment was still part of the eastern North American continental plate in Middle Jurassic time. It had separated by Middle Cretaceous time, but its top remained subaerial or in shallow water until the early Tertiary, when the Knoll began to sink into deep water.

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18. BEDROCK AND SURFICIAL GEOLOGY OF THE NORTHERN GULF OF ST. LAWRENCE AS INTERPRETED FROM CONTINUOUS SEISMIC REFLECTION PROFILES

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Abstract

In the fall of 1969, the author obtained approximately 1,000 nautical miles (1850 km) of airgun reflection profiles in the northern Gulf of St. Lawrence in conjunction with a seismic refraction program carried out by the Exploration Geophysics Division of the Geological Survey of Canada. The

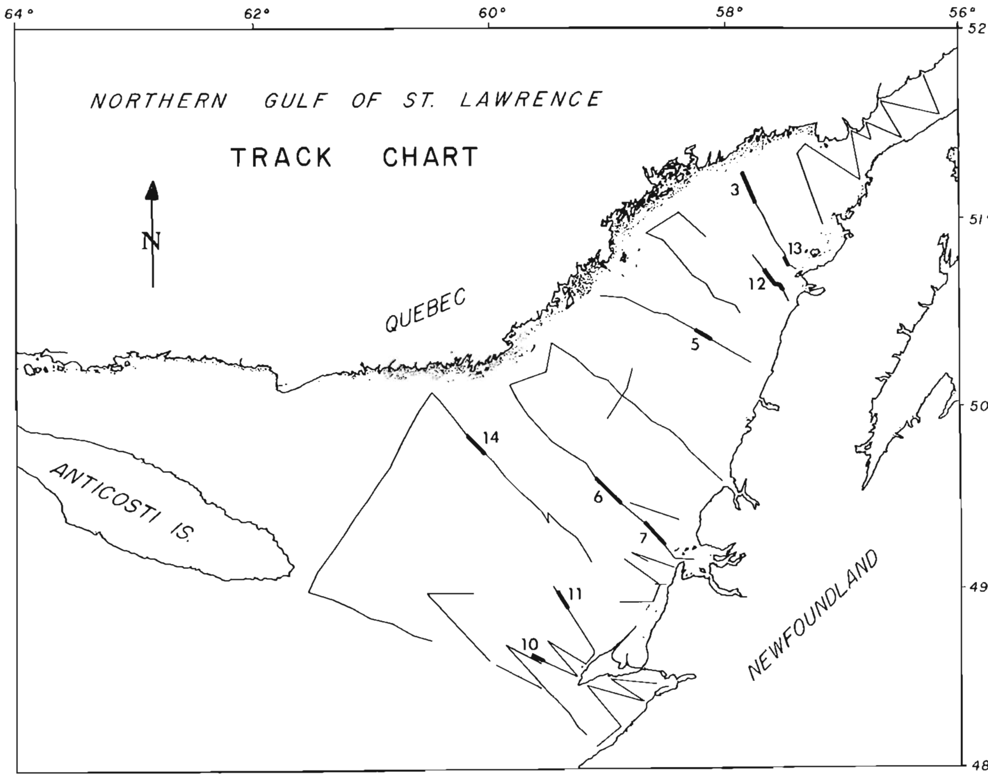

Figure 1. Track chart of 1969 survey by CNAV SACKVILLE in the northern part of the Gulf of St. Lawrence. The thicker lines indicate the positions of the seismic profiles reproduced in subsequent Figures together with the appropriate Figure number. The scale for all these maps is indicated by consecutive degrees of latitude which represent 60 nautical miles.

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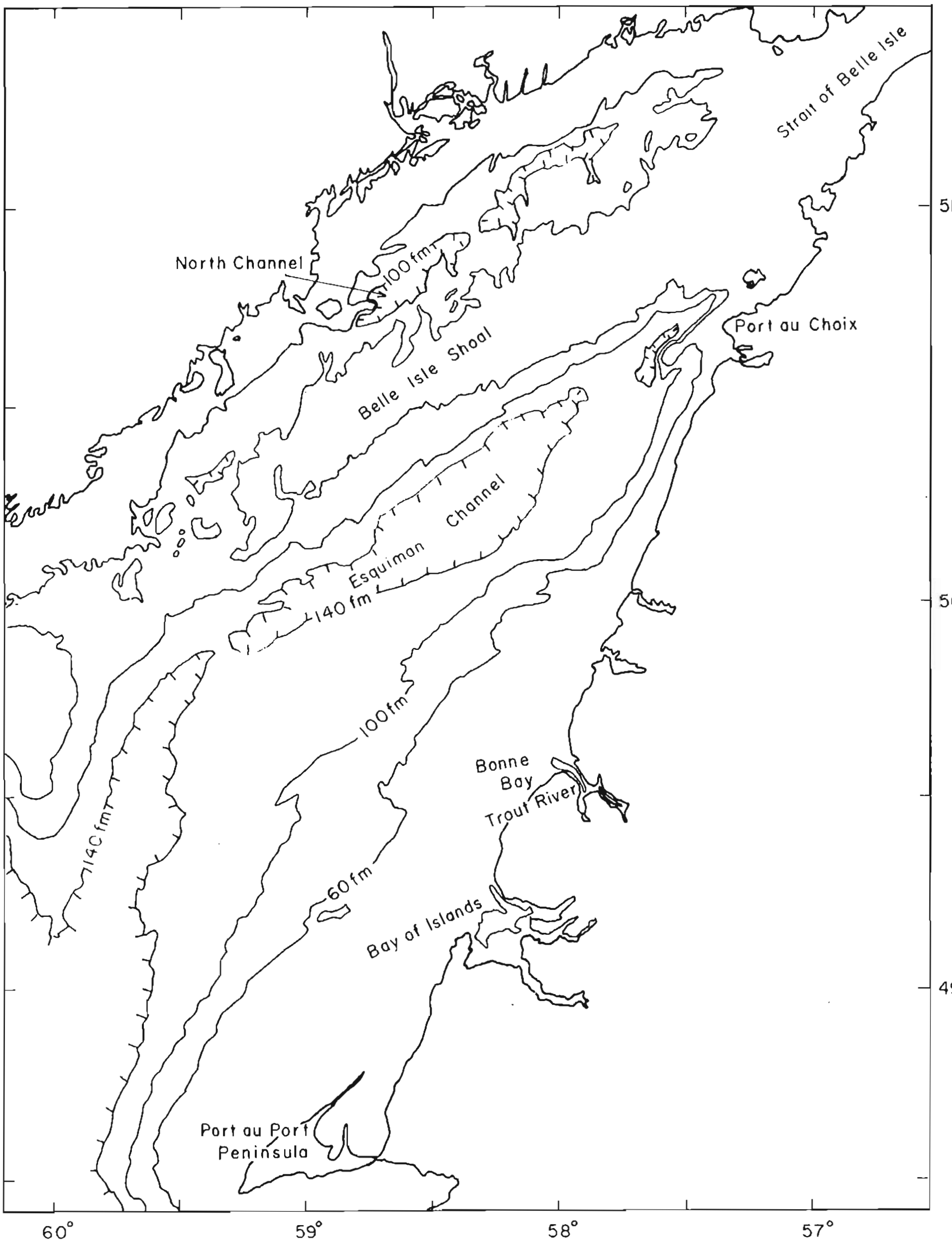

Figure 2. Bathymetric chart of the northern Gulf of St. Lawrence (Advance proof of bathymetric chart 4021 based on compilation from detailed work done by CSS BAFFIN in 1968 and 1969). In the northern part of the chart, much of the north-south linearity observed is thought to be due to sub-glacial, submarine melt-water channels further eroding already glacially-scoured pre-existing fractures in the bedrock.

results indicate that the northern Gulf is underlain chiefly by bedded Cambrian to Silurian platform-type rocks. Gently east-dipping sediments occur over most of the area to a line about 5 miles (8 km) off and parallel to the west coast of Newfoundland, where a pronounced syncline and dip reversal occurs. The west Newfoundland "klippe" extends barely beyond the present coastline (maximum 5 miles) and is overlain by the eastern flank of the syncline mentioned above. The Paleozoic section just west of the Port au Port peninsula is probably about 13,000 feet (4 km) thick; the neoautochthonous sub-section comprises 6,000 feet or about 2 km of this. A gradual decrease in the thickness of the Paleozoic rocks occurs as one moves northward. The presence of till over most of the northern Gulf would seem to indicate that glacier ice covered the study area in Wisconsin time. The apparent absence of till in the shallower depths (down to $\simeq$ 100 m) must be due to reworking by bottom currents or by wave action associated with a much lower post-glacial sea level.

INTRODUCTION

In Septermber 1969, a seismic reflection survey was carried out in the northern Gulf of St. Lawrence (Fig. 1) in conjunction with a refraction seismic program undertaken by members of the Geological Survey of Canada using the CNAV SACKVILLE. A Bolt 600B airgun (1 cu. in. chamber) operating at airpressures between 1,000 and 1,500 lb. sq. in. was employed for the seismic profiling survey.

The study area lies between the shorelines of Quebec on the north and Newfoundland on the east. The northern shore is characterized by rugged, steep and irregular terrain typical of the Grenville and Precambrian shield rocks. The surface irregularity of these rocks is quite apparent offshore and is seen in the bathymetric compilations based on the detailed hydrographic surveys carried out by CSS BAFFIN in 1968 and 1969 (Fig. 2). Likewise, the submarine morphology off the west coast of Newfoundland is characteristic of the onshore geology being gently-dipping lower to middle Paleozoic rocks.

Bedrock Geology

Although the northern Gulf of St. Lawrence is not an ideal area for seismic reflection studies (because most formations have similar high seismic velocities), enough difference in the character of the reflections from these units exists to permit a consistent distinction between them. In this respect, the proximity of the shoreline permitted a tentative identification of the various units by extrapolation from the known onshore geology.

It is pertinent here to summarize briefly the known geology along the shorelines of the study area. Most of the west coast of Newfoundland and Anticosti Island are composed of a somewhat undeformed gently east-dipping sequence of lower Paleozoic (Cambro-Ordovician) platform sediments (autochthonous sequence). In central west Newfoundland, a sequence of highly folded and sheared clastics occurs which belongs to the so-called Taconic "klippen" (allochthonous sequence) described by Rodgers and Neale (1963), Brückner (1966) and others. The klippe composed mainly of the Humber Arm Group rocks (clastics and basic volcanics) and ultrabasic rocks of the Bay of Islands igneous complex, was thought to have slid westward into its present

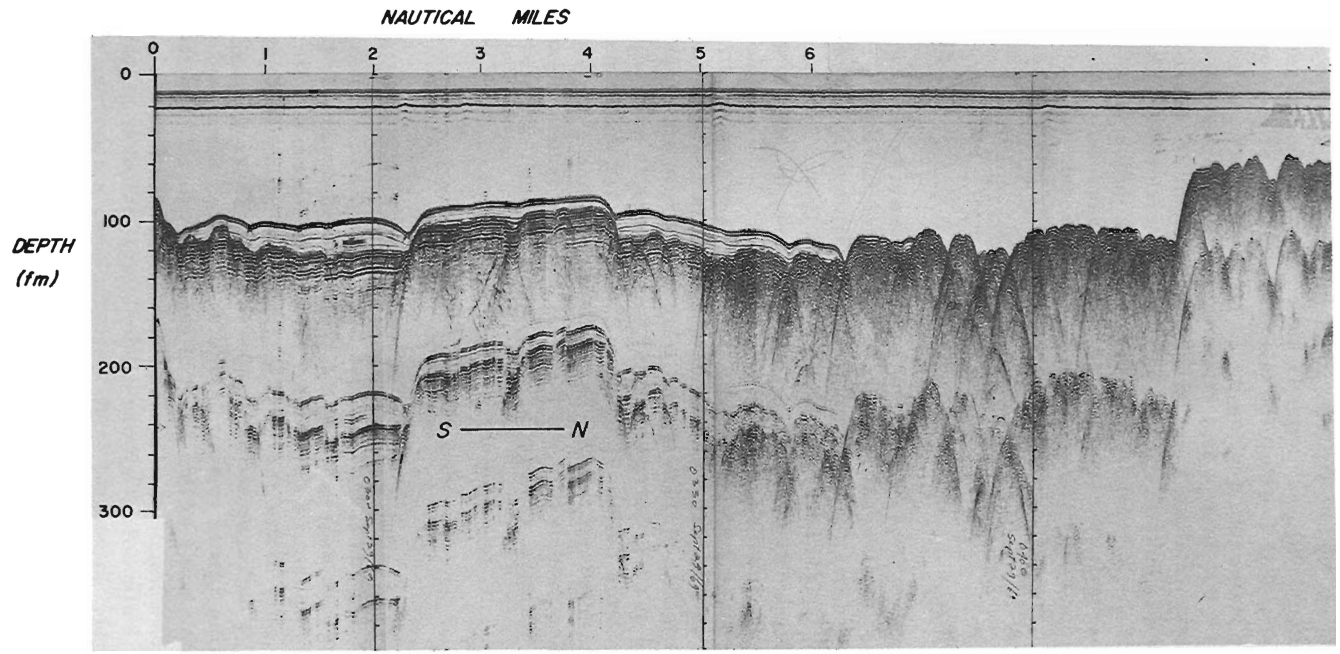

Figure 3. Seismic profile over part of the North channel. At the 5 nautical mile (9.3 km) mark is the contact of the faintly-bedded sedimentary rocks (on the south) and the reflector from the shield rocks. Abrupt deepening in the northern part is possibly due to a fault in the Precambrian rocks. Of interest is the limited extent of outwash deposit, which is absent over much of northern portion of the Figure; possibly due to a recessional ice front at that point.

position in Middle Ordovician times. Overlying the klippe sequence is a relatively undeformed sequence (Middle Ordovician) of platform sediments (neo-autochthonous sequence) which is very similar in character to the autochthonous sequence underlying the klippe (see G.S.C. Map 1231A, Island of Newfoundland, 1967).

Let us now examine the offshore seismic data and attempt an interpretation within this given geological framework. In the Belle Isle area, Cambrian and Ordovician shallow-water sediments outcrop on both sides of the strait. Unfortunately, little penetration of seismic energy into these units occurred which is thought to be due to the high reflection coefficient of the rocks because of their high seismic velocities. However, a somewhat smooth surface with occasional mesa-like structures and faint bedding was apparent on the records, which enabled the lower Paleozoic sequence in this area to be distinguished from the more irregular morphology of the Precambrian basement rocks (Fig. 3).

In this manner the central Belle Isle shoal seen on the bathymetric map (Fig. 2) has been identified as lower Paleozoic and possibly the resistant Upper Cambrian quartzite (Hawke's Bay Formation). The Hawke's Bay

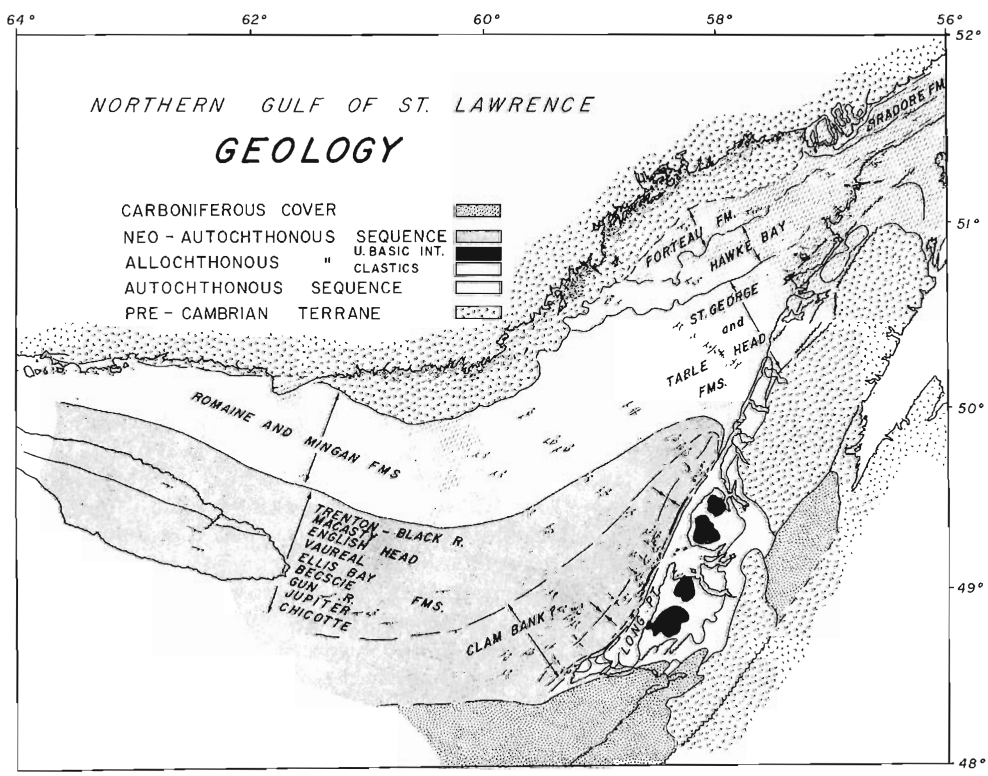

Figure 4. Geological map of the northern Gulf of St. Lawrence. Dips shown are those measured from seismic records assuming a seismic velocity for the sediments three times that of water ($\simeq$15,000 feet/sec or 5 km/sec).

Formation outcrops on the Labrador and Newfoundland coasts and stratigraphically would seem to fit into a position where the shoal now exists (Fig. 4). Continuing southward, the Esquiman channel is likewise thought to be cut into lower Paleozoic sediments (Fig. 5). A very gentle regional dip to the southeast exists over all this area (Fig. 4) and it appears that the rocks in the channel overlie the rocks forming the shoal. This being the case, the rocks in the channel are tentatively designated as Lower Ordovician St. George and Table Head silty dolomites.

Figure 6 shows the contact between the lower Paleozoic rocks described above and a somewhat conformable and well-bedded sequence overlying it. This well-bedded sequence can be followed southeastwards down dip (dip $\simeq$ 1 to 2°) to a point about 5 miles (8 km) off the west coast of Newfoundland where a pronounced dip reversal is observed (Fig. 7). Just east of the dip reversal, this sequence overlies an apparent massive unit in which no bedding or structures are apparent, which is thought to be part of the klippe (allochthonous sequence) bordering the west coast of Newfoundland in this area. This being the case, the bedded sequence overlying the klippe rocks is part of the neoautochthonous sequence which is corroborated by the investigations of Lilly (1966) who made bottom scuba diving observations in this vicinity (5 miles seaward from Trout River) and recovered Clam Bank Formation rocks. It is not known whether the contact observed in Figure 6 is, in fact, that contact between the autochthonous and neoautochthonous rocks on the west side of the northern Gulf. Cumming (1967) has determined the time of klippe emplacement in west Newfoundland to be between the Whiterock and Wilderness stages of the Middle Ordovician. This is the time when a short hiatus between the Mingan and Trenton-Black River Formation existed on Anticosti Island (Table 2, Economic Geology of Canada, 1971). This point in time is slightly older than a prominent seismic reflector (within the Trenton-Black River Formation) described by Rolliff (1968). This would seem to imply that the contact observed in Figure 6 is within the Trenton-Black River Formation and slightly younger than the actual time of klippe emplacement along the west coast of Newfoundland.

With the predominantly northeasterly strike of the beds in the study area, the thickness of the platform sediments increases in a southeasterly direction. The approximate thickness of the neoautochthonous section near the axis of the syncline may be estimated by summing up the individual bed thicknesses on the surface from the travel times to reflectors on the seismic record using a seismic velocity three times that of water ($\simeq$15,000 feet/sec or $\simeq$4.5 km/sec). In the Bay of Islands area (Fig. 7), 3,000 to 4,000 feet ($\simeq$1 km) of post-klippe sediments are present, while further south off the Port au Port Peninsula up to 6,000 feet (2 km) may be present. On the western side of the northern Gulf (north of Anticosti Island), the autochthonous rocks only seem to be between 1,500 and 2,000 feet ($\simeq$0.5 km) thick. Figure 4 of Lilly (1967) shows these pre-klippe platform sediments to be much thicker (1 to 2 km) along the west coast of Newfoundland, presumably because of greater crustal subsidence in this area. The basin of deposition is of contrasting nature to the synclinal structure existing in the neoautochthonous rocks which is thought to be a result of the deformation associated with klippe adjustment and regional metamorphism accompanying Acadian orogenic activity in the central mobile belt of Newfoundland. There is no evidence of large-scale normal faulting (graben structures) as proposed by Kumarapeli and Saull (1966). In essence, a geological section similar to that published by Cumming (1967) is envisaged.

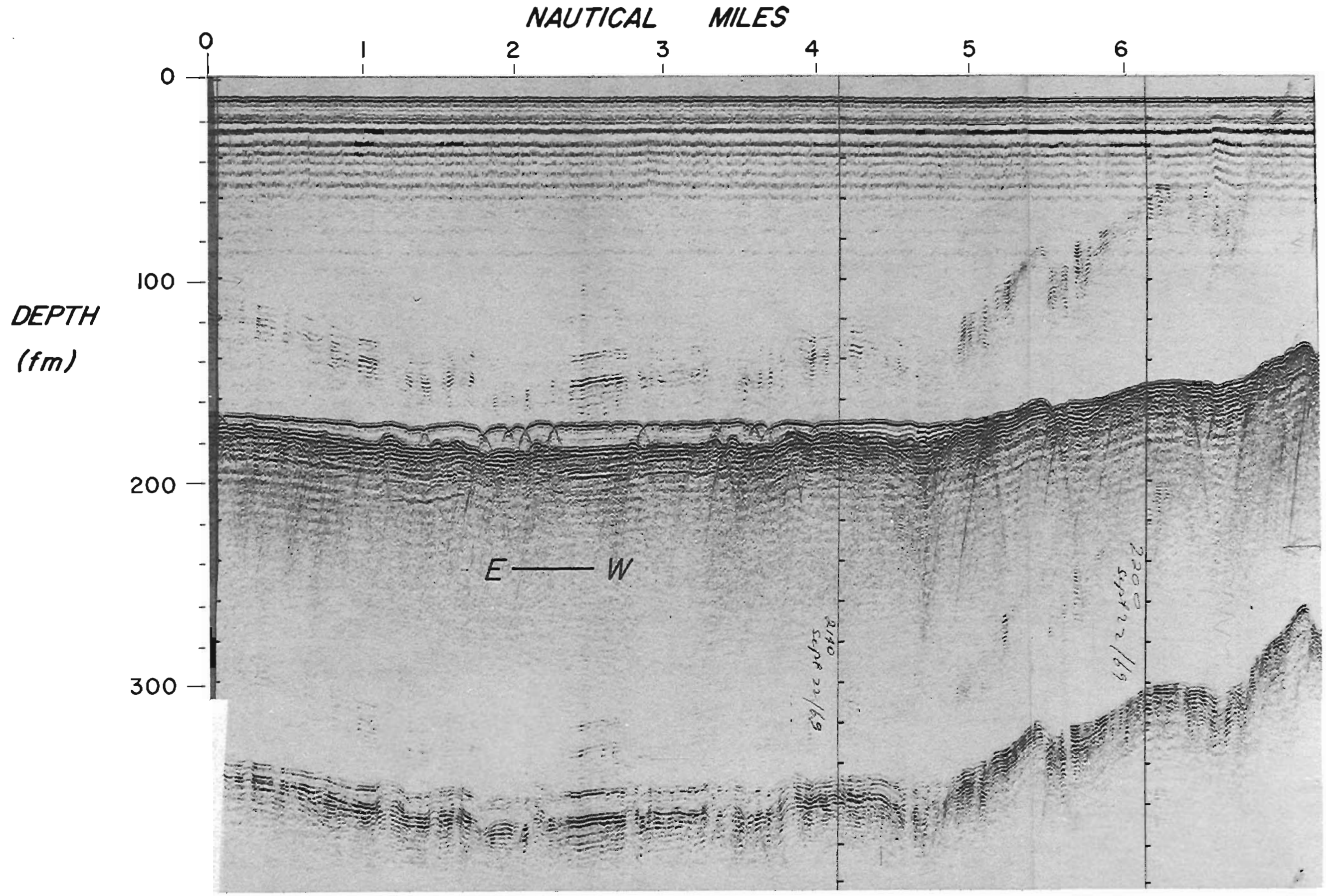

Figure 5. Seismic profile over typical lower Paleozoic sediments. Mesa-like features on right of Figure and faint bedding near the centre indicate a gentle dip to the east. Note the depressions in the recent sediments which are thought to be similar to the pock marks documented by King and MacLean (1970) on the Scotian Shelf.

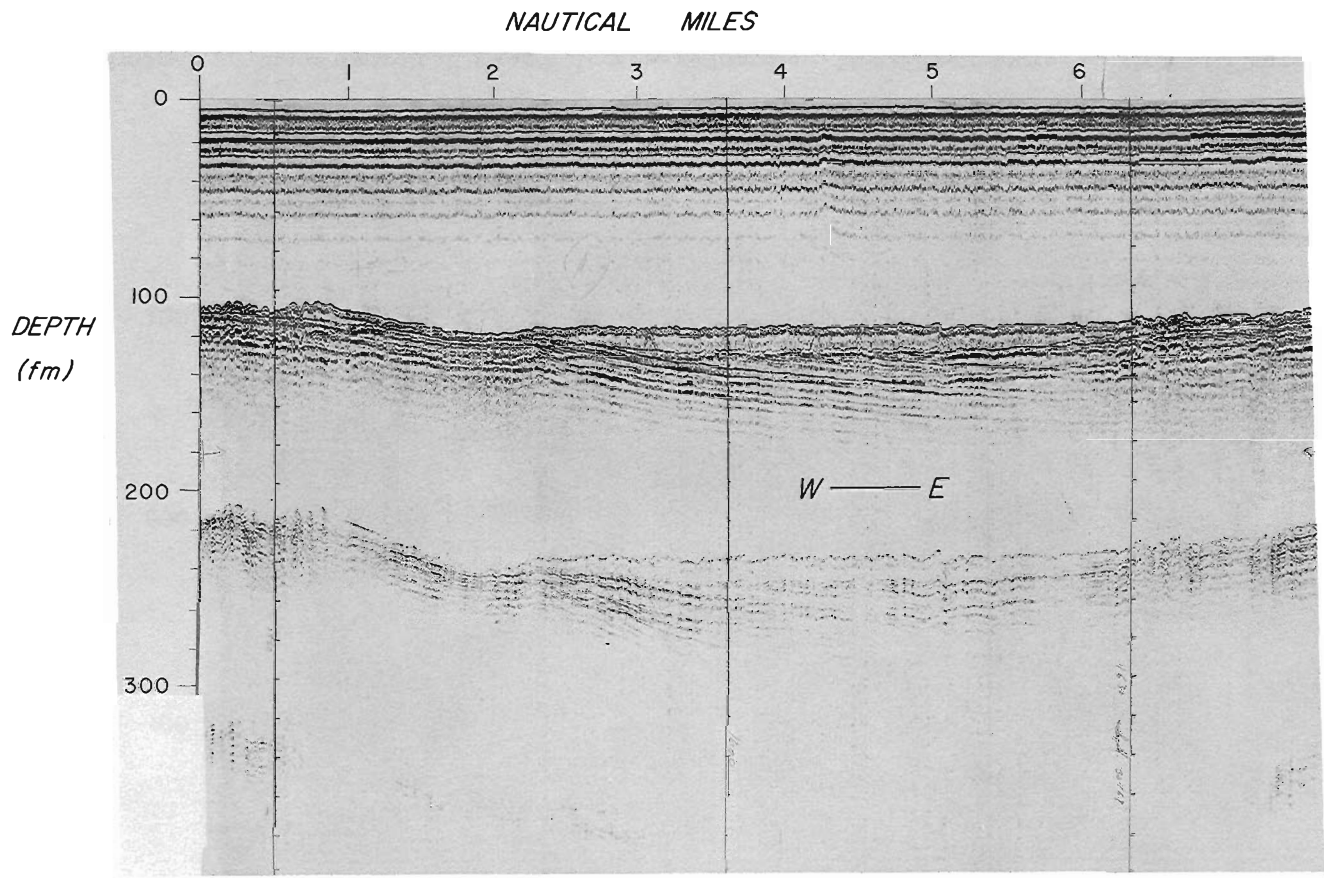

Figure 6. Lithological contact within the Paleozoic sequence near the neoautochthonous and autochthonous time boundary. Gentle eastward dips (≃1 to 2 degrees) exist. Overlying the bedrock is a till and/or ice-rafted deposit.

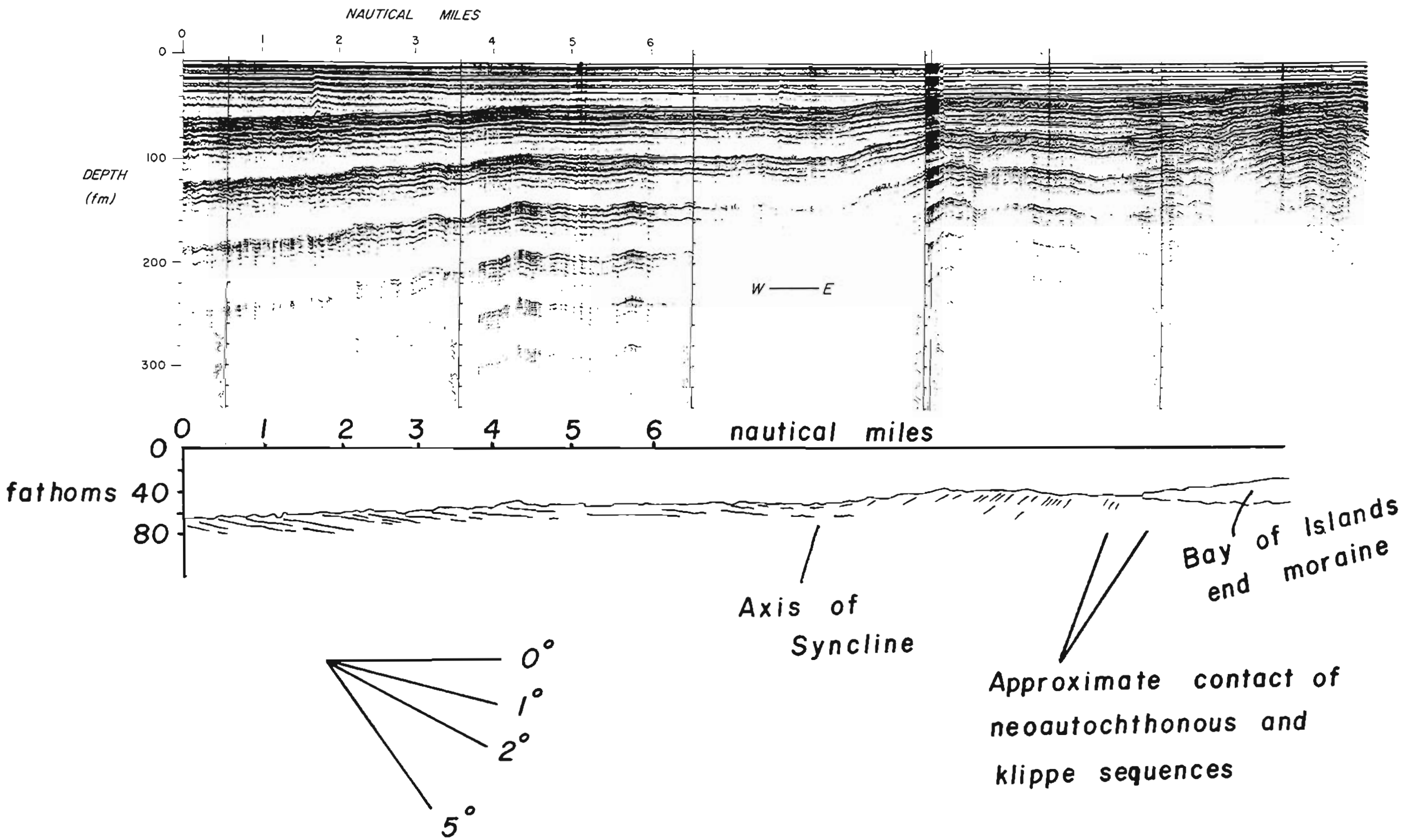

Figure 7. Interpretation cross-section and original seismic record obtained off Bay of Islands, Newfoundland showing syncline in the neoautochthonous sediments. Dips shown do not take into consideration higher seismic velocities (three times that of water) in sediments.

A recently-discovered magnetic anomaly (thick black line, Fig. 4), named "Odd Twins" by Ruffman and Woodside (1970), is located some 6 nautical miles (10 km) off the west coast of Newfoundland between Bonne Bay and Port au Port Bay. They have proposed that it may be caused by a westerly-dipping dyke near or at the contact of the neoautochthonous and allochthonous klippe rocks. The present author concludes from an interpretation of seismic data obtained on the lines across the feature, that it lies within the neoautochthonous sequence. It is inferred that the magnetic signature is due to a sedimentary layer rich in magnetite and chromite located within the Clam Bank Formation. Such concentrations of heavy minerals are possible in areas where an ultrabasic source and a relatively stable environment exist enabling repeated sorting and reworking e.g. a shoreline. The apparent extent of this feature coincides with the distribution of the ultrabasic masses of the Bay of Islands igneous complex (Black bodies on Fig. 4 in western Newfoundland).

<u>Surficial Geology - Glacial and Post-Glacial Geology</u>

A combined interpretation of echo-soundings and continuous seismic records has enabled the distribution of glacial and post-glacial deposits to be approximately established (Figs. 8 and 9). The deposition of clay material has occurred in the deeper channels and basins of the study area (Fig. 8).

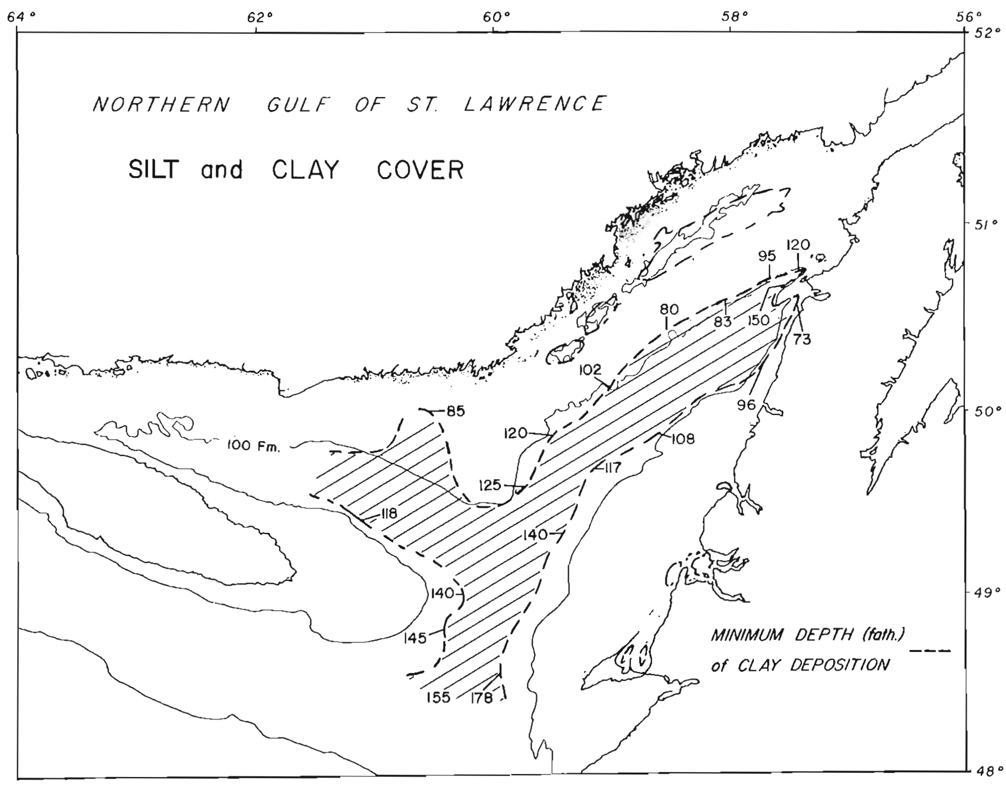

Figure 8. Approximate distribution and minimum depth of the post-glacial silt and clay cover in the northern Gulf of St. Lawrence.

Generally, the upper limit of the silty clay unit deepens to the south; this dis-bribution being dependent primarily upon bottom current velocities. An anomalous distribution of post-glacial clays and silts is shown in Figure 3 where much of the basin apparently contains very little material. It is thought that the present rate of deposition of material from suspension in this area is very slow and that submarine sub-glacial outwash has been deposited outward from a recessional ice-front positioned near the edge of the observed deposits.

The study area is located on the fringe area of the classical Wisconsin ice-front. Wisconsin ice was thought to cover the land masses surrounding the northern Gulf as is indicated by the presence of glacially-modified land forms and glacial deposits. Within the resolution of the seismic equipment used, the whole of the northern Gulf seems to have a thick cover of till-like material with the exception of two areas; the deepest part of the study area and areas where bottom depths are generally less than 60 fm (110 metres).

Within the area of till cover, linear zones occur which are probably end-moraine features. Figures 10 and 11 are reproductions of a section of seismic record over an end-moraine feature located some 20 nautical miles (37 km) west of the Port au Port Peninsula. Similarly, Figures 12 and 13 are sections of seismic records taken over the northern flank of the Esquiman channel near Port au Choix. Even this sketchy distribution of morainal

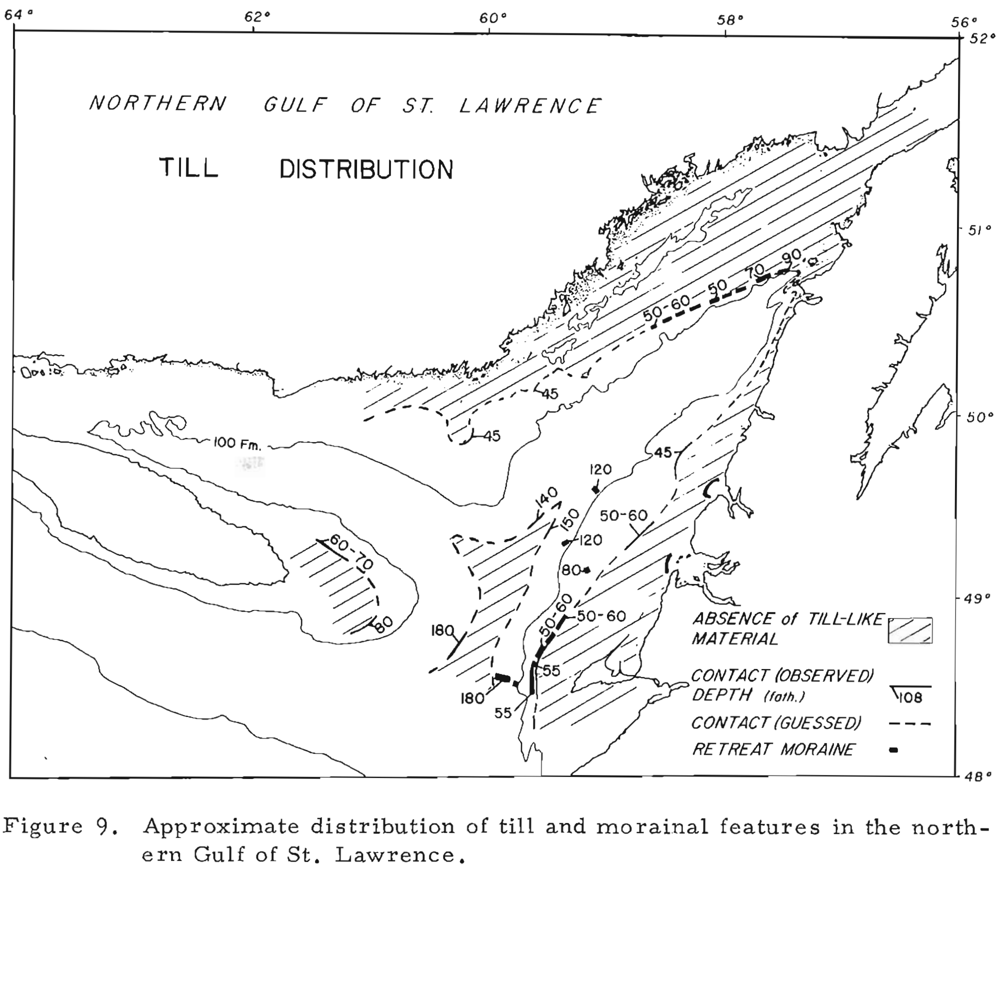

Figure 9. Approximate distribution of till and morainal features in the northern Gulf of St. Lawrence.

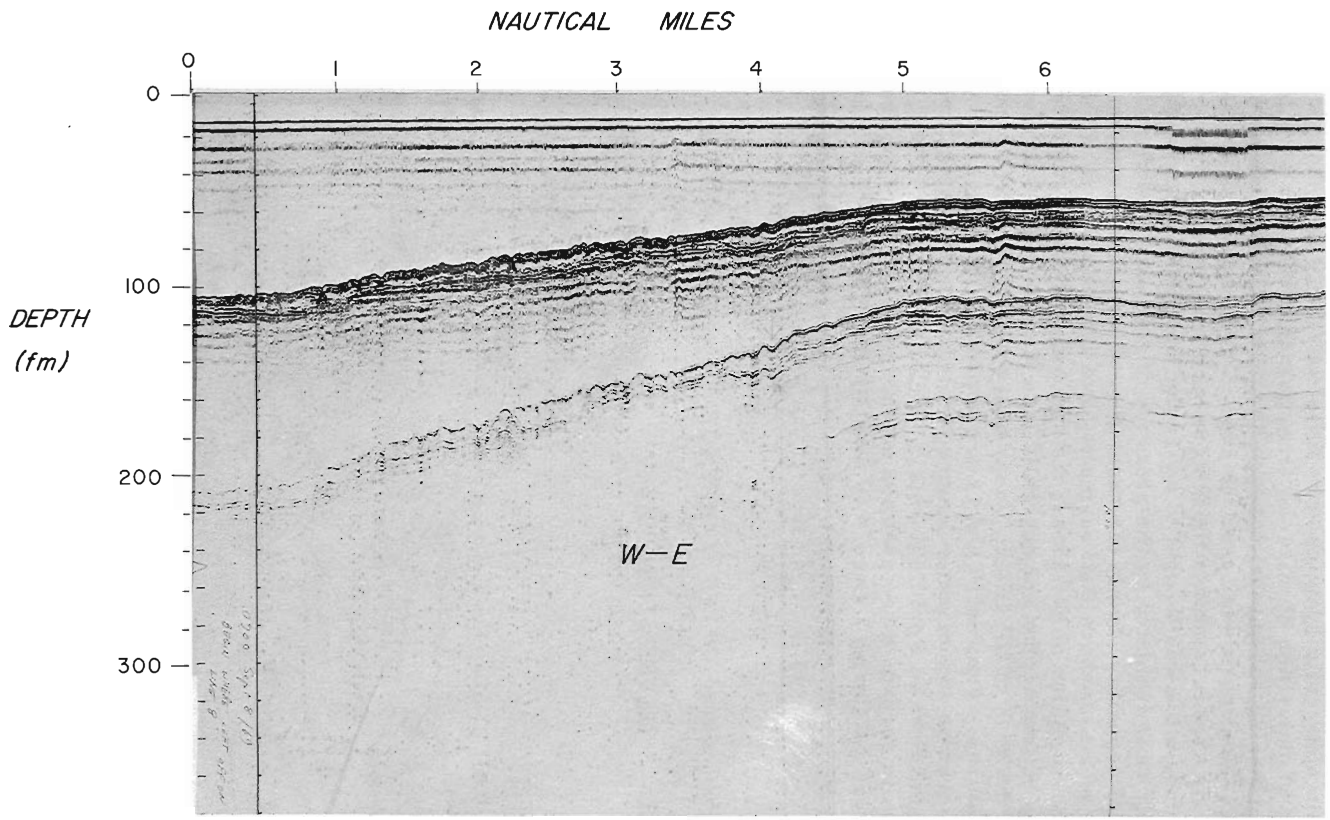

Figure 10. Seismic profile demonstrates presence of an end moraine about 20 nautical miles (30 km) off the west coast of Newfoundland. Irregular surface morphology and the presence of many point reflectors are identifying characteristics of till. Note the contrast in surface characteristics at a depth around 60 fm (110 m) indicating reworking of the original till surface. Bedrock in this area is the gently east-dipping neoautochthonous unit with some small-scale normal faulting having possibly offset the beds near the west end of the profile.

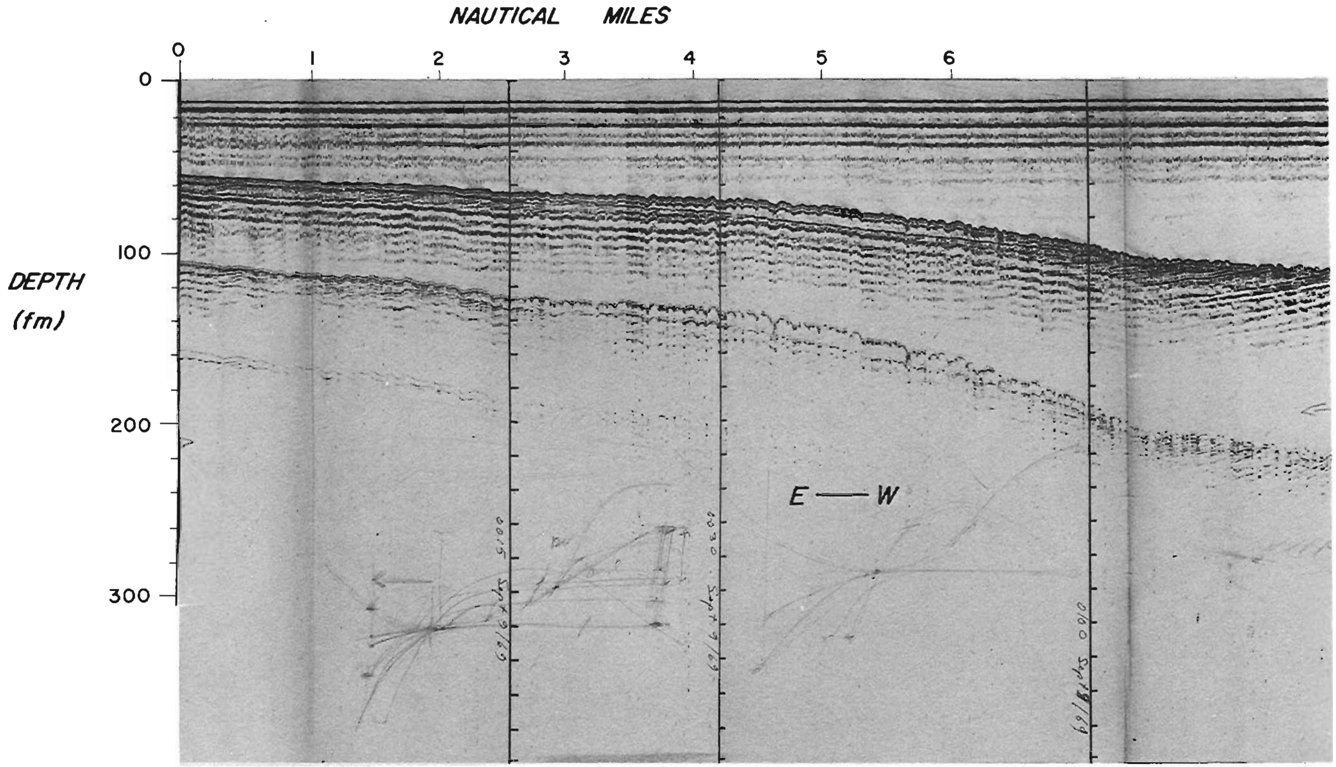

Figure 11. Seismic profile over the same end-moraine system seen in Figure 10. Similar modifications to the original till surface at depths less than 60 fathoms (110 m) seem also to have occurred. Well-defined easterly-dipping beds of the neoautochthonous sequence are apparent.

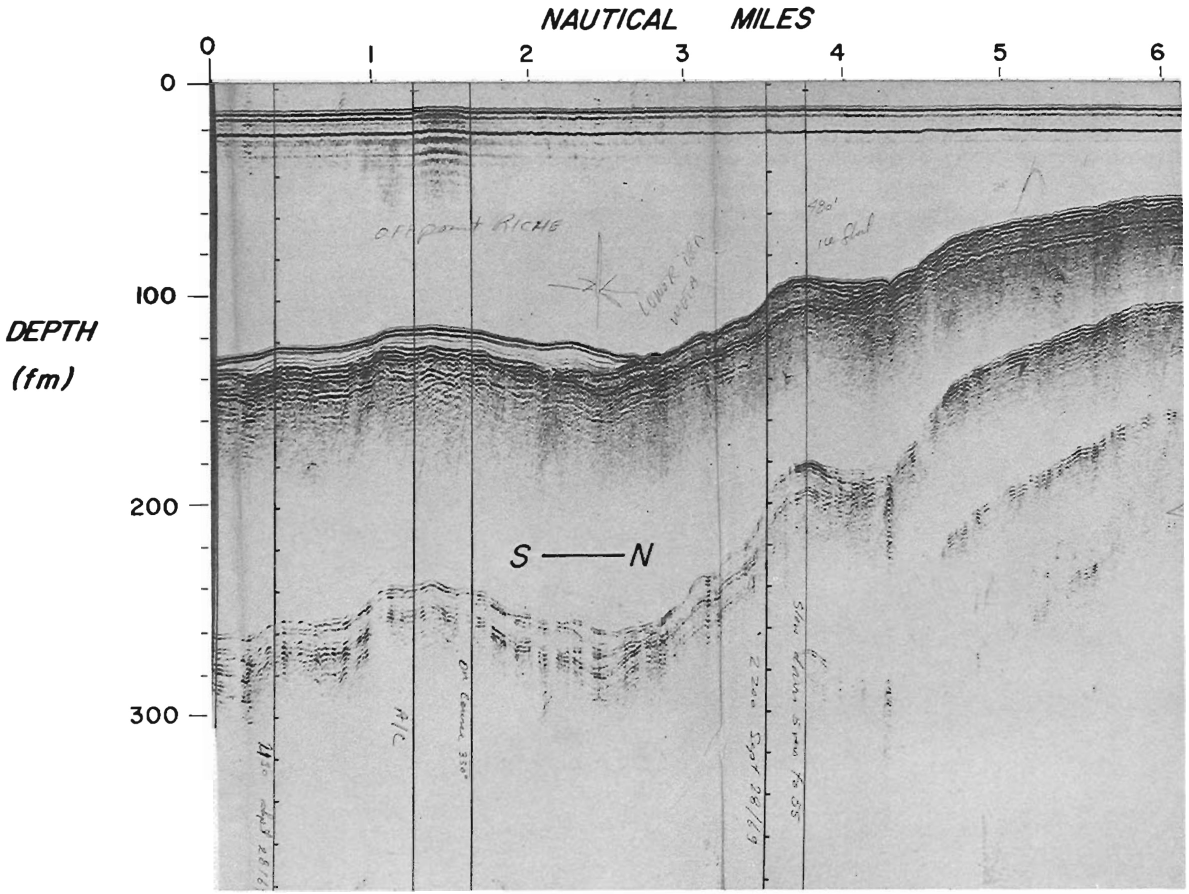

Figure 12. Seismic profile over an end-moraine system on the northern edge of Esquiman Channel. At southern edge (2 nautical mile mark), contemporaneous outwash deposits begin which are overlain by post-glacial clays. Smooth surface of moraine and bedrock (autochthonous units) to the north is attributed to mantling by sands and gravels transported by strong tidal currents through the Strait of Belle Isle.

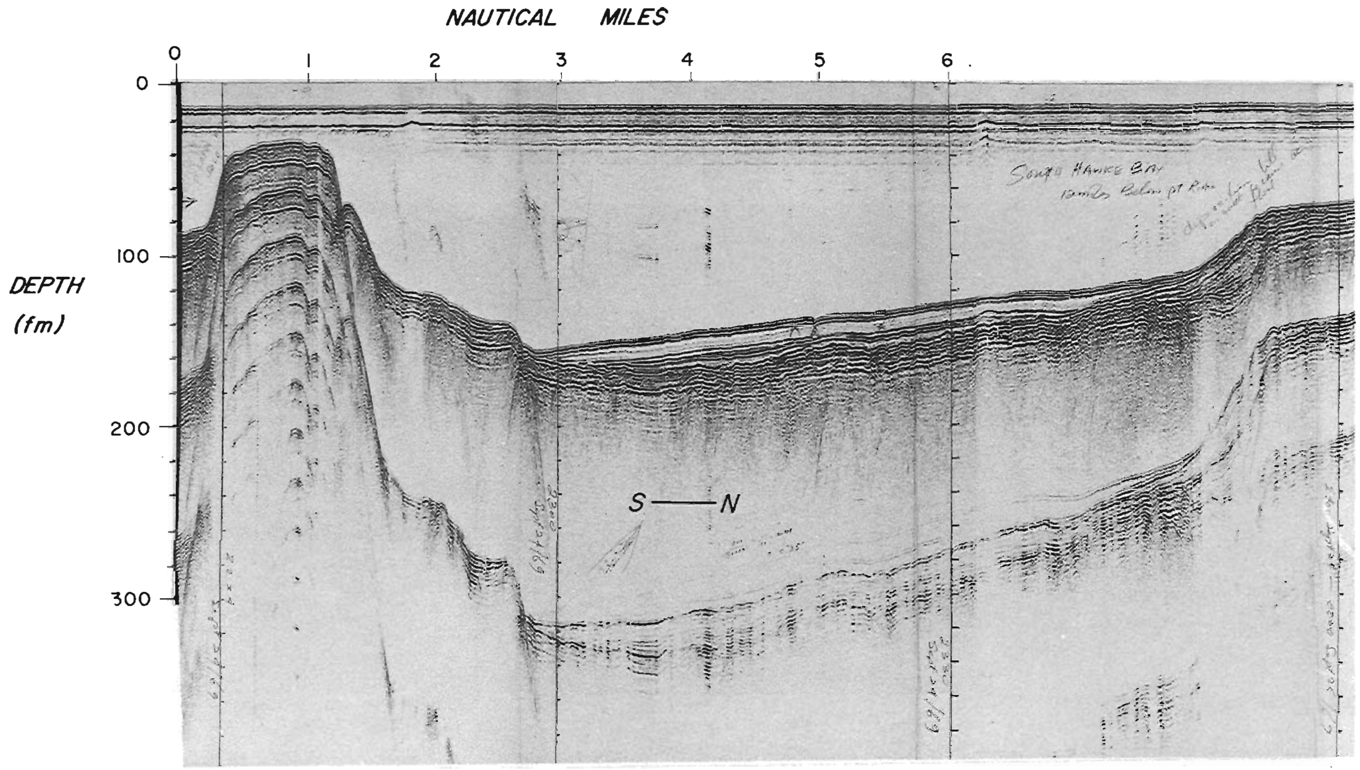

Figure 13. Seismic profile over the same recessional moraine (right hand side of Figure) as seen in Figure 12. Near the base of the moraine, bedding is visible and is thought to be due to the slumping of debris at the time of deposition because of the steep depositional slope. Note the outwash material seen beginning at the southern tip of the moraine, both in turn overlain by post-glacial silts and clays. Bedrock in the area is lower Paleozoic (autochthonous sequence) with the mesa-like feature on the south side being the submarine extension of the Port au Choix Peninsula. Of interest in the interpretation of the record is the first multiple which consists, in fact, of the first return from the bottom, reflected again from the bottom and the sub-bottom superimposed upon the first return of the sub-bottom, reflected again from the bottom and the sub-bottom.

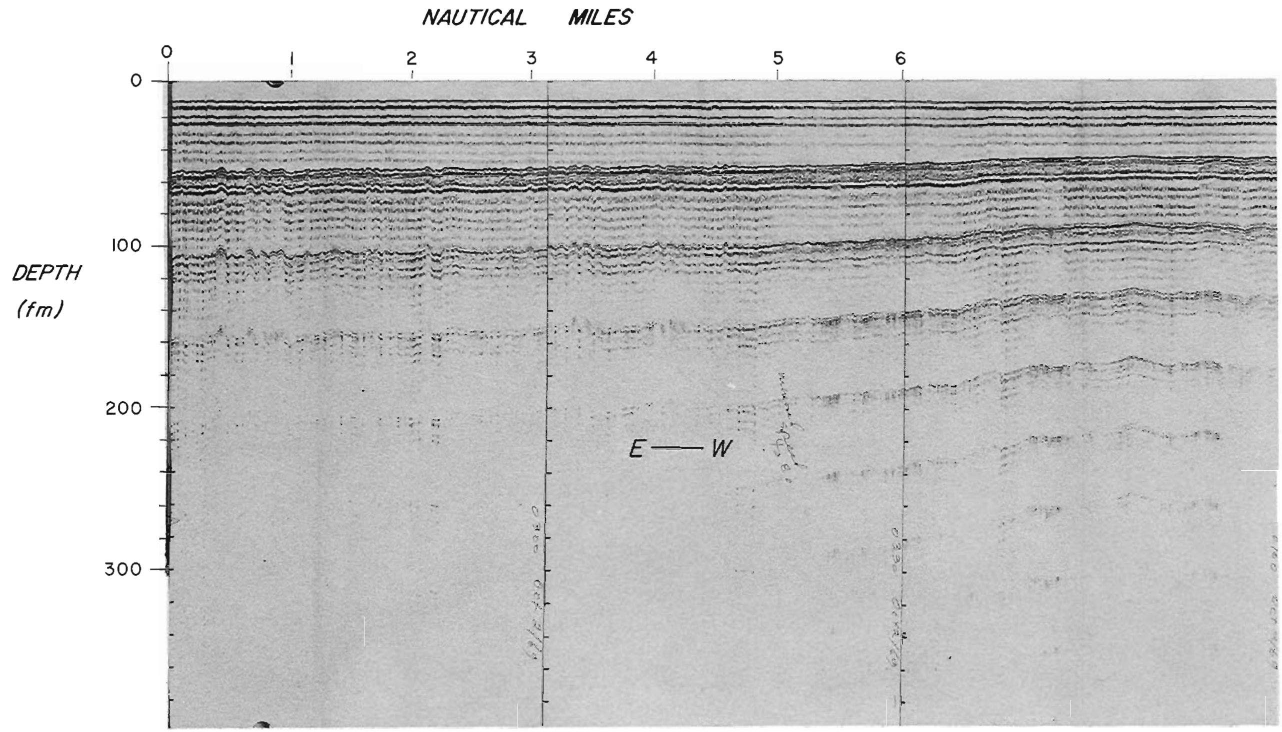

Figure 14. Seismic profiles showing contrast in surface morphology. Glacial deposits have either been removed or were never present at depths less than 50 fathoms (91 m). Bedrock consists in part of autochthonous unit and dips gently (< 1°) to the southeast.

features indicates two directions of ice recessions; one towards the east and another towards the north, respectively. This interpretation is consistent with and similar to that proposed by Prest and Grant (1969) for the southern part of the Gulf and the Maritime Provinces. The apparent absence of till in the deepest part of the study area is also consistent with the recessional ice picture described above; nevertheless, the apparent absence of till does not mean that continental ice was not active in this area.

Examination of Figures 10, 11, and 14 shows a distinct contrast in surface morphology between depths of 50 and 60 fathoms (90-110 m). The morphological break, especially as observed in the till unit in Figures 10 and 11 is indicative of the possible destruction of the original till surface at depths shallower than the break. Two main mechanisms to achieve this are 1) the reworking by strong bottom currents, and/or 2) the reworking by waves during a lower stand of the sea. Strong tidal currents (1-2 knots) are known to exist along the coast of Newfoundland and in the Strait of Belle Isle (Newfoundland Pilot, Canadian Hydrographic Service, 1966). It is not known whether the existing currents are strong enough at this depth and distance from the shore to rework the till unit observed. An alternative explanation is that a lower sea level existed in post-glacial times. The eustatic lowering of the sea during the last glacial maximum was thought to have been somewhere between 100 and 120 metres below the present datum (Milliman and Emery, 1968). Around the fringes and inside the areas covered by glacial ice, isostatic loading of the crust took place. This generally resulted in marine overlap at the time of deglaciation, which is known to have occurred along the shores of Quebec and west Newfoundland bordering the northern Gulf. Unless the behaviour of the crust was much different in the central Gulf area, it would have been impossible to have had a relatively lower sea level there while having a relatively higher one along the present coastline. Crustal warping of this magnitude can only have occurred if the flexural parameter of the crust, in this central Gulf area, as defined by Walcott (1970), was much less than is estimated (100 to 140 km).

ACKNOWLEDGMENTS

The author appreciates the opportunity to have worked with the crew of the CNAV SACKVILLE. The technical and physical assistance given by the Geological Survey of Canada seismic refraction crew on the CNAV SACKVILLE is also acknowledged. Mr. Vern Coady is to be commended for his dedication to the continuous seismic reflection equipment used on this cruise. Discussions with Mr. A. Overton and Dr. L.M. Cumming on the bedrock geology and Dr. D.R. Grant, Dr. Lewis H. King, Brian MacLean and Dr. R.I. Walcott on the post-glacial history of the northern Gulf of St. Lawrence were extremely helpful. The Canadian Hydrographic Service was very helpful in supplying the author with the latest bathymetric data and compilations for the study area. Jan Sklenar is to be thanked for the final drafting of all the maps and line drawings.

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19. MARINE GEOLOGY OF THE GULF OF ST. LAWRENCE

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Abstract

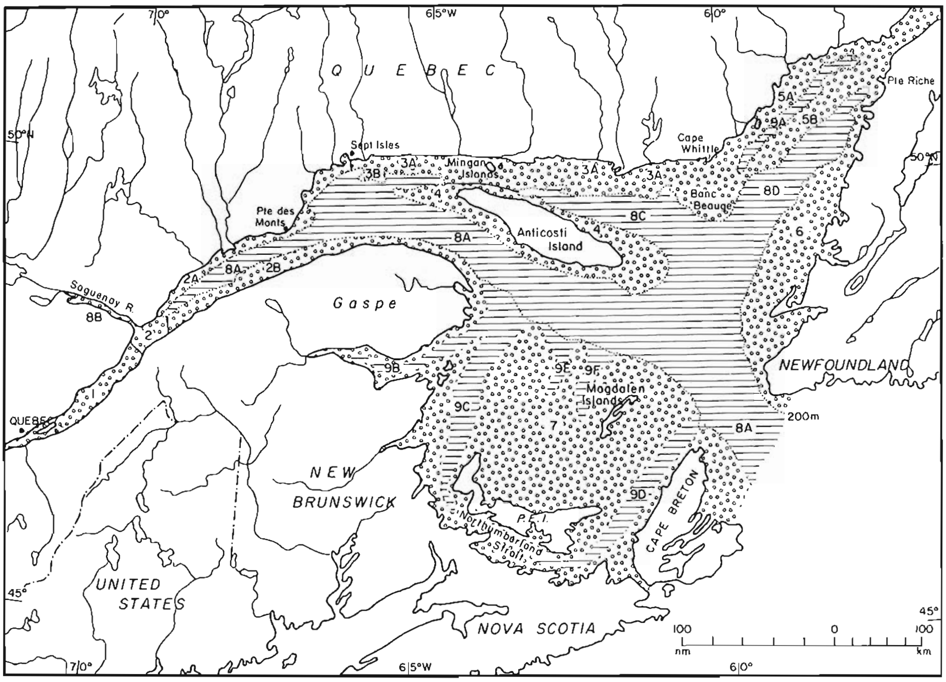
The Gulf of St. Lawrence is an inland sea of triangular shape which occupies an area of approximately 96,000 square miles (250,000 sq. km.). It has an irregular submarine topography composed of long trough-shaped valleys and shallow platforms or shelves of varying widths and relief. Data obtained from continuous seismic profiles, echograms, oblique sonargrams, underwater photographs, bottom and core samples indicate that the major geomorphological features are not related to the present environmental conditions. Instead, the submarine troughs and the adjacent shelves are pre-glacial erosional features developed in bedrock of differing structures (folded, unfolded, faulted), origins (sedimentary, metamorphic, and locally igneous), and ages (Precambrian to Permian).

The valleys were modified into their present form by glacial erosion strongly controlled by the pre-glacial topography and deposition during the Pleistocene Epoch, by post-glacial rises in sea level, and by recent marine sedimentation. The adjacent shelves are bedrock elevations which have a nearly continuous cover of glacial and post-glacial sediments. These sediments, like those in the troughs reflect the pattern of continental and local Pleistocene glaciations, post-glacial rises in sea level involving the reworking of glacial deposits, and the present depositional conditions which involve sorting, redistribution, and deposition in the area.

INTRODUCTION

From 1961 to 1968, a series of marine geology studies of the Gulf of St. Lawrence was made by the Fisheries Research Board of Canada. Most of the geological data collected and interpreted from these studies deals with the geomorphology, surface sediments, and recent depositional processes, but several studies have been made of the bedrock geology and the subsurface stratigraphy of the unconsolidated sediments. The present contribution summarizes the main features of the bedrock, glacial, and post-glacial geology, the details of which are currently being prepared for publication in collaboration with Dr. D.J.G. Nota as a Fisheries Research Board Bulletin. It contains new information on bedrock and glacial geology derived from the interpretation of continuous seismic profiles, as well as a synthesis of previously published data on the geomorphology, surficial sediments and recent depositional conditions.

The interpretations in this paper are based on bottom grab and core samples, echograms, continuous seismic profiles obtained using a Bolt Associates Air Gun (Model 600A), side-scan sonargrams, and underwater



Shelves: (water depth <200 m):

1.	St. Lawrence River floor;	4.	Anticosti Shelf;
2.	St. Lawrence Estuary shelves;	5.	Quebec-Labrador Shelf;
2A.	Les Ecoumains Shelf;	5A.	Inner Shelf;
2B.	Gaspé Shelf;	5B.	Outer Shelf;
3.	North Shore Shelf;	6.	Newfoundland Shelf;
3A.	Inner Shelf;	7.	Magdalen Shelf;
3B.	Outer Shelf;		

Submarine troughs (water depths >200 m):

8.	Laurentian Trough system	9A.	Mécatina Trough;
8A.	Laurentian Trough;	9B.	Chaleur Trough;
8B.	Saguenay Fjord;	9C.	Shediac Trough;
8C.	Anticosti Trough;	9D.	Cape Breton Trough;
8D.	Esquiman Trough	9E.	Western Bradelle Trough;
9.	Shelf valleys (water depth <200 m);	9F.	Eastern Bradelle Trough;

Figure 1. Physiographic divisions of the River and Gulf of St. Lawrence.

photographs. These data were obtained during cruises on board CNAV SACKVILLE (1961: see cruise reports S-56; 1962: S-62; 1963: S-75; 1964: S-79; 1965: BIO-27-65), CSS KAPUSKASING (1966: BIO-13-66, BIO-09-66; 1968), CSS HUDSON (1967: BIO-24-67), and CNAV BLUETHROAT (1967: BIO-20-67).

SUBMARINE TOPOGRAPHY OF THE GULF OF ST. LAWRENCE

The Gulf of St. Lawrence is an inland sea of somewhat triangular shape, having an area of approximately 96,000 square miles (250,000 sq. km.). It has an irregular submarine topography composed of long trough-shaped valleys and shallow platforms or shelves of varying widths and local relief (Fig. 1). A more detailed discussion of the geomorphology of these features is given by Nota and Loring (1964), Loring and Nota (1966), and Loring et al., (1970), and only a few comments of a general nature are given here.

Submarine Valleys

Long, deep, trough-shaped submarine valleys with water depths between 200 and 500 m form the most conspicuous features of the submarine topography (Fig. 2). The largest of these is the Laurentian Channel or Trough, 200 to 500 m in depth, which is about 640 nautical miles (1186 km) long and form 20 (37 km) to 48 nautical miles (89 km) in width. It extends from a position off the mouth of the Saguenay River in the St. Lawrence through the Gulf and across the continental shelf to the shelf edge. Southeast of Anticosti Island this main trough is joined discordantly by two others: the Anticosti Trough or the "Chenal d'Anticosti" (CHS Chart 801), 200-300 m in depth, which enters from the northwest; and the Esquiman Channel or Trough, 200-300 m deep, which enters from the northeast. The main features of the surface morphology of these troughs are their relatively straight, steep sides and their broad undulating floors which contain a number of large elongated depressions as much as 100 metres deep.

Submarine Shelves

The submarine platforms or shelves adjacent to the troughs, with water depths less than 200 m, vary in width and relief (Fig. 1). These are, clockwise around the Gulf: (1) the North Shore Shelf which extends from Sept. Iles to Cape Whittle, (2) Anticosti Shelf adjacent to the Island, (3) the Quebec-Labrador Shelf which extends from Cape Whittle to the Strait of Belle Isle, (4) the Newfoundland Shelf, which lies along the west coast of Newfoundland, and (5) the Magdalen Shelf which occupies the southern embayment of the Gulf. The first two shelves are narrow and are characterized by a narrow rough inner shelf adjacent to the coast of Quebec and Labrador; the inner shelves are separated, in part, from wide outer shelves by longitudinal shelf valleys. The Magdalen Shelf is the largest shelf with an area of about 30,000 square miles (78,000 sq. km.). It has a rather distinct topography composed of long but fairly shallow (water depth from 10-200 m) shelf valleys along its western and eastern margins as well as smaller ones in the central shelf area, and elevated areas or banks and smaller platforms in the central shelf (Loring and Nota, 1966).

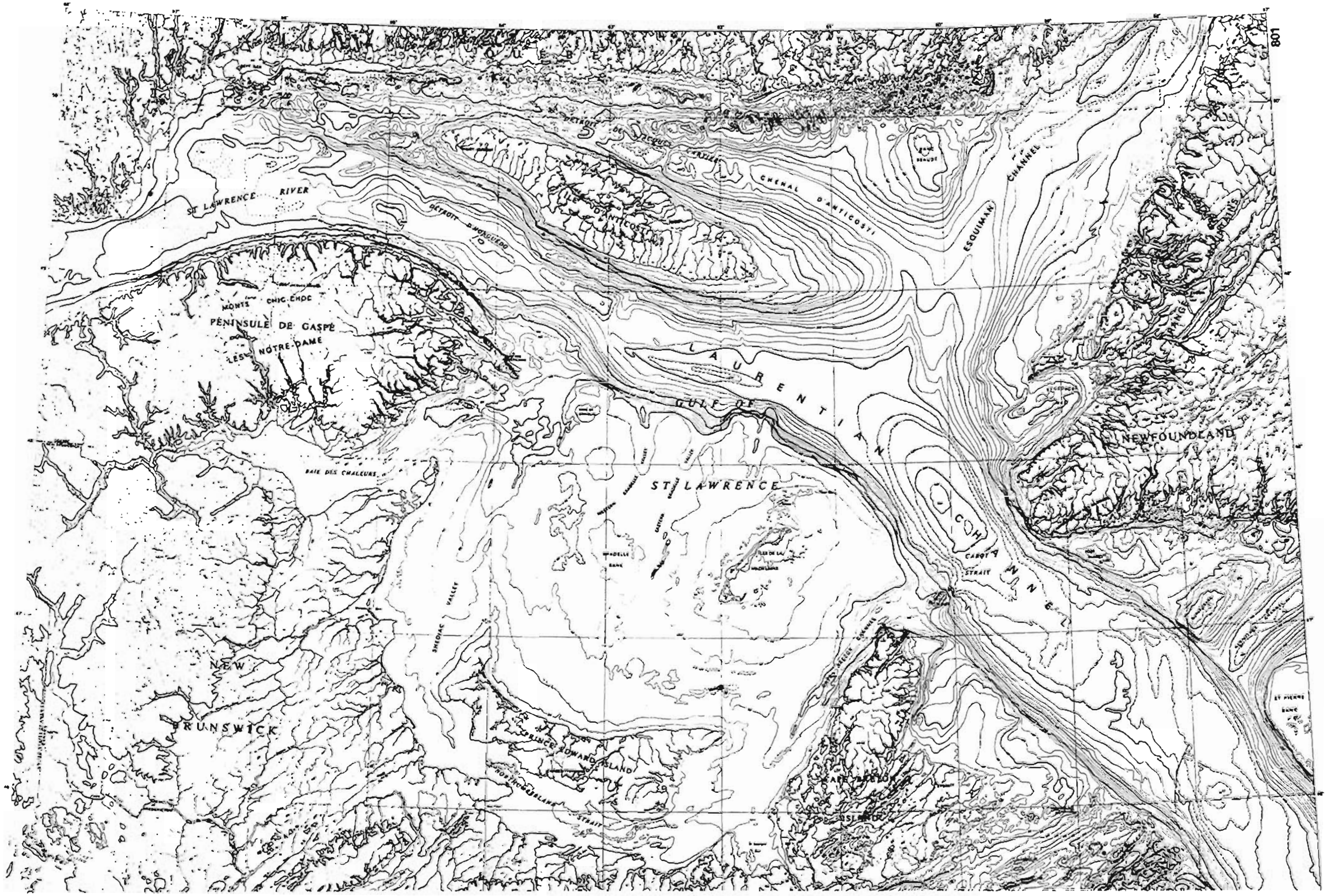


Figure 2. Bathymetry of the Gulf of St. Lawrence (after CHS Chart 801).

Although each of the troughs and shelves has some variety in large and small scale details of form, their surface morphological characteristics have been recognized as those representing submerged pre-glacial bedrock valleys and lowlands. These have been modified by glacial erosion and deposition, by post-glacial rises in sea level, and later by recent depositional conditions (Spencer, 1903; Johnson, 1925; Shepard, 1931; Nota and Loring, 1964; Loring and Nota, 1966, 1969, Loring et al., 1970; King and MacLean, 1970).

BEDROCK GEOLOGY OF THE GULF OF ST. LAWRENCE

Geological Map

A simplified preliminary map of the bedrock geology of the sea floor is shown in Figure 3. This map is based on sampling data (bottom grabs and cores), underwater photographs, and the interpretation of echo-sounding and continuous-seismic records. Also used in the preparation of this map was the data on the adjacent land geology (Stockwell et al., Poole et al., 1970; Poole, 1967) as well as seismic refraction data from Sheridan and Drake (1968).

The Precambrian unit includes all the crystalline rocks that outcrop along the southeastern edge of the Shield and form the basement. These rocks occur offshore along the inner part of the Quebec-Labrador shelf. Elsewhere, except in the very near-shore waters (unmapped) along the north shore of the St. Lawrence River and Gulf, they are overlapped to the southeast by Paleozoic sedimentary rocks.

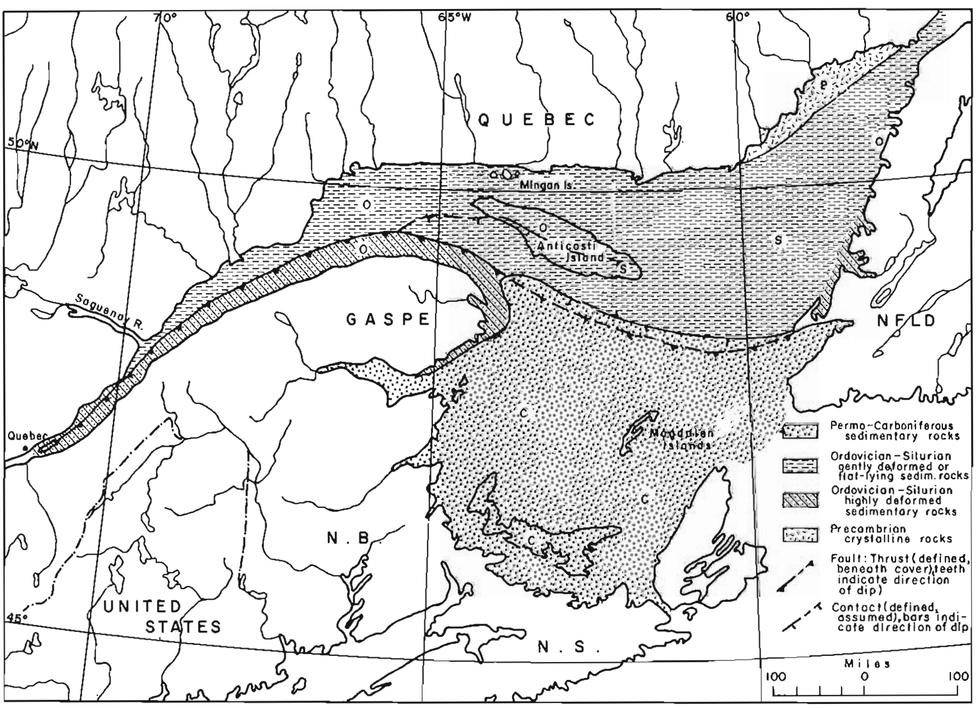


Figure 3. Simplified bedrock geology of the Gulf of St. Lawrence.

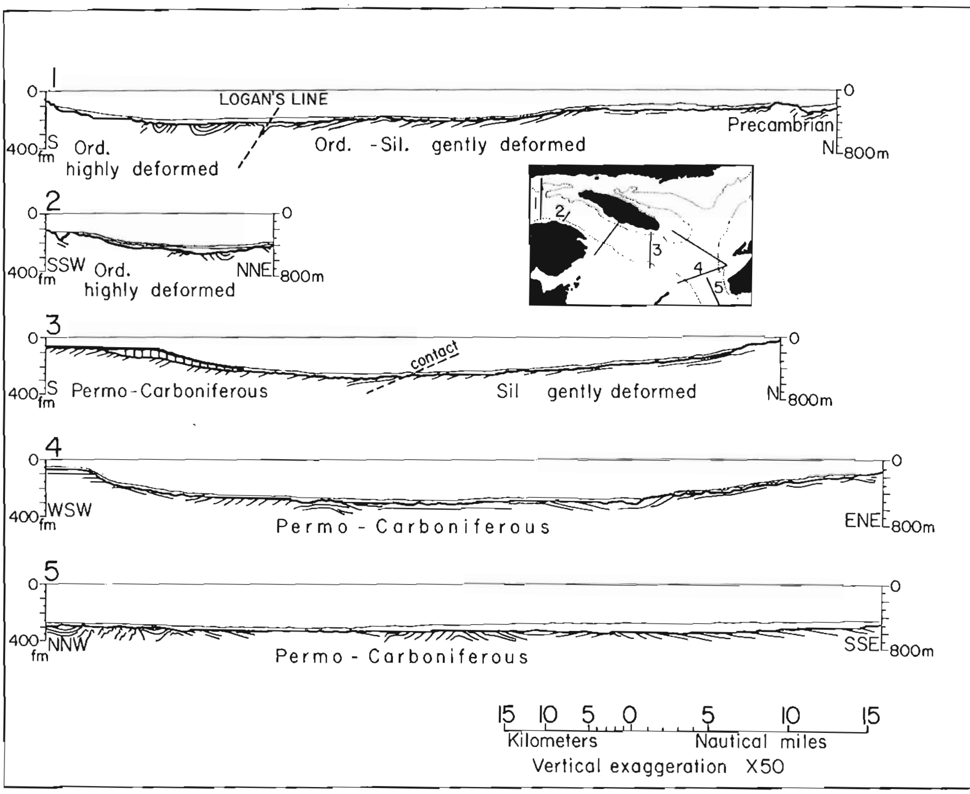


Figure 4. Geological Sections of the Laurentian Trough. The location of the sections is shown in the insert map. Blank area above the bedrock surface (heavy line) indicates the thickness of glacial and post-glacial sediments. Vertical lines in Section 3 refer to glacio-marine-sediments and solid black to glacial till.

Ordovician to Silurian rocks of the St. Lawrence Platform (Poole, 1967) occupy the northern part of the River and Gulf. These rocks consist mainly of limestones, calcareous, shales and sandstones and dip gently to the south and east. They are exposed above water on the Mingan Islands where they are Lower to Middle-Ordovician in age and on Anticosti Island where they form a Middle-Ordovician to Upper-Silurian conformable sequence (Twenhofel, 1927, 1938; Bolton, 1961). Along the north shore of the Gulf from the mouth of the Saguenay to Cape Whittle, and along the line of the Mécatina Trough (9A on Fig. 1) they fault against or overlap the Precambrain crystalline rocks. In the southern part of the St. Lawrence River, these rocks are most likely of Middle Ordovician age and fault against highly-folded Ordovician rocks which make up the coastal exposures along the north coast of the Gaspé Peninsula. The fault zone which separates them is believed to be the eroded edge of the Appalachian Front or "Logan's Line". In the subsurface profiles (Fig. 4, sections 1 and 2) these strata and the fault zone seem to disappear beneath a cover of younger sediments as they are traced southeastward down the Laurentian Trough.

South of Anticosti Island, the northern part of the Laurentian Trough is apparently occupied by Silurian strata which are overlapped near the centre by younger sediments. In the northeastern part of the Gulf, however, the data indicate that the middle Paleozoic strata (mostly Silurian) are warped into a synclinal structure, the axis of which is parallel to and east of the centre of the Esquiman Trough. Along the inshore part of the Newfoundland Shelf these rocks fault against or overlap Ordovician or older strata exposed along the coast.

Permo-Carboniferous rocks comprising red sandstones and shales occupy the southern part of the Laurentian Trough, the southeast corner of the Gulf including Cabot Strait and most of the Magdalen Shelf. To the north they appear to overlap the Silurian strata without any major dislocation.

No younger rocks have been identified on the sea floor of the St. Lawrence River and Gulf, although others (King and MacLean, 1970) have postulated a former occurrence of Tertiary rocks within this region.

Geological Sections of the Laurentian Trough

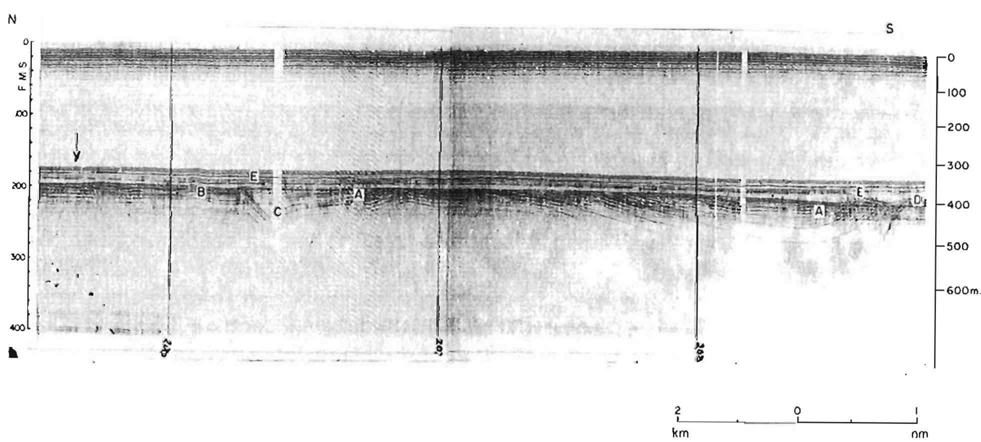
One longitudinal section and four transverse sections of the Laurentian Trough from the St. Lawrence Estuary are shown in Figure 4. These sections which are five of eleven profiles obtained in this region are based on continuous seismic profiles. The record of the upper part of the section that is obscured by the bubble pulse of the seismic display was resolved from echograms obtained simultaneously with the seismic records. The nature of the surface and near-surface sediments was determined by bottom and core sampling. The seismic information on the records was first interpreted visually and then reduced with the aid of a pantograph with independently variable horizontal and vertical scales. The sections in Figure 4 have a vertical exaggeration of about 50 and the sub-bottom depths are uncorrected for variations in sound velocity.

Section 1: This is a northerly transect of the St. Lawrence Estuary from the Gaspé coast to the north shore of the Gulf. The southern wall and part of the uneven bedrock floor is formed by folded strata presumably, as extrapolated from the coastal geology sedimentary rocks of Ordovician (Quebec group) age (Skidmore, 1967). These rocks fault against gently, southerly-dipping strata on the floor about 16 nautical miles (30 km) from the Gaspé coast. The relatively undeformed strata are probably of Middle-Ordovician age (Trenton) similar to the rocks found as smaller outliers along the Estuary (Faessler, 1942). The fault contact between these rocks represents the eroded submerged edge of the Appalachian Front or Logan's Line, details of which are shown by the photograph of the actual record across it (Fig. 5).

North of the fault, the bedrock surface conforms to the shape of the trough floor and slope. Above the trough-shelf break the strata form the outer part of the North Shore Shelf, but near the coast, along the inner shelf, they are pierced by an inlier of Precambrian rocks which they overlap on the north shore. The slightly uneven and perhaps overdeepened bedrock surface is overlain by unconsolidated glacial and post-glacial sediments which are up to 40 m thick where they fill depressions in the bedrock surface.

Section 2: This is a short (15 nautical miles or 28 km) transverse profile from the Gaspé Peninsula towards Anticosti Island. It shows that, as in Section 1, the folded Ordovician rocks form the southern wall and part of the trough floor. Strata at the end of the section are more gently deformed and are most likely of Silurian age (not marked on profile). The bedrock surface eroded across these rocks is very uneven and overdeepened. Near the Gaspé coast the bedrock is incised by a V-shaped channel which is partly filled with unconsolidated sediments. Although the origin of this channel is not apparent, it may represent a glacial overflow channel which was cut into the bedrock when ice masses partly filled the trough. Glacial and post-glacial sediments up to 80 m thick fill the irregularities in the bedrock surface in the centre of the trough.

Section 3: This is another cross-section of the trough between the Magdalen Shelf and the Anticosti Shelf. The edge of the Magdalen Shelf, the southern wall and part of the bedrock floor of the trough is formed by gently, southerly-dipping sedimentary rocks of Permo-Carboniferous age, most likely red sandstones and shales. Near the centre, these rocks appear to overlap southerly-dipping strata of Silurian age which would presumably be similar to those exposed on Anticosti Island. There is no evidence of a distinct structural break as seen in Section 1 (Fig. 4). Presumably the southerly extension of Logan's Line is covered by the Permo-Carboniferous sediments. The unevenly-eroded bedrock surface truncates the strata along the southern wall and floor of the trough but is nearly conformable to the dip slope adjacent to Anticosti Island. It is covered for the most part by a thin (<40 m) layer of unconsolidated sediments, but the southern wall contains a thick wedge of sediment which is almost 130 m thick. Although this wedge resembles bodies of postulated Tertiary material along the south side of the trough east of Cape Breton Island described by King and MacLean (1970) sampling data indicate that at least part of this material is of glacial and post-glacial origin (see Glacial Geology).



(A) folded Ordovician strata;
(B) gently deformed Ordovician-Silurian strata;
(C) Logan's Line or the eroded edge of the Appalachian Front, and
(E) glacial and post-glacial sediments resting on a "planed" bedrock surface

Figure 5. North-south seismic record about 7 nautical miles (13 km) in length from Section 1, Fig. 4.

Section 4: This section which is about 53 nautical miles (99 km) in length, illustrates the sub-surface structure between the Magdalen Islands and the Newfoundland Shelf. The edge of the Magdalen Shelf, the trough slopes, the trough floor, and the edge of the Newfoundland Shelf are apparently underlain by stratified sedimentary rocks of Permo-Carboniferous age. On the southern wall and part of the trough floor the strata are nearly flat-lying or dip very gently, whereas in the centre of the floor they are flat-lying with minor undulations. The smooth bedrock surface truncates the strata on the southern wall and part of the trough floor, whereas on the northern side it nearly parallels the dip of the strata. Here, bedrock is covered by a thin (<40 m) layer of unconsolidated glacial and post-glacial sediments.

Section 5: This is a long section down the central part of the trough through Cabot Strait. The undulating bedrock surface is formed by gently folded rocks most likely of Permo-Carboniferous age, except for one small area of strongly folded strata 5 nautical miles (9.3 km) from the northwest end of the section. This strongly folded strata might represent an outcrop of the Precambrian rocks similar to those found on St. Paul Island off the northern tip of Cape Breton Island. The bedrock surface is uneven and appears to be overdeepened at points along the section. It is covered by up to 100 m of glacial and post-glacial sediments. These overdeepened areas of the bedrock usually coincide with elongated floor depressions outlined on the bathymetry chart (see Fig. 2). They suggest that glacial erosion has also been involved in the development of the bedrock surfaces of the Laurentian Trough.

Pre-Glacial Erosional History

The acoustical data indicate the dominant features of the bedrock topography of the St. Lawrence River and Gulf are a system of long, deep erosional valleys (the Laurentian Trough system) and shallower shelf valleys of a pre-glacial drainage system (Nota and Loring, 1969). These valleys can be traced to the now partly drowned river valleys which occur along the coast. The most prominent of these valleys is that of the St. Lawrence River. Since early Cenozoic times this valley system as we now know it as well as the attendant lowland drainage system developed along lines of structural and lithological weaknesses by a process of sequential fluvial erosion. These erosive processes have resulted in the last stages in the stripping of most of the Paleozoic cover rocks from the Shield, the development of the pre-glacial St. Lawrence drainage system extending from Cabot Strait to the interior, and the development of a broad cuesta-like landscape on the Paleozoic sedimentary rocks of the northern Gulf as well as on the structure of the submarine extension of the Carboniferous lowlands occupying the southern part of the Gulf. These features evolved with only minor interruptions with the latest period of valley entrenchment probably occurring in Pliocene times or early Pleistocene times as evidenced by the suspected presence of eroded late Tertiary strata along the trough at its mouth near the edge of the continental shelf (King and MacLean, 1970) and the erosion of late Tertiary strata on the outer part of the Grand Banks (Barlett and Smith, 1971).

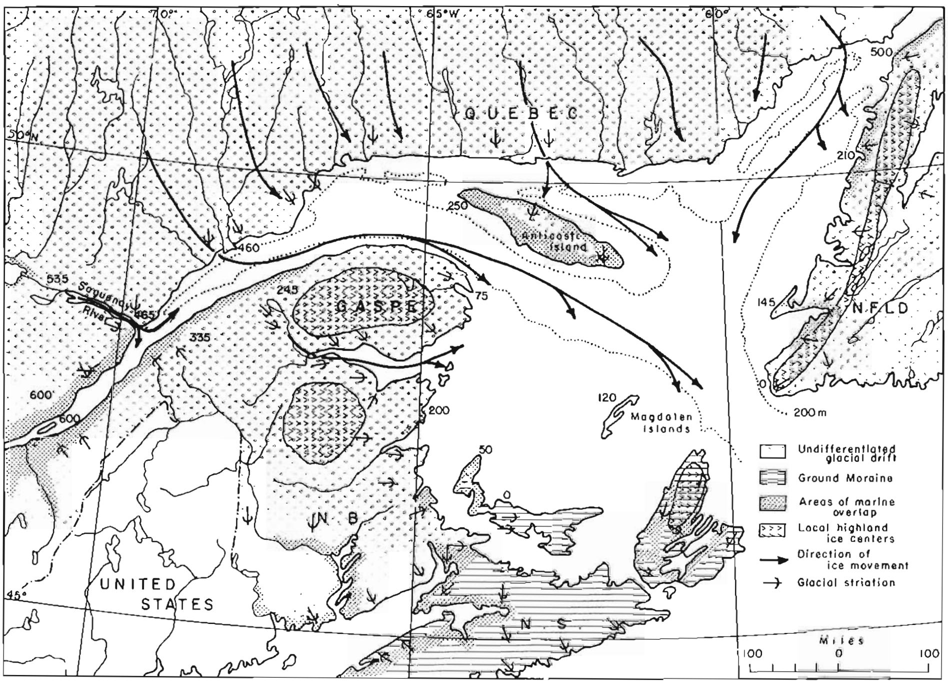


Figure 6. Glacial geology of the adjacent land areas and the direction of the major ice movements within the Gulf of St. Lawrence. Numbers refer to the amount of marine overlap in feet. (Glacial geology modified from G.S.C. map 1253A).

Glacial Geology

The glacial geology of the surrounding land areas (Dresser and Denis, 1944; Prest and Grant, 1969; Prest, 1970) and the submarine morphology and unconsolidated sediments provide ample evidence for the presence of continental ice sheets in the Gulf during the Pleistocene as well as for ice advances from local centres during the retreat stages of the continental ice mass.

Glaciation of the Submarine Troughs and Adjacent Shelves

The evidence of ice movement illustrated by the glacial map of the surrounding land area indicates that the submarine valleys were probably at one time or another invaded by three main ice lobes which later coalesced when the individual troughs were overridden, but separated again when the ice withdrew from the region (Fig. 6). The main ice flow appears to have entered the Laurentian Trough from the north side, from whence one portion flowed down the Laurentian Trough and eventually out through Cabot Strait. In this advance, the ice also overrode parts of the Magdalen Shelf. Part of the main lobe also appears to have overridden the southeastern side of the Trough in the Estuary and eventually flowed down the Matepedia Valley into Chaleur Bay, beyond which it became confluent with ice occupying the

Laurentian Trough (Dresser and Denis, 1944). A second major ice flow must also have moved eastward through the Mingan Trough north of Anticosti Island. The Island, lying across the flow of the ice coming from the north, probably formed a barrier for the basal ice and deflected it in an eastward direction before the Island itself was overridden (Bolton and Lee, 1960). A third lobe from the highlands of Labrador entered the Esquiman Trough and modified its topography. As this ice pushed south and west it came confluent with the Mingan lobe from the north and west and with the Laurentian lobe at the southern end of the Newfoundland Shelf. At its maximum extent I believe that Wisconsin ice covered the whole Gulf. The ice apparently began to wane between 15 and 14.5 thousand years B.P., with the ice front withdrawing up the Laurentian Trough from Cabot Strait and into the tributary valleys in a succession of deepwater re-entrants as the sea invaded the Gulf (Prest, 1969).

Glacial Erosion

The acoustical data indicate that erosion by these ice masses, strongly controlled by the pre-glacial topography resulted in the widening, deepening and straightening of valley walls as far to the southeast as the mouth of the Laurentian Trough (see Figs. 2 and 4). In the narrow parts of the valleys such as the Détroit d'Honguédo (Gaspé Passage) and Cabot Strait (Fig. 2), increased erosion caused by the more rapid flow required for ice to pass through these narrows probably accounts for the overdeepening of the bedrock surface in these areas (Fig. 4, Profiles 2 and 5).

Glacial Deposition

Uneven deposition of glacial till from the ice unevenly charged with debris and detritus, and local deposition from streams that flowed in or under the ice has resulted in the production of an irregular glacial landscape over most of the bedrock surfaces. Local fluvio-glacial activity began to truncate the glacial topography at the end of the Pleistocene period when the ice began to melt and flotation occurred.

The continuous seismic profiles indicate that for the most part the glacial deposits constitute relatively thin ground moraine and glaciofluvial deposits. These vary in thickness from 5 to 30 m except where they fill depressions in the bedrock surface and may be as thick as 80 m (Fig. 4). Echograms also indicate that the ground moraine in its modified state on the slopes and floors of the troughs, has a rough irregular surface composed of hummocks and depressions.

Bottom grab and core-sampling indicates that the till-like sediments which mantle the bedrock vary in texture and lithological characteristics depending upon the parent rocks from which they were derived and their subsequent depositional history.

Figure 7 shows the characteristics of the tills that now partly or completely cover the bedrock surface in the region. These are not necessarily listed in order of age. From this map it may be seen that:

(1) gray, very sandy, pelitic till occupies the river estuary, and the near-shore areas adjacent to the north shore of the Gulf. The material in this till is mainly derived from the crystalline Shield rock;

(2) gray calcareous sandy to very sandy till covers the bedrock surface around Anticosti Island and in the northeastern part of the Gulf. This till is mainly derived from the calcareous Paleozoic rock which underlies the area but the tills also contain crystalline material derived from the Shield;

(3) reddish-brown calcareous very sandy pelitic to sandy pelitic till occurs in the Laurentian Trough, south and east of Anticosti Island. Part of this is moraine and part is of glacial marine origin. It contains a mixture of Permo-Carboniferous, Paleozoic and Precambrian crystalline material.

The Magdalen Shelf to the south has a thin but nearly continuous cover of reddish-brown gravelly pelitic sandy till. This material is derived from the Permo-Carboniferous rocks underlying the shelf, but it also contains small but significant amounts of material from the northern part of the Gulf (Nota, 1968).

The presence of mineral and rock fragments from the Shield area throughout the Magdalen Shelf to the south, indicates that at one stage Laurentian ice invaded the pre-glacial shelf valleys and spilled over the remainder of the shelf. At times part of this ice may have been effectively blocked from some areas of this shelf by ice moving outwards from highland centres on the Gaspé Peninsula and the New Brunswick highlands (Alcock, 1936; McGerrigle, 1952; Prest, 1970).

Glacio-Marine Deposits

The acoustical and sampling data also indicate that a thick wedge of unconsolidated sediments occurs between the 110 m and 440 m contour along the southern side of the Laurentian Trough for a distance of 85 nautical miles

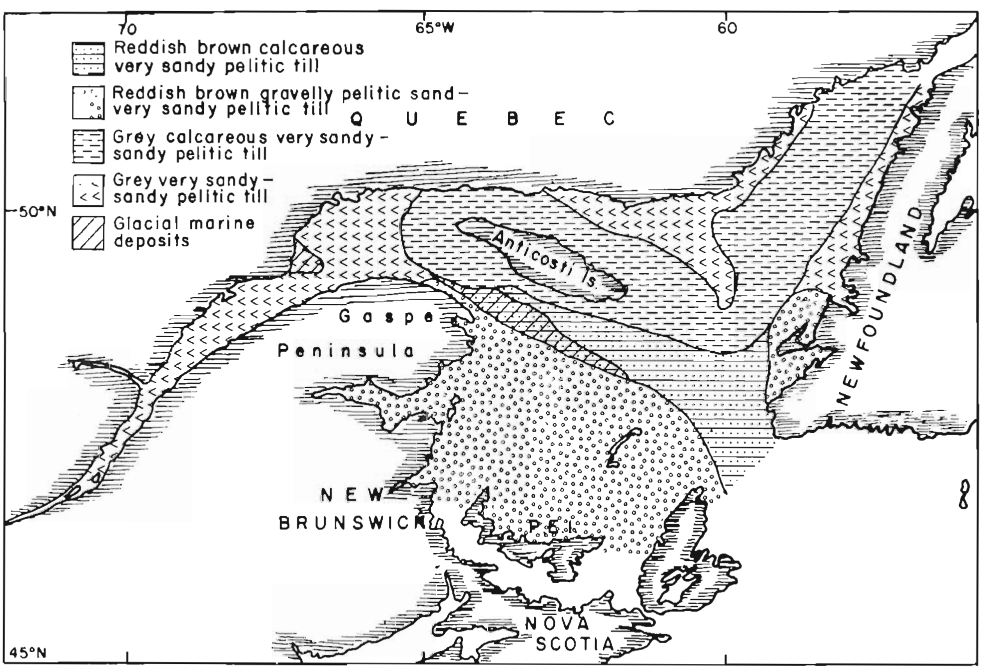


Figure 7. Distribution of till-like sediments in the Gulf of St. Lawrence.

(157 km) between the tip of the Gaspé Peninsula and the mouth of the Eastern Bradelle Valley (Fig. 2).

Near the tip of the Gaspé Peninsula this deposit forms a submarine spur or fan sloping downward from the shelf/slope break at about 110 m to the trough floor at 420 m. Off the mouth of the Eastern Bradelle Trough it is less extensive, but still consists of a thick wedge of unconsolidated sediments resting on an irregular sloping bedrock surface (Fig. 8). In form, it consists of three topographic elements:

(1) a gently sloping terrace surface about 5 nautical miles (9.3 km) wide between the 110 and 132 m level;

(2) a more steeply inclined rather straight face about 5 nautical miles (9.3 km) long between the 140 m and 380 m contour, and

(3) a gently undulating sheet at the toe of the slope which ends in a narrow snout at a depth of 420 m.

Beneath and within this deposit there are four subsurface reflectors:

(1) a sloping terraced bedrock surface eroded across stratified sedimentary rocks of Permo-Carboniferous age (A in Fig. 8) which underlies the deposit;

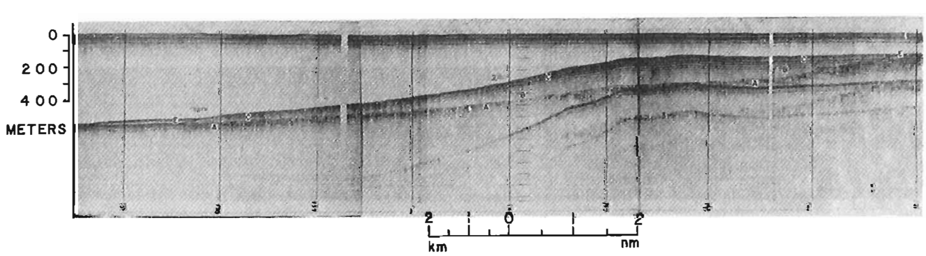
(2) a thick body of acoustically-homogeneous sediments which vary in apparent thickness from 10 to 100 m (B in Fig. 8);

(3) a rather variable (15-30 m) but continuous rough-textured layer which overlies the homogeneous sediments and extends nearly to the surface (C in Fig. 8), and

(4) an acoustically-transparent sediment that covers the snout of this deposit (E in Fig. 8).

Core-sampling data from the inshore edge of this deposit indicates that the top layer, beneath a cover of recent marine sediments, is composed of reddish-brown hard sediments having the textural and lithological characteristics of glacial drift deposited in a marginal marine environment. The sediments beneath this layer are gray, very sandy pelites and appear to be of glacial marine origin.

Deposition of the thick layer of glacial marine sediments along the shelf edge and in the trough relates to the late glacial stages in the Gulf when open water and/or floating ice occurred between the shelf and Anticosti Island



(A) eroded bedrock surface; (C) reddish brown sandy till, and
(B) homogeneous glacio-marine sediments; (E) recent marine sediments

Figure 8. North-south seismic record about 13 nautical miles (24 km) in length (Section 3, Fig. 4).

(<13,000 years B.P., Prest, 1969). At this stage, the ice to the south was grounded along the shelf edge, whereas to the north ice stood in an asymmetrically arced front over the slope of the trough adjacent to Anticosti Island. To the northwest, ice had probably withdrawn up the Laurentian Trough at least to the vicinity of the Saguenay River (~ 12.0 x 103 yrs. B.P., Prest, 1969). Apparently, these glacial marine sediments resulted from the uneven deposition of detritus and debris washed out of the ice fronts and from the ice floating over the troughs. The textural and foraminiferal characteristics of the sediments indicate that the salinities were low at this time, most likely because of large quantities of melt water from the surrounding ice. The thick homogeneous deposits built out over the shelf edge indicate periods of rapid sedimentation in that area. Sea level at this time probably stood at not less than 100 m below present along the shelf edge, the level which deposits now reach about 10,000 yrs. B.P. (*see* carbon date below).

The presence of a layer of reddish-brown till composed of shelf material, over the deposit of glacial marine sediments indicates that the ice standing to the south readvanced over the shelf edge and into the trough reaching almost to the northern slope of Anticosti Island. The presence of marine shell fragments within these sediments indicates that the advance was into the sea. A Carbon^{14} date (G.S.C. 1528) of 10,220 $\pm$440 years B.P. from shell material beneath the till indicates glacio-marine deposition lasted until at least 10,000 years B.P. and that ice readvanced after this period. These readvances may also coincide with the deposition in the deep waters of Cabot Strait of thin brick-red layers of glacial marine sediments which were discovered by Connolly *et al.*, (1967). The sequence of drift and interbedded marine sediments is similar to that found by other workers (Brooks, 1969) on the west coast of Newfoundland. These deposits indicate that local ice readvanced during a period of rising sea level between 13.2 and 12.6 x 10^3 years B.P.

After deposition of the drift layer (<10,000 yrs. B.P.), the ice apparently dissipated rapidly and with drew to the southwest and from other parts of the Gulf.

Post-Glacial Marine Submergence

Parts of the coastal area adjacent to the St. Lawrence River and Gulf were submerged beneath the sea as the deglaciation of the region progressed between 14.5 to 12.5 x 10^3 years B.P. (Elson, 1969). The main cause of this submergence was the isostatic depression of the crust that resulted from the load of the ice, but eustatic rise in sea level was also a contributing factor. Evidence from abandoned coastal strand lines indicates that the amounts of marine overlap varied from nil at the eastern end of Prince Edward Island and the entrance to Cabot Strait to about 76 m (250 ft) on Anticosti Island (Bolton and Lee, 1960), and up to 183 m (600 ft) in the St. Lawrence River and Estuary (Elson, *op. cit.*). As the ice withdrew, and the rebound became greater than the eustatic rise in sea level, these coastal areas re-emerged (Fig. 6).

Acoustical and sampling data from the edge of the shelves adjacent to the trough indicate that above the 110-120 m level the drift surface is almost completely modified. The drift has been reworked into various types of gravels and sands, and wide terraces have formed by marine processes. The transition from unmodified to modified drift surfaces is believed to represent the lowest stand of sea level in the region. Other (higher) sea level standstills during the post-glacial transgression are also recognized by the occurrence of submarine terraces at about the 55 to 62 m level adjacent to the Magdalen Islands and in the coastal waters of Prince Edward Island (Loring and Nota, 1966) as well as the higher levels in the Northumberland Strait (Kranck, AOL, personal communication) which are believed to range in age from 13,000 years B.P. to 5,000 years B.P.

Surface Sediments and Recent Depositional Conditions

The distribution and composition (textural, mineralogical and chemical) of the surface sediments has been determined for the whole of this region (CHS Chart 811G). This map is based on the analysis of about 2,000 bottom grabs from various parts of the St. Lawrence River and Gulf, sediment cores, and echograms representing thousands of miles of sounding lines.

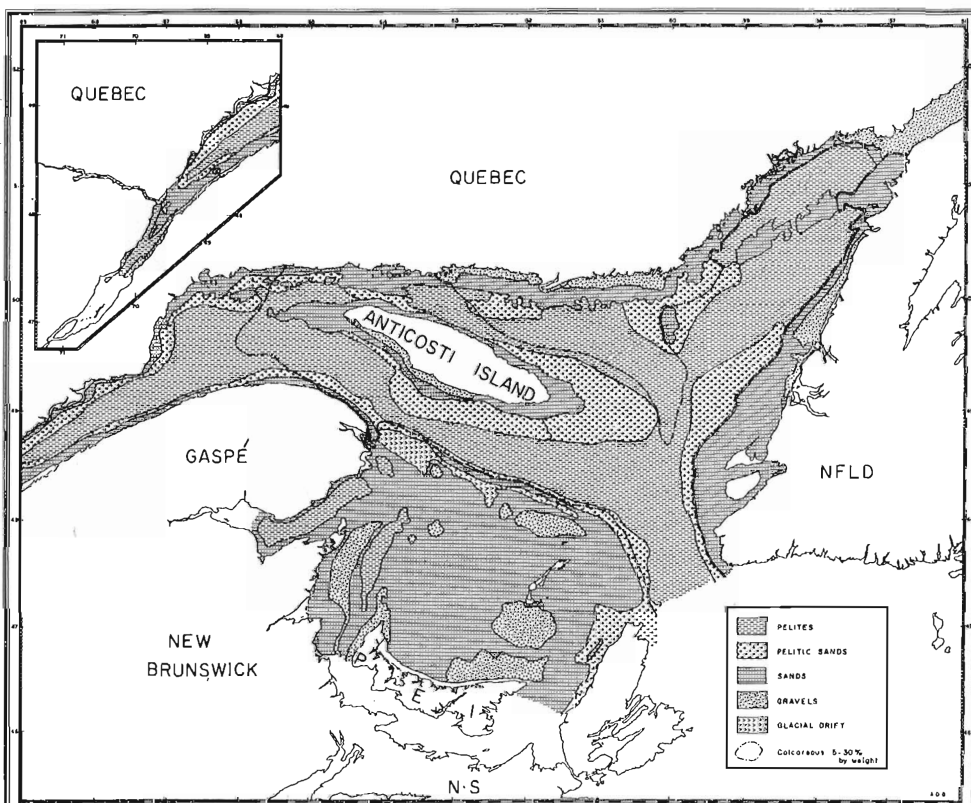


Figure 9. Simplified surface sediment map of the Gulf of St. Lawrence (after CHS Chart 811 G).

A simplified distribution of the surface sediments is shown in Figure 9. The nomenclatural system is modified from that of Nota (1958) in which the critical limit for the different size grades are 5 per cent and 30 per cent by weight. Components representing less than 5 per cent by weight of the total sediment are not considered. Since calcareous debris and detritus from an important part of the central and northeastern Gulf sediments, the area enclosed by the 5 per cent iso-concentration line of $CaCO_3$ is also indicated in Figure 9.

Fine-grained sediments referred to as pelites (material < 0.05 mm) or calcipelites (if containing >5 per cent $CaCO_3$) cover the deep central parts of the major troughs as well as some of the shelf valleys, and vary in thickness from 3 to 50 m. The thickest deposits occur in the floor depressions and the thinnest deposits occur over the rough-surfaced floor rises which are found between these depressions and adjacent to the sides of the troughs. These sediments, which are the last addition of sediments to the area (Holocene), represent deposition of recently supplied suspended material derived from the St. Lawrence drainage system and reworked from older glacial deposits. They also contain a small percentage (<10 per cent) of sand and gravel size material which is supplied annually by ice rafting from the adjacent land areas. The deposition of the pelites and calcipelites has partly modified the glacial landscape beneath it.

The pelites lap onto coarser-grained sediments referred to as pelitic sands and their calcareous equivalents that occur on the lower and middle slopes of the troughs. These sediments are partly composed of coarse-grained glacial sediments, on which they form a thin veneer, and fine material (pelite) supplied from suspension. Their distribution coincides in part with the rough morainal topography on the trough slopes. Below 110 m, small exposures of glacial drift, which has been described elsewhere, occur along the edge of the Magdalen Shelf.

Various types of sands and gravels and their calcareous equivalents (calcarenites, calcirudites) occupy the shelves above the 110 m level, which marks the lowest sea-level stand. These sediments represent reworked, resorted, and redistributed glacial deposits. Mineralogical and chemical analysis of the sands from the Gulf and adjacent shorelines indicate that sands from the northern part of the Gulf are derived from the igneous and metamorphic rocks of the Canadian Shield and locally from the carbonate rock on the sea floor and Anticosti Island by glacial and marine erosion (Nota and Loring, 1964). In the southern Gulf, sands are mainly derived from the underlying sandstone bedrock, although these contain small but significant amounts of crystalline material from the Shield. The dispersal pattern of the sands throughout the Gulf reflects the pattern of Pleistocene glaciation in the region (Loring and Nota, 1969).

At present the material on the shelves is being reworked in response to the present current regime so that a variety of microsedimentary environments, ranging from depositional to active erosional and transport environments, occur in different locations on the shelves (Loring et al., 1970).

In general, on exposed areas of positive relief such as the banks, the drift material has been reworked, sorted, and redistributed to form lag gravel deposits and elsewhere fields of sand waves. In the shelf valleys and to a lesser extent in the intervening lows between the banks, there is an accumulation of varying amounts of fine-grained sediments which has masked older deposits and modified their morphology. The granulometry and mineralogy of the fine-grained sediments indicates that they have been partly derived from the winnowing of the adjacent bank sediments.

SUMMARY AND CONCLUSIONS

The northern part of the St. Lawrence River and Gulf is underlain by relatively undeformed Paleozoic (mainly Ordovician-Silurian) sedimentary rocks (limestones and calcareous shales and sandstones). To the north these fault against or overlap Precambrian crystalline rocks. In the St. Lawrence river and estuary, these rocks are in fault contact with folded Ordovician rocks (mainly slates) of the Appalachian Front. The fault zone which separates these rocks represents the seaward extension of Logan's Line. Further to the southeast the fault zone disappears beneath a cover of younger sedimentary rocks. South of Anticosti Island the middle Paleozoic strata (Silurian limestones) are overlapped by stratified Permo-Carboniferous rocks (sandstones and shales). The Permo-Carboniferous rocks occupy the southern part of the Laurentian Trough and the Magdalen Shelf.

The main feature of the bedrock topography is a well-established system of deep erosional valleys, the most prominent of which is the Laurentian Trough, belonging to a pre-glacial drainage system. These valleys as we know them have developed by sequential fluvial erosion along lines of structural and lithological weaknesses since early Cenozoic times. The establishment of the main valleys in more easily eroded material has left the more resistant rocks to form the adjacent uplands or shelves on which cuesta-like landscapes were developed.

During the Pleistocene the whole area was glaciated. The submarine valleys were modified into glaical troughs by the ice which repeatedly eroded them and filled them with glacial sediments. Glacial erosion by these ice masses, strongly controlled by the pre-glacial bedrock topography, resulted in the widening and straightening of the valley walls and deepening of the floor of the deep and shallow erosional valleys. Uneven deposition of glacial sediments over the bedrock surfaces has resulted in the production of an irregular glacial landscape over most of the bedrock. Additional morphological features such as moraines and thick wedges of glacio-marine sediments buried by glacial tills were also produced by the advance and retreat of the continental ice and by the readvance of local ice masses (<10,000 yrs. B.P.) when the continental ice had almost withdrawn from the Gulf. Post-glacial rises in sea level have resulted in the formation of submarine terraces and the modification of the glacial landscape through the reworking (erosion and redeposition) of the glacial deposits and deposition of recently supplied material.

The surface pattern of sediment distribution in the St. Lawrence River and Gulf reflects the present depositional conditions and to a certain extent the depositional conditions that have been effective during the Pleistocene glaciations and post-glacial rises in sea level.

ACKNOWLEDGMENTS

The author wishes to thank Dr. B.R. Pelletier and A.C. Grant for their interest and constructive comments on the manuscript. He also would like to thank Mr. B. MacLean, Atlantic Geoscience Centre, Bedford Institute of Oceanography, for his help in obtaining the continuous seismic profiles on board CSS KAPUSKASING in May 1968. Mr. S. Deleu (Bedford Institute of Oceanography) prepared the illustrations.

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(Earth Science Symposium on Offshore Eastern Canada
Geol. Surv. Can., Paper 71-23, 1973, p. 325-336)

20. SEDIMENTARY REFRACTION SEISMIC SURVEYS, GULF OF ST. LAWRENCE

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Abstract

From a two-ship marine seismic refraction program, begun in 1964, 94 profiles were obtained from which some aspects of the geology underlying the Gulf of St. Lawrence are interpreted. The seismic refraction results of previous workers, Sheridan and Drake (1968), Willmore and Scheidegger (1956), MacPherson (1962), and Ewing and Hobson (1966) have been incorporated for the present interpretation.

The Carboniferous section about 5 kilometres (16,400 feet) thick northeast of Cape Breton Island remains, with minor modification, as interpreted by Sheridan and Drake. Also remaining essentially unchanged is the sharp rise in the basement across Cabot Strait, separating the seaward Carboniferous trough from the Gulf sedimentary basin. Within the Gulf, Carboniferous sediments exceeding 7 kilometres (22,960 feet) in thickness are indicated within a trough containing the Cumberland Basin, extending northward to the west of the Magdalen Islands and veering sharply eastward to the north of these islands to St. Georges Bay, Newfoundland. These features within the Appalachian Geosyncline are defined by seismic velocities exceeding 5.6 kilometres per second (18,370 feet per second) which are presumed to represent the pre-Carboniferous basement complex. Thicknesses to the same seismic basement (5.6 kilometres per second or 18,370 feet per second) of 3.5 kilometres (11,480 feet) and 4.5 kilometres (14,760 feet) in the Anticosti Basin are indicated to the west of Port au Port Peninsula and Bay of Islands, Newfoundland respectively. These sections include St. Lawrence platform rocks with a Precambrian basement and may also include some Carboniferous or pre-Carboniferous non-marine sediments. The transition between the two types of basement is vaguely suggested by fairly smooth depth contours within the Anticosti Basin grading into the more complex contours of the troughs within the Appalachian Geosyncline and a suggested trough extending into the mouth of the St. Lawrence River. Some indecisive evidence for this basement transition can be found from the areal distribution of basement velocities. Another structure map of depth to seismic velocities exceeding 4.5 kilometres per second (14,760 feet per second) shows, with the exception of the trough extending into the mouth of the St. Lawrence River, generally subdued expressions of the deeper major features already mentioned. Over the survey area, this structure map is also presumed to include some possible non-marine sediments of pre-Upper Pennsylvanian age near southwestern Newfoundland, as well as sediments from Recent to Pennsylvanian in age. Depths approaching 3 kilometres (9,840 feet) are indicated in a roughly northeast-southwest trending basin centered north of the Magdalen Islands.

The results of this work do not support theories of structural origin for either the Laurentian Channel or the Esquiman Channel.

G.S.C. Paper 71-23

INTRODUCTION

In 1964 the Geological Survey of Canada began a program of seismic refraction surveys in the Gulf of St. Lawrence which was completed in 1969 with one-month cruises in the fall of each year. The objectives were to investigate the thickness and type of sediments and configuration of the sedimentary and basement surfaces underlying the Gulf. Two ships were used, one acting as the recording vessel to tow the hydrophone cable to selected positions along the profile while the other functioned as the shooting vessel at the shotpoint marker buoy. The shotpoint remained fixed for each profile. In this way, either single ended profiles with travel-time variation in one direction from the shotpoint, or more commonly split profiles with travel-times recorded in two, approximately opposite, directions radiating from the shotpoint were obtained. Weather conditions usually dictated the choice between a split or single ended profile. The split spreads may be interpreted for dipping beds using the same assumptions used in "reversed" profile interpretations which do not obtain overlapping information from each direction on each velocity layer.

Twelve channels of seismic data were recorded for each shot, using a Texas Instruments model 7000 B amplifier system and two Electro Tech EVP-7 pressure sensitive hydrophones per channel. The hydrophones were attached to a neutrally buoyant marine seismic cable with 19 meters separation between each pair of the twelve groups and 76.2 meters between the groups. The cable was allowed to assume its neutrally buoyant state to minimize noise before recording.

Positions of the shotpoints were usually determined from Decca navigation chain 9 and occasionally also by radar fixes on shoreline features. Distances of the hydrophone groups from the shots were determined from the travel-time of the seismic wave through water using a velocity of 1.45 kilometres per second (4,757 feet per second).

Some land based projects on Newfoundland, Prince Edward Island, Nova Scotia, New Brunswick and Quebec have resulted in seismic velocity determinations for some of the bedrock formations which may also be encountered in the offshore areas. Figure 1 shows the locations of the seismic profiles.

Seismic refraction results of previous workers have been included. The work of Sheridan and Drake (1968) in the Gulf of St. Lawrence was part of a larger project including studies on the shelf areas off Labrador, Newfoundland and Nova Scotia. The work of Willmore and Scheidegger (1956) investigated the concept of a meteoric origin for the circular coast line formed by the north coasts of Cape Breton Island and Nova Scotia, and by the east coast of New Brunswick. The work of MacPherson (1962) consisted of two depth determinations by unreversed seismic refraction profiles off the north shore of Prince Edward Island. Ewing and Hobson (1966) obtained three seismic refraction profiles off the east coast of Nova Scotia in a seismic investigation of the Orpheus gravity anomaly.

Regional Geology

The Appalachian region of Canada comprises southeastern Quebec and the Maritime provinces with the exception of the coast of Labrador, and is the northeastern continuation of the larger Appalachian belt of folded rocks

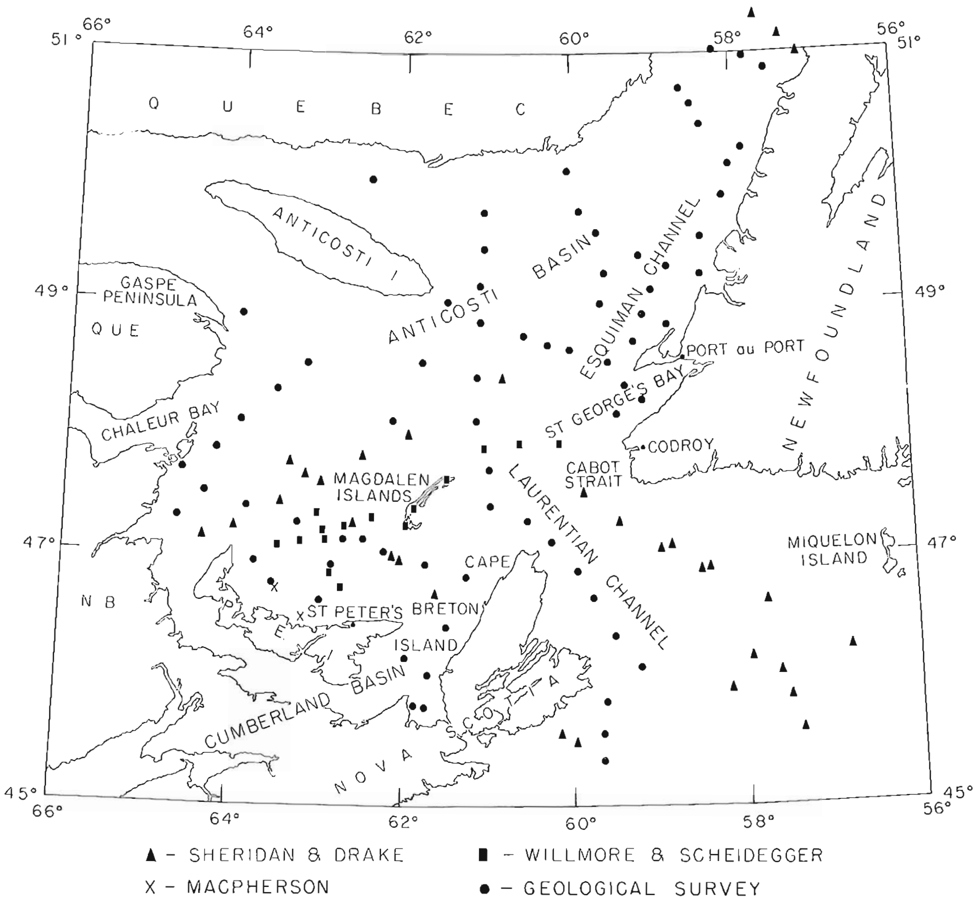

Figure 1: Location of seismic profiles, Gulf of St. Lawrence area.

extending along the eastern United States. The regional geology of the Appalachian region of Canada has been described by Neale et al.(1961, see Figs. 2 and 3), and by Howie and Cumming (1963), Cumming (1967), Poole et al.(1970).

Previous Seismic Surveys in the Gulf of St. Lawrence

A discussion of the published results of previous work based on the seismic refraction method is included inasmuch as certain differences in the mapping of structures, correlation of rock types with seismic wave velocities and considerations in integrating the results must be emphasized.

For the Gulf of St. Lawrence the most extensive refraction seismic program previously undertaken was described by Sheridan and Drake (1968). They present selected cross sections, two subsurface structure maps and an areal distribution of seismic wave velocities in the basement rocks. They found a belt roughly described as extending from New Brunswick, Prince Edward Island and Cape Breton Island across the Gulf to the southwest corner of Newfoundland, within which seismic velocities for basement rocks were

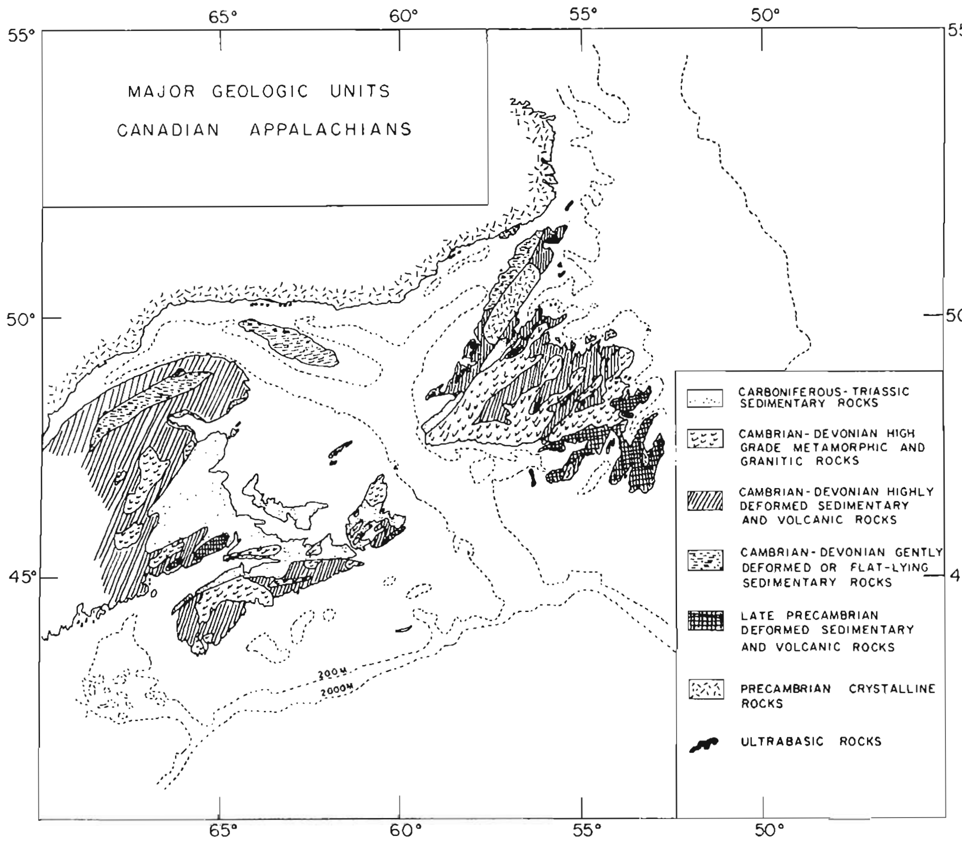

Figure 2: Generalized geological map of Canadian Appalachians, (after Neale et al., 1961).

high with values of 6.0 to 6.6 kilometres per second (19,680 to 21,650 feet per second) compared with basement velocities of 5.1 to 6.0 kilometres per second (16,730 to 19,680 feet per second) outside the belt. Higher velocities within the belt were observed near to, and hence thought to be a manifestation of, areas of Acadian granitic intrusions and metamorphism. The first of their subsurface structure maps (Figure 15 in Sheridan and Drake, 1968) shows depths computed to the base of sediments having velocities within the range of 1.7 to 4 kilometres per second (5,570 to 14,430 feet per second) which include rocks from Recent to Pensylvanian in age. This structure was taken to represent mainly, features of the Carboniferous orogeny which ended in the Pennsylvanian Period, and thus to represent post-orogenic structural development. Their second subsurface-structure map shows depths computed to the base of sediments having velocities of 5.6 kilometres per second (18,370 feet per second) or less. This structure was taken to represent the pre-Carboniferous basement over the limited area of the Gulf of St. Lawrence, Cabot Strait, Laurentian Channel, and Scotian shelf, southeast of Cape Breton Island. Correlation of velocities within the range of 4.5 to 5.5 kilometres per second (14,760 to 18,040 feet per second) outside these areas was uncertain except for being representative of pre-Upper Pennsylvanian rocks.

Willmore and Scheidegger (1956) concluded from their P_1 observations that the circular part of the Gulf contains about 6 kilometres (19,680 feet) of sediments in the centre and probably a greater thickness near St. Peter's on Prince Edward Island, and that there appears to be a change of structure on passing eastward from the Magdalen Islands into the Cabot Strait which is characterized by an abrupt increase in the P_1 time-term. Wilmore and Scheidegger reserved doubts on the latter conclusion inasmuch as the abrupt drop in the basement depends on the accuracy of the time-term and the constant time which may be added to or subtracted from shotpoints or recording stations. This constant time represents the relative structure between shotpoints and recording stations and its determination sometimes must be a compromise of many choices within the limitations demanded by geological constraints. Their P_1 events were characterized by velocities of 6.08±.13 kilometres per second (19,940±425 feet per second).

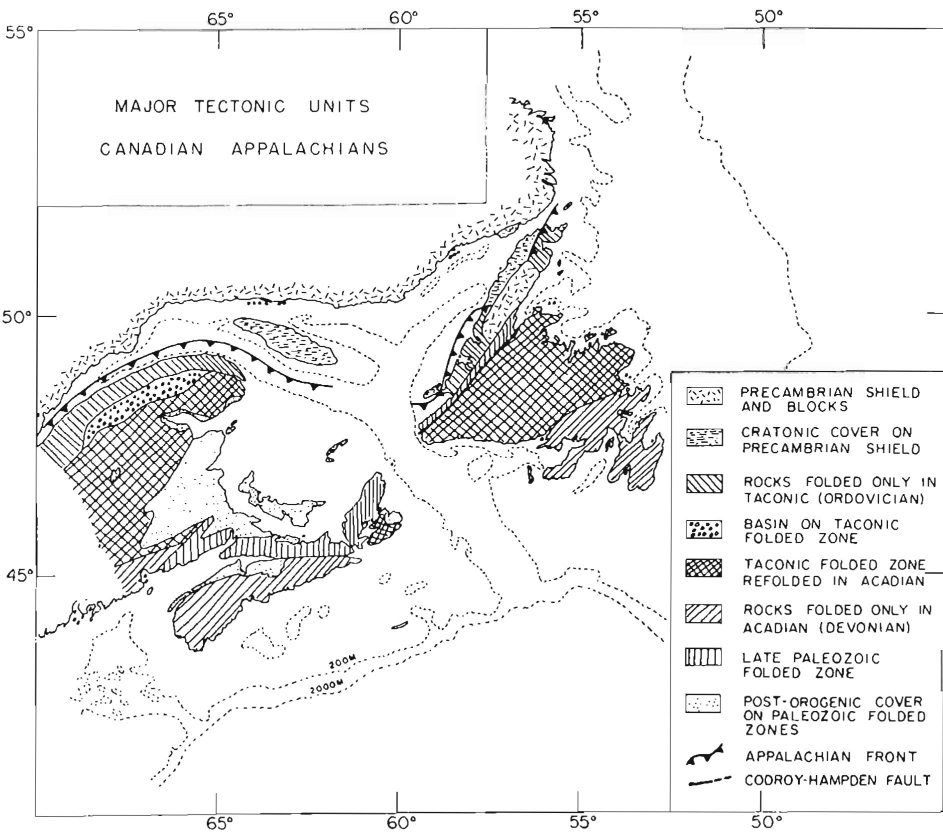

Figure 3: Generalized tectonic map of Canadian Appalachians, (after Neale et al., 1961).

The work of MacPherson (1962) consists of two depth determinations, by the seismic refraction method, off the north coast of Prince Edward Island. He concluded that depths and velocities for the four sedimentary layers detected at both locations were identical within limits of experimental error, and that his fourth layer with a seismic velocity of 5.1 kilometres per second (16,730 feet per second) is probably a thin layer of anhydrite rather than basement.

Velocity Control from Auxiliary Surveys

Hammer seismic projects conducted on Gaspe Peninsula, Quebec, New Brunswick, Nova Scotia, and Prince Edward Island, and similar bedrock studies using conventional seismic equipment on Newfoundland have given an estimate of seismic velocities for various formations which may be expected in the Gulf of St. Lawrence (Hobson 1966, 1970; Hobson and Carr, 1965; Gagne, 1971a and 1971b).

In general, the hammer seismic work shows broadly distributed velocities ranging from 1.5 to 5.7 kilometres per second (4,920 to 18,700 feet per second). Other than sediments of Recent age, which are represented by the lower velocities, this range includes sediments of Permian and Carboniferous non-marine beds. Considerable overlap of velocities exists with no clearly defined contrasts for these sediments.

The velocity determinations on Newfoundland are of particular interest since they include some of the rocks which may be expected to form the basement within the Appalachian Geosyncline. Figure 4 shows the distribution of velocities obtained. It will be seen that with the exception of two measurements exceeding 5.6 kilometres per second (18,370 feet per second) all are included

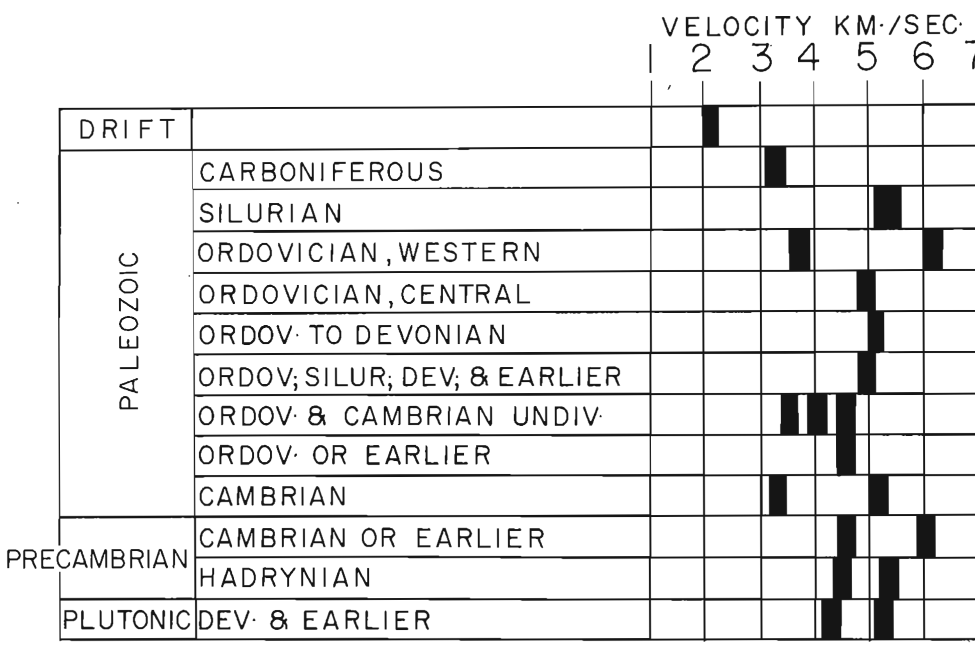

Figure 4: Seismic velocities at selected sites in Newfoundland. (1 km = 3,281 ft).

in the range of 5.6 kilometres per second (18,370 feet per second) or less which presumably represents Carboniferous and later sediments within the Appalachian Geosyncline.

Projected to depths of 4 kilometres (13,120 feet) or greater, and hence pressures exceeding one kilobar, velocities for these rocks are usually well in excess of 5.7 kilometres per second (18,700 feet per second) according to velocity-pressure relationships published by Hughes and Cross (1951), Hughes and Maurette (1956), Birch (1960) and Christensen (1965). Marine facies in the Carboniferous section projected to similar pressures may also be expected to have seismic velocities in excess of 6 kilometres per second (19,690 feet per second). Thus the seismic basement, defined by velocities in excess of 5.6 kilometres per second (18,370 feet per second) may be expected to represent a variety of rock types depending upon lithology, degree of induration and depth of burial.

Seismic Structure Maps of the Gulf of St. Lawrence and Laurentian Channel

The first seismic structure map (Figure 5) shows depths to the base of the 4.5 kilometres per second (14,760 feet per second) velocity layer and

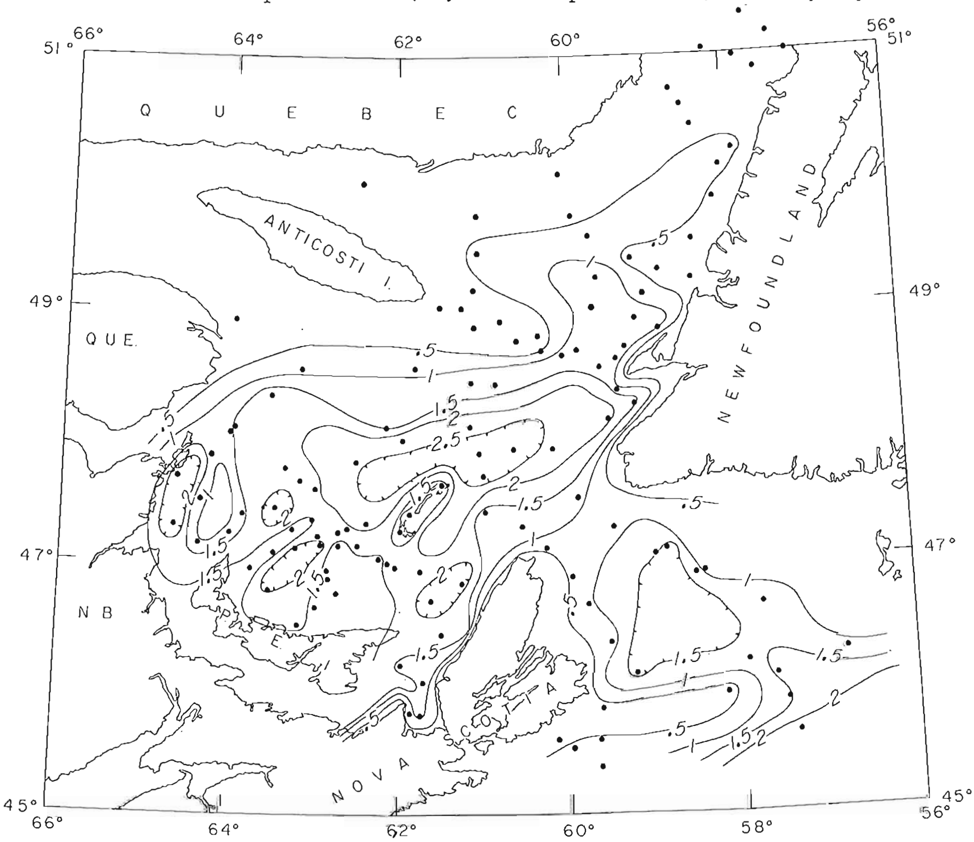

Figure 5: Seismic structure map (thickness of sediments) to base of 4.5 kilometres per second (14,760 feet per second) velocity layer, Gulf of St. Lawrence area. Depths in kilometres (1 km = 3,281 ft).

represents the combined thicknesses of velocities ranging from 1.7 to 4.5 kilometres per second (5570 to 14,760 feet per second). Within the confines of the Appalachian system these contours probably represent depths to well indurated rocks deformed by Paleozoic orogenies and thus includes rocks from Recent to Pennsylvanian in age. The basin exceeding 2 kilometres (6,560 feet) in depth almost surrounding the Magdalen Islands may generally indicate post-orogenic subsidence, but may also include some accumulations of unmetamorphosed non-marine sediments of Pre-Upper Pennsylvanian age which are likely to have seismic velocities within this same range. In the Anticosti Basin the contoured structures probably include non-marine Devonian sediments such as the Clam Bank group which is exposed on Port au Port Peninsula and have been traced offshore in sporadic outcrops for about 80 miles to the northeast, as well as some flat-lying Carboniferous sedimentary rocks overlapping the margin of the St. Lawrence platform as on Port au Port Peninsula (Cumming, 1967).

The second seismic structure map (Fig. 6) shows depth contours to velocities exceeding 5.6 kilometres per second (18,370 feet per second).

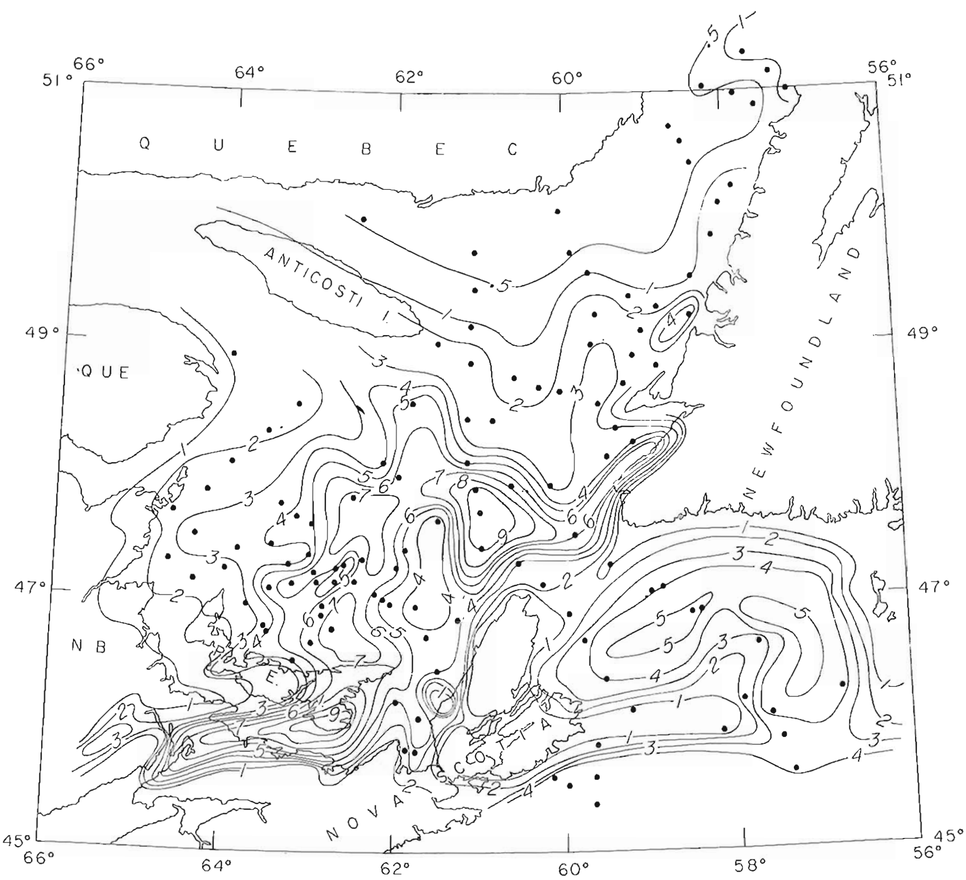

Figure 6: Seismic structure map, pre-Carboniferous basement (velocity greater than 5.6 kilometres per second, Gulf of St. Lawrence area. Depths in kilometres (1 km = 3,281 ft).

Thicknesses of sediments with a velocity range of 4.5 to 5.5 kilometres per second (14,760 to 18,040 feet per second) are thus added to those of the first structure map. These higher velocities are considered to represent the lower part of the Carboniferous section within that part of the Gulf of St. Lawrence coincident with the Appalachian Geosyncline and Cabot Strait. At the depths suggested, however, marine facies such as Windsor evaporites and limestones may well have seismic velocities equal to or exceeding 5.6 kilometres per second (18,370 feet per second). In this area the velocity range may also include non-marine sediments of Devonian age. In the Anticosti Basin, the contours probably represent depths to the Precambrian basement and include a thickness of slightly deformed Ordovician, Silurian and Devonian sediments. Contours onshore are after Howie and Cummings (1963).

The most striking feature of Figure 6 is the deep trough extending from the Cumberland Basin south of Prince Edward Island northward towards Anticosti Island on a line west of the Magdalen Islands and veering sharply eastward to the north of the Magdalens towards the Codroy Basin in St. George's Bay, Newfoundland. Depths in excess of 9 kilometres (29,520 feet) are indicated along this trough. The high angle fault bounding the southeast side of this trough across the Cabot Strait to the Codroy fault on southwestern Newfoundland, the basement ridge across Cabot Strait and the Carboniferous basin 5 kilometres (16,400 feet) deep between Cape Breton Island and Miguelon Island remain the same as that published by Sheridan and Drake (1968). The greatest depth in the trough indicated by seismic control is 9.5 kilometres (31,160 feet) northeast of Magdalen Islands and is one of the values of Willmore and Scheidegger (1956) which they held in doubt. The question of its credibility must be re-examined. As earlier explained, Willmore and Scheidegger reserved doubts about the abrupt increase in depth indicated by this shotpoint as being the weakest of their conclusions, drawn from the basement observations, depending upon the accuracy of time-terms and their choice of α representing the relative structure between their shotpoints and recording stations. The integration of the Willmore and Scheidegger depths with those of the more recent surveys provides one means of reappraisal. The depths for their shotpoints between the Magdalen Islands and Prince Edward Island pose no particular problem in combination with other data; nor do the depths for their recording stations on the Magdalen Islands, lending support to their choice of structural balance for these locations. Direct comparison of the 9.5 kilometres (31,160 feet) depth northeast of the Magdalen Islands with values from the more recent surveys appears to be somewhat doubtful, but for two profiles to the south of this location, a basement velocity of 5.6 kilometres per second (18,370 feet per second) or greater was not encountered even at distances of 16 kilometres (52,490 feet) from the shotpoints. Forcing a velocity of 5.8 kilometres per second (19,020 feet per second) at the ends of these profiles results in minimum depths of 4 to 5 kilometres (13,120 to 16,400 feet). Depths at least intermediate to 5 and 9.5 kilometres (16,400 and 31,160 feet) may therefore be significant for this location.

The transition from the Anticosti Basin into the Paleozoic deformed belt may be suggested by the gradation of the smoother contours of the former into the more complex structures depicted within the Appalachian Geosyncline. Depths published by Roliff (1968) tend to indicate that the trough suggested in Figure 6 extends into the mouth of the St. Lawrence River. Structures are continuous across the Laurentian and Esquiman Channels and do not support theories of structural origin for these features (Kumarapeli and Saul, 1966).

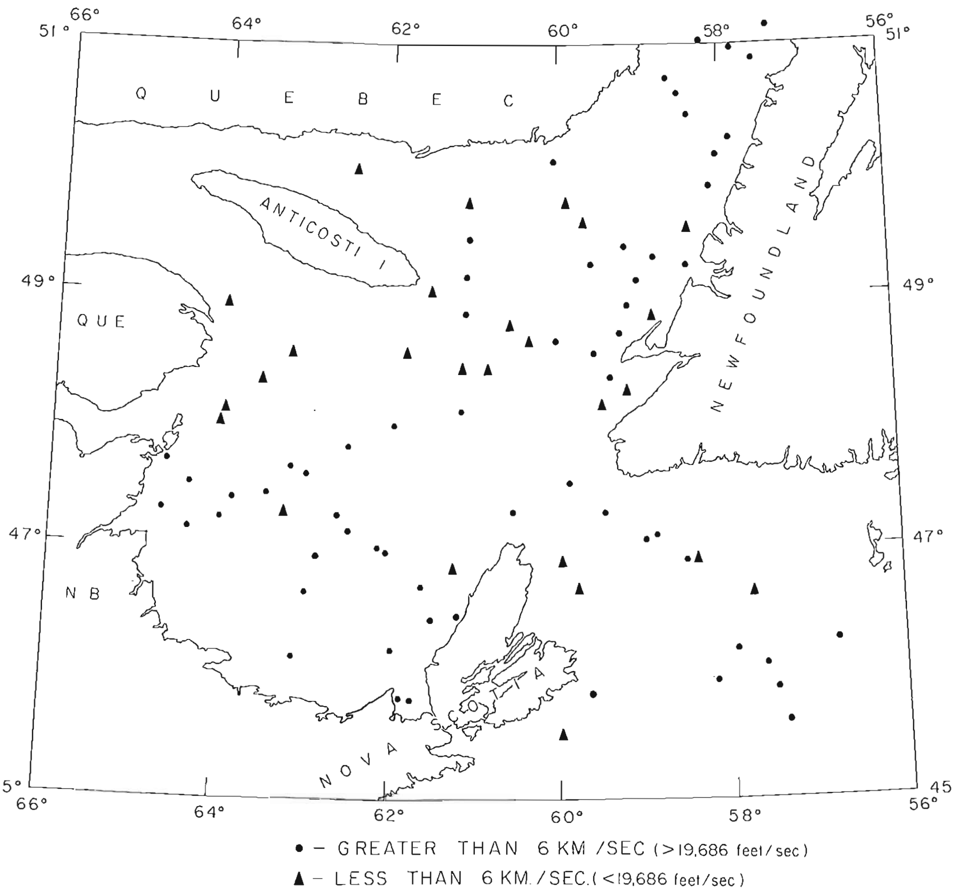

Figure 7: Areal distribution of basement velocities, Gulf of St. Lawrence area.

Figure 7 shows the areal distribution of basement velocities. With few exceptions this map confirms the conclusion of Sheridan and Drake that higher seismic velocities are observed nearer to the areas of Acadian granitic intrusion and metamorphism since velocities are seen to be generally greater than 6 kilometres per second (19,680 feet per second) south of a line extending from the Bay of Chaleur to St. Georges Bay. North of this line, however, velocities may be greater or less than 6 kilometres per second (19,680 feet per second) and do not provide a basis for drawing distinct boundaries. The lower velocities observed during the bedrock studies in Newfoundland suggest that some of the shallower depths to basement velocities exceeding 5.6 kilometres per second (18,370 feet per second) may also include sections of these folded rocks.

CONCLUSIONS

Interpretation of the seismic data is complicated by problems of velocity correlations; a few of the anomalies appear to be explainable in terms of known geological structures, but at this stage no attempt has been made to resolve the less obvious ones. This will be best done by combining the results of all disciplines.

ACKNOWLEDGMENTS

These data were collected, frequently under very adverse conditions, only with the support and enthusiasm of R.A. Hodge and R.M. Youngman in the field and of H.A. MacAulay and R.M. Gagne in the field and in the office. The Captains, Officers and crews of several vessels participated over the years; the personnel of C.S.S. HUDSON, C.N.A.V. SACKVILLE, M.V. THETA, M.V. THERON, M.V. HARENGUS and M.V. FAIRMORSE all contributed significantly in the acquisition of these data through their experience at sea and their interest in our program.

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(Earth Science Symposium on Offshore Eastern Canada
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21. GRAVITY MEASUREMENTS IN THE GULF OF ST. LAWRENCE

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Abstract

During 1968 and 1969 a hydrographic-geophysical survey was carried out in the Gulf of St. Lawrence by the Atlantic Oceanographic Laboratory, Bedford Institute of Oceanography. In surveying the Gulf east of 62° W and north of 47° N, bathymetry, magnetic and gravity data were obtained

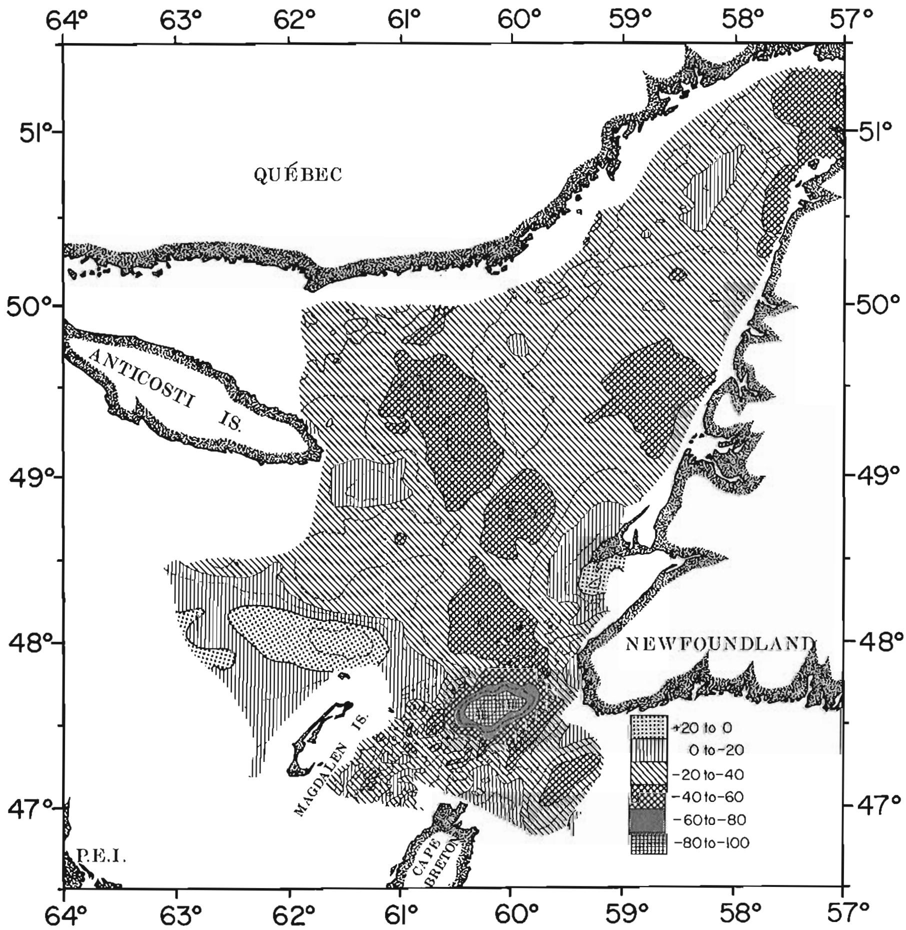

Figure 1: Free air gravity anomaly map of the Eastern Gulf of St. Lawrence as obtained from surface ship measurements by Bedford Institute.

*Now at Lamont-Doherty Geological Observatory, Palisades, New York.

along ships tracks totalling approximately 55,000 km. This corresponds to an average line spacing of 1.2 km. The free air gravity field derived from the surface ship survey measurements has been compiled for publication by the Canadian Hydrographic Service in their 1:250,000 Natural Resource Chart series.

The surface ship gravity survey covers much of the eastern Gulf of St. Lawrence previously surveyed by the Gravity Division, Earth Physics Branch with underwater gravimeters. A comparison between the two sets of data indicates that each survey has an internal repeatability of the order of 2 mgal, the limiting factors being navigation and sea state for the surface ship survey, and navigation and depth measurement for the underwater survey.

Most features of the gravity field in the Gulf have been previously discussed by Goodacre and Nyland (1966) and Goodacre et al. (1969) with reference to the siesmic work of Sheridan and Drake (1968). A notable exception is the series of gravity "lows" of up to 25 mgal amplitude revealed by the continuous gravity measurements of the surface ship survey while not detected by the 15 km spacing of the underwater survey. These "lows" are interpreted as being caused by evaporite bodies within the Carboniferous sequence causing the main Magdalen low centred over the Cabot Strait.

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(Earth Science Symposium on Offshore Eastern Canada,
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22. MAGNETIC SURVEYS OF THE GULF OF ST. LAWRENCE AND THE SCOTIAN SHELF

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Abstract

Magnetic surveys have been carried out over most of the Maritime area of eastern Canada. Aeromagnetic surveys have been flown in the southern half of the Gulf of St. Lawrence as far north as 48° N. Several large wavelength anomalies in the central part of the Gulf have been interpreted using a curve-matching computer technique, and depths to magnetic basement in excess of 12 kilometres were obtained northwest of Cape Breton Island. These interpretations have been supported by measurements of the magnetic properties of rocks collected from the land masses surrounding the Gulf of St. Lawrence. Sea magnetometer and aeromagnetic surveys of the major portion of the continental shelf southeast of Nova Scotia were carried out during the last decade. Close to the shoreline the Meguma Group of slates and quartzites produces a characteristic pattern of sharp linear magnetic contours which parallel the coastline. Several circular granitic intrusions are apparent due to their low intensity of magnetization and the fact that they tend to have magnetic aureoles around their peripheries. Further to the southeast the amplitude of the magnetic anomalies decreases and their wavelength increases because of greater depth to the crystalline basement, i.e. thickness of sedimentary rock. Depth-to-crystalline basement have been carried out on most of the significant anomalies on the Scotian Shelf and it would appear that the greatest thickness of sediments (7.5 kilometres) occurs in the vicinity of Sable Island. The basement depths determined from the magnetic data for the Gulf of St. Lawrence and the Scotian Shelf are generally deeper than has been suggested by seismic refraction surveys. It would seem in some cases that the seismic data represent depths to high-velocity refractors which occur above the crystalline basement.

INTRODUCTION

The magnetic map coverage of Canada to the end of 1972 is shown in Figure 1. The maps resulted from sea magnetometer and aeromagnetic surveys carried out by the Geological Survey of Canada in association with the provincial governments, and by the Bedford Institute of Oceanography.

Offshore magnetic survey data may be interpreted in two different ways. The first method is qualitative, and involves determining lithologies and structural features such as contacts, faults etc., of the crystalline basement rocks in much the same way as is done for the Precambrian shield. The second method utilizes quantitative interpretation techniques to determine the configuration of the crystalline basement surface in order to establish the thickness of the overlying sediments. The degree of success of both of these methods of interpretation is enhanced by ground observations and measurements. Hence it is desirable to obtain measurements of the magnetic

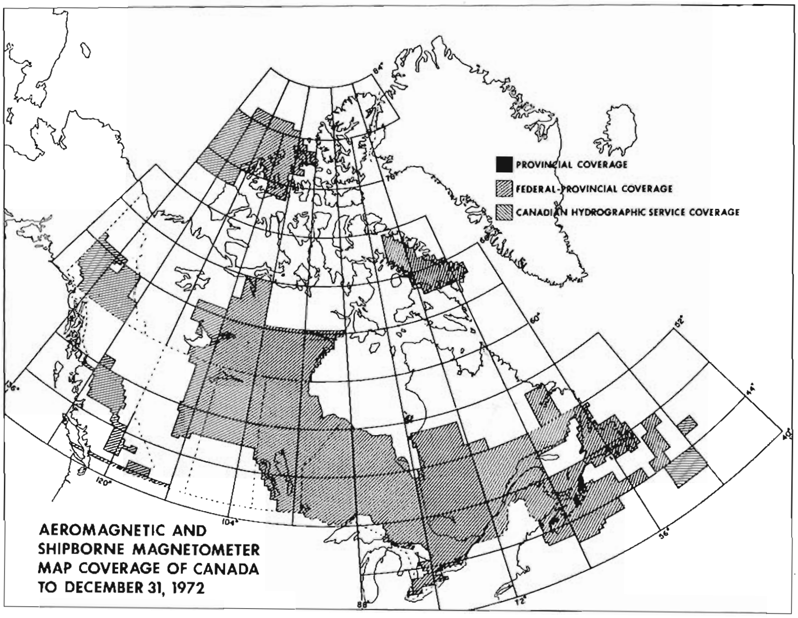


Figure 1. Aeromagnetic and shipborne magnetometer map coverage of Canada to the end of 1972.

properties of rocks from the adjacent onshore areas in order to extrapolate more reliably from the known geology onshore to the offshore. We will deal firstly therefore with the measurement of onshore rock properties.

Magnetic Property Surveys in the Maritimes

Figure 2 shows the locations in southwestern New Brunswick at which ground magnetic information was collected during 1968 (McGrath, 1969); and Figure 3 shows similar locations in New Brunswick and Nova Scotia for the 1969 magnetic property field survey (McGrath, 1970a).

It was found that Silurian and Devonian sediments tend to be non-magnetic whereas many of the Cambrian and Ordovician sediments possess measurable magnetic properties. Also it was noted that in general all magnetic rock units are magnetically inhomogeneous, even the plutonic rock units. The reason for this is unclear. Variations in magnetic susceptibility determined for various rock types for the 1968 data are shown in Figure 4. This figure illustrates the difficulty of relating susceptibility to lithology. There is essentially no difference in the range of susceptibility for acid and basic volcanic and plutonic rocks.

Figure 5 shows a histogram of Königsberger ratios (Q-factors) determined for about 100 oriented drill cores collected from magnetic rock units in southwestern New Brunswick during 1968. The ordinate is the number of samples collected versus the Q-factor which is the ratio of remanent to induced magnetization. It can be seen that the Q-factor of most of the samples falls between 0.1 and 0.2, the modal value being 0.15. This result indicates that in general remanent magnetism is of much less importance than the induced component. It follows that the polarization vector of a given causative body would be essentially parallel to the ambient direction of the earth's magnetic field.

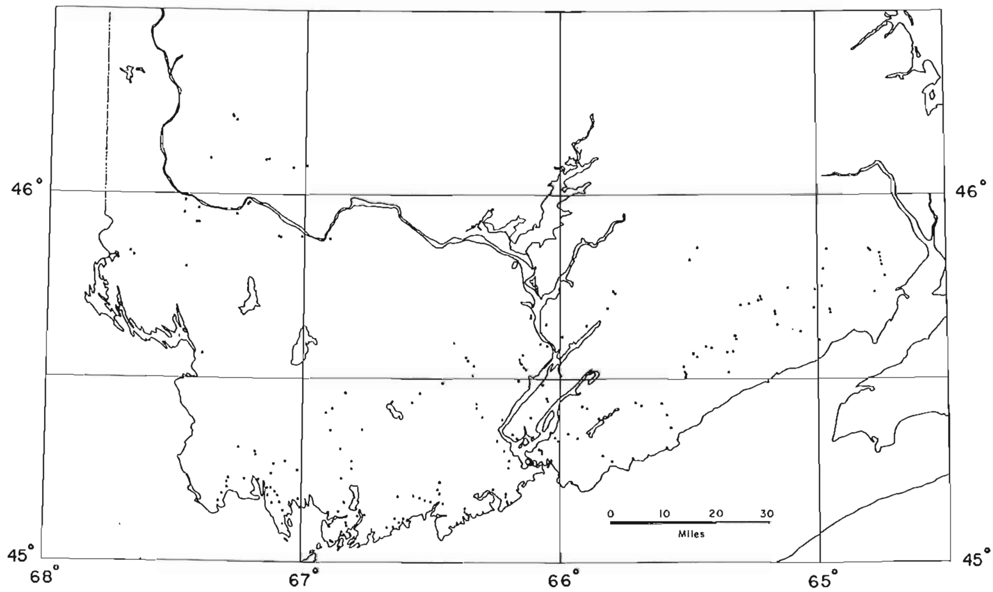


Figure 2. Sampling sites in New Brunswick for the 1968 magnetic properties field survey.

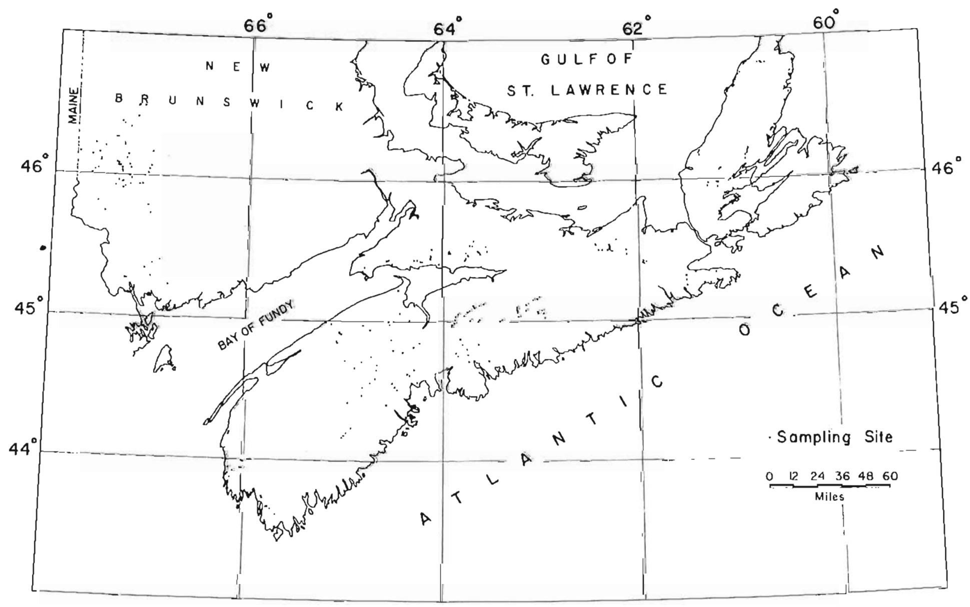


Figure 3. Sampling sites in New Brunswick and Nova Scotia for the 1969 magnetic properties field survey.

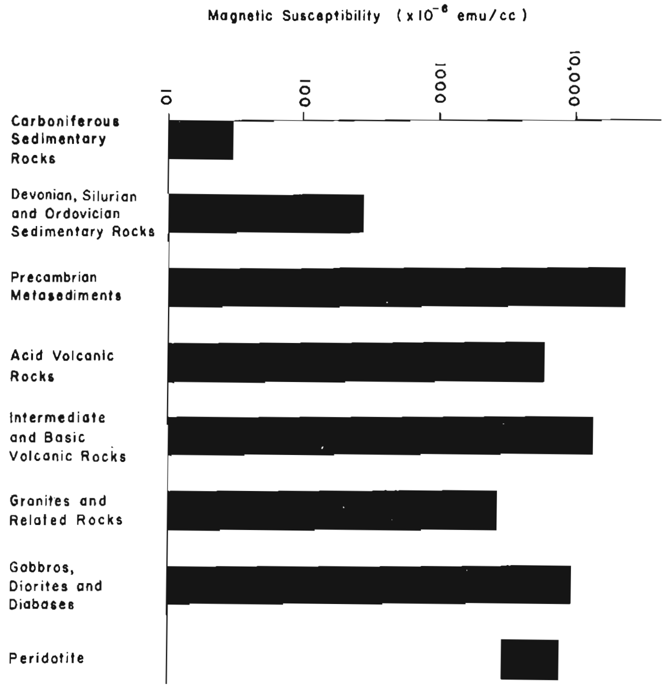


Figure 4. Range of in-situ magnetic susceptibility values in New Brunswick.

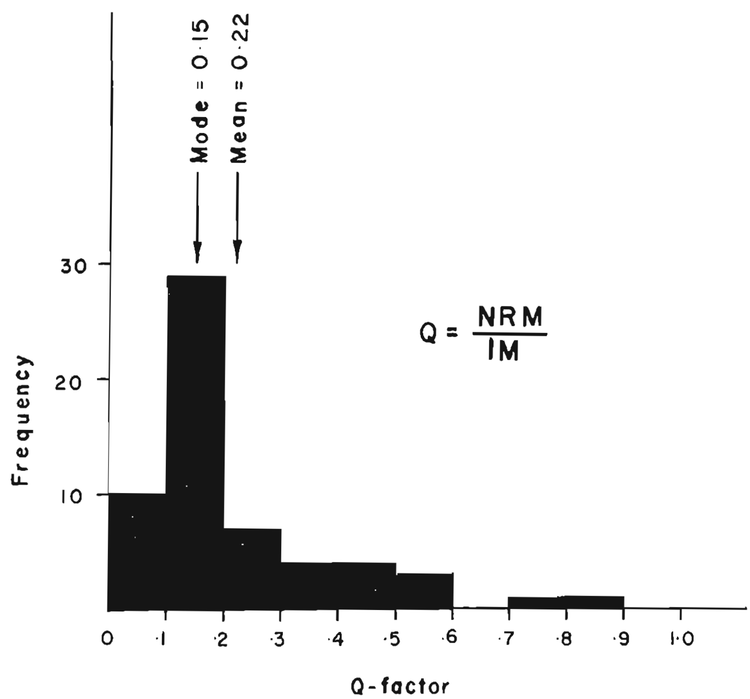


Figure 5. Frequency distribution of Q-factors (Königsberger ratios) for rock cores obtained in southern New Brunswick.

In general, it can be said that the spatial correlation between aeromagnetic anomalies and geological units is quite good in eastern Canada, especially in Nova Scotia. However, deductions with regard to the lithology are usually much more difficult and uncertain. An interesting case in point is the delineation of granites from magnetic surveys. On the left hand side of Figure 6 is part of an aeromagnetic map compiled from survey lines flown at 1000 feet elevation, over a Devonian granite body which intrudes the Meguma Group south of Chedabucto Bay (Hood, 1966). It is obvious that the granite has a very low intensity of magnetization and thus contains an extremely low percentage of magnetite. This fact has also been checked by specimens collected from the granite body. Field investigations have shown that the magnetic properties of the Meguma are due to the magnetic iron sulphide pyrrhotite (Schwarz and McGrath, 1973). The magnetic aureole around the granite body appears to be due to the fact that the pyrrhotite has been changed to magnetite by the intrusion. Thus by direct comparison it is most likely that the anomaly shown on the right hand side of Figure 6 is in fact due to a circular granite intrusion in the Meguma. However, elsewhere in eastern Canada particularly in southwestern New Brunswick and in northeastern Newfoundland, a number of granites occur which have quite significant magnetic properties (McGrath, 1970b). One of these is shown in Figure 7 and is the Pokiok batholith in southern New Brunswick just west of Fredericton. The granite is outlined in black on the figure, and it appears that there are two phases, a magnetic and a non-magnetic phase.

Quantitative Interpretation of Magnetic Survey Data

We come now to a discussion of the quantitative interpretation of magnetic data. A generally applicable method developed at the Geological Survey of Canada will be briefly outlined (McGrath and Hood, 1973), and the results of using this method for depth determinations over the Scotian Shelf and the Gulf of St. Lawrence will then be described.

In order to develop a comprehensive system of quantitative interpretation one requires a variety of geometrical models to choose from. These models may be made up from a basic building block. Our basic building block is the thin dipping rectangular plate. Figure 8 shows an oblique view of the thin dipping plate with the nomenclature used referred to a set of orthogonal axes xyz. J is the intensity of magnetization vector which dips at an angle i and whose declination with respect to the y axis is d. The earth's magnetic field vector T has a dip I and declination D with respect to the y axis. It can be seen that the model is applicable to the case where there is a remanent magnetization component which is not necessarily parallel to the present direction of the earth's magnetic field, so that the model is not limited to the induced magnetization case only.

Figure 9 shows how various geometrical models may be derived from a series of thin plates by numerical integration. At the top of the figure are various infinite strike-length bodies such as the thick sheet, horizontal bar, thin ribbon etc; these models may be used to interpret elongated anomalies. The bottom half of the figure shows various finite strike-length bodies which may be used to interpret more equidimensional anomalies. It should be noted that the thick plate model is a more general case for the two vertical prism cases.

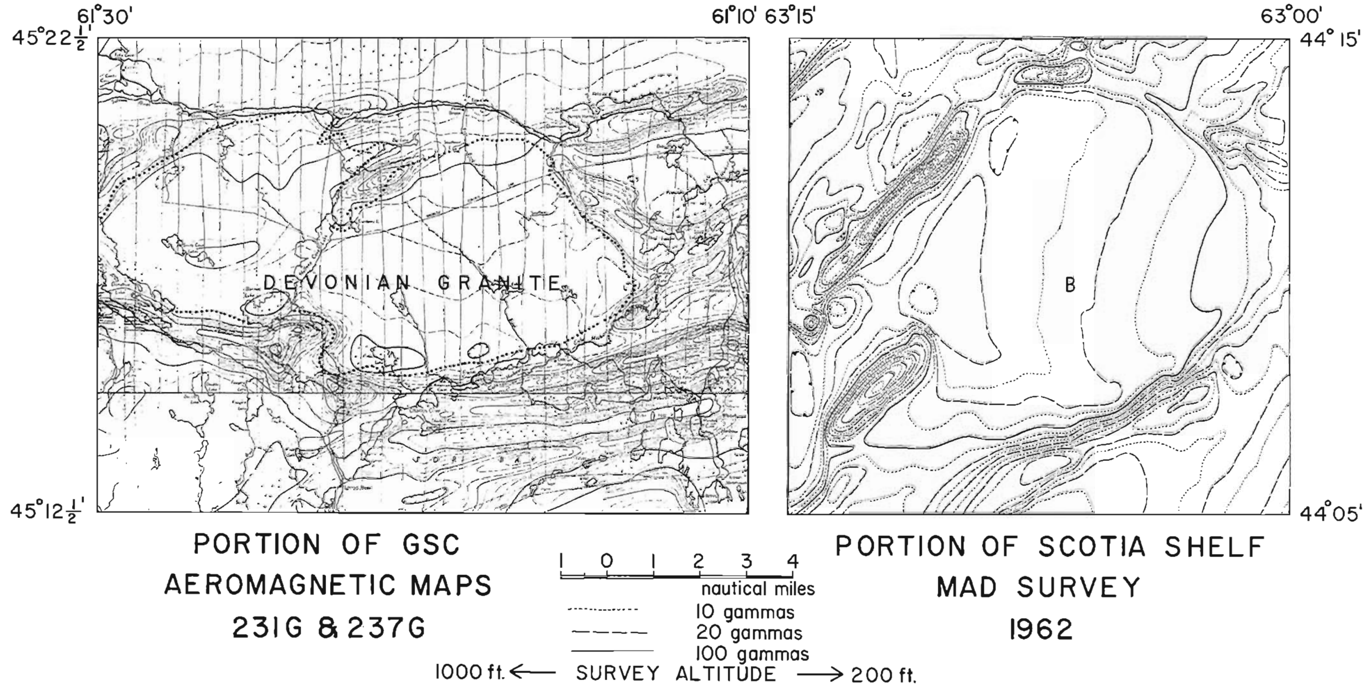


Figure 6. Comparison of circular aeromagnetic features on the Scotian Shelf with similar patterns observed on the Nova Scotia mainland south of Chedabucto Bay.

Figure 10 shows how the vertical prism model may be derived from a series of thin plates by numerical integration. The formula for the numerical integration which is carried out using Simpson's rule is shown at the bottom of the figure.

Magnetic Surveys on the Scotian Shelf

Figure 11 shows the Halifax aeromagnetic sheet (Geol. Surv. Can., Aeromagnetic Map 7031G). Onshore the Meguma is delineated by the linear magnetic anomalies oriented parallel to the coast which are produced by pyrrhotite-rich zones within the Halifax formation. The Goldenville quartzite produces relatively weak magnetic anomalies which in general do not show the same linearity as do the magnetic anomalies associated with the Halifax formation. To the southwest of Halifax is an area of Devonian granite which

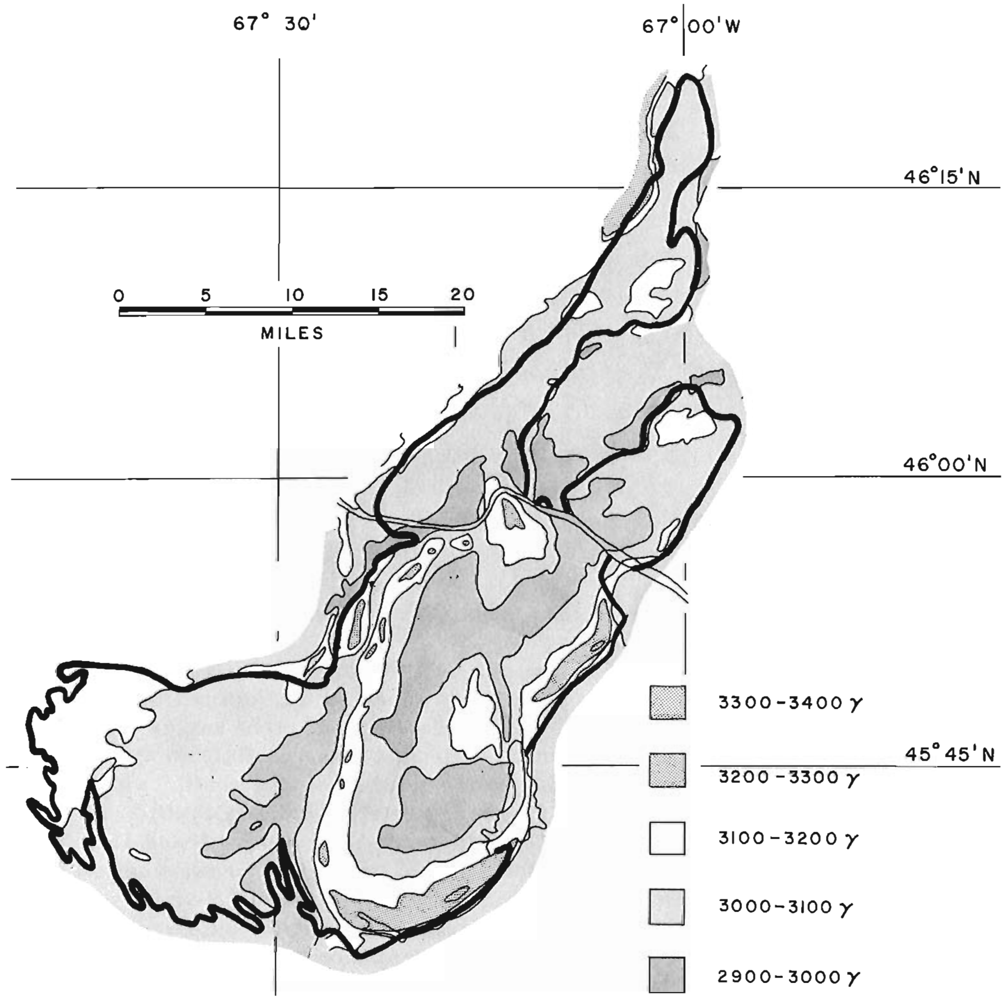


Figure 7. Aeromagnetic map of the Pokiok intrusion, contour interval is 100 gammas.

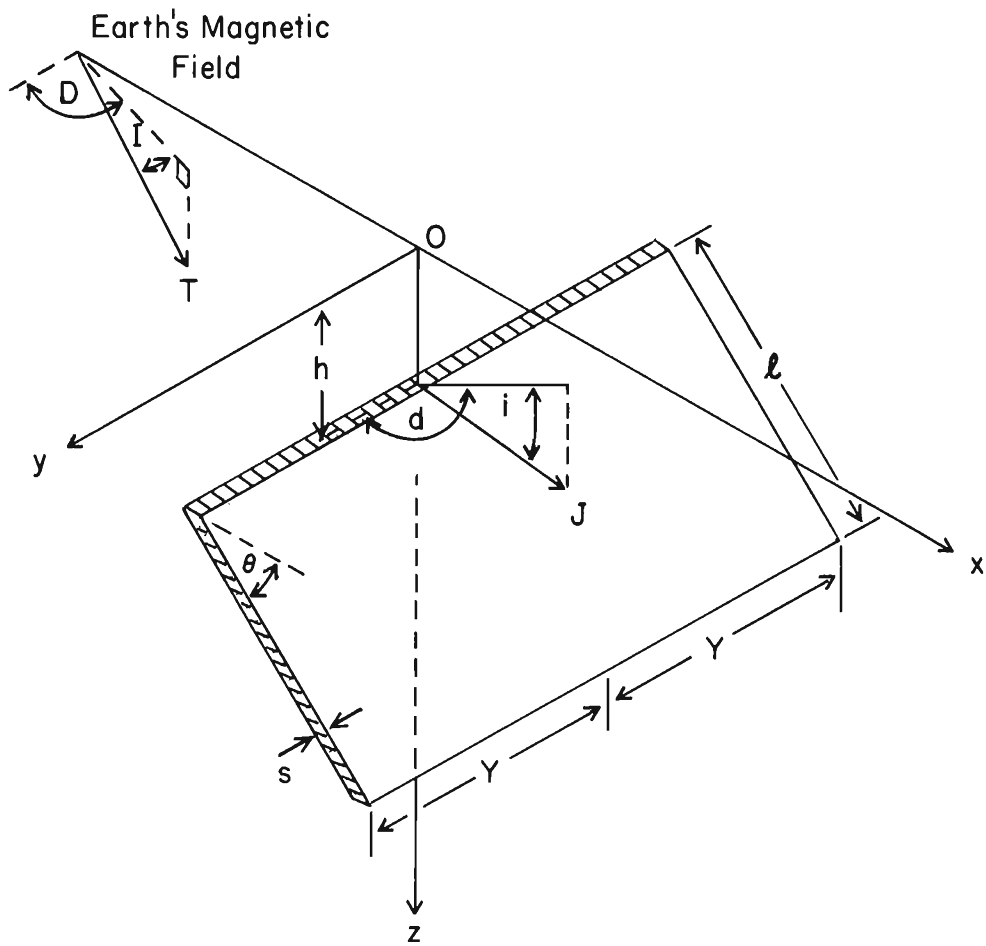


Figure 8. Oblique view of thin plate showing nomenclature used.

truncates the Meguma. The Meguma can be seen to extend offshore, and there are also several circular granite intrusions apparent on the map.

Figure 12 is an aeromagnetic profile flown at right angles to the coast of Nova Scotia across its continental shelf just east of Halifax (Hood and Bower, 1966). Included on the figure is the total intensity profile with the regional gradient removed, a high-frequency bandpass digitally-filtered trace, a low-frequency bandpass digitally filtered second-derivative trace and the underlying bathymetric contour and interpreted geological cross-section. It can be seen that the amplitude of the anomalies on the total intensity profile gradually decrease with the increasing depth to the crystalline basement. The high-frequency bandpass filter is designed to enhance the magnetic effects of the near-surface sediments, whereas the low-frequency second-derivative signature discriminates against shallow causative bodies, and augments the anomalies due to more deeply-buried sources such as the crystalline basement.

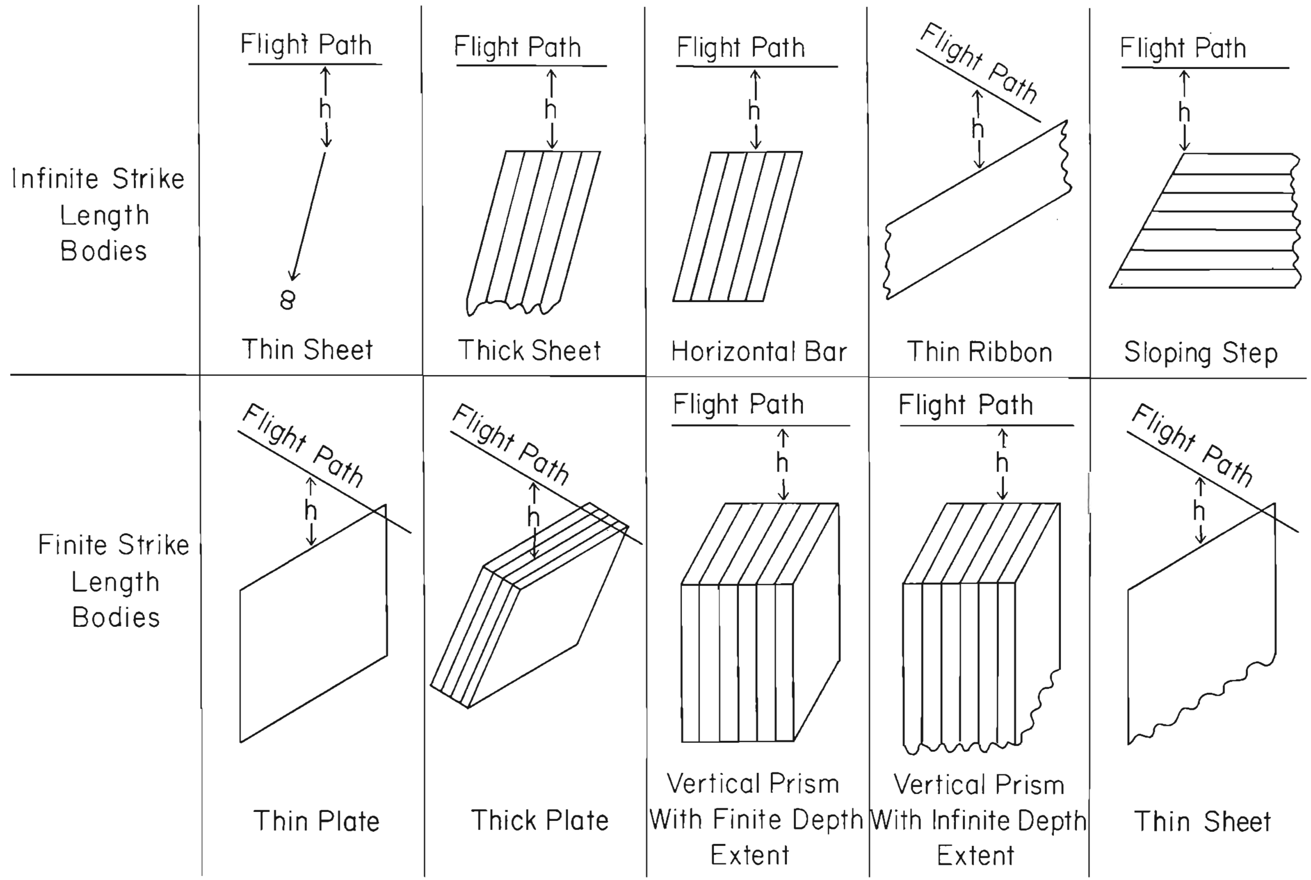
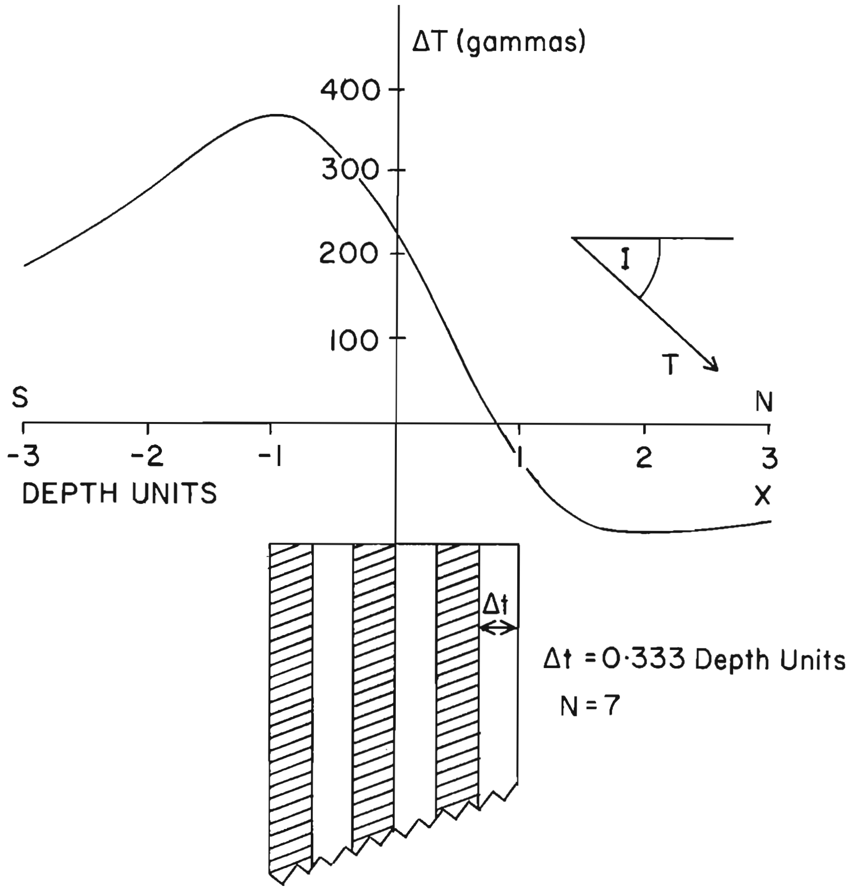


Figure 9. Geometrical models that may be generated from the thin-plate model by numerical integration.



$$\Delta T(x) = 2Jbc \sin\theta \, \frac{\Delta t}{3} \sum_{i=1}^{N} c_i \left[\text{Thin Plate Equation}\right]$$

Figure 10. Cross-section of vertical prism showing derivation of anomaly by numerical integration of thin plate equation.

Thus the edge of the Meguma Group of metamorphic rocks may be inferred from the second-derivative trace, and this interpretation is supported by the fact that the characteristic pattern of linear total intensity contours on the Halifax aeromagnetic map which parallel the coastline terminates at the same location. The left-hand part of the high-frequency filtered trace has been omitted because of the large amplitude spikes which are evident where the crystalline basement is shallow. Some idea of the background noise level of the system can be estimated from the amplitude of this trace over the deep water on the right-hand side of the profile. Note that there is an increase in the amplitude of the high-frequency anomalies close to the edge of the continental shelf.

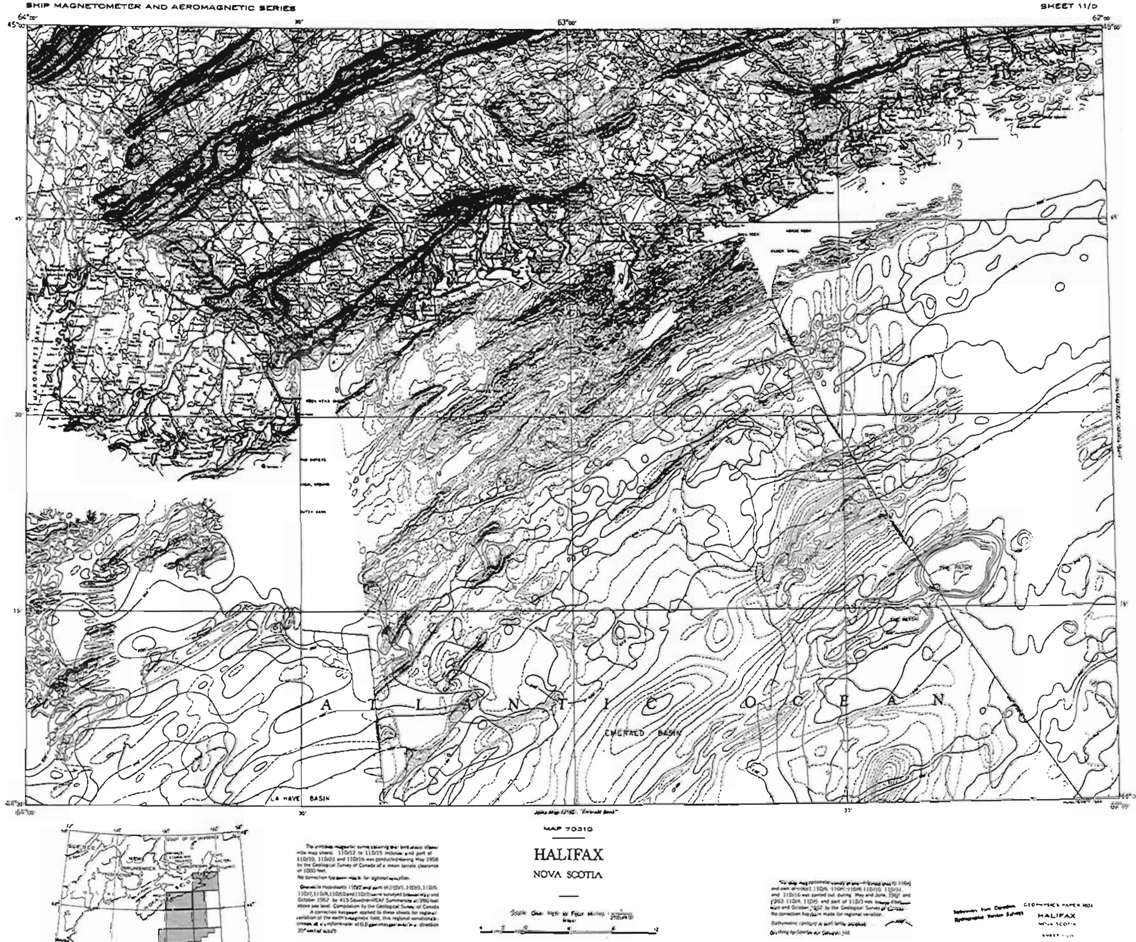


Figure 11. Halifax aeromagnetic map (Aeromagnetic series, Map 7031G, Geological Survey of Canada, 1966).

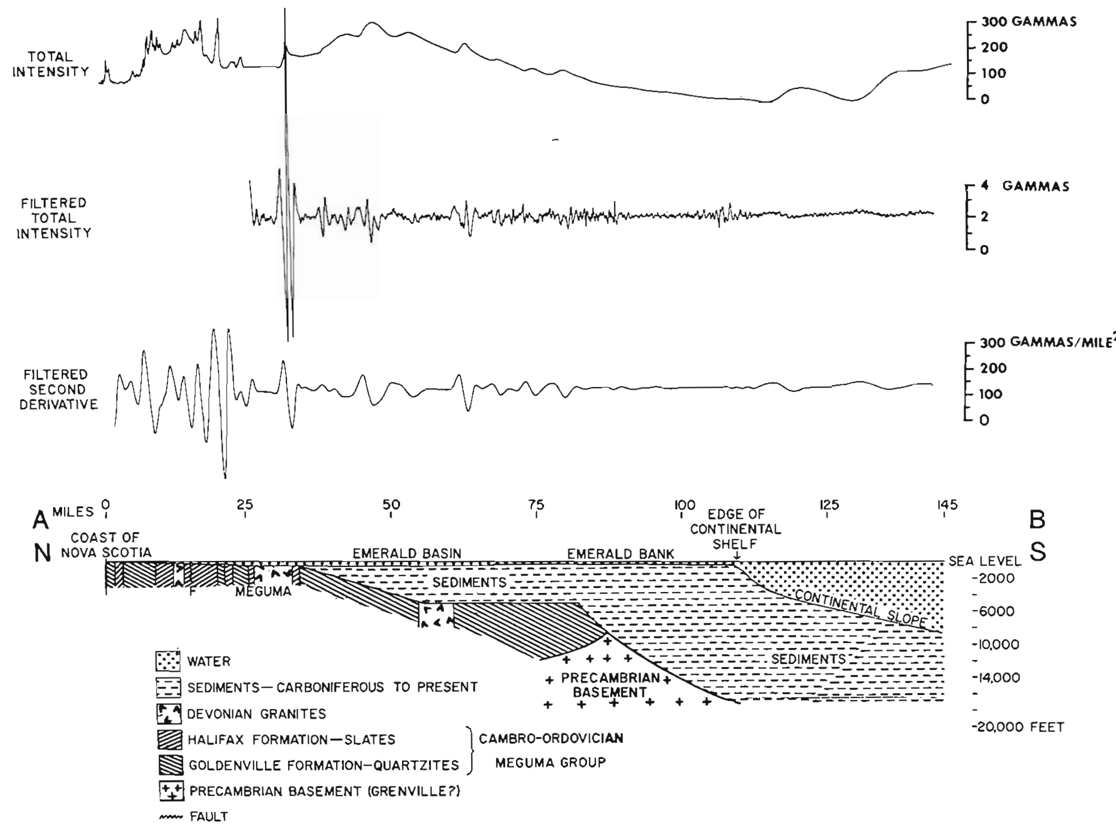


Figure 12. Aeromagnetic profile data and inferred geological cross-section across the Scotian Shelf (after Hood and Bower, 1966).

Figure 13 shows an interpretation of the geophysical results from the Scotian shelf (Hood, 1967). The inferred edge of the Meguma group of rocks which has been obtained from the magnetic maps is given at the top of the figure, together with several granite intrusions which are also indicated. Magnetic depth determinations have been indicated by the letter M and those obtained by seismic refraction investigations by the letter S (Officer and Ewing, 1954). In general the sedimentary section increases in depth out to the edge of the continental shelf and would appear to be deepest in the vicinity of Sable Island where a 15,106 foot hole was drilled by Mobil Oil in 1967 without reaching igneous basement.

Magnetic Surveys in the Gulf of St. Lawrence

Thirty-four magnetic anomalies from the southern part of the Gulf of St. Lawrence were interpreted using the previously described multimodel computer technique. The positions of the magnetic depth determinations are indicated by closed triangles in Figure 14. The closed circles represent the locations from which drilling information was available (Howie and Cumming, 1963; Howie, 1973). These basement depth values were contoured and the results are shown in Figure 15. The contours in the northeastern part of the figure were taken from seismic depth determinations published by Hobson and Overton (1973).

Tectonically the southern half of the Gulf of St. Lawrence is situated in a mobile zone, the Fundy Geosyncline (Poole, 1967) which is bounded on the northwest and southeast by the New Brunswick and Nova Scotia platforms. The Carboniferous comprises six sedimentary rock units which are predominantly continental, except for the Windsor Group which is mainly marine. Hacquebard (1972) on the basis of palynostratigraphic studies has extended the age of the groups downward to Middle Devonian and upward to Lower Permian.

The magnetic depths appear in general to be related to the pre-Carboniferous basement surface which outcrops in the Miramichi and Cape Breton Highlands (Williams et al., 1972). To the east of the Miramichi Highlands the Carboniferous rocks gradually increase in thickness to about 3.5 kilometres in the vicinity of 63° 25'W longitude. At this point a northerly-trending sloping step magnetic anomaly occurs indicating a near vertical fault. This prominent fault, herein called the Bradelle Bank fault, extends from approximately 46° 50'N to 47° 45'N latitude. The eastern side of the fault appears to be downthrown about 2.5 kilometres relative to the western side bringing the non-magnetic Carboniferous sediments into juxtaposition with the more magnetic pre-Carboniferous basement rocks. Additional evidence in support of the existence of the Bradelle Bank fault is the complete loss of the finer detail, i.e. short wavelength variations, from the magnetic field on the eastern side of the proposed fault.

Southeast from the fault the pre-Carboniferous basement gradually deepens from 6 kilometres to 15 kilometres into the proposed axial zone of the northeasterly-trending Magdalen Basin which parallels the northwest coast of Cape Breton Island. This basin probably extends to the northeast as far as St. Georges Bay, Newfoundland where Baird and Cote (1964) reported a composite thickness of 7 kilometres for the Carboniferous sequence. Towards the southwest the basin divides into three parts separated by uplands. One depression of graben extends towards Malpeque Bay, another extends through the Northumberland Strait into Cumberland Basin, and the final depression extends into George Bay near Antigonish, Nova Scotia.

An anomalous result in the magnetic depth determinations is the 4.1 kilometre value determined for the magnetic anomaly centred over Miramichi Bay. Howie and Cumming (1966) have interpreted a northeasterly-trending graben in this area from a consideration of this magnetic anomaly, and seismic and gravity data. The present writers feel however that the Miramichi Bay magnetic anomaly is probably caused by a intrabasement feature since there are no sloping-step anomalies as might be expected at the northern and southern contacts of a graben. There is also the possibility of other intrabasement magnetic sources under the Gulf of St. Lawrence, so that some calculated depths in that area may not in fact represent actual sedimentary thicknesses.

Because of the large wavelength of the magnetic anomalies over the axial zone of the Magdalen Basin there are only a few depth determinations with which to delineate the basin. The magnetic data do suggest however that the axial zone is continuous rather than being composed of a series of isolated basins. There is one drill hole in the axial region which is situated at 61° 37'W longitude and 46° 38'N latitude (Howie, 1973). It was drilled to a depth of 3.5 kilometres and ended in the lower Riversdale formation. This appears to be the only geological evidence that a 15 kilometre deep basin may exist in this area.

Goodacre and Nyland (1966) have interpreted the gravity field in the Gulf of St. Lawrence and note that the gravity anomalies over the Gulf have patterns similar to those observed on land, and hence are likely to have similar sources. For instance, the gravity field suggests that the Kingston lineament may be extended across Prince Edward Island into the Gulf. Goodacre and Nyland have determined that the causative body associated with this particular gravity high in the Gulf is situated at a depth of 6 kilometres which is in agreement with the magnetic data and with Willmore and Scheidegger's (1956) seismic results in the same area. Goodacre and Nyland attribute the negative gravity anomalies in the Gulf to either low density Devonian granite masses within the pre-Carboniferous basement or to low-density Carboniferous strata. The gravity data show a mass deficiency between the Magdalen Islands and Cape Breton Island which could be caused by the proposed axial zone of the Magdalen basin. Watts (1972) shows that density contrasts within and variations in the thickness of the Carboniferous strata can explain many of the negative gravity anomalies in the Gulf. He also showed that some of the smaller negative gravity anomalies are probably caused by salt diapirs. Watts interprets the large wavelength magnetic anomaly east of the Magdalen Islands (region M on Watts Figure 4) as being due to a rise in the pre-Carboniferous basement from 3 to 6 kilometres to a depth of the upper surface of 1.8 kilometres. Because of the smoothness of this anomaly the present writers prefer to treat the anomaly as being caused by a deep source, i.e. one due to a susceptibility contrast within the basement. Generally the anomalies produced by large shallow sources are irregular because of the magnetic inhomogeneity of the causative body. We have calculated a depth of 12 kilometres for this body.

The magnetic results indicate a configuration of the pre-Carboniferous basement northwest of Cape Breton Island considerably different to that given by Sheridan and Drake (1968) and Hobson and Overton (1973) from seismic surveys. However in the region between Prince Edward Island and the Magdalen Islands, the magnetic depths are similar to the seismic depths obtained by Willmore and Scheidegger (1956) for the pre-Carboniferous

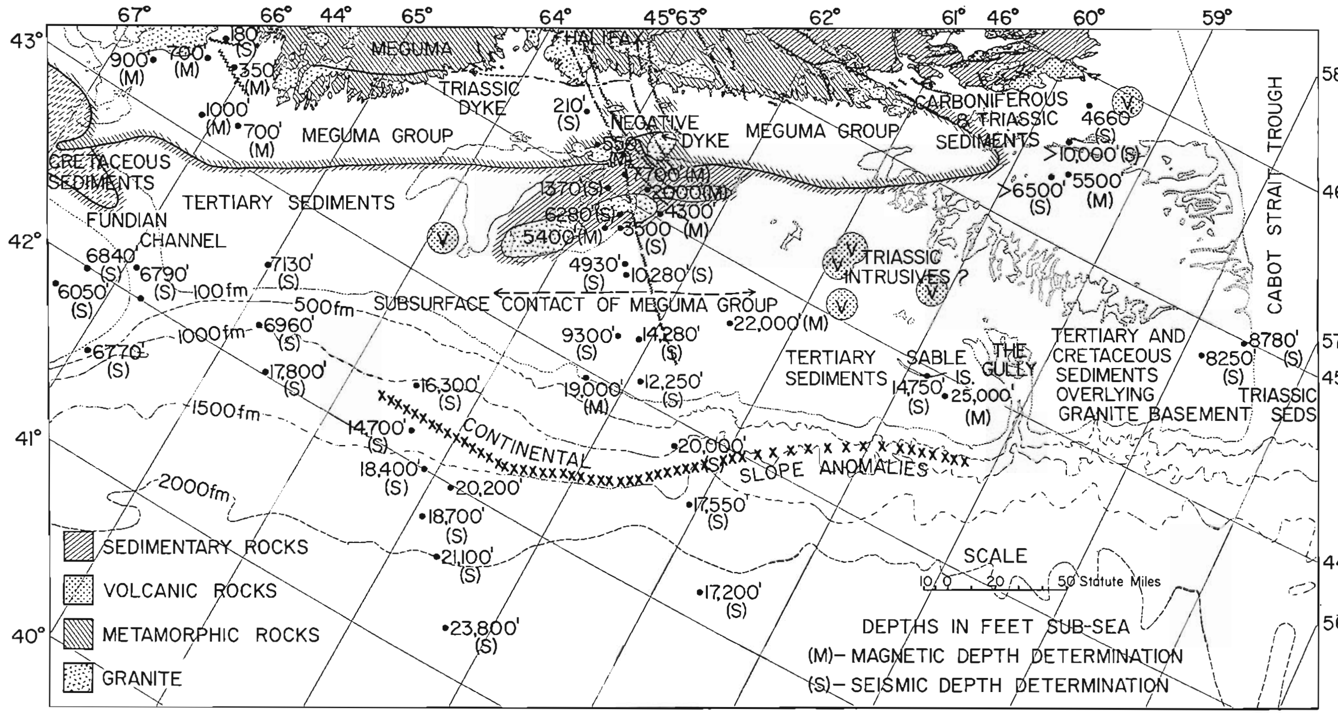


Figure 13. Subbottom geological features on the Scotian Shelf.

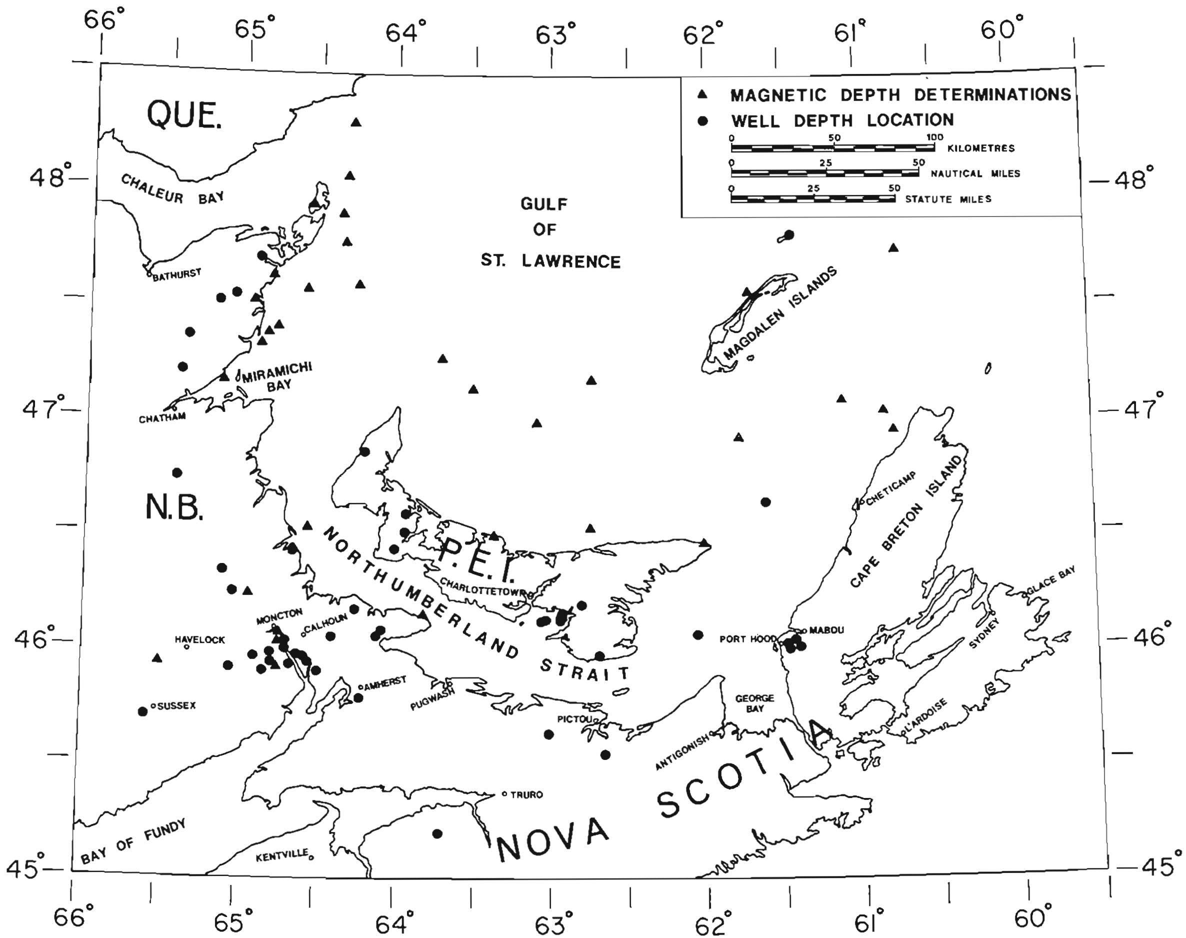


Figure 14. Southern Gulf of St. Lawrence showing locations of magnetic depth determinations and well log information.

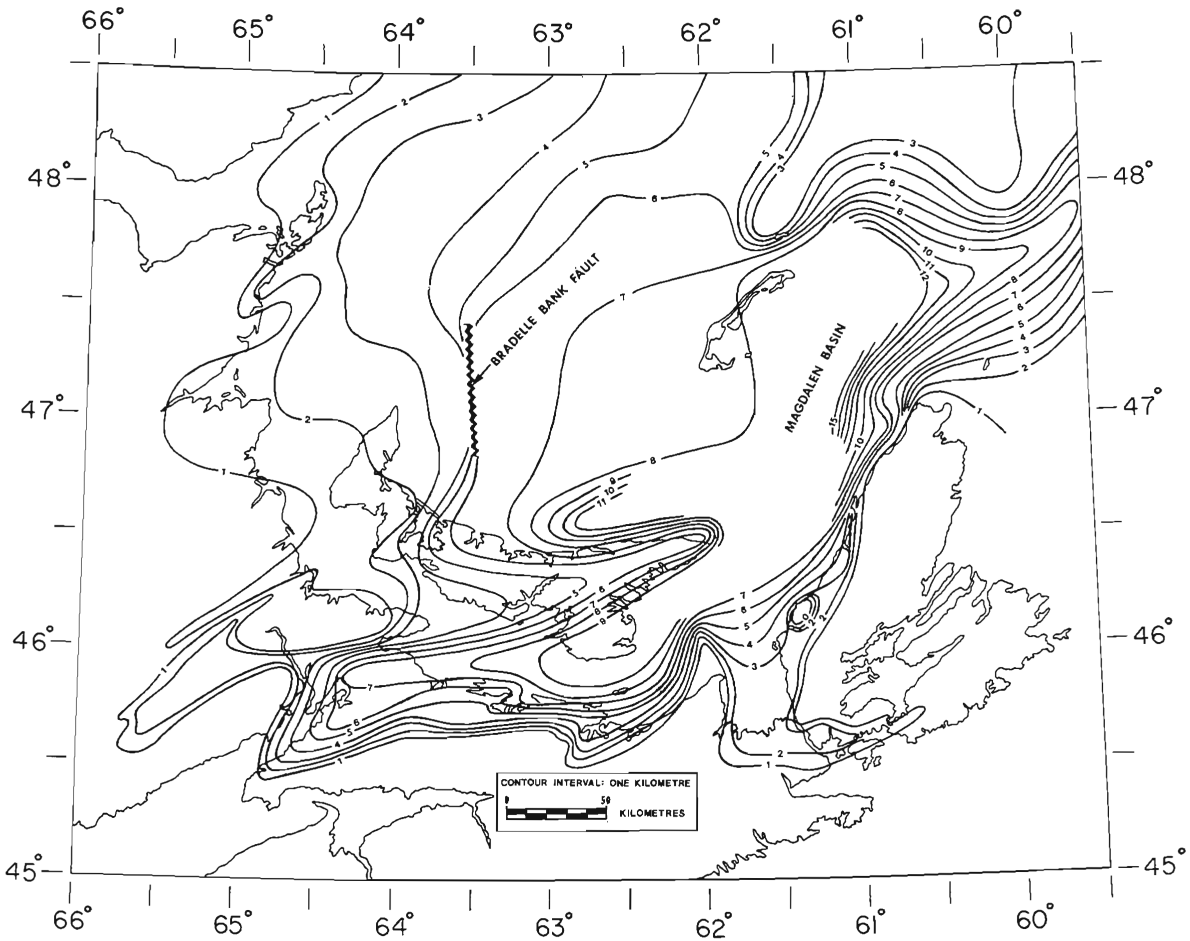


Figure 15. Map of the Precarboniferous basement surface. Contours onshore after Howie and Cumming (1963) and contours in northeast part of map after Hobson and Overton (1973).

basement. Sheridan and Drake assumed that velocities greater than 5.6 km/sec were indicative of basement in the Gulf. Apparently there are higher velocity sediments in the lower part of the Carboniferous section. A similar situation seems to have occurred for the seismic results under Sable Island where Berger et al, (1965) have reported a depth of 14,750 feet to crystalline basement, whereas Mobil Oil Company (McIver, 1972) have drilled to a depth of 15,106 feet on the island, the hole ending in Cretaceous sediments.

The Magdalen Basin probably formed during a critical time in the development of the Appalachian region of Canada. Schenk (1973) states that in North America this was a time when oceans were opening and closing. The axial zone of the basin may represent a rift zone as has been suggested by White (1972) for the Cumberland Basin, or it may be an aulocogen (Nalivkin, 1965).

SUMMARY

It would appear that in both areas discussed in this paper, namely the Scotian Shelf and the Gulf of St. Lawrence, that depth determinations obtained by the magnetic and seismic refraction survey techniques do not agree. A similar situation has occurred in Hudson Bay, see for instance the papers in the Earth Science Symposium on Hudson Bay (Hood, 1969). In general, the seismic refraction depths are shallower than those obtained from the magnetic field values, and it would seem that they usually represent depths to high-velocity refractors within the section above the crystalline basement. Thus magnetic survey results should always be interpreted quantitatively to provide an independent check.

ACKNOWLEDGMENTS

The writers wish to thank R.D. Howie for additional well log information, and L.M. Cumming for assistance during the preparation of this paper.

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(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 359-370)

23. MAGNETIC AND TELLURIC MEASUREMENTS IN ATLANTIC CANADA

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Abstract

Magnetic and electric field measurements in Atlantic Canada have been analyzed for variation periods from 20 seconds to 24 hours using vertical field attenuation, preferred plane or transfer function and magnetotelluric methods. The response of models have been obtained by simple theory and by network analysis using transmission line analogy. The results can largely be explained by electromagnetic induction in shallow continental shelf and Gulf of St. Lawrence seawater (maximum response at about 1-minute period) and in deep ocean seawater (maximum response at 30- to 60-minute period). The notable exceptions are the results for the north shore of the lower St. Lawrence River which suggest a deep resistivity contrast between the Appalachian and Canadian Shield geological provinces, and the results for the coast of Nova Scotia and Newfoundland which indicate a thick low resistivity layer beneath the continental shelf. The existence of the layer can be seen from the attenuation of the vertical field components at Sable Island near the shelf edge and by the large preferred plane or transfer function coast effect that occurs at coastal stations such as Dartmouth and St. John's rather than, as expected at the shelf edge. The layer appears not to extend much to the southwest of the Gulf of Maine but may extend into the Bay of Fundy. We have tentatively explained the low resistivity by highly saline solutions associated with an evaporite or salt layer near the base of the sedimentary section.

INTRODUCTION

Electrical methods have considerable potential for delineating sedimentary structures as well as for determining the electrical resistivity distribution at great depths in the earth. These methods have been used extensively to prospect for metallic orebodies in North America (e.g. see Hansen et al., 1966) but have been little used for oil exploration. In contrast, electrical methods have been used extensively in U.S.S.R. (e.g. Berdichevskiy, 1965) and rather widely by the French petroleum industry to outline sedimentary structures (e.g. Kunetz 1957; Migaux et al., 1955). A few detailed surveys have been reported in North America, for example, the recent magnetotelluric survey carried out by Texas Gulf Sulphur Company in northern British Columbia to map the Triassic and Paleozoic sections (Pamenter, 1971).

Electrical methods do not have the resolution of seismic measurements, generally providing information on regional layering although detailed electrical surveys can outline large highly conductive or resistive bodies such as salt domes or reefs. Because of their low resolution, electrical measurements are most useful in reconnaissance studies. They also can provide information that cannot be obtained in any other way such as detecting the presence of evaporite layers that can give rise to salt domes.

The electrical resistivity of sedimentary rocks depends primarily on their porosity and the salinity of the interstitial fluid. The composition of the solid grains is usually important. Temperature is important above several hundred degrees centigrade and is probably the most important parameter for dry rocks in the deep crust and upper mantle (e.g. Parkhomenko, 1967; Hyndman et al., 1968). The possibility of determining temperatures from electrical resistivity estimates has generated considerable interest. Differences in porosity within sedimentary sections can produce resistivity variations of a factor of 1,000 (about 1 to 1,000 ohm-m). The presence of highly saline solutions can make the differences even larger (e.g. the Paleozoic sections in British Columbia determined by magnetotelluric measurements). A very large contrast in porosity and thus resistivity usually occurs between the sediments and the crystalline basement.

Artificial source electrical methods normally give information to depths of a few kilometers at most. However, the electromagnetic fields produced by currents in the ionosphere can be used to obtain resistivity structure to depths of hundreds of kilometers. Three types of measurement and analysis are used with these fields. Measurements of telluric currents (induced by ionospheric sources) examine just the pattern of surface electric potentials. The current can be observed to diverge away from resistive and toward conductive bodies. Very little can be learned about the depth of the structures or the absolute resistivities. The magnetotelluric method (Cagniard, 1953) involves measurement of time variations in both electric and magnetic fields. The ratio of electric to magnetic fields is computed as a function of frequency. The ratio is related to the resistivity as a function of depth in the earth. It is compared with the ratio with frequency curve for theoretical models, usually for a horizontally layered earth. The theory and analysis are difficult if the structure is not at least approximately horizontally layered. The third method, geomagnetic depth sounding utilizes the variations in the three orthogonal magnetic field components only. Horizontally layered structure can be estimated from the ratio of the vertical (Z) to horizontal (H) magnetic variation amplitudes with frequency (e.g. Caner et al., 1967). This method utilizes the tendency for the conducting earth to bend or refract the incident electromagnetic waves to a horizontal plane, reducing the vertical field component (Fig. 1a).

Transfer function analysis (preferred plane) estimates the systematic dip angle and direction of the variation field lines resulting from lateral conductivity contrasts (e.g. Parkinson, 1959; Schmucker, 1964; Everett and Hyndman, 1967). Field lines will tend to bend around and not penetrate good conductors. Averaged over time, away from auroral zones, the incident variation field lines are nearly horizontal. A simplified representation is given in Figure 1b. For all of the methods, the lower the frequency (longer variation period) the greater is the depth of penetration.

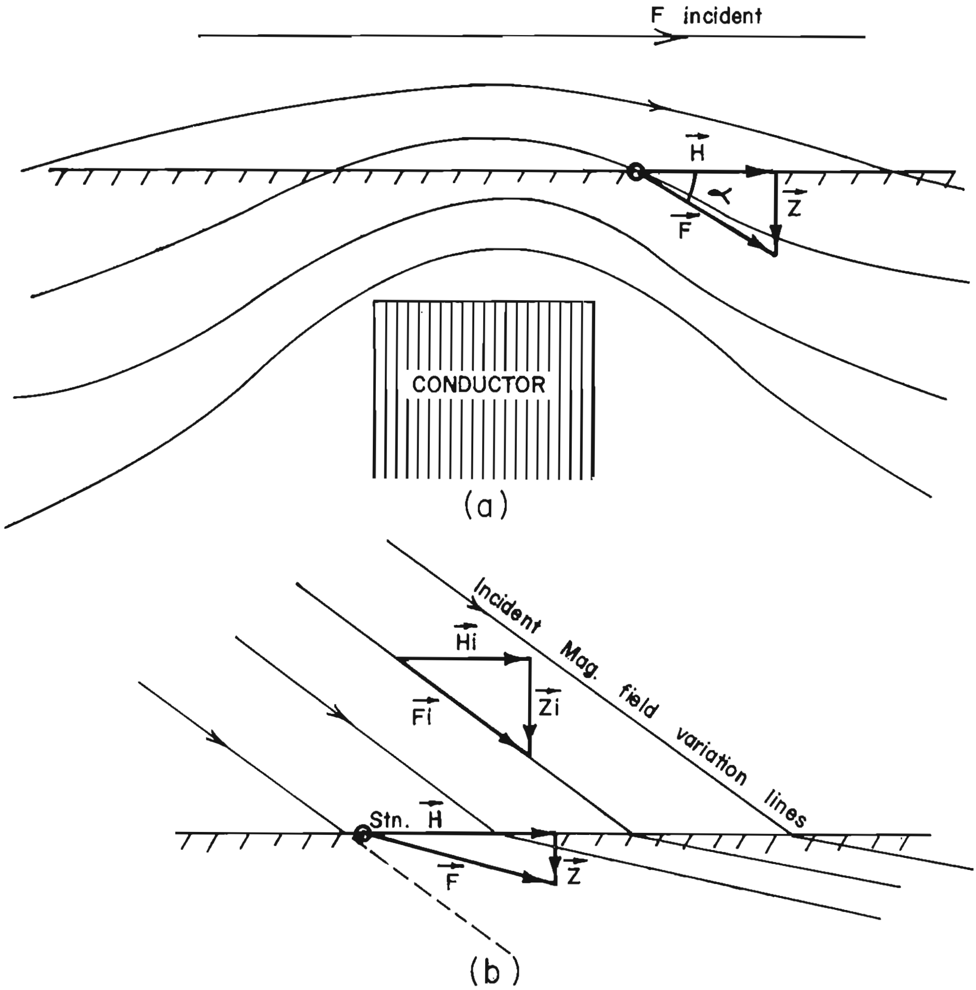

Figure 1 (a) Simplified representation of the transfer function in terms of the systematic dip angle α, where F is the total magnetic field, H its horizontal component and Z its vertical component.

(b) Bending of the indicident field lines towards the horizontal by the conducting earth (depth sounding).

At the continental margin the most obvious electrical conductivity contrast is between the highly conducting seawater and the more resistive rocks. This contrast controls the amplitudes of the surface electric fields, particularly on land, making magnetotelluric measurements difficult to interpret. Geomagnetic depth sounding using the reduction or attenuation of the vertical field involves a large uncertainty from the lateral resistivity contrasts but still gives very useful information as does transfer function (preferred plane) analysis. We have also made measurements that show electric currents produced by tidal water flow through the earth's steady main magnetic field.

We have not yet tried to use these currents to estimate earth electrical conductivity structure. These results will be reported elsewhere. In addition to the resistivity contrast between land and sea, one expects the continental shelf sediments to have a resistivity between those of the sea and land. There is also a difference in depth to the highly conducting low-velocity partial melt layer under oceans and continents. It occurs below about 70 to 100 km under oceans and below 150 to 200 km under stable continental areas. Thermal arguments also suggest higher temperatures and thus lower resistivity at depth under oceans.

The present paper summarizes briefly our findings from the measurements carried out over a four-year period in the eastern part of Canada to determine the resistivity distribution below the land and ocean.

Measurements and Results from Atlantic Canada

The magnetic and electric field variation stations we have studied are shown in Figure 2. Table 1 contains their positional information and symbols used for them in all the illustrations here. Some of these were recorded by the Earth Physics Branch of the Department of Energy, Mines and Resources, Ottawa. Not included are a recent detailed profile of magnetotelluric measurements across western Newfoundland (J.A. Wright, personal communication), and a recent unpublished series of magnetic and telluric measurements around the northern Gulf of St. Lawrence (R.N. Edwards, personal communication). Our data includes short-period three-component magnetic variations (1 cycle in 20 sec. to 1 cycle in 2 hrs), total-field amplitudes of daily magnetic variations (1 cycle in 12 or 24 hrs, also used to correct magnetic surveys), and electric or telluric field measurements on land and at sea.

The presence of low resisitivity under Sable Island was first detected by comparing three-component magnetic variation measurements recorded simultaneously at Sable Island, Dartmouth, N.S., (or Halifax) and Fredericton, N.B. (Srivastava and White, 1971). The amplitudes of the vertical fields were much lower on Sable Island compared to the other stations for frequencies from 1 cycle in 5 minutes to 1 cycle in 4 hours (Fig. 3). The horizontal fields were found to be similar at all three stations. Further measurements have confirmed this pattern. Total field amplitudes for daily variation are also low on the Island which shows that the effect exists at very low frequencies (Srivastava, 1971). These results are in contrast to both theory and measurements on other coasts which have the largest vertical fields at the shelf edges since this is usually where the main resistivity contrast occurs (Roden, 1964; Parkinson, 1962; Schmucker, 1964). These results from the Atlantic coast can only be explained by low resistivity under the shelf such that there is little resistivity contrast between the shelf and land.

Vertical transfer functions (preferred planes or systematic dip angles) were computed as a function of frequency in the manner described by Cochrane and Hyndman (1970). The part of the vertical field that is statistically correlated with the horizontal field is estimated as a function of frequency. The transfer function amplitude is the ratio of this correlated part of the vertical component to the amplitude of the horizontal components for the horizontal field in the direction of maximum correlation. It is usually assumed that the

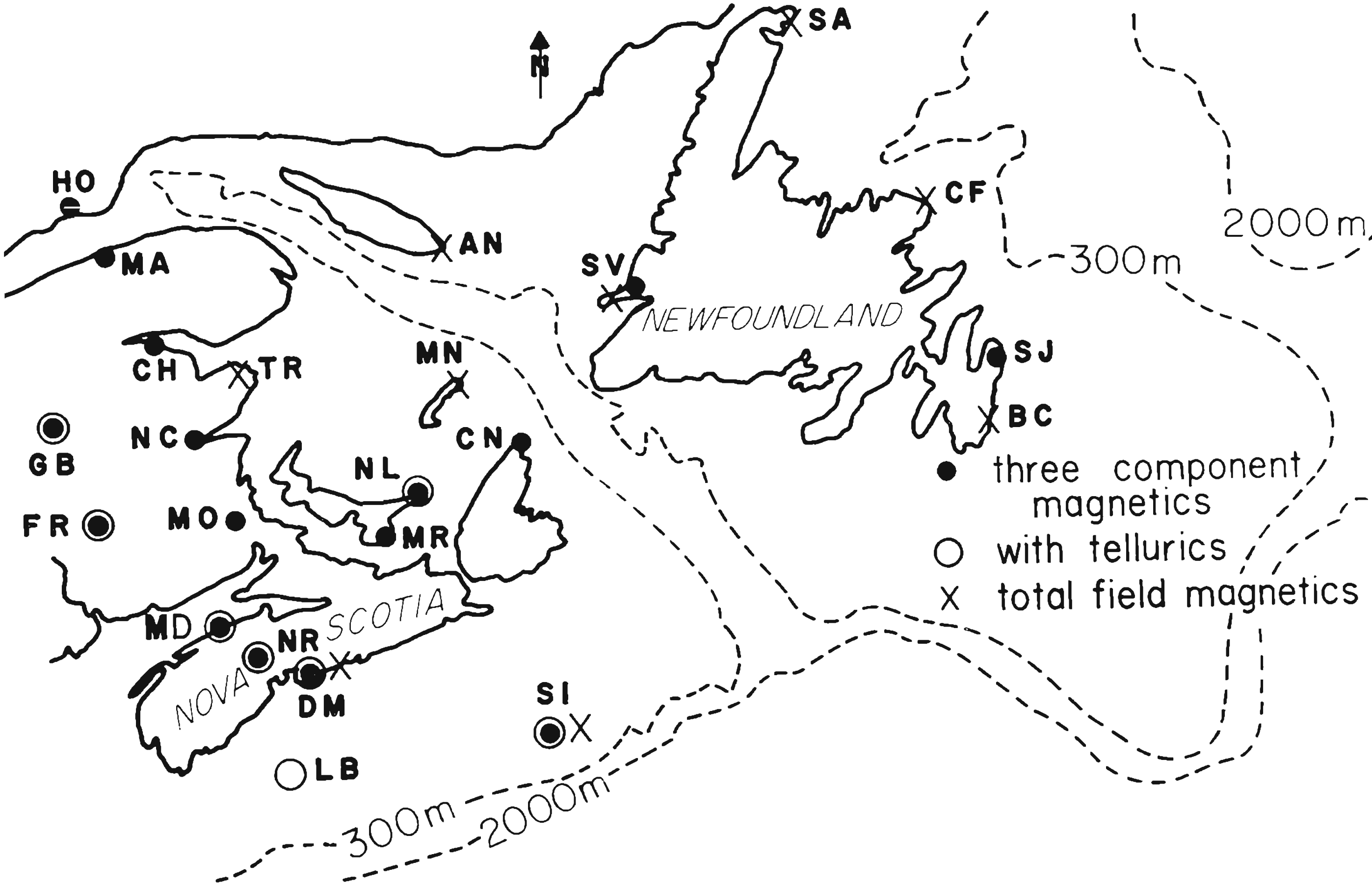

Figure 2. Map showing the location of stations in the Atlantic Canada where magnetic and electric field variation measurements have been made.

Table 1

Location of the Stations

Station	Symbol used	Latitude	Longitude
Anticosti Island	AN	49.133° N	60.672° W
Bear Cove	BC	46.940° N	52.894° W
Cape Freels	CF	49.251° N	53.473° W
Cempbellton	CH	48.000° N	66.666° W
Cape North	CN	47.017° N	60.600° W
Dartmouth	DM	44.685° N	63.614° W
Fredericton	FR	45.966° N	66.805° W
Gordonsville	GB	46.483° N	67.516° W
Hauterive	HO	49.020° N	67.800° W
Le Have Basin	LB	43.500° N	63.666° W
Matane	MA	48.850° N	67.538° W
Morden	MD	45.117° N	64.917° W
Magdalen Islands	MN	47.633° N	61.525° W
Moncton	MO	46.117° N	64.833° W
Murray Harbour	MR	46.050° N	62.566° W
Newcastle	NC	47.000° N	65.583° W
North Lake Harbour	NL	46.466° N	62.000° W
New Ross	NR	44.717° N	64.500° W
St. Anthony	SA	51.371° N	55.600° W
Sable Island	SI	43.980° N	59.750° W
St. John's	SJ	47.583° N	52.683° W
Stephenville	SV	48.537° N	58.583° W
Tracadie	TR	47.500° N	64.920° W

correlated part of the vertical component results from currents induced by the regional horizontal component. Figure 4 shows the Parkinson type arrows (Parkinson, 1962) for periods of 60, 20 and 5 minutes. Their lengths are proportional to the absolute amplitude of the transfer function, and their directions are opposite to the direction of maximum horizontal to vertical correlation or maximum systematic dip angles. The arrows will generally point toward electric current concentrations. The details of this analysis is given by Hyndman and Cochrane (1971).

Figure 4 shows systematic changes in the amplitude and direction of the arrows. The inland stations show large systematic dip angles at shorter periods (less than 20 minutes), resulting from the contrast between the resistivity of the land and shallow water (with some fairly resistive sediments). At periods longer than 5 minutes the magnetic field variations penetrate and are little affected by the shallow water. Our model computations (Hyndman and Cochrane, 1971) for currents induced in the channels show that the measurements can be explained without any major resistivity contrast between land and the rocks underlying the Gulf of St. Lawrence. The one exception was the Houterive station on the north side of the St. Lawrence River where the results indicate a deeper contrast probably associated with the contact between the conductive Appalachian and more resistive Grenville geological provinces.

The transfer function is very small at all periods at Sable Island while there are large coast effect transfer functions at 60-minute periods at Dartmouth and St. John's, contrary to both theory and measurements at other coasts, confirming our other analysis which indicated the major resistivity contrast is between the land and shelf rather than between the shelf and deep water. There must be a highly conductive layer under the shelf.

We have computed the magnetic and electric field variations to be expected for different structures using numerical network analysis of the field equations (transmission line analogy; Madden and Swift, 1969; Swift, 1971; Wright, 1969). Our best-fitting model across the Nova Scotia shelf, Nova Scotia, and the Bay of Fundy is shown in Figure 5. The low shelf resistivity must extend to offshore Newfoundland according to the results obtained at St. John's. It does not extend to the southwest as far as Massachusetts or the Virginia coast according to our analysis of records from Weston, Mass., Cheltenham, Maryland and Fredricksburg, Virginia. However, it may extend into the Bay of Fundy.

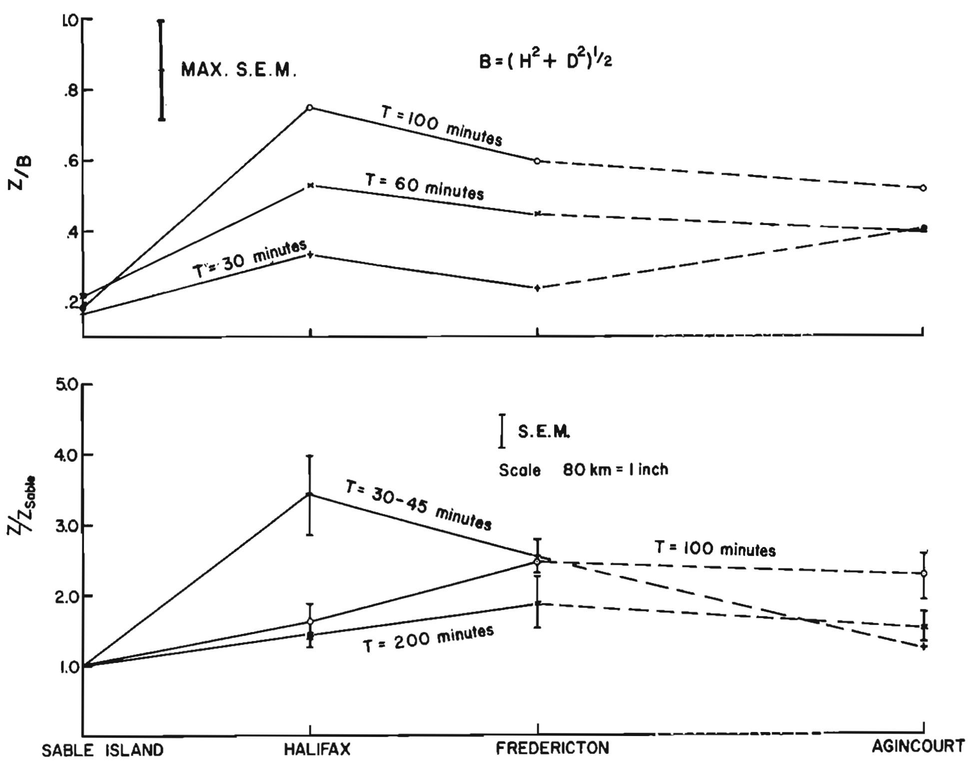

Figure 3. Attenuation of the vertical magnetic field component (Z) across the continental shelf with respect to the vertical magnetic field component at Sable Island (bottom) and the ratio of vertical (Z) to the horizontal (B) magnetic field at each station (top). The vertical bars are the standard error of the mean (S.E.M.).

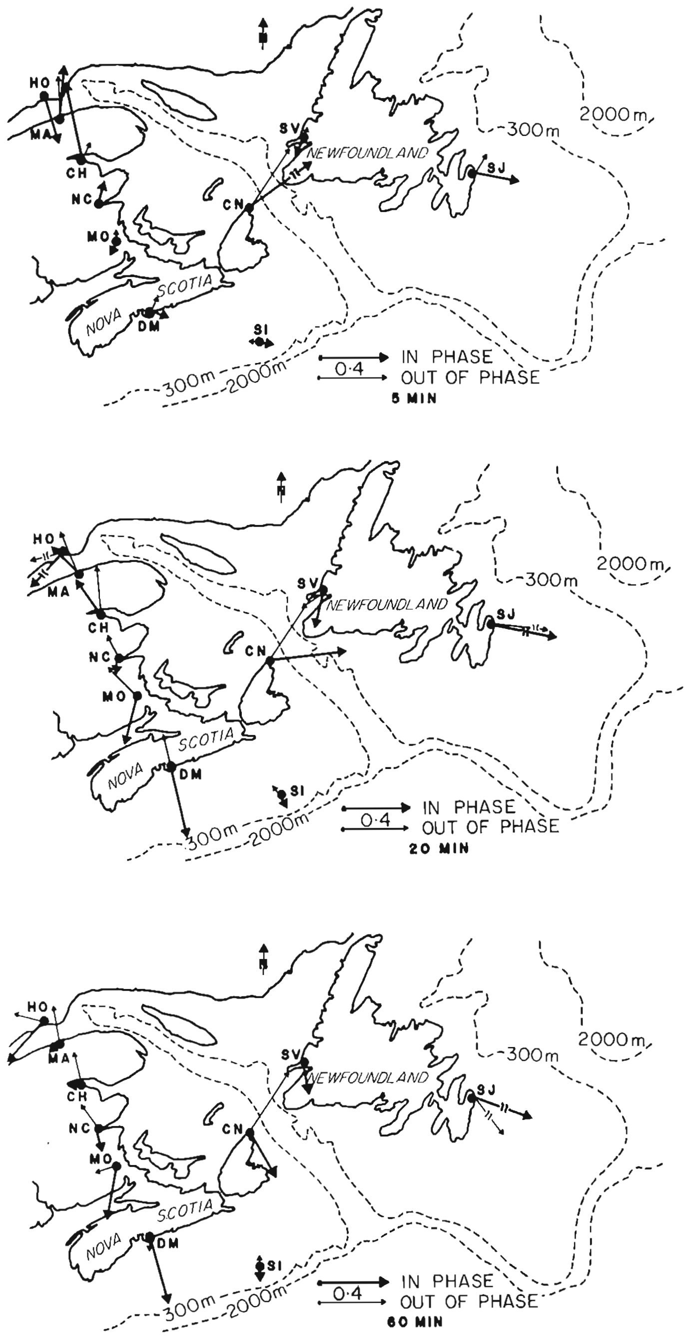

Figure 4. Transfer function arrows at various stations for in phase (thick arrows) and out of phase (thin arrows) magnetic field components for 5-, 20- and 60-minute periods.

The low resistivity under the shelf can be explained in a number of ways. It might be that the general resistivity of the sediments is low but electrical logs down to 5 km in the Mobil Oil Ltd. hole on Sable Island do not show such low values (e.g. Monro and Brusut, 1968). Early seismic measurements estimated some 5 km of sediments below Sable Island but more recent estimates exceed 10 km in thickness. Thus it seems likely that the low resistivity exists in the lower part of the sedimentary section. About 5 km of 0.1 to 1.0 ohm-m material is required. Measurements in Germany have detected sedimentary sections of this resistivity. There, the low resistivities have been explained by the presence of salt (Vozoff and Swift, 1968). As previously mentioned, saline solutions were given as the explanation for low resistivities of Paleozoic sediments in British Columbia. Evaporite diapirs appear to be very common on the Nova Scotia and Newfoundland shelf. Their source must lie in high concentrations of evaporites in the lower part of the sedimentary section. Pautot et al. (1970) have shown that the diapiric structures of the north Atlantic suggest the presence of a continuous salt layer along the margins formed during the early phase of rifting of the Atlantic. Subsidence since rifting has put the layer at considerable depth, probably greater than 5 km deep under Sable Island. It is possible that some of the salt deposits of central Nova Scotia extend under the Bay of Fundy to produce the low resistivity observed in that area.

CONCLUSIONS

The magnetic and electric field variations measured in Atlantic Canada can largely be explained by electromagnetic induction in seawater, (both the shallow shelf and Gulf of St. Lawrence water and the deep ocean) and in known sediments of normal resistivity (10-100 ohm-m). The notable exception is the continental shelf off Nova Scotia and Newfoundland which must have thick layers of very low resistivity. We have tentatively associated this low resistivity with very highly saline interstitial fluid in the bottom part of the sedimentary section. At the bottom of the section there must be

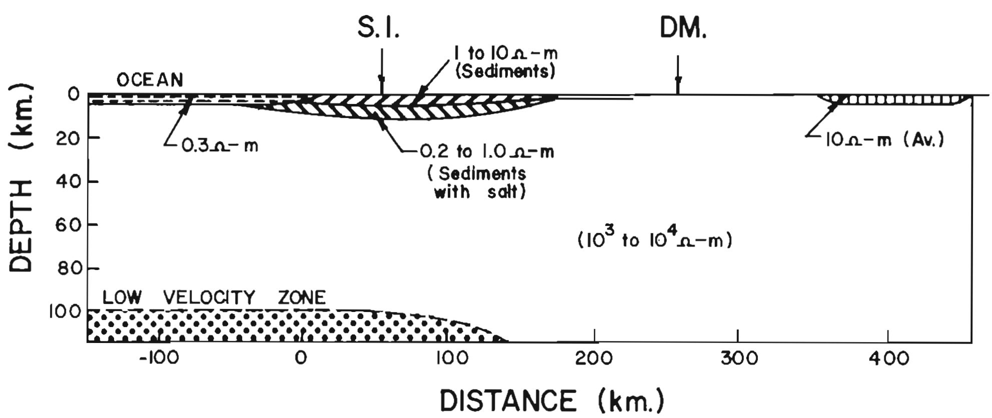

Figure 5. Simplified conductivity structure across the Nova Scotia continental shelf and Bay of Fundy as inferred from the present analysis.

high concentrations of evaporites responsible for the frequent salt domes or diapirs that are present on the shelf. A thickness of at least several kilometers with a resistivity of less than 0.5 ohm-m is required. The data suggests a depth of about 5 to 10 km under Sable Island but there is a large uncertainty. There is some evidence for similar low resistivity under the Bay of Fundy. Measurements on the Massachusetts coast and Virginia coast to the southwest suggest that the high conductivity layer is absent south of the Gulf of Maine. Low resistivity, probably associated with the seismic low velocity layer appear to occur at a shallower depth under the ocean than under the continent. One station indicates a high resistivity contrast between the Appalachian and Canadian Shield geological provinces.

Our studies have outlined the broad pattern of resistivity structure around the Atlantic continental margin, and provide basis for future more detailed measurements.

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(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 371)

24. HUDSON GEOTRAVERSE

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Abstract

New emphasis in oceanographic research is towards a systematic study of selected areas. Several recent developments have made it possible to carry out detailed surveys in distant offshore areas: i) improved means of ship positioning; ii) use of larger, specially designed ships of greater endurance; iii) development of automatic data aquisition techniques suitable for systematic surveys; iv) international co-operation and quick exchange of data. HUDSON GEOTRAVERSE is a co-operative project organized by the Atlantic Geoscience Centre to study a one-degree-wide strip of the Atlantic Ocean between the latitudes of 45° and 46° North and longitudes 18° to 60° West. The 380,000 sq. km area (one and a half times the area of Great Britain) stretches from Cape Breton, N.S. across the Grand Banks of Newfoundland to the eastern flank of the Mid-Atlantic Ridge. This Geotraverse crosses most of the major oceanic provinces recognized so far and spans the North American plate from the continent to its edge at the ridge crest.

(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 373-388)

25. GEOLOGICAL AND GEOPHYSICAL RESULTS BEARING UPON THE STRUCTURAL HISTORY OF THE FLEMISH CAP REGION

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Abstract

Flemish Cap is a continental structure, physiographically isolated from the Grand Banks of Newfoundland by the Flemish Pass. The top of Flemish Cap is a smooth, erosional surface, and seismic and magnetic results indicate that the Flemish Cap consists of a central area of basement material encircled by a zone of outward-dipping sedimentary strata. Geological sample evidence suggests that the central basement area may be underlain by a complex of eroded and intruded sedimentary rocks, possibly equivalent in age to Late Precambrian bedrock on the Avalon Peninsula of Newfoundland. Lower Cretaceous foraminifera have been identified in a limestone specimen from the southeastern flank of Flemish Cap. Flemish Pass, to the west of Flemish Cap, is underlain by a thick accumulation of sediments which is at least in part a continuation of prograding beds on the eastern Grand Banks. Buried ridges of "acoustic basement" occur on either side of Flemish Pass, with depth of burial increasing to the north. Seismic and magnetic evidence indicates that these ridges are composed of sedimentary rock.

INTRODUCTION

Flemish Cap lies due east of the Grand Banks of Newfoundland and due south of the southern tip of Greenland (Fig. 1). Although separated from the eastern Grand Banks by the deep water of Flemish Pass, water depths of less than 140 m over central Flemish Cap suggest that this feature is fundamentally a part of the continental shelf, and as such it constitutes the easternmost element of the continent of North America. The physiographic isolation of Flemish Cap poses intriguing questions as to the geological structure and history of this feature, and its structural relationships to the adjacent Grand Banks region. This paper reviews geological and geophysical data collected by the Atlantic Geoscience Centre, Bedford Institute of Oceanography that may bear upon these questions.

Seismic profiling, geological sampling, and bottom photography were carried out on Flemish Cap in May, 1969, from the C.S.S. HUDSON (Pelletier, 1969). Concurrent seismic and magnetic profiling was conducted from the C.N.A.V. SACKVILLE in July and August, 1969 (Grant, 1969). Earlier marine investigations in the Flemish Cap region include combined hydrographic and geophysical surveying from the C.S.S. BAFFIN in 1967 (Keen et al., 1970), dredging and bottom photography from the C.S.S. HUDSON in 1967 (Gilbert, 1967), and dredging on central Flemish Cap in 1965 by J. Stewart on the M/V THERON. Seismic profiler coverage on Flemish Cap and the Grand Banks is indicated by the dashed lines in Figure 2. The solid

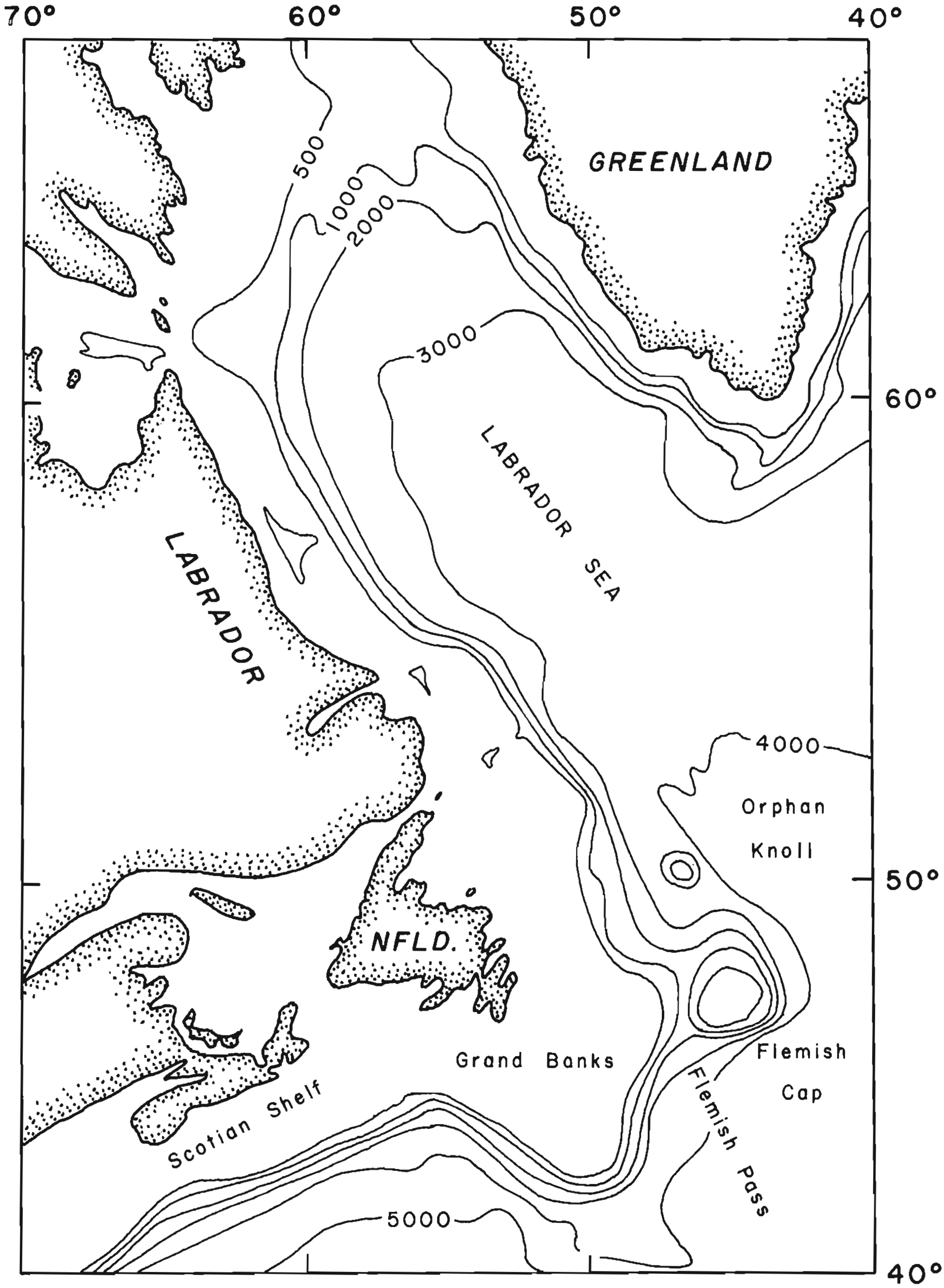

Figure 1. Index chart of the northwest Atlantic. Isobaths are drawn at 500 m, 1,000 m, and at intervals of 1,000 m thereafter.

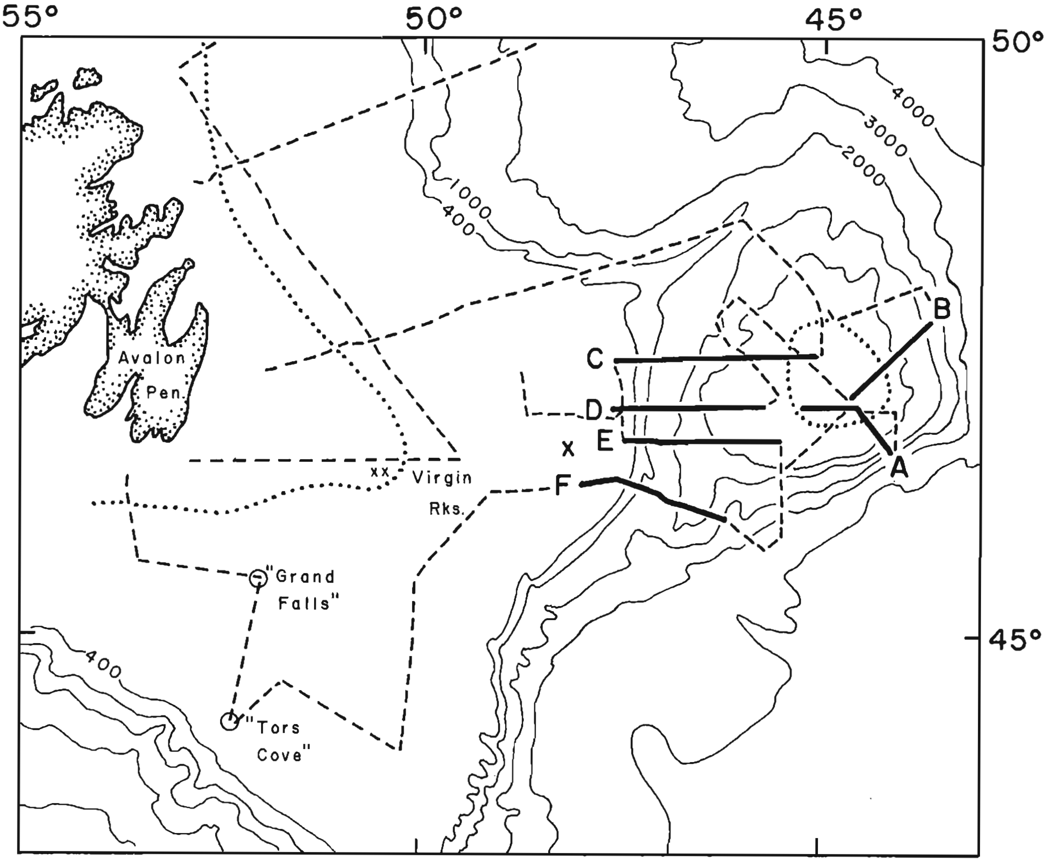

Figure 2. Bathymetry of the Grand Banks-Flemish Cap region. Contours are in metres. The dashed lines indicate the location of continuous seismic profiles. The solid portions of these lines denote sections presented in Figures 3-7. The dotted line on Flemish Cap indicates the extent of the central "basement" area. The dotted line on the Grand Banks traces the inferred landward edge of the coastal plain deposits.

portions of these lines denote sections presented in this paper. All seismic coverage was obtained with a single-channel profiling system using a 10-cubic-inch air-gun energy source. Magnetic measurements were made with a proton precession magnetometer. The total intensity magnetic profiles plotted with selected seismic cross-sections have been referred to an arbitrary zero following removal of the regional field using values from Dominion Observatory Map F-1965.0 (1961).

MARINE GEOPHYSICAL AND GEOLOGICAL RESULTS

Seismic profiler records show that the smooth, gently convex top of the Flemish Cap is an erosional surface, cut on a central area of seismically "hard" basement which is encircled by a zone of outward-dipping strata. The extent of this central basement area as inferred from the seismic data is indicated by the dotted line in Figure 2.

Sections A to F (Figs. 3-6) are generalized seismic cross-sections prepared from tracings of the original profiler records. The vertical to horizontal scale-ratio is approximately 25 to 1. No corrections have been applied for sub-bottom velocity increase. Section A (Fig. 3) is representative of seismic profiler results from the southeast quadrant of Flemish Cap. The layered media underlying the shelf-edge are truncated at both the top and flank of Flemish Cap, and a minor step in the bottom profile occurs at their contact with the seismically "hard" material of the central basement zone. Seismic penetration is occasionally recorded within the central basement area, as indicated toward the western end of section A, perhaps suggesting that this area comprises elements of eroded sedimentary strata.

On section B (Fig. 3) the erosional surface is cut nearly parallel to sub-bottom reflectors within the flank media, and there is no pronounced shelf-edge break. This section is representative of seismic profiler results from the eastern and northern sides of the Flemish Cap.

Sections C to F (Fig. 4-6) provide east-west transects across the Flemish Pass from the eastern Grand Banks to western Flemish Cap. On section C the sediments underlying Flemish Pass are continuous with prograding beds on the slope off the eastern Grand Banks. On sections D and E it is clear that these beds terminate against the western flank of Flemish Cap. Surface irregularity of these deposits on sections E and F probably reflects the action of bottom currents in controlling their deposition. The seismic character of the sediments underlying Flemish Pass is quite unlike that of the layered media on the western flank of Flemish Cap.

Sections C, D and E (Figs. 4 and 5) show a very strong seismic reflector beneath the western flank of Flemish Cap, at the level indicated by the heavy traces on these sections. The surface of this "acoustic basement" dips eastward beneath a thicker cover of sediments, and apparently constitutes a ridge behind which these sediments were deposited. Section C indicates that these sediments overlapped the ridge and spilled into Flemish Pass, prior to infilling of this depression to its present level. Erosion of these

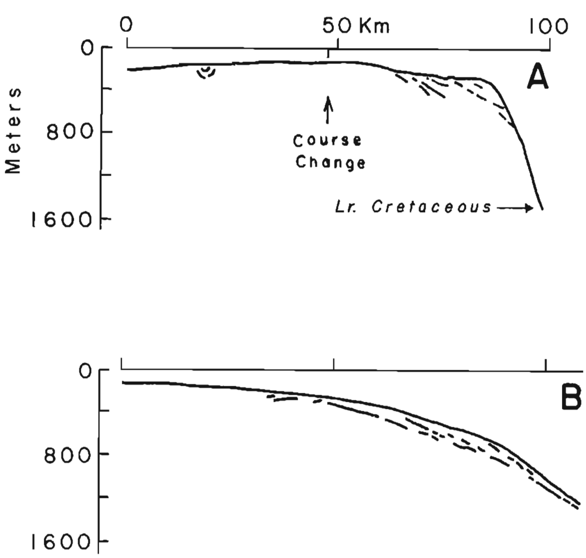

Figure 3;

Generalized seismic cross-sections from eastern Flemish Cap. Vertical to horizontal scale exaggeration is approximately 25 to 1.

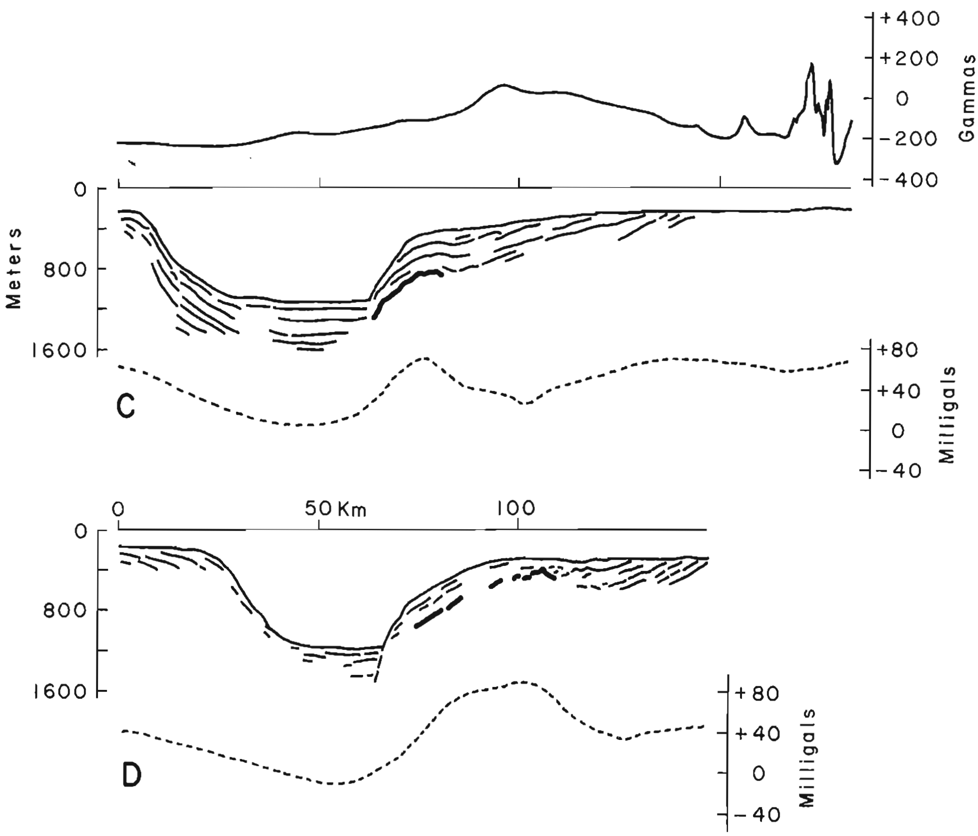

Figure 4. Generalized seismic cross-sections across Flemish Pass. Vertical to horizontal scale exaggeration is approximately 25 to 1. The regional field has been removed from the total intensity magnetic profile (solid line). The dashed lines indicate the free-air gravity anomaly.

sediments at sections D and E has largely obscured this relationship. The disturbed aspect of these sediments to the east of the ridge on sections D and E may reflect their deposition on an irregular surface, or gentle structural deformation subsequent to deposition.

Penetration of seismic energy into the ridge on Section E (Fig. 5) suggests that it is composed of sedimentary rocks. A similar ridge of acoustic basement material occurs beneath the eastern Grand Banks, as indicated by the heavy traces on sections E and F (Figs. 5 and 6).

The irregular magnetic profile over the central basement zone on the eastern end of section C (Fig. 4) is the characteristic magnetic aspect of that area. The rapid attenuation of these high frequency-high amplitude anomalies toward the flank of Flemish Cap denotes burial of this magnetic basement, probably at a greater rate than indicated by the limited seismic penetration. An aeromagnetic profile described by Hood and Godby (1965) crosses the line of section C at the western edge of Flemish Cap. Their calculated depth to magnetic basement in that area is of the order of approximately 7,500 m.

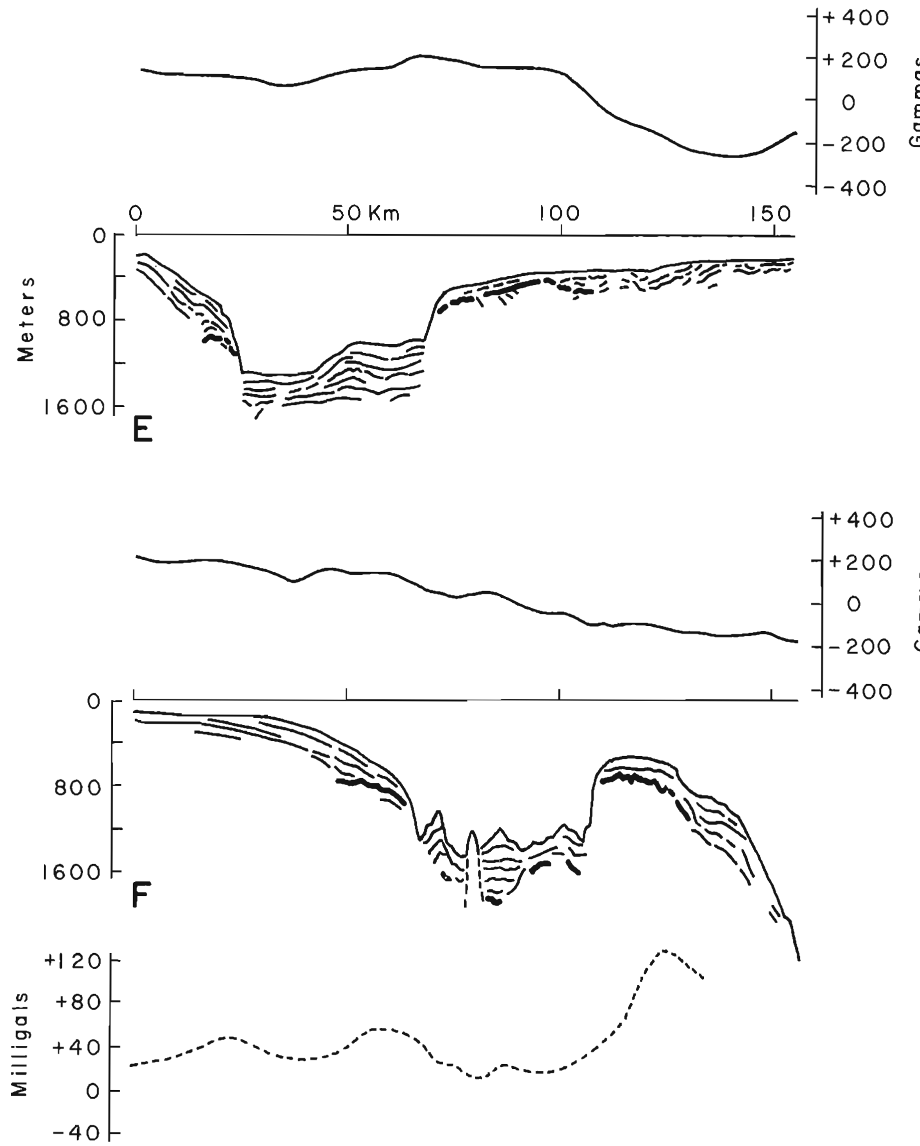

Figure 5. Generalized seismic cross-sections across Flemish Pass. Vertical to horizontal scale exaggeration is approximately 25 to 1. The regional field has been removed from the total intensity magnetic profile (solid line). The dashed line indicates the free-air gravity anomaly.

The lack of magnetic anomalies over the acoustic basement ridges on either side of Flemish Pass is indicative of a sedimentary composition, and indicates that they are not composed of the same material as that of the basement zone of central Flemish Cap.

The free-air gravity profiles plotted under seismic cross-sections C and D (Fig. 4) are derived from the surveys reported by Keen et al. (1970). These profiles reflect the bottom relief, but they also show positive anomalies coincident with the acoustic basement ridge on Flemish Cap relative to negative anomalies over the zone of sediment accumulation behind this ridge.

Section F (Figs. 5 and 6) extends southeast from the Grand Banks across the southern end of Flemish Pass, and across a ridge that projects from the southeast portion of Flemish Cap. A separate knoll (Beothuk Knoll) occurs on this ridge, but the degree to which it is physiographically isolated from Flemish Cap is not yet clearly defined. The heavy lines on this section denote the acoustic basement ridge beneath eastern Grand Banks and a similar feature beneath the physiographic ridge on the east side of Flemish Pass. Possibly this physiographic ridge expresses a southward extension of the acoustic basement ridge on Flemish Cap. The strong seismic events emphasized beneath Flemish Pass may reflect relatively shallow continuation of acoustic basement material across the southern end of this depression. The piercement-like structure in Flemish Pass causes a break in the sub-bottom reflectors, and reflections from the flanks of this structure - if they are not strictly side-echoes - persist to the depth of the acoustic basement events. The magnetic profile shows no indication of disturbance associated with this feature, however, it may be noteworthy that the gravity profile shows a small negative anomaly at this locality. Possibly these points support interpretation of this structure as a salt diapir. The free-air gravity profile on this section also displays positive anomalies coincident with the acoustic basement ridges.

A refraction profile recorded by the ATLANTIS in 1948, at the location marked "X" in Figure 2, indicated a 5.5 km/s refractor at a sub-bottom depth of 500 m (McConnell and McTaggart-Cowan, 1963). Although listed as questionable, C.L. Drake, Dartmouth College (personal communication), has ventured that this result may at least support the occurrence of a high-velocity layer at moderately shallow depth. It is speculated that this high-velocity layer may represent the acoustic basement ridge detected by seismic profiling on the eastern Grand Banks.

Geological data from the region of Flemish Cap are sparse, but they may be significant. Sample material has been collected at 13 locations in the southeastern quadrant of Flemish Cap at depths ranging from 140 m to 2,900 m (Table 1). Apart from a 10-cm core of granitic rock recovered from central Flemish Cap by drilling (Pelletier and Godden, 1970), all sample material has been collected by dredging. The percentages listed for the categories defined in Table 1 are calculated according to the number of representative specimens; roughly equivalent results would derive through calculation of percentages by weight. Obviously, neither measure is of much significance when the total sample consists of only one specimen.

A substantial fraction of most samples consists of well-rounded specimens of igneous rock, generally considered to be ice-rafted. The remainder of the sample, however, tends to consist of one rock-type with predominantly angular specimens, perhaps suggestive of local origin. The important

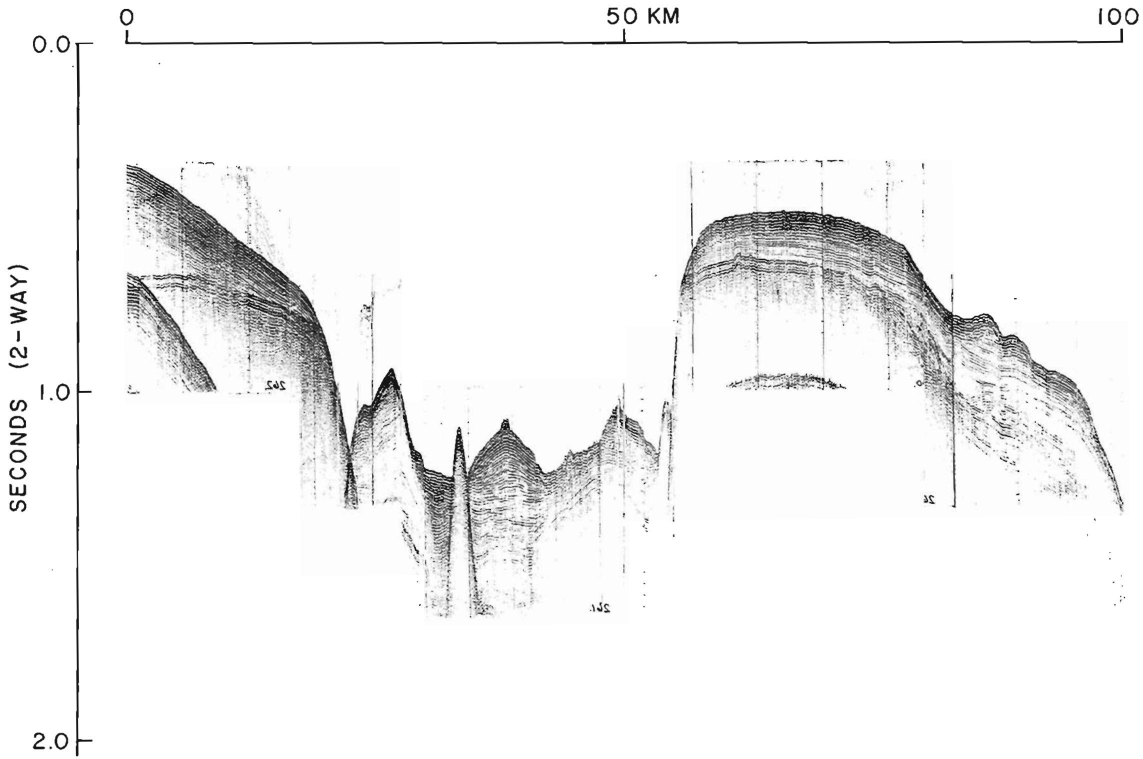

Figure 6. Reduced photograph of original seismic profiler records used to derive the generalized seismic cross-section in Figure 5 (F). The vertical dimension of the record is one second (two-way). The horizontal scale is variable due to changes in recording mode, but has been normalized on the general cross-section.

division of the categories in Table 1 is on the basis of hardness or, seismic properties. The clastics and carbonates are presumably the rock-types that may correspond to the layered media observed on the seismic profiler records. Rocks of this nature are essentially absent in samples from depths of less than 200 m, which is the approximate depth-limit of the central basement zone as defined by seismic profiling on southeastern Flemish Cap. Bottom photographs from the depth interval 800 m, to 1,200 m on the southeast flank of Flemish Cap show apparent outcrops of layered rocks (Fig. 7). Below 200 m, where the seismic profiler results show layered media, carbonate and clastic rocks are an important fraction of most samples. The carbonate material includes large, angular slabs of limestone. Fossil foraminifera were abundant in a limestone fragment from sample Hu-67-28 (Table 1). The consensus as to the age of these fossils is Lower Cretaceous (possibly late Jurassic).

The core of granitic rock from central Flemish Cap has yielded a K-Ar age of 590±20 m.y. (B.R. Pelletier, Atlantic Geoscience Centre, personal communication). Bottom photographs in the immediate vicinity of the drill

Figure 7. Bottom photograph from the depth interval 800 m to 1,200 m on the south flank of Flemish Cap. Estimated length of the fish is approximately 30 cm.

TABLE 1

Lithologic percentages for samples from southeast Flemish Cap

STATION	DEPTH (M)	SPECIMENS	CRYSTALLINE	QUARTZITE	VOLCANIC	CLASTIC	CARBONATE
Th*-65-22	137	1	100%	--	--	--	--
Hu*-69-1	140	90	2	98%	--	--	--
Hu-69-drill	143	1	100	--	--	--	--
Th-65-20	146	27	11	64	19%	--	7%
Th-65-18	154	1	--	--	100	--	--
Th-65-19	157	8	100	--	--	--	--
Th-65-21	209	27	15	7	74	--	4
Hu-69-2	271	9	33	56	11	--	--
Hu-69-3	289	32	50	25	--	3%	22
Hu-67-23	366	101	72	5	9	8	6
Hu-67-25	1472	87	84	2	2	2	10
Hu-67-28	1481	209	49	4	1	6	40
Hu-67-26	2926	21	57	5	10	--	29

* Th = M.V. THERON

* Hu = C.S.S. HUDSON

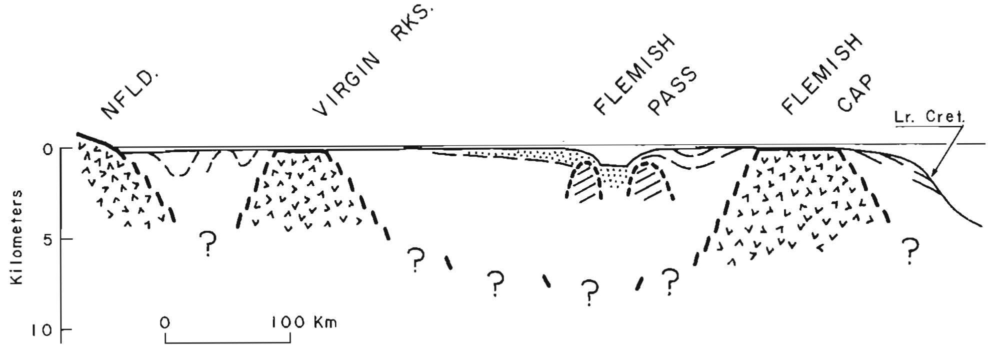

Figure 8. Hypothetical structural cross-section from the Avalon Peninsula east to Flemish Cap. Stipling denotes sediments presumed to be mainly Tertiary in age. Basement media underlying Flemish Cap and Virgin Rocks are tentatively correlated with Late Precambrian rocks on the Avalon Peninsula. Sample evidence indicates that strata on Flemish Cap are at least in part Lower Cretaceous. The buried ridges flanking Flemish Pass may possibly be Jurassic or older in age (see text). The vertical to horizontal scale-ratio of this cross-section is about 14 to 1.

site show what appears to be a smooth, rocky bottom, covered only by patches of sand. Assuming that the cored material was in situ, the central basement zone of Flemish Cap must be underlain at least in part by granitic rocks, and the age determined for the core-sample is close to ages reported for granitic rocks on the Avalon Peninsula of Newfoundland (Wanless, 1969).

DISCUSSION

The geological and geophysical results are summarized in the hypothetical cross-section shown in Figure 8, which extends eastward from the Avalon Peninsula of Newfoundland to the east side of Flemish Cap (Fig. 2), a distance of approximately 700 km. The late Precambrian rocks of the Avalon Peninsula are tentatively correlated with basement at Virgin Rocks, where Lilly (1966) reported similar rock-types, and with the central basement zone of Flemish Cap. The acoustic basement ridges are indicated on either side of Flemish Pass. The level at which the Lower Cretaceous limestone fragment was recovered on the southeastern flank of Flemish Cap is noted at the eastern end of the cross-section.

A zero-edge of the coastal plain deposits presumably occurs in the vicinity of Virgin Rocks. At the AMOCO (formerly Pan American) Tors Cove well on the southern Grand Banks (Fig. 2), about 280 km southwest of Virgin Rocks, the Tertiary section is approximately 1,000 m thick (Bartlett and Smith, 1971). Flemish Pass lies roughly 280 km to the east of Virgin Rocks. Possibly the sediments covering the basement ridge beneath eastern Grand Banks to a depth of about 1,000 m are also Tertiary in age (Fig. 8). These sediments prograde eastward to fill Flemish Pass, but appear to be absent, or are very thin, on the Flemish Cap.

The Lower Cretaceous (Upper Jurassic?) fossils recovered from eastern Flemish Cap may define the age of at least part of the sequence of layered strata that encircle the central basement zone. If strata of this age are included in the deposits behind the basement ridge on western Flemish Cap, then the ridge is indicated as older than Lower Cretaceous in age.

Apart from the cover of supposed Tertiary sediments on eastern Grand Banks, which might be expected to overlie Mesozoic strata similar to the section encountered in the two wells on southern Grand Banks (Bartlett and Smith, 1971), a degree of structural symmetry exists across the axial plane of Flemish Pass. On both sides of this depression there is apparently an accumulation of Mesozoic sediments behind a shelf-edge ridge. It has been mentioned that at least an apparent relationship exists between these ridges and positive gravity anomalies (Figs. 4 and 5). Positive gravity anomalies have been described as a shelf-edge characteristic of the Grand Banks-Scotian Shelf region (Keen et al., 1970). On the Scotian Shelf this gravity high may derive at least in part from an underlying basement ridge (Keen and Loncarevic, 1966). Shell Canada Ltd. may have drilled into Upper Jurassic carbonate rocks in the vicinity of the shelf-edge gravity high on Scotian Shelf (Oil and Gas Journal, 1970). An implication of this report is that the underlying basement ridge may be composed of Jurassic carbonate rocks, rather than of Lower Paleozoic (Meguma) rocks as suggested by previous interpretations of seismic refraction data. Extrapolating these tentative correlations to the Grand Banks region, perhaps an Upper Jurassic or older age for the ridges flanking Flemish Pass may be a plausible estimate.

In terms of the sea-floor spreading hypothesis, the Scotian Shelf in Jurassic time is believed to have bordered a narrow proto-Atlantic ocean. Whatever the origin of the shelf-edge ridge on Scotian Shelf, perhaps its formation was controlled by conditions peculiar to this pro-Atlantic. If the ridges on either side of Flemish Pass have a similar origin, possibly they record the existence of a narrow, northern arm of the proto-Atlantic through Flamish Pass, possibly connecting with a proto-Labrador Sea to the north.

Based on correlations of oceanic magnetic anomalies Hood et al. (1969), Mayhew et al. (1970), Laughton (1971), and Le Pichon et al. (1971) have suggested that the main rifting of the Labrador Sea commenced in Late Cretaceous time. The occurrence of Jurassic rocks on the Labrador Shelf (McMillan, 1973) is strong evidence for the existence of the Labrador Sea prior to Late Cretaceous, so that spreading at that time must have acted to widen an already existing depression. Diapiric structures reported by Pautot et al. (1970) as occurring some 300 km north of Flemish Cap, and Orphan Knoll may perhaps be further vestiges of a Jurassic Labrador Sea. JOIDES drilling results have proven that Orphan Knoll is a foundered piece of continental crust, at least as old as middle Jurassic (Ruffman, 1973). To the south, if major rifting occurred in Late Cretaceous time the axis of spreading was probably located to the east of Flemish Cap, rather than in Flemish Pass. The absence of a magnetic anomaly over Flemish Pass, evidence for spilling of possible Lower Cretaceous sediments into Flemish Pass from the east, indications of an acoustic basement continuity beneath southern Flemish Pass, and the possible salt diapir in this depression may attest its permanence through a Late Cretaceous episode of sea-floor spreading. By correlation with the salt structure in the Tors Cove well on southern Grand Banks (Bartlett and Smith, 1971), the occurrence of a salt diapir in Flemish Pass would imply the existence of this depression as a pre-Cretaceous salt basin.

If Flemish Cap was separated from the adjacent continent by processes of sea-floor spreading, the considerations outlined above may limit the timing of this event to Jurassic or pre-Jurassic dates. The demonstrated foundering of Orphan Knoll may alternatively express vertical crustal movements as an important factor in achieving apparent isolation of blocks of continental crust in this region. Perhaps the extensive erosion of Flemish Cap denotes elevation of this particular crustal block concurrent with Late Cretaceous spreading of the sea floor to the east, since Schneider (1969) has suggested that continental crust adjacent to an axis of spreading may be elevated in the early stages of rifting.

CONCLUSIONS

Geological and geophysical data indicate that Flemish Cap is a continental structure, which constitutes a positive element of the continental margin relative to the adjacent Grand Banks and Flemish Pass. The shelf-edge aspect of the western side of Flemish Cap may denote this feature as a former element of the east Atlantic continental margin. Pre-drift reconstructions of the north Atlantic region, such as that by Bullard et al. (1965), should be revised to include the Flemish Cap.

ACKNOWLEDGMENTS

The author is grateful to Dr. R.K. Banerji, Bedford Institute of Oceanography, Dr. F. Medioli, Dalhousie University, and Dr. B.K. Sen Gupta, University of Georgia, for their work on the fossil material from the Flemish Cap. D.E. Barrett, C.E. Keen, and B.R. Pelletier, Atlantic Geoscience Centre, Bedford Institute of Oceanography, are thanked for their critical review of the manuscript.

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26. A GRAVITY SURVEY OF EASTERN NOTRE DAME BAY, NEWFOUNDLAND

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Abstract

A gravity survey at 2.5 km mean station spacing was conducted on islands and the coast of Notre Dame Bay near the eastern boundary of the Paleozoic Mobile Belt of Newfoundland. The Bouguer anomaly field shows good correlation with the following dominant features of the surface geology: (1) a strong northeasterly structural trend; (2) the Luke's Arm fault; (3) several extensive granitic bodies; (4) diorite/gabbro intrusions. However, no significant gravity anomalies occur over sedimentary areas. The northeasterly regional trend is removable by a 5th-order polynomial approximation.

Geological criteria were used to divide the survey region into 13 blocks, each with a mean density derived from rock samples. Using these blocks as a basis, models were constructed until a good fit to the Bouguer field was attained. The model results lead us to propose two new features: (1) a structural discontinuity, suggested also on published aeromagnetic maps, separating the eastern (Fogo-Change Islands) and western parts of the region; (2) a mafic to ultramafic layer at 5-10 km depth to explain the overall positive character of the Bouguer anomalies. This layer appears to be a landward continuation of Sheridan and Drake's (1968) intermediate layer associated with the Taconic orogeny.

INTRODUCTION

In this paper we present the results of a gravity survey on the islands and the adjoining coast of eastern Notre Dame Bay, Newfoundland. The total area covered is about 2500 km^2, bounded by latitudes 49°00'N and 49°50'N and longitudes 54°00'W and 55°30'W (Fig. 1). An interpretation is also attempted.

Geological Setting of Eastern Notre Dame Bay

The region is part of the Paleozoic Mobile Belt of Newfoundland (Williams, 1964a). The basic structural trend is northwesterly and is normal to that of the continental shelf - a unique occurrence along the Atlantic coast. The region has been studied geologically by Baird (1958), Williams (1963, 1964b), Kay (1967), Eastler (1969, 1971), and Horne and Helwig (1969). The rocks range in age from Ordovician to Devonian (Fig. 1), with major deformation occurring during the Taconic (Ordovician) and Acadian (Devonian) orogenies. The dominant dislocation feature is the transcurrent Luke's Arm fault (Heyl, 1936). A thick and structurally complex sequence of volcanic and sedimentary Ordovician rocks is exposed over a large part of the region.

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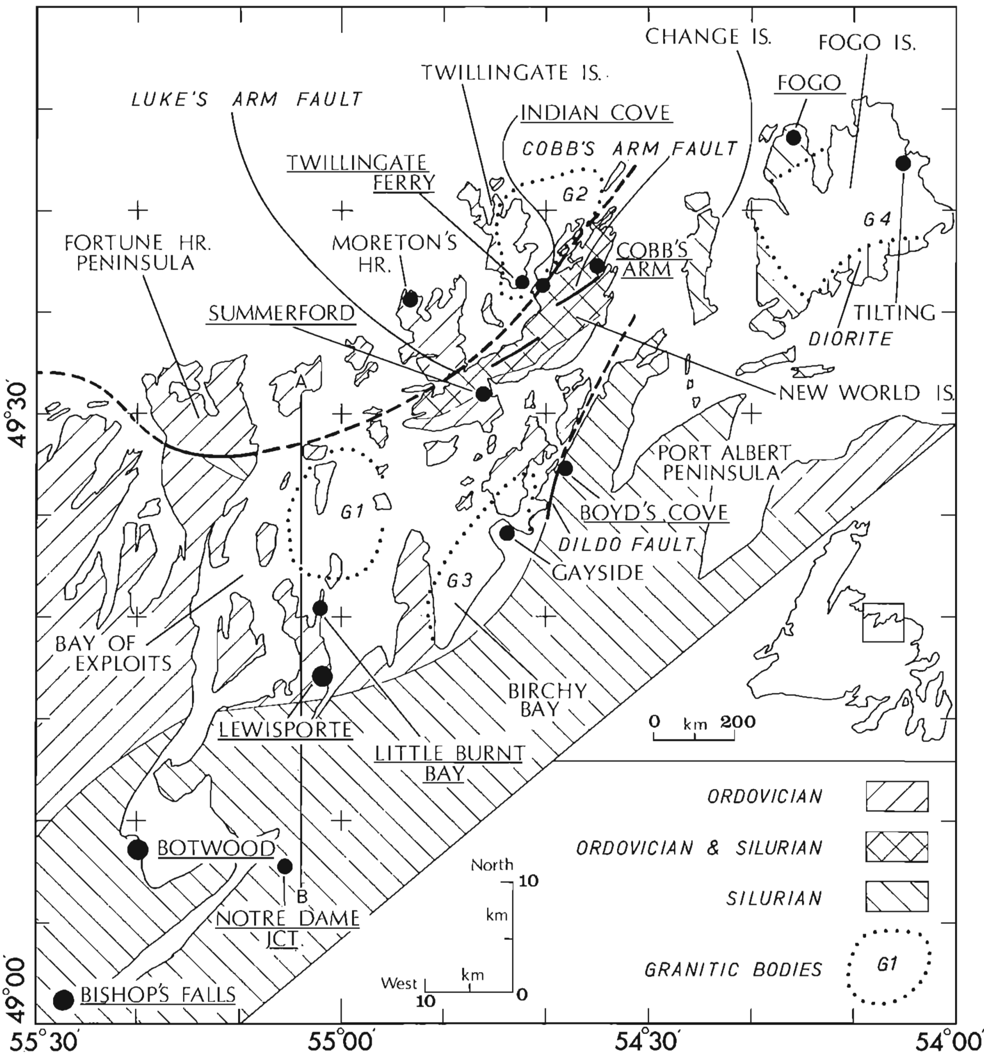

Figure 1. Map of eastern Notre Dame Bay and adjoining area. Geology simplified from Kay (1967), Williams (1967), Eastler (1969) and Horne and Helwig (1969). Outlines of granitic bodies are approximate, representing: G1, Long Island batholith; G2, Twillingate granite; G3, Birchy Bay granite; G4, Fogo granite. For details, see Table 1 and text. Gravity base stations are underlined. AB, gravity profile (Fig. 6).

Mafic, ultramafic and granitic materials was intruded into the existing rocks during the Taconic orogeny, followed during the Silurian by relatively quiet conditions, with deposition of shallow-water sediments and some volcanism. During the Acadian orogeny, most of the granitic material exposed in Notre Dame Bay was intruded. Since the Devonian, the region has been relatively stable.

A fundamental problem in the context of continental drift concerns the composition and shape of the Paleozoic Mobile Belt near the continental margin. Wilson (1962) has discussed the case for Newfoundland which separated from the British Isles. The fit of the North Atlantic by Bullard et al. (1965) places the continental shelves of Europe and North America adjacent to one another, much as Wegener (1921) proposed. This reconstruction is consistent with the proposition that the Caledonian system of Great Britain and the Appalachian system of North America were once a single system. If this is accepted, the area northeast of the Bay of Exploits should provide evidence about the location of the break in the Appalachian-Caledonian system, i.e. the Appalachian structure would be continuous for 300 km or so to the continental margin and end there abruptly. If the system is not continuous to the margin, then it becomes more difficult to accept the hypothesis of continuity of the Appalachian-Caledonian systems. In either case, geophysical rather than standard geological methods of investigation are required, and exclusively so in the region beyond the Bay of Exploits archipelago.

Offshore geophysical surveys

The postulated seaward extension of the Appalachians has been investigated by geophysicists of the Bedford Institute and Dalhousie University (Ewing et al., 1966; Fenwick et al., 1968) and Lamont Geophysical Observatory (Sheridan and Drake, 1968). From results obtained on two orthogonal sets of seismic refraction profiles located between Fogo Island and a location 60 km northeast of that island, Sheridan and Drake conclude that the Taconic orogenic belt extends to the continental margin with no change in axial direction. Since the Appalachian and Caledonian systems were both affected by the Taconic orogeny, while the effects of the Acadian orogeny are observed in the Appalachians and appear to die out short of the continental margin, they further conclude that there is a significant crustal discontinuity at the shelf edge of post-Taconic, pre-Acadian age. Their chief evidence is the presence of a high-velocity (6.68 - 7.39 km/s) intermediate layer encountered at depths varying from 4 to 8 km. Sheridan and Drake assume this layer to be mafic to ultramafic rock associated with the Taconic orogeny. These conclusions elucidate the finding by Ewing et al. (1966) of a high-velocity (7.2-7.3 km/s) intermediate layer in the lower crust beneath the axial zone of the Appalachians.

A number of magnetic and refraction seismic profiles were obtained parallel to the shelf by Fenwick et al. (1968). The seismic results agreed with those of Sheridan and Drake and the magnetic results showed a pattern of positive magnetic anomalies with contours parallel to the shelf edge and with the amplitude decreasing towards deeper water. Fenwick et al. interpret this as evidence for an abrupt discontinuity at the continental margin where the thicker continental crust with the more magnetic intermediate layer grades into oceanic crust with a thin basaltic layer. From these results it appears that the Appalachian system is continuous to the continental margin from the seaward extremity of the present survey region.

Gravity Studies of Notre Dame Bay

Additional geophysical surveys offshore from the mainland coast of Notre Dame Bay are needed to test the continuity of deep-seated structures. Among the most interesting features one could hope to detect is a landward extension of Sheridan and Drake's intermediate layer. Notre Dame Bay, which contains numerous large and small islands, offers a unique opportunity for conducting land-based geophysical surveys in the continental shelf area beyond the outermost fully-exposed part of the Appalachians.

The only published gravity data for Notre Dame Bay are from the Dominion Observatory (now Earth Physics Branch of the Department of Energy, Mines and Resources) survey of Newfoundland (Weaver, 1967), in which the stations were 10 to 13 km apart, with seventeen stations in the present survey region. The results show a rapid change from low to high anomalies, predominantly in a seaward direction perpendicular to the geological strike, including two or three prominent features. Because of the large station spacing of the Dominion Observatory survey, several important features could not be accurately outlined, thus justifying the need for a survey with closer spacing. The present survey meets this need.

Gravity Survey Logistics

The present survey was conducted during July 1968 and May to August, 1969. Details are given in Miller (1970) and will only be summarized here. The Sharpe 'Canadian' CG-2 gravity meter of Memorial University was used. Gravity and elevation determinations were made for 308 stations, and 9 gravity sub-bases were established. The mean station spacing was 2.5 km (Fig. 2), this being the tightest grid permitted by the geography.

Transportation was by Jeep and Land Rover on roads; by locally rented boats to smaller islands; and by aeroplane when tying together the gravity bases at Cobb's Arm and Fogo. Distances along the roads were determined from the odometer of the vehicle used, and the positions of island stations were located with the aid of 1:50,000-scale topographic maps. All station positions are known to ±0.05' latitude or about 100 meters, corresponding to an error of $\lesssim$0.08 mgal in applying the International Gravity Formula.

A total of 223 rock samples were obtained for density determinations (Table 1) by sampling as many outcrops as possible. The density data and their use in gravity interpretation are discussed later.

Altimetry

Elevations were obtained by direct levelling, using standard techniques, for 157 stations and by aneroid barometry for the other 151 stations.

Direct levelling was used on islands and coasts, and as a reference, except where bench-marks were available. As the elevations were often obtained near low tide, all stations were referred to "chart datum", defined as the plane below which the water level seldom if ever falls. Tidal corrections to the chart datum were made by interpolation using published tidal tables of the Canadian Hydrographic Service. Chart datum was not adjusted to mean sea-level, as the difference between these two reference levels was not everywhere known, though we found the former to be typically 0.70 m lower; thus Bouguer anomalies in this survey are estimated to be 0.14 mgal higher than would be obtained in a standard gravity survey.

A "modified single-base method" employing three Wallace & Tiernan altimeters was applied to the readings at barometric stations. Two of the instruments were used for roving along traverses, the third at a base of known elevation. Usually all stations on a traverse were remeasured at least once, starting from bases at opposite ends of it. Errors were estimated by comparing known elevations with barometrically observed values at each of two terminal bases and interpolating the base errors at the intervening stations.

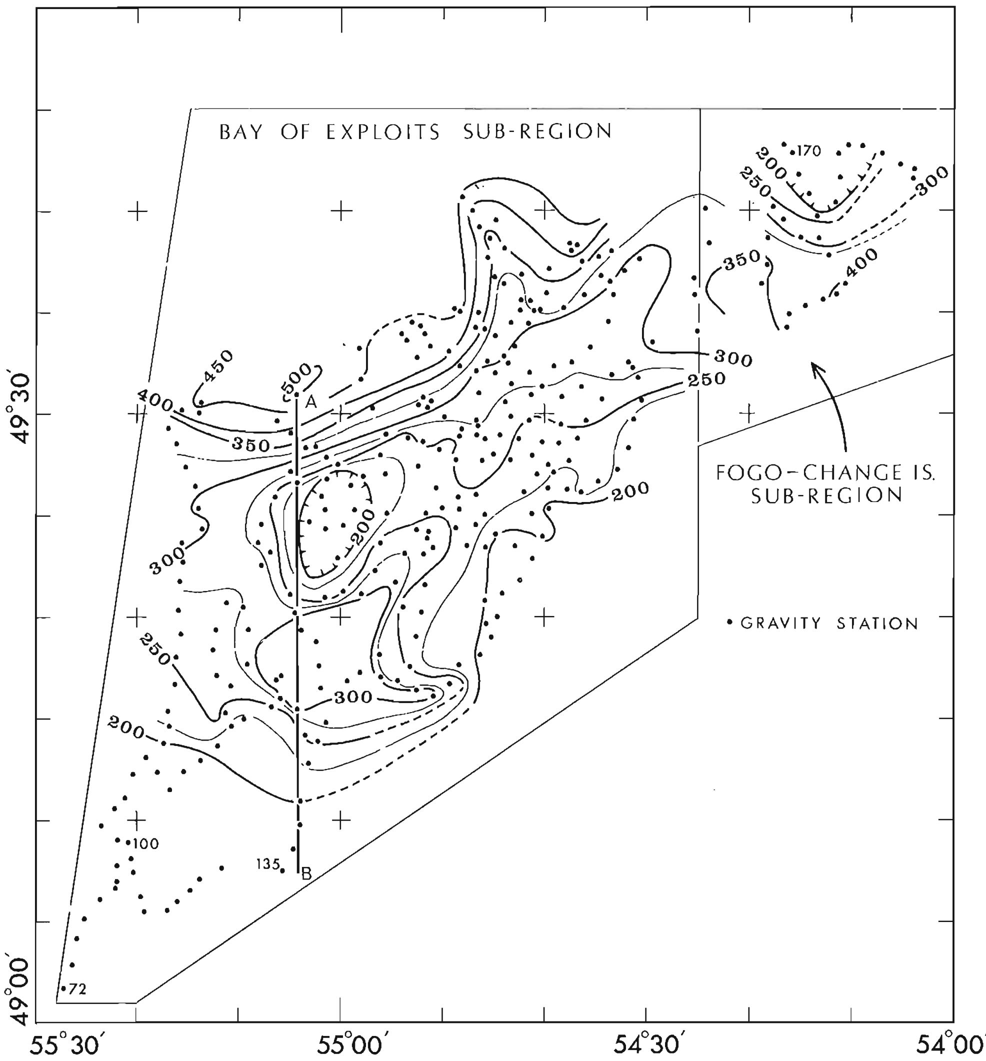

Figure 2. Bouguer anomaly field of eastern Notre Dame Bay and adjoining area. Gravity is quoted in units of 0.1 mgal, with 2.5 mgal contour intervals. Individual anomaly values are quoted for some stations outside contoured area. The two block outlines correspond to the density models of Figs. 3 and 4. AB, gravity profile (Fig. 6).

TABLE 1

Rock sample densities

Block No.[1]	Location and main rock types[2]	No. of samples	Mean density[3] (g/cm^3)
1	Bay of Exploits, granite (Long. Is. batholith, G1)	8	2.68±0.03
2	Twillingate, granite (G2)	10	2.66±0.02
3	North of Luke's Arm fault west of Twillingate, Moreton's Hr. volcanics with small granitic intrusions	21	2.79±0.11
4	North of Luke's Arm fault east of Twillingate, volcanics	3	2.79±0.10
5	Lewisporte area, sediments with diorite intrusions	73	2.79±0.11
6	Birchy Bay, granite (G3)	8	2.66±0.07
7	Port Albert Peninsula, Silurian sediments and volcanics	6	2.69±0.09
8	Fogo Island, granite (G4)	3	2.67±0.04
9	Fogo Island, diorite	4	2.80±0.09
10	North of Botwood, Ordovician sediments	11	2.77±0.12
11	Southern belt, New World Island, sediments and minor volcanics	25	2.73±0.13
12	Central belt, New World Island, sediments and minor volcanics	46	2.70±0.08
13	Gayside, diorite	5	2.95±0.13
	TOTAL	223	2.75

[1] Blocks 1-9 and 11 are used with same numbering in the density models (Figs. 3 and 4).

[2] See Fig. 1 and geological publications. "Granite" here includes granodiorite.

[3] Arithmetic mean. Errors quoted are standard deviations.

Overall error estimates are: Direct levelling error, (i) <0.15 m (102 stations); (ii) 0.15 - 0.4 m (55 stations). Barometric altimetry error, (iii) 0.4 - 2.0 m (85 stations); (iv) 2.0 - 3.0 m (32 stations); (v) undetermined (34 stations). Thus the highest elevation errors in this survey seem to be 3 metres, corresponding to a Bouguer anomaly error of 0.6 mgal. Usually this value will be much less, as shown above.

Gravity Observations

The primary gravity reference point was the Dominion Observatory base at Bishops Falls. The 9 sub-bases were tied together and tied to the primary base through loops of type ABAB... starting and ending with Station A. This base grid was also tied, at Notre Dame Bay Junction, to Memorial University's 0.8 km-spaced gravity profile across Newfoundland (Weir, 1970). The mean standard error for all sub-bases was ±0.03 mgal. A closure error could not be found as the system was tied to only one primary reference. In addition to the sub-base error, there are random observational errors (scale reading, drift), plus a systematic error in the instrument scale constant which we estimate from field tests to be +0.20 mgal or less. At any station, the estimated error in observed gravity due to random observational errors is 0.04 - 0.05 mgal.

Reduction of Gravity Data

Bouguer anomalies were computed by standard techniques, using a density of 2.67 g/cm^3. Principal facts for all stations are given in Miller (1970). Figure 2 is the resultant contour map of the Bouguer anomaly field. Sources of random error in the computed Bouguer value are: (a) the observational errors (0.04 - 0.05 mgal); (b) the latitude error (<0.08 mgal); (c) reduction errors due to (i) the elevation error ($\leq$ 0.4 mgal for 80% of the stations) and (ii) departure of the true density from 2.67 g/cm^3. The error for case c(ii) was 0.6 mgal or less and was usually negligible because of the low topography (always less than 100 m, the usual range being less than 50 m). Despite the low land relief, possible errors due to terrain had to be considered, since much of the survey was conducted close to the land/sea interface with its large density contrast. However, water depths tend to be less than 100 m in the survey region and much less than that close to stations. The terrain effect was estimated to be always less than 0.2 mgal, and was therefore neglected.

With these data, the combined probable error is ±0.5 mgal. Adding the instrument scale error, the estimated overall error in the quoted Bouguer value at any station, relative to chart datum, is +0.20±0.5 mgal.

Separation of Regional and Residual Gravity Fields

Figure 2 suggests a regional trend due to deep-seated masses. It was useful to separate these from the near-surface (small wavelength) structures. Since the station spacing was not uniform, this was done by a polynomial approximation rather than Fourier analysis. The technique was a least-squares fit using a multiple regression program run on Memorial University's IBM 360/40 computer. The solution was carried out for orders

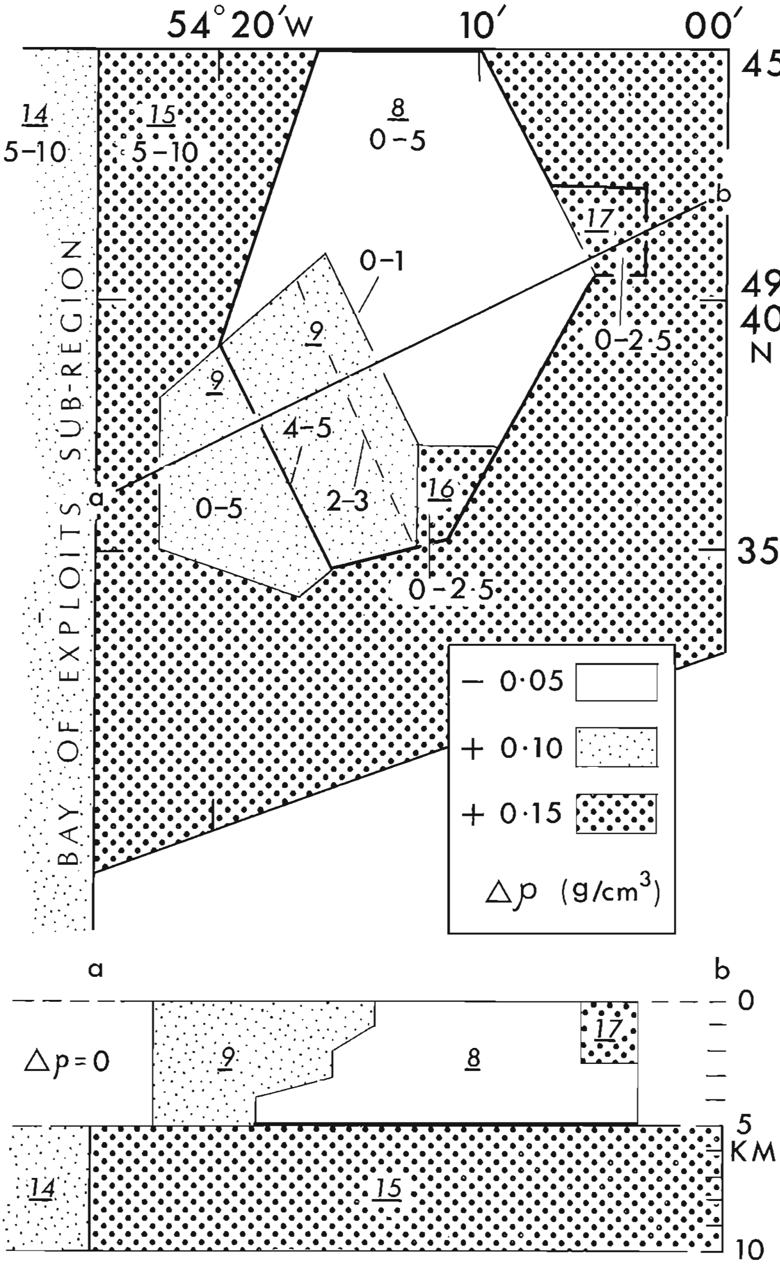

Figure 3. Density model for Fogo-Change Islands Sub-Region. Underlined numbers correspond to blocks described in Table 1 and text. Other numbers (e.g. 0-5) are depth ranges of block surfaces in km below sea level. In void spaces the density contrast (Δρ) is zero.

Top: Regular outlines are boundaries of upper block surfaces; heavey outline is that of Block 8 at 5 km depth.

Bottom: Vertical section through model, to illustrate block shapes at depth.

1 to 5. The regional and residual anomalies were mapped for order 5, which was chosen as the one best representing the regional trend. That trend, striking ENE, has been strongly accentuated on the regional map, compared with Figure 2. The resultant gravity maps are presented in Miller (1970).

INTERPRETATION OF GRAVITY RESULTS

In this section we attempt first to correlate visually some dominant features of the Bouguer map (Fig. 2) with the known geology. Secondly, we discuss a three-dimensional model study intended to take into account those components of the anomaly field that cannot be explained by surface geology alone.

Surface Geology of Eastern Notre Dame Bay

From geological data, the Luke's Arm fault can be traced intermittently from the western boundary of this survey to the northeastern part of New World Island (Fig. 1). That island is demarcated from the most easterly part of the survey region by major faulting in the vicinity of Change Islands (Eastler, 1971; not shown in Fig. 1), where there is also a marked change in the observed Bouguer anomaly trend (Fig. 2). For this reason the survey region was divided into two parts, with the Fogo-Change Islands Sub-Region adjoining the larger Bay of Exploits Sub-Region at longitude 54°25'W. However, for ease in model studies it was more appropriate to use smaller units. Hence the region was divided into 13 "blocks" (Table 1) for which rock density information has been collected. Their outlines are similar to those of blocks numbered correspondingly in Figs. 3 and 4 (e.g. 8). No sample densities have been determined for the four additional blocks (14-17), including two subsurface blocks (14, 15), used in the three-dimensional model study.

Bay of Exploits Sub-Region; (i) North of Luke's Arm Fault (Blocks 2, 3, 4). Ordovician volcanic rocks are exposed east of Twillingate (Block 4) and between Fortune Harbour peninsula and northwestern New World Island (Block 3). Intruded into these volcanics are diorite and gabbro, small granitic bodies, and the Twillingate granodiorite batholith of presumed Ordovician (G2, Block 2).

(ii) Southwestern Part (Blocks 1, 5, 6, 10). A granodiorite batholith (Devonian ?) is centered on Long Island (G1, Block 1). Near Birchy Bay another granitic body, of uncertain age, is exposed (G3, Block 6). Near Lewisporte, a complex of diorite, gabbro, and minor ultramafic rocks is intruded into sedimentary rocks (Block 5). Ordovician and Silurian sedimentary rocks cut by numerous small faults are found in the Botwood area (Block 10).

(iii) New World Island and Southward (Blocks 7, 11, 12, 13). From descriptions mainly by Williams (1963) and Kay (1967), this complex area is cut by a series of northeast-striking faults, separating four repeated, largely sedimentary sections of Ordovician and Silurian age. On New World Island, both the central belt between Luke's Arm and Cobb's Arm faults (Block 12) and a southern belt between Cobb's Arm and Dildo faults (Block 11) contain some volcanic rocks. A small diorite body (Block 13) occurs in the Silurian rocks of the Port Albert Peninsula (Block 7).

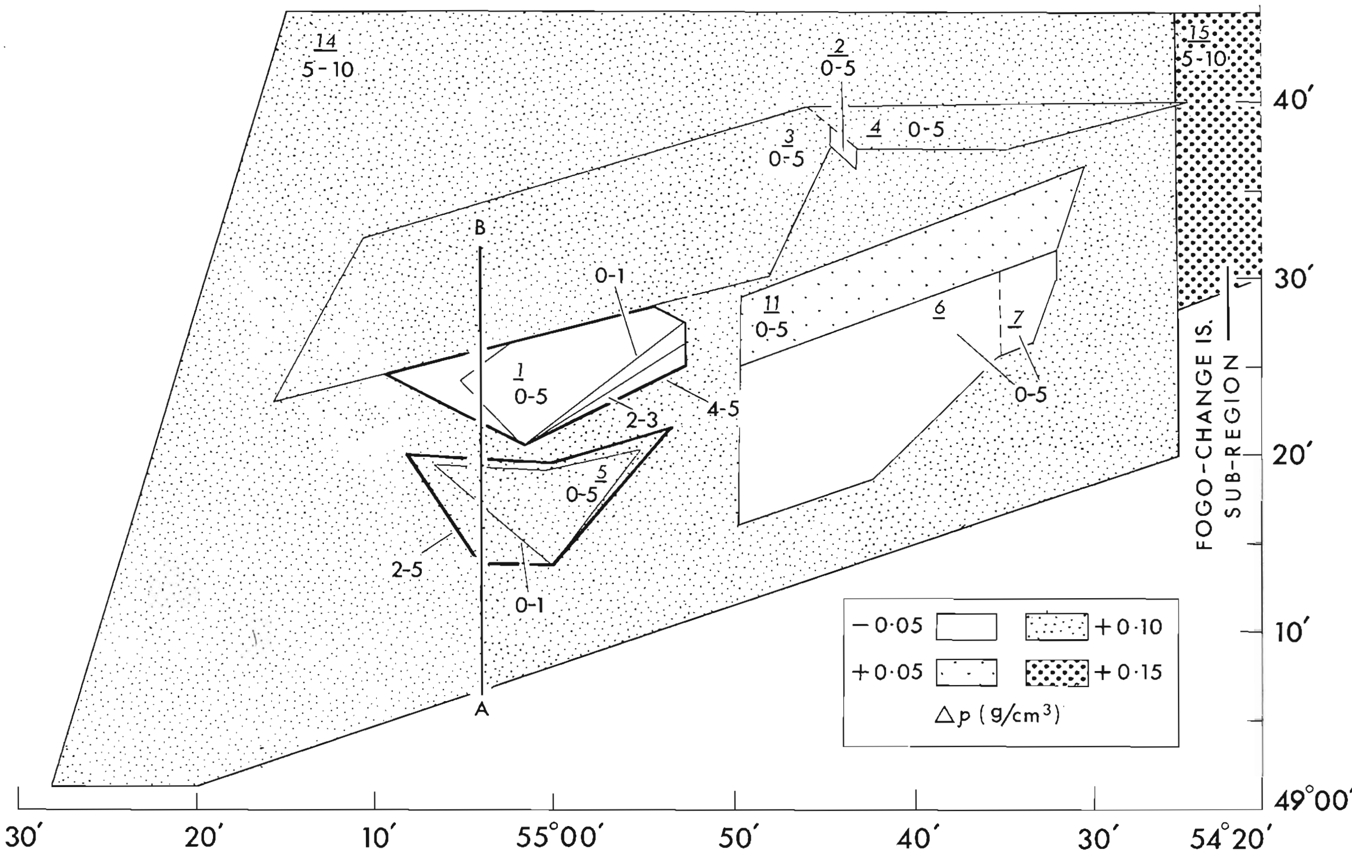

Figure 4. Density model for Bay of Exploits Sub-Region. Regular outlines are boundaries of upper block surfaces; heavy outlines delineate Blocks 1 and 5 at 5 km depth. Notations as in Fig. 3. Note reduction of N-S vs. E-W scale compared with Figs. 1-3. Blocks 10, 12 and 13 in Table 1 were not used and Block 14 is new. AB, gravity profile (Fig. 6).

Fogo-Change Islands Sub-Region (Blocks 8, 9, 16, 17). The geology has been described by Baird (1958) and Eastler (1969). A granodiorite batholith (G4, Block 8) is exposed on more than half of Fogo Island. Block 9 is diorite and gabbro found on southern Fogo. Two blocks (16, 17) used in the three-dimensional model only (Fig. 3) represent smaller diorite bodies in eastern and southern Fogo. Silurian sedimentary rocks exposed in the area are not represented by a block.

Density of Rock Specimens

Rock densities were measured by standard methods using coated surfaces at first. As the density values of samples measured both coated and uncoated were not significantly different, the majority of samples were coated only when highly porous. The results are accurate to within one percent for rocks weighing over 100 g in air (all except 15 samples). It should be noted that the distribution of rock exposures tends to be too irregular to ensure that the mean block densities are always unbiased. Hence the data in Table 1 should be considered the best available but not necessarily the most accurate measure of the average density.

Qualitative Interpretation

Several geological features of the region have associated gravity anomalies (compare Figs. 1 and 2).

(1) Faults. The steep gravity gradient across the Luke's Arm fault is indicative of a high-angle fault marking a sharp discontinuity in density. This interpretation is supported by the large density contrast as measured from rock samples, averaging 0.10 g/cm^3, between Blocks 3, 4 north of the fault and Blocks 1, 12 to the south (Table 1). The Bouguer contour pattern suggests that the Luke's Arm fault, whose locus is known only intermittently from surface geology, is a continuous feature traceable from northeastern Notre Dame Bay, between Change Islands and New World Island, across New World Island and the Bay of Exploits to the western survey boundary. The density contrasts across the Cobb's Arm and Dildo faults are only 0.03 - 0.04 g/cm^3 (compare Block 11 with Blocks 12 and 7 in Fig. 4), and the gravity gradients in Figure 2 are insufficient to delineate these faults accurately.

(2) Granitic Bodies. The granites (Fig. 1, G1 - G4) have densities close to that used for the Bouguer correction (2.67 g/cm^3). Low gravity values are associated with all four granite bodies, which is most strikingly apparent (see Fig. 2) for the Long Island batholith (G1 in Fig. 1). Heyl (1936) mapped the batholith on islands of the Bay of Exploits, estimating its area to be 16 x 8 km^2 at the surface but possibly larger underneath the bay. An eastward extension of the Long Island batholith is indicated by the gravity data, particularly on the 5th-order residual map (Miller, 1970). The western boundary of the body is less certain, due to a paucity of gravity stations in that area, where there are few accessible sites. The other three granitic bodies (G2 - G4) are close to the survey boundaries, making their shapes difficult to determine. The polynomial maps (Miller, 1970) suggest that the Twillingate granite (G2) is a relatively shallow, oval-shaped feature.

(3) Sedimentary Rocks. The blocks (7, 10, 11, 12) composed mainly of sediments have densities near or exceeding that of granite, i.e. higher than

that for typical sediments; this is not unexpected, as these rocks are deformed. The sedimentary rocks seem to have little effect on the gravity field.

(4) High-Gravity Features. The major gravity high near Lewisporte corresponds to the high-density area of Block 5 (mean density of 2.79 g/cm^3, with individual samples as dense as 3.10 g/cm^3). As the exposed rocks are mainly sediments cut by diorite dykes, a likely inference from the gravity data is that the dykes were fed from areally-large magma chambers. The highest gravity value of the survey was measured on Exploits Island (Fig. 1, Point A on north-south profile), where there are diorite and gabbro intrusions; this is not necessarily a maximum value, because no readings could be obtained further north.

(5) Fogo-Change Islands Sub-Region. In Figure 2, this area has a characteristic gravity field pattern distinct from the rest of the survey. The major features is the low associated with the Fogo granodiorite (G4), which composes most of the island. Relatively high gravity values found over eastern and southern Fogo probably are due to diorite bodies, but there are too few gravity stations to determine accurately their shapes and depths.

(6) Gravity Gradient in the Southwest. Figure 2 shows a pronounced southward decrease of gravity in the southwest corner of the survey region. The Bouguer anomalies, though still positive, have the smallest amplitude anywhere in the region surveyed. There is no obvious explanation apparent from the surface geology for this trend.

Quantitative Interpretation - Three-Dimensional Model

It is easily shown that surface geological features cannot account for much more than 15 mgal of the total anomaly field in Figure 2. Thus the major contribution must come from rocks at a depth represented by the regional anomaly. The distribution of this mass can be estimated only by the use of quantitative models. Such a model was developed from a procedure published by Talwani and Ewing (1960). A program written for use with Memorial University's IBM 360/40 computer was tested by calculating the gravitational effect of simple bodies approximated by horizontal laminae. The expected error from modelling actual bodies of unknown shape was found to be of the order of 15-20 percent. Because the average anomaly in our survey is about 30 mgal, this means that an r.m.s. error of 5-6 mgal was as good a fit as could be expected.

Fogo-Change Islands Sub-Region. The survey region was first divided into 13 blocks according to Table 1. Initially the model was applied to Fogo and Change Islands, where the geology is well known. The final block distribution (Fig. 3) evolved as follows: the first version included only two surface masses, Blocks 8 and 9 between sea level and 5 km depth, plus a high-density mass (Block 15) between 5 and 10 km. The presence of Block 15 is consistent with Sheridan and Drake's (1968) postulated mafic to ultramafic crustal layer which, in the area just northeast of Fogo, extends downwards from about 5 km depth. The 5 to 10 km depth span adopted for this layer in our final model corresponds to the best of obtainable fit with the observed Bouguer anomaly field. Density contrasts $\Delta\rho$ == -0.05, +0.10 and +0.15 g/cm^3, respectively, were assigned to Block 8 ("granite"), Block 9 ("diorite") and Block 15 ("mafic to ultramafic rock"). A density contrast of $\Delta\rho$= 0 in this model would correspond to 2.70-2.72 g/cm^3 in Table 1. Mean density values in the latter range seem more realistic than the 2.75 g/cm^3

mean value quoted in Table 1 for the 13 blocks, which is probably biased towards relatively high-density rocks, e.g. diorite, that tended to be more accessible for sampling than other rock types.

Finding that the resulting anomalies, compared with the observed Bouguer anomalies, were too high near the boundary of Blocks 8 and 9 and too low near the eastern and southern shores of Fogo Island, we revised the position of the former boundary to allow Block 8 (i.e. the granite) to undercut Block 9, and added two smaller blocks (16, 17) with $\Delta_\rho = +0.15$ g/cm^3.

This eliminated most of the disagreement with the observed Bouguer field, leaving an r.m.s. misfit of 4.91 mgal. In the final Fogo-Change Island model, Block 9 corresponds to the sedimentary and igneous rocks[1] exposed on western Fogo, while Block 8 is the granodiorite exposed on most of the island which is assumed to spread beneath western Fogo. Blocks 16 and 17 correspond to diorite bodies. The diorite off southern Fogo is inferred from onshore exposures. Change Islands, where the surface rocks are mainly sedimentary, was not modelled above 5 km, i.e. we put $\Delta\rho = 0$.

Bay of Exploits Sub-Region. The model for the 280 western stations was developed by successive trial-and-error in the same way as the Fogo-Change Islands model. In the final version (Fig. 4) the r.m.s. misfit with the observed Bouguer field was 5.46 mgal. The combined model (Figs. 3, 4) resulted in a computed anomaly map (Fig. 5) dominated again by the Lewisporte gravity maximum and the minimum for Long Island granite.

For the Long Island granite (Block 1) the computed and observed anomaly patterns (Figs. 2, 5) are similar, but the exact shape of Block 5 causing the Lewisporte maximum, is questionable: in this case the misfit between calculated and observed anomalies exceeds the 20 percent attributable to the program (see also profile AB, Fig. 6), and the shape and/or depth of the body may have been wrongly chosen. Readjusting the depth would require readjustment of the underlying layer as well. Unfortunately, in the case of Block 5, the geological diversity and the dispersion in density values (Table 1) are exceptionally great. The sediments seem to be a factor only in Block 11 (southern New World Island); the other sedimentary blocks (10, 12) and the Gayside diorite (Block 13) were not needed in Figure 4 to explain observed gravity.

The 5-10 km Layer. The main mass is Block 14, underlying all others. Choosing the same density contrast (+0.15 g/cm^3) as for Block 15, which it adjoins under Change Islands, would have made the computed anomalies 1.5 times too large over most of the Bay of Exploits Sub-Region. We obtained a much better fit by adopting $\Delta\rho = +0.10$ g/cm^3 for Block 14, which is the best value if the model of a layer between 5 and 10 km depth is correct. Alternatively, a subsurface layer having uniform density contrast +0.15 g/cm^3, but thinning westward from Change Islands, would also explain most of the Bouguer field.

Possible shape or depth variations of the 5-10 km depth-layer might resolve model misfits, e.g. the Lewisporte gravity high, but as this is speculative we have not remodelled Block 14. Confining the density-contrast boundary across the Luke's Arm fault to 5 km depth satisfies the model but does not eliminate the possibility that the fault may go deeper. The low-amplitude gravity anomalies in the southwest can be explained by truncating the 5-10 km layer. This was done in the model, but the question of the continuation of this layer beyond the survey region remains open. If it is the layer

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[1] Note: Block 9 density in Talbe 1 is from diorite only.

proposed by Sheridan and Drake (1968), one must ask why there is such a marked discontinuity of density under Change Islands. This question cannot be answered from the results of the present survey.

Allowing for the limitations discussed, the overall agreement between observed and computed anomalies (Figs. 2, 5, 6) is good. The model work helps explain two important new features that could not be interpreted from cursory inspection of the geological maps or Figure 2. These are (i) the decrease in Bouguer values southwards in the southwestern part of the survey

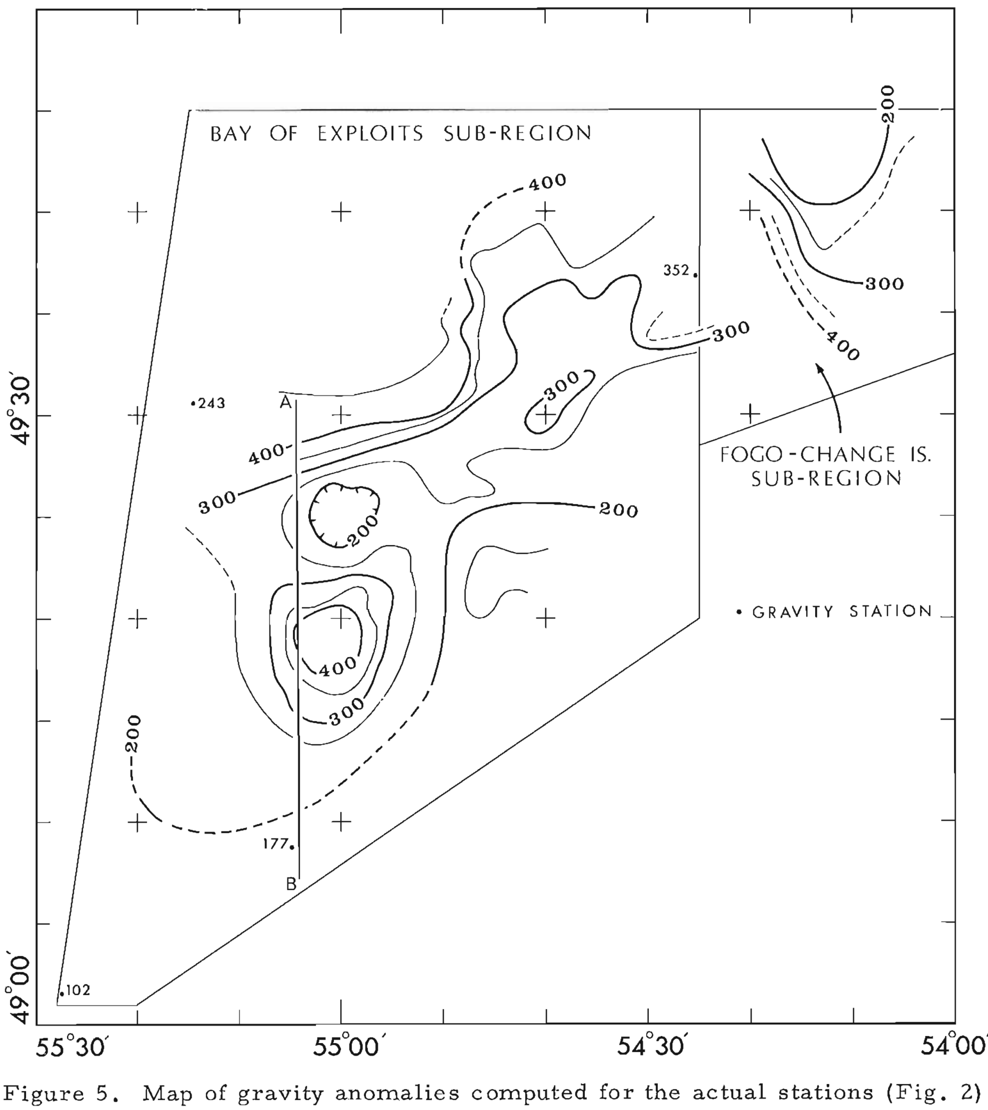

Figure 5. Map of gravity anomalies computed for the actual stations (Fig. 2) from the combined models of Figs. 3 and 4 (shown in outline). Gravity is quoted in units of 0.1 mgal, with 5.0 mgal contour intervals. Stations are not shown except where individual gravity values are given. AB, gravity profile (Fig. 6).

region; (ii) the sharp change in gravity near Change Islands. The study has also shown that the major part of the anomalies cannot be explained solely by features observable at the surface, but requires a subsurface structure with excess mass.

Magnetic Anomalies. We have qualitatively compared the gravity results with anomalies on the total-intensity aeromagnetic map of the region (Geological Survey of Canada, Map 4453G, Ottawa, 1969). The magnetic profile in Figure 6 was obtained by subtracting 54, 600 gammas from all map values and interpolating between contours. The profile sharply depicts the Long Island granite as a magnetic high and shows the Luke's Arm fault; it is also consistent with the postulated subsurface widening of Block 5. The aeromagnetic maps also depict (not shown here) a major break in the anomaly pattern near Change Islands, similar to the gravity maps.

SUMMARY AND CONCLUSIONS

A gravity survey of eastern Notre Dame Bay, Newfoundland, was undertaken with 2.5 km mean station spacing. The random error of the Bouguer anomalies was ±0.5 mgal for more than 80 percent of the stations. Consideration of the known geology, combined with densities obtained from rock samples, leads to a division of the survey region into 13 surface blocks. We conclude:

(1) A northeast-striking regional trend in anomalies can be removed by a polynomial of order 5.

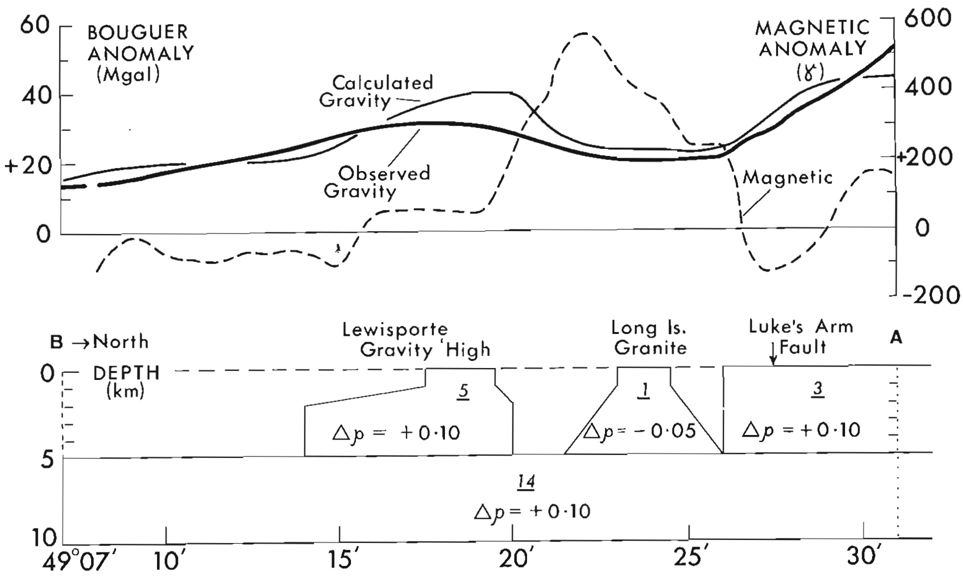

Figure 6. North-south profile, AB, of observed Bouguer anomaly (Fig. 2), computed gravity anomaly (Fig. 5) and block model used for computation (Fig. 4). A magnetic anomaly profile, obtained by interpolation on a published G.S.C. aeromagnetic map (see text) is shown for comparison.

(2) Three-dimensional models can explain the observed gravity anomalies, with an r.m.s. error of less than 6 mgal, by density changes in the upper 10 km of the crust. An excess-density layer between 5 and 10 km depth can account for most of the anomaly field.

(3) The models reveal a major crustal discontinuity near Change Islands at a depth between 5 and 10 km. This feature is seen both on gravity and on aeromagnetic maps. We have found no previous published reference to such a discontinuity, though the existence near Change Islands of the Reach fault and one or two other possible north- to northeast-trending faults reported by Eastler (1971) appears to be compatible with it.

(4) The closed maxima on the Bouguer map are due mainly to diorite/gabbro intrusions and the minima to granitic rocks. Distinctive aeromagnetic anomalies are also produced by these bodies.

(5) The sedimentary rocks have little or no effect on the gravity field.

(6) From the gravity data, the Luke's Arm fault is a dominant, high-angle feature continuously traceable from the western survey boundary through New World Island and perhaps extends underneath the sea beyond.

(7) The proposed 5-10 km layer might be a landward extension of Sheridan and Drake's (1968) inferred intermediate layer off northeastern Newfoundland. However, it should be further investigated under Notre Dame Bay by methods other than gravity. The discontinuity at Change Islands also requires further geophysical study before it can be accepted as a major structural feature. The whole area north and east of the present survey should be investigated before any hypothesis about the seaward extension of the Appalachians can be considered to be verified.

ACKNOWLEDGMENTS

We wish to thank: Dr. M.J.S. Innes and Dr. D. Nagy (Gravity Division, Earth Physics Branch, Ottawa) for information on previous gravity work in Newfoundland and for copies of their computer programs; Dr. M Talwani (Lamont-Doherty Geological Observatory) for a copy of his model program; Dr. D.E.T. Bidgood (Nova Scotia Research Council) for critically reviewing the results; and at Memorial University: Dr. H. Williams for much geological information; Dr. J.A. Wright for proposing the polynomial analysis; H.C. Weir for discussions and suggestions; D.F. Granter, R.F. Kennedy and L.M. Smith for field assistance; P. Gillard and S.C. Pande for technical help. We are grateful to the National Research Council of Canada for providing a Postgraduate Scholarship (to H.G.M.) and for supporting this research through Grant A-1946.

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27.

ORPHAN KNOLL - A "CHIP" OFF THE NORTH AMERICAN "PLATE"

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Abstract

In June 1970 the Research Vessel GLOMAR CHALLENGER surveyed and drilled a seamount off the Labrador shelf: Orphan Knoll. The relatively flat upper surface of the 100 km-diameter mount lies at about 1800 m water-depth and rises 1000 to 2400 m above the surrounding seafloor. The drill-site at 50° 25.57'N, 46° 22.05'W is Site 111 of the JOIDES Deep Sea Drilling Project and was the first to be drilled during Leg 12. This paper presents a summary of the bathymetric, seismic, magnetic and stratigraphic information collected from the Knoll and interpretations of the data are given.

The mount is a continental high at the margin of the North American plate. It has sunk about 2050 m since Jurassic time, whereas the submerged areas to the south and west may have gone down about 3500 m if the conclusion about their continental nature is correct. Paleontological-sedimentological paleo-waterdepth interpretations suggest that the main subsidence took place during the earliest Tertiary. Seismostratic units are dealt with independently from lithostratic, biostratic and chronostratic units, and the Mesozoic-Cenozoic history of Orphan Knoll has been summarized. The Mesozoic sediments, fauna, and flora are directly comparable with those found in NW Europe. Glaciation began in the Pliocene about 3 million years ago.

The data also inspired conclusions and speculations of wider regional impact. Diapirs seem to be absent from the deeper water north of the Grand Bank-Flemish Cap, suggesting that the Atlantic no longer was an evaporite basin by the time the Labrador Sea began to open. This is consistent with a presented revision of the Bullard *et al.* computer fit of Europe and North America. The revision had to be made in order to account for the continental nature of Orphan Knoll and the area to its west and south.

INTRODUCTION

This paper is intended to give interested Canadian earth scientists an opportunity to learn about Orphan Knoll, which was Site 111 (50° 25.57'N, 46° 22.05'W) of Leg 12 of the JOIDES Deep Sea Drilling Project (DSDP), in an inexpensive and readily accessible Canadian publication. The paper is, in part, a condensed version of Chapter 3 of the Initial Report of Leg 12 (JOIDES, 1972), however, the section on stratigraphy has been expanded beyond the above mentioned initial report and more is now known about the pronounced

*Formerly of Bedford Institute, Dartmouth, Nova Scotia.

bathymetric highs that mark the upper surface of Orphan Knoll. In addition, some detailed description is given about the initial site proposal and selection in an effort to encourage other Canadian earth scientists to consider submitting drilling site proposals to the various Advisory Panels of the DSDP and hopefully to promote further scientific drilling off Canada's continental margins. The Leg 12 site proposals were made by scientists of at least 5 countries and the scientific team which manned Leg 12 represented 7 nationalities. It is apparent that the Deep Sea Drilling Project will welcome the suggestions and the participation of Canadian earth scientists. Finally, the article includes a brief sidelight on the naming of Orphan Knoll.

The Deep Sea Drilling Project is managed by a consortium of 5 oceanographic institutions known as the Joint Oceanographic Institutions for Deep Earth Sampling (JOIDES); one member of the consortium, Scripps Oceanographic Institution, administers the Project. The U. S. National Science Foundation has fully funded the Project through the initial phase and two successive extensions and is investing about one million dollars per month. Drilling is done from the dynamically positioned D. V. GLOMAR CHALLENGER operated under contract by Global Marine Inc. of Los Angeles. The GLOMAR CHALLENGER can maintain position over a beacon on the seafloor to within 200 feet in winds of 35 knots and currents of several knots. Drilling is done with a single pipe and seawater is used as the drilling fluid; there is no return circulation to the ship, thus samples are only obtained when a 9 m core barrel is inserted in the drill stem. Recently a re-entry capability has been achieved, but during Leg 12 it was not yet operative and a hole had to be abandoned when the drill bit was worn.

Location and Background Information for Site 111

Initial Site Selection

In late 1969, the first extension of the DSDP was announced and one of the authors (A. R.) while still at Dalhousie University wrote to the Project tentatively proposing Flemish Cap as a drilling site to settle the question of its oceanic or continental nature. A reply in January 1970 encouraged a more detailed proposal. By that time it was realized that the Bedford Institute did have reasonable proof of the Cap's continental nature (Pelletier, 1969, 1971) and that Flemish Cap did not have the prerequisite 100 m of soft sediment cover generally required to let the drill's bottom hole assembly become firmly embedded before penetrating solid rock.

Thus, in February 1970, a proposal for drilling a seafloor knoll some 350 km north of Flemish Cap and 550 km northeast of Newfoundland (Fig. 3) was submitted to the Atlantic Advisory Panel of JOIDES. At the time, little was known about the feature; only very approximate bathymetric contours could be estimated and no seismic profiles were available. The possibility of the knoll being a continental fragment was advanced in the proposal and the name "Orphan Knoll" tentatively applied. Later, in late March 1970, Xavier LePichon of the French oceanographic institution CNEXO also proposed the Knoll as a drilling site and supported it with a Flexotir continuous seismic profile of CHARCOT 5. The site was selected by the Advisory Panel at about this time on the basis of the CHARCOT record which indicated sufficient sediment cover and that solid geology was within reach of the drill string under the sediment cover (DSDP would prefer targets within 1 km of the mudline). Later, in May 1970 a seismic profile of SACKVILLE 69-041 was made available to the Leg 12 scientific team by Alan Grant of Bedford Institute.

Derivation of the name "Orphan Knoll"

There was considerable controversy over the name "Orphan Knoll" in September 1970 shortly after the completion of Leg 12. Indeed the final publication of the Canadian Hydrographic Service's (1970) Chart 802 was delayed by a few days until the Canadian Permanent Committee on Geographical Names settled the matter. One of the authors (A. R.) had formally proposed the name "Orphan Knoll" to the Committee in March 1970 as an appropriate name for such a small continental remnant lying isolated from its parent continental mass and being a generally neglected feature over the last 50 years of oceanographic exploration. In mid-summer 1970, A. C. Grant submitted the name "Sackville Knoll" to the Permanent Committee in the belief that CNAV SACKVILLE was the first research ship to do scientific work over the feature. In late August 1970, before the drilling results were known, the Permanent Committee had the name "Sackville Knoll" recommended to it by its Sub-committee on Underseas Features Terminology and the original colour proof of Chart 802 was made up with the then recommended name. Upon returning from Leg 12, one of the authors (A. R.) successfully challenged the name and Chart 802 finally appeared with the name "Orphan Knoll" as did Chart 800A (Canadian Hydrographic Service, 1971). The results of the drilling establishing the Knoll's continental nature, the prior publication of the name in the Toronto Globe and Mail, the use of the name in two scientific meetings and earlier research by at least four American oceanographic vessels over the feature were some of the factors in the Permanent Committee's final decision to accept the first-proposed name.

Bathymetry of Orphan Knoll

Orphan Knoll was noted as a single 970 fm sounding in 1917 on the British Admiralty Chart 2060A. The more recent GEBCO sounding sheet No. 27 (Deut. Hydro. Inst., 1964) shows a contoured version of the feature but it is distorted by the inclusion of inaccurate data. Until the publication of the chart of Figure 1 (JOIDES, 1972) the U.S. Hydrographic Office. Chart BC0510N (1965) was the most accurately contoured chart for the feature.

The Knoll appears to be separated into two parts (Fig. 1); the larger southern part rises to depths of less than 1800 m while the smaller northern extension only rises to depths of slightly less than 2400 m. The northeast margin falls directly to the abyssal plain at 4000 m and the feature is separated from the Labrador and Flemish Cap shelves by water 2800 m to 3400 m deep. The northeast side has steep slopes of 30 to 40 degrees; whereas the southwest slopes are more gentle, being in the range of 5 to 10 degrees. The northeast margin is quite linear except for the southern extension where a very steep offshore seamount stands slightly seaward of the main body. It is not known if this peak is composed of the same continental material as the main feature. The western margins of the Knoll are broadly semicircular in outline with a canyon-like feature incised into the southern flank (Fig. 1). The full extent of the Knoll to the northwest and to the southeast is unknown.

Almost every traverse across the Knoll shows a series of pronounced bathymetric highs which interrupt the smooth flat top surface of the Knoll (Fig. 4). These have been interpreted (Fig. 1) to form a series of continuous ridges lying subparallel to the linear northeast margin of the main southern part of the Knoll. The narrow ridge features rise in place to 498 m. Devonian limestone was obtained by dredging across these features (Ruffman and van Hinte, in press).

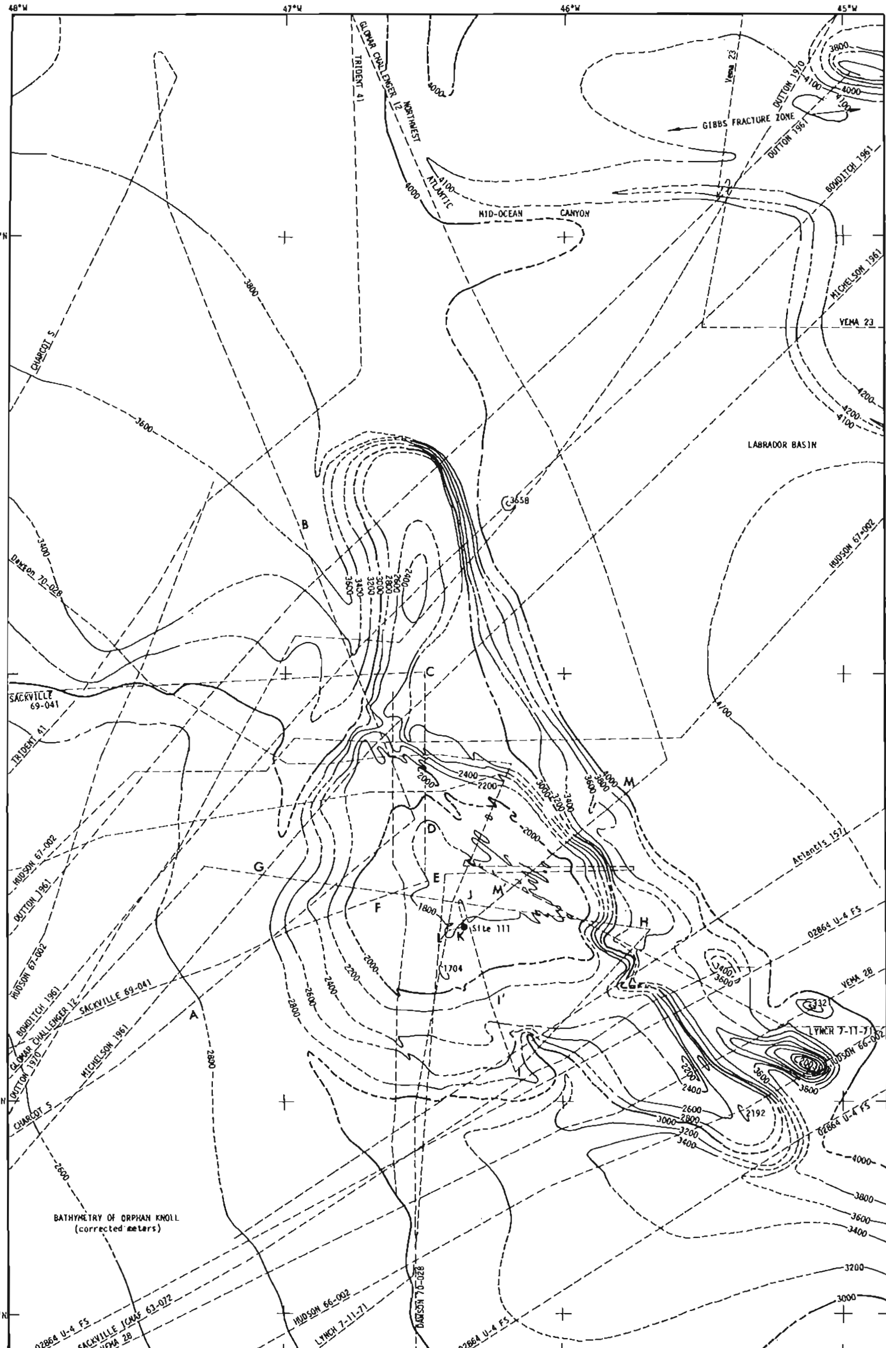

48°W
47°W
46°W
45°W
52°N
51°N
50°N
49°30'N
GIBBS FRACTURE ZONE
MID-OCEAN CANYON
NORTHWEST ATLANTIC
LABRADOR BASIN
BATHYMETRY OF ORPHAN KNOLL
(corrected meters)
Site 111
GLOMAR CHALLENGER 12
TRIDENT 41
VEMA 23
DUTTON 1970
DUTTON 1961
BOWDITCH 1961
MICHELSON 1961
CHARCOT 5
SACKVILLE 69-041
HUDSON 67-002
Atlantis 1571
02864 U-4 FS
VEMA 28
LYNCH 7-11-71
HUDSON 66-002
SACKVILLE (CMAF 63-072
DAWSON 70-028

Figure 1.

The bathymetric map also shows a portion of the Northwest Atlantic Mid-ocean Canyon (Fig. 1). It is located on the basis of reliable data and its position indicates more detail than the map of Heezen et al. (1969).

Magnetic Surveys of Orphan Knoll

A preliminary total magnetic intensity map of the Orphan Knoll area has been prepared (Fig. 2); no secular or diurnal corrections have been applied to the data. The map indicates a linear positive magnetic anomaly lying along the northeast margin of the Knoll and over the small northern extension. However, the southwest margin of the Knoll is not indicated in either the contoured total magnetic intensity map (Fig. 2) or in any individual magnetic profiles that pass over the margin (Fig. 4). The lack of any indication of bottom topography along the southwest margin of the Knoll suggests that there is little contrast in magnetic susceptibility between the main body of the Knoll and the buried seafloor to the west and south. There also seems to be a distinct lack of the oceanic type of magnetic anomalies over the area south and west of the Knoll on the three southernmost aeromagnetic profiles of Hood and Bower (1971) and on the magnetic profiles of ELTANIN 2 (Kroenke and Wollard, 1968). All of the above data strongly suggest that the seafloor between Orphan Knoll and the Labrador Shelf, Newfoundland, Grand Banks and probably Flemish Cap is not oceanic in nature but is rather a continental type of crust.

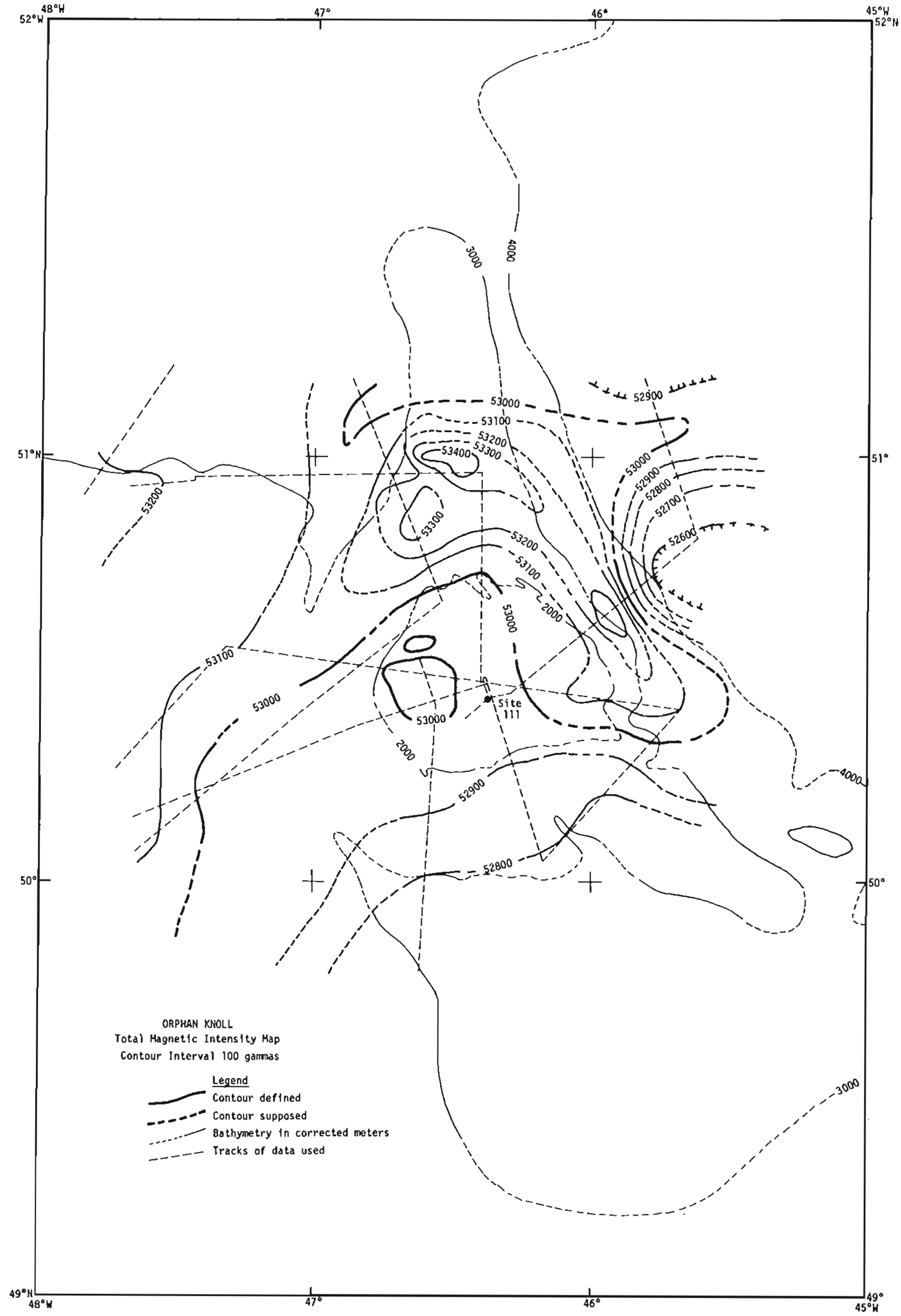
Related Scientific Work in Area

The ATLANTIS in 1949 appears to have been the first scientific ship to obtain a bathymetric profile across the Orphan Knoll (Heezen et al., 1959). In 1961 the U.S. Naval Oceanographic Office (USNOO) obtained bathymetric and magnetic profiles across the feature using the (Avery, 1963). In 1963, DUTTON, BOWDITCH and MICHELSON the SACKVILLE 63-072 cruise (Int. Comm. NW. Atl. Fish, 1968) crossed Orphan Knoll twice in carrying out bathymetric surveys as did HUDSON 66-002 and HUDSON 67-002 cruises in 1966 and 1967 respectively.

There was no further survey work over the Knoll until 1969 when SACKVILLE 69-041 cruise (Grant, 1971) and CHARCOT 5 cruise (Olivet et al., 1970a; LePichon et al., 1971) carried out magnetic and continuous seismic reflection traverses across the feature. LePichon et al. (1971) interpreted the area south of the westernmost extension of the Charlie Gibbs Fracture Zone* (Fig. 1) as foundered continent and Olivet et al. (1970a) noted the Knoll's apparent fault-bounded margins.

Figure 1. Bathymetry of Orphan Knoll in corrected metres. Contour interval is 200 m except over the Labrador Basin where the interval is 100 m. Solid contours indicate that the contour is defined while presumed contours are shown as broken lines. The sources of information are indicated by ship's tracks (thin dashed lines). The exact northwest and southeast extension of the feature is unknown. The minimum soundings to the top of some of the more pronounced peaks are indicated. Letters shown refer to profiles referenced in the text and in the following figures (figure from JOIDES, 1972).

*The use of this name follows the suggestion of Laughton (in JOIDES, 1972).



48°W
47°
46°
45°W
52°N
51°N
51°
50°
49°N
49°
ORPHAN KNOLL
Total Magnetic Intensity Map
Contour Interval 100 gammas
Legend
Contour defined
Contour supposed
Bathymetry in corrected meters
Tracks of data used
Site 111
53000
53100
53200
53300
53400
52900
52800
52700
52600
2000
3000
4000

Figure 2.

In the work of Fenwick et al. (1968) a series of high intensity magnetic anomalies were delineated; they trend at a low angle across the bathymetric contours from the shelf edge at 53° N, 53° W into deeper water at 51° 30'N, 46° W (Fig. 3). These anomalies perhaps should be thought of as the analogue to the shelf-edge magnetic anomalies known along the eastern seaboard except that they now mark the edge of a foundered piece of continental crust. The northern boundary of this foundered block is probably marked by the transform fault of the Charlie Gibbs Fracture Zone (Olivet et al., 1970a, 1970b). The work of Mayhew (1969); Mayhew et al. (1970); Pitman et al. (1971); Pitman and Talwani (1971, 1972) and that of Hood and Bower (1971) along with Laughton (in JOIDES, 1972) is all pertinent in correlating the oceanic anomalies with the original continental margin in the area of Orphan Knoll.

South of Orphan Knoll the Grand Banks and Flemish Cap have been the scene of intense activity by the commercial oil companies and a number of bedrock samples are known. Early work by Hood and Godby (1965) suggested Flemish Cap was oceanic crust, but a recent 15 cm granite core obtained by Pelletier (1969, 1971) at 46° 51'N, 44° 32'W on the crest of the Cap (Fig. 3) yielded an age of 592 ±20 m.y. and has related the Cap to a continental origin. The same cruise dredged quartzites nearby; these were thought to be in place. Carbonates have been dredged in 1481 m (uncorrected at 46° 33.7'N, 44° 25.2'W) (Gilbert, 1967) (Fig. 3); these are thought to be outcrop samples and are Lower Cretaceous in age (Sen Gupta and Grant, 1971). The only other basement outcrops are the Virgin Rocks and the related Eastern Shoals (Fig. 3) that were studied by Lilly and related to late Precambrian rocks on the Avalon Peninsula (Lilly, 1965; Lilly and Williams, 1965; Lilly, 1966a, 1966b; Lilly and Deutsch, 1967). Commercial core holes drilled by CALDRILL 1 on the Grand Banks in 1965 (Fig. 3) have been reported by Swift and Evans (1966) and the two AMOCO test wells, Grand Falls and Tors Cove (Fig. 3) are reported by Pan American Petroleum Corp. (1967a, 1967b).

Deep-water Diapirs

If one considers the locations of all reported deep-water diapirs (Fig. 3) along Canada's eastern seaboard (Watson and Johnson, 1970; Emery et al., 1970; Webb, 1973) and even if one includes shelf structures (King and MacLean, 1970; Keen, 1970; Grant, 1973) and known salt structures (Pan American Petroleum Corp., 1967a; Bartlett, 1969; Bartlett and Smith, 1971) only those diapirs of Pautot et al. (1970) fall north of the Grand Banks. None of Grant's extensive continuous seismic profiling on the Labrador Shelf has recorded diapiric structures north of Grand Banks to Hudson Strait (Grant, 1968; Grant, in preparation).

Figure 2. A preliminary total magnetic intensity map of Orphan Knoll. The ship's tracks along which data was used are shown. No correction was made for secular or diurnal variation. The various surveys were conducted less than a year apart. A pronounced positve anomaly parallels the northeast margin of the Knoll. It is not directly related to the linear ridge structures and may be related to "basement" rocks being brought closer to the surface on the northeast during the mild Jurassic – Cretaceous orogenic period (figure from JOIDES, 1972). See Fig. 1 for correct SACKVILLE 69-041 track.

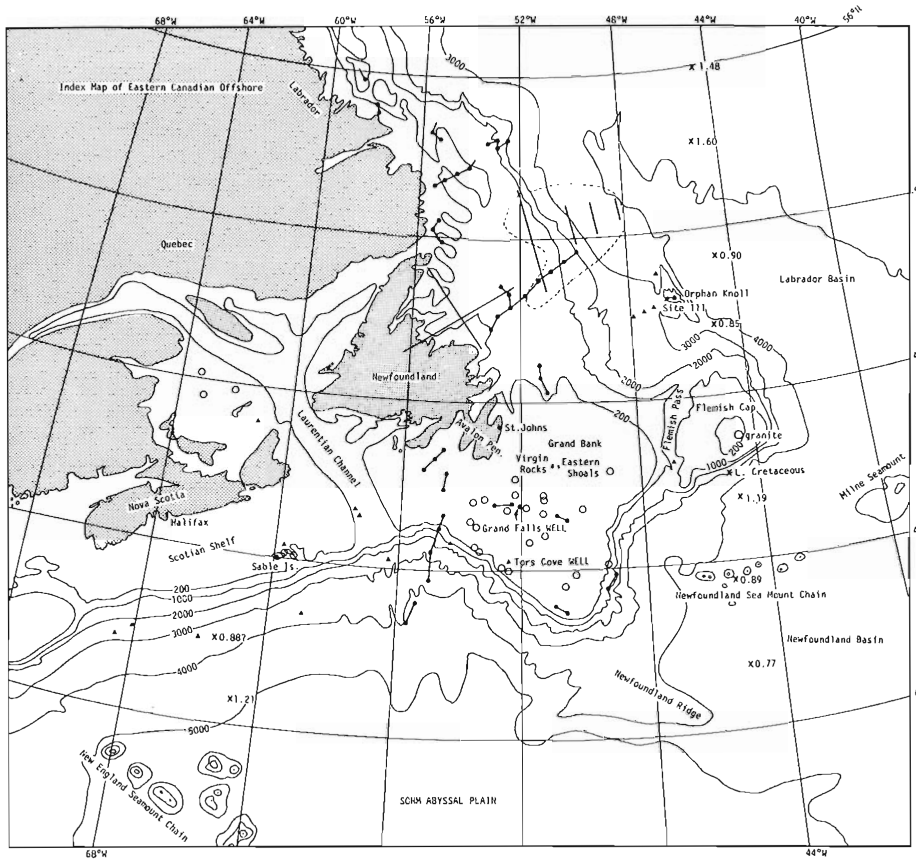


Figure 3. Index map of the area around Site 111 including the Grand Banks showing general bathymetry as taken from the Canadian Hydrographic Service Chart 800A. Contour interval 1000 m except on shelf. Various reported diapiric structures are shown as solid triangles. Lamont seismic refraction lines are shown as solid circles connected by bars (Mayhew et al., 1970). Reported core holes of commercial companies are shown as open circles, and wells are shown as solid circles. Heat flow sites are shown as an "X" with the value noted (Langseth et al., 1970). A Lower Cretaceous dredge site is shown southeast of Flemish Cap along with a granite core site on top of the Cap. The area outlined at 52° N, 50° W is the area of detailed study of Fenwick et al. (1968) and the long solid lines in the area are crustal seismic refraction experiments run by Dalhousie University (figure from JOIDES, 1972).

However, the "diapirs" of Pautot et al. (1970) are not convincing. Their two southernmost structures are very broad and are possibly basement highs; perhaps their structure in Rockall Trough is also a basement high. The northernmost "diapir" of Pautot et al. lies just west of the northern extension of Orphan Knoll (Figs. 1 and 3) on the CHARCOT 5 track. This feature is simply the sub-bottom record of the Orphan Knoll basement as CHARCOT traversed across the buried edge of the Knoll. Thus there are no known deep-water diapirs and perhaps no known shelf diapirs north of Grand Banks to at least the sill of Davis Strait.

It would appear that the shallow saturated saline sea that separated the African and North American blocks during the opening of the South Atlantic had developed a wider circulation by the time the Labrador Sea began to open. As the North American and Greenland - Europe blocks began to separate in Jurassic to Early Cretaceous time, the initial shallow sea separating the blocks was apparently either not confined or not saturated and hence no major amount of evaporite or salt was precipitated over the continental margins and oceanic floor adjacent to the separating blocks. Thus there is no field evidence for deep-water diapirs on the present-day deeply buried and subsided continental edge north of the old African-North American weld. However, north of the Davis Strait sill which appears to be a persistent feature like the Iceland-Faeroes Ridge, the embryonic shallow sea between Baffin Island and Greenland may have been restricted enough to become saturated and to precipitate evaporites.

Seismic Survey Data for Orphan Knoll

Because so little data was available over Orphan Knoll, it was necessary that GLOMAR CHALLENGER spend the better part of a day surveying the site. It was also necessary to tie together the CHARCOT 5 and SACKVILLE 69-041 traverses and to ensure that the site chosen had all important reflectors within reach of the drill string. Hence an approach was made to Orphan Knoll from the west and speed reduced to 8 knots to obtain the best seismic profile possible. The west-to-east traverse passed over the site suggested by the Atlantic Advisory Panel on the basis of the CHARCOT 5 record then continued east to pass over the general site location chosen from the SACKVILLE 69-041 data and off the feature to the east (Figs. 4 and 5).

An approach was then made to the general site location from the south (Fig. 1) and a specific site located while underway using the continuous seismic record. The seismic cables were then recovered, the ship did a slow 180 degree turn and returned to the site by dead reckoning, the ship slowed to zero and a 13.5 KHz beacon was dropped to reach the bottom in about 20 min.

After Holes 111 and 111A were completed the ship made a slow traverse at 4 knots some 10 km southwest of the site, made a 180 degree turn and passed back over the beacon and continued on to the northwest till the abyssal plain was reached before turning north for Site 112 (Fig. 1). The slow speed while profiling southwest of the site provided very clear seismic sections of the whole sediment column drilled at Site 111.

Along all survey tracks run by GLOMAR CHALLENGER 12 bathymetric data was gathered, a proton precession magnetometer was trailed and seismic reflection profiler data was recorded using a 10 cubic inch airgun as a sound source.

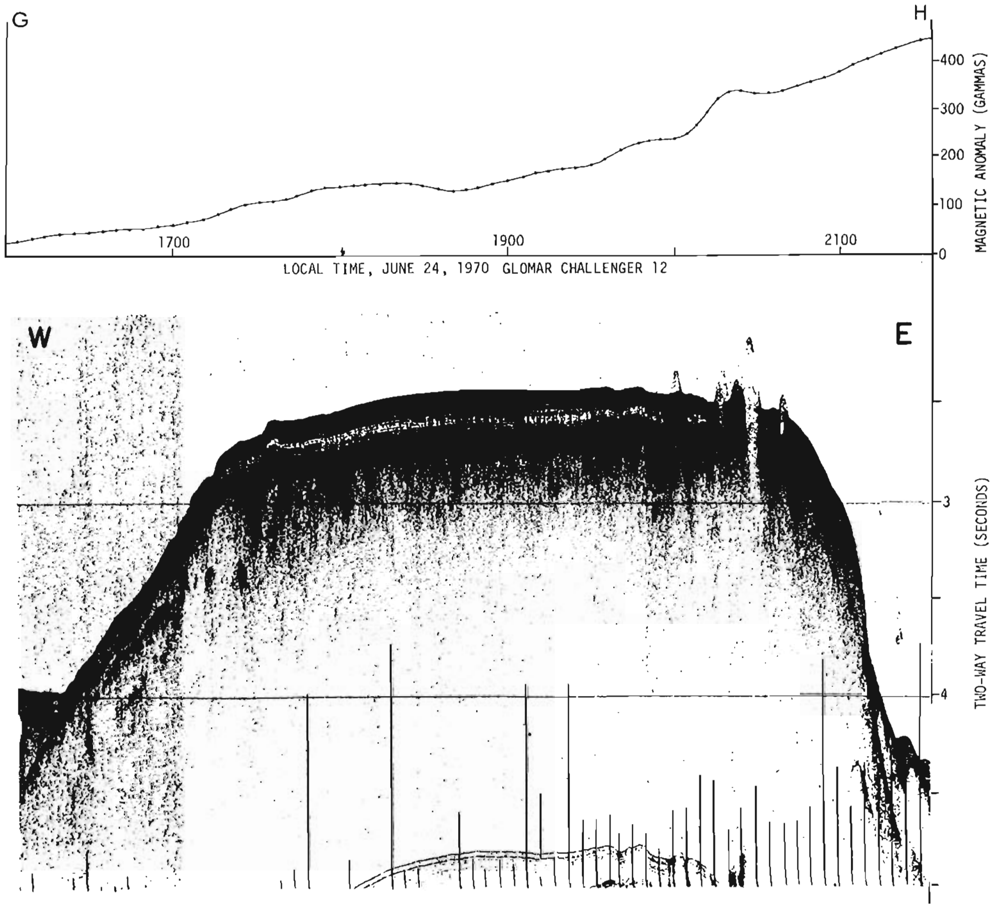


Figure 4. Continuous seismic and magnetic profiles across Orphan Knoll from west to east along line G-H (Fig. 1) of GLOMAR CHALLENGER 12 (vertical exaggeration about 25:1). The narrow ridge-like structures on the eastern edge of Orphan Knoll are very evident on this traverse. The western margin of the Knoll (left) is not evident in the magnetic profile, but there is a slight positive anomaly over the narrow ridge structures on the eastern side (figure from JOIDES, 1972).

Analysis of Seismic Profiles

Various seismic profiles across Orphan Knoll are now available, records from SACKVILLE 69-041 (Grant, 1971), CHARCOT 5 (LePichon et al., 1971; Pautot et al., 1970; R.D. Hyndman, personal communication), VEMA 28 (J. Ewing, personal communication), DAWSON 70-028 (M.J. Keen and I. Park, personal communication) and LYNCH 7-11-71 (Ruffman and van Hinte, in press) are all available in addition to the records of GLOMAR CHALLENGER 12 (JOIDES, 1972). The photograph of the west-east reflection seismic profile (line G-H, Fig. 1) is presented here (Fig. 4)

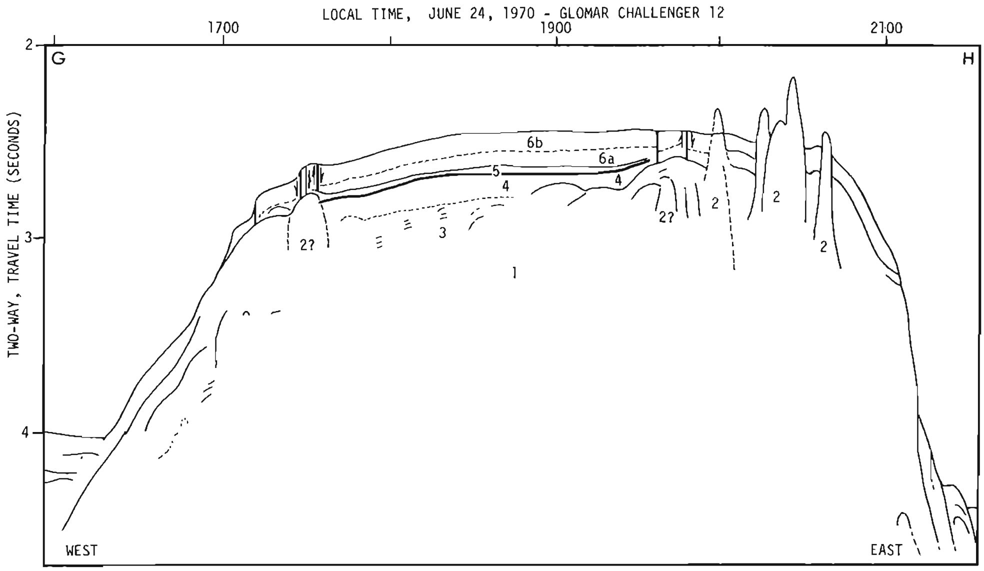


Figure 5. Line interpretation of the west-to-east traverse of Orphan Knoll by GLOMAR CHALLENGER 12 (line G-H, Fig. 1). It is not clear whether a buried ridge-like structure (Seismostratic Unit 2) was seen on the western margin. A series of small normal faults above the possible structure have raised up a low bathymetric ridge on the seafloor. In general the faulting forms two low linear ridges that parallel the trend of the narrow ridge structures (figure from JOIDES, 1972).

along with a line interpretation of the various seismostratic units (Fig. 5). This profile illustrates all of the seismic reflectors except the rather jagged basement, possibly crystalline in nature, that is seen best on the CHARCOT 5 record (JOIDES, 1972) where the low-frequency Flexotir profiling system usually had several seconds of penetration.

In this paper we suggest the terms "seismostratigraphy" and "seismostratic units" to correspond with proposed stratigraphic terminology (Henningsmoen, 1961; Hedberg, 1971). Thus seismostratic units have a distinct characteristic on the continuous seismic reflection profile record and are usually contained between seismic reflectors i.e. are layers with a characteristic seismic velocity. The seismostratigraphy of a site deals with the pile of seismostratic units that comprise the seismic section. The use of these terms recognizes that at times one cannot directly relate a seismostratic unit to a specific lithological unit. This is especially the case in deep sea drilling where the geologist cannot obtain continuous cores and often will never have the opportunity to drill a second hole and record a missing interval. On Leg 12 there was no downhole logging available.

In a later section on stratigraphy the seven seismostratic units will be described and the nature of the reflectors discussed. Suffice to note here that these units are at times very well defined (Unit 5) and at times indistinct and presumed (Unit 1) (Figs. 4 and 5). Very little attempt has been made to correlate the seismostratic units seen on top of the Knoll with those seen off the Knoll. The edges of the feature are often quite steep and appear to be fault-bounded and the beds are discontinuous off the edges.

Southwest of the feature up to 1.7 seconds reflection time of sediment over basement is seen on the CHARCOT record and some apparent slumping is seen south of the site on the margin. To the southwest and west the prograding turbidite flows from the continent are seen infilling against and being dammed up by the barrier that Orphan Knoll presents (Fig. 4). Northwest of the Knoll the CHARCOT record shows over 2 seconds reflection time of sediment. Basement is not seen on the available records beneath the flat abyssal plain northeast of the Knoll.

Drilling Program on Orphan Knoll

Objectives in Drilling Site 111

The objectives in drilling Site 111 may be summarized as follows:

1. to sample, if possible, the crystalline basement and determine whether Orphan Knoll is a foundered continental remnant or a piece of elevated oceanic crust.
2. to sample and date the layers overlying basement and to establish their biostratigraphy and lithostratigraphy.
3. to investigate and identify the prominent seismic reflectors and
4. to establish the time of the change from non-glacial to glacial sedimentation.

All but the first of these objectives were essentially met.

Drilling Logistics for Site 111

The position of the cores taken is seen on the left of the stratigraphic summary (Fig. 8). Drilling operations began at 0445 hours, June 25, 1970 in a corrected water depth of 1797 m at 50° 25.57'N, and 46° 22.05'W. A Christiansen 1927 diamond bit was used below 8 drill collars and 3 bumper subs. The drill stem felt the bottom at 1430 hours, June 25 and a surface core was taken with 55 per cent recovery (Core 1 of Hole 111 is designated as Core III-1). Drilling was easy to 94 m below the mudline where a core of stiff Pleistocene glacial clay was recovered (Core 111-2). Core 111-3 was cut at 189 m and was intended to be above the strong seismic reflector separating Seismostratic Units 4 and 5; however, as soon as coring began the penetration rate slowed markedly and a portion of a very hard layer was recovered; it was a calcareous sandstone of a pre-sinking, shallow-water, non-depositional surface and a short section of glauconitic calcareous sand, all of Cretaceous age. The upper surface of the limestone appeared somewhat nodular and had a weathered appearance and a black manganese-like coating.

Continuous coring was then undertaken (Cores 111-4, 5, and 6) but because of the high pump pressures required, no soft sediments were recovered and only hard rock fragments, often severely rolled and ground, were recovered in the core catcher. In an effort to move deeper in the section an attempt was made to drill ahead 50 m, but after 30 m (at 249 m below the mudline) the progress of drilling was severely retarded and when the centre bit was recovered it was worn flat. A core was immediately taken (Core 111-7) in the belief we were near basement. A short section of apparently unfossiliferous, graded black sandstone over shale was recovered and interpreted to indicate the Knoll's continental nature mainly because of the immature nature of the sediment and its anthracite content.

At 2000 hours, June 26, 1970, the drill was withdrawn to the mudline and within 5 minutes Hole 111A begun. The ship was not offset from a position over the beacon and there was little change in location between Holes 111 and 111A. Hole 111A was intended to recover the missing or highly compressed section of Tertiary between Cores 111-2 and 111-3. At 105 m the first core was taken (111A-1) and a continuous series of cores was taken to 199 m where the presinking limestone horizon was again encountered at 190 m and thus could be correlated with the same horizon in Core 111-3. The rate of coring Hole 111A was considerably slower than that for Hole 111 and it was clear the bit was badly worn. The hole was therefore finished at 2030 hours, June 27, 1970 and the drill string taken inboard and secured by 0300 hours, June 28 and the postsite survey begun enroute to Site 112.

In 19 cores (Fig. 8), 142 m were cut and 74 m recovered for a 52 per cent recovery. Weather varied from rain to sunshine, fog to snow and the winds varied between 15 and 20 knots from the southeast. No ice was sighted. A single 13.5 KHz beacon was required for the 3 days on the site.

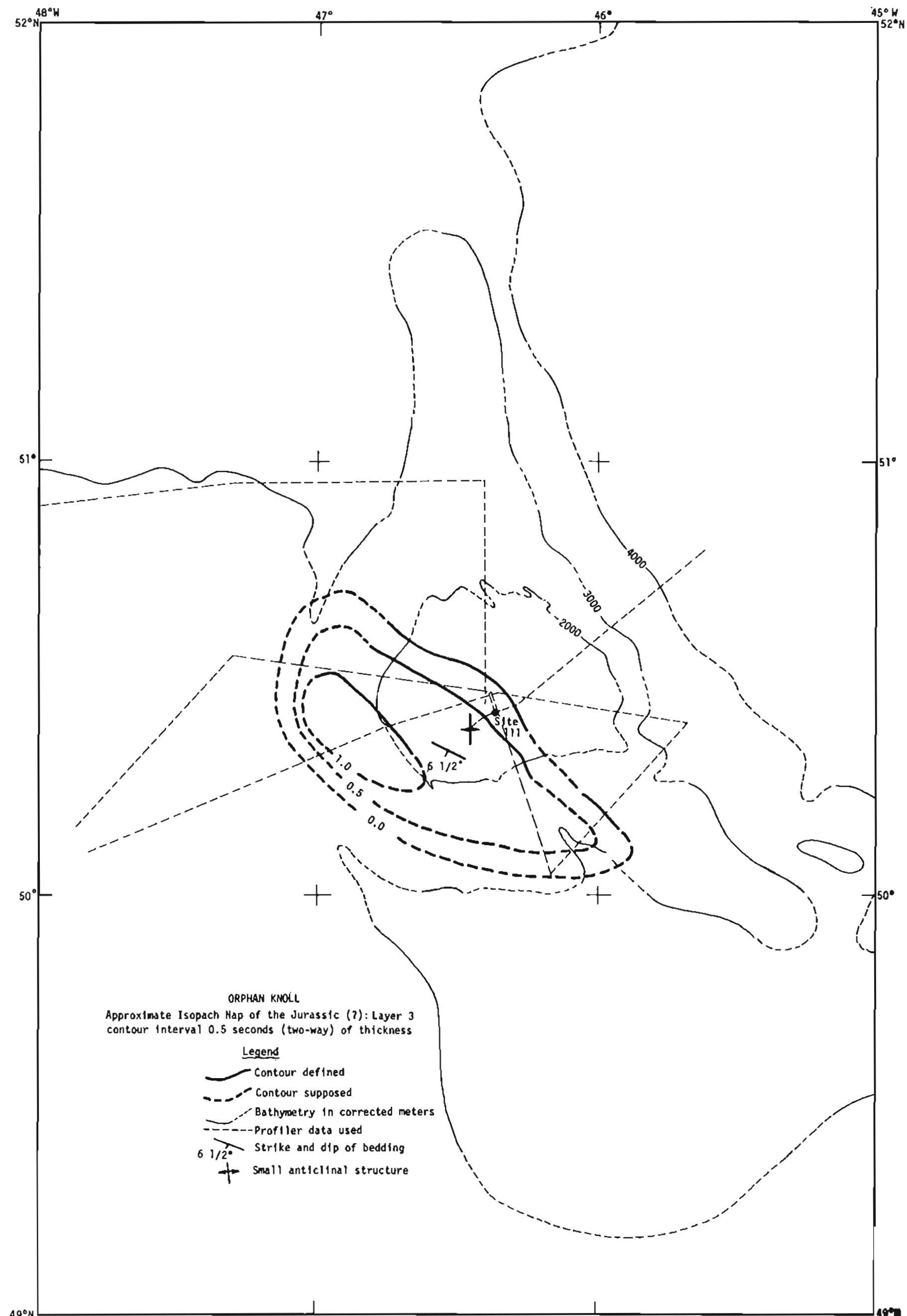
Stratigraphy of Orphan Knoll

Seismostratigraphy is summarized in the profile of Figures 4 and 5. The lithostratigraphy and biostratigraphy of the section drilled on Orphan Knoll is summarized in Figure 8.

Seismostratigraphy

Seismostratic Unit 1 is seldom seen on the GLOMAR CHALLENGER's seismic record and is best seen on the CHARCOT 5 record. There appears to be no seismic energy reflected from within Unit 1 and nothing is known of its structure or its age. The apparent diapiric structures (Seismostratic Unit 2) that rise through all units form the pronounced ridge structures on the Knoll's upper surface. One can clearly see a slight upturning of all reflectors on both sides of this feature (Figs. 4 and 5) down to at least Unit 4. This may well be differential compaction around a series of long standing Devonian erosional remnants (Ruffman and van Hinte, in press; Ruffman, in preparation). Some of these peaks stand 300 m above the flat upper surface (Fig. 1).

Seismostratic Unit 3 is quite indistinct on the CHARCOT records, but shows up quite well on those of SACKVILLE and GLOMAR CHALLENGER. The short record made as the drill ship steamed southwest from Site 111 clearly indicates Unit 3 to be a series of westward-dipping beds that terminate with angular unconformity against the overlying beds of Unit 4. On this short traverse a small anticlinal fold structure was seen in these beds, but it occurs as the vessel turned 180 degrees and is therefore not entirely clear. Using a standard structural diagram and structural contours the strike and dip of the sandstone of Unit 3 over a wide area are averaged to 296 degrees and 6.5 degrees southwest respectively (velocity in sandstone assumed to be 3 km/sec). Both the westward dip and the small fold are thought to be the result of post-depositional tectonic movement. This is perhaps in part confirmed by the approximate isopach map of Unit 3 (Fig. 6). The main body of


48°W
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49°N
4000
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Site
111
6 1/2°
1.0
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ORPHAN KNOLL
Approximate Isopach Map of the Jurassic (?): Layer 3
contour interval 0.5 seconds (two-way) of thickness
Legend
Contour defined
Contour supposed
Bathymetry in corrected meters
Profiler data used
Strike and dip of bedding
6 1/2°
Small anticlinal structure

Figure 6.

the unit lies on the southwestern flank of the Knoll and appears to extend beneath the seafloor close to the southwest wall of the feature. The map is interpreted to indicate that the unit was peneplaned sometime after the orogenic disturbance that induced the westward dip (Figs. 6 and 9). This peneplane may possibly be related to others of the same approximate age (Late Jurassic – Early Cretaceous) known on the eastern seaboard.

Seismostratic Unit 4 is topped by a very pronounced reflector (Fig. 4). The unit overlies the dipping beds with angular unconformity and is a series of nearly flat-lying beds that are generally less than 0.1 second thick and wedge out against basement or the narrow ridge structures (Unit 2). There is evidence of some disturbance by small vertical faults (Figs. 4 and 5) that may be associated with intrusion of the narrow ridge structures or with differential compaction of sediment over the features. Unit 4 is made up of the Albian-Cenomanian sandy carbonates.

The isopach map of seismostratic Unit 4 (Fig. 7) shows the basin to have linear margins on the northeast and southwest. These linear margins could result from post-depositional uplift of the margins and subsequent erosion (uplift perhaps associated with the intrusion of the narrow ridge structures of Unit 2). However, the preferred interpretation is that the Albian-Cenomanian carbonates were laid down in the shallow basin that existed between islands of Jurassic or older rock on top of the Knoll.

Seismostratic Units 5, 6A and 6B could be grouped since their ages almost entirely postdate the sinking and are made up of pelagic material. Unit 5 is quite thin and is seen best on the high frequency records as an even coating over Unit 4 and over some of the sub-bottom ridge structures (Unit 2) that interrupt Unit 4; the more pronounced narrow ridge structures interrupt Unit 4 and all later units.

Seismostratic Unit 6 is everywhere present except over the ridge structures and over the steep margins of the Knoll to the northeast. It can be subdivided into two units, Unit 6A being the seismically transparent lowermost part and Unit 6B the uppermost section seen on the west-east profile (Figs. 4 and 5).

Seismostratic Units 5 and 6 appear draped over the Knoll and conformable with underlying beds and structures as one would expect with pelagic sedimentation. An isopach map of these two units would indicate the section to be thickest on the flat central parts of the Knoll and to wedge out towards the periphery where slopes become too steep to permit accumulation.

Units 5 and 6 are marked by a series of small normal faults that fall over the margins of the Albian-Cenomanian basin (Unit 4). These faults are post-depositional in age and result most likely from differential compaction over the islands of Albian-Cenomanian times. A few of these faults reach the surface and give rise to two low ridges of less than 50 m relief that

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Figure 6. An approximate isopach map of the lowermost seismic horizon identified in Hole 111 (Seismostratic Unit 3). The horizon was dated at Site 111 as Middle Jurassic from 67 cm of core recovered at the bottom of the hole. The tectonically deformed bedding dips at a low angle to the southwest. The whole body has been deformed then peneplaned sometime between Jurassic and Early Albian time (figure from JOIDES, 1972). See Figure 1 for correct SACKVILLE 69-041 track and for CHARCOT tract which was also used.

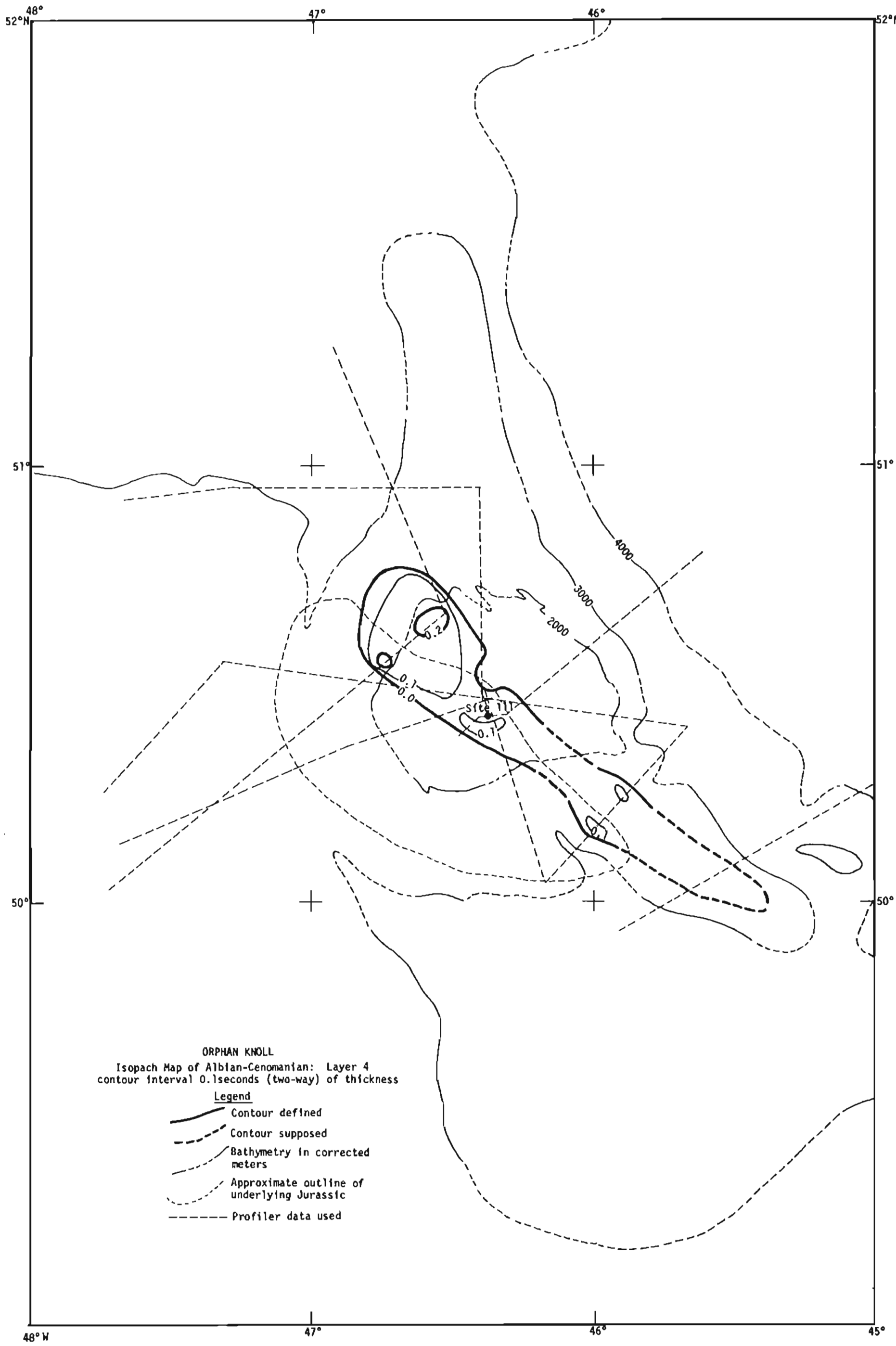
48°
47°
46°
52°N
52°N
51°
51°
50°
50°
48° W
47°
46°
45°
4000
3000
2000
0.2
0.1
0.0
Site 111
0.1
ORPHAN KNOLL
Isopach Map of Albian-Cenomanian: Layer 4
contour interval 0.1seconds (two-way) of thickness
Legend
Contour defined
Contour supposed
Bathymetry in corrected meters
Approximate outline of underlying Jurassic
Profiler data used

Figure 7.

traverse the surface of the Knoll in a northwest-southeast direction (see west side of Fig. 5). It should be recognized that these normal faults in Units 5 and 6 also could result from some post-depositional movement associated with the intrusion of the narrow ridge structures (Unit 2).

Nature of Seismic Reflectors

Prior to drilling, seven seismostratic units were recognized on Orphan Knoll (from bottom to top 1-6B) (Figs. 4 and 5). Four of these (4, 5, 6A, 6B) have been fully penetrated by the drill and were sampled. The well bottomed in the top of the fifth, Seismostratic Unit 3.

The first reflectory at 0.1 sec. (two-way travel time at the drill-site: between Units 6B and 6A) was not cored and its nature is not known. No wireline logging was done. The only continuous record available is the drilling rate. No change was noted in the latter and therefore the presence of a hard layer (for instance, a layer exceptionally rich in erratics) is improbable. Possibly the reflector marks a zone of relative abundance of oozes.

The other reflectors all can be related to unconformities and changes in lithology. It is possible that a reflector at 0.18 sec (between Units 6A and 5) marks the sharp contact between Neogene foraminiferal sand and Paleogene zeolithic clay. It is also possible that the lithologic change marked by the glacially induced increase in terrigeneous content, as cored slightly higher in the section, causes the reflection. This reflector is rather indistinct and is marked by a series of small hyperbolae (Fig. 4).

The very prominent, next lower, reflector at 0.21 sec (between Units 5 and 4) almost certainly occurs at the hard-ground top of a Cretaceous sandy limestone. However, one could also place the reflector slightly higher at the contact between Cretaceous foram sand and Paleogene marl. Both interpretations can be defended. In interpreting the seismic stratigraphy away from the Orphan Knoll, it will make quite a difference which explanation is chosen: either the "top of the Upper Cretaceous" or the "top of the Lower Cretaceous".

The indistinct reflector at 0.29 sec (between Units 4 and 3) is interpreted to be at the top of the hard Jurassic sandstone overlain by softer unsampled sandy dolomitic limestone. The average velocity for the entire section above this reflector is 1.72 km/sec.

Lithostratigraphy

Seven distinctly different lithostratic units were cored; they are from top to bottom (see also Fig. 8):

g) dark and medium grey clay (with erratic pebbles and charcoal) alternating with lighter foram sand (Plio-Pleistocene);
f) white foraminiferal-nannofossil ooze (Plio-Miocene);

Figure 7. An isopach map of the Albian-Cenomanian carbonates (Seismostratic Unit 4). The shallow-water carbonates are interpreted to have been deposited between long linear islands whose margins may, in part, be fault controlled (figure from JOIDES, 1972). See Figure 1 for correct SACKVILLE 69-041 track.

ORPHAN KNOLL - JOIDES DSDP SITE 111

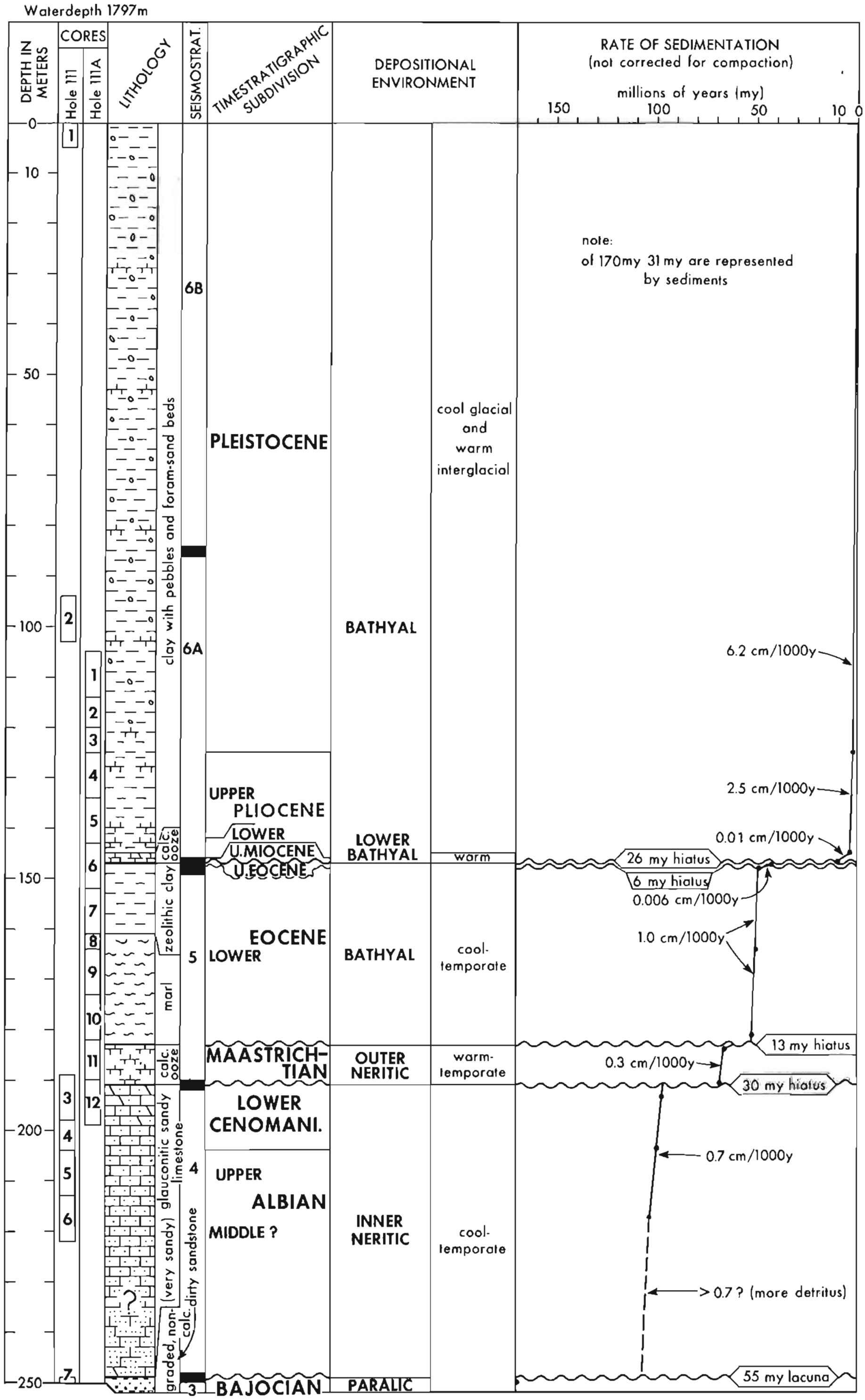

Figure 8.

e) yellowish zeolithic clay with some glauconite bands (Eocene);
d) green-grey siliceous nannoplankton marl (Eocene);
c) brownish white chalky nannofossil-foraminiferal ooze (Maastrichtian);
b) light grey glauconitic sandy limestone topped by a yellowish brown, phosphatic hard-ground (Cenomanian-Albian) with a manganese rich geothite coating on the top;
a) medium to dark grey non-calcareous carbonaceous, immature graded sandstone and shale (Jurassic).

Examples of the different lithologies are shown on Plates I-III. The possible origin of the sediments is briefly discussed below.

The graded sandstone (Unit a) abruptly overlies a silty clay and in its upper part gradually passes into a similar silty clay. The bed could be a turbidite or a channel deposit. The presence of large-scale current bedding and the complete absence of marine fossils from the sands as well as from the shales, lead us to conclude that it is a channel deposit. It might have been a fluvial channel on an alluvial plain (braided river or small point bar), or a fluvio-marine channel on a coastal plain or tidal flat. The carbonaceous nature of the sandstone is due to angular detrital anthracite fragments (reworked, see biostratigraphy) which would not have survived long transport and which suggests near-source fluvial deposition. On the other hand, H. W. Nelson (in JOIDES, 1972) observed that dolomitized micrite is a common constituent in the finer grained part of the bed, which suggests shallow marine conditions. Hence, it can be concluded that Unit a probably was deposited in a channel on a coastal plain not too distant from outcropping older strata; the unit possibly was formed in a paralic realm. It is this short 67 cm section combined with the sinking history that confirms the continental origin of Orphan Knoll. The sandstone can be correlated with the Dogger of Britain and possibly with the recently proposed Mohawk Formation (McIver, 1972) of the Scotian shelf.

The rock of Unit b is a fine-sandy, glauconitic, recrystallized micritic skeletal limestone. The amount of detrital non-carbonate grains (mainly quartz and feldspar) consistently increases downward in the section from about 5 per cent in Core 111-3 to 5-10 per cent in Core 111-4 to 10-20 per cent in Core 111-5 and 20-40 per cent in Core 111-6. Extrapolating this trend it can be assumed that the base of the unit consists of calcareous sandstones. The clean fine sandiness and the skeletal nature of the limestone suggest deposition in relatively quiet shallow marine water. An examination of the fauna confirms this conclusion (see biostratigraphy).

All other sediments cored at Orphan Knoll (Units c to g) are pelagic or at least largely made of the skeletal remains of planktonic organisms. Units c and f are white foraminiferal-nannofossil oozes or chalks. In addition to the skeletal remains the Maastrichtian chalk (Unit c) has minor amounts of very fine silt size terrigenous material; the Neogene ooze (Unit f) is glauconitic, which probably has to do with the very low rate of sedimentation. Unit g, although being pelagic, consists largely of terrigenous material; thick pebbly clays alternate with foram sands. During the Plio-Pleistocene glacial periods

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Figure 8. Stratigraphic summary of description and interpretation of the geological section penetrated at Site 111 on Orphan Knoll (revised from JOIDES, 1972).

large amounts of terrestrial detritus were deposited on the Knoll from the melting ice or icebergs. The foram sands of Unit g probably represent interglacial periods.

Conditions for the formation of calcareous oozes were also unfavourable during the Eocene as appears from Units d and e. The siliceous marl of Unit d does not contain larger grains such as tests of adult foraminifers and it seems probable that the sediment has been sorted, i.e. was deposited under influence of bottom currents rather than being the result of a quiet pelagic rain. Radiolaria and sponge spicules are relatively rare in the marls and if found are corroded. The siliceous material apparently has been mobile after deposition with radiolarian and other siliceous skeletons being dissolved and the silica being concentrated in beds as amorphous chert. This Unit d may be comparable with "Layer A" described in the literature from the Paleocene-Eocene of the NW Atlantic (Ewing et al., 1966; Gibson and Towe, 1971).

The Eocene zeolithic clay of Unit e differs from Layer A by lacking siliceous material. Here the larger and more sturdy foraminiferal tests are preserved and the thinner walled tests are extremely rare; the shells present are fragmented. Apparently this clay accumulated below the lysocline (Berger, 1970; van Hinte, in preparation).

Biostratigraphy and Chronostratigraphy

The lithostratic units distinguished above all have a different, diagnostic microfossil content which makes it possible to date the units and to interpret the environmental conditions under which they formed. The following remarks merely intend to give the reader an idea about the biostratigraphic possibilities and to enable him to judge the degree of reliability of the age determinations (Figs. 8 and 9). Additional details are published in the Initial Reports of Leg 12 (JOIDES, 1972).

The basal clastic Unit a of the section did not yield any calcareous or siliceous fossils. On board ship, the presence of anthracitic coal fragments lead us to conclude a Paleozoic age for the core. Onshore, however, S.A.J. Pocock recovered a mixture of thermally altered Late Paleozoic and an unaffected Middle Jurassic terrestrial microflora from samples of Core 111-7 (in JOIDES, 1972). He concluded that the Paleozoic fossils are reworked with the coal, which fully agrees with the immature lithologic nature of the sediments, and now the core is considered to be paralic Middle Jurassic.

The next higher Unit b (Lower Cenomanian and Albian, including Vraconian) yielded calcareous fossils, foraminifers and ostracods of fair preservation. Most shells are corroded and many glauconite casts are present without the original test being preserved. No calcareous nannoflora was recovered.

Core 111-3, 4, 5, and 6 yielded benthonic forams, ostracods and some planktonic forams. The shallow water fauna has the same species (both forams and ostracods) as found in Western Europe, which is to be expected considering the small size of the Atlantic ocean of those days. The general composition of the assemblage, the plankton/benthos ratio, and the diversity and dominance numbers all suggest that the fauna lived in shallow marine water of a cool-temperate nature. Subsidence may have exceeded sedimentation slightly, since Core 111-6 probably is of inner neritic and Core 111-3 of middle neritic origin.

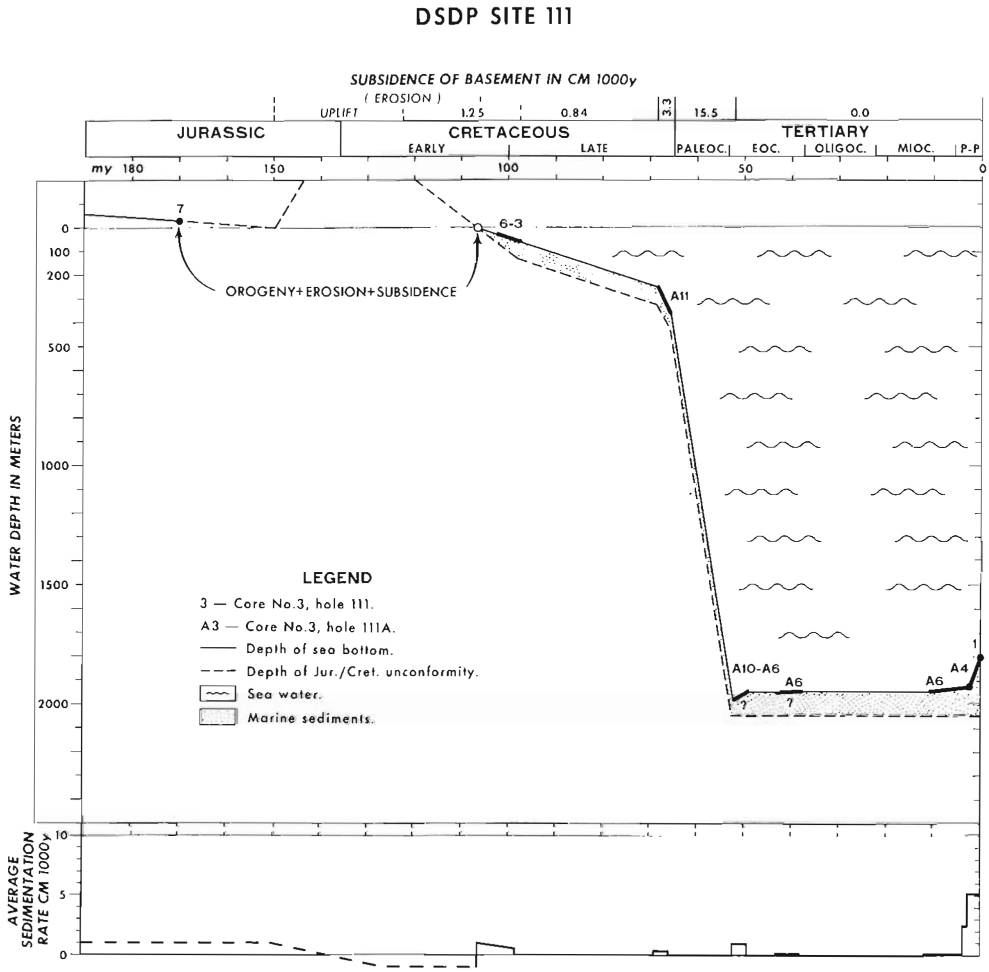

Figure 9. Accumulation of sedimention and sinking history of Orphan Knoll with subsidence and sedimentation rates through time. The Knoll was originally just above sea level 170 m.y. ago then underwent minor structural deformation and tilting; it was then peneplaned by erosional forces. In Early Albian time it sank, in part, below sea level and began to receive shallow water marine sedimentation. The main sinking occurred in the Paleocene and the Knoll has probably remained at essentially the same depth ever since with Paleocene pelagic sedimentation and lately glacial debris adding some 190 m of sediment to build the top to its present depth of 1800 m. The sedimentation rates show the large increase of material during the recent glacial period. Note that the Knoll only has 31 of the last 170 m.y. recorded in the stratigraphic column (modified slightly from JOIDES, 1972).

Planktonic forams in the middle neritic Core 111-3, make it possible to relate the cores to a standard zonation scheme (Rotalipora gandolfii - greenhornensis Zone; van Hinte, 1972). The very shallow marine lower section, however, only yielded very rare (Core 111-5) or no (Core 111-6) planktonics and age determination is based on benthic forams and ostracods.

The age of the section below Core 111-6 is an assumption based on an extrapolation of sedimentation rates (Fig. 9). The rate at which sediments accumulate is expressed in centimetres per 1000 years and is calculated from the assigned absolute age and sediment thickness. It must be emphasized that absolute age assignments for the Cretaceous are not yet well established and that consequently relative age determinations based on the extrapolation of sedimentation rates are not very reliable. However it is better than a mere guess.

The Maastrichtian chalk (Unit c) consists of microfossils and most of these are planktonic forams. Four standard planktonic foram zones can be recognized, upwards these are: Globotruncana gansseri Zone, G. stuarti Zone, G. contusa Zone and Globotruncanella mayaroensis Zone. This succession suggests that three-quarters or more of Maastrichtian time is represented. The faunal assemblage, the foraminiferal plankton/benthos ratio and the diversity and dominance numbers together suggest that the chalk formed in outer neritic (lower part of Core 111A-11) or upper bathyal (possibly upper part of Core 111A-11) rather warm water.

The benthic fauna shows close affinities to Western Europe as well as to the Gulf Coast. The relatively warm nature (diverse plankton includes single keeled Globotruncana and fructicose Pseudotextularia) of the surface water suggests that a proto-Gulf Stream passed over the site. But it must be mentioned that there may have been either climatic fluctuations or changes in the current pattern, for some samples seem to have a cooler fauna with abundant Inoceramus shell fragments. A more detailed study has to be made before conclusions can be drawn.

The actual contact of Cenomanian and Maastrichtian sediments has not been preserved in the cores. But only Maastrichtian faunal elements and no Campanian, Santonian, Coniacian or Turonian are found in core barrel water samples or in the interstices of the hard-ground at the top of the Cenomanian. Therefore, it can be safely assumed that the Maastrichtian unconformably overlies the Cenomanian.

The Cretaceous/Tertiary boundary is not recovered in the cores. However, it is practically certain that the section is incomplete and that most of the Paleocene is missing. The entire Core 111A-10 is of Early Eocene age, whereas all of the immediately succeeding Core 111A-11 is Maastrichtian. However, neither core had full recovery; of the nine metre cored interval, part was lost and the core barrel was not full in either case. If all that was recovered of Core 111A-10 is shifted towards the top of its core barrel and all of Core 111A-11 is shifted towards the bottom of its core barrel, then a gap of 5.4 m is left between the top of Core 111A-11 and the base of 111A-10. This is the maximum thickness of unrecovered section. The minimum is 0.0 m, in which case the contact lies exactly at the base of Core 111A-10; a most coincidental situation. Extrapolating a minimum Eocene sedimentation rate (0.5 cm/1000 y) over the maximum gap (5.4 m) and assuming that only one hiatus is present which is at the top of the Mesozoic and not within the Tertiary, then the oldest possible sediment overlying the Maastrichtian is latest Paleocene. Indeed rare Late Paleocene fossils have been found in core barrel water samples.

The core catcher sample of Core 111A-10 and a sample from the upper part of the same core yielded a few Maastrichtian planktonics. Contamination is a possible, but most improbable, cause for this occurrence and we can assume that the fossils are reworked. Consequently it is not inconceivable that a more complete Paleocene section had been deposited but was eroded.

The Eocene cool-water planktonic and bathyal benthonic foraminiferal fauna is poorly preserved, and confirms in broad terms only the much more detailed age interpretations which were made by K. Perch-Nielsen with calcareous nannofossils (JOIDES, 1972). As determined by her the marl (Unit d) belongs to the Lower Eocene Marthasterites tribranchiatus and Discoaster lodoensis Zones. Sedimentation seems to be continuous despite a lithologic change because the lower part of the zeolithic clay (Unit e) yielded a nannofossil assemblage belonging in the next higher Lower Eocene Discoaster sublodoensis Zone (possibly the sedimentation rate increased from about 1 cm/1000 y to 1.5 cm/1000 y, which is not indicated in Fig. 8). Despite the fact that samples were examined at 2 cm intervals, a number of zones were not detected and about 6 megennia* are not represented, so within the clay a major hiatus occurs. The next younger zone recognized is the youngest Middle Eocene Reticulofenestra umbilica Zone. The lowermost zone of the Upper Eocene is present (Isthmolithus recurvus Zone), but its basal part is missing and a minor hiatus occurs. The Upper Eocene is topped by a rather irregular glauconitic (?) band which represents about 26 megennia; it is covered by an ooze of latest Middle Miocene age (Plate II).

The Plio-Miocene ooze (Unit f) yielded a rich tropical-subtropical well-preserved planktonic foram fauna and calcareous nannoflora. On board ship it was soon apparent that the ooze represented quite a time span and the section was sampled in great detail by K. Perch-Nielsen and W.A. Berggren (JOIDES, 1972). They conclude from their study that the section is complete, at least as far as the stratigraphic resolution of the paleontological tools permit. Therefore a very low rate of sedimentation must be assumed.

The top of the unit is marked by a change in colour, the white ooze abruptly becoming light grey, and 14 cm higher becoming darker and containing obviously ice-rafted material such as lignite. This change represents the beginning of glacial time and thanks to the warm water fossils below it could be well dated to have occurred about 3 megennia ago (base of the Discoaster surculus Zone; planktonic foram ranges).

It is possible that the Plio-Pleistocene boundary was drawn too low in Figure 8 because it is based on the level of extinction of discoasters and these nannofossils are very rare in the glacially depleted flora. Foraminifers suffer from the same depletion problem and with our first investigation it was not possible to make a detailed determination of the boundary. The cold-water flora and fauna are not very diversified to start with and the fossil abundance is further diluted with ice-rafted terrestrial material which brought the sedimentation to the highest rate of the entire section; 6.2 cm/1000 y.

Stratigraphic Conclusions and Discussion

General

The drilling has shown that Orphan Knoll is a part of the North American continent and not an oceanic high. This perhaps will not come

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* 1 megennium = 1 million years (Wickman, 1968).

as a great surprise to geophysicists who are familiar with the magnetic anomaly patterns and bathymetry of the area. The continental nature of Orphan Knoll further necessitates a revision in the Bullard et al. (1965) fit of Europe and North America (Fig. 10). In the 1965 computer fit, Orphan Knoll overlaps onto southwest England and this misfit along with the serious overlaps of Galicia Bank and Flemish Cap (Fig. 10) must be allowed for in future fits. There is further room to spare between the two sides of the Atlantic if one allows for Porcupine Bank and the Hatton-Rockall Basin which both appear continental in nature. Other similar fragments to Orphan Knoll are found south of Hatton Bank. Laughton (in JOIDES, 1972) suggests a revised pre-drift fit (Fig. 11). Any such fit lets one expand the earlier suggestion that large amounts of evaporites do not exist under the deep seafloor north of the Grand Banks to Davis Strait; one can use the fit to suggest that the same lack of evaporites exists off the Greenland margin south of Davis Strait, off Hatton Bank and off the Irish margin.

In fact the drilling results of DSDP Site 111 in conjunction with the magnetic and bathymetric evidence suggest all of the vast area south and west of Orphan Knoll is also continental in nature (Fig. 11). One might argue that Orphan Knoll was itself moved laterally a great distance to the east during the initial rifting phase. If this was the case then the shaded area on Figure 11 would represent in part very early oceanic crust now deeply buried in sediment. This is not the preferred interpretation since none of the necessary east-west transform faults are known between the Knoll and Flemish Cap and since there is no indication of the southern and western margins in the magnetic field. Kroenke and Wollard (1968) earlier suggested a large, deep sedimentary basin north of the Grand Banks.

However, the recovered cores at Site 111 can tell us much more than gross continental detail; they reveal much about what has happened and what has lived at the site during geological time. The information is as relevant in understanding the geology of the Labrador continental shelf and slope as it is in unravelling the geological history of the Labrador Sea and perhaps that of the continental margin of Europe-Rockall Bank.

The structural history of Orphan Knoll and the seamount's biogeographic and ecologic position during time add further interesting details to the description of the phenomenon of continental drift in the North Atlantic, Orphan Knoll is one of the first sources of information on vertical movement of the margin of the North American plate from almost the inception of continental rifting to the present and will be important in the extension of Sleep's work (1969; 1971).

The results obtained from DSDP Site 111 demonstrate for the first time that fragments of continent with their potential reservoir rock can founder to bathyal or abyssal depth; this is quite an encouragement for those petroleum companies who are willing to look for hydrocarbons underneath the deeper seafloor.

The absolute time-scale/fossil zonal schemes used in this paper are 1 a quite reliable scheme presented by Berggren (in press) for the Tertiary and 2. an, as yet, less certain scheme for the Cretaceous compiled by van Hinte (1972).

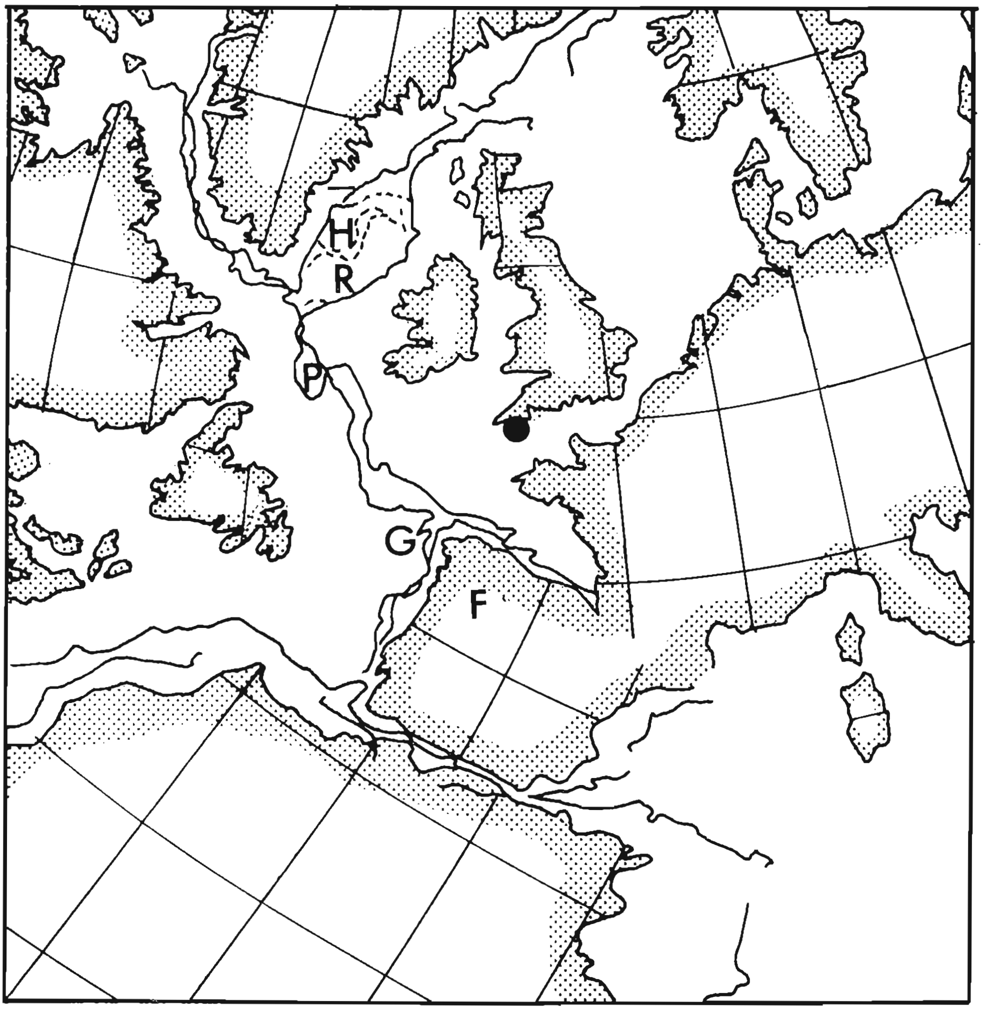


Figure 10. The fit of Europe and North America from Bullard et al. (1965) to give the ancient continent of Laurasia. Orphan Knoll is shown as a black dot overlapping onto the Devonian of Cornwall. Rockall Bank (R) and Hatton Bank (H) are shown fitted together eliminating the Hatton-Rockall Basin. Porcupine Bank (P) and Galicia Bank (G) are both shown overlapping onto North America. Flemish Cap (F) is shown badly overlapping onto northern Spain. These overlaps are now known to generally involve rock of a pre-drift age and this, along with other data, makes the above large overlaps unacceptable (figure from JOIDES, 1972).

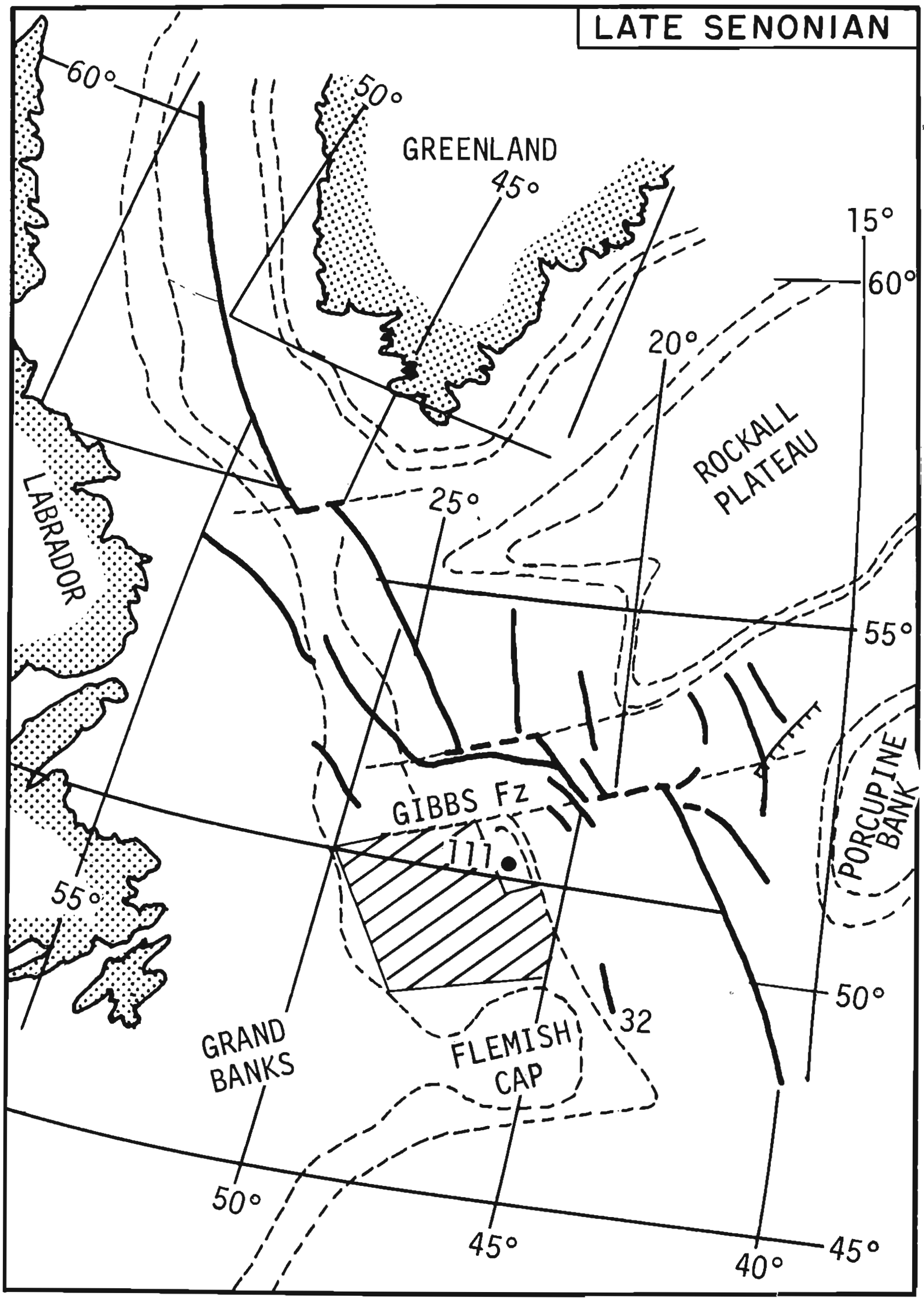

LATE SENONIAN
60°
50°
GREENLAND
45°
15°
60°
20°
ROCKALL PLATEAU
LABRADOR
25°
55°
PORCUPINE BANK
GIBBS Fz
111
55°
50°
32
GRAND BANKS
FLEMISH CAP
50°
45°
40°
45°

Figure 11.

Paleozoic Formations

Seismostratic Unit 1 is possibly made of folded Paleozoic sedimentary rock which forms the "seismic basement" of the Knoll. All we know about the Paleozoic comes from dredge samples (Ruffman and van Hinte, in press) and from reworked material in younger deposits (Pocock, in JOIDES, 1972). During the Devonian, the Knoll was part of a shallow, warm marine sea in which carbonates accumulated. During the Mississippian conditions were continental or paralic and coal was formed. This seems to be in agreement with present theories for the proto-Atlantic given for instance by Kay (1969) and Schenk (1971). The coal is anthracite (Hacquebard, in JOIDES, 1972) and has been tentatively related to a source similar to the coalfields of South Wales (Bloxam and Kelling, in JOIDES, 1972).

Mesozoic Formations

The paralic Jurassic recovered from Core 111-7 probably is part of a shallow marine or marginal continental sequence overlying the Paleozoic with an angular unconformity. The seismic profiles strongly suggest that the Jurassic (Seismostratic Unit 3) is a slightly folded westward-dipping wedge and forms, with the Paleozoic, part of the basement, as far as the younger history of the Knoll is concerned. The wedge shape of the Jurassic can be understood if it is considered to be the peneplaned remnant of a larger body which was preserved from erosion in a depression. The Jurassic may have developed its westward dip when the Knoll became an isolated feature some time between the Jurassic and the late Early Cretaceous. Hence the remainder of the wedge of relict Jurassic has moved relatively downward. It seems reasonable to assume that the various post-Jurassic movements of the Knoll are related to faulting and continental-edge downwarp connected with the initial break-up of the super continent of Laurasia and the subsequent opening of the Labrador Sea between Canada and the Greenland-Europe block.

The Jurassic is in its turn unconformably overlain by a sucession of younger rocks which did not undergo tilting or folding and have remained essentially horizontal until the Recent. The oldest sediments of this section are most likely of Early Albian i.e. Early Cretaceous age. The isolated Knoll probably was subaerially exposed for some time in the Early Cretaceous until sedimentation (marine) recommenced as a consequence of gradual subsidence. The seismic data reveal that the shallow marine sediments filled local depressions on the Knoll. As far as can be ascertained from the seismic reflection results the Cretaceous sediments did not cover the highest parts of the Knoll.

Figure 11. A reconstruction of the North Atlantic taken from Laughton (JOIDES, 1972). The black stripes indicate linear magnetic anomalies on the ocean floor; transform faults are indicated as heavy dashed lines with their older extensions as a lighter dashed line. The estimated shelf edge and bottom of the continental slope are shown as dashed lines paralleling the continental margins. The shaded area west and south of Orphan Knoll (Site 111) is thought to be a vast area of subsided continental crust or possibly an area of primeval oceanic crust now deeply buried by turbidite sequences (figure from JOIDES, 1972).

Early Cenomanian sedimentation possibly ceased because of bottom currents which increased in strength and swept the clastic material as well as the skeletons of organisms away from the Knoll. Although subsidence continued, no sedimentation took place and a hard-ground developed on top of the Cenomanian. However it is possible that sedimentation did continue until as late as the Campanian, and that only then did the current increase and erode everything younger than Cenomanian. Then slower currents just prior to the Maastrichtian might have enabled the hard-ground to form during a relatively short period before the currents further decreased and sedimentation recommenced in the Maastrichtian. However, this possibility is not thought to be the case since no erosional features were seen on the hard reflector believed to mark the top of the Cenomanian (Fig. 4).

In looking at the Cretaceous carbonates, one should remember that the European continent was at the time probably relatively close at hand and at the most only separated from Orphan Knoll by a narrow ocean. One would expect close similarities with the European section and indeed the benthic foraminiferal and ostracod faunae are "all-European", and so is the scarcely recovered macrofauna (Jeletzky, in JOIDES, 1972) besides similarities are seen in the sedimentary succession (James and Hopkins, in JOIDES 1972).

There is some suggestion that currents were still fairly strong during Maastrichtian time because the presence of fine detrital quartz in the Maastrichtian is difficult to explain otherwise if the Knoll was an isolated feature. Furthermore, the assumption that currents were active also makes it easier to understand why skeletal material accumulated at the low rate of 0.3 cm/1000 years rather than at about 1.0 cm/1000 years which would be expected from the rich, relatively warm-water fossil assemblages found to constitute the Maastrichtian chalks.

Mesozoic/Cenozoic Boundary

The Maastrichtian probably was a time of relatively rapid subsidence of the Knoll. However, one of the most interesting findings at Site 111 is that during the Paleocene the site foundered rapidly and subsided over 1000 m (Fig. 9). Unfortunately no Paleocene sediments have been preserved, which makes it impossible to obtain more than a general indication about the rate of subsidence. If subsidence was gradual and took all the time of the hiatus (13 m.y.), the minimum rate was 15.5 cm/1000 years which is not impressive if one considers that horizontal movements of oceanic crust are only of the order of a few centimetres per year. However, it is possible to speculate on an alternative course of events as follows.

If the entire Maastrichtian/Paleocene hiatus is due to nondeposition then a current having a velocity of some 10 cm/sec must have swept the site continuously during Paleocene time in order to prevent deposition and to remove sand-size benthonic foraminiferal tests (Heezen and Hollister, 1971, p. 354). However, the effect of a stronger and erosive current which was active during a shorter time could also have resulted in the same hiatus. In fact such a situation would explain why no glauconitic layer, no other residual deposit, no encrusted layer or no hard-ground covers the top of the Maastrichtian (such features cover the top of the Eocene and the top of the Cenomanian at this site). If in addition one considers that the presence of reworked Maastrichtian fossils in the Eocene Core 111A-10 might suggest that erosive deep-water currents existed during the Early Tertiary then it becomes not inconceivable that more Upper, and perhaps also Middle and Lower Paleocene had originally been deposited, but was eroded. It can be expected (Bartlett and Smith, 1971; Gibson et al., 1968) that such Paleocene would be comparable with the Early Eocene rather than with the Maastrichtian. Therefore it seems to be conceivable that the rate of subsidence in the Early Paleocene was higher or even very much higher than the minimum figure of 15.5 cm/1000 years given above.

One could speculate that the large Paleocene subsidence was not an isolated phenomenon. It coincides, for instance, with the break away of Greenland from Europe, and the development of the triple point in the Labrador Sea (Hood and Bower, 1971; Laughton, in JOIDES, 1972). If the Paleocene subsidence was more than a local phenomenon and happened elsewhere along the Atlantic margin as well, it would account for a large volume of water and may explain the widespread Cretaceous/Tertiary regression known from shallow marine circum-Atlantic and Tethyan areas. Preliminary calculations suggest that, certainly to better than an order of magnitude, this explanation of the regression can be valid.

Post-Paleocene Formations

During the Eocene and later, pelagic sediments quietly accumulated at the site and no post-Paleocene vertical movements are apparent in which basement is involved. This does not mean that a monotonous "continuous" section formed on the Knoll; to the contrary, a variety of sediments is found and extensive hiatuses occur. Deep-water currents must have been passing over the site at velocities that varied with time as a result of changes in climate and in seafloor topography.

The Plio-Pleistocene material collected at Orphan Knoll is extremely interesting and much can be learned about North Atlantic glaciation and related phenomena. It is the subject of special studies by W.A. Berggren (JOIDES, 1972); the onset of glacial material is indicated at about 3 m.y.

Some Speculation on Circulation

The Albian-Cenomanian microfauna is a cool-water fauna, comparable with assemblages found in northwest Germany and the Paris Basin. The Maastrichtian fauna on the other hand has a considerably warmer character; the Early Tertiary again yielded a good planktonic flora and fauna. These differences can reflect climatic changes but may also be explained as a function of oceanic circulation. One could imagine that the shallow-water Albian-Cenomanian was deposited under the influence of a current similar to the present cool Labrador Current (Hachey, 1961). In addition wave action was sufficiently strong to erode the higher parts of the Knoll and the lows were filled with the erosional products and carbonate. Once below wave base the currents merely kept the Knoll clean of sediments and the hard-ground developed; or current velocities varied and the hard-ground developed later.

Cretaceous subsidence (or sea-level rise) brought the Knoll farther away from coastal currents and at the same time made it possible that an early equivalent of the relatively warm Gulf Stream could reach farther north. Possibly the Grand Bank and Flemish Cap were more deeply submerged and acted as less of a barrier. During Maastrichtian time this trend has gone so far that the site possibly was below a zone of convergence between the cool south-going current and the warm north-going current. The convergence caused velocities to decrease and permitted Upper Cretaceous chalks to accumulate.

The Paleocene faulting and consequent emphasis of relief perhaps made the shallow barrier of Grand Bank and Flemish Cap more effective again and the warm Gulf Stream (or equatorial current; Ramsay, 1971) was bent to the east towards its present position and no longer reached as far north as Orphan Knoll. At the same time, cold Norwegian Sea water entered the Atlantic because of the break-up between Greenland and the Rockall Bank, and the basic pattern of the present day deep contour current system, including its Labrador Sea part, was initiated (Berggren and Phillips, in press). The Eocene marls and clays probably were deposited from the margin of such a current.

The Mio-Pliocene fossil assemblages again are indicative of warm surface waters. But, there seems to be no reason to assume that this is the result of an earlier direct Gulf Stream equivalent and the apparent warming may be the result of different climatic conditions which reduced the significance of the Labrador Current and permitted northern migration of the warm Gulf Stream fauna.

ACKNOWLEDGMENTS

The authors would like to recognize the officers and crew of the D. V. GLOMAR CHALLENGER without whom Leg 12 would not have been successful. We thank the National Science Foundation which funded the Deep Sea Drilling Project, JOIDES which manages, and Scripps Institution of Oceanography which administers the Project. The staff of the Deep Sea Drilling Project were most helpful and we would especially like to acknowledge the cooperation of the technicians on board ship. We are most grateful to our own employers at the time, who were very generous in encouraging participation on Leg 12 and supported subsequent research. Of course we salute the most valuable and enthusiastic contributions of the other eight members of the Leg 12 scientific team: the chief scientists A. S. Laughton and W. A. Berggren and the scientists, R. Benson, Th. A. Davies, U. Franz, L. Musich, K. Perch-Nielsen, and R. B. Whitmarsh. We gratefully acknowledge the use of shore reports by T. W. Bloxam and G. Kelling, P. A. Hacquebard, N. P. James and J. Hopkins, J. A. Jeletzky, H. W. Nelson and S. A. J. Pocock. The JOIDES Atlantic Advisory Panel should be credited for having considerable geological foresight and modicum of luck in selecting Orphan Knoll as Site 111 in the face of scanty hard scientific data.

Plates I to III

PLATE I

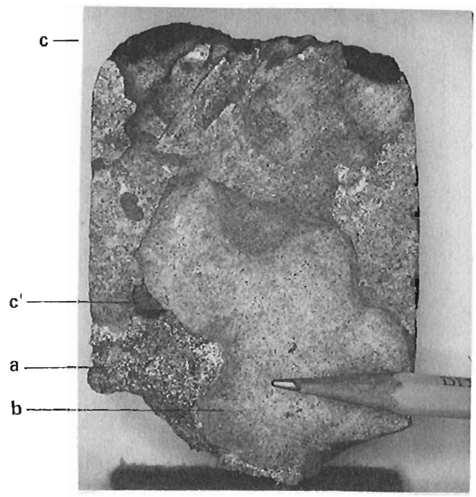
1. PLEISTOCENE, Hole III, Core 2, 94-103 m, Section 6, 81-105 cm (oriented). The composite photograph shows terrigenous glacial clay with pebbles and carbonaceous material, interbedded with light grey interglacial foram sand or ooze (all lithostratic Unit g). The coiling direction of the planktonic foraminifer Globigerina pachyderma is dominantly sinistral in the clay and dextral in the ooze. The original bedding has been disturbed in the coring process.

2. The phosphatized hard-ground at the top of the Cenomanian (Lithostratic Unit b). Hole 111, Core 3, Section 1, top 8.5 cm (oriented).
   (a) Softer material, rich in Maastrichtian fossils, fills original fissures and holes between,
   (b) weathered upper section of the resistant Cenomanian limestone,
   (c) which is coated by a black manganese-rich goethite encrustation. Note the infilling of a similar material in a former hollow (C^1).

3. Shallow marine sandy glauconitic limestone (Unit b) of Hole 111, Core 4 (Section 1, 8-10 cm), CENOMANIAN. The arrows point to a fish scale, a fish tooth and pelecypod shells.

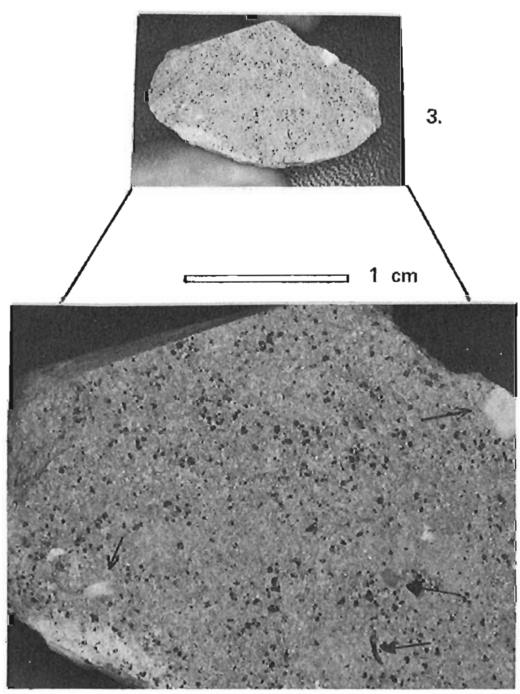
PLATE I

1.



2.

5 cm



3.

PLATE II

Hole IIIA, Core 6, 143-152 m, Section 3, 65-111 cm.

The composite photograph shows the contact between yellowish Eocene zeolithic clay (Lithostratic Unit e) and white Miocene ooze (Lithostratic Unit f). The glauconite band at the contact represents about 26 megennia. The clay is of Late Eocene age, the lower 19 cm shown in the photograph practically cover the complete *Isthmolithus recurvus* Zone which is thought to represent 4 megennia. The ooze is at its base Middle Miocene (*Discoaster hamatus* Zone) and at the top in this photograph Late Miocene (high in the *Discoaster quinqueramus* Zone), and the section in the photograph represents 6 megennia of sedimentation. The age determinations are based on calcareous nannofossils and were made by Professor Dr. K. Perch-Nielsen.

PLATE II

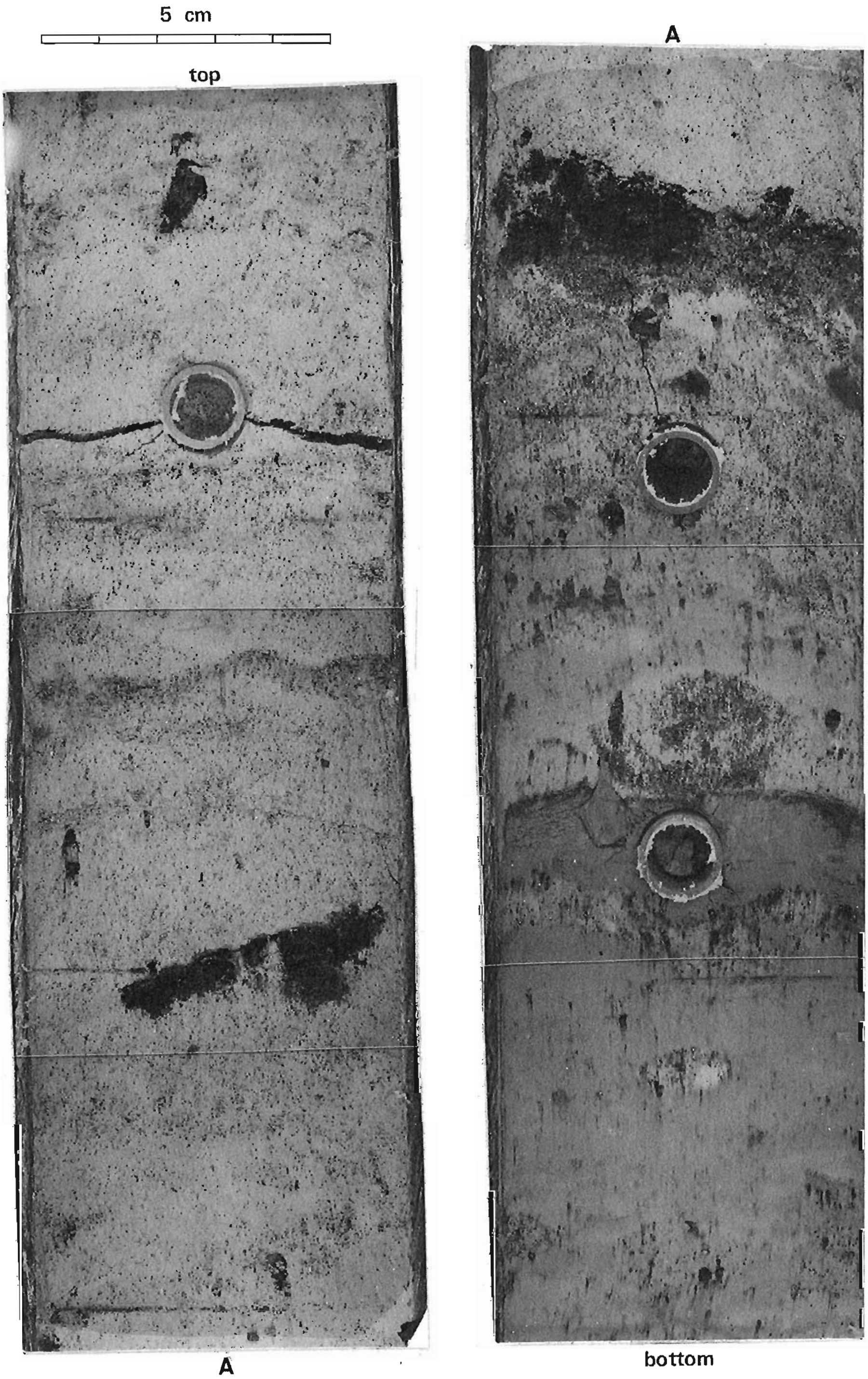
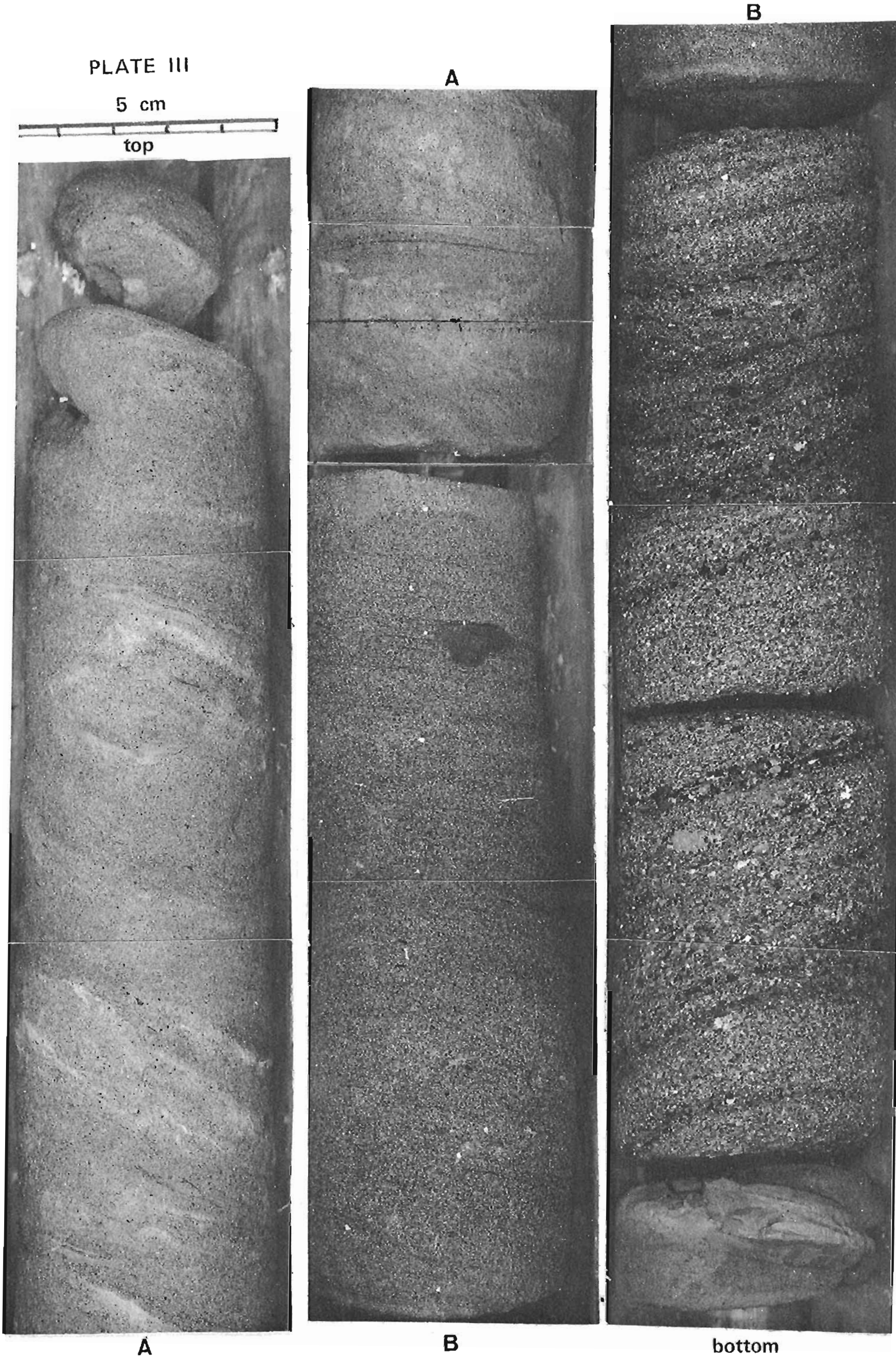


PLATE III

Hole III, Core 7, 249-250 m, total core recovered.

DOGGER. Graded sandstone bed in sharp contact with underlying silty shale and gradationally fining towards shaly top (Lithostratic Unit a). The black spots are coal fragments. The coal is anthracitic and probably originally of Mississippian age. The shales are barren of marine faunal or floral elements. This fact and the presence of large-scale current bedding in the coarser part of the bed suggests that it is a channel deposit rather than a turbidite.

PLATE III



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(Earth Science Symposium on Offshore Eastern Canada, Geol. Surv. Can., Paper 71-23, 1973, p. 451-468)

28. SURFICIAL GEOLOGY OF LABRADOR AND BAFFIN ISLAND SHELVES

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Abstract

The shelves of Labrador and Baffin Island are smooth plains in contrast to the rugged, glaciated and fiord-indented land to the west. These shelves are mainly underlain by sedimentary rocks that have also been glaciated. The abrupt boundary between Precambrian rocks and shelf sediments occurs at tectonic hinge lines which are marked by the marginal channels described by the Holtedahls. Eastward-moving continental glaciers, underloaded with debris, became loaded with available erodable sediments at the trace of the tectonic hinge line thus causing the marginal channels.

Hundreds of bottom samples from Baffin Bay and Labrador Sea have been described in the literature since 1900. The extent of the sedimentary component of these grab samples and gravity cores seems to coincide with the presence of sedimentary bedrock. The idea that icebergs dropped all of the material seems unlikely because sedimentary debris has not been found in previously flooded land areas and because the sedimentary component is absent in certain areas where icebergs are abundant. Furthermore, if all of the icebergs deposited all of their debris on the Labrador Shelf, only a few inches of sediment would have been deposited in the last 10,000 years. It is tentatively concluded that the sedimentary component of bottom samples is part of the ground moraine which has moved only a few miles from its origin elsewhere on the shelf.

Analysis of alleged ground moraine tends to show that in the southern part of the Labrador Shelf at 53° N the bedrock is of Tertiary age. Progressively northward the rocks are older. At 58° N, only Lower Cretaceous rocks are present on the western part of the shelf.

INTRODUCTION

The first purpose of this paper is to review the concepts that have been offered through the years to explain the marginal channels which occur eastward of the mainland of Labrador, Baffin Island, Bylot Island, Devon Island, and Ellesmere Island. In eastern Canada, these submarine troughs approximately mark the boundary between fairly rugged highlands on the west and even, submarine plains on the east. The submerged plains are fragmented by deep transverse channels. The significance of these features in northern latitudes was first mentioned by Olaf Holtedahl (1929) but his best-known work concerning eastern Canada was published in 1950. Later Hans Holtedahl (1958) discussed their geomorphology.

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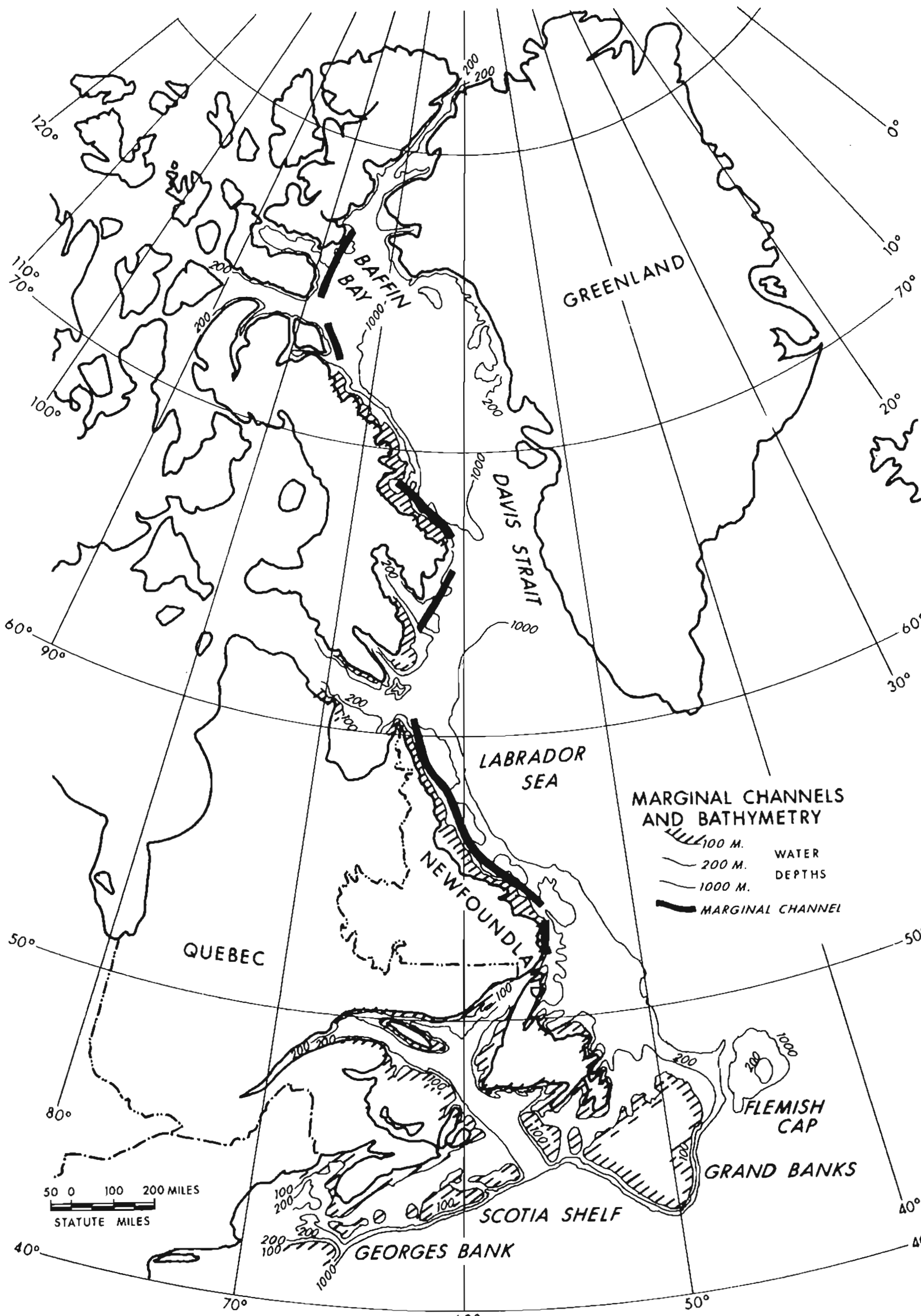


Figure 1. Marginal channels and bathymetry

The second purpose of this paper is to present the ages of rock debris selected from dredge samples and piston cores taken from the shelf and continental slope of Labrador in 1967 jointly by Tenneco and the Atlantic Geoscience Centre.

GEOMORPHOLOGY

The Mainland

From the northern tip of Newfoundland northward to Ellesmere Island, the continental shelf is separated from the highlands of the mainland by a discontinuous marginal channel or trench (Fig. 1). Except for small areas of Mesozoic and Tertiary sedimentary and basaltic rocks on Baffin Island, Bylot Island, and Ellesmere Island the entire region is almost all Precambrian terrain, consisting of acid plutonic rocks and gneiss associated with lesser amounts of diorite and gabbros. The whole is overlain at widely separated intervals by younger sedimentary Precambrian rocks which are mainly unmetamorphosed. This mainland is indented by fiords and deep bays, some of which probably are manifestations of grabens. Some of these physiographic features and their possible relationship to regional tectonics have been described by Kerr (1967). The shore facing Baffin Bay, Davis Strait, Labrador Sea, and Gulf of St. Lawrence to the south is steep. The slope of the land surface away from the sea, that is to the west and north, is a gentle one. The surface was called the "Laurentian Peneplain" by Wilson (1903). Cooke (1929, 1930, 1931) described this surface and concluded that it was uplifted to its present position in the Tertiary. Elevations in Labrador range between 1,000 and 5,000 feet near the coast. In Baffin Island the elevations are much higher. The Tertiary uplift was probably associated with block faulting (Cooke, 1929, 1930, 1931; Kranck, 1947; Ives, 1957).

After the uplift the areas were subjected to several episodes of continental glaciation in the Pleistocene. These ice sheets altered the block-faulted peneplain.

The Continental Shelf

The bathymetry of the continental shelf contrasts markedly with the rugged subaerial territory to the west. The shelf is the submerged coastal plain which has its southern limit in the Gulf of Mexico. South of Cape Cod parts of the plain occur on land but north of Cape Cod it is almost entirely below sea level. South of the 48° N parallel, substantial areas of the shelf are overlain by water less than 200 feet deep but north of this latitude, which passes north of Flemish Cap, the continental shelf is much deeper, being mostly 500 to 600 feet. In general the Canadian portion of the shelf becomes narrower from south to north. Eastward of the northern tip of Newfoundland, for example, the shelf is more than 200 miles wide, whereas off Baffin Island it is barely 20 miles wide in places. The generalization of an ever narrowing shelf as one proceeds northward does not appear to apply north of Bylot Island because a distinct break in slope between shallow and deep waters does not occur (see maps in Pelletier, 1966).

The submerged coastal plain is fairly flat, the relief ranging mainly between 100 and 200 feet. Individual banks are separated from each other by

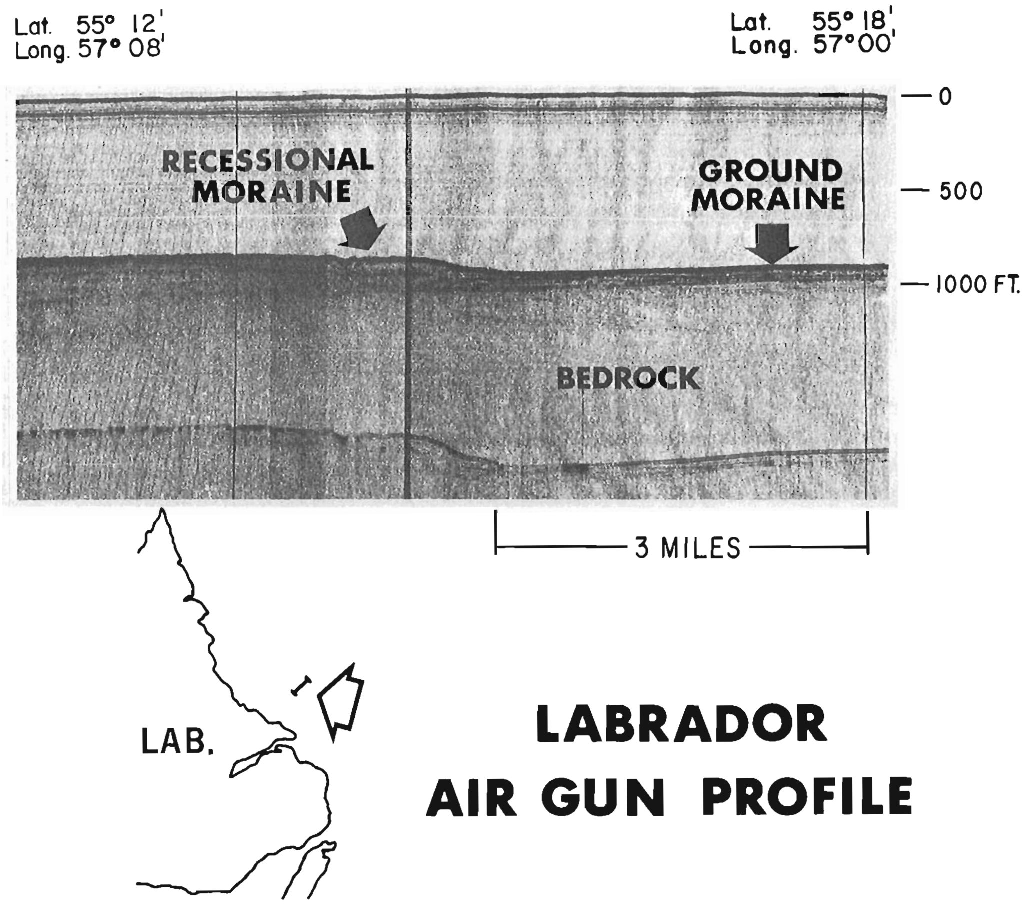


Figure 2. Veneer of ground moraine with recessional moraine on Mesozoic or Tertiary sediments. Profile recorded with an air gun owned by Atlantic Geoscience Centre.

transverse channels 1,200 to 1,800 feet deep. Although there has been no deep drilling on the shelf north of Newfoundland it is considered to be underlain by Tertiary and Mesozoic rocks. The banks contain shallow channels, ridges, and knolls: features suggestive of submerged glacial topography. Figure 2 illustrates a veneer of ground moraine on soft bedrock. A recessional moraine may be present on the profile.

It may not be too far from the truth to liken the whole area to a submerged prairie or steppe, or perhaps a tundra plain of one of the interglacial stages. Trask (1932) described frosted sand grains from sediments from Newfoundland to Davis Strait. He offered no firm conclusion as to their origin but Johnson (1967) stressed that the abundance and prevalence of frosted sand grains in the Ungava Bay area was due to the fact that Pleistocene aeolian sand dunes were submerged and are now being reworked. The explanation is within the range which Trask (1932) allowed.

Marginal Channel

The boundary between the submerged coastal plain and the fiord indented rugged mainland is occupied or marked by a marginal channel. O. Holtedahl (1929, 1950, 1970) and H. Holtedahl (1958) have described this feature in many places at high latitudes. They have suggested that the channels mark faults which separate the mainland from Tertiary and Mesozoic rocks. Grant (1966, 1970) offers an alternative explanation as to the origin of the channels by contending that they have been formed by continental glaciers moving eastward from the mainland.

It can readily be understood that an underloaded, eastward moving, continental glacier passing over mainly igneous rock, would become saturated and loaded with the soft shales and sands of the coastal plain. As the marginal channel was deepened and widened the ice probably flowed northward and southward. Transverse channels served as outlets to the continental slope. The load of debris, acquired while making the marginal channel, was deposited as ground moraine as illustrated on Figure 2.

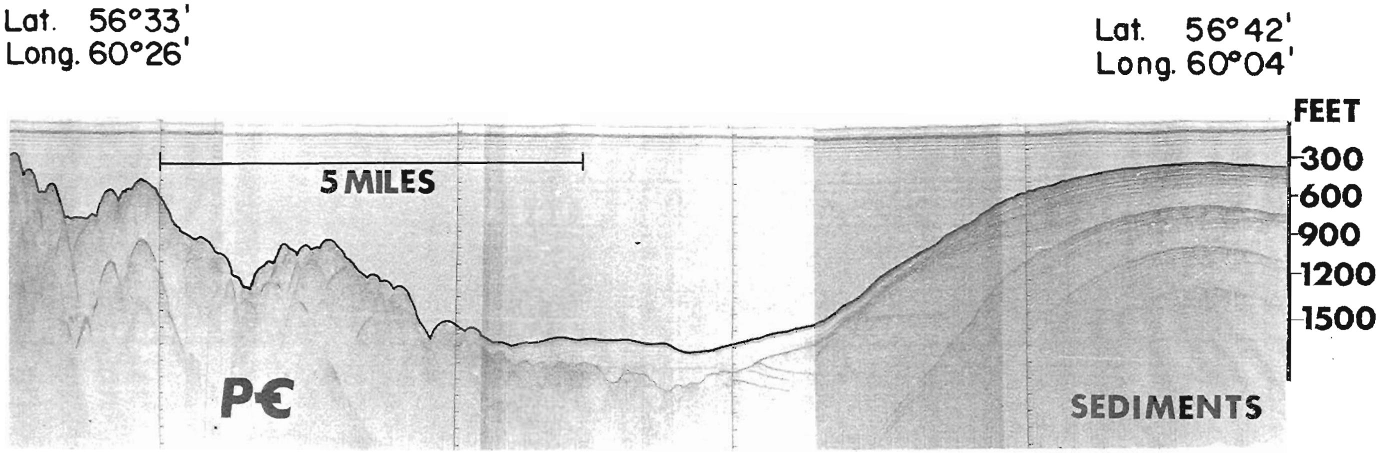
Figure 3 illustrates the main features of the marginal channel in eastern Canada. The channel is well developed along most of the coast from just north of Newfoundland to Hudson Strait. North of Hudson Strait it is not everywhere readily identifiable as a submarine feature. Along Baffin Island and northward to Ellesmere Island channels are represented by areas of deep water separating offshore banks from the mainland. The channels are not developed at all to the same degree as off Labrador (see maps in Pelletier, 1966).

Relationship Between Marginal Channels and Hinge Lines

From the foregoing it appears that for areas in eastern Canada north of Newfoundland, the Laurentian Peneplain was elevated in places to heights of 5000 feet during the Tertiary Period. The movement upward probablv took place along an old, fundamentally weak zone in the crust. This weak zone undoubtedly is adjacent to, and possibly coincident in places with the initial break which initiated rifting and later sea-floor spreading which produced the Labrador Sea and Baffin Bay. The rifting took place mainly within igneous or metamorphic rocks. After the process began, space was made, or basins were formed for the deposition of sediments. A hinge line occurs along the western edge of the new series of basins.

An important aspect of the coastline from Newfoundland to Ellesmere Island is that nowhere south of Cape Dyer have young, soft rocks been found on the mainland. Near Nain on the central Labrador coast and near the northern tip, Lower Paleozoic carbonates outcrop or occur as boulders. At Cape Dyer, and on Bylot Island some Tertiary and Mesozoic sediments are present sparsely. The reason tentatively offered for the general absence of young sedimentary rocks is that Mesozoic and later sedimentation was restricted mainly eastward of the proposed hinge line.

The marginal channel follows the boundary between old igneous rocks and young soft sediments, and this boundary is a hinge line. The westward-moving continental glacier eroded the marginal channels as Grant (1970) has described. It seems clear that marginal channels would not have been possible if there was not a plate of soft sediments abruptly present to the east. Further, there would be no plate of soft sediments if there had been no rifting and subsequent sea-floor spreading. Stating the case in the opposite sense, it must



LABRADOR
AIR GUN PROFILE

Figure 3. Rugged Precambrian surface on west; smooth Mesozoic or Tertiary sediments on east. Profile recorded with an air gun owned by Atlantic Geoscience Centre.

follow that if shield rocks extended appreciable distances offshore a marginal channel could not be made by any process related to glaciation.

To summarize, the evidence indicates that the Holtedahls (1929, 1950, 1958, 1970) are essentially correct in assigning a tectonic origin to the marginal channels. The channels are a good example of a geomorphic feature directly related to tectonics. Grant (1970) however, has clearly shown that, at least off Labrador, present-day faults or faults of recent origin (a possibility stressed by the Holtedahls) are uncommon.

In support of the summary an adaptation of data published by Mayhew et al. (1970) indicates a thick wedge of non-magnetic, soft sediments in abrupt contact with magnetic, hard rocks of the Precambrian at about the centre of the marginal channel along the middle part of the Labrador coast (Fig. 4) (see also Hood et al., 1967, p. 3). The hard, magnetic, basement in this locality has an eastward slope of 10 degrees. A hinge line must surely be present here.

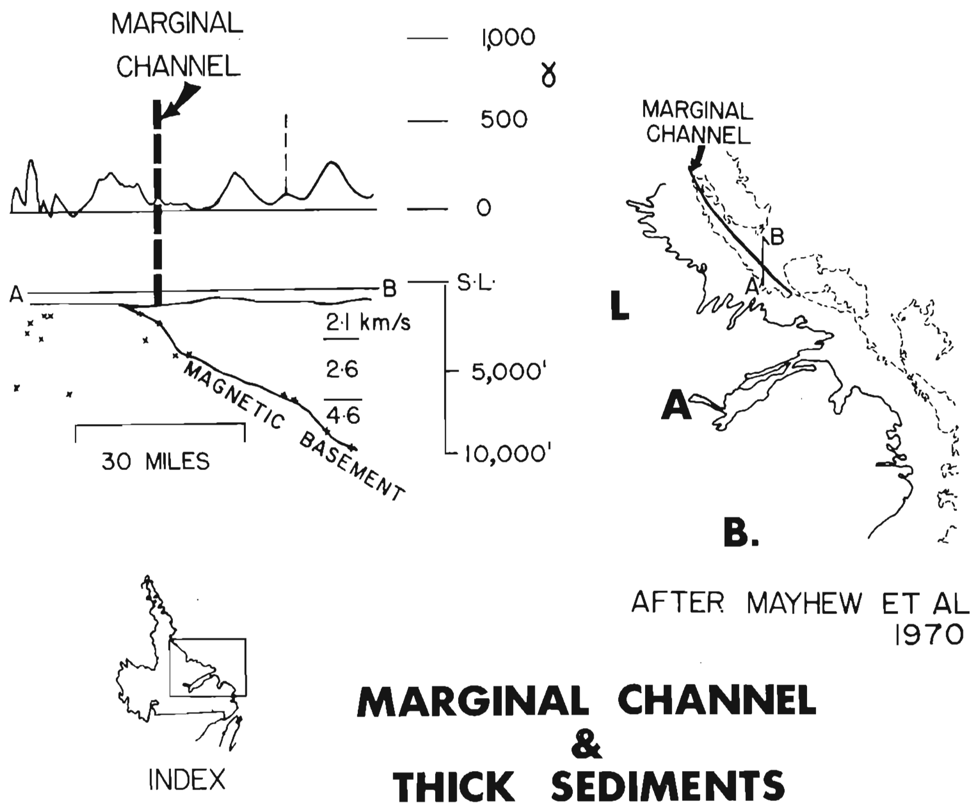


Figure 4. Relationship of non-magnetic low-velocity sediments and magnetic Precambrian basement. True dip of "basement" on average is 10 degrees but near the marginal channel it is steeper.

SEDIMENTARY VENEER OF THE CONTINENTAL SHELF

Records of the various types of unconsolidated sediment found on the shelves of this part of the northwest Atlantic have been reported since 1900. Figure 5 is a summary of some of the results. Publications used for the compilation on the Canadian portion are Trask (1932), Litvin and Rvachev (1963), Grant (1965), and Kranck (1966). The reports used for the Greenland side of the Labrador Sea include Boeggild (1900), Dibner et al. (1963), and Rvachev (1964). The sampling method was by grab, dredge, or piston core. Almost all samples contain pebbles and cobbles ranging from wheat-grain size to tens of centimetres across. Granite and carbonate are the most common. Almost all authors stressed the presence or absence of carbonate pebbles.

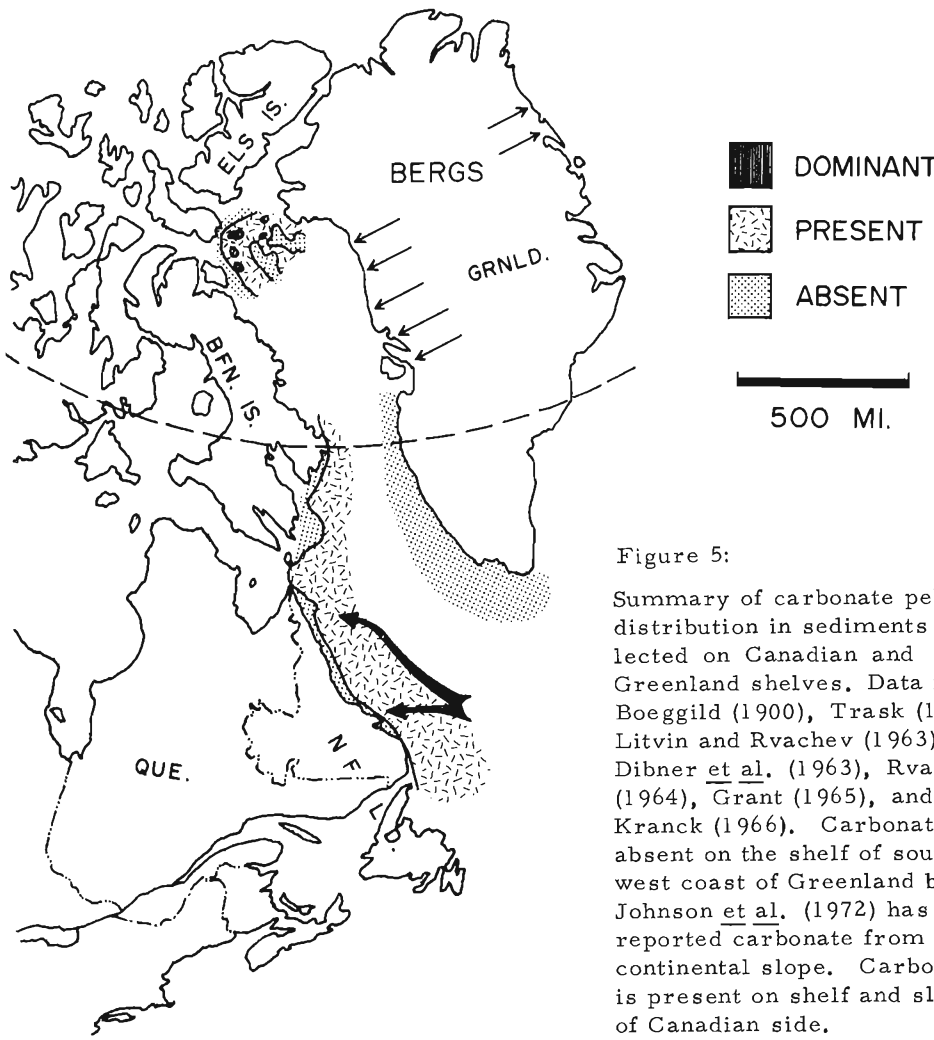


Figure 5:

Summary of carbonate pebbles distribution in sediments collected on Canadian and Greenland shelves. Data from Boeggild (1900), Trask (1932), Litvin and Rvachev (1963), Dibner et al. (1963), Rvachev (1964), Grant (1965), and Kranck (1966). Carbonate is absent on the shelf of southwest coast of Greenland but Johnson et al. (1972) has reported carbonate from the continental slope. Carbonate is present on shelf and slope of Canadian side.

These are variously called shaley limestone, marl, vuggy limestone, dolomite, and crystalline limestone. A problem common to several of the authors is to explain the widespread occurrence of so much carbonate or limestone in the bottom debris of the shelf. Almost all authors state that the iceberg-distribution process is perhaps one of the likely causes and some point out that perhaps floe ice is one of the agents of transporation for this sedimentary component of the bottom debris. The obvious difficulty that is stressed in one way or another is that icebergs are calved in an area that is mainly underlain by Precambrian rocks (Fig. 5). Furthermore, westward-lying areas are an unlikely provenance, except in the Arctic Islands which are a considerable distance away. Therefore, it is tempting to rely on the icebergs as the main agent for the distribution of the carbonate component. There are valid objetions to this explanation in addition to the apparent lack of carbonate at the iceberg source. A second problem is that limestone pebbles are uncommon on the shorelines of Labrador Sea and Baffin Bay between the present water's edge and the raised beaches. Presumably 10,000 years ago, icebergs were floating over present-day land areas and if their enclosed rock material included appreciable amounts of carbonate, substantial traces of this component have not been found. A third problem, which is similar to the second, is that near shorelines frequented by icebergs, carbonate as a relevant component of the bottom material does not appear to be present (Kranck, 1966; Trask, 1932). Fourthly, if it is assumed that the carbonate does arrive in the western part of Baffin Bay and on the Labrador shelf by iceberg transportation then the surface of the southwest shelf of Greenland should also be littered with the same material. Icebergs from northern Greenland calve on the east coast of the island and find their way to the west coast (Fig. 5). Carbonate pebbles are absent on the surface of the shelf of southwest Greenland. *

Relationship of Bedrock and Composition of Unconsolidated Veneer

No results of deep coring on the continental shelves (if any has been done other than piston coring) have been published. Therefore the relationship between bedrock and the veneer of loose sediment above cannot be effectively ascertained. Figures 6 and 7 are copies of aeromagnetic profiles published by Hood et al. (1967) with sampling locations reported by Trask (1932) superimposed on the maps. On the western end of the aeromagnetic profiles the sharpness of the anomalies indicates that the magnetic source is close to the sea bottom. On the eastern part of the lines where the profiles are smooth, it is apparent that the magnetic source is at an appreciable depth and is overlain by non-magnetic rocks. Where non-magnetic rocks occur at sea-bottom Trask (1932) reports a carbonate component in the samples collected. Where magnetic rocks outcrop at sea-bottom Trask reports that carbonates, or any other sedimentary rock, are absent.

Figure 8 illustrates that near the shoreline of Baffin Island (Exeter Bay) carbonates are absent but eastward of the two islands pebbles and fragments of carbonate are common in the bottom debris. The rocks outcropping on the islands and the mainland are of Precambrian age, so presumably the rocks comprising the sea bottom between are also the same age. The areal

* Recently Johnson et al. (Nature, vol. 236, p. 86-87) report the widespread distribution of Mesozoic carbonates on the continental slope of the southwest shelf of Greenland. Most of the samples collected were in situ.

situation seems to be that where carbonate is absent in the bedrock (between islands and mainland), it is also absent in the overlying debris.

The examples given indicate that there may be a positive correlation between the composition of bottom debris and underlying bedrock.

Origin of Bottom Debris

The main objections to the theory that floe ice and icebergs are the principle transporting agents of bottom debris have already been cited. On the other hand there is supporting evidence for the argument that on the shelves there is a positive correlation between bottom outcrop and debris composition, and not a randomness of distribution of material.

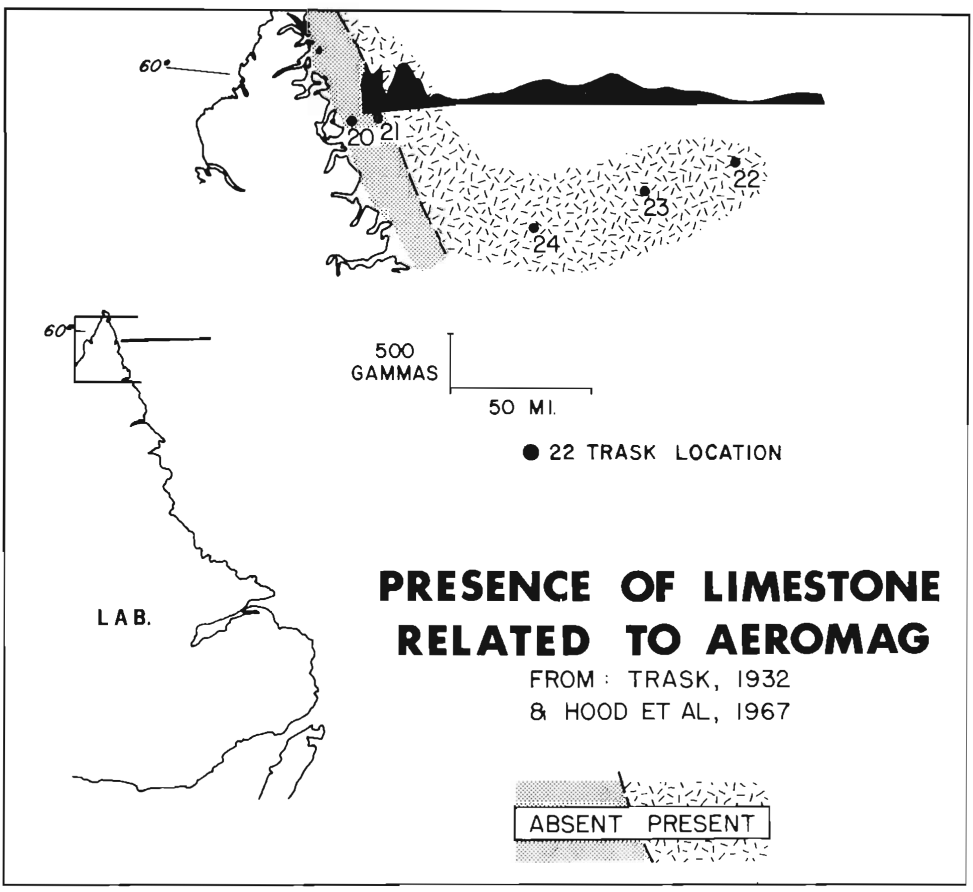


Figure 6. Carbonate pebbles in sedimentary veneer absent where aeromagnetic profile indicates a relative absence of non-magnetic rocks off the northern tip of Labrador. On eastern end of profile where non-magnetic rocks at sea-bottom are indicated carbonate pebbles are present.

Grant (1965) described the bottom sediments of the northern part of Baffin Bay and from the tabulated data it is possible to show that the distribution of carbonate pebbles is not random (Fig. 9). Carbonate pebbles are a dominant component in some areas near the Canadian side and are mainly absent on the Greenland side. Additionally, most sediment-ladened icebergs calve from the continental glaciers of northern Greenland in this vicinity. It seems unlikely that icebergs transported all of the debris now found on the bottom of the bay.

International Ice Patrol reports show that about 2,500 icebergs per year cross the 60th parallel on the drift southward. If we assume that the average size of an iceberg at that latitude is 140,000 tons and that 2 per cent of the berg by volume is morainic material, and that all of the bergs melted over the Labrador shelf, it is easy to calculate that only four inches of sediment would be deposited in 10,000 years in an area 100 by 800 miles, it being

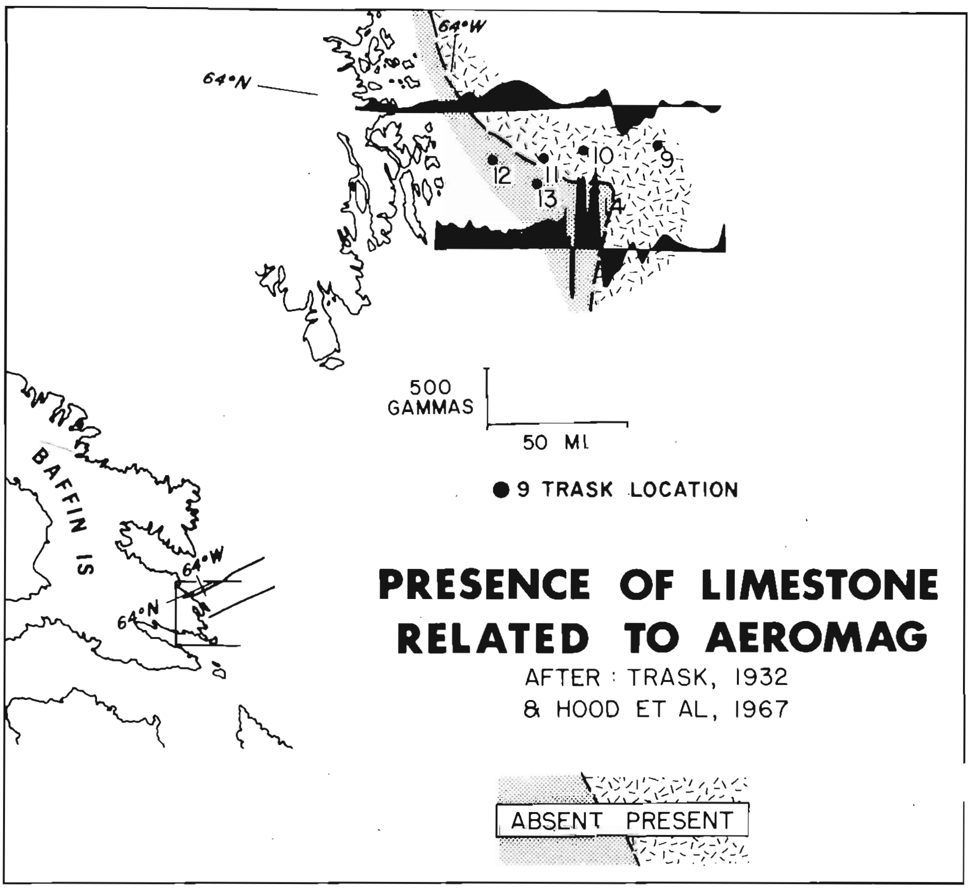


Figure 7. Carbonate pebbles in sedimentary veneer absent where aeromagnetic profile indicates a relative absence of non-magnetic rocks off southern Baffin Island. On eastern ends of profiles where non-magnetic rocks at sea-bottom are indicated carbonate pebbles are present.

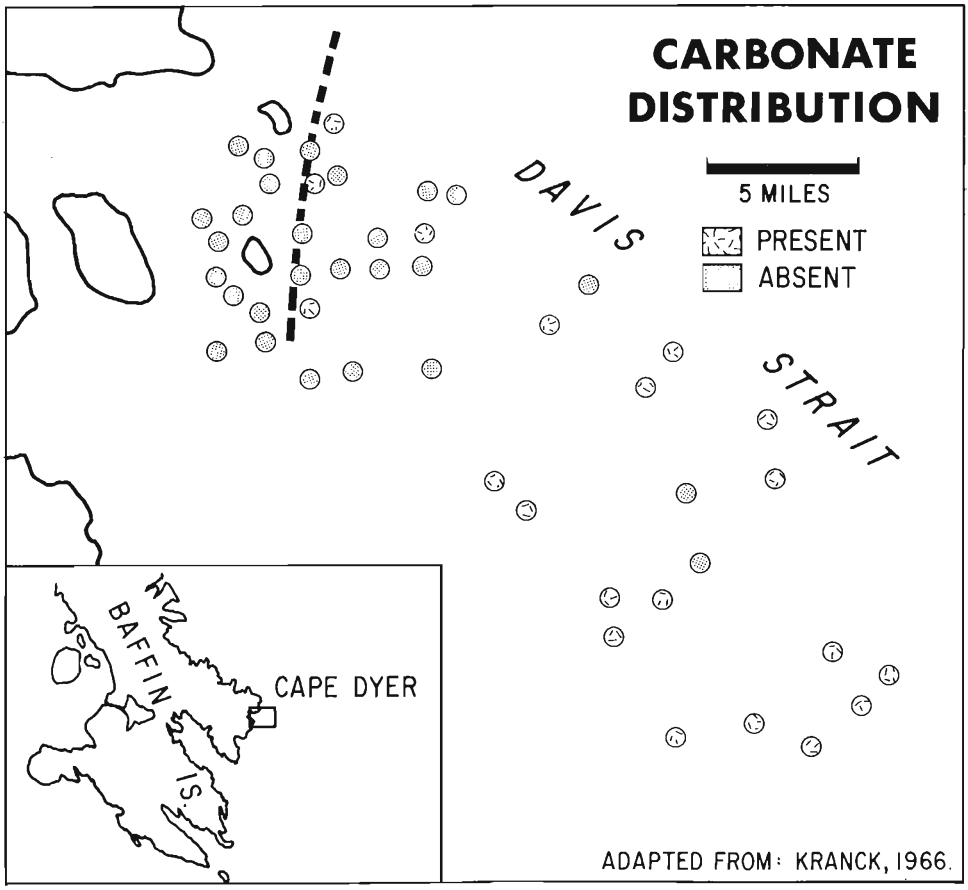


Figure 8. Boundary line between Precambrian Shield and the sedimentary wedge to the east is adjacent to and directly eastward of the two small islands.

assumed that no clay and silt drifts to the open Atlantic. Thus whereas icebergs have deposited vast amounts on the shelves, it is probable that the main part of the veneer is ground moraine of the continental glaciers.

Ground Moraine

Ground moraine is probably the main surficial deposit on the shelves of Baffin Island and Labrador. It has been reworked by bottom currents and added to by iceberg transport. Ground moraine composition, in bulk, is a reliable average of the composition of the underlying bedrock. Continental glaciers move material only 5 or 10 miles. Accordingly, the composition of the ground moraine can be examined, if bedrock itself is not available, and a reasonably accurate regional picture of the geology established. Veltheim (1962) was able to make a geological map of the Gulf of Bothnia using this technique.

Carbonate pebbles in bottom debris are absent on the surface of the southwest Greenland shelf, probably because no carbonate component is present in the submerged uppermost outcrops. On the Canadian side, the absence or presence of carbonate pebbles in the surficial deposits depends on the composition of the underlying bedrock. Carbonate, in most of its forms, is the most of its forms, is the most durable sedimentary rock in the bottom of a glacier so it is the one most commonly reported. All non-soluble rocks survive with their contained fossil remnants however.

Bottom Sampling and Seismic Profiling on the Labrador Shelf

In July and August 1967, A. C. Grant and V. F. Coady of the Atlantic Geoscience Centre, Bedford Institute of Oceanography, Dartmouth; A. Reinsbakken and the author, both of Tenneco, conducted a fathometer, air gun, and bottom sampling survey on the Labrador Shelf. The cruise was conducted aboard the M. V. FREDERICK L. BLAIR.

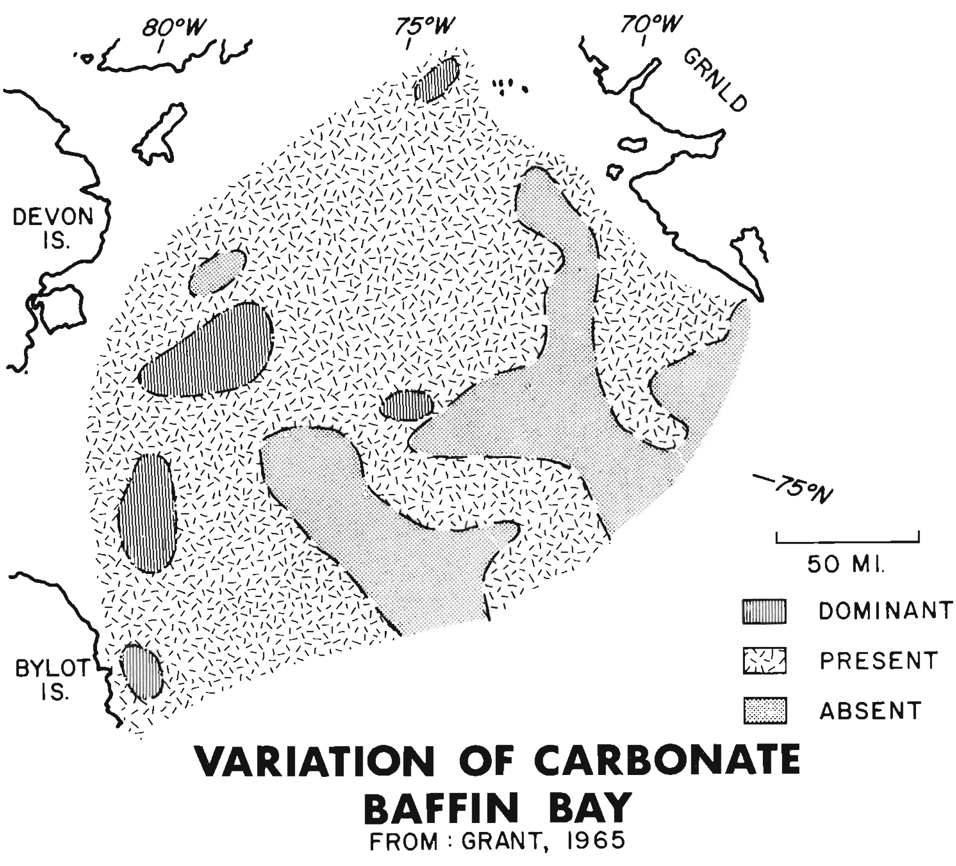


Figure 9. Carbonate pebbles in bottom sediments are dominant in places in the western part of Baffin Bay. Near source of icebergs, and near Greenland carbonate pebbles are substantially absent.

AGES OF PISTON CORE SAMPLES BASED ON PALYNOMORPHS

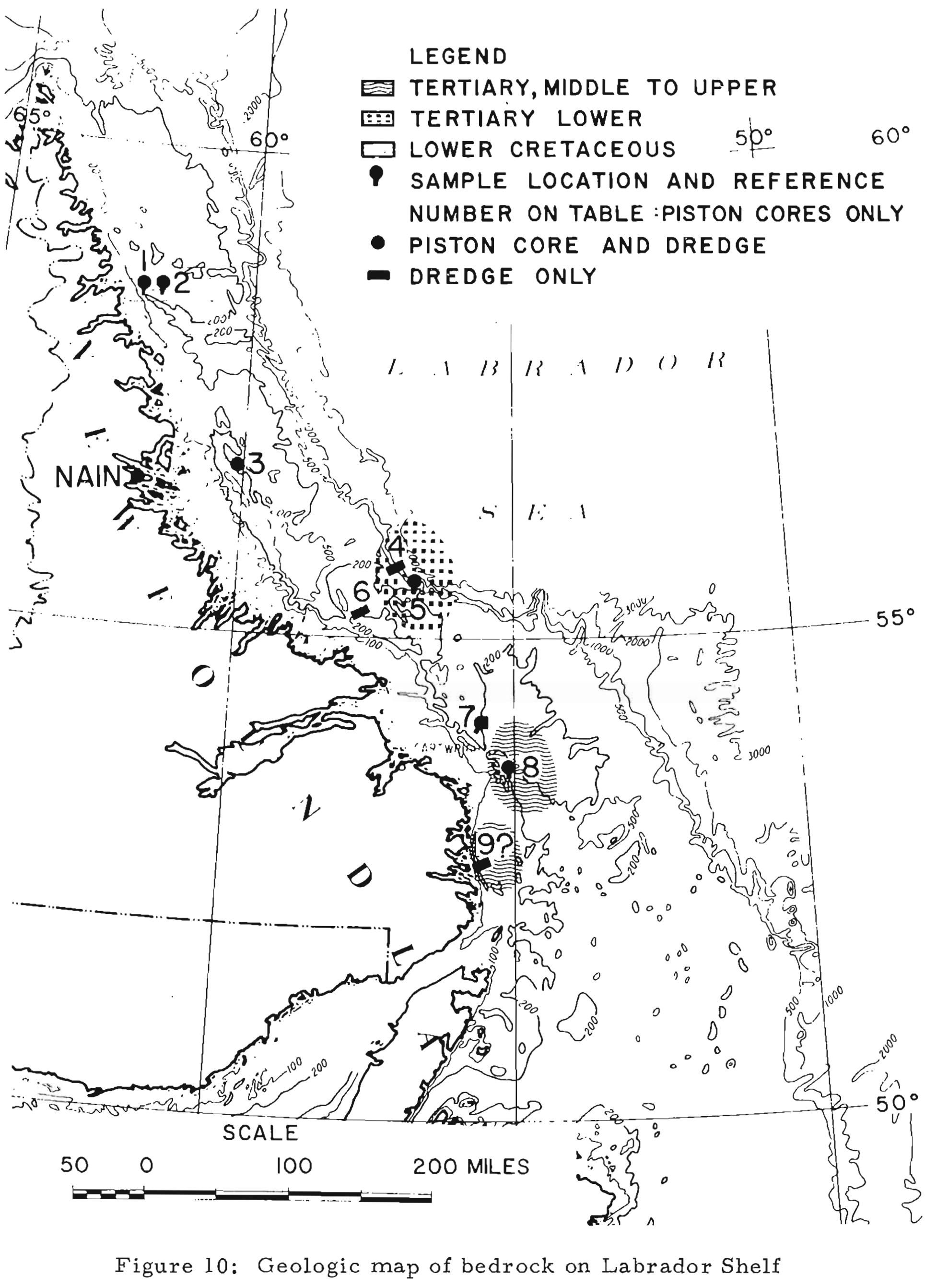


Figure 10: Geologic map of bedrock on Labrador Shelf

The main object of the fathometer and air gun survey was to locate steep topographic slopes which may have submarine outcrops. Several were located and they were sampled with a pipe dredge, piston coring device, or both. At the time it was hoped that some bedrock samples or at least submarine "talus" would be obtained, or failing that it was hoped that the ground moraine samples would be useful. However it is unlikely that bedrock was sampled.

Sedimentary pebbles, mainly vuggy limestone, chalk, and dense limestone, were obtained from the dredge hauls. The few pebbles in the piston cores were analysed for palynomorphs as well as clayey matrix. The ages determined for the cores are remarkably consistent whereas those determined from dredge samples are mixed. The ages reported in the accompanying Table were determined by C. R. Pickett, a palaeontologist of Tenneco Oil Company in Houston, Texas, using palynomorphs from sedimentary pebbles. Individual dredge samples weighed between 100 and 400 pounds. Core recovery ranged between 6 inches and 6 feet.

The accompanying Table is a summary of the results of the sampling. Figure 10 depicts the areas on the shelf that were sampled as well as the general results of the ages determined on the cores. It is possible from the results, to tentatively conclude that the bedrock at Locations 1 and 2 is Lower Cretaceous. At Location 3 probably the bedrock is also Lower Cretaceous but the possibility of overlying Upper Cretaceous and/or Tertiary material cannot be discounted, judging from the results of dredge-sample analyses. On the upper part of the continental slope off the central part of Labrador (Location 4 and 5) the bedrock may be Lower to Middle Tertiary, but on the inner part of the shelf at Location 6 Cretaceous rocks may outcrop. Location 7 yielded no results. At Locations 8 and 9 the age of submarine outcrops may be Middle to Upper Tertiary.

In the south the submarine outcrops are generally of Middle to Upper Tertiary age. In the central part of the area the age of the rocks may be Upper Cretaceous and Tertiary on the west but eastward, to the depths sampled (1950 feet), the rocks are Lower to Middle Tertiary. In the north it is likely that only Lower Cretaceous rocks occur at the surface of the inner part of the shelf. Lithologically the Tertiary to Lower Cretaceous rocks are sands and shales with interbedded discontinuous lenses of carbonate of various sorts.

On the continental slope of the central part of Labrador, Upper Jurassic rocks have been identified. This discovery is not included on Figure 10 because the determination is from a single location. Bartlett (Queens University, Personal Communication) reports Jurassic rocks from bottom samples collected farther north on the same continental slope. Henderson (in press) reports that a Lower Jurassic sample was collected from the sea-bottom south of Disko Island in 1859.

The Paleozoic is represented by Devonian occurrences at Locations 3, 5, and 6. These discoveries are, like those of the Jurassic, not readily explainable.

It is very probable that Jurassic rocks were deposited in the Labrador Sea and the dredge sample represents them. Devonian rocks which were collected may be representatives of true glacial erratics; however their source has not been found in the ice-shed area. Ordovician and perhaps Silurian rocks are known however from near Nain and near the northern tip of Labrador (Bell, 1884; Roy, 1932; Little, 1936). Other erratics include most kinds of igneous rocks that outcrop on mainland Labrador.

SUMMARY OF DREDGING AND PISTON CORING RESULTS, LABRADOR SHELF (1967)

(See Map on Figure 10)

LOCATION		CORES		DREDGE HAULS		
NO.	CO-ORDINATES	DEPTH FT.	AGE	DEPTH FT.	AGE OF PEBBLES	GEOMORPHIC POSITION
1	58°35'N 61°58'W	(a) 492 (b) 498 (c) 522 (d) 522 (e) 576	(a) None possible (b) None possible (c) L. Cret. (d) L. Cret. (e) Cret.	None		On western upper lip of shelf
2	58°37'N 61°45'W	(a) 510 (b) 510	(a) L. Cret (b) L. Cret	None		On shelf
3	56°40'N 60°09'W	(a) 552 (b) 558 (c) 654 (d) 768 (e) 960	(a) No Diagnostic Palynomorphs (b) L. Cret. (c) L. Cret. (d) L. Cret. (e) No age	(a) 630 (b) 960	a) 2 slides Dev. 1 sli. L. Cret.; 2 sli. U. Cret.-L. Tert. 10 sli. M. Tert. b) 1 slide Eocene; 2 slides M-U Tert.	Cores and dredges on westward facing scarp of marginal channel
4	55°43'N 57°06'W	None		(a) 1020 (b) 1080 (c) 1620	a) M. Tert. b) U. Tert. c) 2 slides U. Jurassic 10 slides M. Tert.	On upper part of Continental slope.
5	55°40'N 57°00'W	(a) 1566 (b) 1650(1 (c) 1650(2 (d) 1950	(a) Lower Mid. Tert. plus Cret. and Dev. (b) (1) Barren (c) (2) Lower Mid. Tert. (d) Pre-Miocene?	None		On upper part of Continental slope.
6	55°13'N 57°48'W	None		Single haul between 480 & 720	55 samples proc. 15 U Cret-Tert. 5 Mid. Tert. 1 L. Dev.; 34 not useful	Haul on westward facing scarp of marginal channel
7	54°08'N 55°30'W	(a) 588 (b) 702 (c) 738	All contained Palynomorphs no age possible	None		On upper part of west facing scarp of marginal channel
8	53°43'N 55°04'W	(a) 576 (b) 780 (c) 984	(a) U. Tert. to Recent (b) Mid. Tert. (c) Mid. Tert.	None		On west facing scarp of marginal channel
9	52°44'N 55°38'W	None		Single haul at avg. 600 feet deep	33 samples proc. 24 barren; 4 age unknown; 3 Pleistocene; 2 Mid. Tert.	On west facing scarp of marginal channel

The mixing of ages can tentatively be explained by the likelihood that ground moraine, in contrast to bedrock or talus, was sampled in many places. Not only was there a constant flow of glacier ice eastward to cause mixing but parts of the ice sheet moved laterally – northward and southward – along the marginal channel, further complicating the pattern.

ACKNOWLEDGMENTS

The presentation is offered through the offices of Tenneco Oil & Minerals Ltd., Total Petroleum (North America) Ltd., Amerada - Hess Petroleum Corporation, and AGIP Canada Ltd. This group of companies is actively pursuing a program of exploration on the offshore of eastern Canada.

The idea of a joint program (Bedford – Tenneco) was mainly that of Dr. B.R. Pelletier, the Chief Marine Geologist, and Dr. A.C. Grant both of the Atlantic Geoscience Centre, Bedford Institute of Oceanography, Dartmouth, Nova Scotia. Dr. Grant and V.F. Coady, with mainly Bedford equipment, participated on the cruise of the M.V. FREDERICK L. BLAIR. I wish to thank them very much.

In addition I want to thank the Drafting Departments of Tenneco Oil & Minerals, Ltd. and Aquitaine Company of Canada Ltd. of Calgary for making the figures. Also Gilroy Henderson, Geological Survey of Greenland; the staff of the Geological Survey of Canada, Ottawa and Calgary; and C.R. Pickett, Tenneco Oil Company, Houston. I benefitted from criticisms of Dr. R.U. Best, University of British Columbia. Dr. J.W. Murray, also of University of British Columbia provided me with a pipe-dredge design.

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29. SEDIMENTATION IN AN ARCTIC MARINE ENVIRONMENT: BAFFIN BAY BETWEEN GREENLAND AND THE CANADIAN ARCTIC ARCHIPELAGO

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Abstract

Analysis of cores and grab samples from Baffin Bay indicates that sediment texture reflects the local bottom topography and the random distribution of sediments by ice-rafting. The distribution of ice-rafted sand and gravel can be related to the surface currents in Baffin Bay. The shelf sediments are composed by and large of sands and gravels intermixed with mud. Large depressions on the shelf contain fine-grained sediments similar to those that floor the deep bathyal basin in the central part of Baffin Bay. Fine-grained sediments extend from the mouths of the fiords north of Disko Island across the continental shelf to the bottom of the basin. These fine-grained sediments are inferred to have been deposited as a hyperpycnal deltaic flow that issued from the margin of the glaciers.

The sediment currently being deposited in many areas of the bay is not representative of the underlying sediments. Many cores exhibit alternating intervals of coarse- and fine-grained sediments indicating varying processes of sedimentation in the past.

Sediment colour shows a marked shift in oxidizing conditions from the northern parts of the bay southward. On the shelf the colour of the bottom sediments suggests reducing conditions but the colour of subbottom sediments suggests oxidizing conditions. During times of maximum glaciation subaerial exposure of the shelf led to the formation of colouration suggestive of oxidation; however, as the sea later transgressed, reducing conditions followed. In the deeper areas of the bay colours indicative of oxidizing conditions may be related to bacterial consumption of organic material primarily in the upper 200 metres of the water column. Bottom currents are discounted as important oxidation agents because the presence of significant percentages of fine-grained material at almost every sampling locality indicates very slow current velocities. Sandy layers which occur in the deeper areas of the bay probably were formed by localized downslope movements of sediment rather than by the winnowing action of bottom currents.

INTRODUCTION

Purpose of Study

The purpose of the present study was to map the distribution of the bottom and subbottom sediments of Baffin Bay and to use this information for interpreting the depositional processes that are at work in Baffin Bay, especially the extent to which ice plays a part in the transport of the sediments.

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A total of 47 cores, taken with an 11-foot "modified Ewing" gravity corer equipped with 2.5-inch (t.35-cm) inside diameter polycarbonate plastic-liners, and a total of 43 grab samples, taken with an "orange peel"-type sampler, were collected for this study. The cores were taken on a grid pattern to cover most of Baffin Bay (Fig. 1). This study has provided the densest core control yet available from Baffin Bay. Preliminary results on this study have been published (Blee, Baker and Friedman, 1968); radiographs of sedimentary structures in Baffin Bay cores are illustrated elsewhere (Baker and Friedman, 1969), and data on interstitial waters are part of a separate study (Friedman, Ali and Amiel, in press).

Physical Setting of Baffin Bay

Bathymetry of Baffin Bay

Baffin Bay is an enclosed, elongated basin, lying between Greenland on the northeast and Baffin Island of the Canadian Arctic Archipelago on the

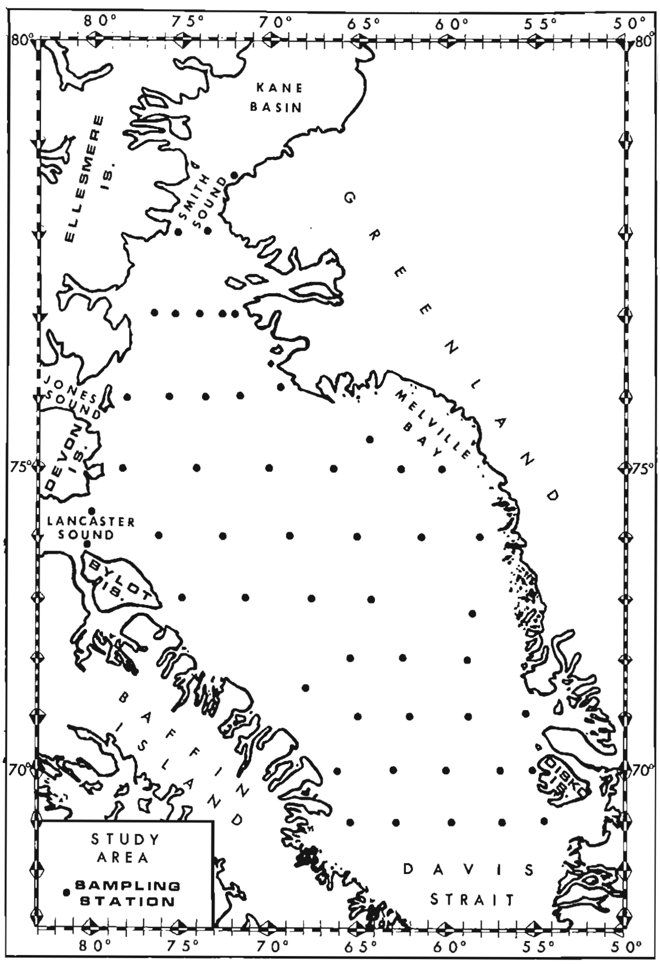


Figure 1.

Index map of Baffin Bay showing location of sampling stations.

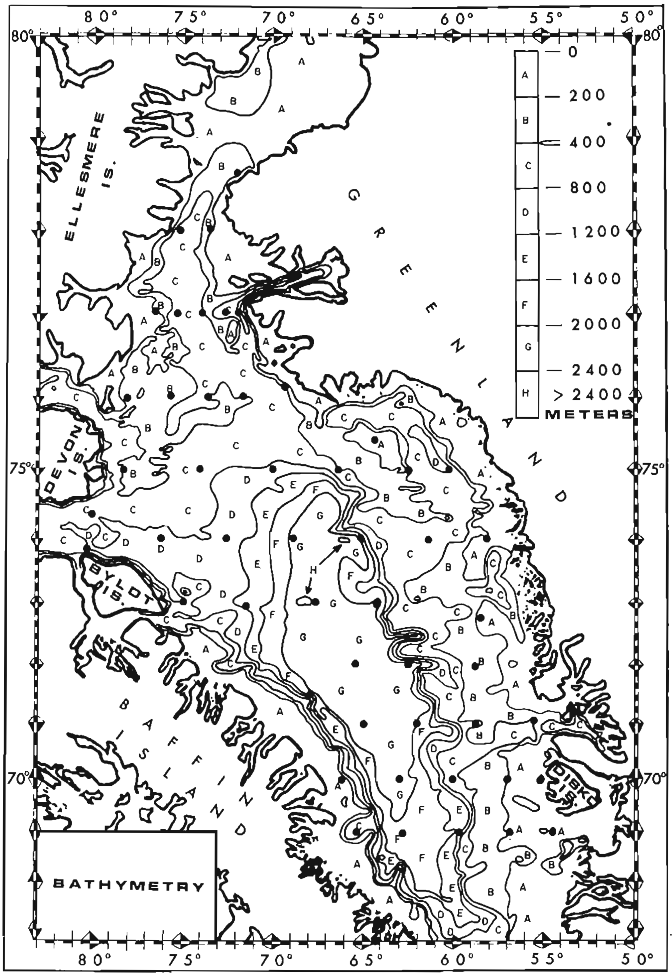


Figure 2.

Bathymetric map of Baffin Bay study area. The depth intervals below 400 metres are in 400-metre increments.

southwest (Fig. 1). The maximum depth of Baffin Bay is approximately 2,400 metres, which is roughly 550 metres below the 1,000-fathom limit for the bathyal-abyssal boundary. However, as most of Baffin Bay is of true bathyal depth we refer to the floor of the basin as a bathyal plain to avoid confusion in nomenclature. The regional extent of the bay is shown by its dimensions, about 800 miles (1,300 km) long by 280 miles (450 km) wide. The long axis of the bay runs northwest to southeast with slopes between the shelf areas and the bathyal plain in this direction ranging from 1:250 to 1:90 (Marlowe, 1968). Steeper gradients, as high as 1:4, occur locally. Figure 2 shows the bathymetry of Baffin Bay. Submarine canyons incise steep slopes in places. These canyons appear to have resulted from erosion by streams and glaciers during the lower stands of sea level.

The term "shelf" used in this study refers to the submerged area bounded by the shore and the first significant break in slope seaward from shore. The eastern shelf of Baffin Bay is approximately 140 miles (225 km) wide with a change in slope occurring at water depths ranging between 400 and 600 metres; the western shelf is much narrower and the break between shelf and slope occurs at the 200-metre water depth (Fig. 2). At the north end of Baffin Bay between Ellesmere Island and Greenland, no discernible slope-change from shelf to bathyal depth occurs. Pelletier (1966) described the relatively flat shelf topography extending from the northern part of the bay, through Smith Sound, and into Kane Basin (Fig. 1) as resembling a submerged headland weathered during sub-aerial exposure.

Baffin Bay is connected to the Atlantic Ocean by Davis Strait where the bathymetry reaches a limiting threshold "sill" depth of about 700 metres. Connections of the bay to the Arctic Ocean are through Smith Sound to the north and Lancaster and Jones Sound to the west, the first two being the main channels with effective sill depths of 200 metres and 175 metres respectively (Hachey et al., 1956).

Water Circulation in Baffin Bay

In general, the surface water circulation is anti-clockwise (Fig. 3), with warm Atlantic Ocean water entering Baffin Bay through the eastern part

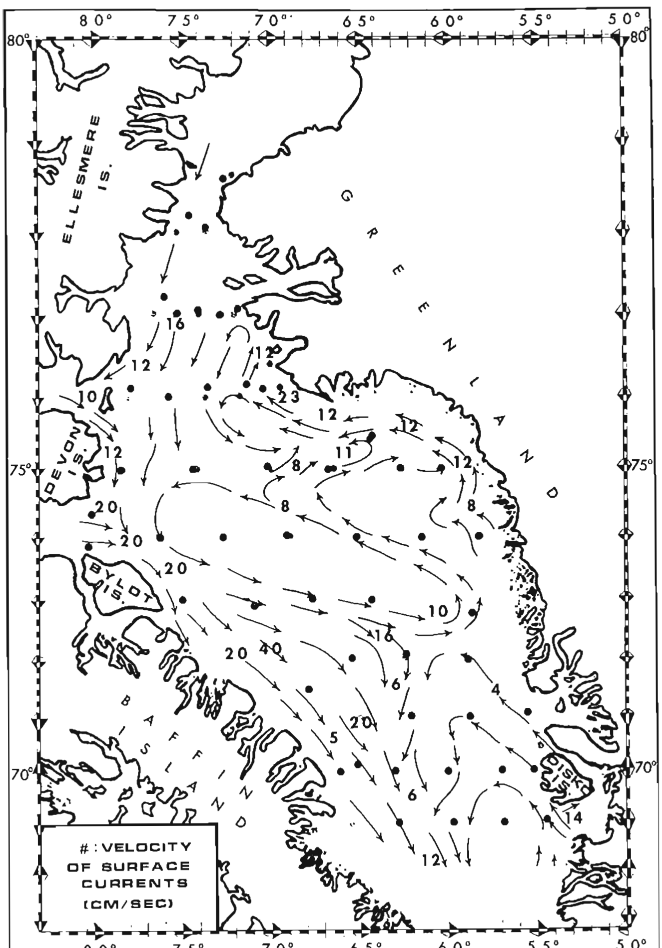


Figure 3.

Surface currents of Baffin Bay during August-September; velocity of surface currents in cm/sec (after Dunbar, 1951).

of Davis Strait and travelling along the west Greenland coast while cold Arctic Ocean water enters the bay through Smith, Jones, and Lancaster Sounds (see Fig. 1), travels down the eastern coast of Baffin Island and passes through the western part of Davis Strait (Dunbar, 1951). In a large elliptical area, in the central part of the bay, exchange of surface water between the two currents takes place (Fig. 3). Iceberg movement within the bay follows the pattern of surface currents (Smith, 1941; Dunbar, 1954).

Below 700 metres, water leaving Baffin Bay through Davis Strait is restricted by the limiting sill and is reflected back at depth along the west Greenland coast. At depths greater than 700 metres, currents are very weak.

Ice-rafting in Baffin Bay

Various studies have clearly demonstrated the importance of ice as an agent of sediment transport throughout the entire eastern Arctic region. The mechanism of loading ice with sediment in fiords by the interplay of glacial streams, wind, and the ice during the calving process and the effect of grounded ice in stirring up fine sediment and abrading the shoreline was first described by Tarr (1897). Sverdrup (1938) discussed the upward movement of sediment through sea-ice by the action of bottom freezing and surface thawing, with the sediment being washed away by waves as it becomes exposed on the upper surface of the ice. The relationship between ice and sediment in the Arctic Basin was recorded by Emery (1949). The incorporation and transport of fine sediment into sea-ice by suspension-freezing was discussed by Campbell and Collin (1958). Further evidence of the sediment-ice relationship is provided in studies by Forgeron (1959), Perry (1961), Leslie (1963), Kranck (1964), Grant (1965), Marlowe (1965, 1966, and 1968), Codispoti and Kravitz (1968), and Kravitz and Sorensen (1970).

Ice carries sediment ranging in size from clay to boulders. The site of final deposition of this sediment depends on such widely diverse factors as where the ice was formed, the velocity and pattern of the surface water currents, the temperature of the water, the thickness of the ice, and even the possible collision between two icebergs. Therefore, the size frequency distribution of sediments beneath the paths of floating icebergs or sea-ice is most likely to be polymodal, with a random distribution of various-sized ice-rafted sedimentary particles being superimposed on a continuous distribution of water-deposited sediment.

Previous Studies of Bottom Sediments

The earliest expeditions to Baffin Bay produced qualitative reports on the bottom sediments. Boeggild (1900), analyzing sediment taken from Davis Strait and Labrador Sea during the 1895 and 1896 Danish INGOLF expedition, suggested that the "gray deep sea clay" found in the sampling area also extended north of 63°N latitude, covering the entire sea-bottom of Baffin Bay. Knudsen (1899) described two samples taken from the deep water of Disko Bay as "clay" and "gray clay".

A few samples recovered during the second Norwegian expedition in the FRAM from 15 fathoms (27 m) of water in Goose Fiord off Jones Sound were described by Kiaer (1909) as "soft brown clay" with pebbles, molluscan shells, and plant material.

Riis-Carstensen (1931) reported on the bottom sediments collected during the 1928 expedition of the GODTHAAB. Grey clay was more commonly found than brown or yellow clay. At stations closer to shore, gravel, sand, and shells occurred in greater amounts.

The first quantitative analysis of sediments was made by Trask (1932), using material collected by the 1928 MARION expedition in Davis Strait and into Baffin Bay as far as 70°N latitude. The size distribution and mineralogy of the bottom sediments were statistically analyzed and the samples were classified by median grain-size and "coefficient of sorting". A sediment range from clay to fine sand occurred at most localities. Ice-rafted pebbles were common in most of the samples.

A general decrease of coarse sediment outward from the shore was reported by Vibe (1939) in a study of the shallow water deposits (10 to 64 m) of northwest Greenland.

Perry (1961) described a similar relationship of finer sediment texture to increased distance offshore in the eastern Canadian Archipelago. The material contained particle sizes from clay to cobbles and was predominantly poorly sorted and olive-grey.

In 1965, Grant studied the surficial sediments of northern Baffin Bay. Marlowe (1968) added the third dimension by analyzing cores in a reconnaissance study of the bottom sediments over most of Baffin Bay.

SAMPLING AND LABORATORY PROCEDURE

Sampling Program

The sediment analyzed in this study was recovered from Baffin Bay during the late summer and early autumn of 1967 (Blee, Baker and Friedman, 1968; Codispoti and Kravitz, 1968). The sampling area runs from just north of Davis Strait to the southern part of Kane Basin (Fig. 1).

Laboratory Procedure

X-ray Analysis

The cores were X-rayed within their plastic liners to determine the sedimentary structures present, the amount of sediment heterogeneity, and the location and dimensions of any pebbles, thereby establishing the most desirable core-splitting orientation for the sedimentological study to follow (Baker and Friedman, 1969).

Granulometry and Colour

After each core was split, the material was logged for colour (National Research Council, 1948) and sedimentary structures and then subdivided into samples based on differences in texture or colour. The samples were wet-sieved with demineralized water through a number 230 mesh sieve (62 micron opening). The coarser than 62 micron fraction was dried and sieved, using 1/4 φ calibrated sieves in an Allen-Bradley Sonic Sifter.

The <62 micron fraction was diluted to 1,000 ml and treated with 0.2 to 2.0 gm of Calgon. The size distribution of the fine-grained fraction was obtained by using the pipetting procedure outlined by Folk (1968, p. 37-38.

After each sample was pipetted an additional colour examination was made on the 9φ and finer clay material and compared with the colour of the untreated sediment.

Statistical Analysis of Data

Moment measures determined by computer were used in the analysis of the grain-size data (Friedman, 1962). Only the first and second moments (mean and standard deviation) gave meaningful results. The masking effect of coarse, ice-rafted grains on the size frequency data provided a problem in interpretation.

To perform meaningful statistical analysis on the size frequency distribution of sediments which have resulted from two such diverse and distinctive processes as ice-rafting and water transport, it becomes imperative to separate the two size populations that were derived from these two distinctive processes. Effective separation is made possible by considering that under normal circumstances, ice has a much greater competence for carrying large particles than water or wind. So while ice, water, and wind can deposit similar fine-grained sediment, indistinguishable as to agent of transport, sieving analysis of the coarser fraction of the samples frequently reveals a discontinuous distribution of sizes in sediments that have been subjected to ice-rafting. By contrast, water and wind-laid sediments, including those that are strongly polymodal, show a more or less continuous spectrum of sizes, even if some of the sizes are quantitatively deficient (Friedman, 1967). The discontinuity at the coarse end of the size frequency distribution in sediments that have been subjected to ice-rafting is represented in this study by a lack of grains on two or more sieves interrupting the continuous size spectrum within the medium sand or coarser size range. This discontinuity in the size frequency distribution is taken as the break between the coarser ice-rafted material (hereafter referred to as "random gravel") and the finer-textured water-laid sediment. Samples with continuous distributions are treated as having been deposited by water only.

Although this method of differentiating between ice and water deposited sediment is admittedly imperfect, it does allow a rough quantitative separation of the masking effect of ice-rafting from the sediment distribution attributed to water transport.

Presentation of Data

Various quantitative mapping techniques exist for illustrating sediment size distributions (Pelto, 1954; Forgotson, 1960). For this study, the "distance" or "D" function was found to be the most effective (Fig. 4) (Pelto, 1954).

The use of more than one triangular diagram on any one map allows more than three end-members to be represented (Figs. 5 and 6). Two triangular diagrams on a single map were used to provide a better understanding of the sediment textures within Baffin Bay. The gravel-sand-mud (G-S-M) triangle was used to differentiate the coarse part of the size-frequency

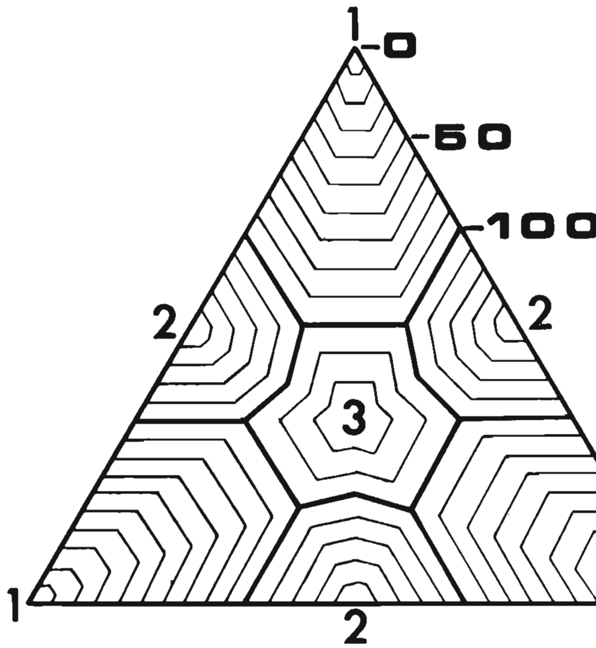


Figure 4. Three-component system, known as D-function, used in mapping bottom sediments of Baffin Bay. The system is divided into 7 classes by the heavy 100 line. The "1" classes are areas characterized by the predominance of one end member; the "2" classes are areas characterized by the mixture of 2 of the 3 end members; and the "3" class is the area characterized by the mixing of all 3 end members. The contour interval is 10 units and decreases in all 7 classes from the 100 line (after Pelto, 1954).

distribution (sediment size larger than 4Ø or 62μ), whereas the gravel plus sand-silt-clay (G&S-Z-C) triangle served to differentiate the fine part of the distribution (sediment size smaller than 4Ø or 62μ). In this study the size terms are defined as follows: gravel > -1Ø or larger than 2.000 mm, sand -1Ø to 4Ø or 2.000 mm to 62μ, silt 4Ø to 8Ø or 62 to 4μ, and clay > 8Ø or <4μ. The term mud follows Folk (1968) in combining silt and clay.

To illustrate clearly the masking effect of ice-rafting in different parts of the study area, the texture of the bottom sediments was presented in two ways. In Figure 5, the effect of the ice is included by incorporating the ice-derived "random gravel" into the weight percentage calculations. In Figure 6, the influence of ice as an agent of sediment transport has been removed by excluding the "random gravel" in the weight percentage calculations.

RESULTS

Bottom Sediment Characteristics

D-function

Figure 5 shows the distribution of the bottom sediments in Baffin Bay. Ice-rafted sediment is included in this figure. The numbers identifying the various classes of sediment texture in the two D-function triangles may be thought of as a rough measure of the relative degree of sediment coarseness with "1" representing the gravel class and "9" representing the clay class.

In general, coarse-textured sediment is restricted to the shelf areas, especially the Greenland shelf, whereas finer-grained "silty clay" and "clay" occupies the deeper basin. Similar fine-textured sediment is also found in shallower water within the large depression of Melville Bay, part of

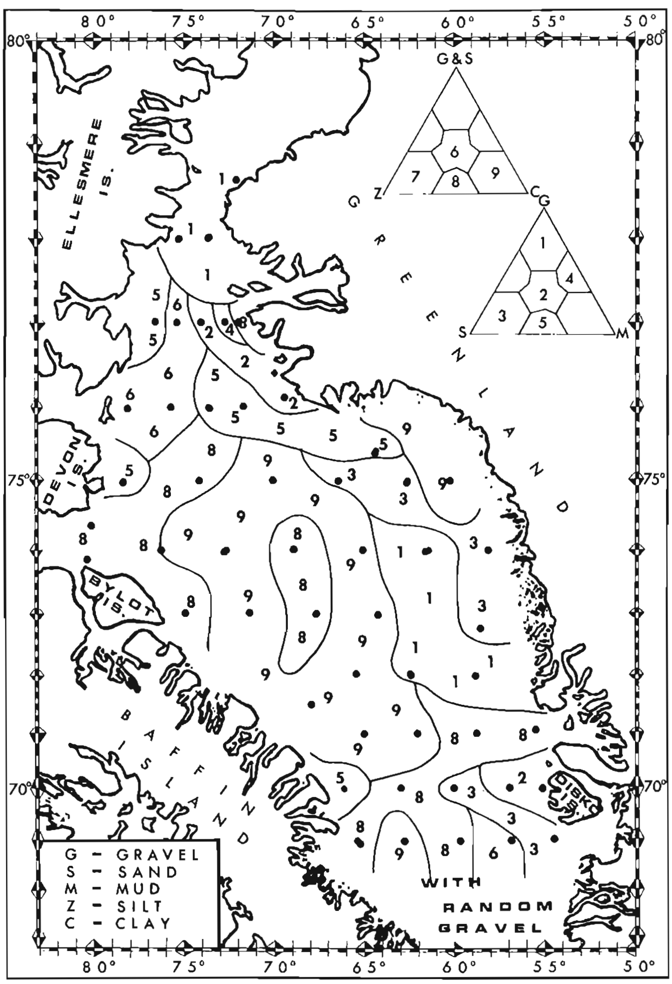


Figure 5.

Distribution of bottom sediments in Baffin Bay study area. Ice-rafted "random gravel" (see "Statistical Analysis of Data" in text) is included in the size-frequency distribution. The two triangles in upper right corner are D-function triangles (see Fig. 4). For grain-size ranges see text; the term mud combines silt and clay (Folk, 1968).

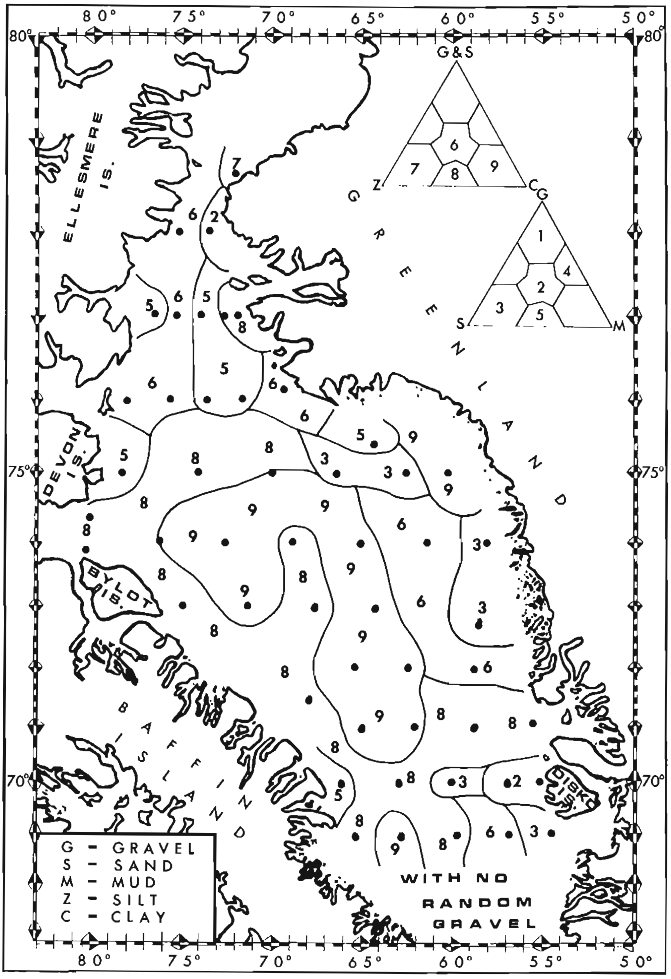


Figure 6.

Distribution of bottom sediments in Baffin Bay study area. Ice-rafted "random gravel" is excluded; compare this figure with Figure 5.

the west Greenland shelf (for location see Fig. 1), and in the deep fiord off New Thule in northwest Greenland. The "silt" textural class is absent in the bottom sediments.

On the shelf northwest of Disko Island in the southeastern part of the study area (Fig. 5), a strip of "silty clay" (designated by number 8) is found bounded by much coarser sediment to the north and south.

The "gravel" area contiguous to and north of this "silty clay" strip trends essentially northwest to southeast, is bounded on the east and north by "sand", and extends roughly as far west as the slope break. The other "gravel" occurrence is in Smith Sound and in the southern part of Kane Basin (for location see Fig. 1).

In Figure 6, the effect of the coarse ice-rafted material has been removed from the size frequency distribution of the sediment. The "gravel" areas of Figure 5 have been replaced by much finer-grained sediment types, but other areas of the bay are not appreciably affected by excluding the ice-rafted material.

Colour of Bottom Sediments

Figure 7 shows the distribution of colour in the bottom sediments of Baffin Bay. The map legend uses five classes of shade and colour to represent relative degrees of oxidizing or reducing conditions in the sediments.

Colours indicating oxidizing conditions (class 1) occur only below the "shelf" break, restricted to the deeper parts of the bay. Intermediate colour classes 2, 3, and 4 are found mostly south and west of Melville Bay (for location <u>see</u> Fig. 1), west of Disko Island, and in the southern part of the deep bathyal plain. Influence of reducing conditions is represented by colour class 5 and is a characteristic of sediments in the remaining shelf and upper slope areas.

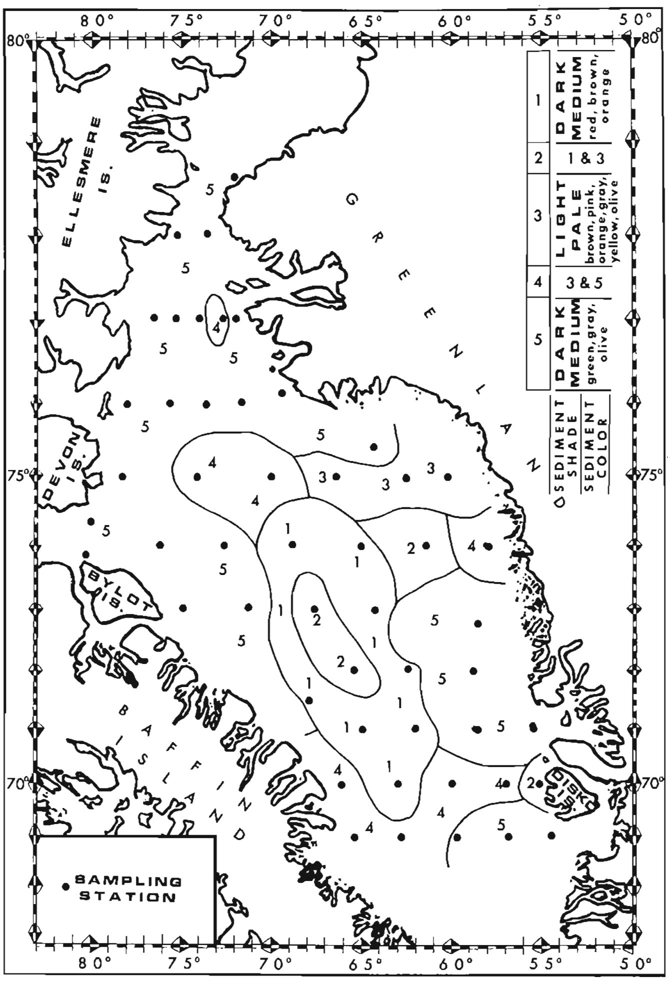


Figure 7.

Distribution of colour in bottom sediments of Baffin Bay study area. Colour class 2 represents a mixture of colours from classes 1 and 3. Colour class 4 represents a mixture of colours from classes 3 and 5.

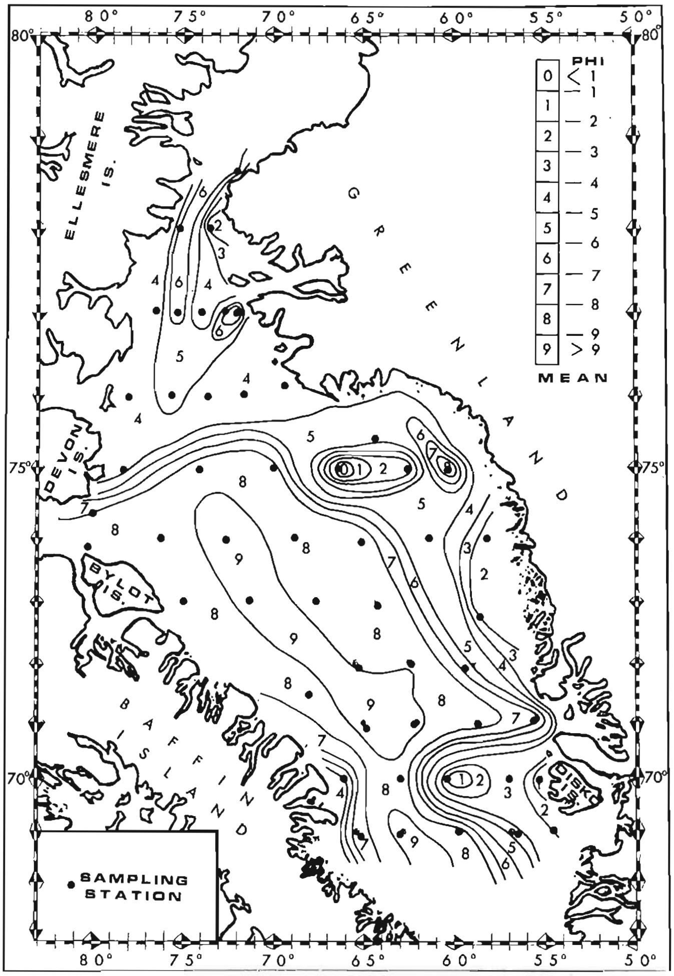


Figure 8.

Distribution of mean grain sizes for Baffin Bay study area.

Mean Grain Sizes of Bottom Sediments

Figure 8 illustrates the mean grain sizes for the bottom sediments of the Baffin Bay study area, devoid of the ice-rafted "random gravel" part of the size frequency distribution.

Figure 8 shows that the deep basin contains the finest sediment in Baffin Bay with a coarsening of the mean grain size towards the shelf areas. The eastern shelf of the bay has a coarser mean grain size than the western shelf. In the shelf depression of Melville Bay (for location see Fig. 1) the sediment is as fine-grained as in a large part of the deep basin. Local areas of coarse mean grain sizes are west of Disko Island and Melville Bay.

Sorting of Bottom Sediments

Figure 9 shows the distribution of sorting (standard deviation) values for the bottom sediments in the Baffin Bay study area. The ice-rafted "random gravel" part of the size frequency distribution was excluded from the computation of the standard deviation.

Standard deviation values range between 2.03 and 5.29, hence the sorting is very poor to extremely poor (Friedman, 1962). Standard deviation values between 2 and 3 for the size frequency distribution of the bottom sediments occur mainly on the inner (proximal) west Greenland shelf, on the Baffin Island shelf, and over a large part of the bathyal plain. Values exceeding 3 are mostly confined to Smith Sound (for location see Fig. 1) and contiguous areas and to the outer west Greenland shelf.

Subbottom Sediments of Baffin Bay

Subbottom Sediment Distribution

Subbottom sediment characteristics will be discussed along a roughly north-south profile through the centre of the bay (profile A-A', Fig. 10). The

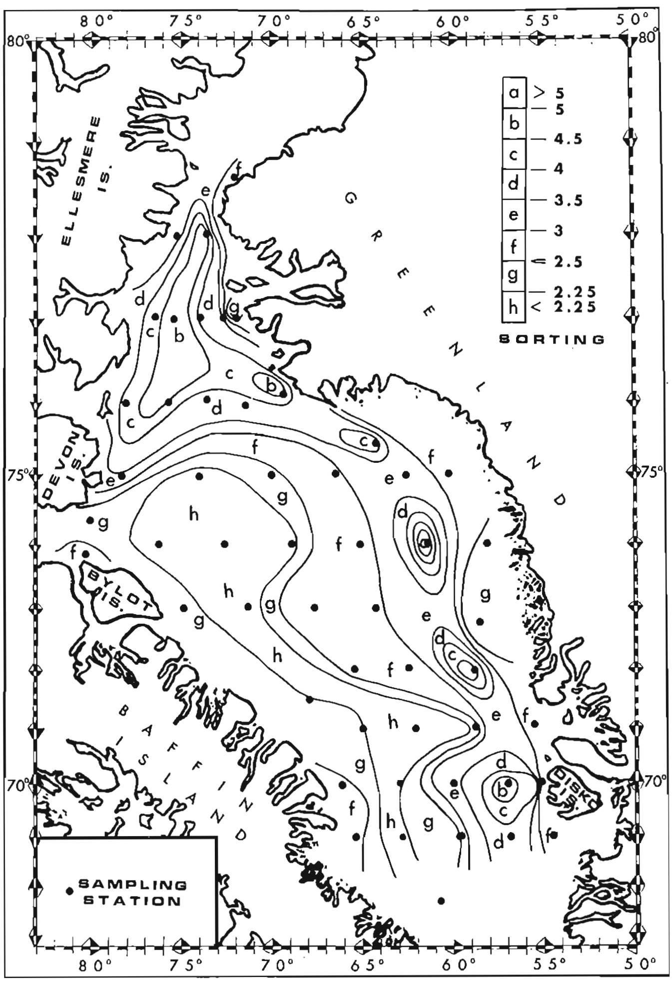


Figure 9.

Standard deviation (sorting) values for Baffin Bay study area.

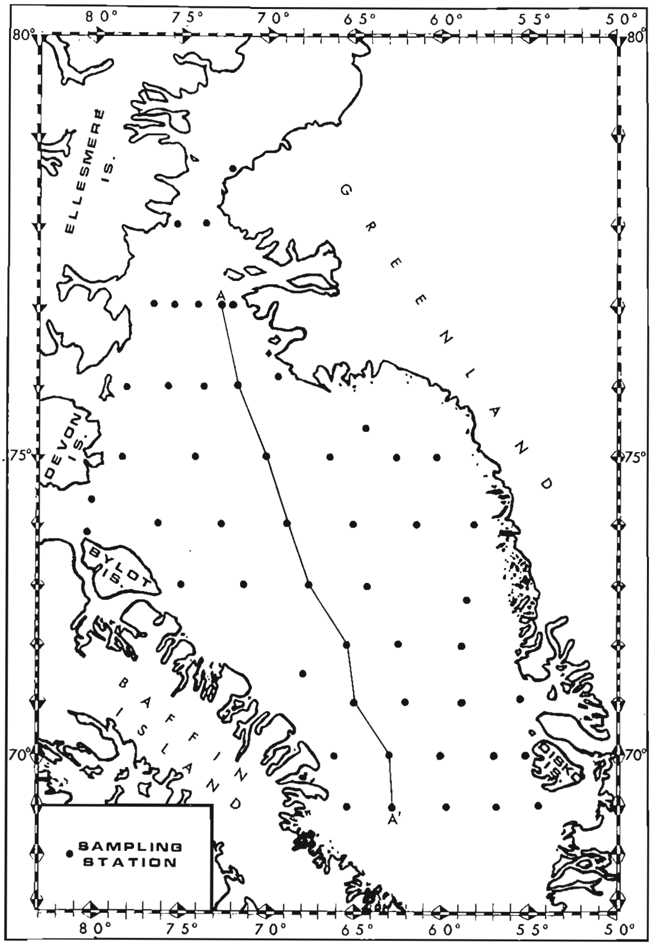


Figure 10.

Index map of Baffin Bay showing location of sediment profile A-A' (<u>see</u> Fig. 11).

bottom topography and the location of the cores along profile A-A' are shown in the top half of Figure 11.

Statistical analysis of sediment texture at depth within cores is made difficult by the presence of ice-rafted "random gravel" (Fig. 12). Ice-rafted particles, "dropped" in random fashion by passing ice, may penetrate into the subbottom sediment. The amount of penetration depends on the velocity of impact and the relative size and shape of the ice-rafted particles, as well as on the mass physical properties of the bottom and subbottom sediments.

Even with X-ray analysis, accurate determination of the amount of penetration of the ice-rafted particles is frequently impossible. Therefore, in Figure 11 the ice-rafted "random gravel" has not been included in the sediment profile. The sediments shown are those which we consider to have been deposited concurrently and hence excludes the ice-rafted "random gravel" which may have become incorporated in the subbottom sediment by later penetration.

Profile A-A' in Figure 11 illustrates the degree of sediment heterogeneity within each core and the wide variation in sediment texture between

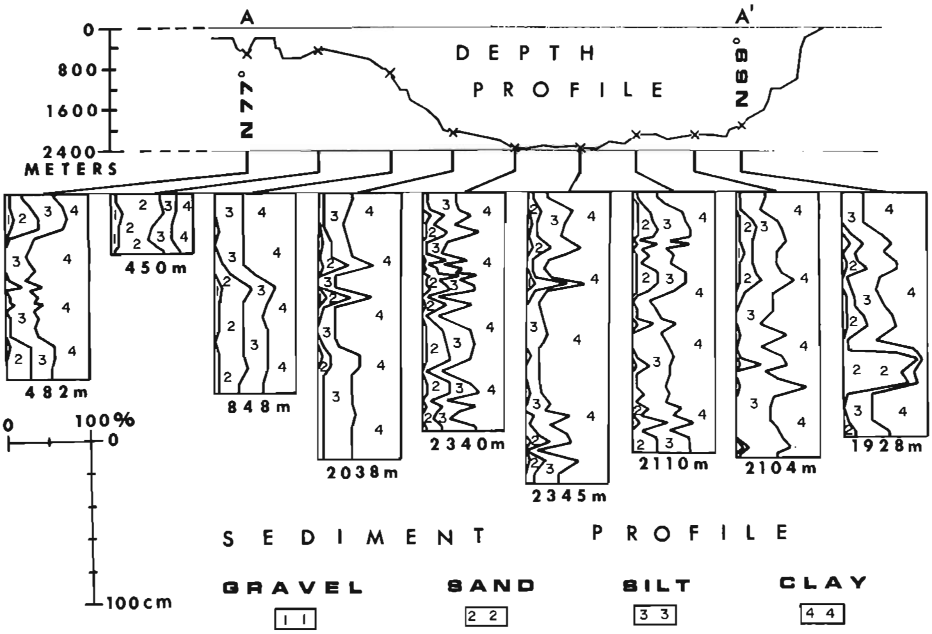


Figure 11. Depth and sediment profile along A-A' in Baffin Bay study area (see Fig. 10).

cores. Although a general relationship between coarser sediment texture and shallower water depth exists in the bottom sediments (Figs. 5 and 6), no such relationship is shown in the subbottom sediments (Fig. 11). Equally fine and coarse sediment textures occur in the cores of both shallow and deep water. Generally, the texture of bottom sediments is not representative of subbottom sediments.

Texture and Colour of Subbottom Sediments

In Figure 13, variations in sediment texture using the D-function classification (Fig. 4) are compared with colour variations in the cores along profile A-A' (Fig. 10). No apparent relationship exists between sediment texture and colour.

Although changes in colour are common within the core profiles, a relatively distinct shift is shown in oxidizing conditions from north to south. The first and third most northern cores in Figure 13 show "reducing" sediment colours (classes 4 and 5) overlying colours indicating more "oxidizing" conditions (classes 1 and 2); the second core was too short to show any change. The fourth core shows a dominance of oxidizing conditions throughout its length. The fifth to eighth cores inclusive indicate an oscillating colour pattern, yet trend toward "oxidizing" colours closer to the surface. The ninth core shows a variable colour pattern without any distinct trend.

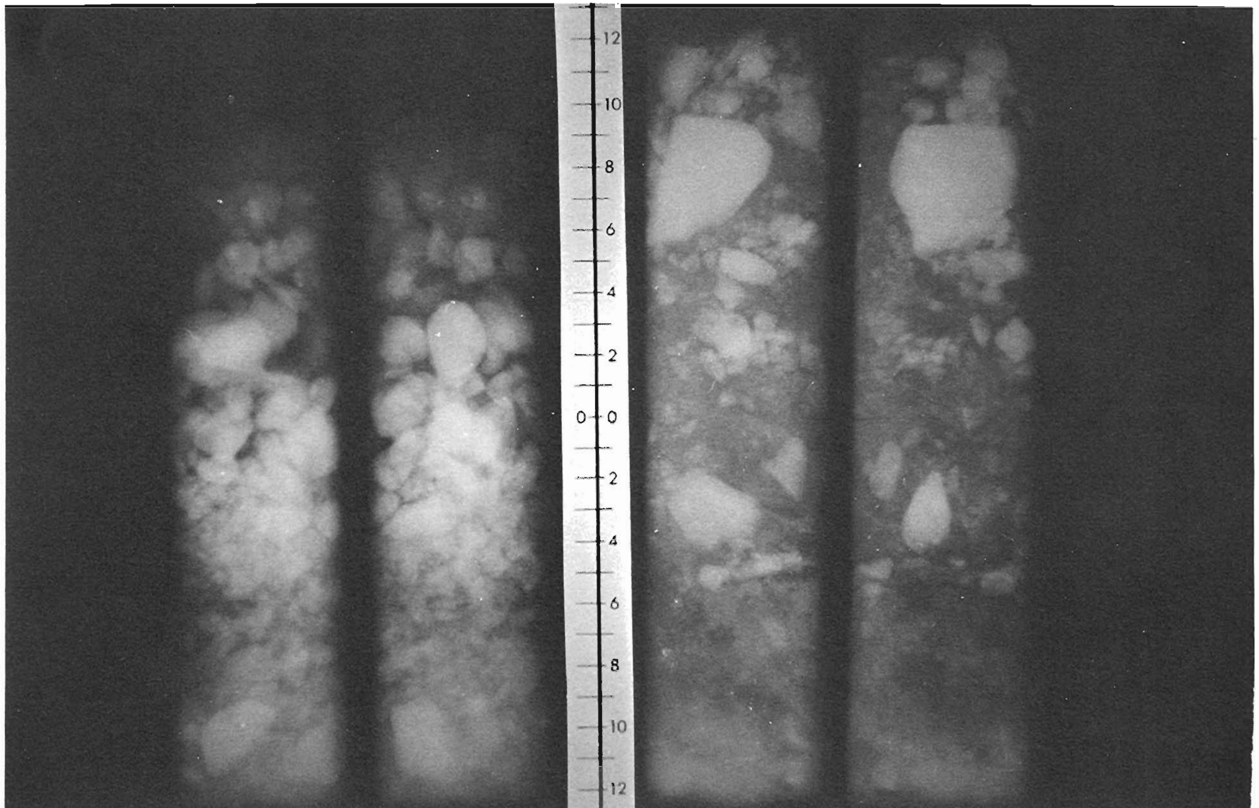


Figure 12. Radiographs of two pairs of core segments showing ice-rafted pebbles in subbottom sediment. The two radiographs of each pair to the left and right of the centimetre scale are at 90 degrees to each other.

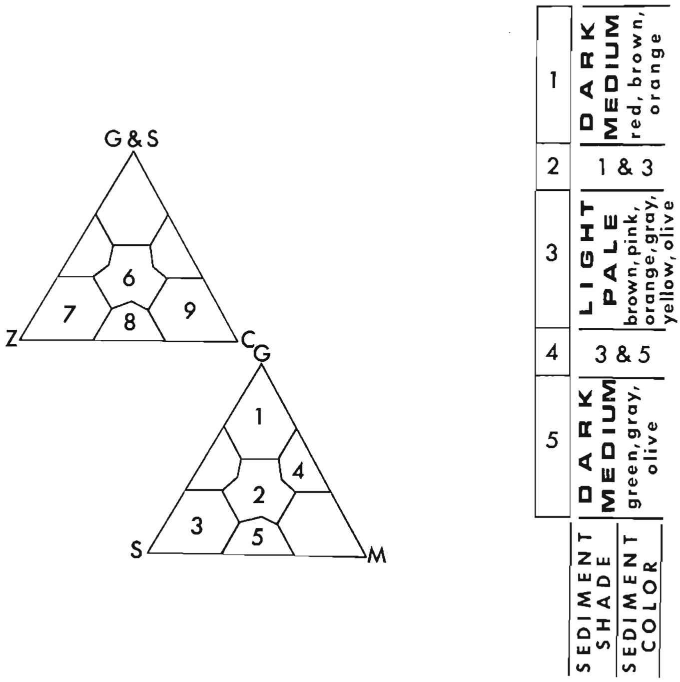
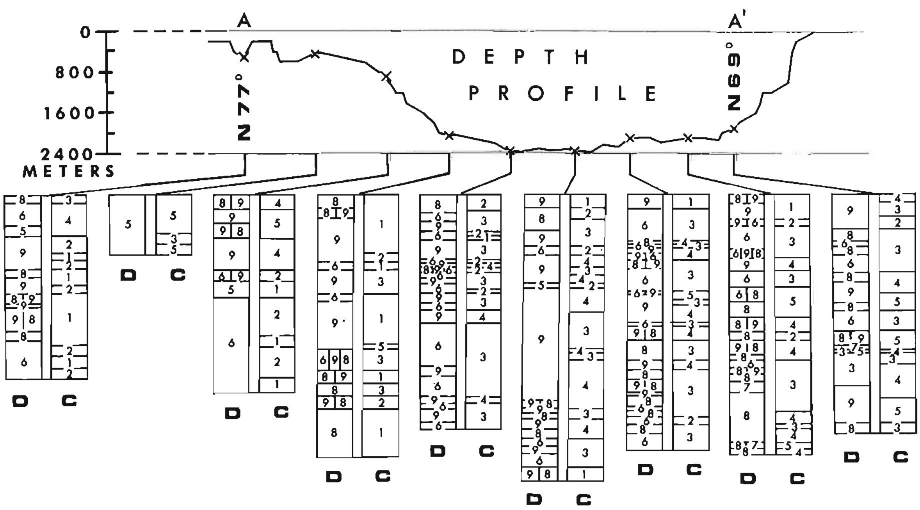


Figure 13.

Depth and sediment profile along A-A' showing variations in texture (using D-function classification of Fig. 4) and colour. The two triangles in upper part of figure are D-function triangles. For end members of triangles see Figures 5 or 6. More than one number in the same D-function core interval indicates a borderline sediment texture.



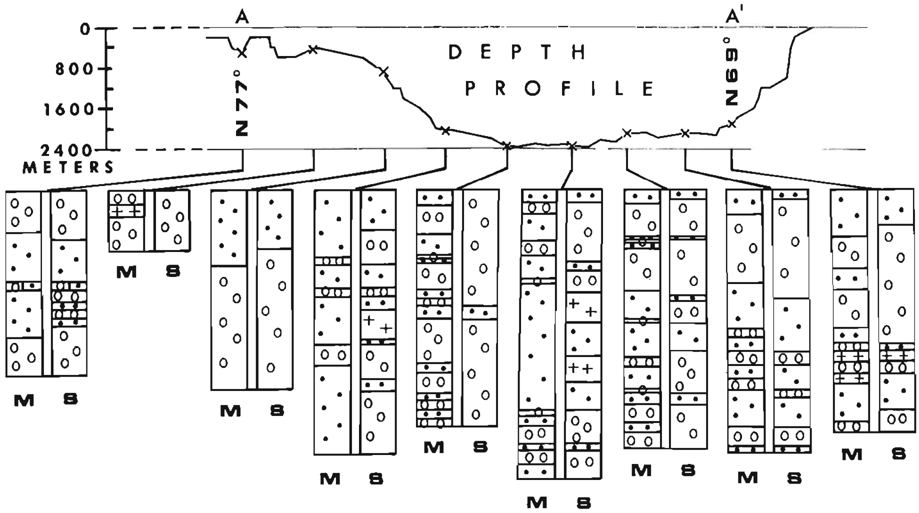
D-FUNCTION (D) AND COLOR (C)

PROFILE

Mean Grain Size and Sorting in Subbottom Sediments

In Figure 14, variations in mean grain size are compared with those of sorting in the cores along profile A-A'. In general, where the mean grain-size falls in the clay class, sorting is better than where mean grain-size falls in the silt class.

For the bathyal basin of Baffin Bay (Fig. 11), the cores show on the average more clay in the bottom sediments than in the subbottom sediments; conversely the bottom sediments contain on the average less sand and gravel than the subbottom sediments. Hence the mean grain-size of bottom sediments



MEAN (M) AND SORTING (S)

PROFILE

MEAN PHI	SEDIMENT TYPE	DIAGRAM SYMBOL
−1		
	SAND	+ +
+4		
	SILT	o o
+8		
>+8	CLAY	• •

PHI UNITS	SORTING	DIAGRAM SYMBOL
1.4		
	POOR	+ +
2.0		
	VERY POOR	• •
2.6		
>2.6	EXTREMELY POOR	o o

Figure 14. Depth and sediment profile along A-A' showing mean and standard deviation (sorting) values in sub-bottom sediments. More than one symbol in the same core interval indicates a borderline value for mean or standard deviation (sorting).

on the average is finer-grained than that of subbottom sediments and the sorting of bottom sediments tends to be better than that of subbottom sediments. No distinct trends are apparent on the shelf areas.

Subbottom Sedimentary Structures

X-ray analysis commonly illustrates heterogeneity in sediment density that is frequently indistinguishable to the eye after a core is split (Baker and Friedman, 1969). Of the 28 cores X-rayed, only five cores contained essentially homogeneous sediment texture throughout the core length. Stratification in the cores is primarily horizontal, but cross-bedding occurs in six cores taken at stations in deep water (Fig. 15). Distinct micro-crosslaminated forests have thicknesses as small as 1 mm. Some of the intervals of micro-cross-lamination occur with contorted or wavy laminations. Worm burrows are found in the cores from several stations.

DISCUSSION

Relationship Between Sediment Texture and Bathymetry in Baffin Bay

Classically, sediment texture is supposed to reflect bottom topography. The higher energy at shallow water depths results in the deposition of coarser-grained sediments on topographic highs while finer-grained sediments settle into lower energy topographic lows. In general, this relationship is found in the bottom sediments of the Baffin Bay study area (Fig. 6). Depressions on the west Greenland shelf contain sediment as fine-grained as that found in the deep bathyal basin. Depressions as little as 200 to 400 metres below the surrounding topography, as in Melville Bay and Smith Sound (for location see Fig. 1), may cuase a change in sediment mean grain-size on the order of 3ϕ (Fig. 8).

Local variations in the distribution patterns of statistical parameters, however, appear to be more related to sediment source and mode of transport than to bottom topography. In specific areas the distribution of sediment mean grain-size, such as west of Melville Bay (for location see Fig. 1) and west of Disko Island (Fig. 8), and sorting, mostly on the west Greenland shelf (Fig. 9), disregards bathymetric control. Figure 12 shows the presence of gravel on the shelf approximately 70 miles (113 km) west of the Greenland coast. Gravel this far from shore indicates transport by ice-rafting.

Interplay of Ice and Bottom Sediment

The ice which composes the terminus of a glacier may contain many tons of sediment either as a superglacial coating or as an internal meltwater deposit. As a given iceberg breaks off from the front of a glacier, it moves into the fiord and seeks an equilibrium configuration in the water due to its irregular shape, thereby shifting the location of the sediment load. The calving iceberg may strike the bottom of the fiord and stir up fine sediment which may then be carried out of the fiord by the water currents caused by the calving.

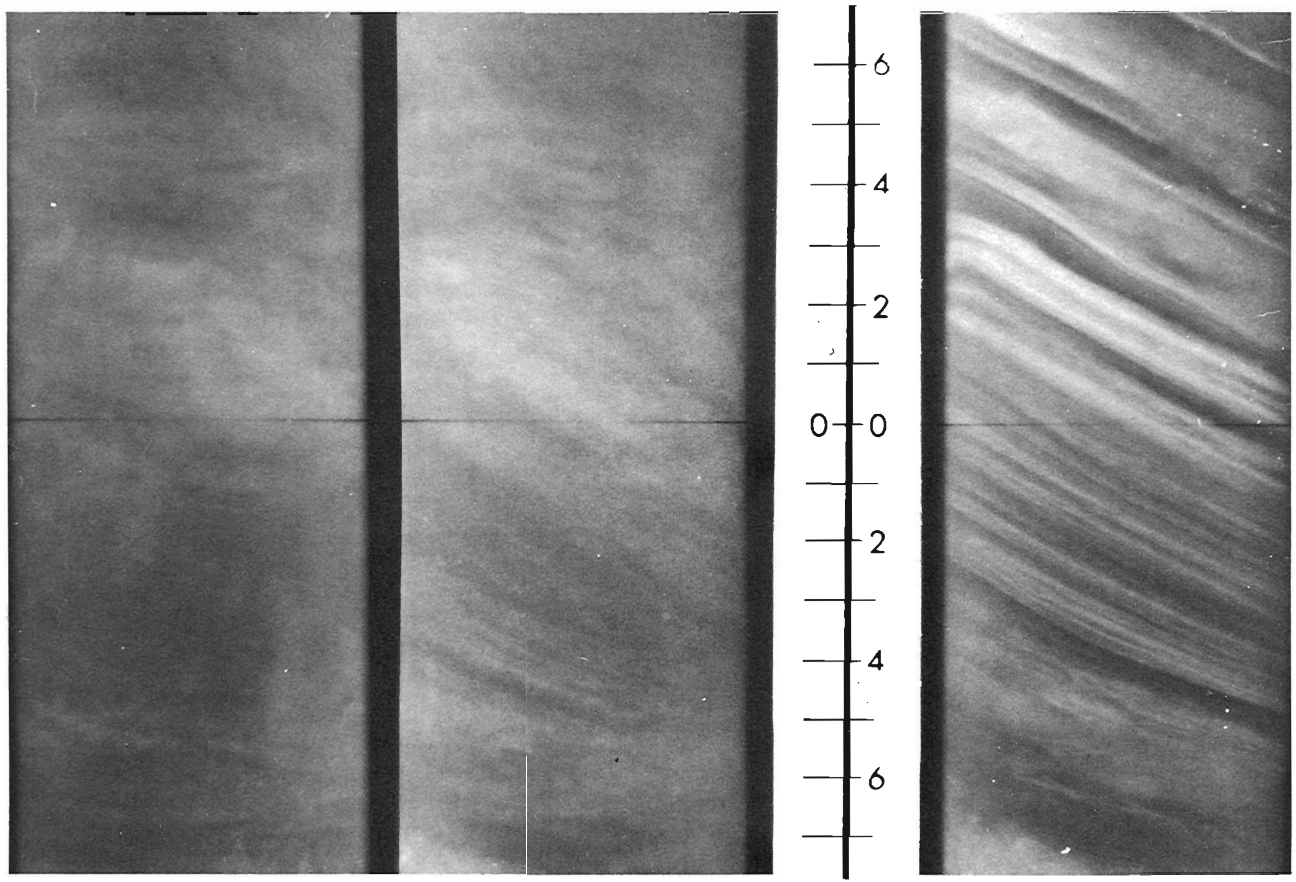


Figure 15. Radiographs of the same core segment taken at the 0° (perpendicular to strike of inclined laminae), 45°, and 90° (parallel to strike) core orientation (scale in cm).

Icebergs in Baffin Bay have heights above water level ranging up to 400 feet (120 m); their lengths may be up to 5 miles. As many as 6,000 icebergs may be jammed into a single fiord, as is frequently the case in Jacobshavn, Umanak, Rink, and Umiamako fiords of west Greenland. In 1940, the USCG NORTHLAND estimated a total of 6,000 icebergs in the open water of Baffin Bay, indicating that it may take five years or more between the time an iceberg calves to the time it leaves the fiord (Hawley et al., 1941). Tarr (1897) reported that before travelling a great distance from the fiord, icebergs had lost much of their sediment load, as indicated by the clean appearance of icebergs observed in the southwestern part of the bay.

Depending on the nature of the sediment distribution within the iceberg and the temperature of the water through which the iceberg travels, which in turn is related to the surface current pattern and to a lesser extent to unusual shifts in wind direction, the sediment is released in an unpredictable, random manner practically anywhere in the bay. However, the fact that most icebergs follow a well-defined distribution pattern and unload their sediment load fairly quickly does allow the statistical parameters of size frequency distribution determined in this study to be discussed as an interplay of ice and sediment.

Most of the icebergs in Baffin Bay are calved from fiords between Cape York (latitude 76 degrees, longitude 66 degrees) and Disko Island. The icebergs are carried northward by the surface currents (Fig. 3) along the west Greenland coast to Melville Bay, westward past Cape York to the vicinity of Devon Island, and southward along the east coast of Baffin Island (Hawley et al., 1941). Comparison of Figures 5 and 6 shows how ice-rafting controls the texture of bottom sediment. The effect of ice-rafting is especially evident in two areas, one on the west Greenland shelf (where the sediment is classified as class 1 on Fig. 5 and as class 6 on Fig. 6) and the other in Smith Sound and contiguous areas to the southeast (for location see Fig. 1) (where the sediment is classified as classes 1 and 2 on Fig. 5 and as classes 5, 6 and 7 on Fig. 6). Classes 1 and 2 in these two areas of Figure 5 are sediments with abundant ice-rafted gravel and sand.

The coarse texture of shelf sediments southwest and west of Disko Island (size classes 2 and 3 of Fig. 5) is probably related to the icebergs from Jacobshavn ice fiord (southeast of Disko Island).

The fine-grained sediment strip northwest of Disko Island (size class 8 of Fig. 5) is considered to have been caused by rapid transport and deposition of fine-grained glacial material during late spring meltwater floods from Umiamako, Rink, and Umanak fiords east of this area. The mode of transport for this fine-grained material is suggested to be a high density hyperpycnal flow (Bates, 1953) forming a submarine deltaic deposit between latitudes N70° and N72°. This hypothesis is supported by the mass physical properties of the fine-grained sediment within the cores taken in this area. Rapid rates of sedimentation are indicated by high water content, low wet unit weight, and high void ratio in these sediments as shown on Table 1. Further evidence that this fine-grained sediment comes from the east is derived from the bathymetry (Fig. 2). This area of the shelf is incised by a distinct relatively deep submarine canyon which extends shoreward almost to the mouths of the fiords. The canyon extends westward to a dissected part of the slope (latitude N71°) which may have acted and probably still acts as a funnel for the fine-grained sediments that are brought down into the bathyal basin. The absence of ice-rafted material in the cores of the area northwest of Disko Island is due to the infrequent passage of icebergs in the immediate area. Analysis of surface currents (Fig. 3) and ice observations (Hawley et al., 1941) indicate that the icebergs from Jacobshavn fiord that travel into Baffin Bay south of Disko Island are deflected westward by surface currents south of latitude N71°.

Table 1

Mass physical properties of cores at three easternmost stations on latitude N71°, northwest of Disko Island, Baffin Bay area (strip indicated as sediment texture class 8 in Figs. 5 and 6) Core stations 1, 2 and 3 are from east to west (Station 1 is contiguous to fiords).

Stations	1	2	3
Water content (%)	50 - 85	84 - 113	60 - 171
Wet unit weight (gm/cm^3)	1.51 - 1.62	1.42 - 1.52	1.30 - 1.65
Void ratio	1.39 - 2.28	2.30 - 3.06	1.62 - 4.67

The unusually coarse sediment mean grain-size values southwest of Melville Bay (for location see Fig. 1) (Fig. 8) are postulated to have resulted from the concentration of icebergs as indicated by the confluence of surface currents (Fig. 3). However, no such concentration of icebergs in this area has been reported in the literature.

The map patterns for mean grain-size and sorting (Figs. 8 and 9) in the deep bathyal plain appear to have little relationship to bathymetry (Fig. 2) or to the surface current pattern (Fig. 3). The suggestion by Marlowe (1968) that the median diameter of bottom sediments in the deep basin is related to the trend of the Baffin Land Current is not supported by this study.

Sediment Colour

Dark or medium red, brown and orange sediment (called class 1 on Fig. 7) floors the deep central part of the Baffin Bay study area, an observation already made by Marlowe (1968). In the present study sediment colour is thought to be related to the degree of biochemical decomposition of organic material which determines the amount of organic material incorporated into the subbottom sediments. The amount of biochemical degradation of organic material is reflected by the valence state of iron in the subbottom sediments. Emery (1960) suggested that the red masking colour effect caused by the ferric iron state is probably more important in sediment colouration than the contribution of the ferrous iron state to the normal green colour of clay minerals.

In the subbottom sediments, the oscillation of sediment colour throughout the length of most of the cores (Fig. 13) indicates variations in oxidizing and reducing conditions and discounts the possibility that the colour changes are post-depositional. Sediment being buried below an oxygenated water/sediment interface would show only a uni-directional change from ferric to ferrous valence states.

Figure 13 indicates a shift in oxidizing conditions from the shallow northern parts of the study area to the deep bathyal plain farther to the south. In the north, sediments with colours associated with reducing conditions overlie those having colours suggestive of oxidizing conditions. The "oxidized" colour is considered to be related to a sea level that was about 300 to 400 metres lower during times of maximum glaciation. This change in sea level has been proposed by physiographic and sedimentological studies in the western Arctic Archipelago (Pelletier, 1962 and 1966; Horn, 1963; Marlowe and Vilks, 1963; Marlowe, 1964 and 1965). Much of Smith Sound and the northern part of Baffin Bay should have been subaerially exposed during these glacial maxima resulting in the oxidation of the sediment that floored this part of the bay. With the warming trend and the melting of the ice front back to its present location, it is suggested that the resulting rise in sea level would have submerged the previously subaerially exposed areas and sediment colour would have become more indicative of reducing conditions.

In the deep basin to the south, shifts in sea level and the extent of glaciation played more indirect roles. During the glacial maximum, the decrease in sea level of 300 to 400 metres would put the effective depth of the Davis Strait ridge at 300 to 400 metres, thereby greatly reducing the interchange of water between the bay and the Atlantic Ocean. In addition, the ice probably cut off much of the inflow from the Arctic Ocean, further decreasing the water circulation in Baffin Bay. The oscillation of colours suggestive of intermediate to reducing conditions in the sediments of the cores (Fig. 13) is probably caused primarily by either variations in the amount of organic material settling to the bottom of the bay or to changes in the effective threshold depth of Davis Strait with fluctuating sea level.

The rise in sea level, marking the end of the glacial period, caused a greater interchange of water between Baffin Bay and the Atlantic Ocean while the shift in the ice front landward opened the water channels with the Arctic Ocean. The enlarged water column in the bay and increased oxygen content resulted in a trend toward a more oxidized colour in the sediment.

Deep Bay Sandy Layers

Sandy layers are found within the cores from the deep basin of Baffin Bay (Fig. 11). The sorting of most of these layers ranges from very poor to extremely poor (Fig. 14) and is characteristic of downslope density currents or slumps. Marlowe (1968) discusses the mechanism for transporting such material.

The colour of many of these sandy layers indicates intermediate or oxidizing conditions whereas the colour of the interbedded finer-grained layers indicates intermediate or reducing conditions. It is suggested that these more oxidized sandy layers indicate deposition by meltwater streams of subaerially exposed sediment from the shelf or the uppermost part of the slope during the time of maximum Pleistocene glaciation when the ice front had advanced to cover all of the Baffin Island shelf and approximately half of the west Greenland shelf (Emery, 1949).

Bottom Current Characteristics

Various lines of evidence indicate that water circulation in the deep basin is restricted. At water depths of 2300 meters, dissolved oxygen content is 3.2 ml/l as compared with a surface value of 10 ml/l in northern shelf areas (Collin, 1964). Riis-Carstensen (1936) found dissolved oxygen content of 3.5 ml/l at a 2,000 m depth and 7.6 ml/l at the surface.

No sedimentary structures other than horizontal bedding occur within the top 47 cm in any of the cores studied. The scarcity and limited thickness of intervals found containing cross-laminations and possible ripple marks indicate that bottom currents when they did occur were probably of short duration and localized. These sedimentary structures occur in layers containing 71% to 96% material finer than 62 microns. This observation suggests that the structures may be formed by currents caused by small down slope slumps of fine-grained sediments.

Of the 736 samples analyzed, 99% were composed of more than 20% by weight of sediment finer than 62 microns. The occurrence of this fine-grained material in most parts of the bay discounts winnowing by bottom currents as an important process at any water depth.

ACKNOWLEDGMENTS

This study was sponsored by the Office of Naval Research under contract No. N00014-67-A-0117-0004. The Baffin Bay survey was a joint effort by the Naval Oceanographic Office and Rensselaer Polytechnic Institute.

Sincere thanks are extended to Mr. Joseph H. Kravitz of Navoceano for making the study possible and for invaluable aid during the survey and research stages. The outstanding cooperation of the officers and men of the U.S. Coast Guard Icebreakers SOUTHWIND and EDISTO in recovering the cores from Stations 1-48 and 49-51 respectively is gratefully acknowledged. The authors also wish to thank Mr. John J. Blee, formerly of Rensselaer Polytechnic Institute for his assistance aboard ship and during the early research phases of this study.

Additional thanks are extended to Curtis Norwood and John Fisher of Rensselaer Polytechnic Institute for assistance in computer analysis.

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30. A PRELIMINARY ACCOUNT OF BENTHONIC AND PLANKTONIC FORAMINIFERA IN BAFFIN BAY, DAVIS STRAIT AND THE LABRADOR SEA

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Abstract

Benthonic and planktonic foraminifera were obtained in both sediment samples and plankton tows from Baffin Bay, Davis Strait, and the Labrador Sea. Benthonic foraminifera show little order in their species distribution. Form variation in certain species shows some potential for future paleo-oceanographic work. Only two species of planktonic foraminifera were observed in the study area.

INTRODUCTION

During July, August and September of 1970, collections of foraminifera were made in the Labrador Sea, Davis Strait, and Baffin Bay areas from the CSS Dawson. Both bottom collections and plankton tows were conducted, in order to look at aspects of both planktonic and benthonic foraminifera. This is a preliminary report regarding the samples collected. In the near future, the results of more rigorous study of the samples will be published.

Benthonic Foraminifera

The benthonic foraminifera were collected by means of a conventional Van Veen grab and gravity core. The sampling coverage of the entire Baffin Bay, Davis Strait and Labrador Sea areas (Fig. 1) was not ideal due to ice conditions. In certain areas, however, there was excellent coverage. One would be cautious to generalize from such widely separated stations; indeed, even the data from areas of much denser sampling has not permitted much generalization.

The material representing the topmost few centimetres of sediment was washed free of mud, after which splits were taken for further sedimentary analysis. From these 65 stations, over 100 species of benthonic foraminifera and two species of planktonic foraminifera were identified. The abundance or foraminiferal number of benthonic foraminifera was consistently low (5-10/g) in all samples with the exception of those from Pond Inlet and Godhavn Inlet. The observed fauna was generally consistent with faunas reported from Arctic and subarctic regions by Phleger (1952), Brady (1884), and Loeblich and Tappan (1953) and was not too dissimilar from recent faunas found along the coast of Nova Scotia. Phleger's work (1952) in Arctic areas implied a lesser faunal diversity than was encountered, but the supplementary species found have been reported in Arctic studies elsewhere.

A qualitative interpretation of the species present at various stations throughout the sampling area showed little change in diversity. Most of the 100 species were found at all stations. Qualification of the data will be conducted in future studies. Species dominance in contrast to diversity, varied with the station. It was quite difficult to draw any conclusions, when variation in dominant species between two closely and geographically-related stations was greater than between widely-separated or geographically-dissimilar stations, such as locations in the north and south, or in deep or shallow water.

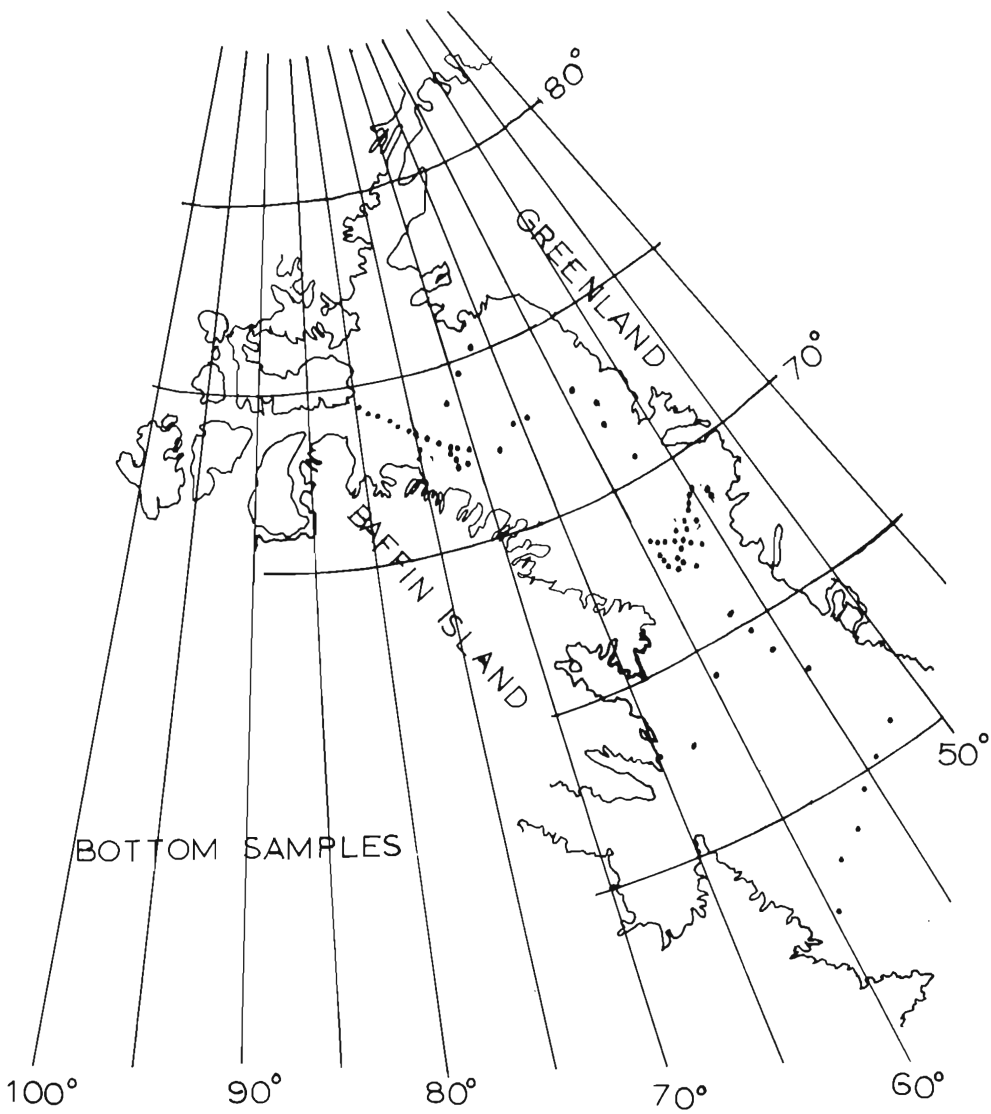

Figure 1. Location of bottom samples in the study area.

With the following exceptions, there was no visual difference in the sediments from localities of differing faunal dominance. Very large arenaceous foraminifera were much more conspicuous in abyssal depths (though still not dominant) than in shallow water; for example, Reophax pilulifer (Brady) and Rhabdammina abyssorum (Sars). Also, the large foraminifera were more predominant in the southern part of the survey area. Generally, too, all the arenaceous foraminifera identified were more commonly found in finer sediments. A phenomena which Gregory (1971) has attributed to destruction in higher energy environments. Another trend observed, was a sudden absence of Eggerella advena below a depth of 200 metres. Vilks (1969) in his studies of Arctic foraminifera found this species to occur in only a shallow water environment.

Generally, the initial investigation showed the foraminiferal distribution to consist of irregular patterns in dominant or persistent species, which fail to conform to any other obvious parameter. Examination of the cores revealed certain graded and sandy textures which could be attributed to bottom transport of sediment. It was therefore hypothesized that the confused distribution pattern of benthonic foraminifera is due to transportation and faunal mixing.

Planktonic Foraminifera

Planktonic foraminifera from both sediment samples and net tows were studied firstly, to investigate their abundances in Baffin Bay, Davis Strait, and the Labrador Sea, and secondly, to look at variations in form of the species Globigerina pachyderma, expected in this area. The form study was done to further investigate a paleo-oceanographic technique first proposed by Kennet (1968). Planktonic foraminifera have never been studied extensively in this area before.

Vertical plankton tow collections to 200 metres were made at the stations shown on Figure 2, using a plankton net with a 75 cm diameter. To quantify plankton abundances, a Ogawa Seiki TS flow meter was attached to the plankton net to measure the volume of water filtered. The plankton samples were preserved in buffered formalin till they were ashed and washed to concentrate the non-combustable constituents (shells, tests, spiculas, etc.).

Figure 3 gives data on the abundance of planktonic foraminifera in the study area. Values are given in numbers of individual's per 1,000 cubic metres of water. No station was devoid of a planktonic foraminiferal population. A majority of the stations have values between 100 and 1,000 individuals per 1,000 cu m of water, which is about an order of magnitude lower than abundances at stations in temperate waters of the North Atlantic Ocean such as those taken by Be and Hamlin (1967) and others. Considerable variation was observed from station to station, possibly resulting from patchy distribution, or perhaps from oceanographic parameters. Studies have not been conducted to date attempting to correlate any of the data from this study with chemical or physical oceanographic parameters.

The interesting part of this study, however, is not in the distribution of the higher and lower abundances of planktonic foraminifera, but in a comparison of their abundance values in the water column to values in the sediments directly below. In general, the number of planktonic foraminifera preserved in the sediment was very low, and sometimes a washed 100-gram sample yielded no planktonic foraminifera at all, while a tow through 400 cu m of water directly above this sediment yielded over 1,000 forams. Assuming that the sediments have taken a few years to be deposited, and that the foraminifera are replaced by new individuals at a fairly rapid rate, the net tow abundances observed in this study would imply that there should be much larger numbers of foraminifera found in the sediments. From this, it is hypothesized that a certain number of the planktonic foraminiferal tests are dissolved or somehow removed from the area after death and never preserved in the geologic record. Thus, the use planktonic foraminifera as a measure of sedimentation rates in the study area is highly questionable.

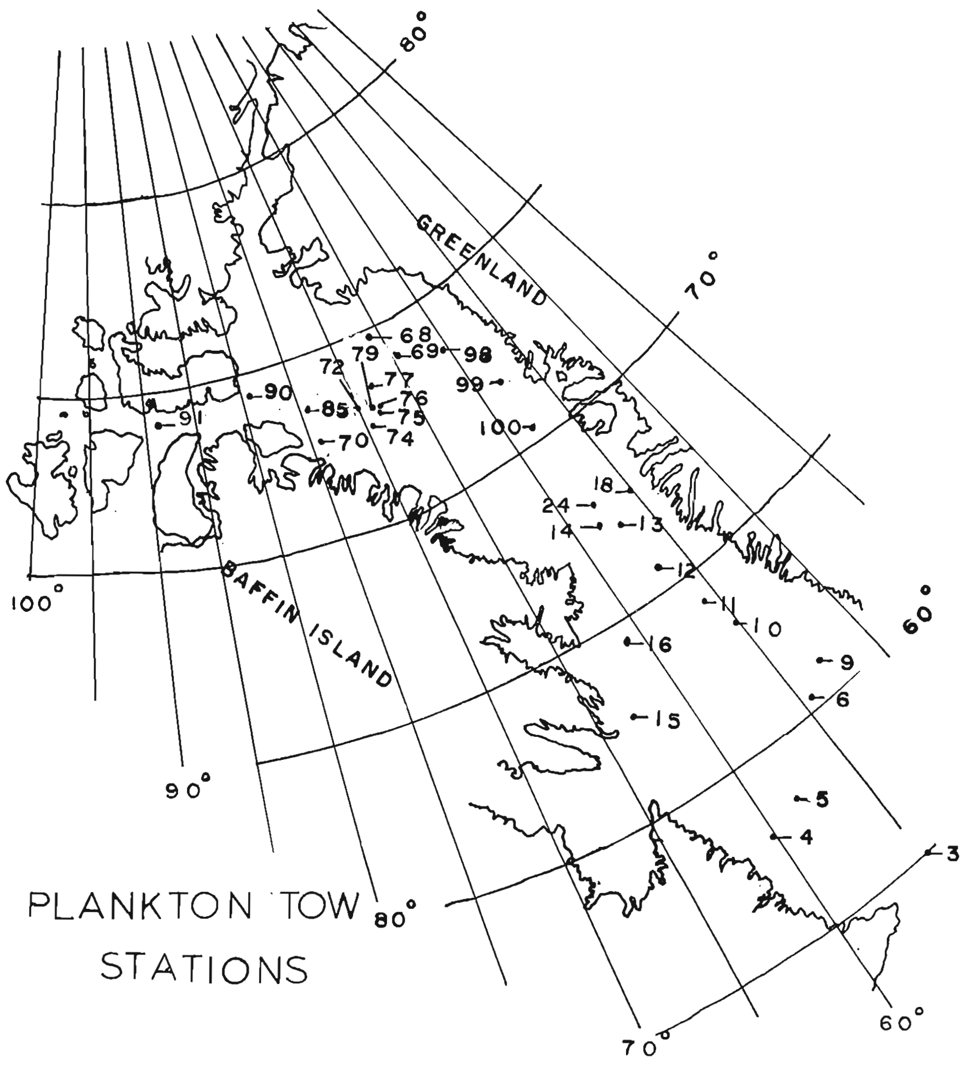

Figure 2. Location of plankton tow samples in the study area.

The examination of the planktonic foraminiferal fauna revealed only two species for the entire study area: Globigerina bulloides d'Orbigny and Globigerina pachyderma (Ehrenberg) of which the latter was in much greater abundance. G. bulloides decreases rapidly northward and is absent above the Davis Strait area. Three form variants of G. pachyderma were found in both plankton tows and sediment samples from the study area. These are shown, along with a topotype of G. bulloides, in Plate 1. All these forms of G. pachyderma are left coiling, which is the previously established cold water form. Rarely were any right-coiling forms encountered in either the tows or the bottom sediments, therefore no studies of coiling ratios were made.

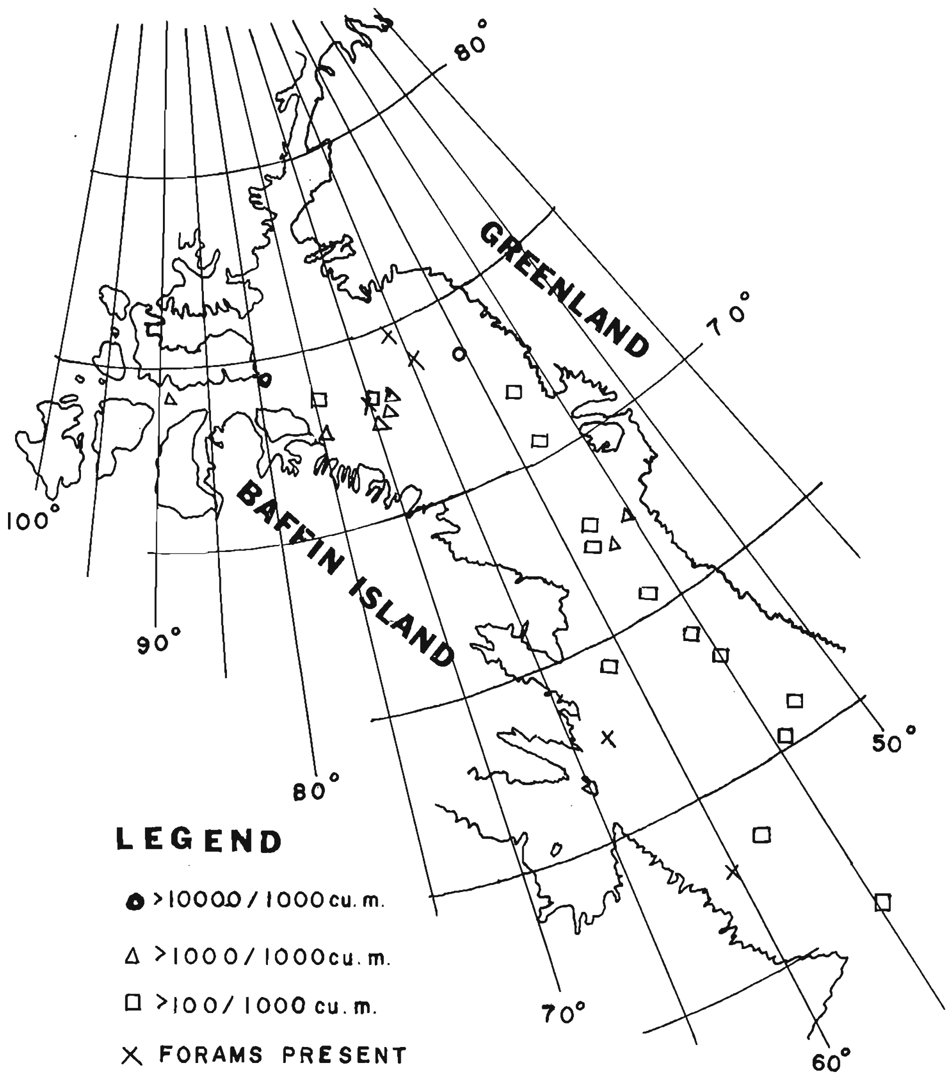

Figure 3. Density of planktonic foraminifera in the top 200 metres of water at stations in the study area.

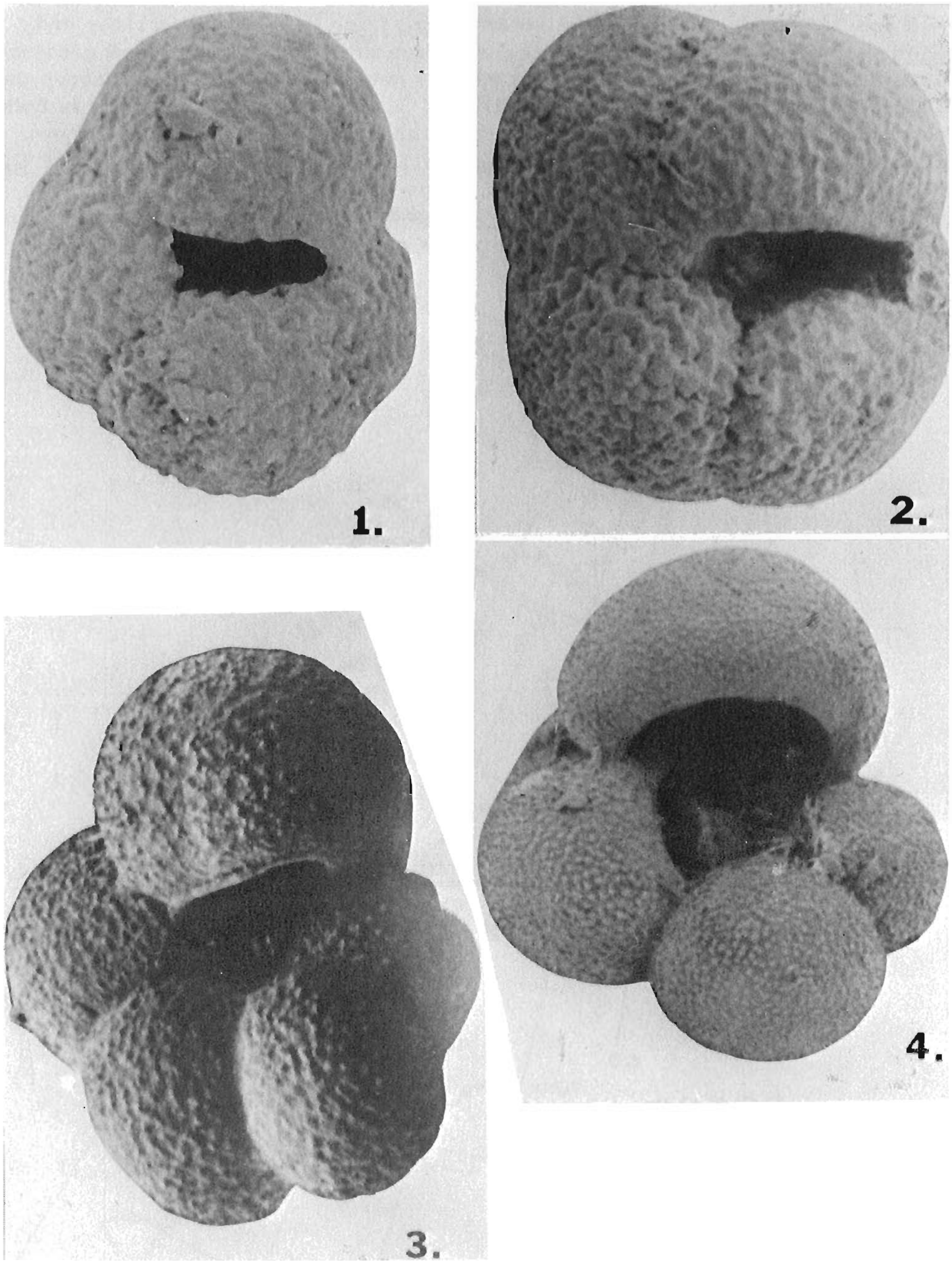

Plate 1.
1. Globigerina pachyderma form 1
2. Globigerina pachyderma form 2
3. Globigerina pachyderma form 3
4. Globigerina bulloides

STATION NO. / FORM	005	006	016	018	024	099	074	070	079	090	091
G. pachyderma  Form 1	40%	32%	35%	00%	14%	00%	08%	05%	12%	00%	02%
G. pachyderma  Form 2	20%	18%	31%	33%	53%	63%	08%	15%	46%	28%	22%
G. pachyderma  Form 3	00%	15%	08%	00%	19%	36%	83%	79%	41%	71%	75%
G. bulloides 	39%	34%	25%	66%	29%	00%	00%	00%	00%	00%	00%

INCREASING LATITUDE →

Figure 4. Species and subspecies distribution in per cent of whole planktonic foraminiferal fauna at selected stations in study area.

Kennet in 1968 described a intra-species variation of form in G. pachyderma which corresponds to changes in latitude in the South Pacific and Indian Oceans. A similar type of variation was found in the study area in both plankton tow samples and sediment samples. These three distinctive G. pachyderma variations are characterized as follows: a) variation 1 has a compactness with small aperture and heavy calcification which masks any pore structure at high magnifications; b) variation 2 has quadrate or square appear ance and heavy calcification; and c) variation 3 has five chambers on the dorsal side and much less coarse calcification. Variation 1 is found in highest concentrations in the southernmost part of the survey area and diminishes northward. Variation 2 is found to increase northward to the Davis Strait area, and then diminish farther to the north. Variation 3 is found in highest percentages in the northernmost part of the study area. Figure 4 contains quantitative information from several plankton tow stations of increasing latitude demonstrating this relationship with the form variation. It must be noted that this is not exactly the same form variation found by Kennet (1968) in the southern hemisphere. However, because this somewhat similar latitudinal variation does occur in both plankton tow samples and sediment samples, it does present a potential tool for paleo-oceanographic studies.

ACKNOWLEDGMENTS

The author wishes to thank Michael J. Keen, Chief Scientist and fellow researcher on the Dawson Cruise BI-70-028 for his cooperation in the sample collecting program. Thanks are also extended to Jim Johnson, and Robert Gerstein, who assisted in the collecting and analysis of the samples.

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(Earth Science Symposium on Offshore Eastern Canada,
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31. BATHYMETRIC OBSERVATIONS ALONG THE EAST COAST OF BAFFIN ISLAND: SUBMARINE MORAINES AND ICEBERG DISTRIBUTION

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Abstract

Detailed studies of the submarine trough which extends across the continental shelf off Clyde Inlet show that a terminal moraine lies across the outer part of the trough and that there is further evidence of four (4) other ice marginal positions, all seaward of the present shoreline. The oldest submarine feature is correlated with a supramarine moraine. The irregular topography of the trough differs sharply from the very smooth bottom topography on either side. The submarine moraine along the northwest side of the Clyde trough forms a barrier to the southward flowing icebergs which become stranded, thus accounting for the well-known field of icebergs characteristic of the Cape Christian area; which has an important stabilizing influence on the ice cover in the area. Studies of this naturally stabilized ice cover will be relevant to plans for artificially constructed ice islands in other areas.

INTRODUCTION

A Kelvin Hughes 26B echo sounder and a grab sampler were used to obtain most of the survey data on which this report is based. The speed of sound in water was assumed to be constant at 1463 m/sec. This instrumentation is very modest in comparison with that found on most current oceanographic platforms. Yet the bathymetric map (Fig. 1) shows many previously uncharted bathymetric features; in addition, several features depicted on the basis of old and incorrect data have been eliminated. The Canadian Hydrographic Service, furthermore, is preparing greatly improved charts of the area. These facts emphasize our lack of knowledge about large areas of the Canadian offshore. It is urgent that we fill these gaps now, when offshore oil and mineral exploration is developing at a rapid pace and when government agencies with regulatory responsibilities are being called upon to implement new or revised regulations in order to meet the public demand for environmental protection. International law and its application to the sea beds is also under careful consideration by international bodies.

Basic bathymetric information is also imperative in detailed studies of ocean circulation, for the shape of the ocean basins, i.e. the container in which the fluid moves, is clearly of significance. Furthermore, any interpretation of ocean sediment cores must consider the morphology of the ocean floor as an important parameter of the sedimentary environment.

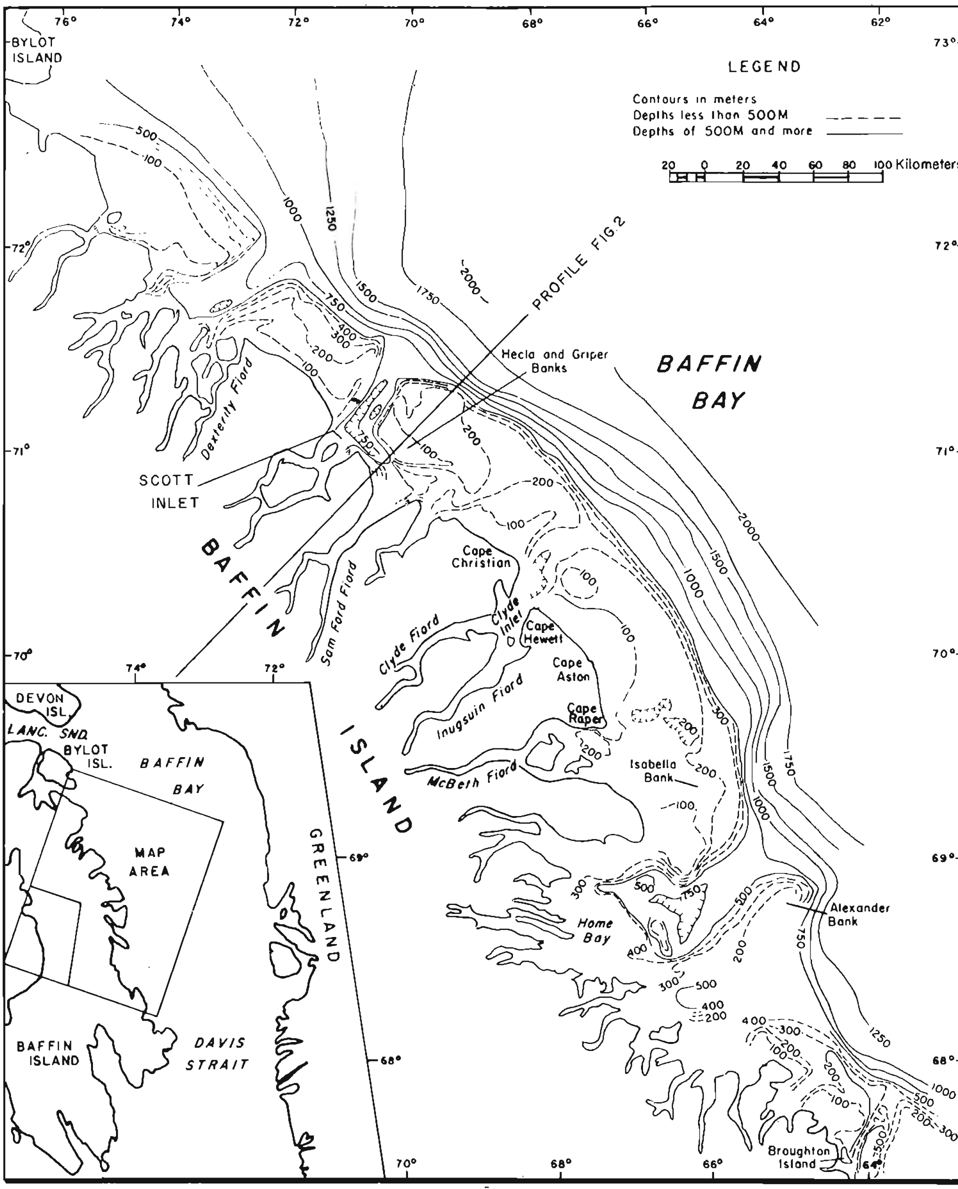

Figure 1: Bathymetry of the offshore area along the east coast of Baffin Island.

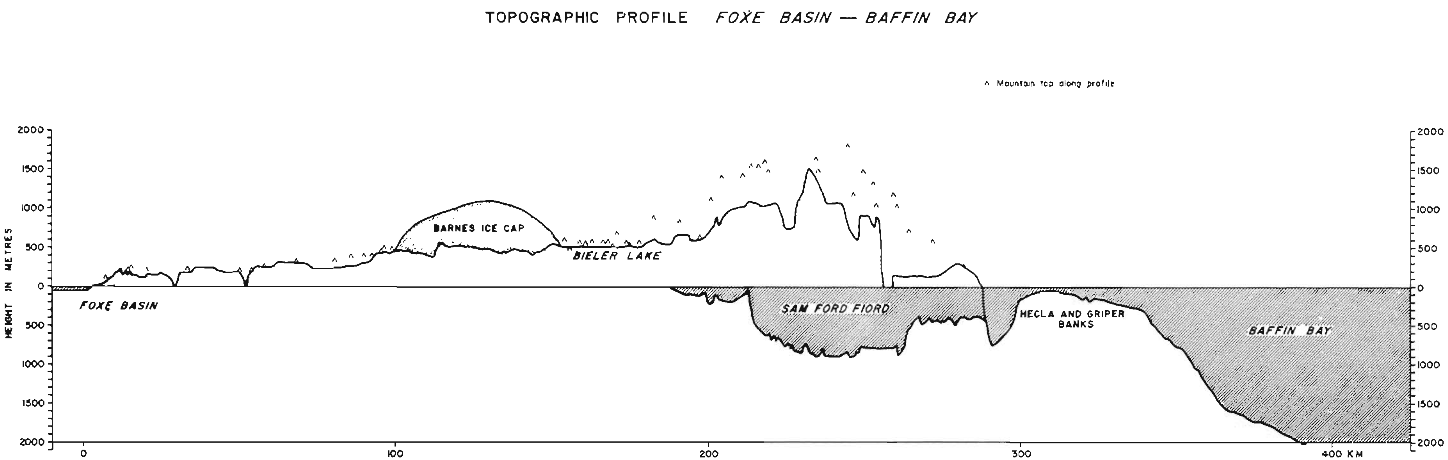

Figure 2.· Topographic profile across north central Baffin Island. See Figure 1 for position of profile. Inverted V's (Λ) show the positions of significant mountain tops within 20 km on either side of the profile line. Bottom profile under the Barnes Ice Cap was kindly supplied by Dr. J.R. Weber, Earth Physics Branch, Ottawa.

The main features of the eastern seaboard of Baffin Island are evident from Figure 1 (Løken and Hodgson, 1971); the continental shelf, which is generally between 30 and 50 km wide, reaches its greatest width in the Home Bay area and tapers off rapidly near Bylot Island to the northwest. The edge of the shelf occurs at varying depths, normally at about 200 m in this area; the continental slope usually has an inclination of less than 4 degrees, and it is generally gentle off Clyde Inlet, with steeper sections to the north and south, off Hecla and Griper and Isabella banks respectively; deep troughs extend seaward from the mouth of the major fiords and two of them (off Scott Inlet and in Home Bay) reach depths of more than 750 m.

The topographic profile in Figure 2 shows the location of the eastern seaboard in relation to Baffin Island in a typical transection across the central part of the Island. A few areas along the profile are covered by Pleistocene sediments, but otherwise the entire supramarine part of the profile is underlain by Precambrian rocks, with the exception of the extreme southwestern part where the Paleozoic formations of the Foxe Basin-Hudson area reach Baffin Island. Little is known of the lithology of the formations underlying the continental shelf, but seismic refraction studies carried out between Cape Raper and Cape Christian indicated the occurrence of thick low-velocity sediments (D. L. Barrett, personal communication). Interpretation of aeromagnetic surveys gives similar results (Hood and Bower, 1972).

The terrain of Baffin Island slopes gently to the west but falls off sharply to the east of the height-of-land (see Fig. 2). The east coast fiords extend inland of the high glacier-covered mountains along the crest of the island. No ready explanation can be given for the steep slope of the terrain along the east side of the island; there is no clear evidence of faulting nor is such evidence necessary because an oceanic crust underlies the central part of Baffin Bay (Barrett et al., 1971).

The position of the Barnes Ice Cap at a relatively low level and in the middle of the island is of particular interest (Løken and Sagar, 1969). A similar relation between the height-of-land and the centre of ice dispersal existed during the Wisconsin period when the major ice sheet was centred over the Hudson Bay-Foxe Basin area and large outlet glaciers flowed along the glacially-sculptured valleys to Baffin Bay, where they formed submarine moraines on the continental shelf (Løken and Hodgson, 1971). The areas north and south of Cape Christian were subjected to extensive deposition during the Pleistocene and the cliffs along the shore expose an exceptionally long time series of fossil-rich sediments (Løken, 1966).

Bathymetry off Clyde Inlet

The glacial troughs at the mouth of the major fiords and the associated submarine moraines were mentioned by Løken and Hodgson (1971). Figures 3 (a) and 3 (b) show the results of an echo-sounder survey of the trough and the adjacent parts of the shelf off Clyde Inlet. The trough is not very deep, with a maximum recorded depth of approximately 330 m; this represents a maximum overdeepening of about 260 m below the overall level of the adjacent shelf. The bathymetric profiles across the trough show that the feature is best defined near the shore where it is narrowest and its sides steepest.

The northwest side of the trough is relatively well-defined along its length but the boundary along the opposite side is somewhat diffuse (Fig. 3(a)). The northwest side is characterized by a sharp increase in depth which in

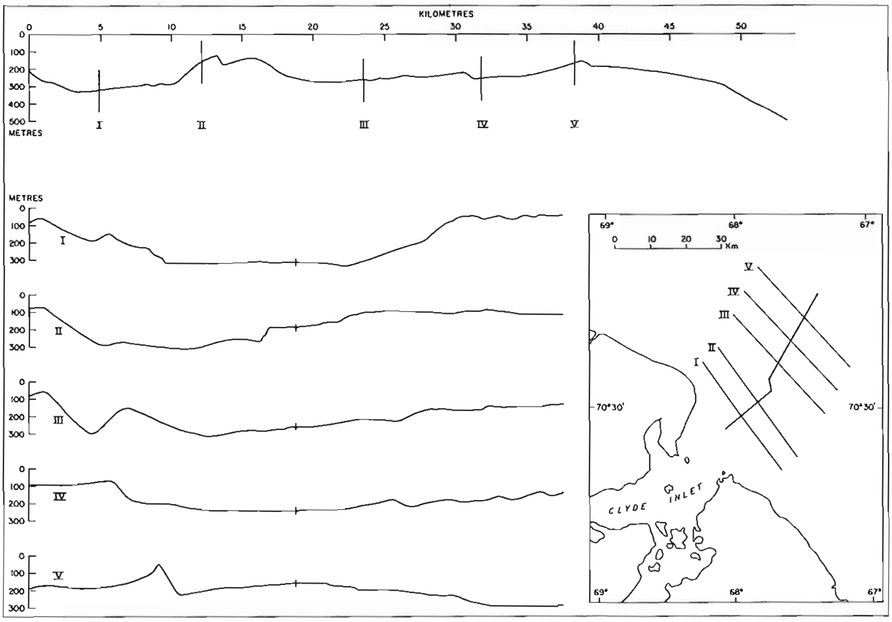

Figure 3 (a) Bathymetric profiles along (at top) and across the Clyde trough.

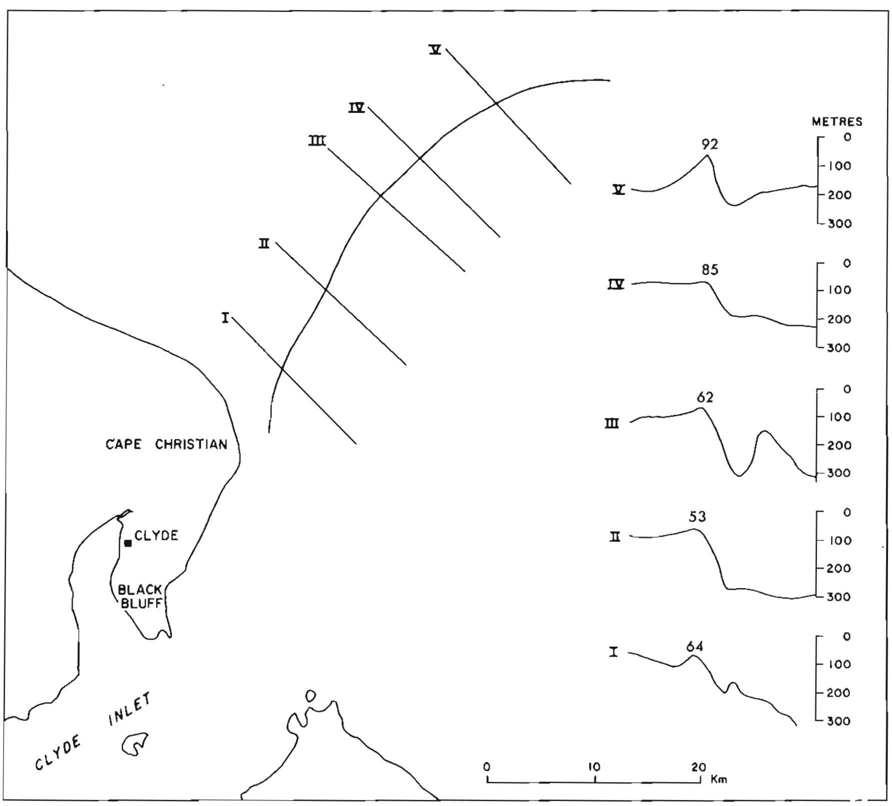

Figure 3 (b) Bathymetric profiles across the submarine moraine off Cape Christian, supramarine moraine is also shown. Numbers above profiles show minimum depth in metres.

some places is associated with a ridge-like feature. The northeasternmost profile is an exception because only a ridge is apparent. The ridge(s), as well as the sharp southeast-facing escarpment, are interpreted as being ice-marginal features; specifically the ridges were lateral moraines along the side of the ancient outlet glacier in Clyde Inlet.

This interpretation is in agreement with onshore observations, a broad ridge of till can be followed southward from near Cape Christian to the east side of Black Bluff. In the cliffs at Cape Christian a heavy, bouldery drift is exposed whose composition is in sharp contrast to the finer glaciofluvial material farther to the northwest (Løken, 1966; Feyling-Hanssen, 1967).

The ridges along the side of the trough are younger than the step on their distal side and there is thus clear evidence that there has been more than one glacial phase. This is also apparent from the profile along the trough which shows several ridge-like features (Fig. 3(a)), notably two seaward of profile II, one near profile IV, and a fourth near profile V. Four distinct periods of moraine formation may thus be inferred. However, the sharp ridge in profile V cannot have been formed contemporaneously with the ridge that occurs near its intersection with profile V in the along – strike profile. This is obvious from an examination of profile V which has a broad ridge in its central part. We must therefore conclude that the Clyde outlet glacier at some time had its terminus at least as far out as the edge of the continental shelf; furthermore there have been at least five periods during which moraines were formed on the continental shelf in this area. On the basis of the smooth form of the crest-line shown in Figure 3(b) the oldest of these periods occurred when the ice margin lay along the Black Bluff-Cape Christian line.

These conclusions are partly based on the results of an examination of grab samples collected on the north and the south side of the escarpment. To the north only sand and silt were found whereas small boulders, or nothing at all, were recovered from the area immediately south of the steep slope.

The fact that there were several periods of moraine formation, which may or may not have been separated by long intervals of time is confirmed by studies of the stratigraphy exposed in the 40 m cliff that extends northwestward for more than 20 km from Cape Christian (Løken, 1966; Feyling-Hanssen, 1967). Fossiliferous sands and clays, glaciofuvial deposits, and till-like material form the major formations, and Feyling-Hanssen recognized at least three glaciogene formations in the upper parts of the sections near Cape Christian.

The age of the submarine moraines is difficult to determine but postglacial in situ shell samples collected from the shores at the mouth of Inugsuin and Clyde fiords give maximum C-14 ages of around 8,000 years (Løken, 1966). The offshore moraines are obviously older than that. Several shell samples from the Cape Christian cliffs have been dated, by the C-14 method, but they yield indefinite ages e.g. >50,000 years (see Løken, in Andrews and Drapier, 1967); it is therefore clear that the moraines may be very old. Paleontological studies of the cliff section and of sediment cores collected from the main depressions in the trough should be of considerable interest in unravelling the glacial chronology.

The irregular topography shown in the southern part of the profiles discussed above is replaced by a smooth bottom topography farther south. This is particularly evident in the inshore profiles where one profile, some 30 km long between Cape Aston and Cape Hewett, shows a gradual shallowing from about 45 m at the south end to a minimum depth of 36 m, and then increasing depths to about

53 m at the north end. The profile extends between 5 and 10 km offshore and nowhere does depth change more than 3 m over any distance up to about 1 km. The smoothness probably reflects the gradual prograding of the coast between Cape Aston and Cape Hewett, which is shown by the shore flats and low beach ridges that are forming along this section of the coast. The beach flats indicate that the relative position of sea level is not changing substantially at the present time.

The longshore profiles near Cape Aston are all smooth and the bottom dips seaward at a rate of approximately 4:1000, based on a comparison between two parallel track lines. No profiles were run perpendicular to the coast in this region and no evidence of submerged beach ridges was found.

Near Cape Raper the topography is again more irregular as the influence of the McBeth outlet glacier becomes evident. Lateral moraines are found along the outer part of this fiord and the arcuate depressions between ridges that occur in the offshore area are interpreted as submarine continuations of these features. The McBeth trough itself is shallower and even less well-defined than the Clyde trough.

Iceberg Distribution

The shallow submarine moraine extending northeastward from Cape Christian will prevent the southward flow of large icebergs, which clearly explains the presence of a consistently large number of icebergs in this area. This iceberg concentration, although not the reason for it, has been known for a long time and is clearly demonstrated in Figure 4, which shows a section of a map of iceberg sightings (from Murray, 1969).

In interpreting Figure 4, it is important to note that it shows the number of sightings of icebergs but not the actual numbers of bergs. This is a significant distinction because the number of sightings depends on several factors, e.g., the number of bergs, the number of opportunities to observe the bergs, prevailing visibility, etc. However, in the area covered by Figure 4 the opportunities for observations are assumed to have been almost uniform and the figures are therefore a meaningful indication of the actual number of icebergs.

Grounded icebergs have been reported from many areas (Smith, 1928 and Arnold, personal communication), but the grounding areas have not been mapped and analyzed in detail in the context of iceberg drift and for use in iceberg forecasting for the eastern seaboard. This type of study will now become increasingly important as growing offshore exploration activity will make icebergs an important hazard to be considered, and not only a shipping problem as in the past.

The grounding areas are effective filters through which only icebergs below a certain size will pass, e.g. bergs with a draft of more than 92 m will not pass the moraine off Cape Christian. This limitation in size will in turn significantly influence the time an iceberg can survive in water of a certain temperature and hence determine the number of bergs that reach lower latitudes, e.g. the Newfoundland banks. Since the number of bergs that are trapped in a shallow area is dependent on the wind and current conditions, variations in these parameters in Baffin Bay will have important downstream influences, e.g. along the continental shelf of Labrador and Newfoundland. A closer examination of the changing meteorological and oceanographic conditions in Baffin Bay may show that these environmental factors and not the changes in

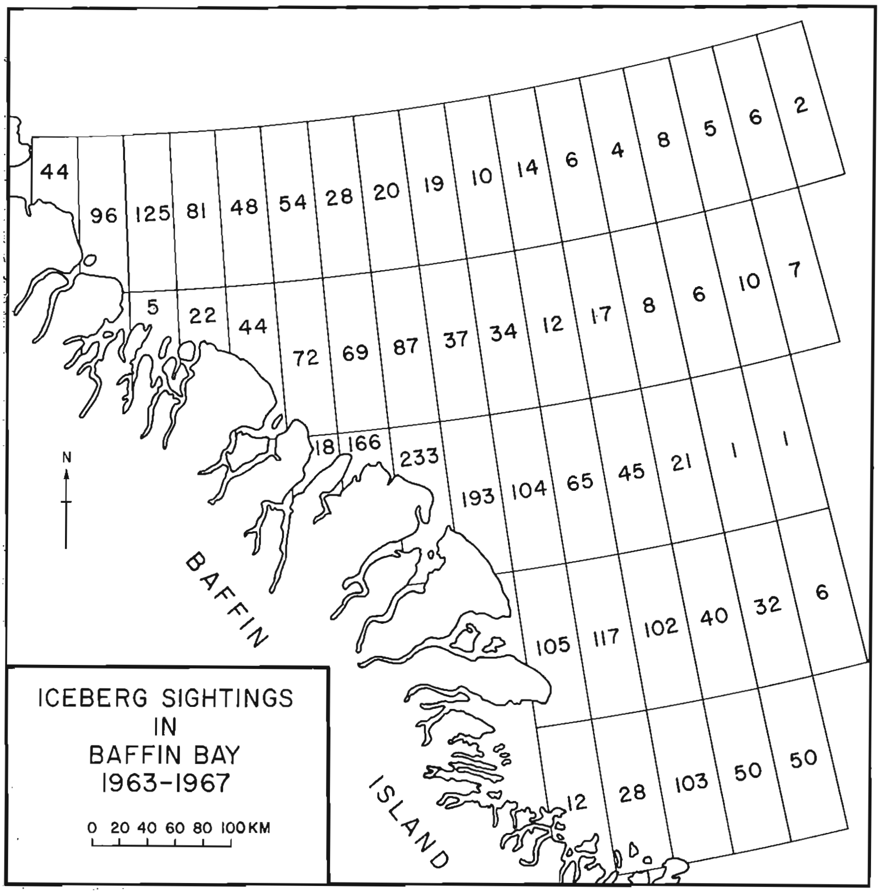

Figure 4. Iceberg sightings 1963-67 in area near Cape Christian (after Murray, 1969).

the rate of supply of icebergs from the calving glaciers are the main reasons for the observed variations in the number of icebergs reaching the Newfoundland banks.

Because great numbers of bergs are concentrated in a small area, a large number may be released and set adrift if storms and/or tidal effects create a particularly high water level. In this case an event equivalent to a major calving in the iceberg source area may take place in any one of many localities along the coast where no calving glacier exists.

It is therefore urged that greater attention should be given to the study of such grounding areas, and how the number of bergs change with time and with meteogological and oceanographic conditions. The Glaciology Division of the Inland Waters Directorate in cooperation with units of the Department of National Defence and other federal agencies has started such a study in the Cape Christian area in the expectation that it will result in better understanding of the variations in the southward flux of icebergs. Other grounding areas should be investigated as well. The grounding areas also provide excellent opportunities for the study of bottom scouring by icebergs, such as the depth and rate of scouring under different conditions.

While grounded on the banks, the icebergs will form pillars which in turn stabilize the sea-ice cover. Figure 5 shows a plot of the fast-ice distribution on a RCAF trimetrigon photograph. It can be seen that the fast-ice extends 5-10 km out from the shore and about the same distance downstream from the submarine moraine.

Sea-ice charts prepared by the Meteorological Service show that this distribution pattern is typical and that an ice cover normally exists north of Cape Christian well into the summer. In 1968, for example, 10/10 ice cover extended from the moraine to Cape Adair until August 22. Better knowledge of the areas where many bergs are normally aground may in this way be used in ice-forcasting routines.

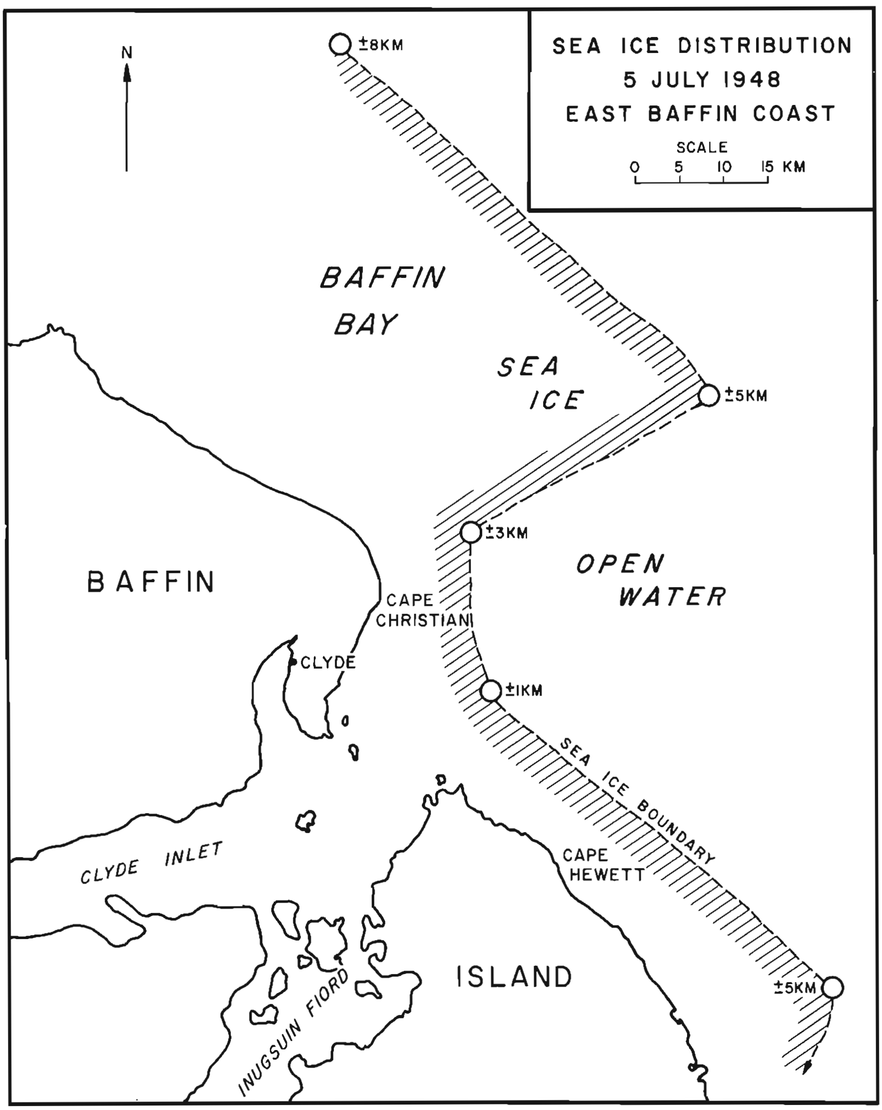

Figure 5. Sea ice distribution near Cape Christian on July 5, 1948.

This discussion of grounding areas for icebergs has important implications for all offshore exploration activities; it points out areas where icebergs may be very frequent, but also how one can take advantage of such areas, e. g. by operating on the shallow side of them. Due to melting, breaking and changing meteorological and oceanographic conditions the grounded icebergs are naturally unstable. However, the stability can be increased using the bergs as cores in artificial ice structures similar to those formed off Prudhoe Bay along the north shore of Alaska (Behlke and McKay, 1970). The cost of stabilizing an iceberg for a long time may be excessive, but to do it for the duration of a test-drilling operation on the leeward side of the iceberg may be very profitable.

In addition to providing protection for offshore drilling operations, the grounded icebergs may be stabilized in order to prevent ice sheets from breaking up. Normally one thinks of an ice sheet as something that ought to be broken up, and in areas of active shipping operations this is the most effective way to deal with it. However, outside the shipping lanes it may be better to preserve the ice sheet thus preventing the broken floes from drifting into the shipping lanes. If the ice sheet between some of the islands in the Arctic Archipelago was preserved, one would prevent the local ice as well as much heavy multi-year ice from the Arctic Ocean from reaching the Northwest Passage. A stable ice sheet in the shallow sounds of the archipelago would furthermore reduce fog formation that is frequently associated with open water, and it could have other climatological effects. Artificial ice structures formed around naturally occuring icebergs may become useful in this context.

CONCLUSIONS

A survey of the glacial trough off Clyde Inlet shows that the Clyde outlet glacier formed moraines in what is now the offshore area during at least five different periods of the Pleistocene. During its maximum extent the calving front of the outlet glacier was located at or seaward of the edge of the continental shelf.

Icebergs are grounded in several localities along the Baffin Bay coasts and the variations in the number of grounded bergs will influence the southward flux of icebergs. The meteorological and oceanographic conditions that cause frequent groundings should therefore be investigated. The grounded icebergs stabilize the sea ice sheet in the grounding areas and they may provide natural protection for offshore operations. Artificial stabilization of the iceberg will further enhance these effects and this may be used in large scale stabilization of the sheet between the Arctic Islands.

ACKNOWLEDGMENTS

The Marine Sciences Directorate were helpful in obtaining ship time for the project on the icebreakers D'IBERVILLE and LABRADOR, and the captains and crews of these vessels gave full cooperation during the surveys. The Canadian Hydrographic Survey compiled the field survey data and Mrs. Lyn Arsenault assisted in preparing the illustrations. I wish to express my gratitude to these individuals and organizations for their help.

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32. THE GEOLOGICAL SETTING OF THE WEST GREENLAND BASIN IN THE BAFFIN BAY REGION

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Abstract

The sediments and lavas of Cretaceous-Tertiary age exposed onshore in the peninsulas and islands of central West Greenland form only part of a more extensive sediment/lava pile that is largely concealed beneath the waters of Baffin Bay, Davis Strait and the Labrador Sea. The well-exposed and deeply dissected rocks preserved onshore in West Greenland between the Svartenhuk area in the north and Grønne Ejland in the south are considered to be of prime importance for the interpretation of the offshore area.

Geophysical work by Canadian institutes in the Melville Bugt area has shown the presence of a wide, coast-parallel graben containing about 6,000 m of sediments. Geophysical work and bottom sampling undertaken by Canadian institutes in the area offshore from central West Greenland have shown the approximate extension of the basalts in the offshore area.

A continuation of the fault system associated with the Melville Bugt graben occurs on land in the Thule-Dundas area of western North Greenland, and it is suggested that an equivalent to this system may exist in the Lancaster Sound - Jones Sound area. It is suggested that the sediments of the West Greenland basin have been deposited in a continuation of this graben that terminates southwards as a fault-bounded embayment at the southern end of Disko Bugt, while a second graben or fault is developed on the west side of Disko. The embayment is considered to have remained static since the Danian, the continuation of the Melville Bugt graben and the graben or fault on the west side of Disko then being linked by a subsidiary graben that developed between Ubekendt Ejland and Disko. The very large thicknesses of basalt in the western part of the onshore area probably accumulated in an almost continuously subsiding graben, and it is suggested that the graben itself was the source of the basalts.

There is evidence that the sediments offshore from West Greenland could range in age down to the Jurassic.

INTRODUCTION

The area between 68°50'N and 72°40'N in central West Greenland is the only part of West Greenland where sedimentary and volcanic rocks of Cretaceous to Tertiary age are present on land. Recently published geophysical work offshore from West Greenland has shown that sedimentary and volcanic rocks occur extensively below the sea floor off West Greenland, and these are very thick in some areas. They must, at least in part, be equivalent in age to the rocks found on land within the West Greenland basin.

In view of the unique opportunity afforded by the presence of these rocks on land it is clearly vital that the rocks of this eminently accessible area should be examined in considerable detail. Much work has already been done in the area, but much remains to be done.

The present paper is an attempt to interpret certain aspects of the geology of the West Greenland basin in terms of what is already published on the offshore area, particularly that part of the offshore area situated north of 68°N.

Previous Work in the West Greenland Basin

A review of Cretaceous-Tertiary stratigraphy and tectonics in this part of Greenland has been published by Rosenkrantz and Pulvertaft (1969). The marine Upper Cretaceous and lowermost Tertiary deposits are described by Rosenkrantz (1970), while a review of fossil floras and nonmarine deposits in this area is provided by B. E. Koch (1964).

The Tertiary basalts of the Svartenhuk peninsula are described by Noe-Nygaard (1942), Munck (in Rosenkrantz et al., 1942) and by Pulvertaft and Clarke (1966). Those of Ubekendt Ejland are described by Drever (1958). A preliminary account of the basalts of northern Disko is given by Pedersen (1969). Clarke (1970) discusses the chemistry and petrogenesis of the Tertiary basalts of Baffin Bay (including those of Svartenhuk). A paper on the West Greenland basalts is being prepared by V. Münther of the Geological Survey of Denmark (manuscript in course of being translated).

The petroleum prospects of the area are discussed by Henderson (1969) and by Henderson and Stevens (1969).

Summary of Stratigraphy of the West Greenland Basin

The West Greenland basin (see Figs. 1 and 2) contains a thick sequence of marine and non-marine shales and sandstones, which were deposited during the Cretaceous and earliest Tertiary (Danian). The marine sediments reach a maximum of 2,000 m in thickness. The oldest marine sediments are of probable Cenomanian age (about 100 m.y. old[1]) and are found in Svartenhuk. The thickest sections of marine sediments exposed in any one area (total thickness about 1,450 m) occur along the north coast of the Nûgssuaq peninsula. The nonmarine sediments reach about 1,500 m in thickness. The oldest nonmarine sediments are of Lower Cretaceous (Barremian-Aptian) age. These occur on the north coast of Nûgssuaq, east of the marine sediments mentioned above. The sediments of the southernmost part of Nûgssuaq and of east and northeast of Disko are nonmarine.

The sediments are overlain by a thick pile of Tertiary basalts. The lower basalts are usually developed as sub-aquatic pillow breccias of mainly picritic composition. These are overlain by picritic subaerial lavas, which grade upwards into olivine-poor and olivine-free feldspar-phyric basalts. The subaerial basalts are up to 8 km thick in the western part of the area.

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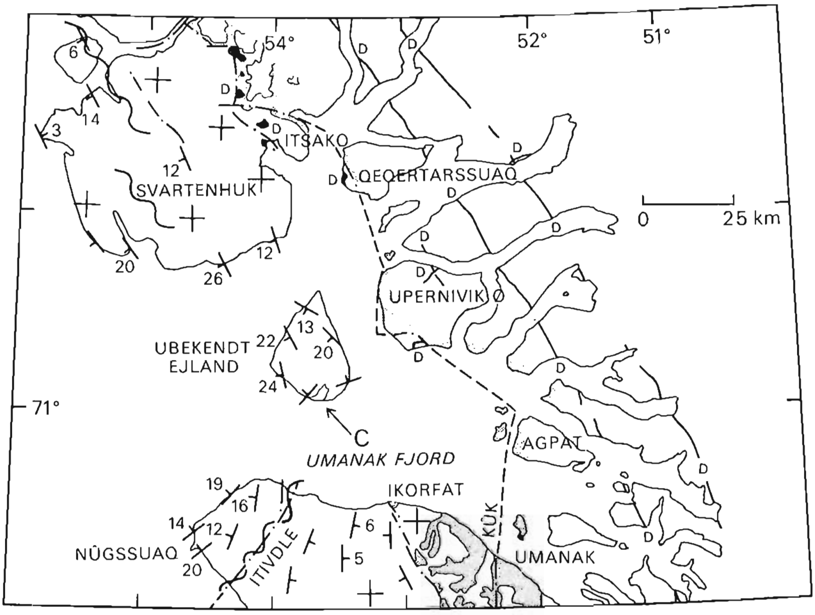
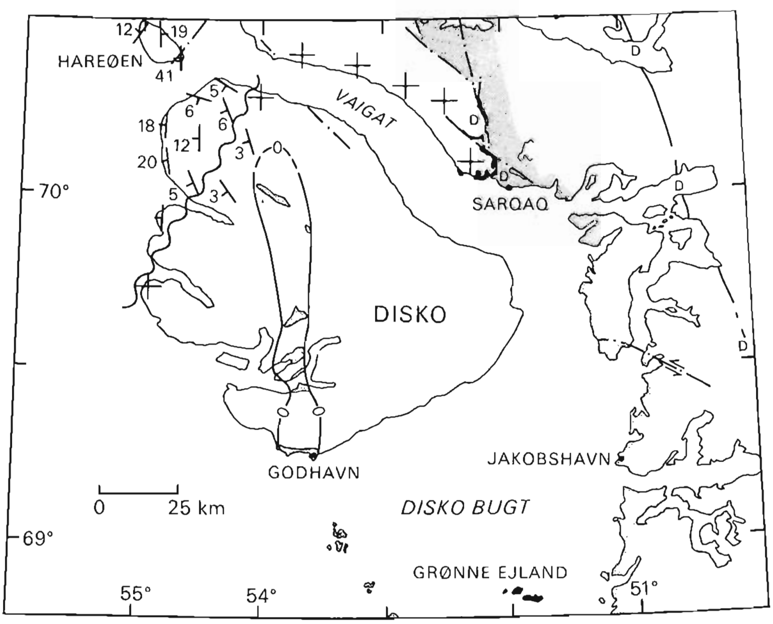
[1] Ages used for stratigraphic units in this paper are based on the Geological Society Phanerozoic time-scale, 1964 (Quart. J. Geol. Soc. London, v. 120S, p. 260-262).





Black areas are dolerite sills and sheets. D = dolerite dyke.
Unshaded areas to the west are sediments and basalts.
The border of the Precambrian rocks is shown in grey.
Wavy lines show lines of flexure in basalts. Broken lines are faults.
Dips and strikes are from the basalts (+ = horizontal to subhorisontal).

Figures 1 and 2. Simplified geological maps of the northern and southern parts of the West Greenland basin.

Structure of the West Greenland Basin

This paper concentrates on the structure of the West Greenland basin and the geology is therefore shown in greatly simplified form in Figures 1 and 2. The sedimentary and volcanic rocks are shown together. The main structural elements are shown, the strikes and dips plotted being from the basalts.

Boundary Fault System

Rosenkrantz and Pulvertaft (1969) have already described in considerable detail the boundary fault system that limits the Cretaceous sedimentary basin to the east. This system probably determined the limits of most of the Tertiary sediments in the southern part of the area as well. However, in the peninsula immediately north of Svartenhuk nonmarine sediments of presumed Tertiary age are found northeast of the main fault system. Likewise, in the fjord east of the peninsula, nonmarine sediments of presumed Tertiary age extend for some distance inland, filling pockets in the Precambrian surface. There is no evidence that they are limited by faults (T. C. R. Pulvertaft, pers. comm.). The Tertiary basalts are however found overlying the Precambrian rocks over considerable areas northeast and east of the fault system in the northern part of the area (see Fig. 3) and east of the Kûk – Sarqaq fault system in Nûgssuaq. Many of the abundant sheets and sills of mafic picrite, olivine dolerite and quartz dolerite found in the sedimentary strata and particularly along the Cretaceous-Precambrian boundary (Rosenkrantz and Pulvertaft, 1969, p. 892) are clearly spatially related to the boundary fault system. Rosenkrantz and Pulvertaft (1969, p. 892) also draw attention to the two olivine dolerite master dykes that trend NW-SE and cut the Precambrian rocks east of the basin. A prominent dyke parallel to those mentioned was mapped by Henderson and Pulvertaft (1967) in the northern part of Upernivik Ø and the western end of the peninsula to the north. This dyke and the easternmost of the two master dykes referred to are cut by NE-SW-trending dykes.

The prominent N-S-trending gneiss ridge in Disko is overlain and flanked by basalts. Although it is believed that this ridge is of tectonic origin (i.e. an old horst), there does not seem to have been post-basalt movement along its immediate margins where these are exposed on land. Field evidence shows that the ridge was in existence when the basalts were laid down. No faults have been seen along the eastern margin of the ridge and none have been seen along the immediate western margin either, but V. Münther (pers. comm.) has seen one fault affecting basalts 1 - 2 km west of the western margin. The small islands to the south-southeast of Godhavn are composed of Precambrian rocks and are considered to be a continuation of the Disko ridge.

Pillow Breccias

The pillow breccias that are normally found at the base of the lava pile show large-scale crossbedding. No detailed work has yet been done to determine the direction of provenance of the basalts forming these breccias. The crossbedding is often very large-scale and best seen from a distance, so that the strike of the foresets is not easy to determine. Nevertheless, there are many areas where the approximate direction of provenance is

Figure 3. Basalts resting on Precambrian rocks, inner part of Svartenhuk. Note the irregularity of the pre-basalt topography. (Reproduced by permission (A. 158/71) of the Geodetic Institute, Copenhagen).

known. Figure 4 shows part of an aerial oblique view of a section of the south coast of Nûgssuaq, some 75 km northwest of Sarqaq. Along this part of the coast the pillow-breccia foresets show a substantial dip component to the southeast, indicating that the source lay somewhere in the area westwards from the outcrops of the breccias. It is important to note that this does not necessarily mean that the source was offshore from the western limit of the basalt area, since breccias are not exposed west of the Itivdle valley. In Svartenhuk, the pillow-breccia foresets also dip mainly inland (T.C.R. Pulvertaft, pers. comm.).

Examples have been seen at two localities of crossbedded pillow breccias showing two directions of dip of the foresets. Figure 5 shows one of these occurrences. The photograph was taken close to the Ikorfat fault system, on its southwest side (see Fig. 1). This fault system, when viewed in detail (see Fig. 6), can be seen to consist essentially of a number of en echelon NW-SE faults, with downthrow to the southwest. The system can be

Figure 4. Part of an oblique aerial photo showing the succession on the south coast of Nugssuaq 75 km northwest of Sarqaq. Immediately above sea level are Cretaceous sediments, largely covered by talus and solifluction deposits. Note the landslip on the right. The sediments are overlain by crossbedded pillow breccias whose foresets dip to the right. The top of the section comprises subaerial basalts. (Reproduced by permission (A. 158/71) of the Geodetic Institute, Copenhagen).

Figure 5. Thick crossbedded pillow breccias with a cap of subaerial basalts overlying Cretaceous-Tertiary sediments southwest of the Ikorfat fault. Note the two directions of crossbedding, with the foresets dipping left and right respectively (GH.08.05.1968).

followed from the north coast of Nûgssuaq into the interior of the peninsula over a distance of at least 30 km. It is a major fault system and is described by Rosenkrantz and Pulvertaft (1969, p. 893) and Henderson (1969, p. 23). The hill face seen on Figure 5 trends NE-SW. The strikes of the foresets are not known, but pronounced dip components to the northeast and southwest are clear. The writer would suggest very tentatively two sources for the breccias in this hill, one to the westwards, possibly that which gave rise to the breccias of the south coast of the peninsula and the other to the eastwards, possibly the Ikorfat fault system. A second example of cross-bedded pillow breccia showing two directions of dip of the foresets was found close to the fault system at the margin of the sedimentary basin in Svartenhuk by T.C.R. Pulvertaft (pers. comm.). It seems significant that both occurrences are found close to important faults.

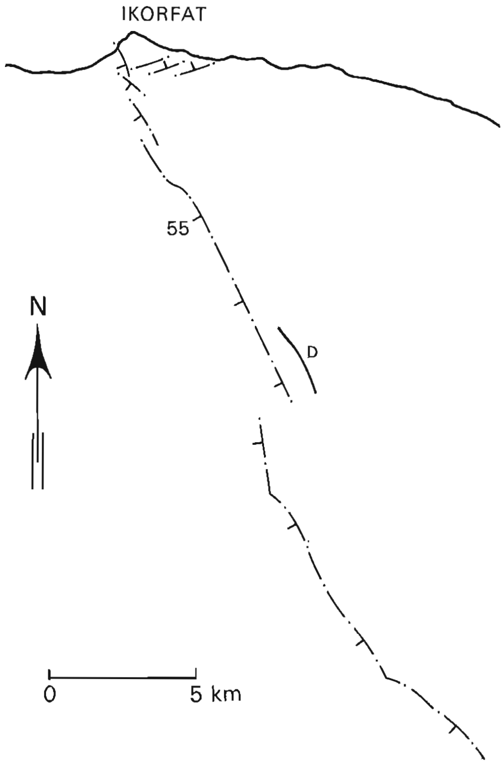


Figure 6. Details of the Ikorfat fault system.

Downwarping of Basalts

The zone of downwarping of the basalts is shown on Figures 1 and 2. The structural picture in Svartenhuk is complicated (V. Münther, in prep.). Along the northwest coast of the peninsula and in the island off the coast the attitude of the basalts changes rapidly across a northwest-trending flexure. Northeast of the line the basalts are subhorizontal to horizontal. Southwest of the line the basalts dip southwest at angles of up to 14°. At the western end of this coast section the basalts dip gently inland. Along the south coast there is a steepening of the dips of the basalts towards the west, and there is no distinct line of flexure. In part of southern Svartenhuk, if the strikes of the southwest-dipping basalts are projected inland, they encounter an area of subhorizontal to horizontal basalts.

The basalts of Ubekendt Ejland form a syncline with axis plunging westwards. The strike is NW-SE along the coast facing Svartenhuk, but changes to NE-SW along the coast opposite outer Nûgssuaq. This is structurally a key area and links Svartenhuk and Nûgssuaq.

The structure of the basalts of outer Nûgssuaq can be seen on a photogeological interpretation of this area (Fig. 7). The blank area extending NE-SW across the peninsula is the Itivdle valley. It contains Cretaceous-Tertiary sediments in its southwestern and central parts, and at the northeastern end. There is much Quaternary cover in the valley. The faults shown northwest of the valley are only part of an extensive fault system that was subsequently mapped in more detail in the field. The cumulative effect of this system has been to downfault the basalts to the northwest. They are also tilted and dip northwest to west. Southeast of the fault there are extensive areas of pillow breccia, which are not shown on the interpretation. The few traces shown are of basalts dipping gently east to southeast. Dykes are abundant southeast of the fault and only a few of these are shown to illustrate the two important trends. One system is parallel to the Itivdle valley and the other is normal to it or parallel to the length of the Nûgssuaq peninsula.

The downfaulting has been accompanied by the development of a system of antithetic faults, which are very clearly displayed along the outermost part of the north coast of Nûgssuaq. Figure 8 shows a sketch of part of the north coast some 10 km west of the Itivdle valley drawn by V. Münther. However, although antithetic faulting is important in Nûgssuaq, there is no reason

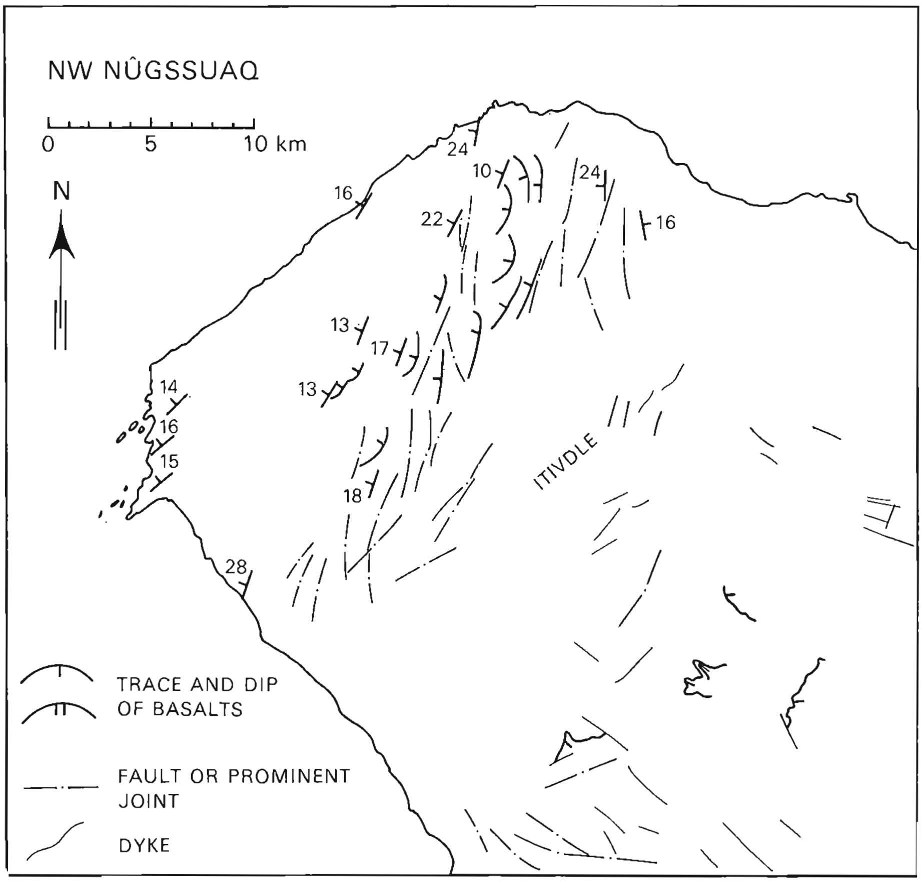


Figure 7. Photogeological interpretation of northwest Nûgssuaq.

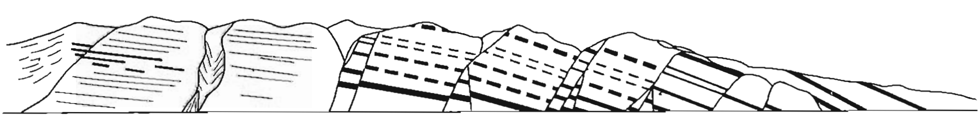


Figure 8. Antithetic faults cutting basalts on the north coast of Nûgssuaq, 10 km west of the Itivdle valley. (From sketch by V. Münther.)

to consider that its presence can be used as an argument for making large reductions in the calculated total thickness of the basalts in the western part of the area as a whole. The total thickness of the subaerial basalts in the western part is given as about 8 km by Rosenkrantz and Pulvertaft (1969) while in an earlier paper, Noe-Nygaard (1942, p. 71) estimated a total thickness of about 10 km for the basalts of Svartenhuk.

Although the thickness estimates given are quite realistic for the western part of the area under discussion, this paper stresses that these large thicknesses are local, and confined to a marginal zone along the coast and inner part of the continental shelf in this part of central West Greenland.

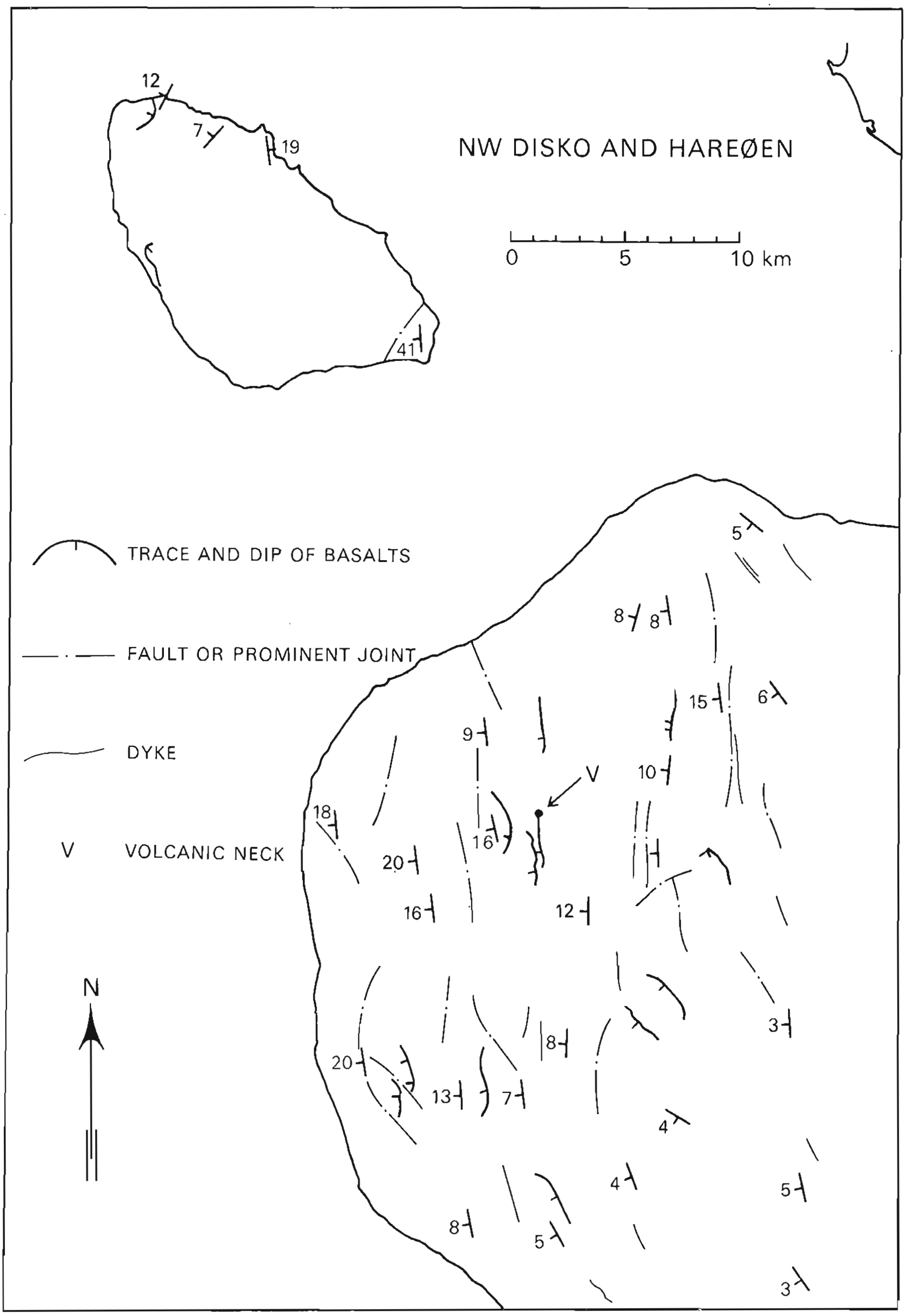


Figure 9. Photogeological interpretation of northwest Disko and Hareøen.

The structure of the basalts of Hareøen and northwest Disko is shown in detail in Figure 9. In Hareøen, there is a continuation of the Itivdle fault zone of Nûgssuaq. Southeast of the fault in Hareøen the basalts dip obliquely towards the fault at about 40°. Northwest of the fault there is an anticline with its axis parallel to the fault. This situation is in contrast to that found on the coast of Nûgssuaq opposite, where the basalts dip northwest between the fault and the end of the peninsula (see Fig. 7).

In Disko the axis of downwarping can be seen in Figure 2 to trend NE-SW across the northwest end of the island. This axis does not coincide with the Itivdle fault zone of Nûgssuaq, but is displaced relative to the Itivdle fault zone some 25 km to the southeast. In the detailed interpretation of northwest Disko shown in Figure 9 it can be seen that the dominant strike of the basalts is N-S, and the dip is westwards, i.e. the strikes are oblique to the axis of downwarping. Furthermore, the figure illustrates the way the dip steepens from south to north. This may be related to a system of N-S en echelon faults northwest of the axis of downwarping. The pattern resembles the structural pattern of central and southern Svartenhuk, where the dip of the basalts decreases inland along the strike.

Structural Interpretation of the West Greenland Basin

It is clear from the previous sections of this paper that the structure of the Cretaceous-Tertiary rocks of onshore West Greenland is very complicated. However, it is possible that the structure may be explained in a simpler manner than was previously thought possible. Recent geophysical work offshore, although only of a reconnaissance nature, has provided evidence about the constitution and structure of the rocks below the sea floor that suggests, albeit in very general terms, a possible explanation for the structure and sedimentary/volcanic history of the onshore area.

Melville Bugt

In Figure 10 a very generalized picture is given of the area between Thule and Disko. Aeromagnetic surveys undertaken by the Geological Survey of Canada in 1967 (Hood et al., 1968, p. 79) showed the presence of a substantial thickness of sediments underlying Melville Bugt (Melville Bay). In recent papers by Hood and Bower (1970; 1972) this area is described in more detail. The authors mention the presence below the Greenland shelf of a deep sediment-filled graben whose width is approximately 35 miles (56 km) with its eastern margin located about 35 miles from the coast of West Greenland. The graben is filled with about 20,000 feet (6 km) of sediment.

During the years 1963 to 1966 extensive marine geophysical surveys were undertaken in Baffin Bay and the Davis Strait by the Bedford Institute, Dartmouth, Nova Scotia. The graben is very clear on marine magnetometer profiles across Melville Bugt, and the presence of a deep sedimentary basin was mentioned at a meeting of the American Geophysical Union in 1969. (For abstract see Barrett and Manchester, 1969.) The results of the earlier surveys and of more recent work undertaken by the Institute in collaboration with Dalhousie University are referred to by Manchester and Clarke (in press), who discuss the graben in some detail. The outlines of the graben as shown on Figure 10 were plotted by the present writer on the basis of the very prominent negative magnetic anomalies in Melville Bugt, using a copy of the profiles presented to the Geological Survey of Greenland.

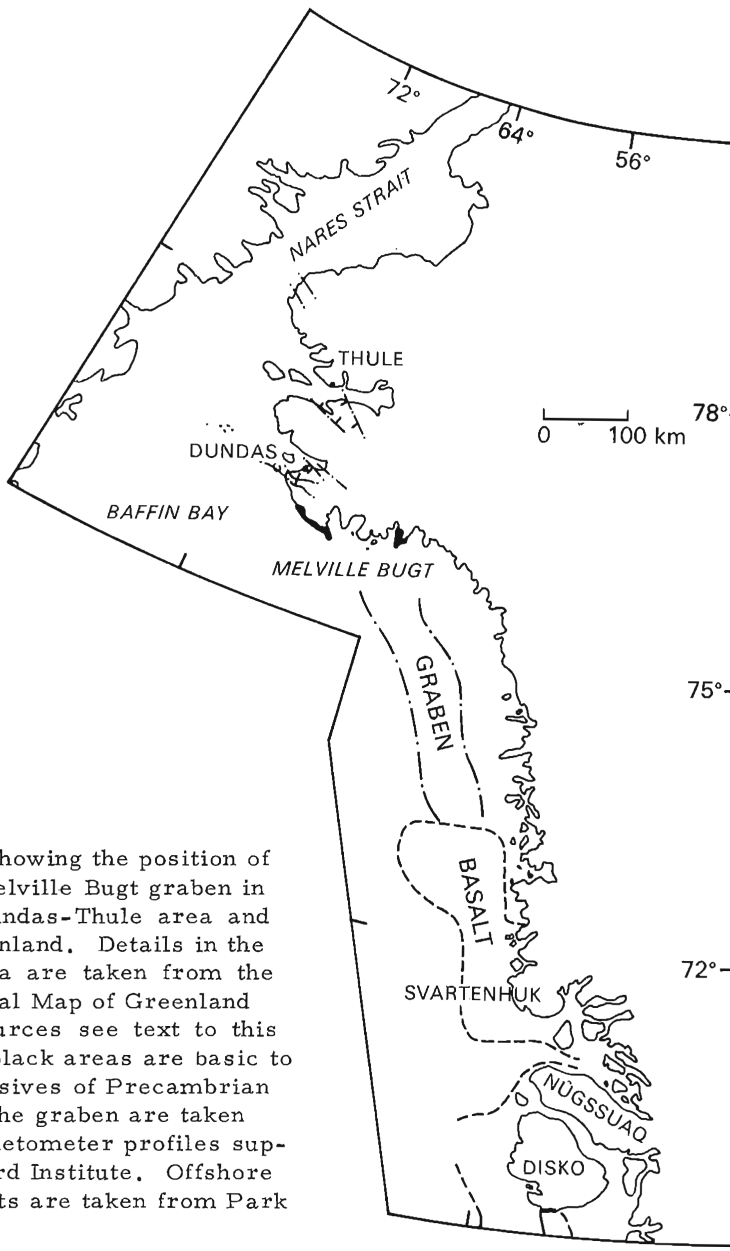


Figure 10. Map showing the position of the Melville Bugt graben in relation to the Dundas-Thule area and central West Greenland. Details in the Dundas-Thule area are taken from the Tectonic/Geological Map of Greenland (for details of sources see text to this paper). The two black areas are basic to intermediate intrusives of Precambrian age. Outlines of the graben are taken from marine magnetometer profiles supplied by the Bedford Institute. Offshore limits of the basalts are taken from Park et al. (1971).

Thule-Dundas Area

The Melville Bugt graben is in direct alignment with an area on land to the northwest which contains Eocambrian rocks of the Thule Group and older Precambrian rocks (see Davies et al., 1963). The area is cut by numerous normal faults that let down the younger rocks into the older rocks. One fault described by Davies et al. (1963, p. 46) has had a vertical movement of at least 7,000 feet (2,100 m). Numerous diabase dykes and sills cut the older and younger rocks.

The details shown on Figure 10 in this part of onshore Greenland were compiled from the new Tectonic/Geological Map of Greenland, published by the Geological Survey of Greenland, and occupy an area larger than that described by Davies et al. (1963). The two black areas are shown on this map as basic to intermediate intrusives of Nagssugtoqidian age (1,790 - 1,650 m.y.). The presence of Precambrian intrusive rocks in these two areas was originally noted by L. Koch (1920). The western occurrence is described by Davies et al. (1963, p. 40-41) as a hypersthene-bearing quartz diorite. These bodies are shown on Figure 10 because of their situation along the trend of the graben margins. If their location has any significance in terms of the graben, they could indicate old lines of weakness in the area that now shows graben-type faulting. In this connection it could be noted that Fahrig et al. (1971) suggest that Baffin Bay and Davis Strait may have started to form as early as the late Hadrynian (latest Proterozoic) and may contain Paleozoic strata.

The fault system in the Thule-Dundas area was originally discussed by L. Koch (1926). He considered that faulting in this area occurred simultaneously with faulting in the Disko Bugt/Umanak Fjord area, and that the age of the movements was Tertiary.

A fossiliferous siltstone sample allegedly from the Parker Snow Bugt area 35 km south of Dundas (see Fig. 10) was submitted many years ago to the University of Copenhagen and proved to contain an Upper Triassic (Norian) assemblage. The area has since been visited by numerous geologists, but none of these have found Upper Triassic rocks in this area (A. Rosenkrantz, pers. comm.).

Kerr (1967, p. 509) states that the graben at Thule Air Base (close to Dundas on Fig. 10) is in alignment with a deep submarine canyon that trends into Melville Bugt, and suggests that the faults may extend under the bay. He also cites the mapping of Kurtz and Wales in the Thule area where 2,600 m of Proterozoic-type sediments are preserved in a graben, the vertical movement having thus been at least 2,600 m. The mean trend of the faults in the Dundas area is 110° as opposed to the 130° shown on Kerr's tectonic map of the Nares Strait region (1967, Fig. 6). Nevertheless, recent geophysical work in Melville Bugt confirms the suggestion put forward by Kerr about the probable extension of faults under the bay. The graben as shown on Figure 10 is probably over-simplified, but the movements associated with its formation have had a profound effect on the rocks of this corner of Greenland. There is a strong suggestion that the fault system bordering the graben offshore splays out on land.

Possible Structural Correlation across Baffin Bay

Kerr's map (1967, Fig. 6) shows that the structure of the Nares Strait area is complex. Dawes and Soper (in press) conclude that the structural history of the Nares Strait linear feature is long and complex. During discussions on this area in Ottawa the present writer suggested that a pre-drift extension of the Melville Bugt-Thule fault system should be sought on the Canadian side of Baffin Bay. If found, this would provide an important key to the interpretation of the Nares Strait linear feature. D. L. Barrett of the Bedford Institute tentatively suggested that the fault system in Lancaster Sound could be such a pre-drift extension of the fault system on the Greenland side.

If the outermost faults in the Dundas-Thule area shown on Figure 10 are extended to meet a line down the middle of Nares Strait, the distance between the two intercepts is about 220 km. This is the distance between the centre of Lancaster Sound and the centre of Jones Sound. The outermost faults diverge to the northwest and this divergence could reasonably be expected to have continued for some distance before displacement occurred. Thus a correlation between the fault system on the Greenland side and the zone between the southern coast of Lancaster Sound and the northern coast of Jones Sound is possible. This paper urges that this possibility be investigated.

Such a correlation would be broadly in accordance with the fit of North America and Greenland proposed by Bullard et al. (1965, Fig. 7) in the sense that this fit shows the Dundas-Thule area opposite the Lancaster Sound-Jones Sound area.

Central West Greenland

The Melville Bugt graben and its continuation on land in western North Greenland is clearly a very major structural feature. It is therefore natural to look for an expression of the fault system to the south, i.e. towards central West Greenland. Figure 10 shows the region of offshore volcanics as given by Park et al. (1971), excluding the area at the southern end of Disko Bugt. The graben is in perfect alignment with the northern extension of the offshore basalts. The remaining part of this paper will be devoted to a proposed interpretation of the structural development of this part of West Greenland in the light of this alignment.

Proposed Model for Central West Greenland and the Offshore Area

In Figure 11 the writer offers his present ideas about the structure of central West Greenland and the offshore area. The model proposed is highly speculative, and will remain so pending more detailed work onshore and offshore. Nevertheless, there is a growing amount of evidence that an explanation along these lines will be acceptable.

It is postulated that the West Greenland basin owes its existence essentially to interference between two grabens, one extending from Melville Bugt across the West Greenland basin to the southern part of Disko Bugt, and the other extending from west of Disko southwards. In the Thule-Dundas area of western North Greenland (Fig. 10) the movements have resulted in the development of grabens with a mean trend of about 120°. The large graben of

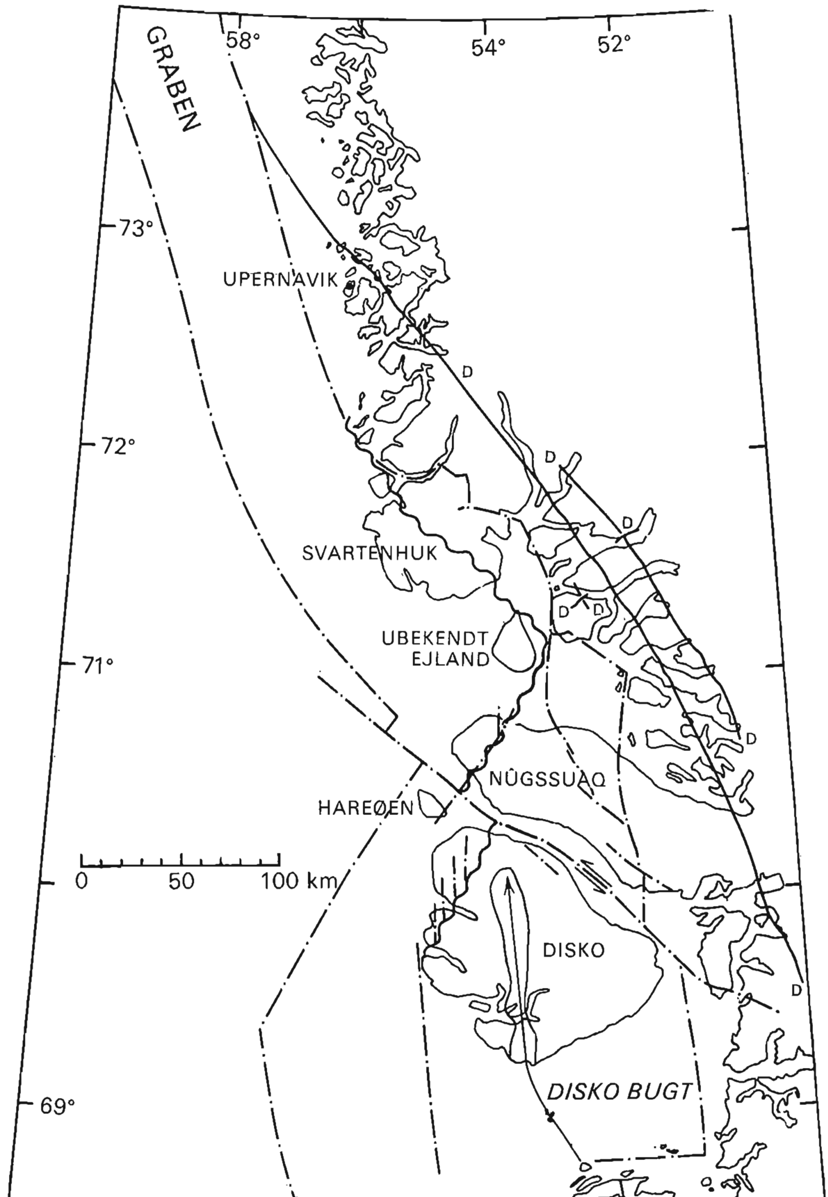


Figure 11. Proposed model for central West Greenland and the offshore area. The graben margins and other faults are shown with broken lines. Wavy lines show the axis of downwarping of the basalts and are generalized. The crest of the Disko ridge and its northwards plunge are shown by a line with an arrow. D = dolerite dykes, the longest being extrapolated out to sea.

Melville Bugt, which is a continuation of this fault system, has a trend of about 150°. The graben continues across the West Greenland basin, and ends in the southern part of Disko Bugt, where the trend is about 180°. The sediments of the southern part of the West Greenland basin are thus believed to lie in a fault-bounded embayment representing the southern termination of this graben. N.B.H. Stevens of Olexcon International, The Hague (pers. comm.) has pointed out that the presence of an embayment would be a very favourable factor from the point of view of the petroleum prospects of the area.

Concurrently with the ending of this graben, another graben, with a trend of about 175°, is believed to develop southwards from the west coast of Disko. This reasoning is based on the abrupt downwarping of the basalts in this area, analogies being drawn with the structure of the basalts farther north. However, the structures in northwest Disko could perhaps be explained in terms of a single fault with downthrow to the west.

Now, for the first time, the full significance of the two dolerite master dykes cutting the Precambrian rocks to the east can be understood. The western dyke is not continuously exposed (in the area inland from Svartenhuk it enters an area with cover of Tertiary basalts whose relations with the dyke have not yet been established), but there is no doubt that it is the same dyke that extends from northeast of Jakobshavn (see Fig. 2) to north of Upernavik, a distance of 444 km. These two dykes are very well-exposed in the field and there is no displacement of marker horizons in the Precambrian rocks (see Henderson and Pulvertaft, 1967). They thus represent infillings of tension fractures which merge with the east side of the Melville Bugt graben. A dolerite dyke on the southern margin of Figure 10, which was mapped by Ellitsgaard-Rasmussen (1952) has a mean trend of 5° and has been traced over a distance of 60 km. This could also be an infilling of a tension fracture parallel to an offshore graben or fault.

Other elements that fit the proposed model are the faults. The fault system bounding the sediments and extending from northern Svartenhuk to Agpat (see Fig. 1) although irregular in form, has the general trend of the Melville Bugt graben, as has the Ikorfat fault system in Nûgssuaq. The Kûk-Sarqaq fault system and its presumed continuation down the eastern side of Disko Bugt has a N-S trend parallel to the Disko gneiss ridge and the second graben (or fault) west of Disko.

The concept that the sediments found onshore were deposited in a subsiding graben ending southwards as a fault-bounded embayment would be in accordance with evidence from the sediments. The oldest marine sediments are probably of Cenomanian age (base Cenomanian 100 m.y.), and there are marine sediments of definite Upper Turonian age (base Turonian 94 m.y.). These sediments are found in Svartenhuk, in the north of the onshore area. Sediments of comparable age farther south are nonmarine, the oldest marine sediments in Nûgssuaq being of Coniacian age (base Coniacian 88 m.y.) (see Rosenkrantz, 1970, Fig. 2 and p. 421-422). The marine transgression has thus been from the north.

The thick, tilted basalts between Ubekendt Ejland and Disko are considered to be lying in a subsidiary graben linking the Melville Bugt graben and the graben west of Disko. This subsidiary graben probably developed as a result of interference between the two main grabens and may have developed at a late stage (Danian?). The evidence for this is as follows. Field evidence has shown that there was considerable subsidence between the Disko gneiss ridge and the area east of Disko Bugt after the deposition of much of the sediments, but before the basalts were erupted. The main fault movements in the southern part of the area from the start of the volcanism (Danian) took place west of Disko. The cessation of subsidence in the Disko Bugt area may be related to the development of the graben farther west, where movements were concentrated from the Danian on. It is considered that continued movements in the southern extension of the Melville Bugt graben, as far as Nûgssuaq, and movements offshore from western Disko have caused the development of one long sinuous graben in which the thickest sections of basalts accumulated.

Possible Relation between Volcanism and Graben Development

Whereas there has been limited igneous activity along the fault system bounding the sediments to the east (mainly expressed by sills and dykes, but with the possibility for some extrusive volcanism) the graben system in the west of the area could be the major source of the West Greenland basalts.

It is considered that this basaltic volcanism was accompanied by a nearly continuous subsidence within the graben, so that most of the basalt pile was accommodated within the graben. This can be expressed in a general way, if it is assumed that the maximum thickness of basalt cover on more stable areas outside the graben is 2,000 m and the maximum thickness of basalts within the graben is 8,000 m. If it is postulated for illustrative purposes only that all flows have the same thickness, then three out of four flows would only be found within the graben, while every fourth, extruded during a temporary break in the subsidence or during a period of particularly prolific magma production, overlapped the graben margins on to the adjoining areas.

Thus with more or less continuous subsidence within the graben, no marked unconformities would be expected within the basalts. However, increase in thickness of the basalt pile in areas of downwarping would be expected. Rosenkrantz and Pulvertaft (1969, p. 895-896) note that this is seen in Svartenhuk, where the increase in thickness is due to an increase in the number of flows, not a thickening of individual flows. In mountainsides showing a vertical section through the basalts a decrease in dip upwards would be expected. This has been seen by V. Münther (pers. comm.) in Svartenhuk.

However, in areas where a major and sudden subsidence has occurred, angular unconformities would be expected. The Nûgssuaq-Hareøen area is such an area, there having been considerable subsidence, accompanied by tilting, along the northwest side of the Itivdle fault and its continuation across the southeast end of Hareøen. In the summer of 1970, a colleague working in Hareøen found the youngest basalts in this part of the area - olivine-porphyritic basalts - resting unconformably on tilted feldspar-phyric basalts (Hald, 1971).

Evidence for a Major Fault between Nûgssuaq and Disko

A sinistral transverse fault has been postulated between Nûgssuaq and Disko. There is a major sinistral wrench fault in the Precambrian rocks to the southeast (Escher and Burri, 1967) and the zone of downwarping is offset between Nûgssuaq and Disko. Moreover, V. Münther, who has worked on the basalts of this area for many years, postulates a fault in the channel.

Nevertheless, it seems prudent at this stage to be cautious in accepting a transverse fault with 25 km displacement of the Tertiary basalts, and other suggestions may have merit. Old zones of weakness (not necessarily faults) in the Precambrian may have been instrumental in determining the shape of the pull-apart structure, and could give rise to an irregular, zig-zag shape. Rosenkrantz and Pulvertaft have already pointed out (1969, p. 895) that the segments of the zig-zag fault system between northern Svartenhuk and Kûk do not continue into the Precambrian terrain as branches of the main fault.

Westward Continuation of the Umanak Fjord

The graben system shown in Figure 10 has been drawn without detailed reference to the region of offshore volcanics as shown by Park et al. (1971). These authors mention that the shelf west of the western boundary of the Disko basalt extension appears to consist of nearly flat-lying sediments draped over the basalts and that it is not known how much farther west the basalts extend beneath this sediment cover. If the graben concept is correct, then basalts would be expected to overlap the margins of the graben to the west as well as to the east, and no close correlation would be expected between the boundaries of the basalts and the limits of the graben below. However, if the ship's track (Park et al., 1971) is plotted on Figure 10 there is an approximate correlation between the two interpretations. Sediments are bounded to the north and south by basalts in the westward continuation of the Umanak Fjord. According to the model shown in Figure 10 this would be possible without postulating the absence of basalts in the channel between Ubekendt Ejland and Nûgssuaq. At time of writing, plans are being made for a marine geophysical survey by the Bedford Institute, with participation by the Geological Survey of Greenland, in this part of Umanak Fjord, so it is hoped to resolve the problem.

Evidence of Feeders for the West Greenland Basalts

If the postulated graben is the source of the basalts of West Greenland, how strong is the evidence of feeders for the flows? The evidence in the past has been limited, but it is now accumulating.

Rosenkrantz and Pulvertaft (1969, p. 892) consider that dykes acted as feeders to the lower basalt unit and that small central eruptions gave rise to the upper basalts. They mention the discovery by Drever in Ubekendt Ejland of four picrite dykes passing up into flows and one small neck found by V. Münther to pass up into a series of flows. While undertaking a photogeological interpretation of northwest Disko (see Fig. 9) the writer discovered a small volcanic neck. This was subsequently checked in the field by A.K. Pedersen (see Pedersen, 1969, p. 23). Pedersen (1969) also discovered a number of small volcanoes in northern Disko and mentions the presence of others in northwest Disko originally observed by V. Münther. Hald (1971) discovered scattered dykes of aphyric basalt cutting the feldspar-phyric basalts of western Hareøen. These are similar in mineralogy to the basalts found to overlie the feldspar-phyric basalts unconformably, and although no direct connection with these youngest basalts was observed, it seems probable that these dykes have acted as feeders.

Indirect evidence is also valuable in this context. Noe-Nygaard (1942, p. 69) considered that field evidence from Svartenhuk suggested that the dykes of this area should be regarded as feeders. A photogeological interpretation of the area southeast of the Itivdle fault in Nûgssuaq showed numerous dykes and parts of dykes in a wide zone southeast of the fault system.

Evidence from the pillow breccias should also be considered. The direction of crossbedding in the breccias of southern Nûgssuaq indicates a westward source, whereas in Svartenhuk it indicates a southwest to southward source. However these breccias do not outcrop in the westernmost parts of the land areas, so this evidence does not necessarily mean a source offshore in Davis Strait. An alternative explanation could be a source in the westward parts of the land areas and in the fjords in between. The possibility that some extrusive volcanism could be related to the faults bounding the sedimentary area to the east has already been mentioned, and is mentioned again here to stress the possibility that faults could have been feeder channels.

Possible Explanation for the Distribution of the Basalts

Although a graben is considered to be the source of the West Greenland basalts, some further factor is necessary to explain the restricted area within which the basalts occur. The paper by Park *et al*. (1971) shows the termination of the basalts north of Svartenhuk, and the Disko basalt extension is shown to terminate at 68°N. The limits of the province to the west, as distinct from the limits of the basalts on the sea floor, were not defined by this survey, but it is noted that in one place the basalts appear to give way to or are underlain by folded sedimentary strata (Park *et al*., 1971, p. 237). Basalts occur as far as the Inland Ice in the northeastern part of the area, but recent investigation of aeromagnetic profiles over the Inland Ice by Elizabeth King of the United States Geological Survey (E. King, pers. comm.) indicate that the basalts end some distance inland under the ice. The West Greenland basalts may thus be restricted to an elongated area with N-S trend on the coastal belt and on the inner part of the continental shelf. However, it is noted that Park *et al*. (1971) have left the boundary open due west of Disko. The results of further work in this western area are eagerly awaited.

It is suggested very tentatively that a second deformation, possibly a system of regional warps with E-W trend and very large wavelengths, may have been superimposed on the structures already described, to control production of magma from the graben in this area of limited N-S extent. This deformation could be a reflection of the commencement at 60 m.y. of the second phase of the opening of the Labrador Sea as proposed by Le Pichon, Hyndman and Pautot (1971).

Possible Ages of the Sediments Offshore

Based on surface outcrops, the maximum thickness of marine sediments onshore in West Greenland is 2,000 m. The maximum thickness of nonmarine sediments is 1,500 m, but these are in part equivalent in age to the marine sequence. However, the oldest nonmarine sediments are of Barremian-Aptian age (base Barremian 118 m.y.) and are thus older than the oldest marine sediments. The thicknesses displayed onshore are very much less than the thickness of sediment in the Melville Bugt graben (6,000 m). Considerable thicknesses of sediment have been reported from other parts of the marine area off West Greenland. Hood and Bower (1970; 1972) report that Baffin Bay west of the graben is underlain by sediments of thickness comparable to that found in the graben. Drake et al. (1963, p. 1086) show two seismic profiles across the Labrador Sea to the Julianehaab area off South Greenland and state that near the continental margins of Greenland and North America the surface of the underlying rock is deeply buried beneath the sedimentary cover and did not appear on the profile records. Mayhew et al. (1970) report the results of two seismic refraction and magnetic profiles off Julianehaab. One profile runs NE-SW from near the shelf break out into the basin, while the other is normal to the first profile, somewhat shoreward of its northeast end. The NE-SW profile shows that the high-velocity basement apparently comes very close to outcropping midway down the continental slope and that a thick layer of sediments (2.0 km/s) overlies a 3.9 km/s layer northeast of this point. Hood and Bower (1972) report considerable thicknesses of sediment, in places more than 16,000 feet (5 km), below parts of the West Greenland continental slope in the area southwards from Disko Bugt. Hood et al. (1967) have already reported the presence of more than 20,000 feet (6 km) of sediments below parts of the outer Labrador Shelf.

The Melville Bugt graben probably contains Cretaceous-Tertiary sediments equivalent in age to those occurring on land. In addition, Tertiary sediments may have been laid down at the time basalts were being erupted in West Greenland, and younger Tertiary sediments may also be present. The seismic profile given in the paper by Manchester and Clarke (in press) shows about 150 m of nearly acoustically-transparent sediments overlying an erosional surface. These sediments are presumably of Quaternary age.

The very substantial thicknesses of sediment in parts of the marine area between Greenland and Canada pose the question whether sediments substantially older than those seen on land in West Greenland are present below the sea floor. McMillan (1972) reports the abundant presence of carbonate rocks in bottom samples collected from Baffin Bay and the Labrador Sea. On the Labrador shelf these samples range in age from mid-Tertiary at 53°N to Lower Cretaceous at 58°N. Two Upper Jurassic pebbles were found. In this connection it may be mentioned that D. Bridgwater (pers. comm.) noted the abundant presence of cobbles and boulders of fossiliferous carbonate rocks along the shoreline in the Saglek, Nain and Makkovik areas of the Labrador coast (i.e. between 55°N and 59°N) during field work in 1969. It has not yet proved possible to date these rocks.

A. Rosenkrantz, in a footnote to a paper on the Lower Jurassic rocks of East Greenland (Rosenkrantz, 1934, p. 7) made a statement that may well prove to have been prophetic:

"In this connection it deserves mention that SAM HAUGHTON (10)* in 1859 by dredging on the sea-bottom off Godhavn in West Greenland secured a fossil that was determined as Cardinia ovalis (STUTCHBURY), a species belonging to the angulatus zone of the Lower Lias. If the determination is correct, this find must be said to be of great interest, Jurassic beds not otherwise being known from West Greenland. The fossil is assumed to have been carried to West Greenland by the pack-ice, but as the nearest place from which the Liassic beds are known is the Scoresby Sound area, this view seems not very probable. A closer investigation of the sea-bottom off Godhavn is therefore very desirable."

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*The reference (10) in the above quotation is: Haughton, S., 1862. On the fossils brought home from the arctic regions in 1859, by Captain Sir F.L. M'Clintock. J. Roy. Dublin Soc., v. 3, p. 53-58.

An indirect pointer to the possible age of rocks offshore from West Greenland is provided by Watt (1969), who states that the Jurassic age of the coast-parallel dykes of southwestern Greenland suggests that initial rifting between Greenland and Labrador took-place at about the middle of the Mesozoic Era. Henderson (1969, p. 9-11, 23) mentions evidence from the rocks of the West Greenland basin suggesting that rocks older than those seen onshore could be present away from the margin of the basin.

The conclusion must be that in areas offshore from West Greenland where the sedimentary section is thick (this includes the Melville Bugt graben), it would seem reasonable to expect Jurassic rocks in the lower part of the section. The possibility that even older sedimentary rocks could be present should not be overlooked.

ACKNOWLEDGMENTS

The writer would like to thank Professor A. Rosenkrantz, Dr. P.R. Dawes, Dr. D. Bridgwater and Dr. W.S. Watt for critical reading of the manuscript, and colleagues at the Bedford Institute and Dalhousie University for discussions on the concept presented.

Thanks are also due to the Bedford Institute for consenting to the use of previously unpublished marine magnetometer profiles, to Dr. P.J. Hood, Dr. X. Le Pichon, Mr. K.S. Manchester and Dr. N.J. McMillan for allowing the writer to read manuscripts of papers submitted for publication, and to the Director of the Geological Survey of Greenland for permission to publish this paper.

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33. MAGNETIC PROPERTIES OF ROCK SAMPLES FROM THE BAFFIN BAY COAST

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Abstract

Various magnetic properties of about 200 Tertiary basalt samples from the coasts on both sides of Baffin Bay have been measured. The main collections are from (i) three profiles representing over 30 lava flows on southern Disko Island, West Greenland, and (ii) one profile of six flows near Cape Dyer on Baffin Island. In both areas, the natural remanence (NRM) of the samples is predominantly stable, being reversed in Disko and normal in Cape Dyer, but unstable components dominate in some of the samples. The stable component of magnetization appears to be primary, i.e. acquired when the lavas cooled from a molten state. A paleomagnetic pole position at 62°N, 169°W, with an error oval of $\delta p = 8°$, $\delta m = 9°$, was obtained from two of the South Disko profiles; the significance of this result to local paleogeography is discussed.

Various magnetic properties of Baffin Bay Archean metamorphic rock samples have been measured and are discussed, with other available results on the magnetism of local Tertiary and Precambrian rocks, in the context of magnetic anomaly interpretation.

INTRODUCTION

The basic volcanic rocks found on both sides of the Davis Strait (Fig. 1) have been described by Rosenkrantz and Pulvertaft (1969), Clarke and Upton (1971) and Henderson (1972), among others. From fossil evidence and K-Ar dating, these volcanics are believed to be mostly of Paleocene-Eocene age, and they are probably related to the opening of Baffin Bay by a process of sea-floor spreading. The Baffin Island volcanics, exposed only along a narrow coastal strip between Cape Dyer and Cape Searle, consist of a few hundred metres thickness of olivine-rich basalt breccia and lavas, with minor intrusives in the north. The much thicker volcanic suite in West Greenland consists mainly of olivine lava flows overlain by olivine-poor, usually feldspar-porphyritic lavas. It rests partly on Precambrian basement and partly on Cretaceous-Tertiary sediments, subaqueous breccia often occurring as the first volcanic facies. Much intrusive activity and faulting has taken place in west Greenland, and distinct interbasaltic beds are rare, which makes regional stratigraphic mapping difficult.

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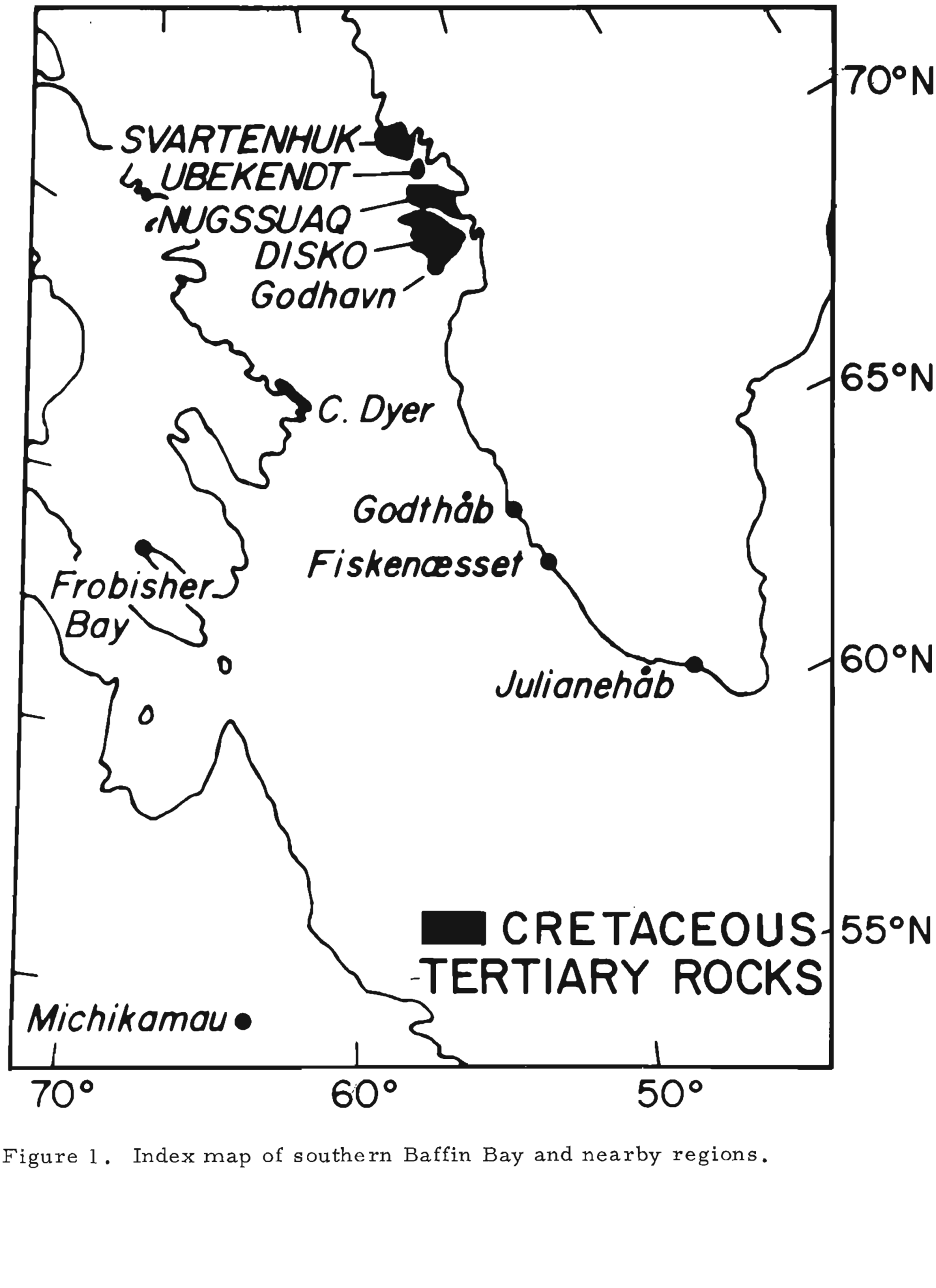


Figure 1. Index map of southern Baffin Bay and nearby regions.

EARLY TERTIARY PALEOMAGNETISM IN BAFFIN BAY

Volcanics on Cape Dyer, Baffin Island

In 1968, a Memorial University expedition collected 38 oriented samples from six horizontal lava flows at Cape Dyer (Fig. 1). The following account is mainly a summary of results from a more detailed description by Deutsch et al. (1970).

Except for three samples out of six from one flow, the samples were all magnetized in the same (normal) sense and with a similar direction as the Earth's present magnetic field, i.e. their natural remanence (NRM) was inclined steeply downward. The three anomalous samples were reversely polarized, with steep upward NRM directions that were not very stable to alternating-field (AF) or thermal demagnetization; we suspect that they have suffered secondary magnetization in situ, although otherwise all samples appeared to be fresh and unaltered. The normal NRM directions were found to be very stable to AF and thermal treatment and are therefore believed to be primary remanence directions.

The average paleomagnetic field direction, obtained from 5 of the 6 Cape Dyer flows after treatment to 400 paek Oersteds, corresponds to a pole position at 83°N, 55°W, with a 95 per cent confidence oval of mean radius 12 degrees. Because of the small number of flows used in this and in other published Lower Tertiary paleomagnetic studies from North America, no definite conclusions about the paleogeography of Baffin Bay can yet be drawn from these results.

Thermomagnetic and AF curves and microscope examination indicate that the remanence of the Cape Dyer lavas resides in titanomagnetite. Both in our samples and in several samples of olivine-rich basalt from Baffin and Svartenhuk examined by Clarke (1968), this mineral constitutes 2 per cent or less of the rock by volume. The arithmetic average intensity of our samples (excluding one exceptional value of 59×10^{-3} Gauss) is about 4×10^{-3} Gauss, a Gauss being the c.g.s. electromagnetic unit of magnetic moment per unit volume. Their average susceptibility in c.g.s. units is 0.6×10^{-3} Gauss/Oe; therefore remanent magnetization dominates induced magnetization in situ. Buildup of viscous remanence on storage in the laboratory over a period of 1-2 months was small ($\lesssim 5\%$ of the NRM).

We have also made magnetic experiments on eight unoriented samples from Tertiary basalt outcrops north of Cape Dyer, collected and kindly given to us by Dr. D. B. Clarke of Dalhousie University. Their NRM intensities (average 2.3×10^{-3} Gauss) and susceptibilities (average 0.5×10^{-3} Gauss/Oe) and high magnetic stability are comparable to those obtained from the Cape Dyer samples just described.

Aeromagnetic maps of the Cape Dyer coastal area, published by the Geological Survey of Canada (1970) show positive magnetic anomalies at almost all the twenty or so main exposures of Tertiary basalts (as identified from aerial photographs). The amplitude of these anomalies averages about 1,000 γ at 300 metres above ground. We conclude that the Tertiary volcanics outcropping at Cape Dyer are predominantly normally magnetized, and were probably extruded during one geomagnetic epoch of normal polarity.

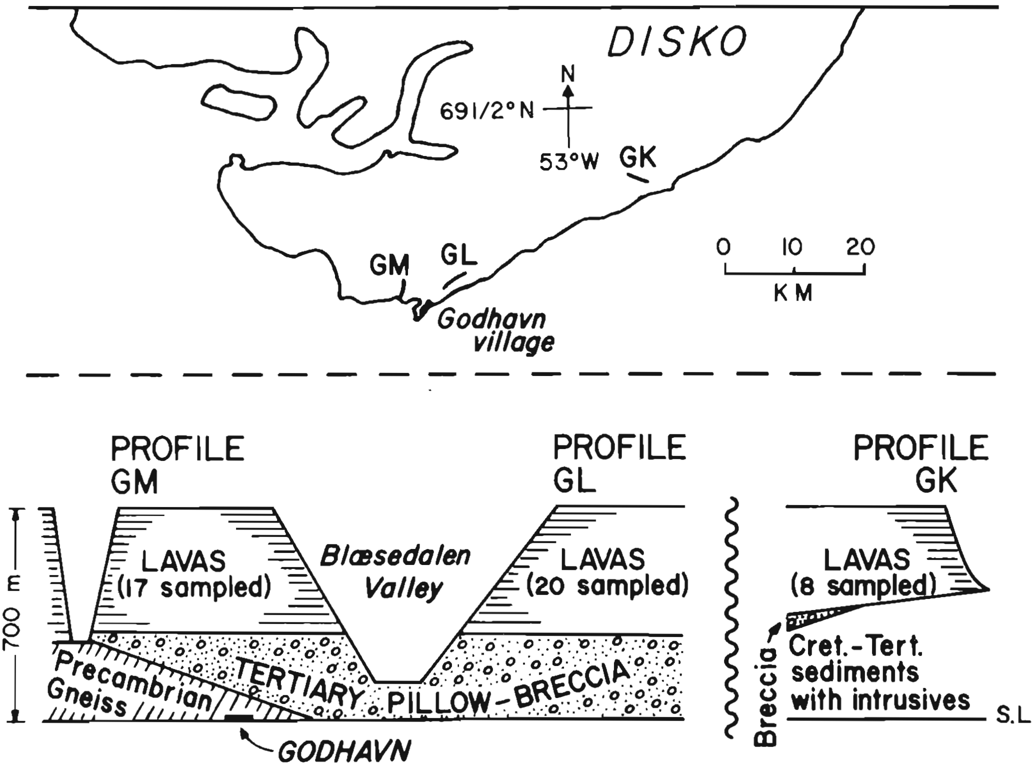


Figure 2. Map and schematic cross-section of paleomagnetic sampling localities on the south coast of Disko Island.

Volcanics at Disko Island, West Greenland

In the summer of 1970, a Memorial University expedition collected oriented samples from feldspar-porphyritic basalt lavas in three profiles, GL, GM and GK of Figure 2, on the south coast of Disko Island. Samples were also obtained from the underlying breccia and from basalt intrusives. Most of the samples were obtained using a portable core drill and were spaced several metres apart where possible. They were oriented either by suncompass or geographic sightings, depending on weather conditions; the orientation is believed to be accurate within 2 degrees in inclination and 3 degrees in azimuth on the average. Few or no lavas were missed within the profiles where sampled, but on the plateau above both profiles GL and GM, a few poorly exposed lavas were not sampled. The massive parts of the lava flows are fresh and almost free from zeolites, and they are horizontal within 1-2 degrees.

The remanent magnetization of 110 samples (one specimen per sample) from the lava profiles, and of about 30 samples from the breccia and intrusives was measured with a PAR spinner magnetometer. We also used an AF demagnetizer (Pearce, 1967); a pendulum balance (Deutsch et al., 1971) for obtaining strong-field thermomagnetic curves; a Scintrex SM-4 1000-Hz bridge to measure susceptibility; and a Zeiss Universal Standard microscope for petrographic work.

The NRM directions of the Disko basalts were of both polarities, but were mainly reversed, and on progressive AF demagnetization in steps of 50 Oe (Fig. 3), all samples were found to possess a stable or very stable reverse component of remanence. Treatment to 200 peak Oe appeared to give the best grouping of remanence directions, the r.m.s. value of the within-site dispersion angle δ_w (Sanver, 1968) then being about 5 degrees in the lava profiles. The low value of this dispersion as compared to the between-site dispersion δ_k (Fig. 4 and Table 1) clearly indicates that this stable component is a primary remanence.

All mean site paleofield directions are shown in Fig. 4. Table 1 gives the mean overall directions and statistical parametres for the lava flows of profiles GL and GM. As the lavas are nearly horizontal and these profiles are only 4 km apart, they probably were extruded during the same interval of time, which appears to span at least several secular variation cycles of that magnetic epoch. As it has not been possible to correlate the magnetic field directions in these two profiles, we will assume that no flow occurs in both of them, i.e. that the horizontal extent of each flow is less than 4-6 km. On this assumption the last entry in Table 1 represents the best average local paleofield for the time interval. This field corresponds to a geocentric magnetic dipole with a northern pole at 62°N, 169°W, which has error oval semi-axes $\delta p = 8°$, $\delta m = 9°$. The pole is shown in Figure 4 along with a pole from the reversely magnetized Lower Tertiary basalts in East Greenland (Tarling, 1967), and a pole from the Faeroe Islands basalts (Tarling, 1970), which are also Lower Tertiary but of mixed polarities. A mean pole from various British Tertiary formations has also been given by Tarling (1970) and is very near the Faeroes pole position.

Figure 4 shows the site mean field directions obtained from most of our sites in the breccia, intrusives and in profile GK in Disko. Their mean field direction is seen to be similar to that of GL and GM, but its significance is low, both because of the occurrence of shallow directions in GK due to the uncertain age of the intrusives, and because the breccia sampling sites were spaced unevenly in the sampling area. Table 2 lists other averaged magnetic properties for all these formations.

In profiles GL and GM, the samples collected seem to fall into two magnetically distinct groups. The larger group has a much higher intensity of primary remanence than the other and has less secondary magnetization (compare GL 10-1 and GL 8-2 in Fig. 3), lower mean susceptibility, higher oxidation state in polished section, and higher strong-field Curie point (560°C vs. 520°C). In any particular flow, however, both these types of magnetic behaviour may occur within metres of each other horizontally or vertically. This points to variable oxidation conditions within the flows during or subsequent to cooling, and it may preclude the application of magnetic properties other than paleofield directions and polarities, to stratigraphic correlation in Disko. About a half of the samples collected from profile GK, the breccia and intrusives fitted these two magnetic groups, but the rest behaved differently, possibly because of rapid cooling, and will be discussed below. In several polished sections of South Disko lava samples we find a titanomagnetite content of 5-8 per cent by volume, and similar values have been reported by Clarke (1968) from the feldspar-porphyritic basalts of Svartenhuk Peninsula.

TABLE 1

Mean remanence directions for two lava profiles on Disko Island, west Greenland

Profile	N	n	D	I	$\alpha 95$	δ_k
GL	20	3	133	-64.4	7.0	16.6
GM	17	2	150	-69.2	9.6	20.5
Both	37	-	140	-66.7	5.7	19.0

All samples measured after AF treatment at 200 peak Oersteds:

N = number of flow-mean directions averaged;

n = number of samples measured per flow (usually one specimen per sample);

D = declination; degrees east of north;

I = inclination; degrees, positive down;

$\alpha 95$ = radius of 95 per cent confidence circle in degrees;

δ_k = observed between-flow dispersion angle in degrees (Sanver, 1968).

TABLE 2

Some magnetic properties of igneous rocks on Disko

Rock group	N	n	$\lvert J_o \rvert$ (10^{-3} G)	J_{200} (10^{-3} G)	K (10^{-3} G/Oe)
GL flows	20	3	2.9	2.3	3.4
GM flows	17	2	3.5	2.7	3.1
GK flows	8	2	8.6	3.0	2.4
Breccia sites	7	1-3	6.5	2.7	1.5
2 dykes and 5 sills	7	2	6.6	1.7	2.4

Arithmetic averages of site-mean values are shown:

N = number of sites averaged;

n = number of samples measured per site (usually one specimen per sample);

$\lvert J_o \rvert$ = NRM intensity (averaged without regard to sign);

J_{200} = remanence intensity after 200 Oe demagnetization (all directions have reverse polarity);

K = volume susceptibility.

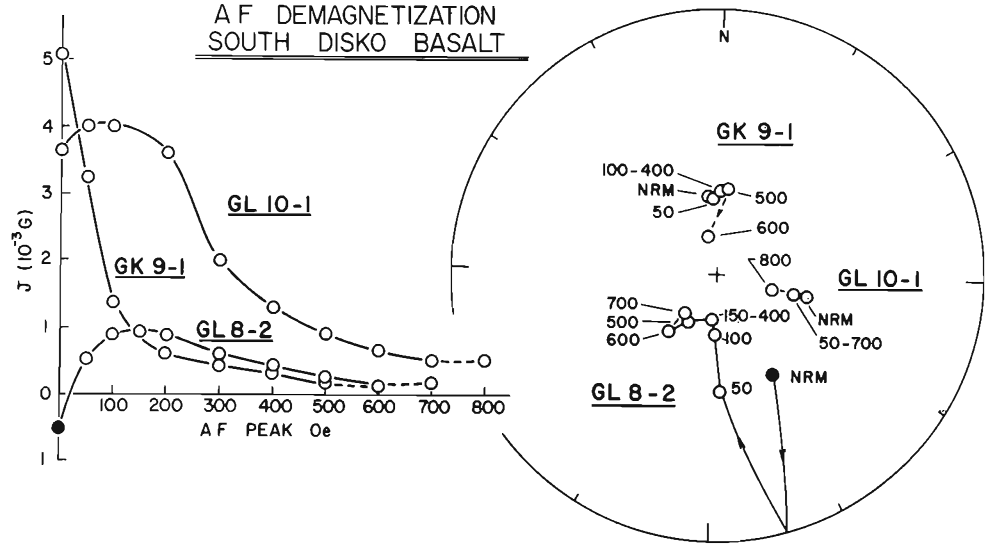


Figure 3. Left: Remanent intensity J as a function of AF demagnetizing field for three typical South Disko basalt samples. Right: Polar equal-angle projection of remanence intensities for the same samples. The declination of GK 9-1 has been altered by 180 degrees to avoid overlap. Numbers refer to demagnetizing fields. In both parts of the figure, open (filled) circles refer to reverse (normal) magnetization directions. Broken lines indicate that in the last step of each demagnetization, spurious components appeared in J.

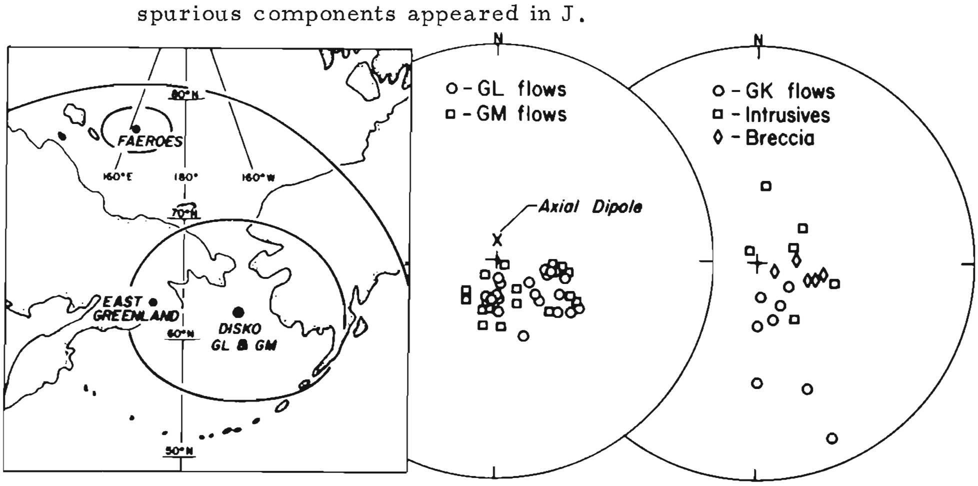


Figure 4. Left: Paleomagnetic pole position for Disko, calculated from the overall flow-mean remanence direction of profiles GL and GM (Table 1), with 95 per cent confidence oval. Also shown are pole positions and 95 per cent confidence ovals from 28 Tertiary flows in East Greenland (Tarling, 1967) and from 253 flows in the Faeroe Islands (Tarling, 1970). Middle: Individual flow directions for profiles GL and GM, after 200 Oe treatment. Right: Individual site directions for profile GK, breccia and intrusive sites after 200 Oe treatment. Polar equal-angle projection.

Paleomagnetism and Rotation of Greenland

One possible use of magnetic data from rocks bordering Baffin Bay is paleogeographic, for example in a test of the proposed separation of Greenland from Canada (Wegener, 1929). In planning future field work, the minimum sampling required for such a test may be roughly estimated.

For this purpose we assume that (1) the pole reported here for Disko Island (62°N, 169°W) is the "true" early Tertiary pole relative to present-day Greenland; (2) the Tertiary rocks on Baffin Bay became magnetized in a geocentric axial dipole field of either polarity; (3) after the rocks were magnetized, Greenland rotated away from Canada according to the model by Bullard et al. (1965); and (4) no other crustal displacements involving these rocks occurred. Rotating Greenland back to Canada shifts the Disko pole into a new position that should coincide with the hypothetical early Tertiary pole for Baffin Island. The angular distance p between the latter pole and the original Disko pole then defines the limiting sensitivity of the test. For a specified rotation angle R, p depends only on the paleocolatitude θ of the rotation pivot and is given by (Deutsch, 1969): $1 - \cos p = \sin^2\theta(1 - \cos R)$. In the Bullard model for Greenland, R = 18.0° and the pivot is at 70.5°N, 94.4°W. The distance of the Disko pole from this pivot is $\theta = 29°$, giving p = 9°. Though small, p is more than twice the present distance between Cape Dyer and Disko.

One may judge the hypothetical poles for Greenland and Baffin Island to be distinct if their 95 per cent error ovals do not intersect. For a maximum pole separation p = 9°, this means that the sum of the major semi-axis δm of the two ovals should not exceed about nine degrees; hence, on average, the oval dimensions should not be more than one-half those of the present oval for Disko ($\delta m = 9°$). We now assume for simplicity that the dispersion of the N remanence vectors averaged in calculating future pole positions will be the same as the dispersion we observed for Disko; then δm varies approximately as $N^{-1/2}$. A significant test would thus require on average four times the present number of Disko flows (N = 37), i.e. 150 or so rock units, on each side of Baffin Bay. This minimum requirement is large, but probably can just be met. We conclude that a paleomagnetic test of the opening of Baffin Bay is marginally feasible, but that the same rock collections needed to investigate these purely directional features of the early Tertiary field can probably yield more significant results on other characteristics of that field, such as its secular variation, strength and polarity.

ROCK MAGNETISM AND MAGNETIC ANOMALIES

Magnetic surveys in the Baffin Bay area

In addition to the aeromagnetic mapping by the Geological Survey of Canada referred to above, several magnetic surveys have been carried out in Baffin Bay and nearby regions (Hood et al., 1967; Johnson et al., 1969; Mayhew et al., 1970; Haines et al., 1970; Park et al., 1971; Vogt, 1970; Hood and Bower, 1972). Attempts at interpreting anomalies found in these surveys have suffered from a lack of knowledge about the nature or magnetic properties of the causative bodies, since they are often submarine, subglacial or otherwise unexplored. Below, we will discuss two examples of how such knowledge may provide constraints on the interpretation of magnetic surveys in the Baffin Bay area. The presently available rock magnetic results from the area are somewhat fragmentary (see legend to Fig. 5) but they may serve as guidelines for future work on establishing the origin of the magnetic anomalies.

Example 1: Disko Island anomalies

As stated above, all the samples collected in South Disko by Memorial University possess a stable reverse remanence of presumably primary origin, and one may therefore expect a broad negative aeromagnetic anomaly to occur over Disko. Such appears to be the case in the results of Hood and Bower (1972) but in the magnetic maps of Haines et al. (1970) a small positive anomaly in total field F occurs over the south coast. This anomaly may perhaps be explained by the following results from our profiles GL and GM:

The NRM of one sample from each flow (total 37) was measured soon after they arrived at Memorial University. They were then stored for a period of 4-6 weeks, remeasured and then AF demagnetized in steps of 50 Oe. It was found that in all samples the remanence shifted towards the reverse sense on storage, and still further, up to a maximum (Fig. 3) on AF treatment. The maximum usually occurred between 50 and 150 Oe. These results clearly indicate that on the primary remanence there was superimposed a very soft viscous remanence (VRM) due to the present field in Disko, and that this VRM decayed during storage, being replaced with one of the opposite polarity because the samples were stored upside down. By extrapolating the observed change in VRM back to the date of collection (assuming simple expotential decay, though the changes involved are mostly small and the choice of extrapolation method makes little difference), we were able to estimate the average in situ VRM and hence the in situ NRM. It was assumed that the maximum in each AF demagnetization curve (as in Fig. 3) represents the in situ primary remanence, i.e. that this remanence is unaffected by AF treatment to 50-150 Oe. The volume susceptibility K of all samples was also measured. The calculations shown in Table 3 then yield an estimate of the total in situ intensity of magnetization J_t for each profile. Because the paleofield and present field directions are not far from the vertical, we have added all magnetization values as scalar quantities, with due regard to sign, but for improved accuracy in application to local magnetic surveys they may be multiplied by correction factors as indicated by Kristjansson (1970).

From the results of Table 3 it is seen that the lava flows of profiles GL and GM, although reversely magnetized to begin with, would cause a small positive anomaly in total-field geomagnetic intensity, as found by Haines et al. (1970), if surrounded by effectively non-magnetic material. If surrounded by similar basalts of a normal primary magnetization, the magnetization contrast would be twice the value of J_p in Table 3, i.e. about 6×10^{-3} Gauss.

Because the primary remanence in the other basalt formations sampled in South Disko, especially in those samples that appeared to have cooled rapidly, was often very soft to AF treatment (in intensity, e.g. GK 9-1 in Fig. 3; the direction of remanence was stable) their in situ VRM could not be estimated by extrapolation as described above, but only indirectly from repeat measurements after storage. As shown in Table 4, the VRM is small and the mean primary remanence intensity in these formations is much higher than in profiles GL and GM, so that these formations could, by themselves, cause considerable negative magnetic anomalies.

Since the Tertiary areas in Greenland are partly made up of olivine-poor basalts as described above and partly of olivine lavas as described in the section on Cape Dyer, overlying a variable thickness of breccia, which in turn is partly made up of randomly magnetized material, it may prove difficult to correlate local geology and magnetic anomalies in detail.

Two other results from South Disko may have some relevance in this discussion: First, we have measured the susceptibilities of three samples of Cretaceous-Tertiary sediments underlying the lavas of profile GK (Fig. 2) and found them to be essentially non-magnetic ($K \sim 0.01 \times 10^{-3}$ Gauss/Oe). Secondly, it is known that iron inclusions occur in the Disko basalts (e.g. Melson and Switzer, 1966) and these may be highly magnetic: selected and cut pieces (~ 1 cc) from two samples of basalt, kindly sent to us by Mr. E. Fundal, were found to have remanence intensities of the order of 500×10^{-3} Gauss. However, we believe that the iron inclusions are sufficiently rare and small as not to contribute appreciably to local aeromagnetic anomalies.

Example 2: Baffin Bay coastal and oceanic anomalies

Figure 5 summarized data available to us on the magnetism of Baffin Bay and nearby coastal rocks. Each dot represents a sample or a site average of NRM intensity, plotted without regard to sign, or of susceptibility. The lengths of the arrows indicate the magnitude of the present local geomagnetic field, i.e. the distance that the susceptibility values should be moved along the axis to become values of induced magnetization in Gauss.

It may be assumed that the Precambrian basement rocks bordering Baffin Bay consist predominantly of metamorphic rock types, magnetic results for which are shown in the bottom part of Figure 5a. It may also be assumed, as has been demonstrated by Puranen et al. (1968, Fig. 9), that the r.m.s. amplitude of magnetic anomalies observed over a region is directly related to the standard deviation of total-magnetization values in the underlying rocks. On comparing the gneiss values in Figure 5a with the values for Tertiary basalts shown in Figure 5b, it is apparent that the former will, under similar conditions, produce much smaller anomalies on the average. This has been used by Park et al. (1971) in mapping the extent of Tertiary volcanics off West Greenland. It must also be noted, however, that both in the metamorphic rock types and especially in the less common basic Precambrian rocks (Fig. 5a), magnetization values comparable to those in Tertiary basalts do occur. This observation is supported by the fact that magnetic anomalies of several hundred gammas are commonly recorded in magnetic surveys over areas where only Precambrian rocks are known to occur. Interpretation of single magnetic anomalies in regions of unexposed bedrock, such as on the Labrador Sea and Baffin Bay continental shelves, in terms of the age and lithology of the causative bodies, may therefore be rather speculative.

In such qualitative interpretation of magnetic anomalies over the presumed submarine basalts underlying the central parts of Baffin Bay and the Labrador Sea, it would be very useful to know their average remanence but this cannot be inferred directly from our own or from other subaerial basalt results. It is known that rapid cooling, such as would take place in

TABLE 3

Estimated in situ intensity of magnetization of two Disko lava profiles

Component	Intensity (10^{-3} G) GL	Intensity (10^{-3} G) GM	Net Sense
Observed NRM, J_n'	1.7	2.1	reverse
Primary remanence, J_p	2.8	3.3	reverse
In situ VRM, J_v	1.5	1.8	normal
In situ NRM, J_n	1.3	1.5	reverse
In situ induced magnetization, J_i	2.2	2.0	normal
Total in situ magnetization, J_t	0.9	0.5	normal

The arithmetic average of N flow intensity values in each profile (Table 1) is shown; one sample was measured per flow:

J_n' is the mean of two measurements;

J_p is the maximum J in the AF curve (Fig. 3);

J_v was obtained by extrapolation in time (see text);

$J_n = J_p + J_v$;

J_i = KF, where K is volume susceptibility and F the present field at Disko (0.56 Oe);

$J_t = J_n + J_i$.

TABLE 4

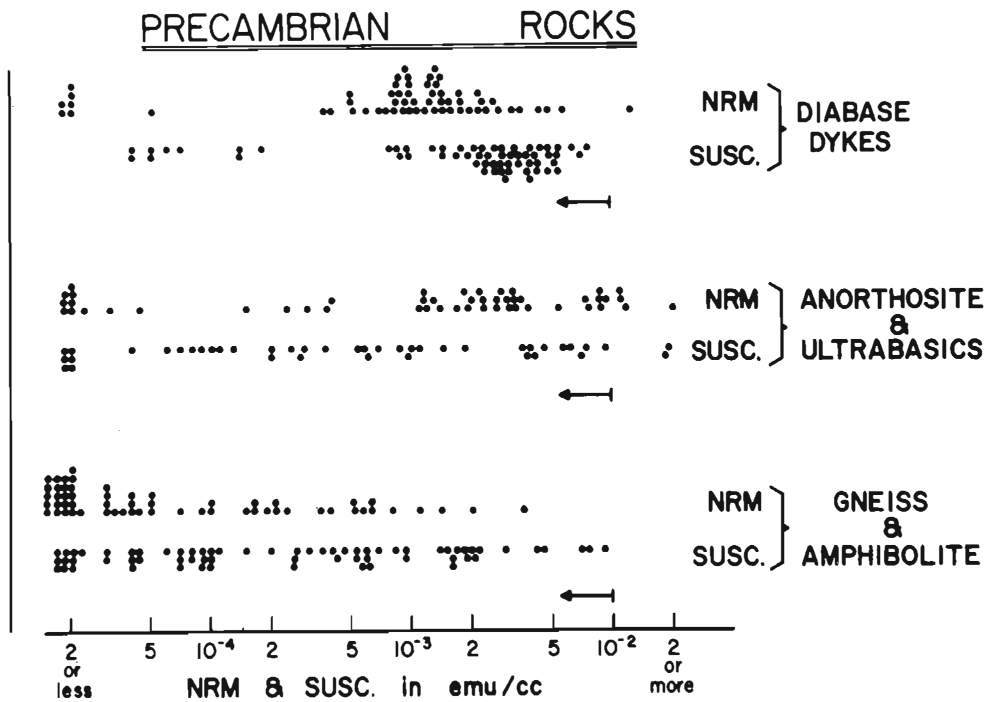
Estimated average in situ magnetization of some igneous rocks on Disko

Component	Intensity (10^{-3} G)	Net Sense
Primary remanence, J_p	7.3	reverse
In situ VRM, J_v, estimated	0.5	normal
In situ induced magnetization, J_i	1.2	normal

The arithmetic average of 22 site-mean intensity values (lava profile GK, breccia and intrusives; Table 2) is shown.

For most sites, 2 samples were averaged. See also explanatory notes in Table 3.

extrusive submarine basalts, generally tends to produce high remanence intensities. Also, some of the samples collected by us in Disko show evidence of rapid cooling (pillow and block-jointed lava structures; small skeletal homogeneous titanomagnetite grains in a glassy groundmass) and these tend to have



Top: Site mean values for 44 diabase dykes between Julianehab and Godthab, west Greenland, possibly including a few Mesozoic (TD) dykes (Bullock, 1967); plus 16 sites in north Baffin Island (Fahrig et al. 1971, and Dr. W. Fahrig, pers. comm.).

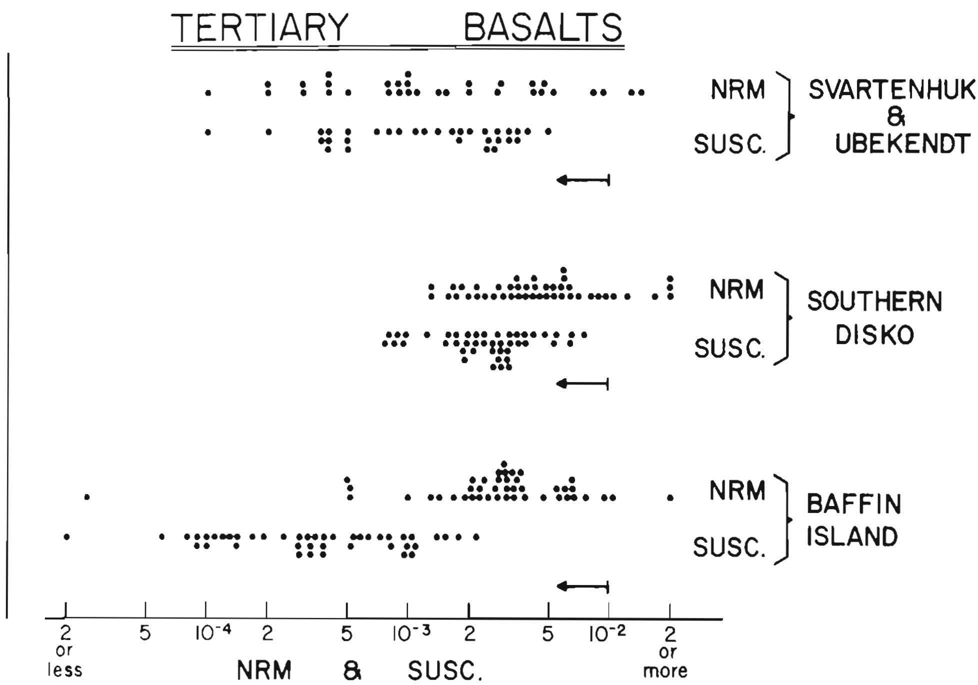
Middle: 18 anorthosite and ultrabasic samples, Fiskenaesset complex (Ghisler and Sharma, 1969); 29 anorthosite samples from Michikamau (Murthy, 1969), NRM only; 22 Michikamau anorthosite specimens (kindly lent by Dr. G.S. Murthy), susceptibility only; and 2 samples of basic rock near Julianehab.

Bottom: 31 gneiss samples collected by Memorial University between Julianehab and Godhavn, and in Frobisher Bay and two other localities on Baffin Island; 7 gneissic gravel samples from Baffin Bay (solid-rock susceptibility only, inferred by us from measurements on this gravel; 17 Fiskenaesset gneiss and amphibolite samples (Ghisler and Sharma, 1969); and 10 gneiss samples from the Labrador coast east of Michikamau (collected and kindly given to us by Dr. J.S. Sutton).

Figure 5a. NRM intensity and volume susceptibility of Precambrian basement rocks from Baffin Bay (Fig. 1).

remanence intensities a few times higher than the subaerial basalts of profiles GL and GM. Similar results have been obtained from the subglacial volcanics in Iceland (Kristjansson, 1970). On the other hand, the magnetic effects of magmatic oxygen fugacity and of secondary alteration mechanisms in suboceanic basalts are not well known.

An empirical approach to the problem of estimating the average magnetization of submarine Baffin Bay basalts may, however, proceed as follows. First, it may be found from Table 2 and the other data making up Figure 5b that the arithmetic mean remanent intensity (without regard to sign) in South Disko, Ubekendt Island and Cape Dyer (see Fig. 1) Tertiary basalts is everywhere of the order of 3-5 x 10^{-3} Gauss. Secondly, the average NRM intensities of Tertiary subaerial basalts in East Greenland (Tarling, 1967) and in Iceland (Kristjansson, 1970) are also of similar magnitude. Thirdly, Irving (1970) has shown that the mean remanence of basalts dredged from the Mid-Atlantic away from the axial rift zone is of the order of 5-8 x 10^{-3} Gauss.



Top: Average values for 6 lava sites and 21 dyke sites, Ubekendt Island (Dr. D.H. Tarling, pers. comm.), plus 5 samples from Svartenhuk Peninsula (collected by Dr. D.B. Clarke).

Middle: One sample each from all sites of lava profiles GL and GK and 2 other lavas, 7 intrusives and 7 breccia sites, South Disko.

Bottom: 35 samples from the Cape Dyer lavas, plus 8 from other Tertiary basalt exposures on Baffin Island (see text).

Figure 5b. NRM intensity and volume susceptibility of Tertiary basalts from Baffin Bay (Fig. 1)

Since the exposed Baffin Bay and North Atlantic subaerial rocks are so similar in this aspect, one might, by analogy, expect the submarine Baffin Bay basalts to have an average remanence not very different from that of Mid-Atlantic rocks, i.e. 5-8 x 10^{-3} Gauss, or slightly lower because of more advanced alteration.

Irving (1970) has proposed that the magnetic "smooth zones" bordering the North Atlantic are due to the presence of underlying intrusives and subaerial basalts, emplaced in the initial stages of continental drift. The average remanence of these, according to Irving's estimates, would be of the order of 0.6 x 10^{-3} Gauss. As this is much less than the average NRM of the Tertiary basalts just mentioned, we conclude that any basic rocks underlying the smooth zone would resemble the Mesozoic volcanics tabulated by Irving, and not be Tertiary basalts. Because of the high susceptibilities occurring in some of the Tertiary rocks, this conclusion would be valid even if the smooth zones represent a time interval when the geomagnetic dipole moment was small, and even if alteration has taken place in the smooth zone rocks.

The Magnetization of Gneiss

Finally, we consider the origin of the magnetization of the Baffin Bay basement gneiss. Strong-field thermomagnetic curves obtained by us from five unweathered gneiss samples were reversible (in air) with Curie points about 565°C, which is evidence that their magnetic properties reside primarily in pure magnetite. In polished sections made from the more magnetic of the gneiss samples, magnetite grains were indeed seen and have diameters of the order of 100μ. The remanence of Baffin Bay gneiss is weak (Fig. 5a) and of normal polarity, and is therefore most probably of viscous origin, though in some samples the median destructive field was found to exceed 200 Oe. Since magnetite crystal structure in gneiss may be expected to be much more uniform and ordered than in rapidly-cooled basalt, it is reasonable to look for correlations between magnetite content and susceptibility in the former. This we have done for gneiss samples from several widely-scattered sites around Baffin Bay, and the results are illustrated in Figure 6.

The magnetite percentage in each of these samples was estimated by counting 6000 or more squares in a 2-3 sq. cm. polished section with a square-grid microscope eyepiece. Our results, supplemented by results obtained with similar techniques by Ghisler and Sharma (1969), strongly indicate a linear relationship, and a best-fitting straight line through the data points and the origin has a reciprocal slope of 2.8 x 10^{-3} units of susceptibility per volume percent of magnetite. As these gneiss samples are coarse-grained and often foliated, the possible error in the above value is $\pm 0.5 \times 10^{-3}$ Gauss/Oe/volume per cent.

Very similar relations, giving about 3 x 10^{-3} units of susceptibility per percent of magnetite for magnetite contents of less than 5 per cent, have been obtained e.g. from Finnish Precambrian rocks by Puranen et al. (1968), from Quebec serpentinites by Gaucher (1965) and from Minnesota iron formation by Mooney and Bleifuss (1953). These authors also discuss many aspects of magnetic anomaly interpretation in Precambrian regions.

Theoretical and laboratory studies on the susceptibility of dispersed magnetite (e.g. Strangway, 1967) show that 3 x 10^{-3} Gauss/Oe/volume per

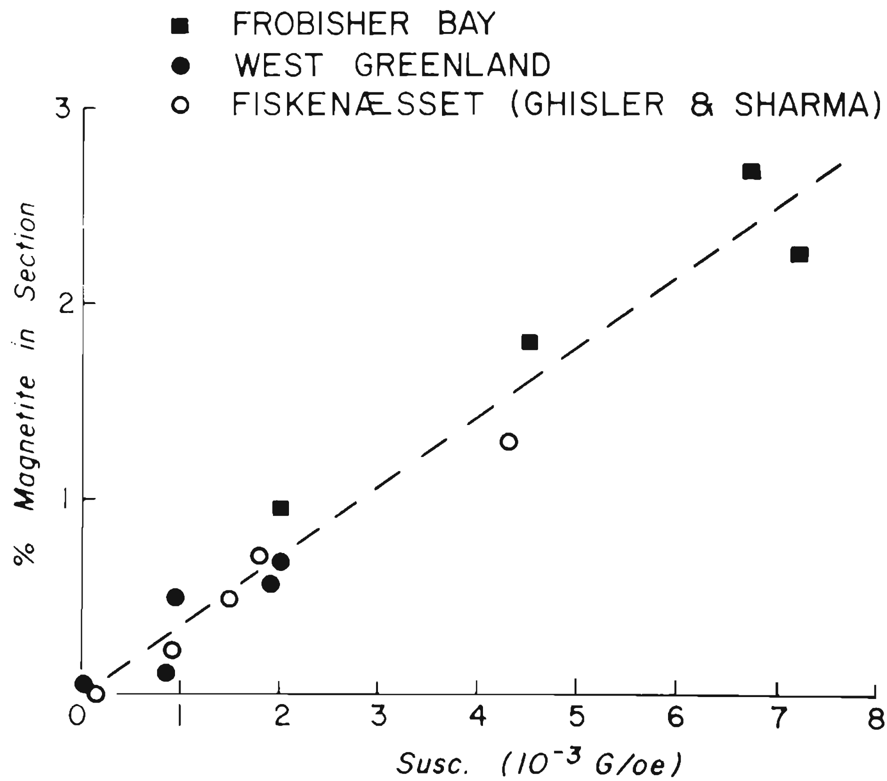


Figure 6. Relation between volume percentage of magnetite and initial volume susceptibility for gneiss samples from the Baffin Bay coast. Each entry represents one sample. The broken line is a least-squares approximation, assuming direct proportionality between the variables.

cent is definitely an upper bound for the susceptibility of rocks containing magnetite as the only magnetic mineral, especially so in fine-grained or magnetite-rich rocks.

We conclude that where magnetic anomalies of the order of hundreds of gammas occur over Precambrian basement formations in the Baffin Bay area, they are most likely due to induced magnetization in local concentrations of magnetite, of the order of a few per cent by volume.

SUMMARY

The results presented above show that magnetic property studies in the Baffin Bay area may aid in the mapping of its geological structure in three ways. First, the study of magnetic polarities in the Tertiary and other igneous rocks is a valuable stratigraphic tool; secondly, magnetic pole positions derived from these rocks may furnish key information on continental displacements; thirdly, the study of magnetic remanence and susceptibility is a relatively inexpensive way of obtaining important constraints on the interpretation of costly magnetic surveys. So far, however, only limited information of this nature is available from the Baffin Bay area, though the results are promising.

ACKNOWLEDGMENTS

We wish to thank: the Ministry of Greenland for permission to carry out field work in Greenl and; the Arctic Station in Godhavn and the Federal Electric Company for providing logistical assistance during field work; Mr. W.J. Drodge and Mr. R. P. Kennedy for invaluable help in field and laboratory work; Dr. D.B. Clarke (Dalhousie University), Mr. E. Fundal (Godhavn Observatory) and Dr. J.S. Sutton (Memorial University) for providing rock samples; Mr. P.W. Bullock (B.M.R., Australia), Dr. W. Fahrig (Geological Survey of Canada), Dr. G.S. Murthy (Memorial University) and Dr. D.H. Tarling (Newcastle University) for providing unpublished data included in Figure 5; Memorial University of Newfoundland and the North Atlantic Treaty Organization for financial assistance to one of us (L.G.K.); and the National Research Council of Canada who supported this research through Grant A-1946.

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34. MAGNETIC AND GRAVITY SURVEYS IN THE ITIVDLE VALLEY OF NÛGSSUAQ, WEST GREENLAND*

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Abstract

Ground magnetic and gravity surveys were made in the Itivdle valley of Nûgssuaq in West Greenland as part of the Geological Survey of Greenland (G.G.U) field program to study the thickness of marine Cretaceous-Tertiary sediments and the configuration of the underlying basement in that area which would bear on the assessment of its potential for oil and gas prospects.

Tedious operational and computational problems were involved in the reduction and correction of the observed geophysical field data because of the rugged and undulating topography, the disturbing effect of basaltic masses and to the frequent occurrence of magnetic storms in the surveyed area.

Maps of the observed magnetic and gravity anomalies in the Itivdle valley are presented. Due to the rather small lateral extent of the area, the data are too meagre for making reliable depth estimates for the basement, but the magnetic data indicate that the sediments may be around 1,500 m or more thick in the southeast part of the valley. Seismic work along key profiles is suggested which may provide a better picture of the basement configuration.

INTRODUCTION

The recent increase in geophysical exploration activity in offshore eastern Canada has provided an incentive for initiating similar geophysical work along the west coast of Greenland. This incentive coupled with the encouraging geological data compiled by the Geological Survey of Greenland (Rosenkrantz and Pulvertaft, 1969; Henderson, 1969) on the West Greenland basin has led to the commencement of geophysical programs in that area.

As a first step, in the summer of 1969 magnetic and gravity surveys were made in the Cretaceous-Tertiary sedimentary basin of Nûgssuaq with the objective of estimating the thickness of sediments and the configuration of the basement in that area.

This paper presents the results of magnetic and gravity surveys made in the Itivdle valley (northwest part of the peninsula Nûgssuaq, see Fig. 1) which has a rather small lateral extent because of the presence of basaltic mountains on both sides of the valley.

*Publication authorized by the Director of the Geological Survey of Greenland.

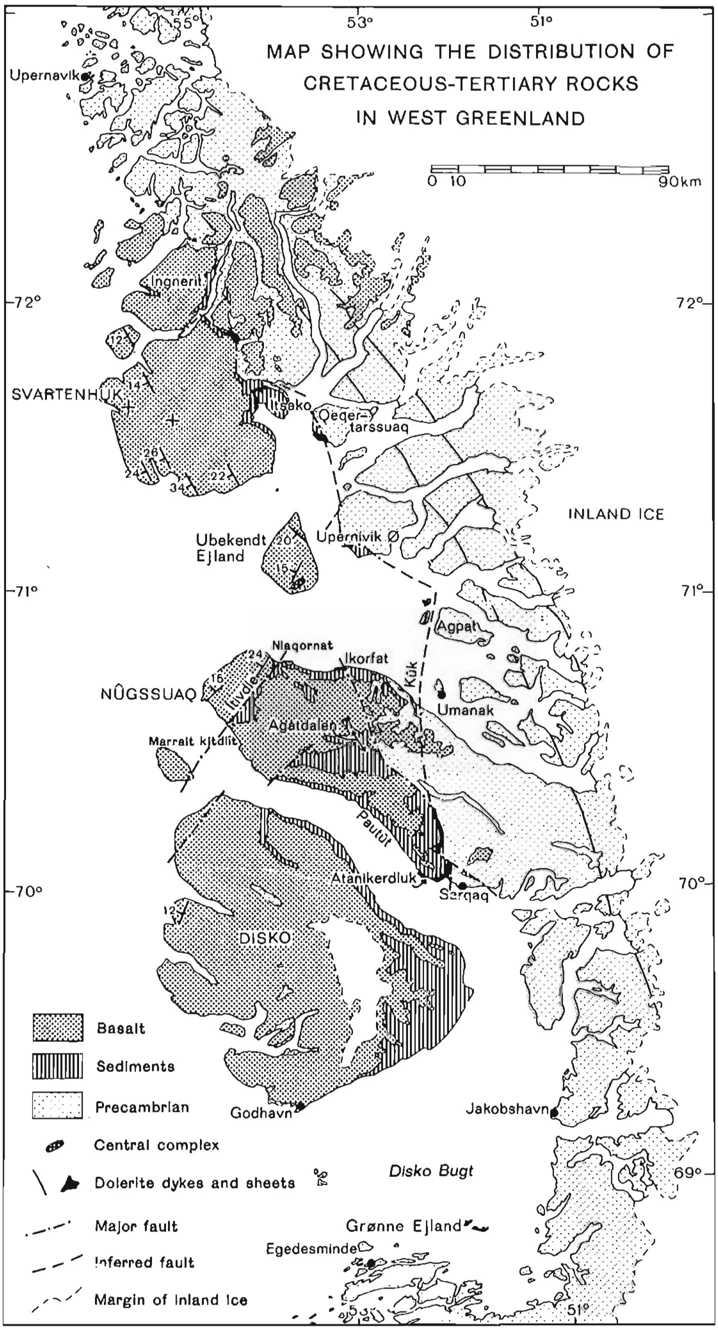

Figure 1. Map showing the distribution of Cretaceous-Tertiary rocks in West Greenland (after Rosenkrantz and Pulvertaft, 1969).

Geological Setting of the Itivdle Valley

The geological setting of the area has been described in detail by Henderson (1969). Nevertheless, it might be worth summarizing the main, points concerning the geology of the Nûgssuaq area for a better understanding of the implications of the geophysical results.

Figure 1 shows the distribution of the Cretaceous-Tertiary rocks in the West Greenland basin. The sediments outcropping onshore in this basin are of marine and nonmarine origin and range in age from Cretaceous to lowermost Tertiary. The thickest section of marine sediments exposed are seen along the north coast of Nûgssuaq where the total thickness is estimated to reach a maximum of 2000 m (Henderson, 1969). One important fault on the north coast of Nûgssuaq shows a displacement of 900 m with respect to the marine sediments with a downthrow to the southwest. The sedimentary succession throughout the area consists mainly of sandstones and shales.

In a considerable part of the area (including the sides of the Itivdle valley), the sediments are overlain by Tertiary basalts which locally attain a thickness of over a thousand metres and prevent the effective application of geophysical methods in the area. Also dykes and sills which occur in lavas and sediments cause local disturbances in the geophysical anomalies.

Geophysical Field Surveys in the Itivdle Valley

Ground magnetic and gravity surveys of an area 6 km long by 4 km wide of the Itivdle valley were carried out. The topography of the survey area is very undulating, the elevation differences from one part to another being as much as 450 m. Tedious operational and computational problems were anticipated beforehand in the reduction and correction of the field data, due mainly to the rugged and undulating topography, disturbing effects of basaltic masses and to the frequent occurrence of geomagnetic field disturbances in this Arctic area.

Notwithstanding these difficulties, a network of about 100 stations with an average spacing of 0.5 km was laid out. The gravity measurements were made using a Worden gravimeter and the vertical component of magnetic field intensity was measured using an Askania (GFZ) torsion magnetometer.

During the greater part of the field season in the summer of 1969, diurnal variations of the geomagnetic field during normal day working hours were of extremely large magnitude. Figure 2 shows the magnitude of the diurnal variation in the vertical component of the geomagnetic field as recorded by the base station magnetometer on July 14, 1969. Such disturbances appeared to be more frequent on bright sunny days, so that the field measurements were restricted to evening hours (from 6 p.m. until midnight) when the diurnal variations were observed to be much quieter in general.

Interpretation of Magnetic and Gravity Maps of the Itivdle Valley

Figure 3 shows the vertical magnetic intensity (ΔZ) anomaly map of the Itivdle area which has been drawn using a contour interval of 25 gammas (γ). The ΔZ anomaly values are corrected for the regional geomagnetic gradient (using the 1969 International Geomagnetic Reference Field) but not for the undulations in the topography. Unfortunately none of the anomalies has been well-defined because of the relatively small lateral extent of the survey area, so it is not possible to make unambiguous estimates of basement depth. However, rough estimates of depth have been made using the two profiles AA' and BB' (see Fig. 3). Figure 4 shows the plot of the two magnetic profiles.

Considering first the profile AA', because the complete set of characteristic points commonly employed in depth calculation is not available, the straight-slope distance has been used to estimate a probable depth of 1,700 m. In addition, the constantly decreasing amplitude of the anomaly in the southeast direction may be indicative of the basement depth increasing in that direction, which is in conformity with the dips of exposed sediments along the northwest part of the valley.

The second magnetic profile BB' (see Fig. 4) appears to be indicative of a basement depression or a flexure in the magnetic basement. Using the sloping step model as a source for this anomaly, its depth can be roughly

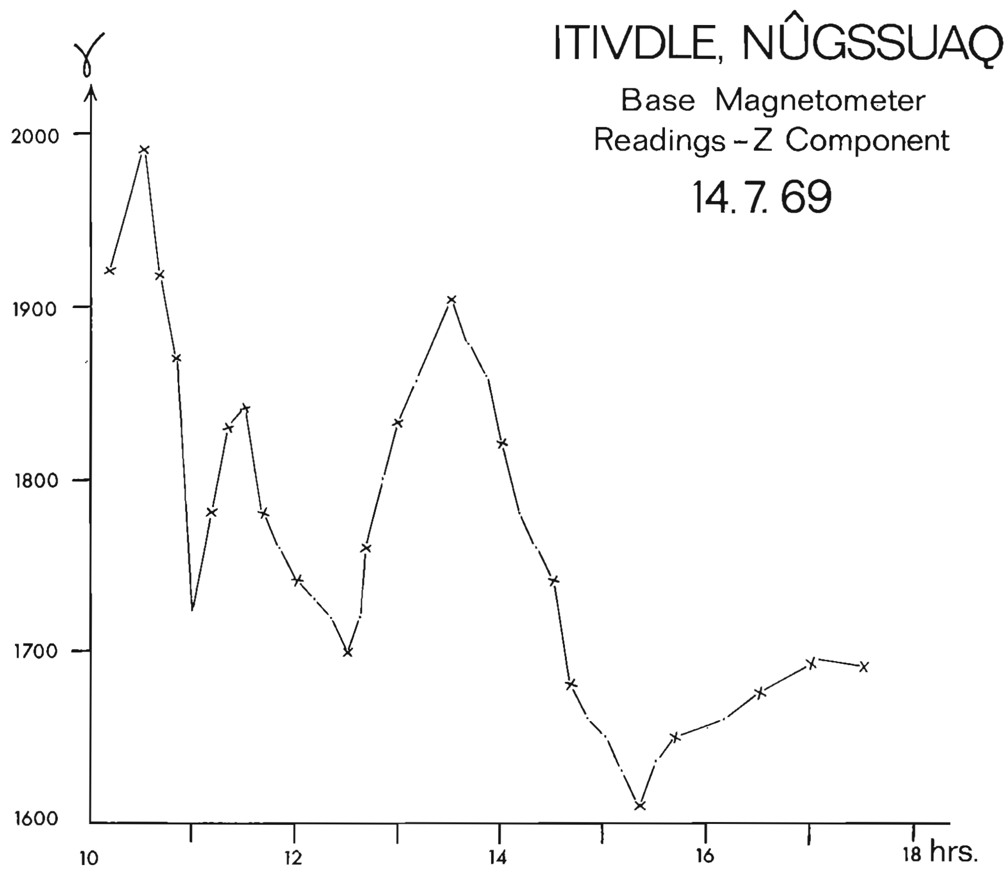

Figure 2. Diurnal variations in the vertical field intensity recorded at the base station in the Itivdle valley, Nûgssuaq during a magnetic storm on July 14, 1969.

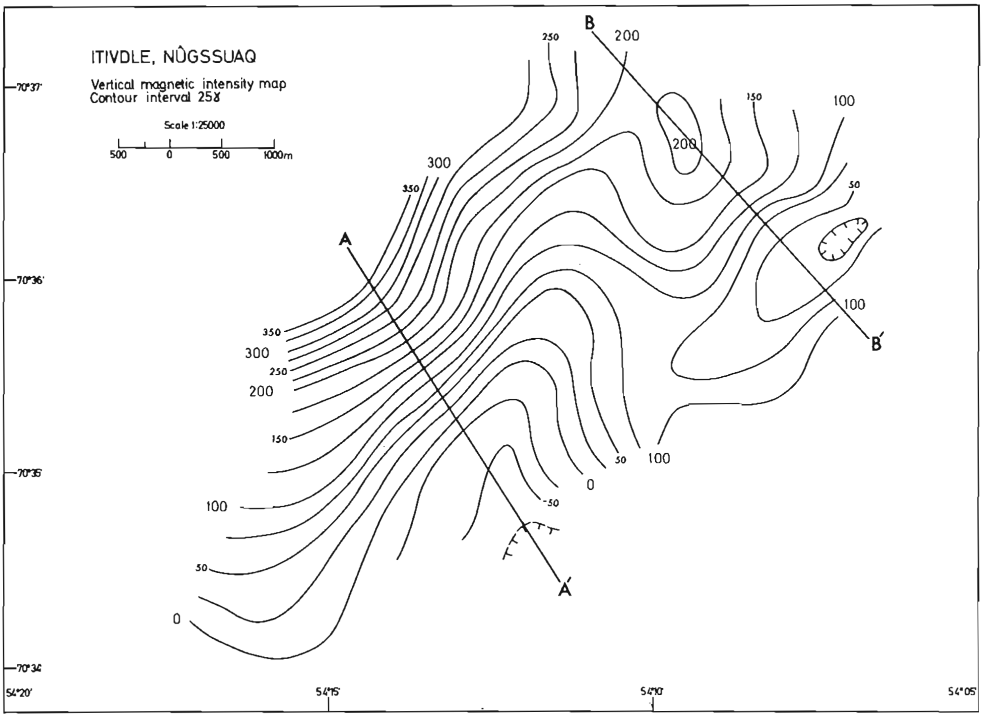

Figure 3. Vertical magnetic intensity map, Itivdle valley, Nûgssuaq, West Greenland.

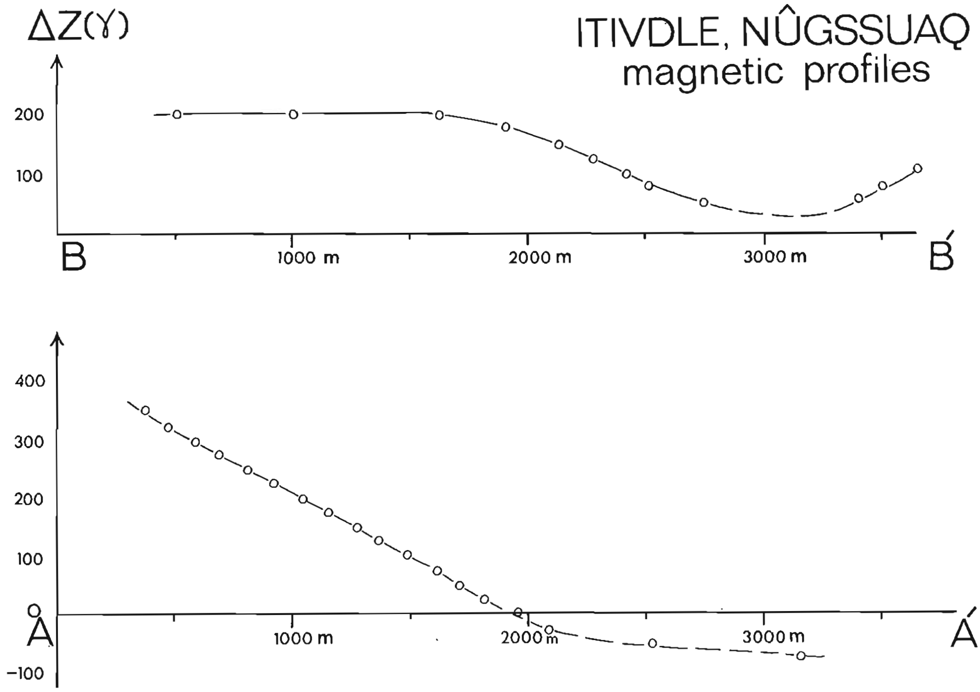

Figure 4. Magnetic profiles, Itivdle valley, Nûgssuaq, West Greenland.

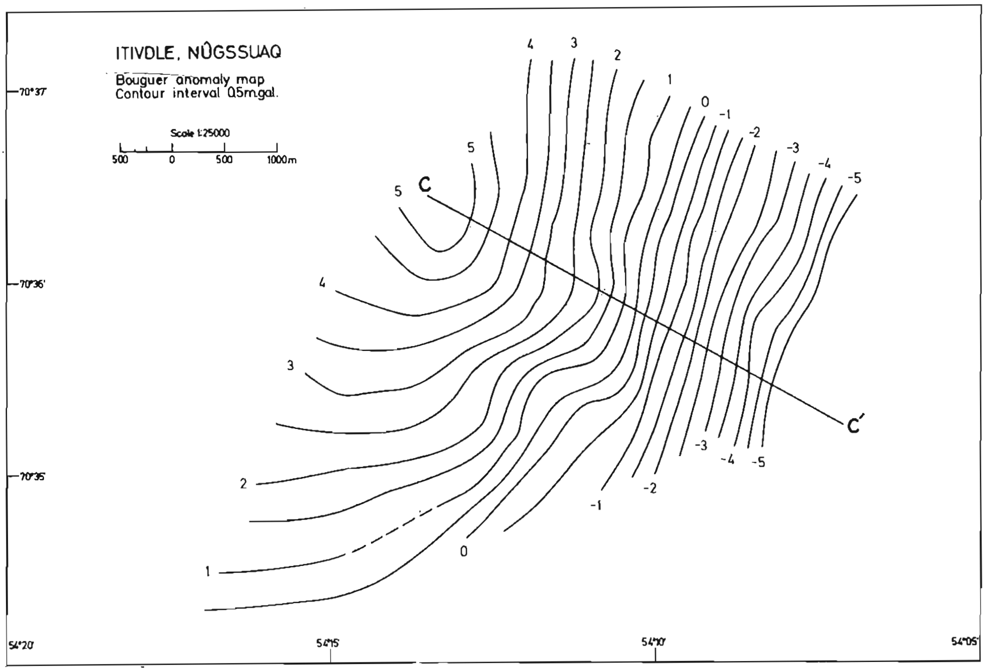

Figure 5. Bouguer gravity map, Itivdle valley, Nûgssuaq, West Greenland.

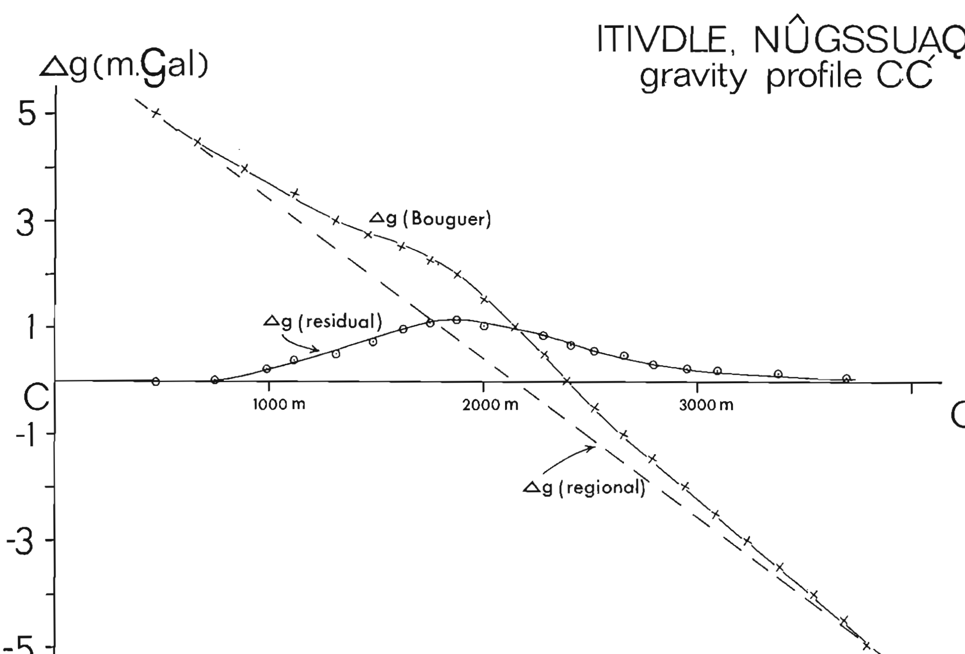

Figure 6. Gravity profile, Itivdle valley, Nûgssuaq, West Greenland.

estimated by the half-width of the anomaly and gives depth to the centre of the slope of 1,000 m, and a total basement relief of about 250 m.

In the above depth estimates, induced magnetization has been assumed for the granitic gneiss basement. Further, the various normal and reversely magnetized dykes which intrude in the sediments are too narrow (usually not more than 2 m in width) to account for the observed anomalies. However, the possibility of an intruding sill or plug as a causative body (over BB' in Fig. 3) cannot be ruled out.

Figure 5 shows the Bouguer gravity map of the same area of the Itivdle valley. The correlation with the magnetic map is poor. Although masked by a strong regional gradient in the northwest to southeast direction, the nosing of the contours is indicative of a dome-like excess mass whose edge is located at the -0.5 milligal contour of the gravity map. By removing the regional effect the anomaly can be better resolved and is found to have an amplitude of about one milligal. Figure 6 shows the residual anomaly obtained over the profile CC' (see Fig. 5). This interpretation is very speculative because the area of the coverage is much too small for analysis. The residual anomaly shows a half-width of about 600 m indicating shallow depth for the causative higher density feature, which might be a volcanic intrusive feature, although this interpretation is not supported by the magnetic results.

CONCLUSIONS

In conclusion, it should be emphasized that the gravity and magnetic survey data obtained so far in the Itivdle valley do not cover a large enough area for reliable depth estimates to be made. The best one can say at present is that the magnetic data indicate that the sediments may be approximately 1500 m or more thick in the southeast part of the valley.

More geophysical survey data is needed to obtain a reliable picture of the depth and configuration of the basement. However, due to the disturbing effect of rather thick basaltic cover on both sides of the valley, an extension of gravity and magnetic surveys to the flanks would be inadvisable. Seismic surveys along key profiles should provide a better picture of the basement depth contours in the Itivdle valley.

ACKNOWLEDGMENTS

Sincere thanks are extended to Mr. G. Henderson of the Geological Survey of Greenland for supplying the necessary geological information and for his various suggestions in the initial planning of the geophysical work. Thanks are also due to all the collaborators who assisted in the field surveys.

The author is very much indebted to Prof. Henry Jensen of the University of Copenhagen for the loan of the Worden gravimeter, to Prof. A. Rosenkrantz, Mr. T.C.R. Pulvertaft and Dr. D.S. Parasnis for very stimulating and helpful discussions, and to Dr. S. Watt for the critical reading of the manuscript.

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35. LOW-LEVEL AEROMAGNETIC SURVEYS OF THE CONTINENTAL SHELVES BORDERING BAFFIN BAY AND THE LABRADOR SEA

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Abstract

Since 1962, the Geological Survey of Canada and the National Aeronautical Establishment have co-operated in joint low-level aeromagnetic surveys of the continental shelves and deep-ocean basins adjacent to Canada. Reconnaissance aeromagnetic profiles at about 60 mile intervals have been obtained from the southern tip of Greenland to the Kane Basin between Ellesmere Island and northern Greenland. Correlation of the anomalies on the profiles indicate that a triple-spreading junction existed in the Labrador Sea at approximately 56°50'N, 41°40'W. Moreover from a correlation of the anomalies with those on the flanks of the Reykjanes Ridge it is deduced that sea-floor spreading terminated in the Labrador Sea just after Anomaly 13 was generated some 38 million years ago. Several transform faults are indicated in the Labrador Sea.

Over the continental shelves, there is a marked change of character in the aeromagnetic profiles some tens of miles from shore. The anomalies are relatively sharp close in to shore and then quite abruptly the wavelength of the anomaly increases and the amplitude decreases. This change is due to a sudden increase in the depth to the crystalline basement, and therefore marks the edge of the wedge of sedimentary rocks which underlie the outer part of the continental shelf. The double asymmetrical anomaly which marks the transition from continental to oceanic rocks is also recognizable on most profiles enabling this boundary to be delineated.

Depth determinations on the profiles indicate that the thicknesses of sedimentary rocks on the continental slopes and rises exceed 20,000 feet over wide areas. On the Labrador Shelf, the sediments extend all the way along the outer part of the continental shelf and rise deepening towards Hudson Strait and are underlain to a large extent by oceanic rocks. Two sedimentary basins have been delineated on the Labrador Shelf; the Nain Basin contains more than 30,000 feet (9 km) of sedimentary rock. The Saglek Basin contains in excess of 50,000 feet (15 km) of sediments. In both basins the sediments would appear from the sea-floor spreading sequence to be mostly of Mesozoic age making these areas attractive for the petroleum industry to prospect.

In central Baffin Bay the magnetic anomalies are of low amplitude but anomalies up to 50 gammas in amplitude having a wavelength of 20 km are readily discernible in the central part indicating that the basement rocks are oceanic. Two sediment-filled grabens have been delineated in the northern part of Baffin Bay. The Melville Bay graben is at least 200 miles (320 km) long, some 35 miles (56 km) wide and is filled with more than 15,000 feet (4.6 km) of sediments. The Baffin Shelf graben is about 350 miles long, 30

miles (48 km) wide and contains a minimum of 20,000 feet (6 km) of sediments. Basalt extrusives cover large areas on both sides of Davis Strait and extend as far south as Frobisher Bay presumably over a thick sedimentary cover. Sedimentary cover in the central deep-ocean part of Baffin Bay appears to exceed 20,000 feet (6 km) over large areas.

INTRODUCTION

Since 1962, the Geological Survey of Canada and the National Aeronautical Establishment have cooperated in a low-level aeromagnetic reconnaissance of the continental shelves and deep-ocean basins adjacent to eastern Canada. Aeromagnetic profiles at about 60 mile intervals have been obtained from the southern tip of Greenland to the Kane Basin between Ellesmere Island and northern Greenland.

Magnetic profiles have also been obtained in the Labrador Sea and Baffin Bay by the marine geophysics group at Bedford Institute of Oceanography, Dalhousie University, the Lamont Geological Observatory and by ships of U.S. naval agencies. High-level aeromagnetic profiles have also been flown by the Geomagnetic Division of the Earth Physics Branch and by Project Magnet. Except for the systematic survey carried out by the U.S. Naval Oceanographic Office in southern Labrador Sea during 1966 (Anonymous, 1970), the profiles obtained from the foregoing surveys are somewhat random so that the coverage is non-existent in many areas. The G.S.C. - N.A.E. surveys were however a systematic reconnaissance and go in an approximately orthogonal direction from one coastline to the other so that for a given area the lines are usually parallel to one another. The aeromagnetic profiles described in this article were obtained using the Northstar aircraft of the National Aeronautical Establishment which was equipped with an optical absorption magnetometer and the total magnetic field values were recorded both on digital and analog tape. The survey aircraft was usually flown at 1,000 feet above the ocean surface, except in a few cases where fog or poor visibility necessitated that the aircraft be flown at a somewhat higher or lower elevation for safety reasons. Navigation was by astro fixes, Loran A and in more recent years by Omega. Navigation has been the most serious problem encountered and has been of constant concern throughout the surveys. Diurnal monitor stations were usually established at each of the base of operations which were Goose Bay, Frobisher Bay and on the Greenland side, Narsarssuaq, Sondrestrom and Thule.

Magnetic Profiles across the Labrador Sea

Figure 1 shows most of the total intensity profiles obtained across the Labrador Sea. The presence of oceanic-type magnetic anomalies and their correlation over reasonably large distances was initially established by two low-level lines across the Labrador Sea from the Labrador Coast to the southern tip of Greenland and back to the Strait of Belle Isle which were flown by an Argus aircraft in February 1963 (Hood and Godby, 1964). With the indication that there was magnetic striping in the Labrador Sea a more extensive program was commenced both to delineate the configuration of the magnetic stripes and also to ascertain where large thicknesses of sedimentary rocks existed on the adjacent continental shelves which would be of

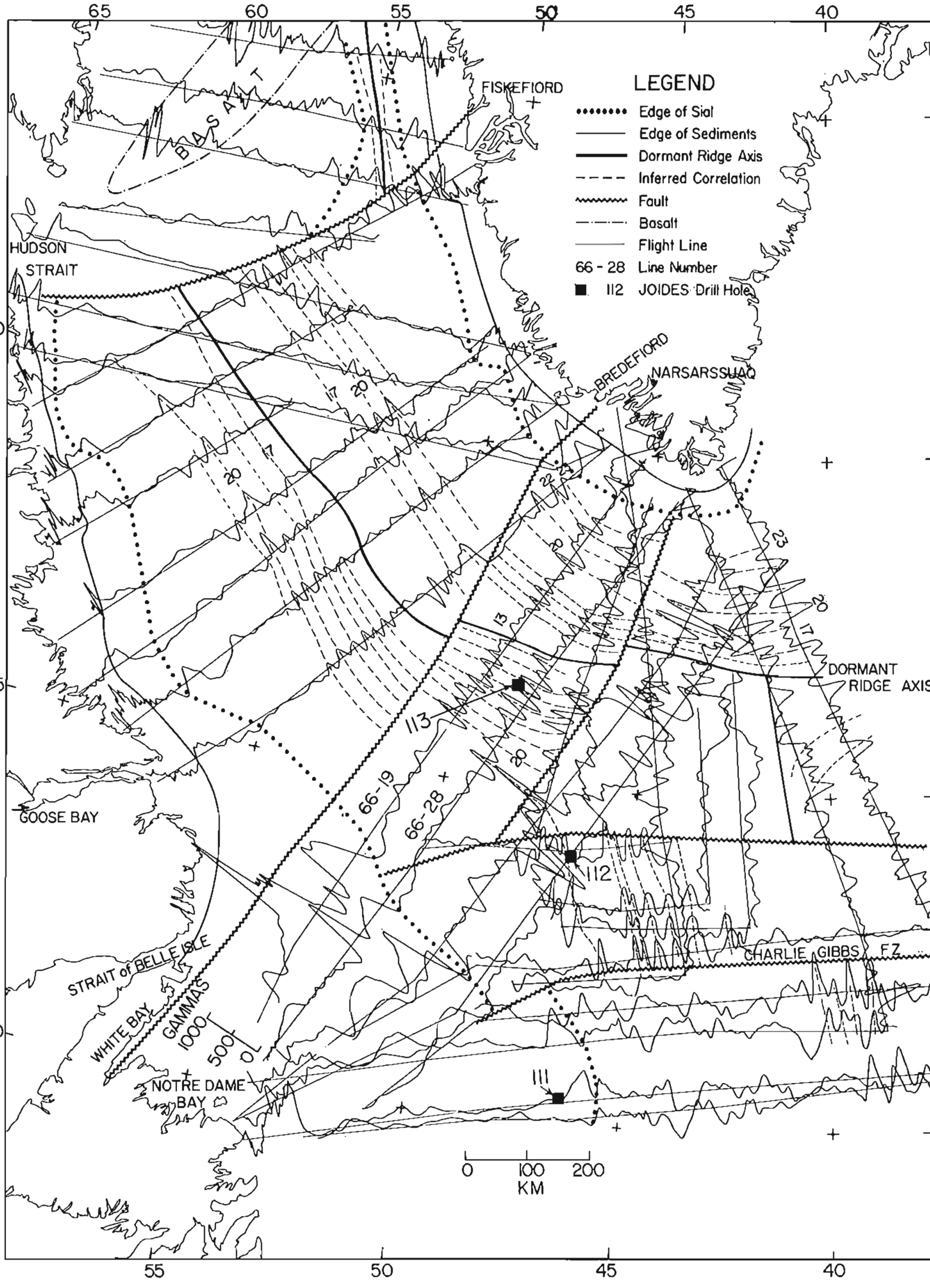


Figure 1. Aeromagnetic profiles in the Labrador Sea.

interest to the petroleum industry. This article is mainly concerned with the latter topic; however, the separation of Greenland from North America is of interest in the search for hydrocarbons because the age of the sedimentary column will be determined by the timing of the sea-floor spreading sequence.

On Figure 1, the inferred correlation between the flight lines has been indicated by dashed lines. In central Labrador Sea, the correlation does not differ significantly from that given by Godby et al. (1966), and in that earlier article there was some indication that the correlation could be extended around the southern tip of Greenland and northward towards Denmark Strait. Immediately south of Cape Farewell on the southern tip of Greenland the picture is somewhat confused and so additional surveys with a closer line spacing were carried out in this area in order to be more certain of the correlation. One of the main difficulties has been to make an unequivocal correlation of the anomalies with the Heirtzler geomagnetic polarity time scale (1968) in order to deduce the timing of the sea-floor spreading sequence in the Labrador Sea. To do this, we have carried the anomaly correlations from the Reykjanes Ridge southwards into the Labrador Sea, so that the numbers on the profiles in Figure 1 are intended to identify those anomalies on the Heirtzler time scale. The position of the inferred dormant ridge axis has been indicated on Figure 1, by a thick continuous line; and it coincides with a central U-shaped anomaly. There is a marked symmetry about this central U-shaped anomaly on a number of profiles and this is reasonably good evidence that sea-floor spreading has in fact occurred in the Labrador Sea and that it terminated at a time when the earth's magnetic field was reversed.

Figure 2 shows one of the profiles (66-19) obtained in the Labrador Sea (see Fig. 1 for location) and the same profile reversed. It can be seen that there is excellent symmetry on this particular profile about the central U-shaped anomaly which appears to mark the locus of the rift valley at the centre of the dormant ridge in the Labrador Sea (Le Pichon et al., 1971).

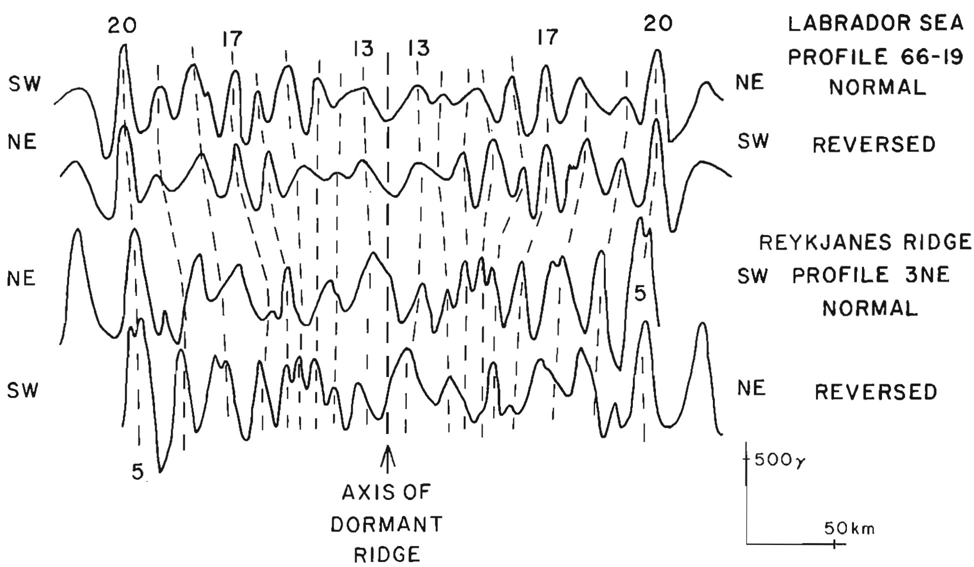


Figure 2. Comparison of Profile 66-19 with the same profile reversed and with an aeromagnetic profile across the Reykjanes Ridge.

Also presented on Figure 2 is a portion of the aeromagnetic profile across the Reykjanes Ridge which has been published earlier (Godby et al., 1968). The presence of transform faults make it rather difficult to be certain of the correlation from the Reykjanes Ridge into the Labrador Sea, but we conclude that the sea-floor spreading in the Labrador Sea terminated just after Anomaly 13 on the Heirtzler geomagnetic polarity time scale, which would correspond to the end of the Miocene about 38 million years ago. As Figure 2 indicates, there is a reasonably good correlation between the Labrador Sea profile 66-19 and the Reykjanes Ridge profiles which, if correct, would mean that a double ridge system existed in the North Atlantic Ocean which became dormant after Anomaly 13 was generated. It would also mean that no spreading occurred in the North Atlantic Ocean between Anomaly 5 (9 m.y.) and Anomaly 13 (38 m.y.). In the Labrador Sea, anomalies can be recognized with certainty as far back as Anomaly 24 which represents 60 m.y. B.P. The thick sediments near the continental margins appear to contribute to the difficulty in the recognition of earlier anomalies.

The conclusion that sea-floor spreading in the Labrador Sea ceased 38 million years ago receives support from the continuous seismic profiler results published by Jones et al, (1970) and Le Pichon et al. (1971), and from the Joides drilling results in the southern Labrador Sea (Laughton et al., 1972). Two prominent reflectors can be traced on the seismic profiles across the Labrador Sea and terminate on the sides of the mid-Labrador Sea ridge. The Joides drilling has enabled the upper reflector to be identified as uppermost Eocene to Lower Oligocene so that the inferred date of cessation of spreading would indicate that the reflector probably marks the boundary between sediments of Eocene and Oligocene age, that is the 40 m.y. isochron.

Consideration of the pattern formed by the dashed lines on Figure 1 leads to the conclusion that a triple-spreading junction has existed south of Cape Farewell at approximately 56°50'N, 41°40'W. Thus sea-floor spreading would have taken place in three directions about the triple junction which together with the presence of transform faults makes the magnetic anomaly pattern difficult to unravel in that area. The locus of the dormant Labrador Sea ridge can be traced around the southern tip of Greenland using the other aeromagnetic profiles obtained in the North Atlantic Ocean and appears to join with the dormant ridge earlier postulated (Godby et al., 1968), which formed part of the double ridge system generated in the North Atlantic during the late Mesozoic and early Tertiary period.

Several transform faults are apparent on Figure 1. At the north end of the Labrador Sea, a transform fault appears to run between Fiskefiord on the Greenland side and the south side of Hudson Strait. Thus the Hudson transform fault offsets the magnetic striping in the Labrador Sea in a right lateral sense into the Davis Strait. A basement ridge has been noted by Le Pichon et al. (1971) at about 63°30'N, 55°30'W (see also Johnson et al., 1969) and this ridge has the morphology characteristic of a fracture zone. A second transform fault runs from Bredefiord on the Greenland coast across the Labrador Sea towards Hawke Saddle between Belle Isle and Hamilton Banks on the Newfoundland shelf (see Canadian Hydrographic Chart 813, 1972) and may possibly join with the Cabot fault which runs along the west side of White Bay. The Cabot fault is clearly traceable out to sea on GSC aeromagnetic maps (7349 G, 7357 G and 7366 G). We would therefore suggest the name Cabot transform fault for this feature. On the Greenland side the transform fault is marked by what appears to be a very linear graben, namely

Bredefiord, and the Greenland coast itself appears to have been offset. Off the Greenland shelf, the transform fault is detectable from the broad lows which appear on the aeromagnetic profiles. The Cabot transform fault offsets the dormant ridge axis in a left lateral sense a distance of 25 miles (40 km) approximately. It is also apparent on the continuous seismic profiles published by Le Pichon et al. (1972, see Fig. 3) at approximately 56°N, 51°10'W and at 55°N, 52°20'W. These points correspond to the basement highs at 1680 km and at 1350 km approximately on Figure 3 in Le Pichon et al. (1971) which are on the southeast side of basement lows. Both these basement highs coincide with magnetic lows. Further towards Newfoundland the transform fault offsets the broad magnetic high and low that strike in a northwest direction parallel to the Newfoundland continental margin in a left-handed sense. This offset of magnetic contours is best seen on Figure 2 of Fenwick et al. (1968) at about 53°30'N, 53°W. Profile 151 of Sheridan and Drake (1968) on the Newfoundland Shelf shows a steepening of the basement rocks at about 53°30'N, 54°30'W which may be due to the transform fault (see also Fig. 15 of Grant, 1972). An earthquake occurred on 3rd August 1962 at approximately 52°N, 54°20'W (Milne and Smith, 1963), which is close to the locus of the transform fault. The Charlie Gibbs fracture zone is also apparent on Figure 1. It has been mapped to the edge of the Newfoundland shelf by Olivet et al. (1970) using a continuous seismic profiling system. The dormant ridge is offset in a left-handed sense approximately 360 km, which is almost exactly the same distance that the Charlie Gibbs fracture zone offsets the mid-Atlantic ridge, so that the dormant ridge is offset to a position which is in alignment with the Reykjanes Ridge.

The locus of the inferred contact between continental and oceanic rocks has also been drawn on Figure 1 using a dotted line; the determination of this boundary is discussed in a later section of this paper. The position of Joides Hole 111 on the Orphan Knoll and Holes 112 and 113 in the Labrador Sea (Laughton et al., 1972) are also located on Figure 1.

Labrador Shelf

Examination of the aeromagnetic profiles across the Labrador Shelf which are shown in Figure 1 indicates that there is an abrupt change in character on each of the profiles some distance from the coast. The anomalies are relatively sharp close in to shore and some tens of miles from the coastline the wavelength of the anomalies increases and the amplitude decreases. This is demonstrated more clearly in Figure 3, which shows an aeromagnetic profile flown approximately at right angles to the Labrador Coast at about 58°N some twenty miles south of Saglek. At the top of the figure is the total intensity profile with the regional gradient removed. It can be seen that near the coastline at the left hand end of the profile, the anomalies have a noticeable high-frequency component. At a certain point, the high-frequency anomalies stop abruptly and the profile becomes quite smooth. This phenomenon is demonstrated in a much more conclusive manner by the digitally-filtered aeromagnetic profile in which only those frequency components of the anomalies having wavelengths between 0.185 to 1.85 kilometers, that is 0.1 to 1.0 nautical miles, have been allowed to pass. It should be noted that the gamma scales of the filtered and total intensity profiles in Figure 3 differ by a factor of 32. On the filtered profile the high-frequency hash, which is due to the close proximity of Precambrian crystalline rocks, stops abruptly at a

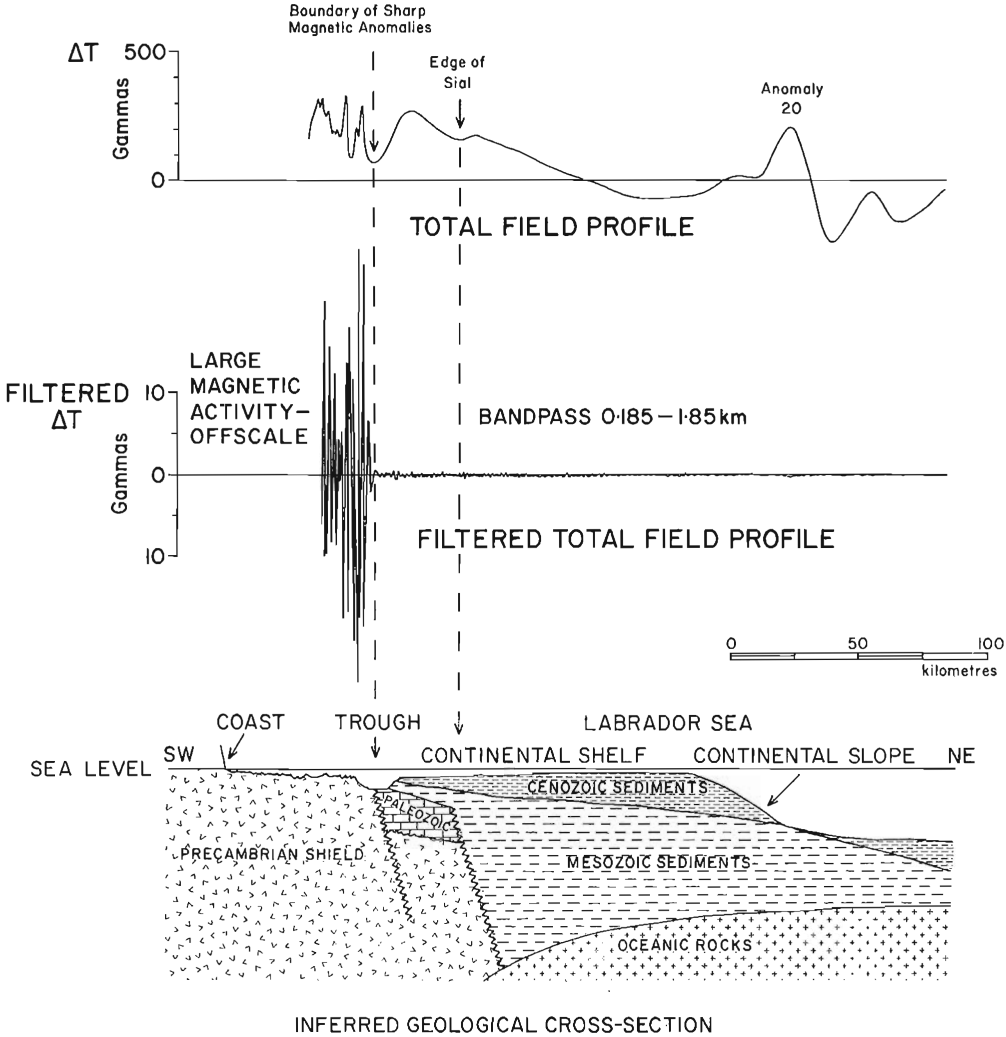


Figure 3. Aeromagnetic profile data and inferred geological cross-section across the Labrador Shelf near Saglek.

point some 70 km (43.5 miles) from the coast which also coincides with the trough running for many hundreds of miles parallel to the Labrador coast (Holtedahl, 1958), whose position is indicated on the geological cross-section at the bottom of Figure 3. The reason for the abrupt termination of the high-frequency anomalies must be that the crystalline basement begins to dip steeply at that point possibly due to faulting. Grant (1966) published an interpreted Sparker profile immediately to the south of this particular aeromagnetic profile at approximately 56°30'N which confirms that the contact between the Precambrian rocks of the inner shelf and the sediments of the outer shelf coincides with the position of the trough. The distinctive anomaly on the northeast end of the profile appears to be Anomaly 20 on the Heirtzler geomagnetic polarity time scale.

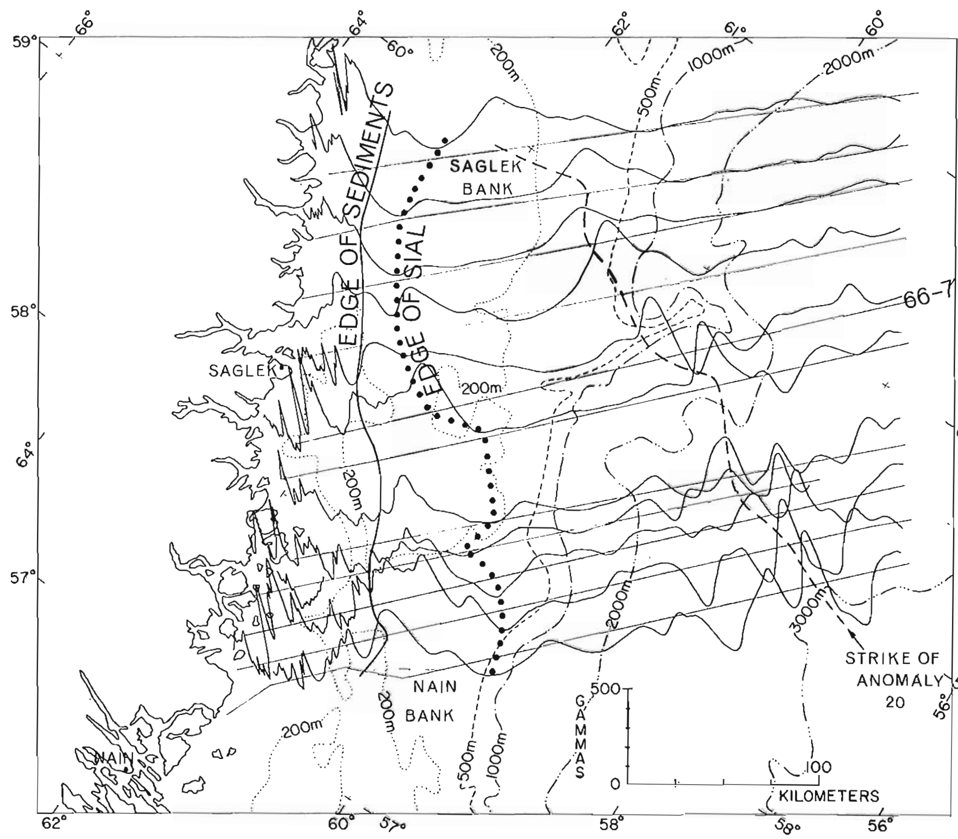


Figure 4. Low-level aeromagnetic profiles across the northern Labrador Shelf showing the inferred loci of the edge of the shelf sediments, the sial-sima boundary and Anomaly 20.

Actually this profile is one of a group of eleven low-level profiles obtained over the northern Labrador Shelf during 1966 and 1967 that reveal a most interesting situation (Figure 4). All the profiles which are spaced an average distance of 13 miles apart have the same characteristics as the one described in Figure 3, in that the high-frequency component of the anomalies terminates abruptly some distance from the coastline. The profile presented in Figure 3 is the fifth profile from the top of Figure 4. Thus on Figure 4, a line has been drawn to mark the locus of the edge of the sedimentary wedge where the Precambrian rocks are downfaulted. This locus agrees quite well with the same boundary interpreted from continuous seismic profiler results by Grant (1972; see Fig. 16), and with the position of the marginal trough described by Holtedahl (1958).

Another interesting feature on each of the profiles, is the double anomaly which appears to mark the transition zone from continental to oceanic rocks. The double anomaly consists of two highs of unequal amplitude separated by a minimum and is due to the magnetic effects of a relatively thin but strongly magnetic oceanic crust overlying a non-magnetic oceanic mantle abutting a less magnetic but much thicker continental crust. From a consideration of the studies by Fenwick et al. (1968, see Fig. 9) of the edge effect north of Newfoundland (see also Keen, 1969), the minimum between the two highs would appear to approximately designate the edge of the sial at high magnetic latitudes (see Figs. 5 and 6). The locus of the edge of the sial has also been drawn on Figure 4 using a dotted line, together with the solid line which marks the locus where the thick sedimentary sequence begins. Between these loci the Precambrian crystalline rocks are, of course, present as basement rocks at depth. The situation has been represented on the lower half of Figure 3 by a diagrammatic geological cross-section. Sedimentary rocks of Paleozoic age are inferred to overlie the downthrown Precambrian block because Grant (1972) has concluded that a thin layer of Paleozoic rocks underlies the inner continental shelf off southern Labrador and might extend northward to at least the transverse trough between Nain and Makkovik Banks. Their presence near Nain and near Cape Chidley (Bell, 1884; Roy, 1932; Little, 1936), on Akpatok Island in Ungava Bay (Cox, 1933), at the head of Frobisher Bay (see also GSC Map 1250A), coupled with the fact that Paleozoic rocks were dredged from the northern part of the Labrador Shelf by the USCGS MARION (Trask, 1932) and by McMillan (1972), would strongly indicate that Paleozoic rocks extend all the way up the Labrador Shelf, and moreover that they outcrop on the shelf. Sample 3 of McMillan (1972) which contained Devonian and Cretaceous pebbles was collected at 56°40'N, 60°09'W; it lies close to the western edge of the marginal trough. Grant (1972) also noted that formations giving good reflections occurred between the Precambrian rocks and the coastal plain sediments, and he concluded that these might represent either Paleozoic or Mesozoic rocks. Thus on Figure 3, Paleozoic and Mesozoic rocks have been shown as outcropping on the western side of the marginal trough in accordance with the foregoing field evidence. Presumably the Paleozoic sediments were deposited in a narrow graben which formed the proto-Labrador Sea prior to rifting. Mesozoic rocks have also been dredged from the continental slopes on both sides of the Labrador Sea (Johnson, pers. comm.; McMillan, 1972) so this fact has also been incorporated on Figure 3 in the geological cross-section.

On Figure 4 the linear magnetic anomalies west of the dormant ridge axis (see Fig. 1) can be correlated across the Labrador Shelf and the strike of Anomaly 20 has been drawn on the figure. The wavelength of the anomalies appears to increase going northward, which indicates that the Precambrian basement plunges to the north. This deduction is confirmed by depth determinations carried out on the anomalies (see Fig. 7). Thus the oceanic rocks of the Labrador Sea extend underneath the sediments of the northern part of the Labrador Shelf striking towards Hudson Strait. It would therefore appear that subduction of the oceanic rocks occurred at least during the latter part of the Tertiary off the northern coast of Labrador in a similar situation to that taking place off the west coast of North America at the present time.

Continental Shelf north of Newfoundland

Consideration of Figure 1 shows that a series of high amplitude magnetic anomalies runs approximately parallel and close to the edge of the continental shelf north of Newfoundland. These have been mapped in more detail by Fenwick et al. (1968). The highest amplitude composite anomaly occurs on Profile 66-19 which runs between the southern tip of Greenland and Notre

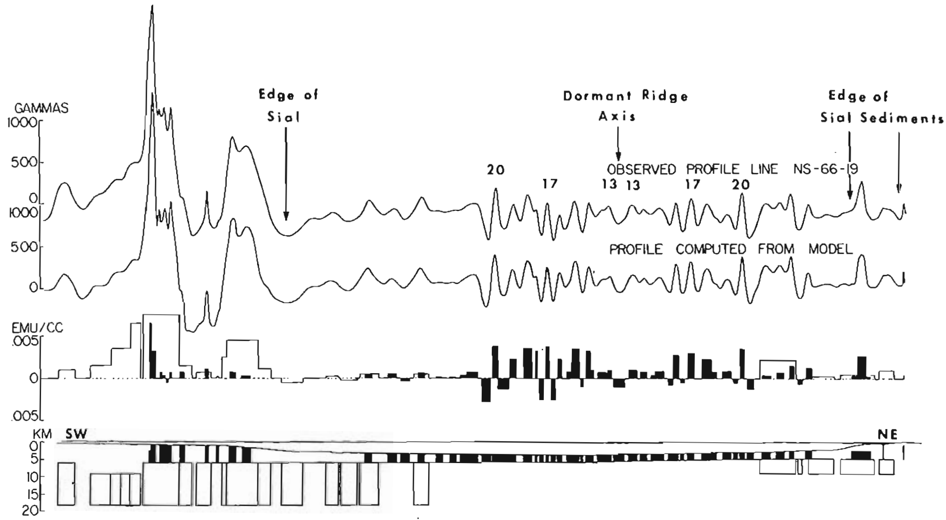


Figure 5. Interpretation of Labrador Sea Profile 66-19.

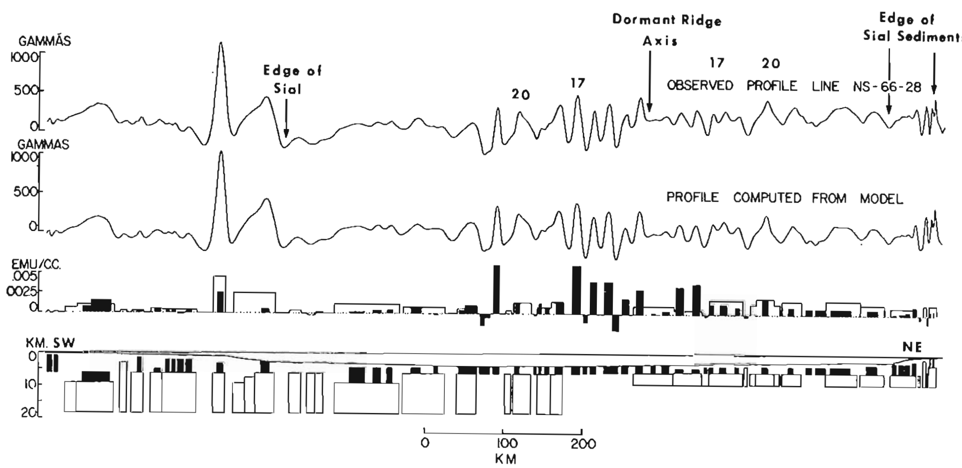


Figure 6. Interpretation of Labrador Sea Profile 66-28.

Dame Bay, Newfoundland. The profile is seen in more detail on Figure 5, and part of the profile was presented in Figure 2 to illustrate the excellent symmetry about the central U-shaped anomaly which exists on this profile. A quantitative interpretation of the profile has been carried out by computer matching by adjusting the widths and intensity of magnetization of individual blocks. It is assumed that the anomalies in the deep ocean are due to causative bodies 2 km thick which are at no great distance below the ocean floor. The values calculated for the intensities of magnetization of the various two-dimensional blocks are shown in the third profile from the top and these magnetization values also possess good symmetry about the dormant ridge axis.

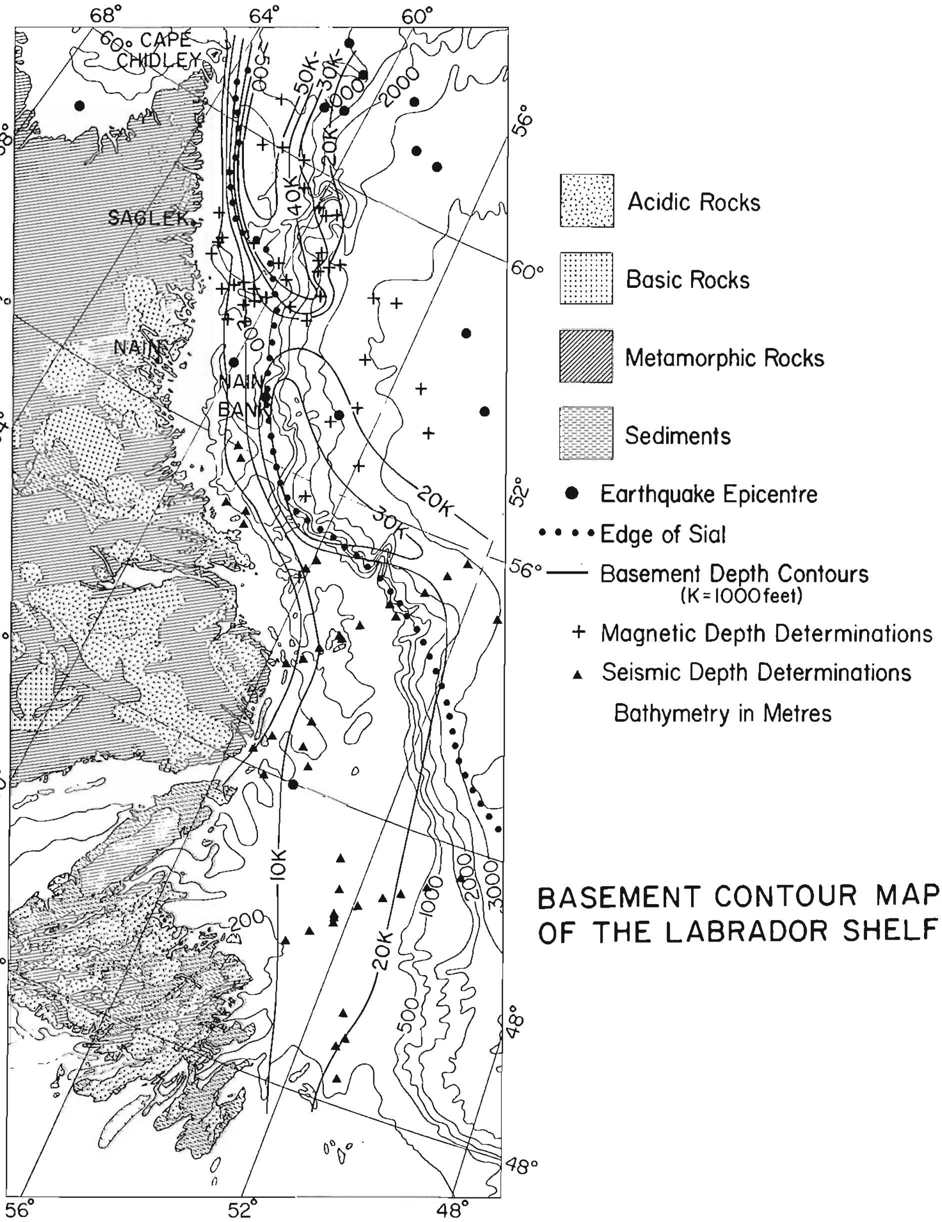


Figure 7. Basement contour map of the Labrador Shelf.

At the bottom of Figure 5 the geometric outlines of the causative bodies have been drawn, and it can be seen that two magnetic layers were required to produce the observed anomalies. Note that there are about 9 km i.e. approximately 30, 000 feet of sediments indicated on the Newfoundland Shelf in a basin-like feature. The previously mentioned high-amplitude anomalies on the left-hand side of the profile seem to be due to intrusive bodies probably diabase dykes which strike in a westerly direction towards southern Labrador. The position of the edge of the continent has been indicated at each end of Profile 66-19. Note the substantial thicknesses of sediments which exist on the continental rises of both Newfoundland and Greenland.

Figure 6 shows a quantitative interpretation carried out on Profile 66-28 immediately to the east of the previous one described. An axis of symmetry, the dormant ridge axis, in the deep-ocean part is less apparent. The dyke anomalies on the left-hand side of the figure are much more subdued than on Profile 66-19. About 20, 000 feet, that is 6 km of sediments is indicated on the Newfoundland Shelf and a lesser thickness on the continental rise along Profile 66-28.

Figure 7 shows the compiled results on the Labrador Shelf superimposed on a bathymetric map together with the generalized geology of Labrador. From the onshore geology, there is essentially no evidence that a thick sequence of young sediments exists offshore. The first solid line offshore from the Labrador coast (labelled OK) indicates the edge of the sedimentary formations and corresponds approximately with that given by Grant (1972, Fig. 16). The boundary of continental and oceanic rocks has been indicated by a dotted line; it coincides generally with that given by Grant for the Newfoundland Shelf and southern Labrador Sea, but differs in the northern Labrador Shelf. The sial-sima boundary of Grant swings away from the northern Labrador coast, and it is difficult to reconcile Grant's interpretation with the fact that the magnetic anomalies on the western side of the mid-Labrador Sea ridge can be traced under the continental shelf of northern Labrador.

The depth determinations were carried out on the magnetic anomalies using an automatic least-squares computer method of interpretation developed by McGrath and Hood (in press); the positions of these depth determinations have been indicated on Figure 7 by crosses. The depths obtained from the seismic refraction results published by Sheridan and Drake (1968) and Mayhew et al. (1970) are also indicated on Figure 7 by solid triangles. The results have been contoured at the rather coarse interval of 10, 000 feet (about 3 km) because the results of the present study do not warrant a greater accuracy. Two major sedimentary basins are thus delineated on the Labrador Shelf; the southerly one lies in close proximity to Nain Bank although the deepest part appears to underlie the continental slope. The northerly sedimentary basin on the Labrador Shelf underlies Saglek Bank and it extends northward towards Baffin Island. It would appear logical therefore to call these sedimentary basins respectively Nain and Saglek sedimentary basins. It is readily apparent that the deepest part of the sedimentary section lies seaward of the sial-sima boundary. In the Nain basin the sedimentary section exceeds 30, 000 feet (9 km), whereas the Saglek basin is much deeper being more than 50,000 feet (15 km). Note that the offshelf depths to igneous basement become shallower to the east indicating that the oceanic crust rises towards the mid-Labrador Sea ridge (see also Fig. 3).

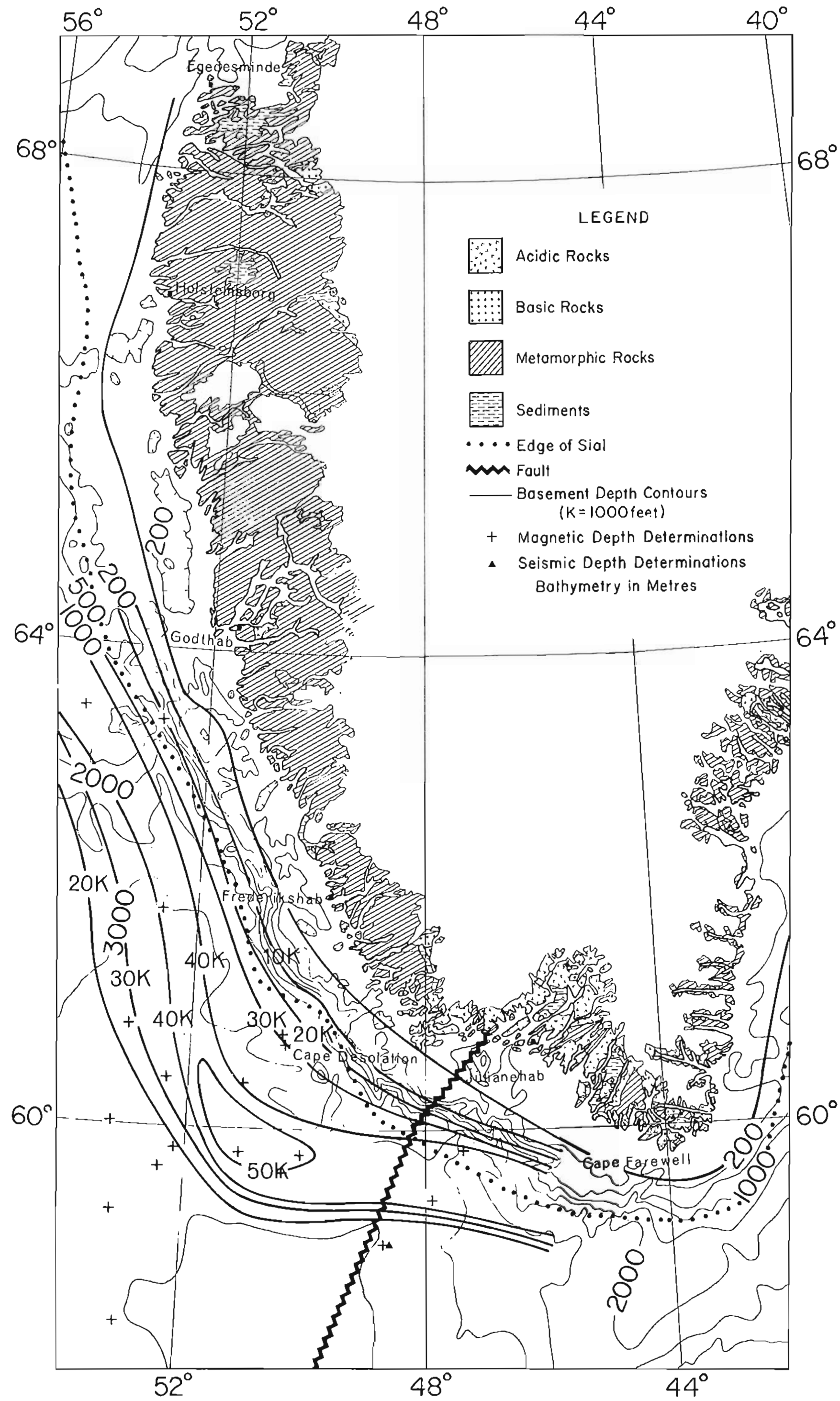


Figure 8. Basement contour map of the southwest Greenland Shelf.

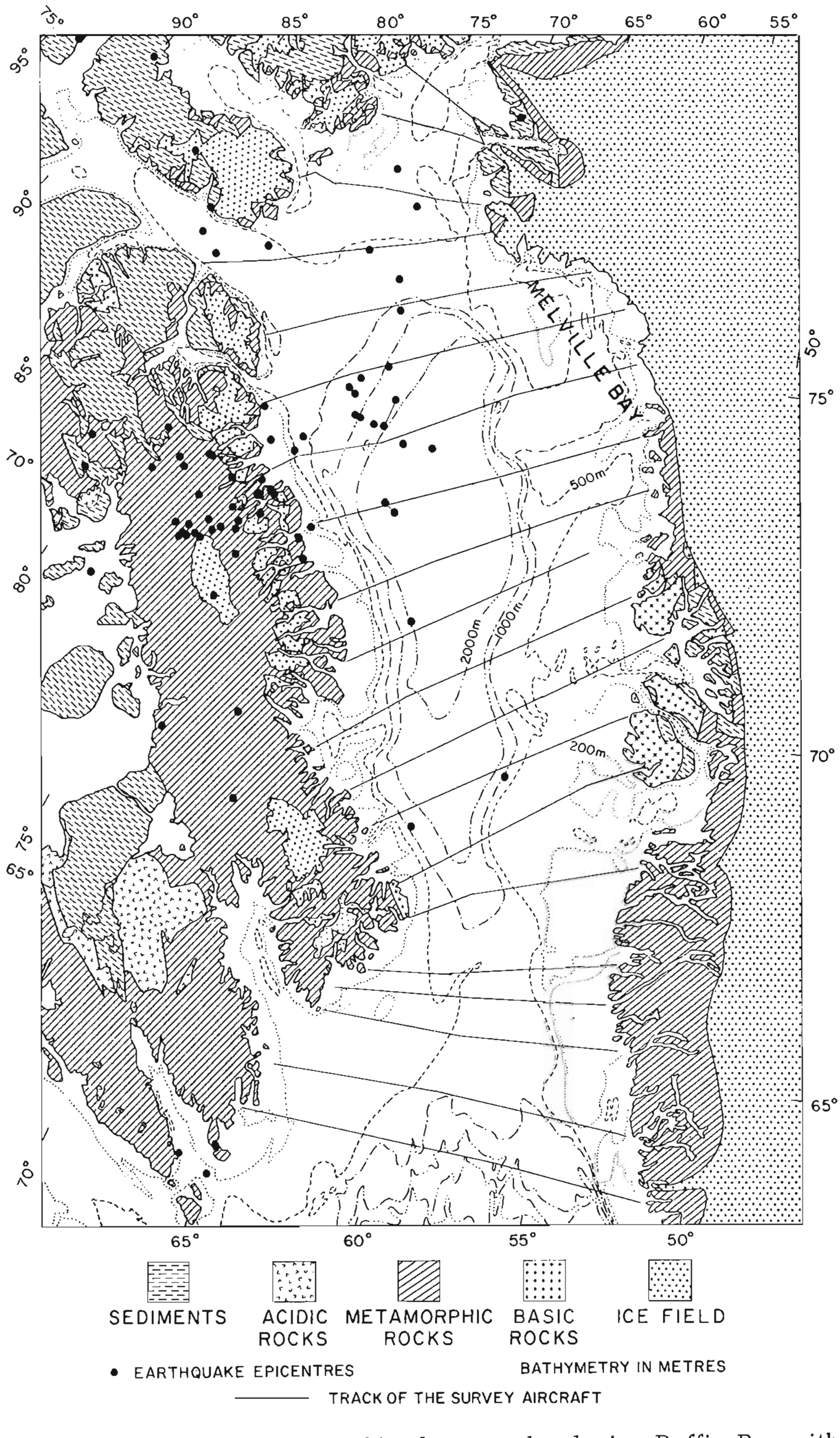


Figure 9. Generalized geology of landmasses bordering Baffin Bay with tracks of survey aircraft and earthquake epicentres.

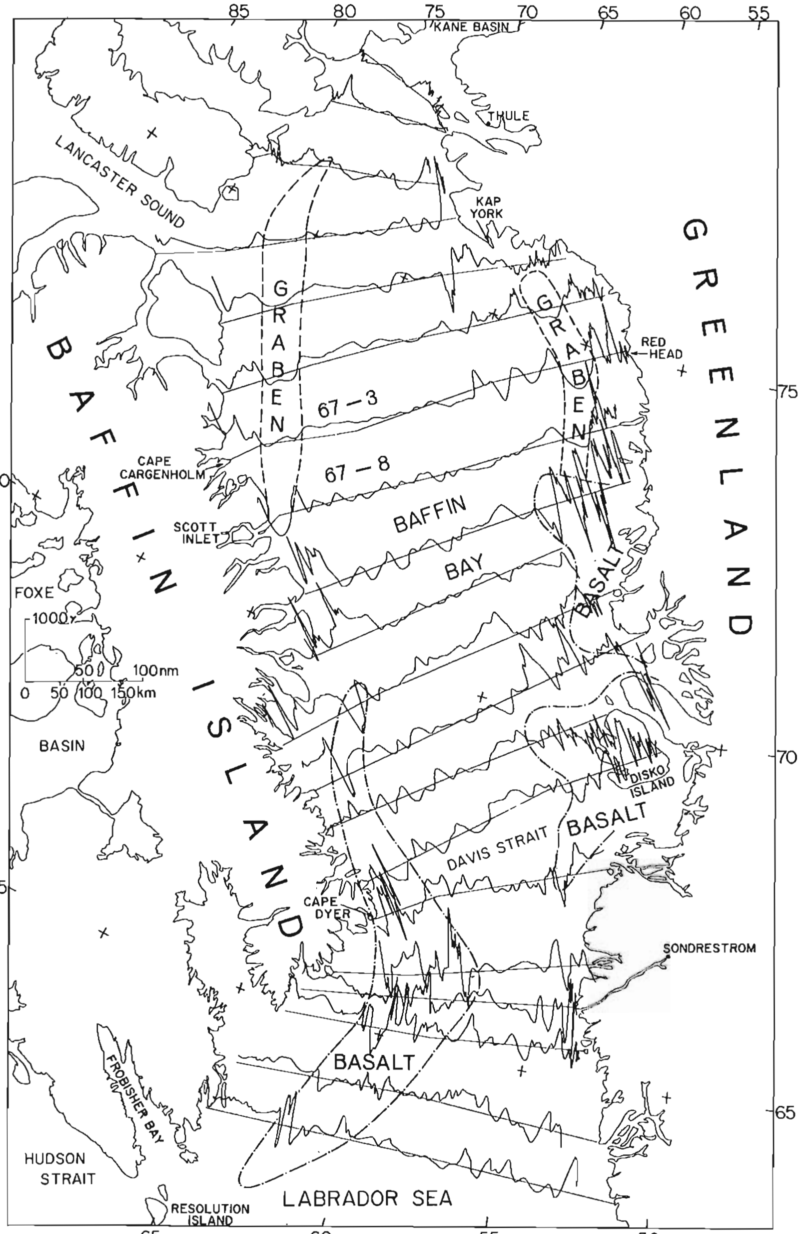


Figure 10. Aeromagnetic profiles obtained in Baffin Bay showing areas interpreted as being underlain by basalt together with the positions of two sediment-filled graben.

Continental Shelf of Southwest Greenland

Figure 8 shows the mainland geology and bathymetry of southwest Greenland which has been modified from Geological Survey of Greenland publications. The limit of the high-frequency anomalies is shown by the first solid line offshore from the Greenland coast, which marks the edge of the sediments on the southwest Greenland shelf. The inferred boundary between continental and oceanic rocks is shown by the dotted line. A number of depth determinations on the anomalies have been carried out using the technique of McGrath and Hood (in press) and the positions of these are indicated by the crosses. The sediments appear to form a belt running parallel to the continental slope and immediately seaward of it. The depth contours shown on Figure 8 are depths below sea level obtained from an interpretation of the aeromagnetic profiles so that the depth of water must be subtracted to obtain sediment thickness. Southwest of Cape Desolation more than 50,000 feet (15.2 km) of sediments are indicated at the base of the continental slope in a small sedimentary basin, which it would be appropriate to call the Desolation sedimentary basin.

Magnetic Profiles Across Baffin Bay

Figure 9 shows the aeromagnetic profiles obtained to date together with the bathymetry and generalized geology of the land masses bordering Baffin Bay. It can be seen from the figure that the land masses mostly consist of metamorphic rocks which are of Precambrian age except for the young basalts and sediments in the Cape Dyer and Disko Island areas. Positions of earthquake epicentres are also indicated by black dots. These have been obtained from a variety of sources particularly Dominion Observatory (now Earth Physics Branch) Bulletins. Some of the earthquake epicentres appear to fall in straight lines; for instance, several fall close to the faulted side of the half-graben in Lancaster Sound (Barrett, 1966).

Figure 10 shows the total intensity profiles obtained in Baffin Bay and these have had the regional gradient removed. The character of the profiles is essentially the same as the profiles obtained in the Labrador Sea; in general the anomalies are sharp and of high amplitude close to the coast and then at a certain point the high frequency component disappears and the profiles become much smoother. There are several features of particular interest. The first is the large U-shaped anomaly which bends around Melville Bay, and would appear to be due to a 56 km (35 mile) - wide graben filled with sediments. A similar but less distinct anomaly is apparent on the northern Baffin shelf which strikes across the entrance to Lancaster Sound.

The anomalies in the vicinity of Cape Dyer and Disko Island, where extensive sequences of Tertiary basalts occur, have a distinctive character which is quite different to that observed close to the coastline in other areas. The published GSC aeromagnetic maps of the Cape Dyer area (Map 7637G and 7659G) show very characteristic anomalies which are associated with each of the outcrops of Tertiary basalts mapped by Clarke and Upton (1971). The anomalies generally consist of a sharp positive high with negatives on either side and the total amplitude exceeds 1,000 gammas at 1,000 foot flight elevation in a number of cases. Thus the anomalies have a much larger amplitude and a tendency to oscillate sharply from positive to negative values in comparison to those occurring over Precambrian rocks, and this fact has been

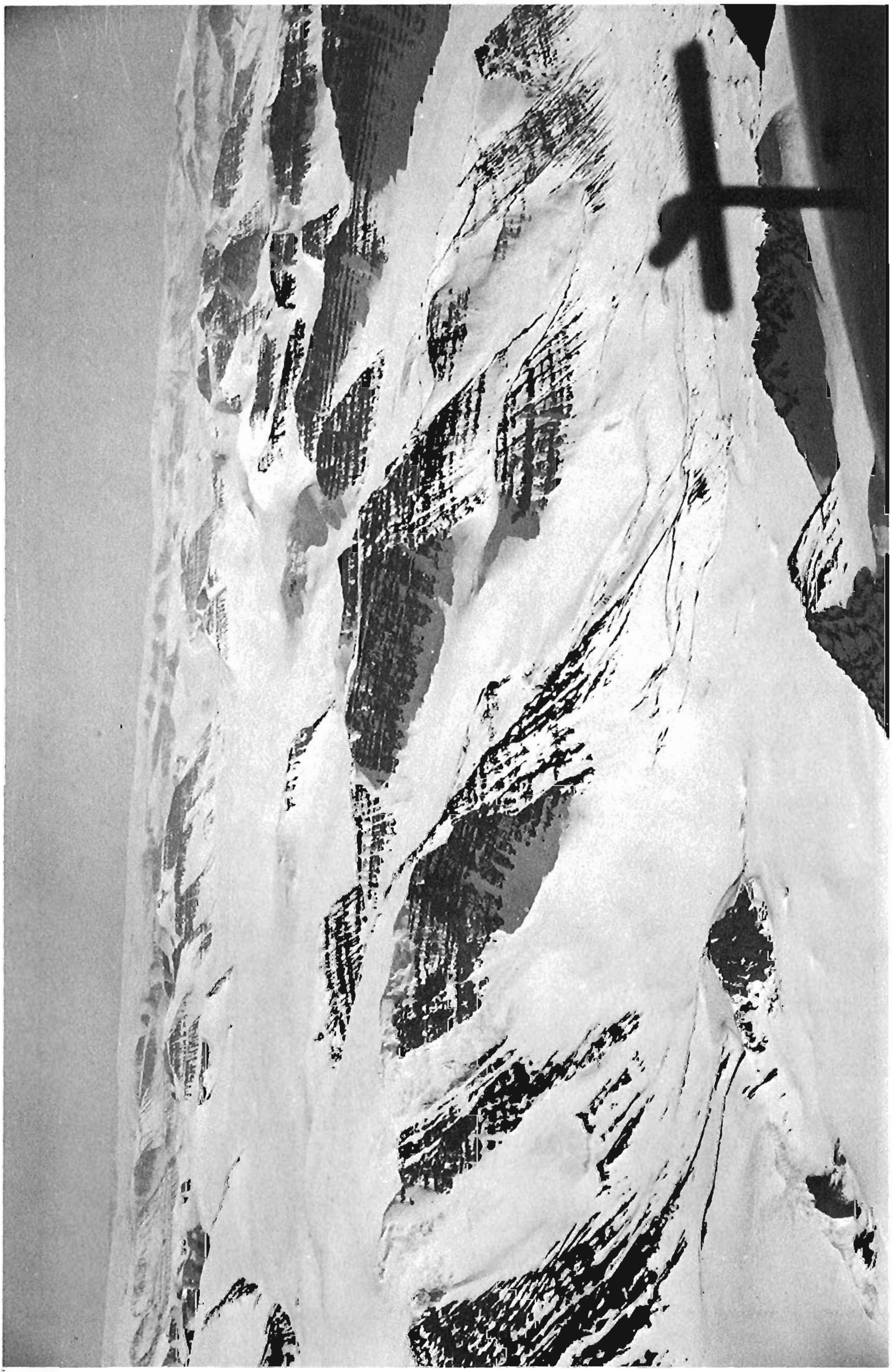


Figure 11. Basalt flows on Disco Island, West Greenland.

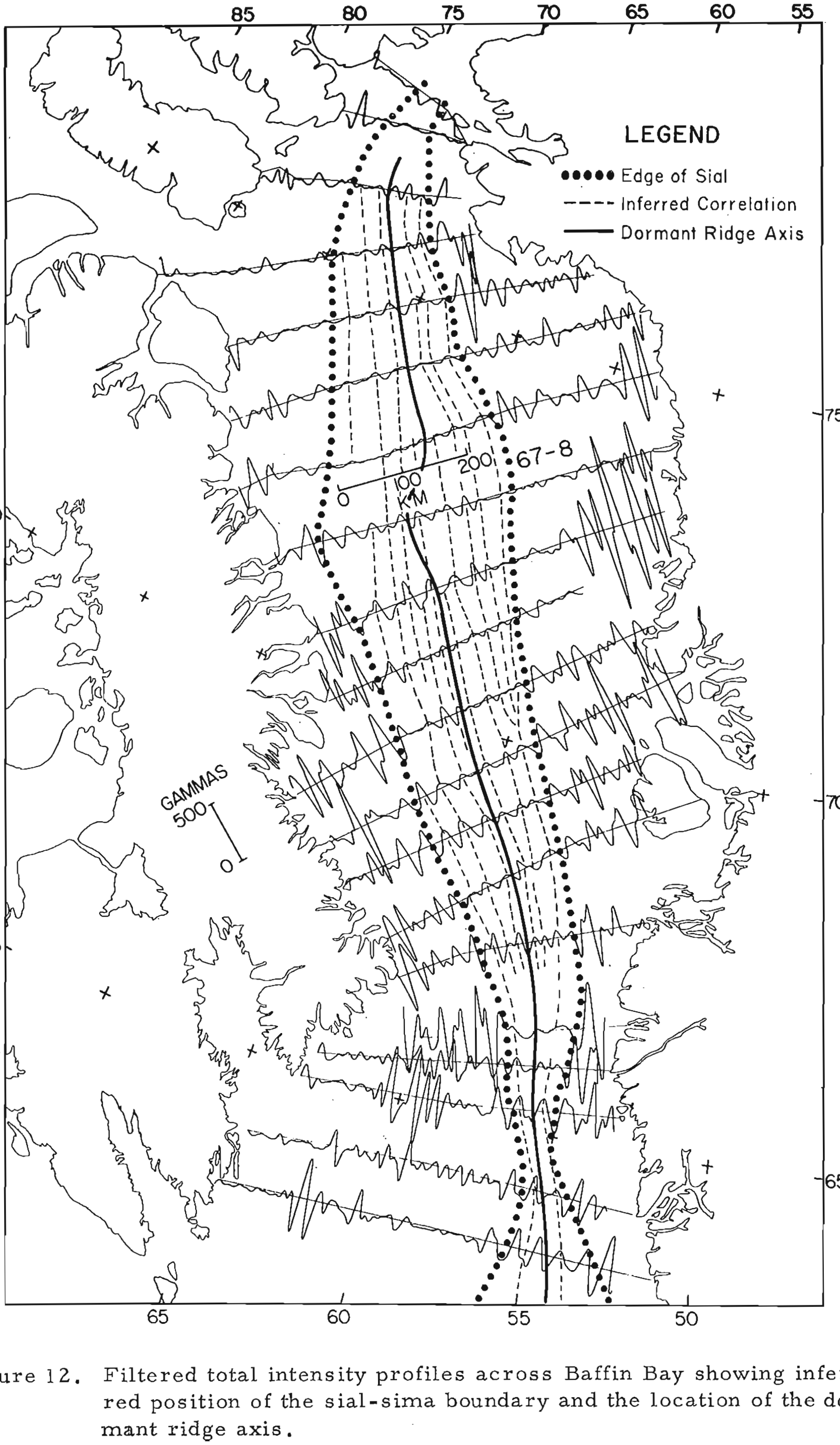


Figure 12. Filtered total intensity profiles across Baffin Bay showing inferred position of the sial-sima boundary and the location of the dormant ridge axis.

documented by Park et al. (1971) and by M.J. Keen et al. (1972) for the basaltic rocks occurring on the Greenland side. Using this pattern recognition criterion, the areal extent of the basalt intrusions has been inferred from the magnetic profiles; this qualitative interpretation is also shown on Figure 10. Thus the high frequency anomalies which occur offshore and southeast of Cape Dyer appear to be caused by the same Tertiary basalts. Similarly on the Greenland side the sharp high-frequency anomalies appear to be due to shallow basalt extrusions which also extend offshore. Note that the anomalies go well below the base line indicating that extensive areas of the basalt on Disko Island (Fig. 11) must have a negatively-polarized remanent magnetization. Moreover there is an area in the central part of Davis Strait where the anomalies are comparatively smooth indicating that the basaltic extrusions are not present at a shallow depth although presumably the basement must be oceanic rock.

The profiles in central Baffin Bay have a relatively smaller magnetic expression than those occurring in the centre of the Labrador Sea, although it is possible to discern some correlation between the anomalies. The dominant wavelength appears to be about 25 kilometres in the deep ocean part of Baffin Bay, and in order to improve the correlation between the anomalies occurring on adjacent flight lines, the data was digitally filtered using a pass band of 13.2 - 29.1 kilometres. The resultant profiles are shown in Figure 12. The reader should however note that in filtering geophysical data, the resultant end product does acquire some of the characteristics of the filter itself.

It would also appear possible to delineate the boundary of the continental and oceanic rocks by recognition of the previously discussed assymetrical anomaly and this boundary has been drawn on Figure 12 using a dotted line. The certainty of recognition is very much less than in the Labrador Sea however, although in the northern part of Baffin Bay between 70° and 76°N, the boundary agrees reasonably well with that deduced by C.E. Keen et al. (1972; see Figs. 4 and 6) from the gravity 'shelf-edge' anomaly supported by shipborne magnetometer data. The filtered profiles (Fig. 12) also appear to delineate the sial-sima boundary much better because the amplitude of the concomitant anomaly is enhanced.

In addition correlations between the profiles on Figure 12 have been made over that central part of Baffin Bay between the sial-sima boundaries. Thus there would seem to be reasonably good correlation over large distances indicating that the crust under Baffin Bay is in fact oceanic. A seismic refraction profile has been obtained in Baffin Bay centred at approximately 72°N, 66°W (Barrett et al., 1971; C.E. Keen et al., 1972) which indicates that the basement rocks are indeed oceanic at that location.

Because of the fact that some symmetry appears to be evident in the filtered profiles in the oceanic part of Baffin Bay, a possible dormant ridge axis has been drawn on Figure 12 through the central U-shaped anomaly to indicate where such an axis might be found by further field studies.

Figure 13 shows one of the profiles (Profile 67-3) across Baffin Bay which was flown from Cape Cargenholm on Baffin Island to Red Head in Melville Bay, a distance of 382 miles (615 km), and which was previously published by Hood and Bower (1970) (see Fig. 10 for location). At the top of the figure is the total field profile with the regional gradient removed. A number of short wavelength anomalies appear at either end of the profile on the respective continental shelves, and in order to bring these out in a more

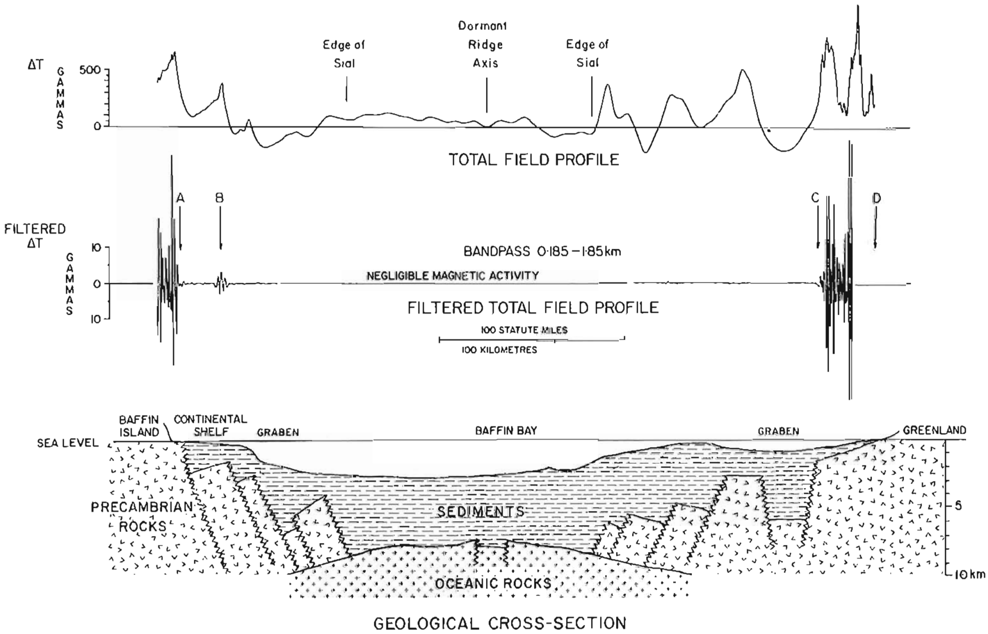


Figure 13. Aeromagnetic profile data and inferred geological cross-section across Baffin Bay from Cape Cargenholm, Baffin Island to Red Head, West Greenland.

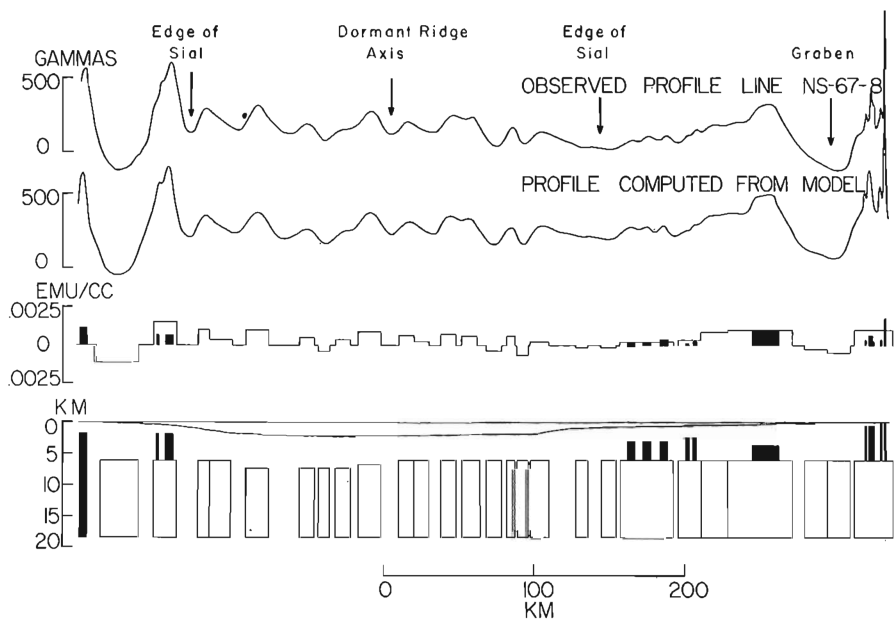


Figure 14. Interpretation of Baffin Bay Profile 67-8.

definitive way, a digital high-frequency bandpass filter which did not attenuate frequencies between 0.1 to 1.0 nautical miles (0.185 to 1.85 km) was applied to the data. The resultant filtered total-field profile is shown between the geological cross-section at the bottom of the figure and the total-field profile. The reader should note that the vertical scales of the filtered and total-intensity profiles differ by a factor of 32. It can be seen that at the coast of Baffin Island (Point A), the high frequency activity on the filtered profile stops abruptly and that there is another short length of high frequency anomalies about 16 miles from the coast, the centre of which has been labelled point B on the filtered profile. The slope of the continental shelf appears to steepen at this point. As in the case of the Labrador Sea, the fine structure is caused by the proximity of crystalline basement rocks of Precambrian age which occur on Baffin Island and as these basement rocks deepen offshore the high frequency activity consequently diminishes. Because the amplitude of the high frequency anomalies abruptly diminishes at the coast of Baffin Island, it is reasonable to conclude that a considerable sedimentary section exists on the Baffin Shelf. Moreover the presence of the high-frequency anomalies on the outer part of the shelf would also strongly suggest that a basement ridge runs along the outer part of the shelf which is probably similar to that found along the eastern sea board of North America; the inferred structure is depicted in a diagrammatic geological cross-section at the bottom of the figure.

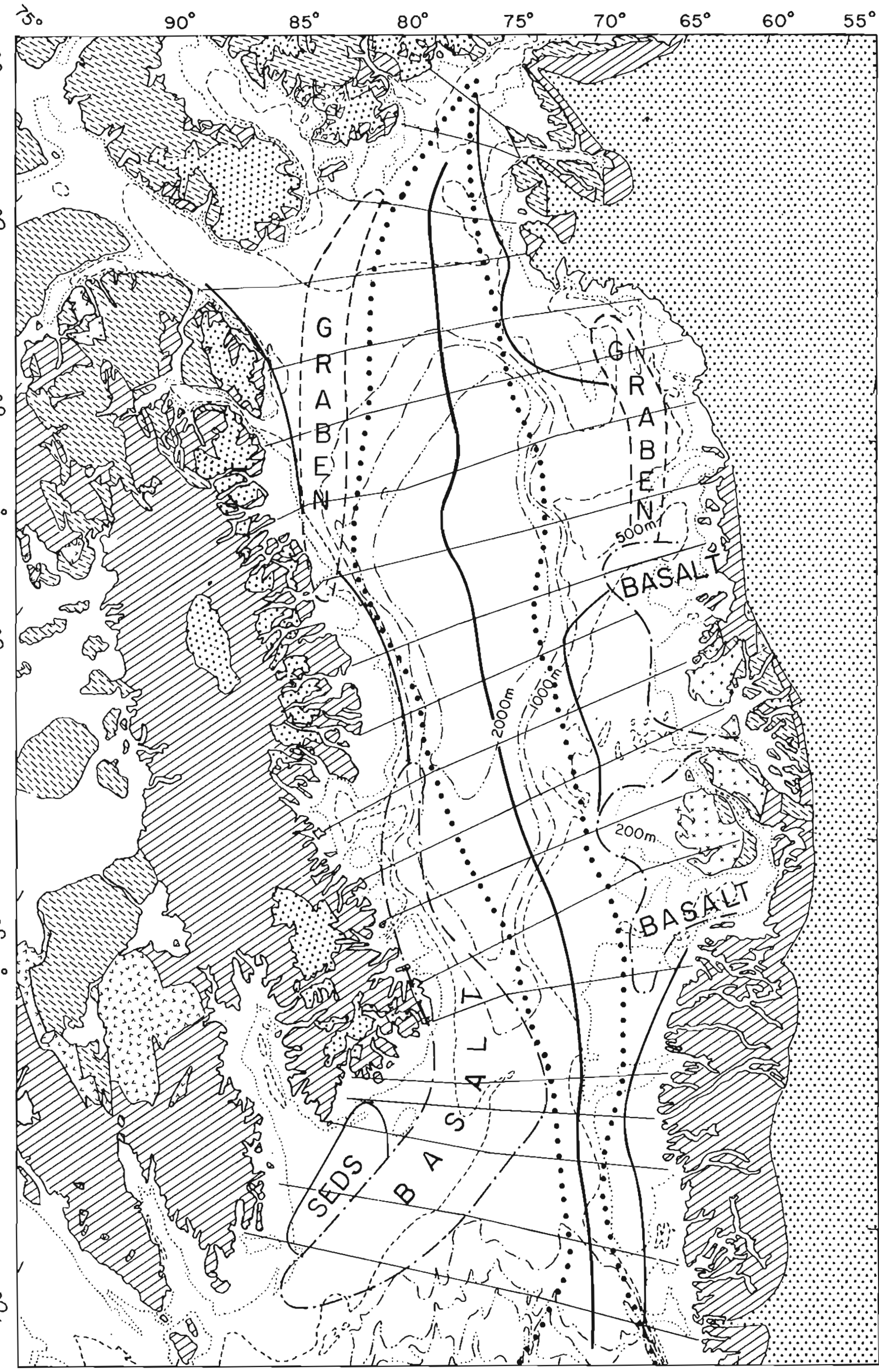
Depth determinations carried out on the profile presented and on the adjacent profiles indicate that the thicknesses of sedimentary rock exceeds 10,000 feet (3 km) in this area of the Baffin Shelf, and probably exceeds 20,000 feet (6 km) in the inferred graben.

On the Greenland side, the crystalline basement rocks appear from the filtered trace to extend for about 35 statute miles (56 km) offshore from the coast (Point D) before abruptly deepening at Point C. From this and adjacent profiles the distinct U-shaped anomaly immediately west of Point C would appear to be due to a deep sediment-filled graben whose width is approximately 35 miles (56 km). The sequence of anomalies further west appears to be produced by downfaulted blocks whose tops are buried more than 20,000 feet (6 km) below the ocean surface.

Figure 14 shows a quantitative interpretation of the profile (Profile 67-8) immediately to the south of the previous one (see Fig. 10 for location). The interpretation has also been carried out by computer curve-matching to arrive at the two-dimensional causative bodies shown in the cross-section in the bottom of the figure using the magnetizations indicated immediately above. The graben on the Baffin Shelf is much better defined on the magnetic profile and it would appear to be approximately 30 miles (48 km) wide and at least 20,000 feet (6 km) deep. The measured width of the Melville Bay graben on this profile is approximately 37 miles (60 km) wide and is filled with a similar depth of sediments in that location. The basement ridges near the edge of the continental shelf which produce a series of short wavelength anomalies are apparent from the quantitative interpretation.

The thick sedimentary cover in the central part of Baffin Bay appears to exceed 20,000 feet (6 km) over wide areas.

The qualitative interpretation of the aeromagnetic survey results obtained in Baffin Bay has been consolidated in Figure 15. The edge of the sedimentary wedge is indicated by a solid line together with the inferred limits of the basalt extrusions.


GRABEN
GRABEN
BASALT
BASALT
BASALT
SEDS
500m
1000m
2000m
200m
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60°
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95°
90°
85°
70°
80°
75°
65°
70°
50°
75°
70°
65°
65°
60°
55°
50°

LEGEND FOR FIGURE 15 (opposite)

Acidic Rocks

Basic Rocks

Metamorphic Rocks

Sediments

Ice Field

——— Track of the Survey Aircraft

• • • • • Edge of Sial

——— Edge of Sediments

—·—·— Basalt

— — — Graben

——— Dormant Ridge Axis

Bathymetry in Metres

Figure 15. Geological interpretation of aeromagnetic profile data in Baffin Bay.

On the Greenland side the boundary of the basalt agrees in general with the area delineated by Park et al. (1971) and by M.J. Keen et al. (1972) south of Disko Island. The basalt may however extend further south than is shown in Figure 15 in a narrow band parallel to the Greenland coast.

The basalt outcropping in the vicinity of Cape Dyer appears to be but a very small part of those occurring offshore which extend at least 100 miles (160 km) in a northwesterly direction along the coast and in a southwesterly direction at least as far as the entrance to Frobisher Bay. There is some indication from the magnetic profiles that there are basalt extrusives further south seaward of Cape Chidley at the northern tip of Labrador. Moreover Trask (1932) identified basalt fragments in the bottom samples collected by the USCGS MARION in 1928 in the centre of Hudson Strait south of Resolution Island and also on the northern Labrador Shelf.

The position of the grabens on either side of the northern part of Baffin Bay which were delineated from the magnetic profiles (see Fig. 10) are also given on Figure 15. The Melville Bay graben, which is at least 200 miles (320 km) long, may extend to the south beneath the basalt extrusions; from a consideration of its arcuate shape and the bathymetry, the graben may extend to the west of Kap York. It is perhaps tectonically significant that the graben on the Baffin Shelf appears to terminate near Scott Inlet opposite a linear swarm of earthquake epicentres located on Baffin Island (see Fig. 9) which are aligned in a direction that intersects the southern end of the graben. The Baffin Shelf graben is at least 350 miles (560 km) long and averages 30 miles (48 km) wide. It contains a minimum of 20,000 feet (6 km) of sediments.

SUMMARY AND CONCLUSIONS

The aeromagnetic reconnaissance of the Labrador Sea and Baffin Bay has indicated that long linear sedimentary basins are found along the outer edges of the continental shelves, slopes and rises which border these oceans. On the northern Labrador Shelf more than 50,000 feet (15 km) of sediments are to be found in the site of an extinct subduction zone. An extensive sedimentary cover is found in the central part of Baffin Bay which exceeds 20,000 feet (6 km) over large areas. In addition two major sediment-filled grabens have also been located in northern Baffin Bay each of which contains more than 15,000 (4.6 km) of sediments. However, in the Davis Strait area basalt extrusions cover large areas of young sediments. Most of the considerable thicknesses of sediments in the Labrador Sea and Baffin Bay would appear from the timing of the sea-floor spreading sequence to be Mesozoic in age, making these areas attractive for the petroleum industry to prospect.

ACKNOWLEDGMENTS

The authors wish to thank the various members of the crew of the North Star aircraft who took part in the various surveys over the years, and also to E. A. Godby, R. C. Baker, M. Strome and N. Davis formerly of the National Aeronautical Establishment who were participants in the GSC-NAE project. We also wish to thank N. J. McMillan of the Aquitaine Company of Canada Ltd. for many helpful discussions, and G. Cameron for his assistance in carrying out the depth determinations necessary to produce Figures 7 and 8.

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36. GEOLOGICAL HISTORY OF BAFFIN BAY - CONTINENTAL DRIFT BEFORE SEA-FLOOR SPREADING AND THE EXPLORATION FOR HYDROCARBONS

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Abstract

The Cretaceous to Eocene sedimentary/volcanic sequence of the Disko Island area and southern Baffin Island indicates that the opening up of Baffin Bay by Greenland and North America drifting apart involved this sequence: 1) land-locked rift valley filled with terrestrial deposits, Barremian to Turonian; 2) widening of valley and intermittent marine connections, Upper Turonian to Danian; restricted bottom circulation with deposition of bituminous shales, Santonian through Danian; 3) deepening of rift, opening of magma chambers and effusion of basaltic lavas, Paleocene and Eocene (limited to Davis Strait area); 4) widening and deepening of rift valley to present Baffin Bay and Davis Strait; very thick young sedimentary sequence in deepest part of graben. There is no evidence of sea-floor spreading in Baffin Bay, nor is there a mid-Baffin Bay ridge. In this respect it is similar to several other deep marine basins believed due to continental drift. Furthermore, the tectonic-depositional-volcanic history of the Rhine Valley and East African rift systems closely parallels stages 1 to 3 of Baffin Bay. Rifting and continental drift apparently can take place independently of sea-floor spreading, which would be the last phase of the sequence. If sea-floor spreading had created Baffin Bay, its sediments would not predate its volcanics by some 50 million years, nor would these be so limited in area. Sea-floor spreading in the Atlantic started only about 60 million years ago, long after the initial (Jurassic) phase of sedimentation. Practical implications to the petroleum industry of considering rifting the primary cause of continental drift are: 1) the same sedimentary sequence is common to many coasts broken by drift, e.g. the oil provinces of western Africa; 2) this sequence includes both source rocks and thick reservoir rocks; 3) the antithetically-rotated fault blocks and cross-faults caused by rifting created suitable structural traps; 4) since volcanism started only at the end of the sequence, there existed little danger of volatilization of hydrocarbons; 5) the central depression, the last stage of rift widening, was filled with thick clastics without source rocks.

INTRODUCTION

The striking parallelism of the coasts on either side of Baffin Bay and of the Labrador Sea (Fig. 1) has caused many geologists to believe that Greenland was at one time connected to the North American continent, and became separated from it through the mechanism of continental drift (Wegener, 1924; Taylor, 1928; DuToit, 1937; Carey, 1958; King, 1958; Wilson, 1965; Bullard et al., 1965; Demenitskaya and Dibner, 1966). Most

*deceased

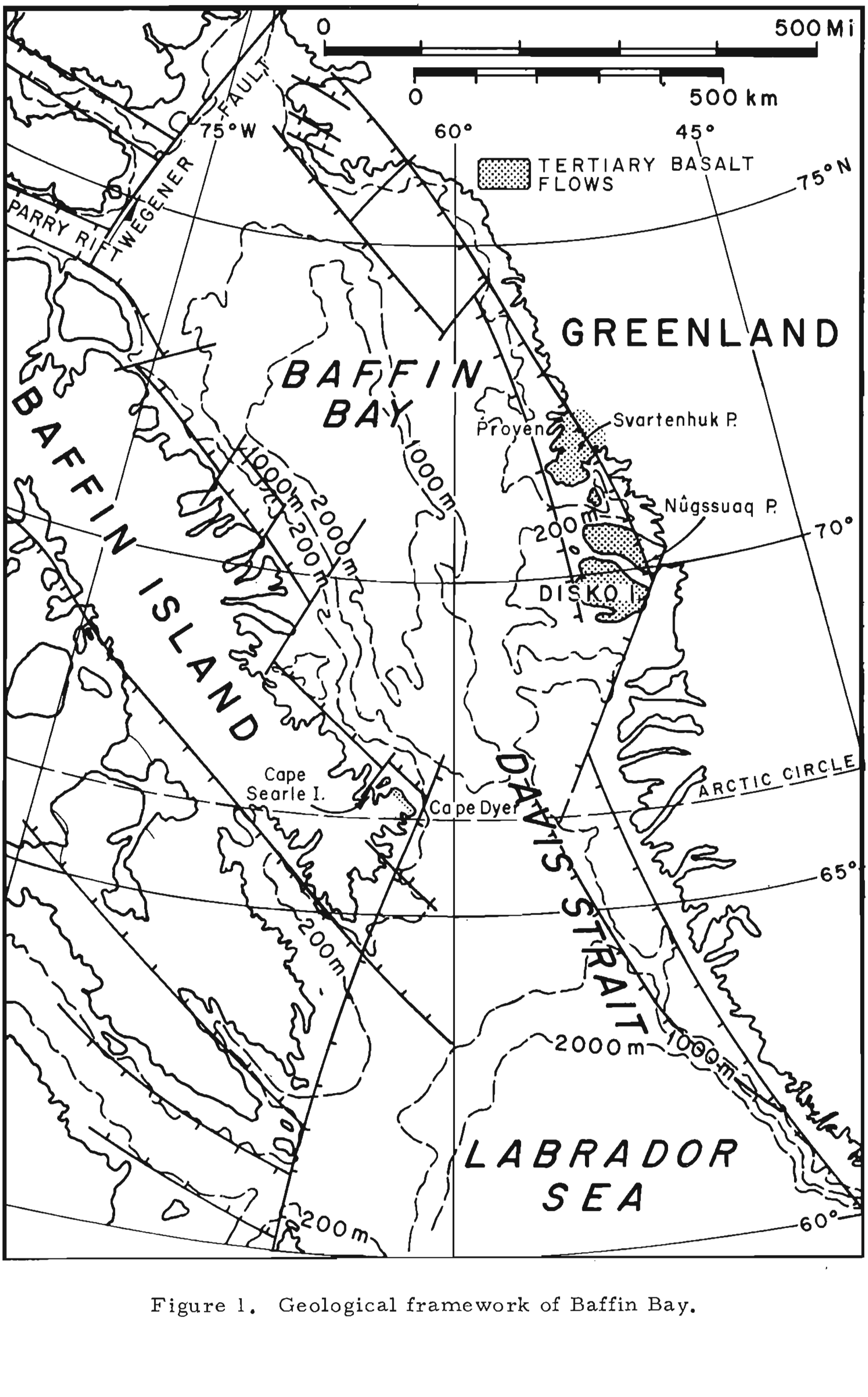


Figure 1. Geological framework of Baffin Bay.

of these authors saw the Nares Strait (Smith Strait, Kennedy and Robeson Channels) between Greenland and Ellesmere Island as a transform fault (Wilson, 1965), along which Greenland had moved northeastward, away from Baffin Island. Wegener was the first one to show it as such; Carey coined the hybrid term "Robeson Megashear" for it, and Wilson named it the Wegener Fault. A third fault, or pair of faults, in the area has defined Lancaster Sound. It, too, was originally shown on Wegener's map; it was recently named the Parry Rift by Pallister and Bourne (1971) and Kerr (1971). Fig. 1 shows the broad outline of the resulting fault system, which was elaborated upon further by Kerr (1971).

In recent years it has become fashionable to discuss continental drift – an old idea, for which Wegener deserves more credit than he nowadays receives – in terms of "sea-floor spreading" and so-called "plate tectonics". The oceanic crust recently discovered beneath the deepest portion of Baffin Bay (Barrett et al., 1971) shows that the sialic, or continental, crust has separated to form the Baffin Bay rift, but no conclusive proof of sea-floor spreading has been found to date.

Along with Wegener (1924) and, more recently, such authors as Bartlett (1971) and Trümpy (1971), the author of this paper wishes to stress the need for analyzing the mode and timing of continental drift movements in the light of well-known geological, stratigraphic and paleontological facts instead of using only geophysical data. Wegener's approach, in writings dating back to 1912, was geological, stratigraphic, structural, palaeontological, palaeoclimatological, zoological, botanical and a lot more, but he lacked strong geophysical arguments. It is to the derogation of the geologists present at that memorable symposium on continental drift held in Tulsa in 1928 that they did not stand up for the significance of Wegener's arguments and backed down before geophysical counter-arguments that, as was subsequently shown, were devoid of substance. Dating from this meeting, there was for thirty years hardly a mention of the concept of continental drift in the geological literature of this continent.

In the case of Baffin Bay, Greenland was one part of the world that Wegener knew better than most, through the field work he carried out there, and where he subsequently perished. Greenland has been the subject of intensive geological studies by numerous authors for the past hundred years or so, which culminated in the recent publication by the Geological Survey of Greenland (1970) of a new tectonic-geological map of Greenland. Work on the Cretaceous-Tertiary sedimentary volcanic rocks of the Svartenhuk to Disko Island area of west Greenland has been summarized in recent years by Koch (1964), Henderson (1969), Rosenkrantz and Pulvertaft (1969) and others. The excellent work of these and many other authors is freely and gratefully drawn upon in the following pages.

Geological History of Baffin Bay

Thick (approximately 2000 metres) clastic sediments, Barremian to Paleocene (Koch, 1964) in age, capped by Paleocene to Eocene plateau basalts, occur on and north of Disko Island on the west coast of Greenland. These are matched by small outliers of sand and shale (about 150 metres thick) of similar age along the shores of Baffin Bay on southern Baffin Island and adjacent small islands (Kidd, 1953; Wilson and Clarke, 1965; Clarke and Upton, 1971). In both areas the sediments lie on an eroded and deeply-weathered old land surface of Precambrian rocks. The sequence in west Greenland (Henderson, 1969) starts with terrestrial sands and coal seams of Barremian to Coniacian age, with some marine intercalations from the Upper Turonian onwards. These are followed by a mixed terrestrial-marine clastic sequence from the Lower Santonian through early Tertiary age levels. Bituminous shales with possible to definite hydrocarbon-source rock characteristics occur from the Lower Santonian to the Upper Danian (Henderson, 1969).

The basaltic lava flows that have been mapped in this part of west Greenland as well as in the vicinity of Cape Dyer, Baffin Island, overlie the bulk of these sediments (Fig. 1). Whereas, on Baffin Island, the sediments do not cover much more than the immediate edges of the coasts, the basalt flows extend considerably farther inland (up to ten kilometres); they overstep onto the old Precambrian surface on both sides of Baffin Bay. Since many, and especially the oldest lava flows appear to be of subaqueous origin, this overlap indicates that the entire area of their deposition subsided prior to their extrusion.

Some clastic beds are intercalated between the basalt flows. Detailed dating of nonmarine sediments found as intercalations in the basalts of west Greenland has established a Paleocene to Lower Eocene age for these lavas (Munck and Noe-Nygaard, 1957). The only older evidence of volcanic activity are tuffs (ash beds) of Danian age, which have an andesitic to andesito-basaltic composition (Munck and Noe-Nygaard, 1957). The origin of these lavas is indicated to lie west of Disko and east of Cape Dyer, so that at the time of their formation a sizable gap must already have existed across Davis Strait (Clarke and Upton, 1971; Brooks, 1971).

Using the Geological Society of London (1964) Phanerozoic time scale, the sequence described above is broken down in Table 1 in terms of age.

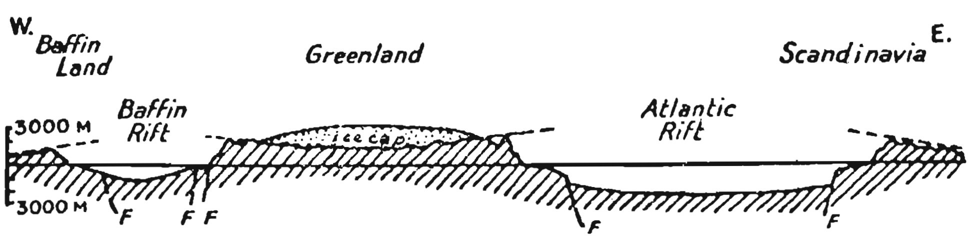
Structural movements during the Cretaceous and Tertiary were predominantly in the form of block faulting, accompanied by a certain amount of tilting. Unconformities at four levels have been described by Henderson (1969). The principal movements took place at the beginning of the Danian and of the Paleocene; both are represented by heavy conglomerates. Beneath the more or less flat-lying Tertiary, the Cretaceous of the Disko Island area dips at various angles in different directions. A major fault, with a throw of approximately 900 metres, traverses the Nûgssuaq Peninsula from northwest to southeast. The oldest member of the stratigraphic sequence, the Barremian-Aptian Kome Beds, occurs exclusively on the upthrown (east) side of this fault (Munck and Noe-Nygaard, 1957).

Subsequent to the termination of the depositional (sedimentary/volcanic) sequence, the areas on both sides of Baffin Bay were uplifted to a considerable height. Both west Greenland and eastern Baffin Island now tilt slightly away from Baffin Bay (Fig. 2, after DuToit, 1937). Since many, and especially the basal lava flows are of subaqueous origin, and all those now on land occur up to considerable elevations, the amount of uplift is 150 to 500 metres on Baffin Island (Clarke and Upton, 1971), and at least 500 metres on the Greenland side (Brooks, 1971). The tops of the basalt flows now reach more than 1200 metres above sea-level on Baffin Island, and close to 2000 metres in the Disko Island area.

Presumably during and after this phase, the Baffin Bay rift opened up to its present width, and the graben deepened to its present depth. In the northern and most depressed area, the water depth is almost 2400 metres. Murray _et al._ (1970) have calculated magnetic basement depths of up to

TABLE 1

Million years B. P.				
45		(		
	TERTIARY	(Eocene to	plateau basalts (K-Ar age: 54-58 m. y.)	
		(Paleocene		
54		(		
		(Paleocene	marine and terrestrial sediments, some basalt	
		(		
60		(		
		(Danian	marine and terrestrial sediments with some andesitic-basaltic tuffs	
		(		
65				
	CRETACEOUS	(Maestrichtian	marine and	intercalated
70		(		
		(Campanian	terrestrial	bituminous shales
76		(		
		(Santonian	sediments	marine
82		(		
		(Coniacian		intercalations
88		(		
		(Turonian	terrestrial	
94		(		
		(Cenomanian		
100		(	deposits	
		(Albian		
106		(		
		(Aptian	terrestrial deposits (Kome Beds)	
112		(		
		(Barremian	(only east of Nûgssuaq fault)	
118				
	PRECAMBRIAN			



Cross section 3500 km. in length from Canada to Scandinavia, vertical scale greatly exaggerated, showing tilting of plateau-surface away from oceanic rifts. *F.F.*, Hypothetical Faults.

Figure 2. Schematic cross-section from Baffin Island through Greenland to Scandinavia. Vertical scale greatly exaggerated; length of section approx. 3500 km. After DuToit (1937).

18,000 metres along the west side of Baffin Bay (Fig. 3), and sediment thicknesses in excess of 9000 metres. Hood and Bower (1971), however, have estimated this latter value only at more than 5000 metres. A refraction seismic profile (Barrett et al., 1971) established the presence of 4300 metres of sediments and a basement depth of 6600 metres.

The opening up of the Baffin Bay rift as a result of the drifting apart of Greenland and North America has thus led to the following sequence (Fig. 4):

First. Land-locked rift valley filled with fluviatile, lacustrine and other terrestrial deposits (Kome Beds).

Second. Widening of the rift and further terrestrial deposition in the newly-formed central graben.

Third. Deepening of the graben and intermittent connections with the open sea. Restricted bottom circulation leading to deposition of bituminous shales.

Fourth. Deepening of the rift, opening of first magma chambers and extrusion of andesitic and andesito-basaltic tuffs.

Fifth. Subsidence of a broader area than before, and effusion of basaltic lava flows.

Sixth. Widening and deepening of the rift valley to the present Baffin Bay and Davis Strait. Very thick, young clastic sedimentary sequence in the deepest part of the graben.

In terms of duration, using the same time scale as before, these various stages extended over a considerable amount of time:

Millions of years		Stage
?	Uplift and block faulting	6
9	Plateau basalts	5
11	Marine and terrestrial sediments Andesitic — basaltic ash layers	4
(17	Restricted marine deposits)	
(	)	3
(9	Some marine intercalations)	
53 (		
(15	Nonmarine deposits in new central graben	2
(		
(12	First nonmarine deposits on (later) flank blocks	1

Structural History of Baffin Bay

The Tertiary volcanics represent only the fourth and fifth stages in this sequence. In addition, their occurrence is limited to the Disko Island area of western Greenland and the Cape Dyer area of Baffin Island, on either side of Davis Strait. No lavas have been reported north of a line from Cape Searle Island to Proven, Greenland (map, Fig. 1). There has also been no

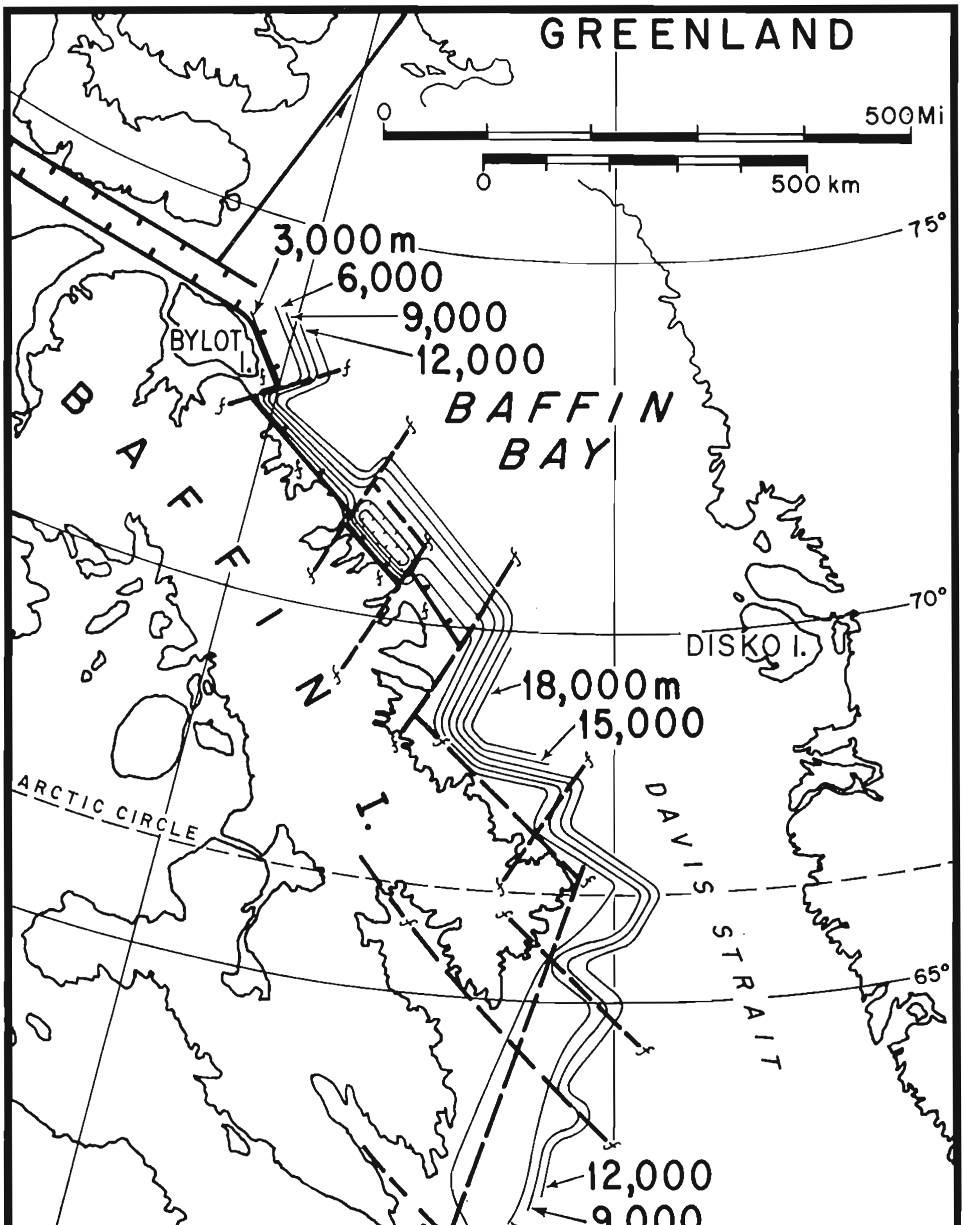


Figure 3. Depth to magnetic basement, west side of Baffin Bay (in metres). Interpreted from data by Murray et al. (1970).

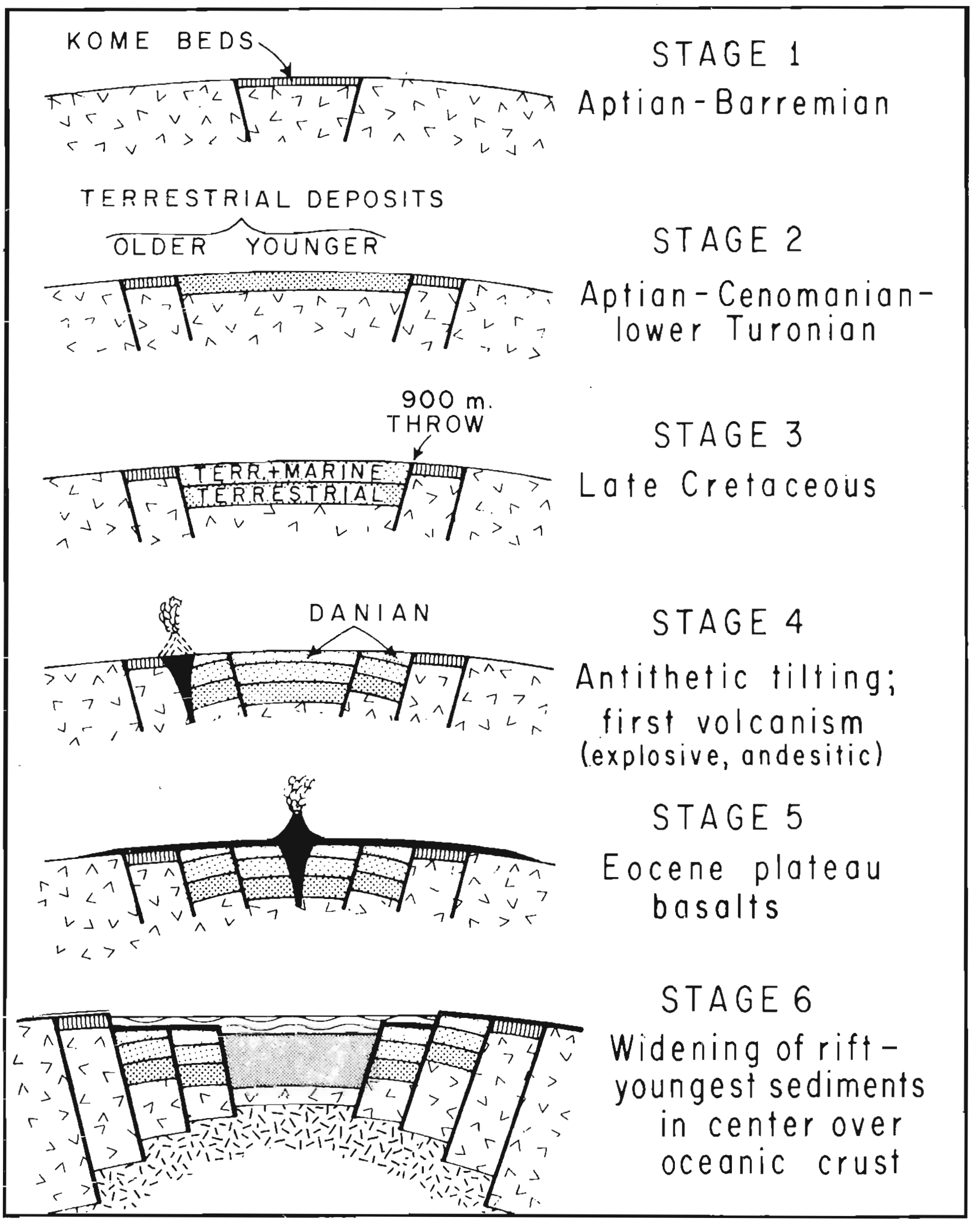


Figure 4. Cretaceous to early Tertiary geological history of the Baffin Bay rift (schematic).

conclusive evidence to date of sea-floor spreading in Baffin Bay, even though some authors believe that the shallow (630 metres and more) Davis Strait passage may be underlain entirely by plateau basalts that might have preserved a record of such spreading (Clarke and Upton, 1971), but others do not share this opinion (Henderson, 1971). No mid-Baffin Bay ridge has been observed that might be an extension of the submerged ridge (Drake et al., 1963; Hood and Godby, 1964) or "sunken continent" (Kerr, 1967a) beneath the Labrador Sea.

During the last stage, the rift opened wide enough to expose some oceanic crust at the base - the continental crust having been withdrawn to either side of some parts of Baffin Bay, or having been subjected to chemical and physical alteration. The presence of such oceanic crust has been established below the cover of the youngest sediments in at least one area (Barrett et al., 1971). On the other hand, the propagation of Lg seismic waves through the northern part of Baffin Bay was earlier interpreted by Oliver et al. (1955) as an indication of the presence of continental crust, except possibly in the small deep areas of the Bay. If both observations are correct, one for the deepest and one for the shallower parts of Baffin Bay, then the latter would be very similar to the Red Sea. Here, too, oceanic seismic velocities have been noted only in the axial zone where the sea has both its greatest width and its greatest depth (Girdler, 1969).

If the oceanic crust that underlies the deepest portions of Baffin Bay is flanked by blocks of continental crust that have measurably subsided, then the principles of isostasy require that the latter blocks become thinner as they lie deeper. A cross-section may be expected such as has been established on the Labrador Shelf (Mayhew et al., 1970) and south of Newfoundland (Bentley and Worzel, 1956). A similar thinning of the continental crust, accompanied by downward block-faulting, has been shown for the southern Red Sea by Lowell (1971). To what extent this thinning involved processes of alteration ("oceanization"), tectonic erosion by mantle currents, or subaerial erosion and isostatic adjustment will not be discussed in this paper. The writer prefers to believe, however, that the continental and oceanic crust are transitional into one another. The oceanic crust beneath the deepest part of Baffin Bay has a thickness of 4200 metres (Barrett et al., 1971). It is evident from a comparison of the Mayhew and Bentley and Worzel profiles that the older sediments are either restricted to, or are at least thickest, in the fault blocks closest to the shore (primarily the area underlying the continental shelf), whereas the youngest sediments are thickest in the deep-water area far offshore (Fig. 4, Stage 6, centre).

A number of shipborne magnetometer, aeromagnetic and seismic refraction profiles have been published for the Labrador Sea (Manchester, 1964; Godby et al., 1966; Grant, 1966; Hood et al., 1967; Johnson et al., 1969), and for Baffin Bay (Hood and Bower, 1970; Manchester and Clarke, 1971). Fig. 5 represents an attempt at interpreting a profile across northern Baffin Bay from these data and from the general considerations discussed above.

The process of crustal thinning makes it difficult to decide which bathymetric contour to use for fitting the two sides of Baffin Bay together in a reconstruction of pre-drift conditions. Lowell (1971) correctly stated that the present opposing coastlines of the Red Sea were never in contact, and the same may be assumed for Baffin Bay.

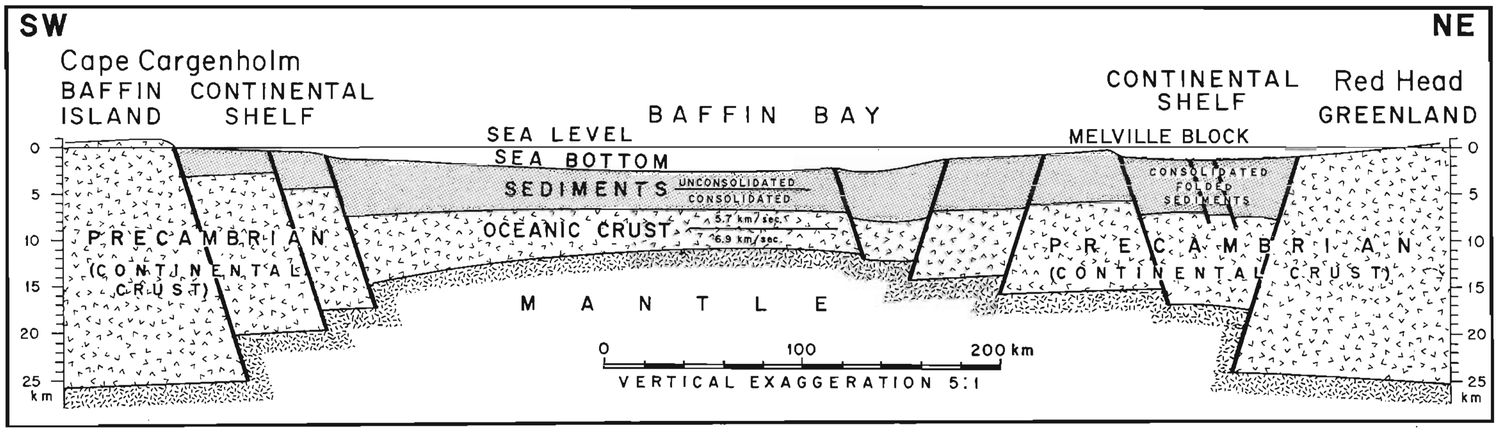


Figure 5. Schematic geological cross-section across northern Baffin Bay. Interpreted after Hood and Bower (1970), Barrett *et al.* (1971) and Manchester and Clarke (1971).

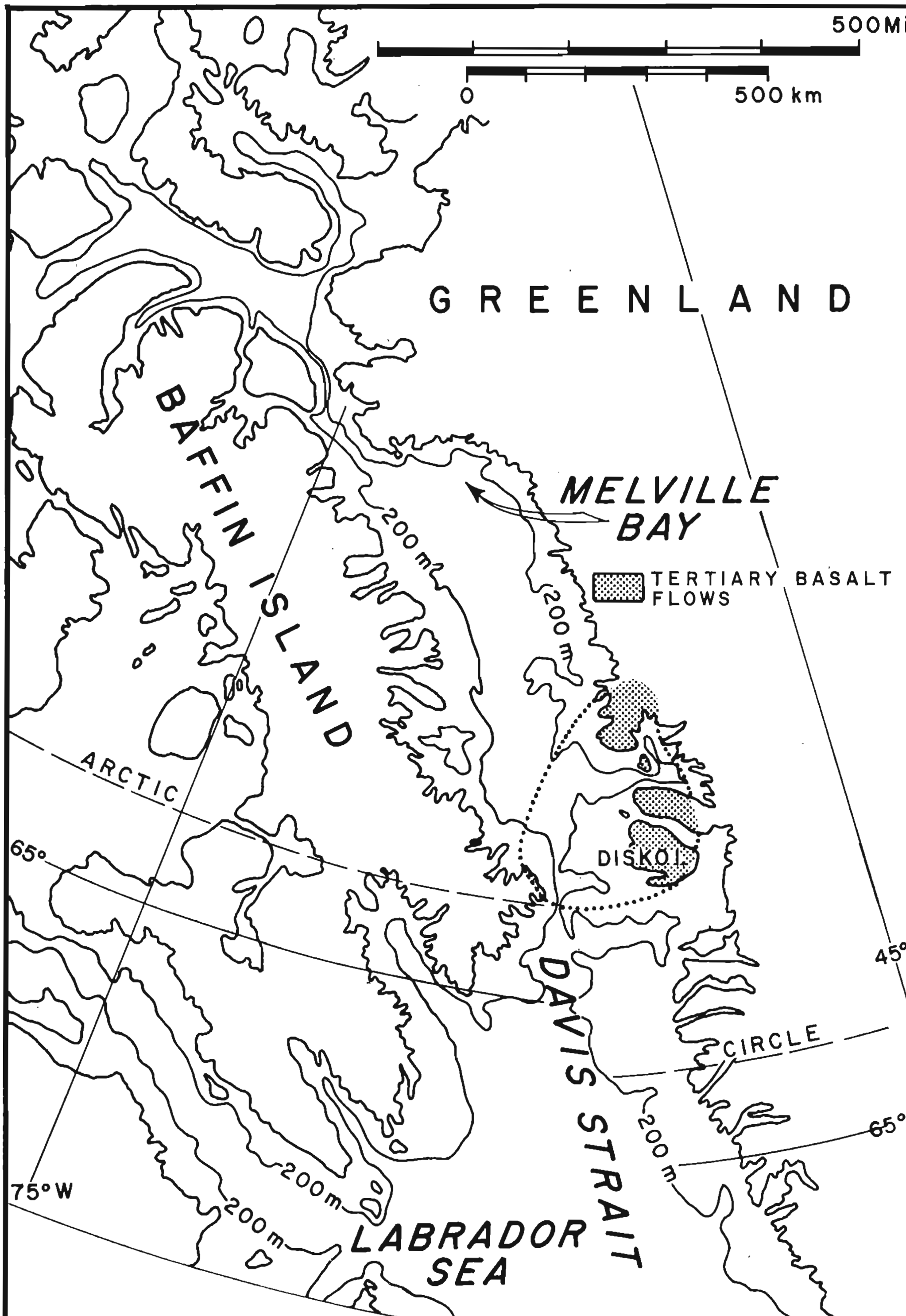


Figure 6. Reconstruction of approximate original position of Greenland and Baffin Island, I. Closing on the 200-metre depth line.

One reconstruction that is frequently referred to is the one offered by Bullard et al., (1965). Since they show a wide gap between Ellesmere Island and the original position of Greenland, and since that gap was originally apparently minimal (Kerr, 1967b), their reconstruction is not complete without assuming that the Parry Rift represents a fault of left-lateral displacement, most of the Canadian Arctic Islands having been much farther east originally than at present. Bullard et al. (1965, p. 48) recognized this problem themselves, and the author would certainly prefer a reconstruction of the Atlantic rift that has less far-reaching consequences for the Canadian Arctic Islands.

Baffin Bay, while to all appearance a northerly extension of the Labrador Sea, is actually narrower and is separated from the latter by the Davis Strait and the adjacent volcanic extrusions. Another reconstruction, different from that of Bullard et al. (1965) which assumes the Wegener Fault to be a transform fault, shows that Baffin Bay may virtually be closed by moving the 200-metre depth lines on both sides together, but that the same amount of translation still leaves a major gap in the Labrador Sea (Fig. 6). If it is assumed that the shallow Davis Strait area is underlain by volcanics that are related to the rifting, and the 200-metre depth line is disregarded in that area, the Labrador Sea gap still persists (Fig. 7). This gap may well, wholly or in part, correspond to the mid-Labrador Sea ridge (Kerr, 1967a). An additional gap corresponds to Melville Bay, the widest part of Baffin Bay (see Fig. 6), which appears to represent a sunken continental block with an interesting history of its own (Manchester and Clarke, 1971; Henderson, 1971). In the opinion of Kerr (1967b), the displacement of Greenland along the Wegener Fault is considerably less than is generally assumed. If this view is correct, the sunken "Melville Block" (see Fig. 7) might have been that much more extensive.

An attempt has been made on Figure 1 to outline some of the possible rift faults (parallel to Baffin Bay), transform faults (Wegener Fault; separation between Baffin Bay and the Labrador Sea) and cross-faults that have affected the area. Detailed aeromagnetic maps reveal additional longitudinal as well as cross-faults (Geological Survey of Canada, 1970). Wilson and Clarke (1965) and Clarke and Upton (1971) noted several large normal faults striking parallel to the coast at Cape Searle (see Fig. 1) and on Padloping Island, the coastal blocks being down-faulted. On Greenland's west coast, normal faults parallel to as well as at an angle to the coast have been mapped, trending NNW and NE, respectively (Geological Survey of Greenland, 1970). That some of these faults are quite recent was shown for the Labrador shelf by Grant (1970). Several earthquake epicentres are known from the northern portion of Baffin Bay (Hood et al., 1967).

Some folding is indicated in the area, although most of this is probably related to the faulting. Kidd (1953) noted a south dip near Cape Searle and a north dip on northern Padloping Island, which would place a syncline between them, striking northeast.

Baffin Bay and the New Global Tectonics

Although new data to the contrary may become available in future, there are several other deep marine basins generally believed to be caused by continental drift that have not, at least to date, yielded conclusive evidence of sea-floor spreading and/or mid-basin ridges, e.g. the Red Sea, the

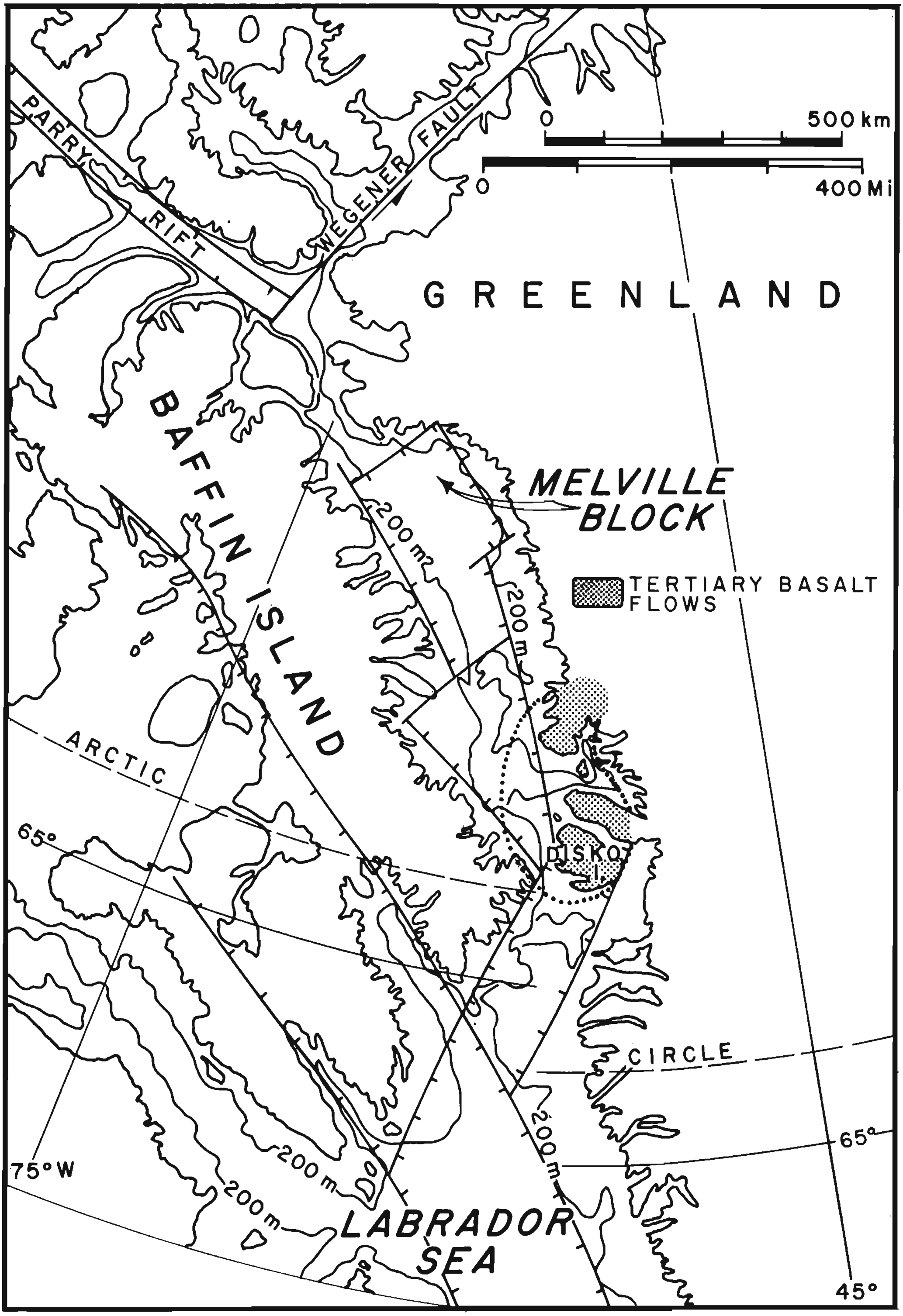


Figure 7. Reconstruction of approximate original position of Greenland and Baffin Island, II. Disregarding 200-metre depth line in the area of Disko Island.

Mozambique Channel, the Gulf of Mannar between Ceylon and India, the rift between the Faeroe Plateau and the British Isles, and doubtless others. Furthermore, the tectonic-depositional-volcanic history of the Rhine Valley and East African rift systems closely parallels the first three stages of the geological history of Baffin Bay given above. Cloos (1939) has compared the Rhine Valley and Red Sea graben systems, and it is easy to make a similar comparison for Baffin Bay (Fig. 8). The similarity between Baffin Bay and the Red Sea is particularly noticeable. Arabia has moved northward along the Gulf of Aqaba-Dead Sea transform fault in a manner very similar to the movement of Greenland along the Wegener Fault. Volcanism in the Red Sea area, too, is present primarily at its southern (Afar triangle) and northern extremities (Sea of Galilee). Comparable centres of volcanic activity in the Baffin Bay area are Davis Strait in the south and the Cape Washington area of northern Greenland at the northern end (Dawes and Soper, 1971). The Red Sea also is underlain only by a limited area of oceanic crust, has yielded no conclusive (magnetic) evidence of sea-floor spreading, and also contains no mid-Red Sea ridge (Girdler, 1969; Lowell and Genik, 1971; compare also Falcon et al., 1970). Its central portion, where the oceanic crust occurs, is topographically, like Baffin Bay, its deepest part. Of particular interest to the petroleum industry is the fact that one of the branches of this three-pronged fault system, the Gulf of Suez, is the locus of several singularly prolific oil fields. Considering that the Red Sea proper is also the scene of intensive exploration for oil and gas, it seems only logical that Baffin Bay is also getting its share of attention. For some years now, there has been considerable exploration interest on both the Greenland and the Baffin Island sides of the Bay.

Thus there appear to be strong arguments in favour of the view that rifting and continental drift can take place independently of sea-floor spreading; and that sea-floor spreading, where it has taken place, has been the terminal phase of the sequence given above. If sea-floor spreading had caused the formation of Baffin Bay, its sediments would not predate its volcanics by some fifty million years, nor would the latter be limited to a small area around

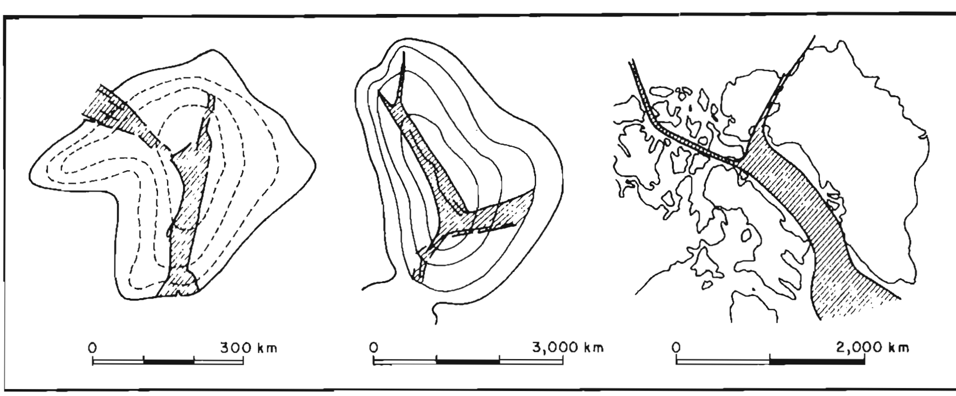


Figure 8. Comparison of Rhine Valley, Red Sea and Baffin Bay rift system (two drawings at left after Cloos, 1939).

Davis Strait. In addition, it is significant that the first volcanic sediments are andesitic tuffs, indicating, firstly, the presence of a sialic crust, and secondly, explosive volcanic vents that opened only rarely – in contrast to the subsequent more or less continuous flow of basaltic lavas.

Without in any way denying the existence of sea-floor spreading in the Atlantic Ocean and certain other areas, the writer considers it erroneous to credit its mechanism with being the driving force in continental drift. It is obvious from the geological history of Baffin Bay that a rift started to open, then widened and deepened, long before any basaltic magma was extruded. It also appears that the locus of the extrusion that did take place was very limited in area, and that it could therefore not have been responsible for the formation of the entire Baffin Bay rift. There is no conclusive evidence for the existence of a "spreading centre" in the middle of Davis Strait (Clarke and Upton, 1971; Henderson, 1971).

On a much larger scale, the Atlantic Ocean must have started to open at least 170 million years ago, since Bajocian and other Middle to late Jurassic sediments were recovered in a core taken on Orphan Knoll, north of Flemish Cap, and in the Grand Falls well drilled off Newfoundland (Laughton et al., 1970; Bartlett and Smith, 1971), and other Jurassic occurrences are known from various parts of this ocean. Yet the evidence for sea-floor spreading in the Atlantic does not reach farther back than about 60 to 70 million years before the present.

The mechanism of sea-floor spreading, as proposed by its adherents, also envisages subduction zones at the edges of the continental plates that were moved against other plates by the spreading force. There is no such subduction zone or other evidence of collision with a neighbouring plate on either side of Greenland. Consequently, the mechanism that caused the opening of the Baffin Bay rift and, more significantly, that allowed oceanic crust to be exposed on the floor of the rift, is not likely to be sea-floor spreading.

It would appear therefore that much of the theorizing that has been done during the past decade with respect to the causes of continental drift is in need of revision. A thorough examination of the geological field evidence should be a good investment in a better future understanding of the problem. The writer does not intend at this stage to formulate yet another opinion with respect to the mechanism of continental drift. He does wish, however, to register his objections against much recent writing and speaking that tries to over-simplify the problem.

CONCLUSIONS

It is concluded from this survey of the geological history of Baffin Bay that modern concepts of sea-floor spreading and "plate tectonics" fail adequately to explain certain well-documented geological data on Baffin Bay:

1. The sialic crust was affected by extension rifting over a period of some 50 million years before simatic magma chambers were opened up. This can only be explained by independent movement of the sialic crust along the Mohorovičić discontinuity, and not by "plate tectonics", if the base of the plates is assumed to be the asthenosphere.

2. The graben formation that accompanied rifting and the sinking of such continental crust blocks as the Melville Block, possibly the mid-Labrador Sea ridge (if indeed continental), and similar blocks in the Atlantic Ocean (Orphan Knoll, Rockall Plateau, Vøring Plateau) can only be explained isostatically by a thinning of the sialic crust from the bottom, e.g. by sub-crustal tectonic erosion (Gilluly, 1964), which implies convection currents in the sima. The early, explosive andesitic-volcanic phase in Baffin Bay indicates the creation of a sialic magma chamber before simatic magma started to rise.

3. There is no evidence of subduction zones that could be related to sea-floor spreading in Baffin Bay, neither on the Canadian side, nor on the Greenland side. On the contrary, there are other rifts on either side which have formed the Arctic and northern Atlantic Oceans.

Implications for the Origin and Accumulation of Hydrocarbons

The practical implications to the petroleum industry of viewing continental drift and the opening up of such basins as Baffin Bay as the result of rifting, and of considering rifting the primary cause and sea-floor spreading, a secondary effect, are numerous.

1) The sedimentary sequence of Baffin Bay is common to many coasts broken by continental drift, such as the west coast of Africa and the east coast of South America, and a comparison of Baffin Bay with the rich oil provinces of both sides of the South Atlantic is therefore warranted.

2) This sedimentary sequence includes both hydrocarbon source rocks and great volumes of clastic reservoir rocks derived from the adjacent land masses during the initial stages of their separation. Because of later widening of the rift, these promising sediments now underlie the fault blocks closest to the shoreline, i.e. primarily the continental shelf.

3) The fault blocks broken off by the rifting have invariably been rotated antithetically. This, in combination with cross-faults, has created suitable structural traps for hydrocarbons.

4) The fact that lava flows occurred only after the sedimentary series had been laid down, limits the danger of volatilization of hydrocarbons to the immediate vicinity of the feeders. If sea-floor spreading were the primary cause of continental drift, volcanic extrusions would have contaminated all sediments from the very start. In addition, the area in which volcanic activity took place is quite limited and decidedly does not extend the full length of the Baffin Bay rift.

5) The central depression which formed the last stage of widening of the rift was filled with a thick series of younger Tertiary and Quaternary sediments. These are not underlain by any significant amount of older sediments, such as those that include source rocks. Their locus roughly coincides with the deepest offshore waters. The deeper part of the basin beyond the edge of the continental shelf thus does not appear an attractive exploration objective.

Baffin Bay is filled with sediments originating from the two adjacent land areas, particularly Greenland. The potential reservoir rocks that were deposited in the general Davis Strait area might be of better quality than those off the Labrador coast and southwestern Greenland because of the presumed distribution of the Cretaceous and early Tertiary sediments. The major valleys delineated by seismograph surveys beneath the Greenland ice-cap all drain, or rather used to drain, westward through a narrow gap near Disko Island. A deep fossil river course of submarine channel of Paleocene age was mapped on the south coast of Nûgssuaq Peninsula; the major outlet of Greenland drainage may well have been between this peninsula and Disko Island. Sediments in this part of the Baffin Bay area are to a large extent of terrestrial origin, and primarily consist of sandstone. Farther out to sea, as well as farther north (e.g. northern Nûgssuaq), marine intercalations, including shales, increase in number and thickness, thus providing better chances for caprock development in the areas north and south of Disko Island and near the Baffin Island coast. Farther out again, in southwest Greenland and near the Labrador coast, the section may well consist predominantly of shale, as it does in the Grand Falls well on the Grand Banks off Newfoundland (Bartlett and Smith, 1970, 1971).

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Editor's Note

Dr. Rudolph Martin died in Calgary last April at the age of 60 years. He was born in the Hague, Holland, and obtained his Ph.D. in geology from Leiden University in 1936. That same year he joined the Royal Dutch/Shell Group as a field geologist in Columbia. He also served in Venezuela, the U.S.A. and Indonesia before transferring to Shell Canada in 1955. At the time of his retirement from Shell in 1961, he was in charge of special studies. In 1960, he became a Canadian citizen.

Upon his retirement, Dr. Martin started a consulting practice, Rudolf Martin and Associates, and a year later a private oil company, Nitram Exploration. Subsequently, the activities of both companies expanded into all the continents of the world.

Starting in 1931, Dr. Martin published a total of 24 scientific and technical papers, including six on tectites one of his hobbies since his student years, and six on paleogeomorphology, a subject on which he was considered to be an authority. The list of geologists who received reprints of his papers included well over 500 names, reflecting the time and effort he spent keeping in contact with his colleagues around the world.

Dr. Martin belonged to twelve professional and technical societies in Canada and abroad, including fellowships in the Geological Society of America, the Institute of Petroleum, and the Geological Society of London.

The foregoing obituary was written by H.K. Roessingh of Calgary who was a personal friend of Dr. Rudolph Martin.

(Earth Science Symposium on Offshore Eastern Canada,
Geol. Surv. Can., Paper 71-23, 1973, p. 621-631)

37. GEOPHYSICAL AND GEOLOGICAL STUDIES IN BAFFIN BAY AND THE LABRADOR SEA

R.D. Hyndman, D.B. Clarke, H. Hume, J. Johnson, M.J. Keen, I. Park, and G. Pye, Departments of Oceanography, Geology and Physics, Dalhousie University

Abstract

Geophysical and geological studies of the 1970 cruise of CSS DAWSON to Baffin Bay, Lancaster Sound and Labrador Sea included seismic reflection profiling, magnetic surveys, rock dredging, sediment coring and heat flow measurements. The offshore extension of the Greenland - Davis Strait Tertiary basalts was outlined by magnetic profiling and by dredging. Seismic profiles showed deep sediment-filled graben structures in Melville Bay and Lancaster Sound. The continental-shelf edge off Baffin and Bylot Islands is marked by a pronounced magnetic anomaly and by a transition from disturbed landward to undisturbed seaward sedimentary structures. Central Baffin Bay is filled with at least 2 km of flat-lying sediments. Normal geothermal heat flow in Baffin Bay and Labrador Sea indicates that there has been no significant spreading or opening of the area for at least 20 million years.

INTRODUCTION

This paper will describe some of the geophysical and geological measurements obtained on the 1970 cruise of CSS DAWSON to Bafffin Bay, Lancaster Sound, and Labrador Sea (Fig. 1). The measurements included seismic reflection profiling, magnetic surveys, rock dredging, sediment coring and heat flow. Two areas were studied in detail; the Davis Strait area where there are occurrences of Tertiary basalts partly over-lying older sediments, and north-central Baffin Bay where there is a concentration of earthquake activity and where previous surveys had suggested the possibility of weak ocean-floor magnetic lineations.

Baffin Bay, like Labrador Sea, has been shown to be a small ocean basin. Bedford Institute seismic refraction measurements indicated that there is oceanic crustal structure under the central basin (Barrett et al., 1971). Studies in the Labrador Sea, the ages of continental tectonic deformation and the age of the basalts on either side of Davis Strait all suggest that Baffin Bay opened between about 75 and 50 million years ago from a now extinct mid-ocean ridge (see Keen et al., 1970; LePichon et al., 1971).

We would like to consider four aspects of the Bay and relate them to the Cretaceous-Tertiary history. These are: 1. the basalts of Disko Island and Baffin Island and their extension offshore in Davis Strait; 2. The sediments and structures of the west Greenland shelf; 3. The sediments and structures of northern Baffin Bay and Lancaster Sound; 4. Heat flow.

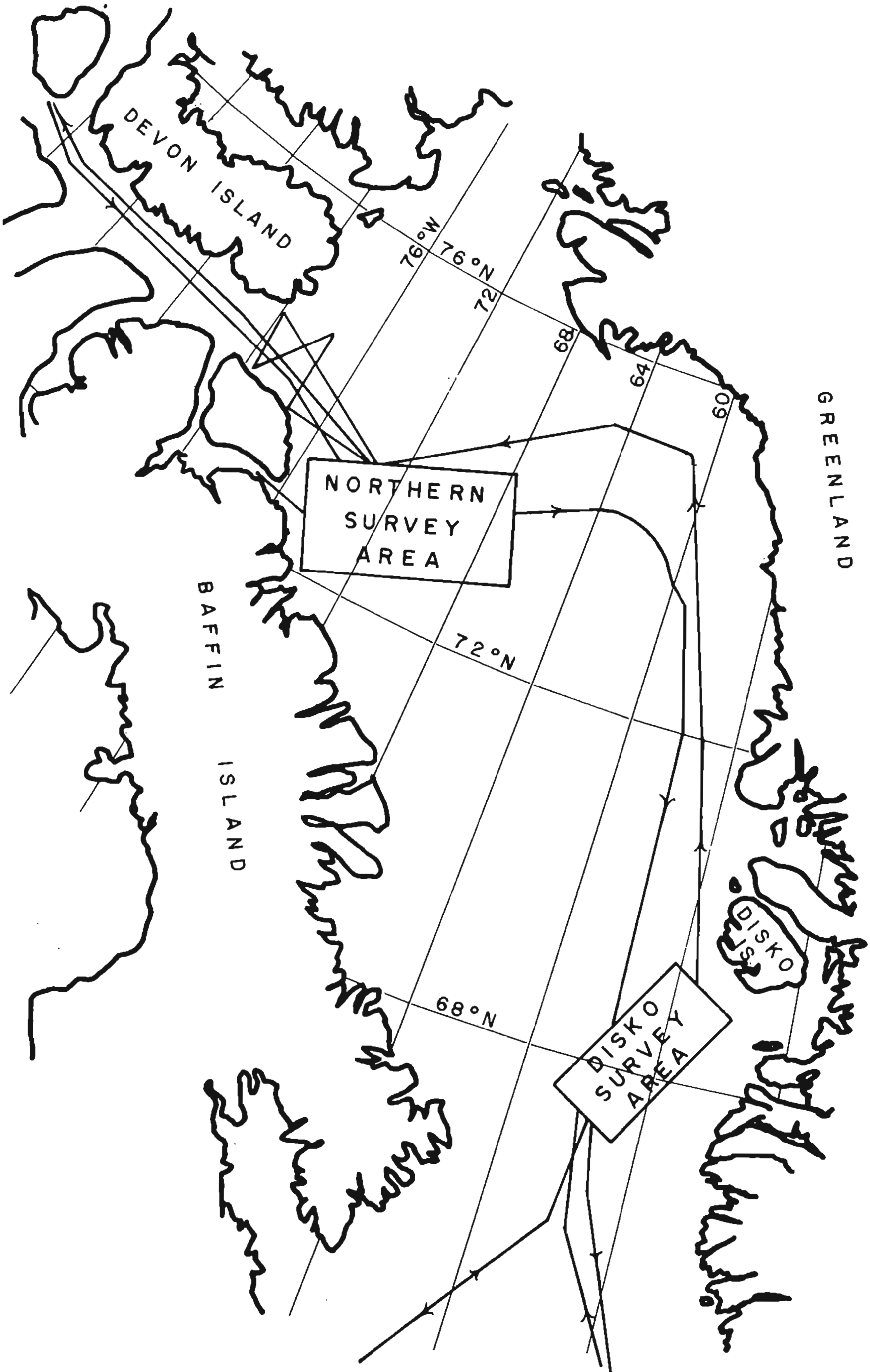

Figure 1. Location of measurement areas in Baffin Bay and Davis Strait.

Basalts associated with rifting across Davis Strait

The Davis Strait area appears to hold several important keys to the opening of Baffin Bay. In any rigid-plate reconstruction of the opening of the Bay, there is a serious overlap at Davis Strait (e. g. Bullard et al., 1965). Consequently there must have been adition of new material in the Strait area during or since opening or there must have more extension of the continental crust in this region than elsewhere along the rift. Both of these are indicated by the Tertiary (54-58 m. y.) basalts found in two restricted land areas on either side of the Strait (Clarke and Upton, 1971; Clarke and Pedersen, in press). Clarke (1970) has shown that the two occurrences are petrologically very similar. Thus it seemed important to delineate the structure and offshore extent of the basalts.

Manchester (1964) first pointed out that the magnetic anomalies in Davis Strait were large, often suggestive of highly magnetic rocks at shallow depth and perhaps having normal and reversed magnetizations. Sharp high-amplitude anomalies were also apparent in aeromagnetic profiles (Hood et al., 1967). It seemed likely that we could map any offshore extension of the land basalts by a magnetic field survey. We chose the west Greenland side of Davis Strait because of the difficulties of surveying in the ice cover off south-eastern Baffin Island. Lines were run at a spacing of about 12 km.

The first magnetic profile off Disko Island showed very large (500-2000 gammas) amplitudes with short wavelengths (0. 5 to 2 km). The wavelengths were obviously too short to permit contouring the data with the line spacing for the survey. However, the contact between the region of large short-wavelength anomalies and that of small broad anomalies is sharp so that a map delineating the basalt boundary could readily be prepared (Park et al., 1970) (Fig. 2).

It seemed likely that the intense anomalies corresponded to the offshore extension of the basalts and the "quiet" regions to Precambrian rocks or to sediments overlying normal deep-ocean floor. This correlation was confirmed on an approach to Disko Island where the Precambrian basement could be seen onshore. Also three dredge stations were attempted on the margins of the intense anomaly area, where the scarps were steep enough to obtain rock samples. All three attempts dredged fresh basalts.

Consequently it seems likely that a large part of the shelf off Disko Island is underlain at shallow depth by Cretaceous-Tertiary basalts. Furthermore the magnetic survey results show that they extend much further north than had been suspected, considerably to the north of the northernmost onshore outcrop. The offshore basalt area outlined is thus greater than the onshore occurrences (Fig. 3). Probably the only difference between the terrestrial-continental shelf basalts of Davis Strait and basaltic crust of the deep part of Baffin Bay is that in the former area consistently more magma was generated during spreading. It is possible that this early Tertiary 'hot-spot' in Davis Strait is now located by lithospheric plate motion under Iceland. Morgan (1971) has suggested that such hot spots represent vertical convective plumes in the mantle.

The eastern boundary of the extension of the Disko basalts is partly marked by a longitudinal trough separating the basalts from the Precambrian rocks. This trough could be the downfaulted eastern boundary of Cretaceous-Tertiary sediments (e. g. Rosenkrantz and Pulvertaft, 1969; Henderson, 1969) which underlie the basalts. On its western boundary the basalt outcrop is

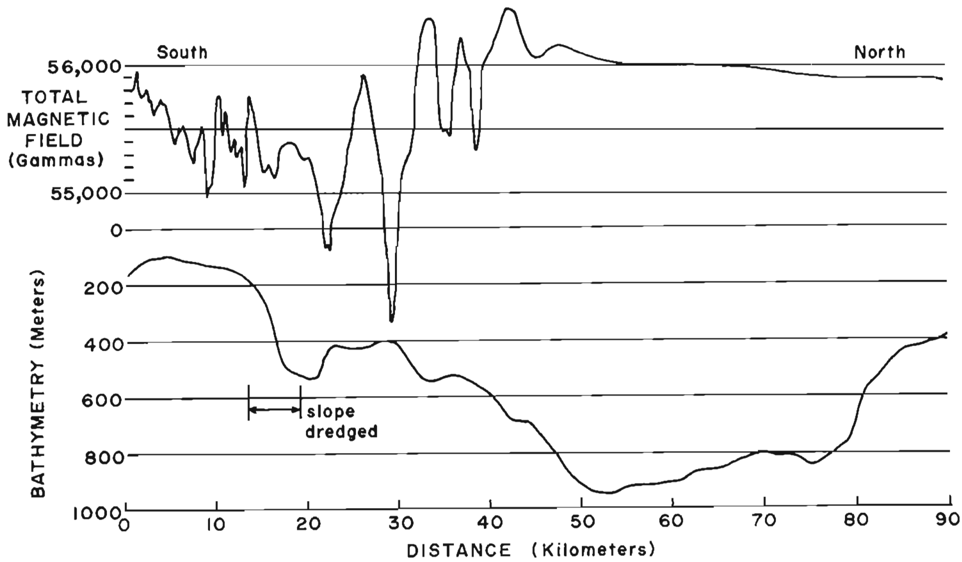

Figure 2. Bathmetric and magnetic profile across the edge of the Tertiary basalts off Disko Island (after Park et al., 1971).

shown from seismic profiling to abut flat-lying sediments; the basalts in places dipping under the sediments. Under the flat-lying sediments and probably under parts of the basalts are folded sedimentary strata. There is thus a major sedimentary basin in this region of the west Greenland shelf.

Fault-bounded Basins at the Margins – Melville Bay and Lancaster Sound

Baffin Bay, as an ocean basin, probably has rifted margins. We might expect to see evidence for fault-bounded sedimentary basins on the shelves as are found on other continental margins, for example off Nova Scotia (Loncarevic and Ewing, 1967). Because of the extensive glacial erosion one might expect such basins to have a topographic expression. Consequently we carried out a bathymetric survey and seismic profiling in Melville Bay off northwestern Greenland and in Lancaster Sound.

The profile obtained along the length of the deep water trough of Melville Bay shows folded sedimentary strata of unknown age. The profile from the centre of the trough to the shallow water on the west towards the centre of Baffin Bay suggests that the sediment abuts a faulted basement ridge which in turn forms the eastern margin of this part of the main deep basin of Baffin Bay (Fig. 4).

Lancaster Sound has a similar geological structure to Melville Bay. Folded and faulted sedimentary strata lie between Devon and Bylot Islands and generally dip to the north. This field evidence supports the suggestion of Barrett that the Sound is a faulted structure but the sediment thickness is much greater and the faulting more intense than previously suspected. The similarity of these two sediment-filled graben structures supports the suggestion of G. Henderson (personal communication) that they were joined before the opening of Baffin Bay.

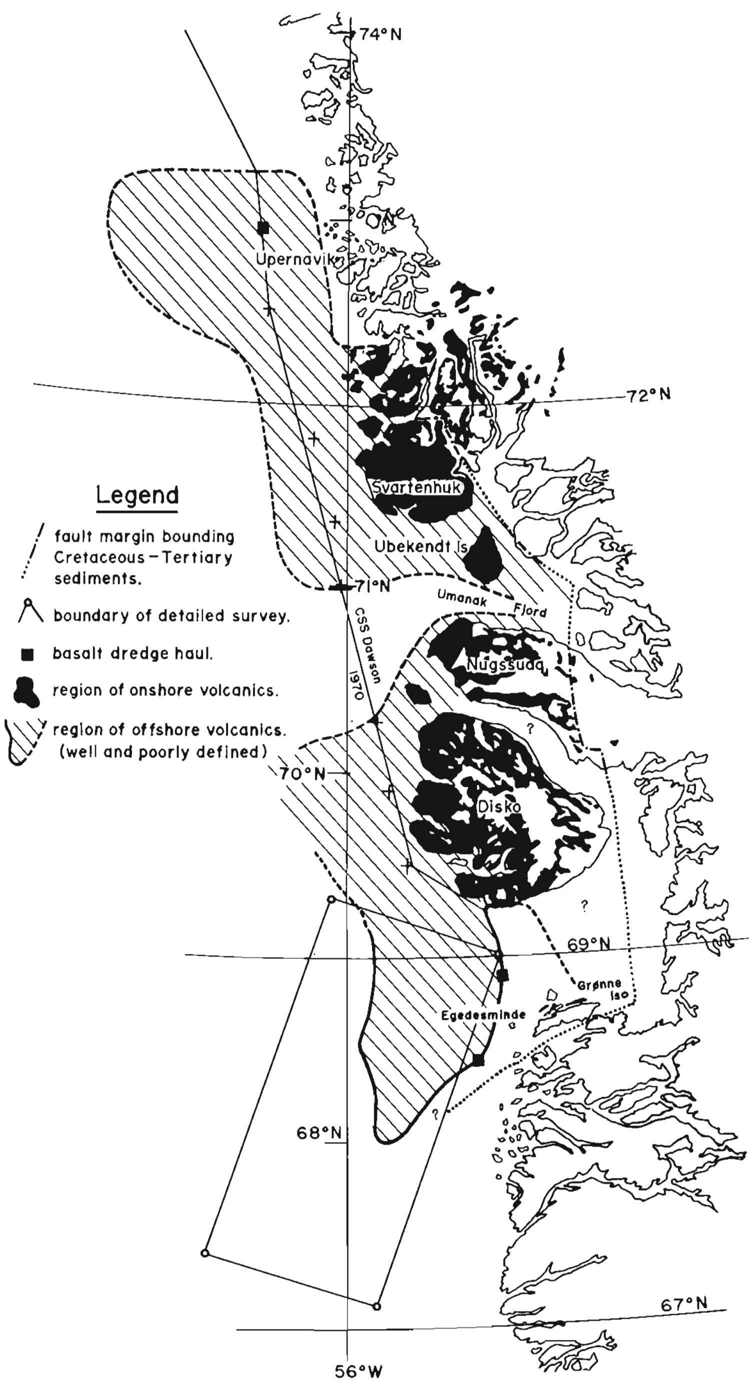

Figure 3. Distribution of Tertiary volcanics on the west Greenland shelf (after Park et al., 1971).

Structure of Sediments in Northern Baffin Bay

Seismic reflection profiles across central Baffin Bay showed a great thickness of undisturbed flat-lying sediments. No clear basement reflections were seen by the thickness must be in excess of 2 km. This is in agreement with the Bedford Institute refraction results and with an age of opening of around 50-75 million years ago. There is no evidence of the effects of the earthquakes with the sedimentary structure. The earthquakes thus do not indicate major, recent tectonic activity.

The magnetic profiles obtained for the northern Baffin Bay survey area were quite flat with no very obvious magnetic lineations which is in agreement with previous measurements (e.g. Keen et al., 1970). However, approaching Baffin and Bylot Islands there is a pronounced magnetic anomaly, to the west of which the sediments are distinctly disturbed while to the east they are undisturbed (Fig. 5). The presence of the magnetic anomaly indicates that the change from undisturbed to disturbed sediment is associated with the structure or topographic relief of the basement. The picture is somewhat different for the mouth of Lancaster Sound. There, the magnetic anomaly is less pronounced. At the mouth of Lancaster Sound there are great thicknesses of prograding sediments indicating large sediment transport out of this channel.

Recent Sediments in Baffin Bay

The distribution of surface sediments in Baffin Bay is largely determined by ice-rafting of material, where the ice movement is controlled by surface currents. There is also indication of turbidite sequences from cores obtained to the east of Lancaster Sound.

Detailed sediment sampling was undertaken to the southwest of Disko Island in Davis Strait. The pattern of sediments is very irregular, but in general the average grain-size decreased toward the central basin. The route of icebergs in the west Greenland Current is delineated by a large fraction of gravel and boulders. This path is in agreement with iceberg sightings: the movement is into Disko Bay, along the south coast of Disko Island and up the west coast of the Island. More detailed work is being carried out on sediment-size distribution, colour, organic content and foraminiferal content.

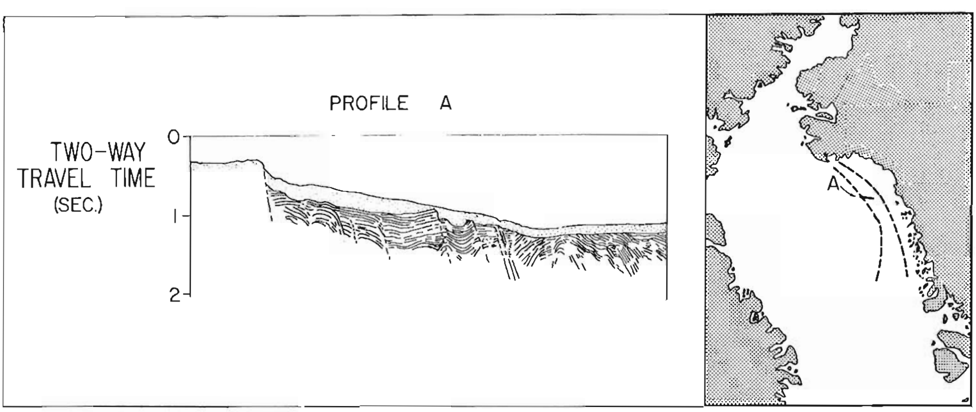

Figure 4. Seismic cross-section in Melville Bay.

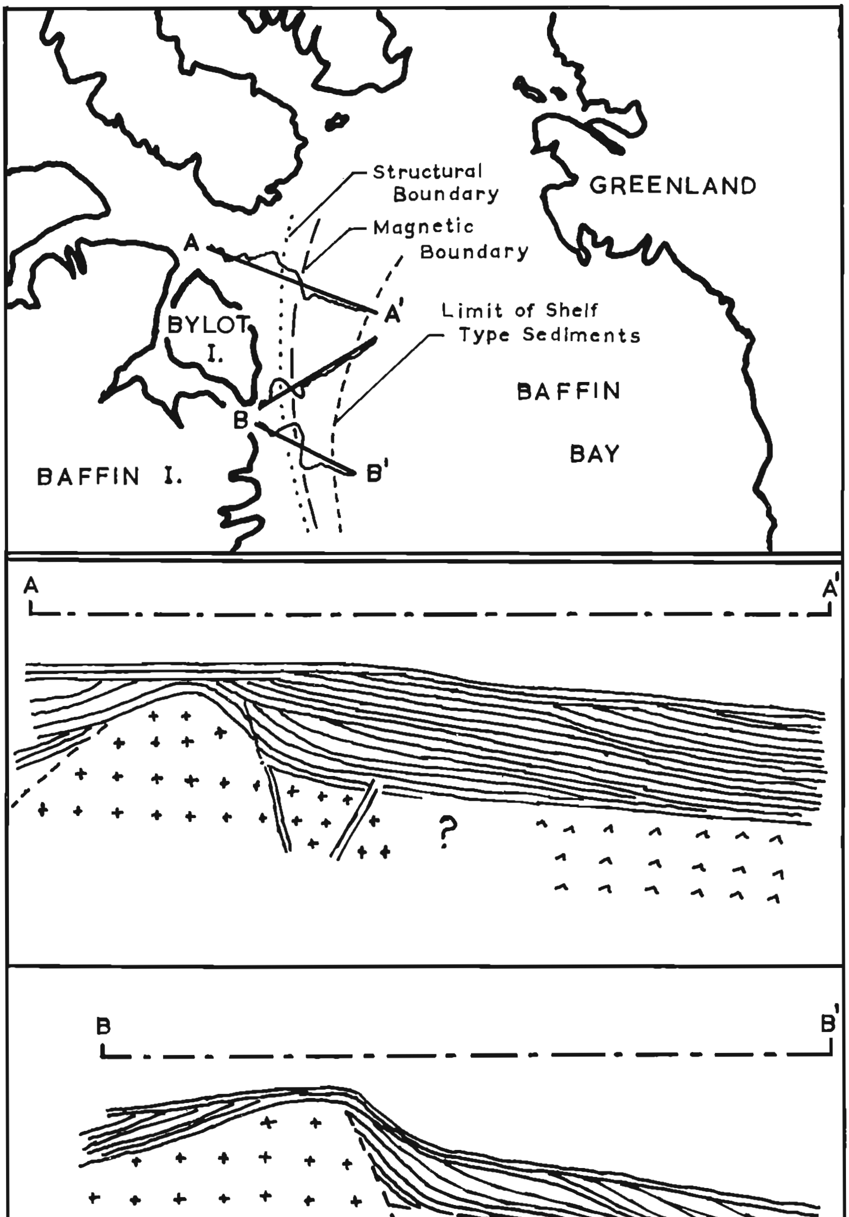

Figure 5. Seismic cross-sections off Bylot Island, with the associated magnetic profiles.

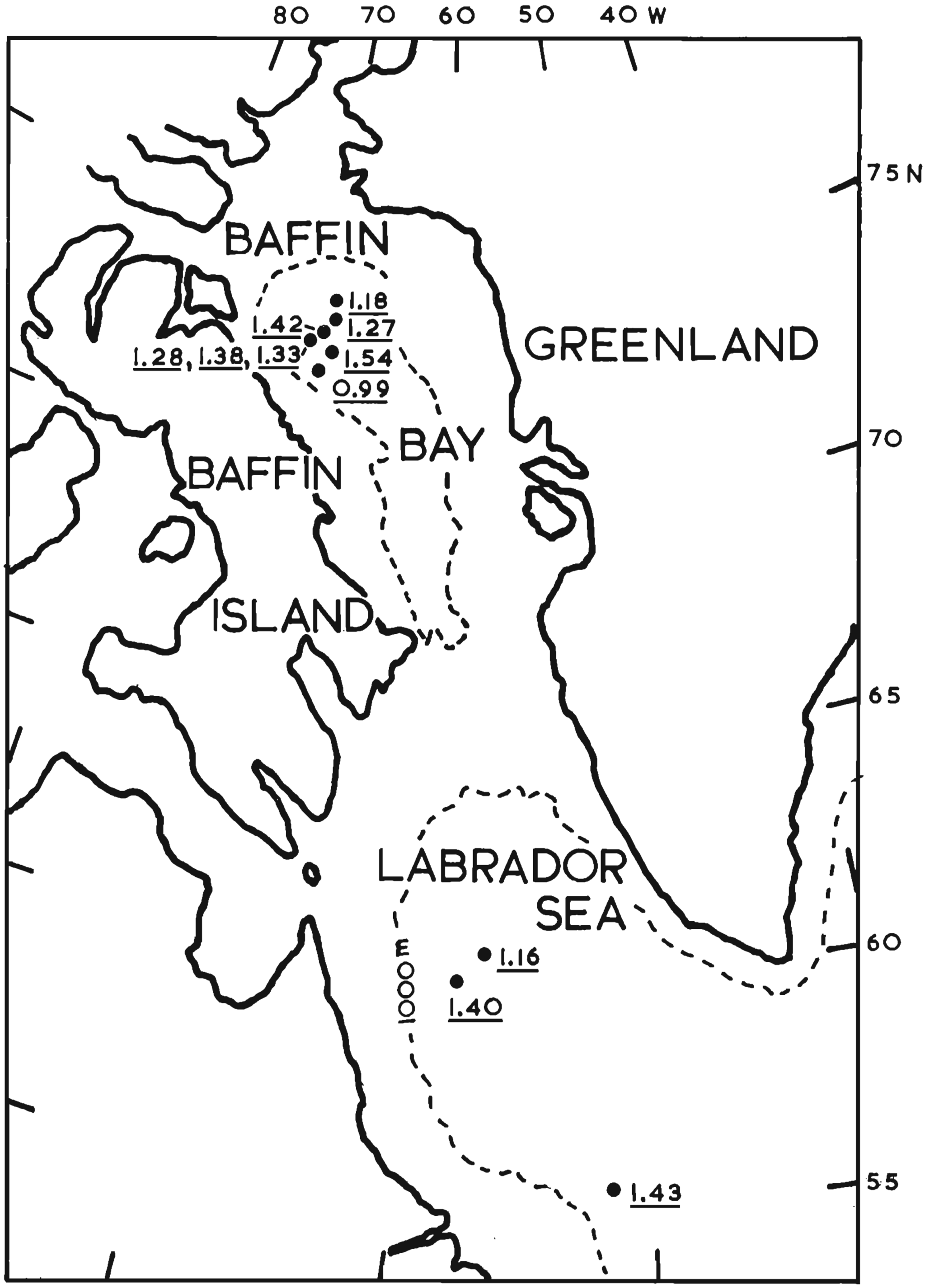

Figure 6. Heat flow measurements from Baffin Bay and the Labrador Sea (after Pye and Hyndman, 1972).

Heat Flow in Baffin Bay and the Labrador Sea

High heat flow is characteristic of all actively spreading ridges. A crucial test for recent active spreading in Baffin Bay and the Labrador Sea was to see if heat-flow values were normal. A series of 7 successful heat-flow measurements were made in Baffin Bay and 3 in the Labrador Sea (Pye and Hyndman, 1972). Each of the Baffin Bay results showed a non-linear temperature profile with depth in the sediment. Three repeated measurements at the same location over a period of 17 days gave different gradients. We were able to deduce that a single bolus of warm (0.30° C) water moved through the area 6 to 7 weeks before the first measurement with a duration of 2 weeks. There is some evidence that the bolus moved from north to south. Water at this temperature has been measured to the west of the sill at the entrance to Lancaster Sound and we postulate that perhaps once a year during major storms boluses of deep dense water move over the sill. Such water masses had been suspected in Baffin Bay by physical oceanographers but were never previously detected. This is a similar but less complicated situation to that observed by Lachenbruch and Marshall (1967) in the Norwegian Sea. Heat flows in Baffin Bay corrected for this disturbance and for the high sedimentation rate have a mean value of 1.35 ± 0.14 $\mu cal/cm^2$ sec. (Fig. 6). This normal heat-flow value indicates that there can have been no very recent spreading and there is no extensive vulcanism associated with the concentration of earthquakes in north central Baffin Bay.

In the Labrador Sea, the mean of 3 heat-flow values is 1.33 ± 0.15 $\mu cal/cm^2$ sec. which is also normal. One of the measurements is located where our seismic profiling results and those of Le Pichon et al., (1971) indicated that a portion of the buried mid-Labrador Sea ridge to be situated. These heat-flow values confirm that there has been no significant ocean-floor spreading in Baffin Bay and the Labrador Sea for at least the past 20 million years.

ACKNOWLEDGMENTS

We wish to acknowledge the assistance of the officers and crew of CSS DAWSON during the cruise. Mr. R. Gerstein and Mr. J. O'Byrne provided the best of technical assistance. We would also like to express our appreciation to the Atlantic Oceanographic Laboratory of Bedford Institute. The work was support by the National Research Council, Defence Research Board and Dalhousie University.

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(Earth Science Symposium on Offshore Eastern Canada
Geol. Surv. Can., Paper 71-23, 1973, p. 633-638)

38. GEOPHYSICAL STUDIES ON THE STRUCTURE OF BAFFIN BAY

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Abstract

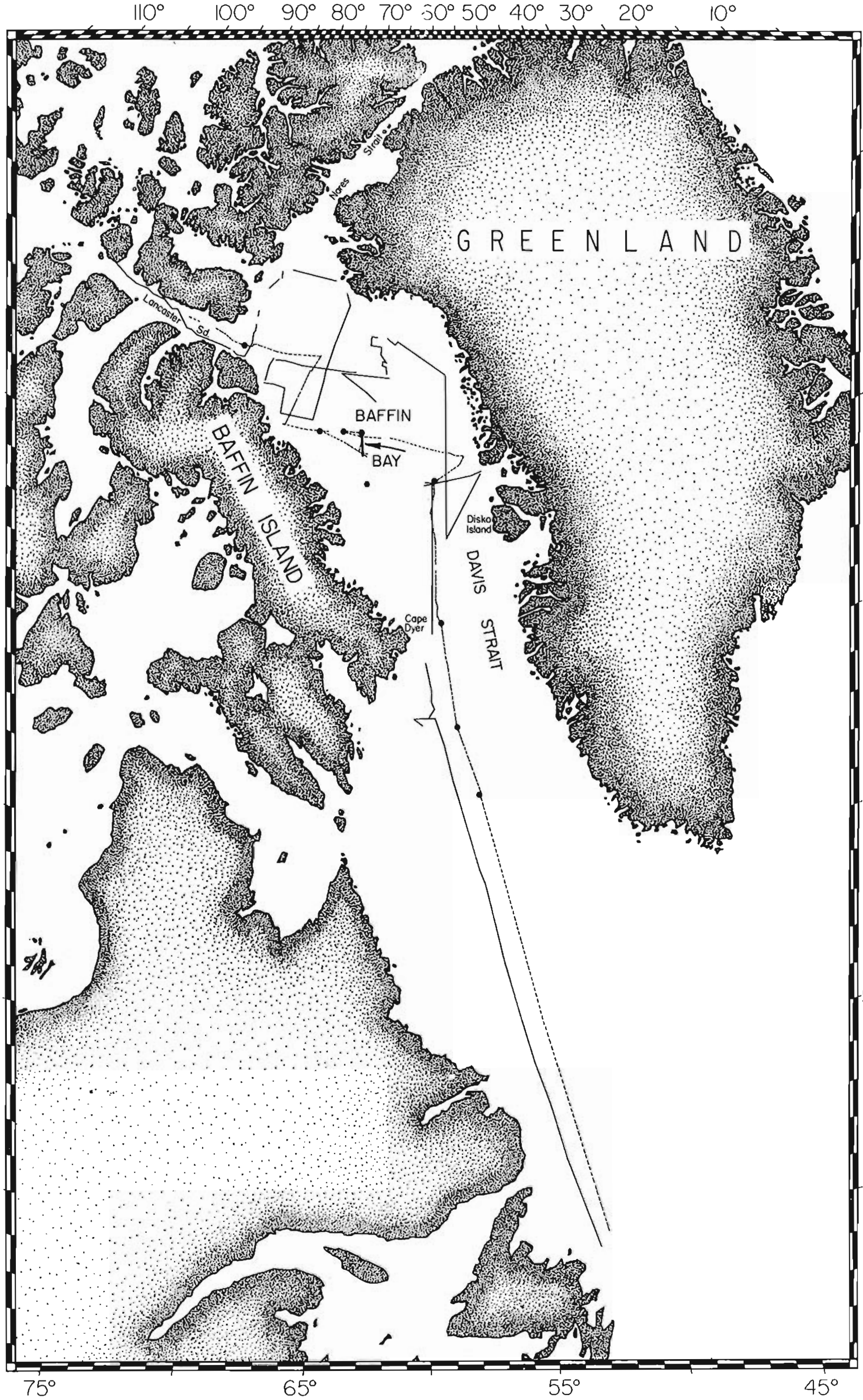
The addition of marine gravity and seismic data obtained during the summer of 1970 to the already extensive collection of magnetic data held at the Bedford Institute has made it possible to define the structure of central Baffin Bay, obtain information on the oceanic-continental transition zone on both sides of the Bay and attempt a preliminary reconstruction of the opening of the Bay. This note briefly summarizes the findings to date. A more complete discussion of the results and their implications has been given by Barrett et al. (1971), Keen et al. (1972) and Manchester and Clarke (1971).

INTRODUCTION

Baffin Bay provides an important geographic link between two large oceanic areas, the Arctic Ocean to the north and the Labrador Sea to the south. Reconstructions of plate motions in both regions must be reconciled with the continental geology surrounding the Bay and with geophysical measurements defining its structure and history. Past attempts to explain its origin and subsequent evolution have been limited by a lack of geophysical measurements within the area; the evidence being substantially confined to the geology of the surrounding land masses and to some knowledge of the opening of the Labrador Sea to the south. Continental geology is of limited usefullness in defining Cenozoic and Mesozoic plate motions because, with a few important exceptions, only Precambrian and Lower Paleozoic rocks are exposed along most of the surrounding coastline. Previous geophysical studies of the region have been inconclusive. No definite evidence of a buried median ridge or of magnetic lineations indicative of sea floor spreading have been found (Keen et al., 1970 ; Barrett, 1966). The fact that Baffin Bay is a sea of intermediate depth, reaching 2000 metres in its central region, has led some investigators to suggest that the area may not be underlain by oceanic crust. An additional complexity is the substantial number of shallow focus earthquakes in the area; the activity on Baffin Island and the northern portion of the Bay being significantly greater than elsewhere in the Eastern Arctic (Whitham et al., 1970) or in the Labrador Sea.

During October 1970 the Bedford Institute ships C. S. S. HUDSON and C. S. S. BAFFIN conducted geophysical investigations in Baffin Bay en route from survey projects in the Western Arctic. The measurements made were designed to answer three basic questions.

1. Is the crust beneath Baffin Bay oceanic or continental?
2. If it is oceanic, what is the geometry of the oceanic area and the nature of the oceanic-continental boundaries?
3. Is it possible from considerations of the known geology and geometry of the area to determine the probable history of the formation of the Bay?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GREENLAND
Nares Strait
Lancaster Sd
BAFFIN
BAY
BAFFIN ISLAND
Disko Island
DAVIS STRAIT
Cape Dyer

Geophysical Results

The nature of the crust beneath the deep central region of Baffin Bay was determined by a two-ship seismic refraction experiment carried out in cooperation with U.S.C.G.C. EDISTO (Fig. 1). The results of this work have been described by Barrett et al. (1971). The layer velocities and thicknesses obtained on the reversed line are reproduced in Table 1. The velocities of the upper layers are uncertain and have been assumed to be 2.3 km/sec and 3.5 km/sec respectively for purposes of the calculations. However, the other layers have been adequately defined and the depth to the M-discontinuity, 10 km, is unlikely to be in error by more than 20 per cent because of these assumed velocities.

The results of this refraction experiment have proved that the central region of the Bay is underlain by oceanic crust. The outline of the oceanic crust can be mapped if the oceanic-continental transition zone can be determined. Gravity measurements across the margins of the Bay have enabled the transition zone to be mapped on the east and west sides of the Bay. More information is required to accurately define the northern and southern boundaries. The tracks on which the gravity data were obtained are shown in Figure 1. The profiles across the transition zone show large, positive "shelf-edge" anomalies (Fig. 2). Theoretical models which satisfy the gravity observations have been calculated and these confirm that the anomalies can only be explained reasonably by the transition from continental to oceanic crust (Keen et al., 1972). Seismic reflection records obtained along with the gravity data across the Baffin Island shelf show that a change in the nature of the sediments occurs at the transition zone; from highly deformed on the landward side to gently prograding to seaward.

The continental-oceanic boundary determined from the gravity measurements (Fig. 2) corresponds also to a boundary across which there is a distinct change in the character of the magnetic records. Landward of the boundary the magnetic field contains large amplitude, high frequency anomalies while seaward the anomalies are of much lower amplitude (less than 300 gammas) with longer wavelengths, typically around 30 km (Hood and Bower, 1970; Manchester and Clarke, 1971; Keen et al., 1972).

Seismic reflection data obtained on a line across the Bay confirmed the presence of a thick sequence (2-3 seconds two-way travel time) of flat lying sediments in the deep region. Underlying this sequence there is some indication of a rough reflector. This reflector may correspond to the boundary between the two upper most layers (assumed velocities of 2.3 km/sec and 3.5 km/sec respectively) determined from the refraction data. Because of the weak and discontinuous nature of the reflector and the poor definition of the two upper layers obtained from the refraction measurement, it is impossible to determine whether this interface represents layering in the sedimentary sequence or the top surface of the oceanic basaltic layer.

Gravity measurements at the entrance to Lancaster Sound (Fig. 2) show that this area corresponds to one of large negative free-air gravity.

Figure 1. (opposite)
Track charts of C.S.S. HUDSON and C.S.S. BAFFIN in Baffin Bay. In addition to gravity and magnetic measurements carried out by both ships, seismic reflection data were recorded on HUDSON whenever possible. The location of the reversed seismic refraction line is indicated by the heavy solid line.

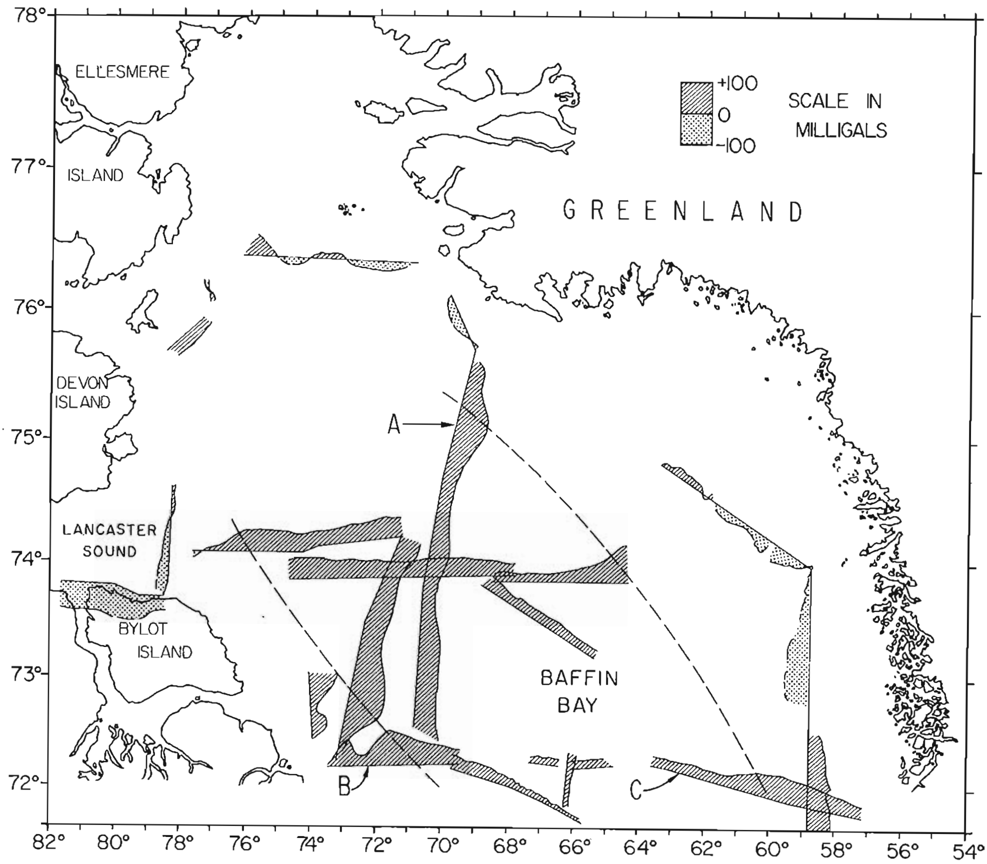

Figure 2. Gravity profiles across the margins of Baffin Bay obtained on HUDSON and BAFFIN. The dashed lines outline the oceanic-continental boundaries as defined by the gravity "shelf-dege" anomalies. The letters A, B, C refer to profiles for which model calculations have been carried out.

Seismic reflection data indicates considerable deformation of the sediments at the entrance to the Sound, the transition from a deformed sequence to an undeformed prograding shelf corresponding to the change from negative to positive free-air gravity associated with the continental edge. In this area the continental-oceanic boundary as defined above occurs about 100 km to the west of the increase in water depth corresponding to the topographic expression of the continental margin. Seismic velocity determinations from a short sonobuoy refraction line in the area of deformed sediments at the entrance to the Sound indicate velocities of 3 km/sec or less in the top 1 km of the section, suggesting that these sediments may not be older than Mesozoic.

TABLE 1

Layer Thicknesses and Velocities

Layer	Velocity km/sec	Thickness in km Southern Station	Northern Station
Water	1.464	2.3	2.3
Unconsolidated Sediment	2.30	2.1	2.0
Consolidated Sediment?	3.47	2.0	2.3
Layer 2	5.72	2.0	2.1
Layer 3	6.93	1.9	2.1
Mantle	7.68	-	-

NOTE: The positions of the southern and northern stations are 71°39'N, 65°55'W and 72°18'N, 66°13'W respectively. These are median positions taken over the time that C.S.S. HUDSON was on station.

CONCLUSIONS

Keen et al. (1972) have discussed the implications of existing geophysical measurements in Baffin Bay in terms of sea floor spreading and the history of the opening of the Bay. Although there is no conclusive evidence for a buried median ridge beneath the Bay or definitive evidence for the existence of lineations in the low amplitude oceanic magnetic anomalies a preliminary attempt to define the geometry of the opening of the Bay can be made. The reconstruction discussed by Keen et al. (1972), suggests two stages in the formation of the Bay. The first motion occurred around a pole (77°N, 100°W) north of Bathurst Island. During this stage an appreciable portion of Labrador Sea was formed but very little opening of Baffin Bay occurred. The second stage resulted in the formation of the major part of the oceanic crust in Baffin Bay. The present geometry of the oceanic area requires that Nares Strait acted as a transform fault during this phase of the opening and that a displacement of 150 km occurred along it. The restriction that Nares Strait acted as a transform fault puts the pole for this phase of opening in the region of the central Pacific (e.g. 19°N, 173°W).

ACKNOWLEDGMENTS

We are very grateful to the Masters, Officers and Crew of C.S.S. BAFFIN, C.S.S. HUDSON and U.S.C.G.C. EDISTO for their help and cooperation in obtaining the information reported here. We thank all our colleagues, in particular T.B. Smith and J.M. Woodside, who helped collect and process the data and participated in subsequent discussions.

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39.

ICEBERGS: A NEW PROBLEM FOR OIL EXPLORATION AND PRODUCTION

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Abstract

In offshore Eastern Canada and Western Greenland, the sea ice limits the drilling season for floating platforms and during the drilling season a major danger is the threat of icebergs. When an iceberg is coming towards the drilling ship, and if a decision is made to move off site, drilling operations must be stopped at a certain time before the drilling platform can be moved. This time multiplied by the speed of the icebergs provides the concept of warning distance. It is necessary to detect all the icebergs trespassing the warning area. The path of an iceberg is very erratic and it is difficult to predict its short-term route and to make the subsequent decision to move or not. It is feasible to tow the icebergs before they enter the warning area but provisions must also be made to move should there be a failure in the towing operations.

During the winter, there are the cumulative problems of ice plus icebergs. For development of oil or gas fields, the use of permanent platforms is not feasible. A suggested solution is to drill in the summer and to use submarine completion, separation and pipes. This submarine equipment must be protected from the scouring of icebergs.

INTRODUCTION

One of more important considerations associated with offshore exploration on the East Coast of Canada and the West Coast of Greenland (Fig. 1) is the occurrence of icebergs. For the past two years, Total Compagnie Française des Petroles assisted by Marine Exploration Ltd., England, has been studying the many problems posed by the threat of icebergs to offshore oil operations. From these studies it became apparent that the first question to be answered was the probability of their occurrence.

OCCURRENCE OF ICEBERGS

Grand Banks of Newfoundland

In 1968, Blenkarn and Knapp (1969) published a study of iceberg conditions in the Grand Banks between 43° and 47°N. For their calculations, they used the sighting data published by the International Ice Patrol Service. This agency has for many years covered the area south of 50°N extensively during the period March to July. For much of this area, they concluded the

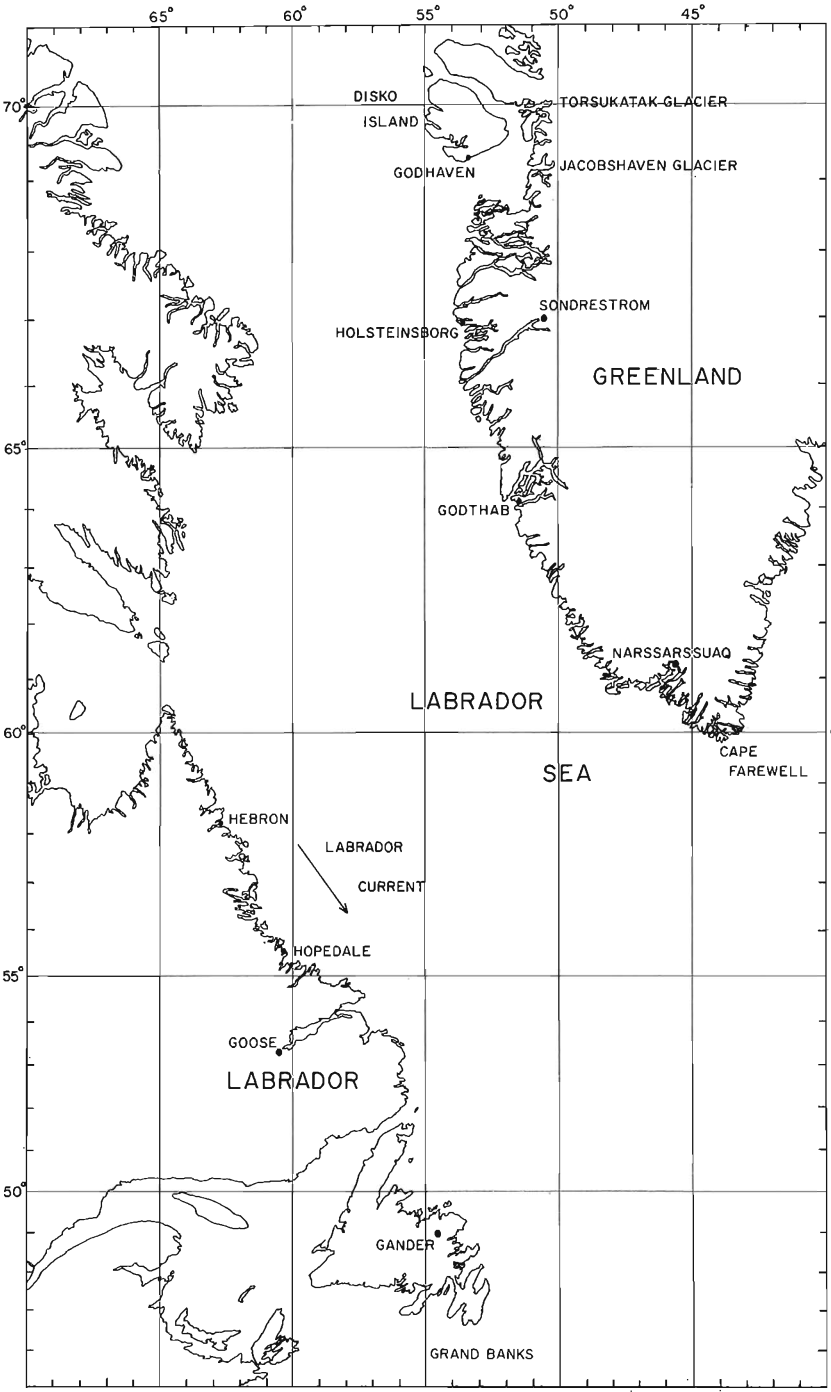


Figure 1. Ports and airports bordering the Labrador Sea.

likehood of iceberg collision is not great and that a berg collision would not always cause catastrophic loss of a fixed platform. Their conclusions are probably reliable due to the fact that the data were consistent and as most of the icebergs (average 380) pass Lat. 48°N between March and June, an exploration drilling program conducted during the other months largely eliminates the iceberg problem.

Labrador Sea

For the continental shelf of Labrador (50°N to 60°N), there was very little data available prior to 1965. Subsequently the area has been covered by the International Ice Patrol about once a month from February to June but the data still remains sparse for the other months (International Ice Patrol, 1965-1969; and Murray, 1969). Based on available data, calculations of the average lifetime for a fixed drilling structure resulted in indicating an average lifetime in the area which increased from two years at 60°N to 10 years at 50°N. As most of the icebergs occur in the months when good data is available, the calculations are reliable and it may be assumed that permanent platforms are not feasible for most of the Labrador Sea.

In planning a drilling program in the vicinity of 55°N in the Labrador Sea, the ice-free season is expected to be from July to December (Fig. 2) and due to the heavy seas which can occur in November, it is preferable to arrive

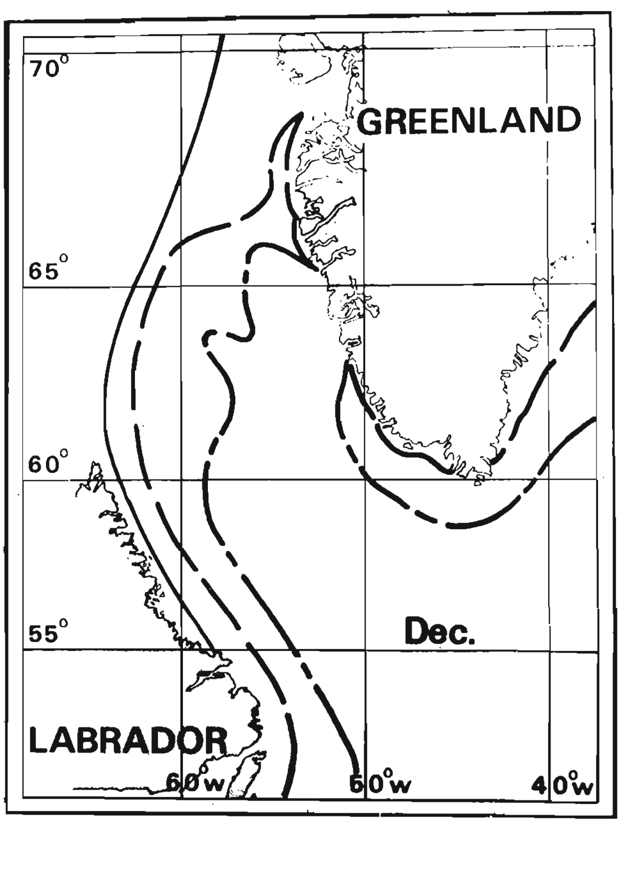


Figure 2.
Mean ice limits in December (U.S. Navy H.O., 1968). The solid line refers to the minimum ice limit, the dashed line to the mean ice limit and the long-dashed, short-dashed line to the maximum ice limit.

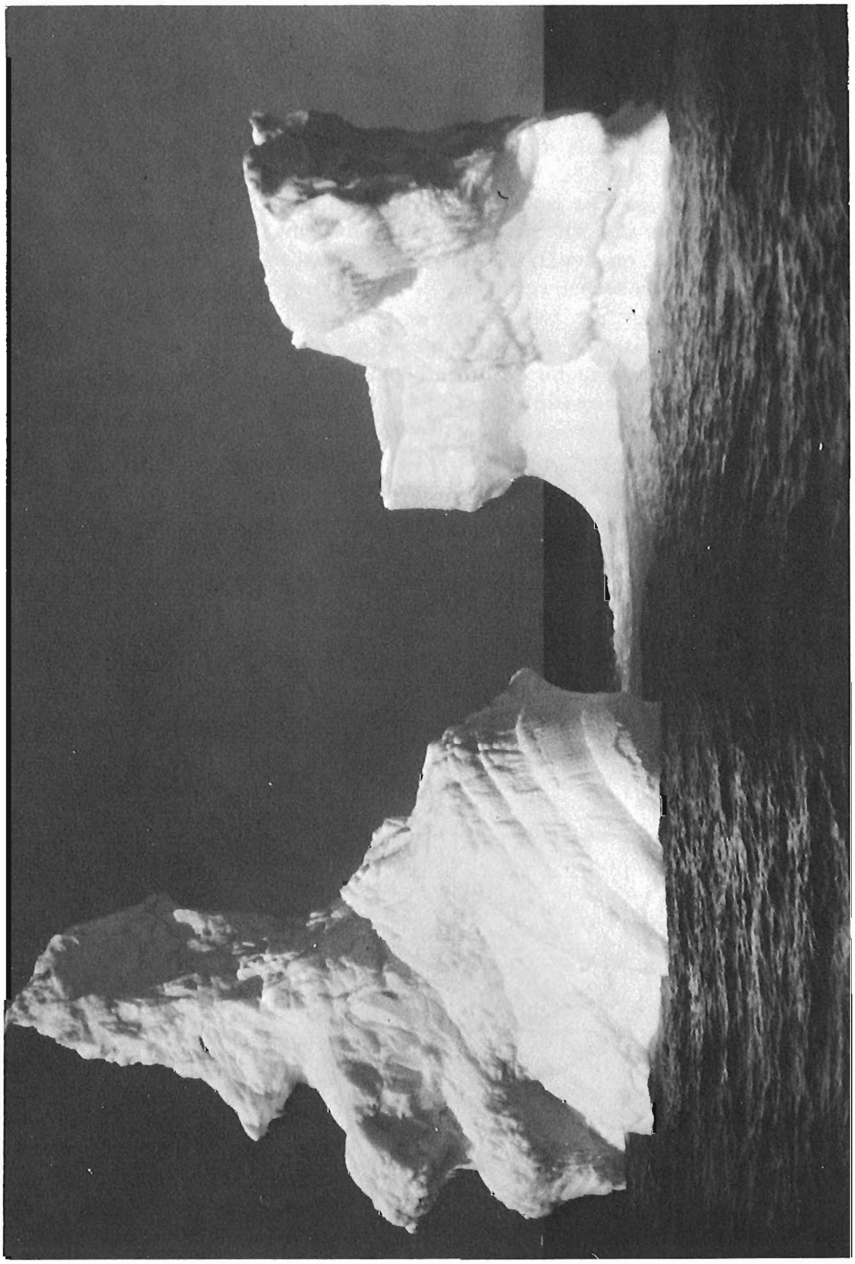


Figure 3. Small eroded iceberg (40 m long).

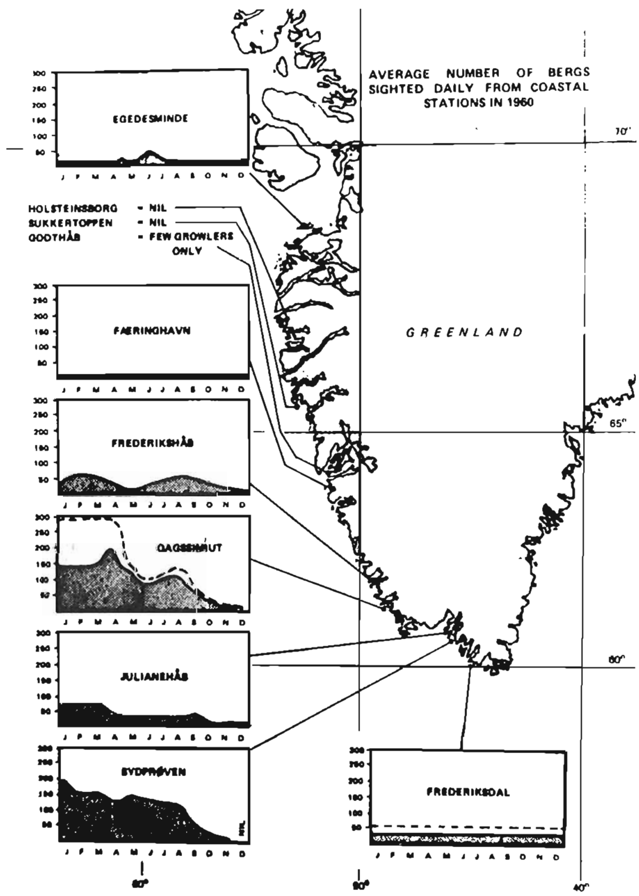


Figure 4. Average number of icebergs sighted daily from coastal stations in 1960.

on the site as soon as ice conditions will permit. Estimates of the risk involving an iceberg collision with a drillship in this area is interpreted from the poor and sparse data which exists for the summer months (Duval, 1970). For a summer operation, the probability of an impact with a drilling ship is less than 0.10 and if it is assumed that the ship must move when an iceberg is closer than one-half mile, the probability of not having to move is about 0.60. Thus, for any planned drilling programs in this area, the drilling ship must be ready to move quickly to avert an iceberg collision.

Southwest Greenland Shelf

For southwest Greenland, results of flight and coastal station observations have been published in "Ice Conditions In Greenland Waters" (Meteorologiske Institute, 1957-1962). Flight data is sparse and coastal station data is not very reliable because the number of icebergs actually counted depends on the angle of visibility, visibility, and density of grounded icebergs.

Icebergs calve on the east coast of Greenland, go south to Cape Farewell and then north with the West Greenland current. Their size decreases considerably while going north due to the melting caused by this relatively warm current, and the breaking and the grounding on the banks. At 64°N, some of these bergs continue north and others go west and then south under the effect of the Labrador current. These icebergs are very eroded (Fig. 3). Our interpretation of the data in this area has shown that the iceberg density per square mile varies from 0.07 at 60°N to less than 0.001 at 67°N in May. The maximum density is encountered in September, October and November (Fig. 4). Yearly variations from 0.5 to 2.5 can be expected.

Many icebergs calve from Torsukatak and Jacobshavn Glaciers into Disko Bay and these icebergs are usually bigger than those found further

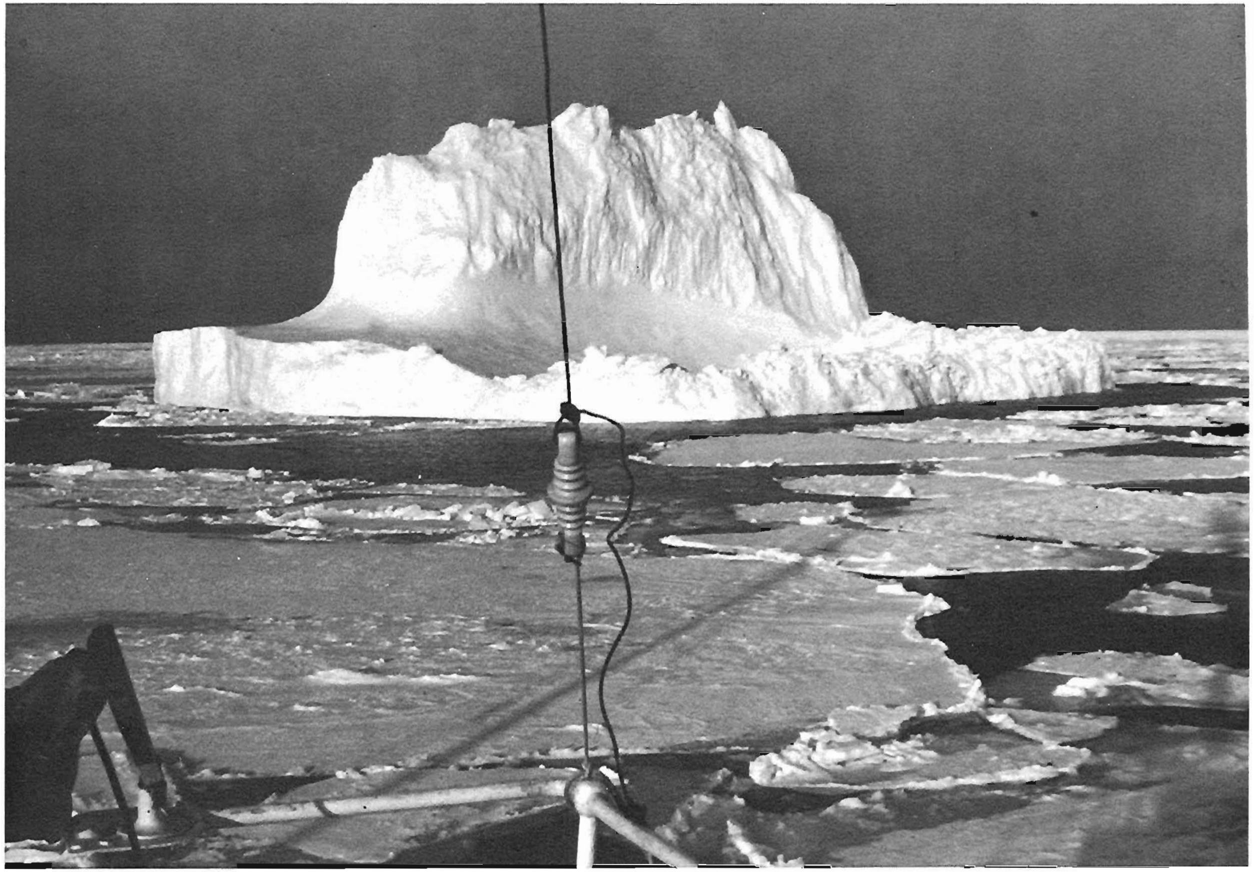


Figure 5. Large iceberg (142 m long, 34 m high).

south (Fig. 5). They leave the Bay and usually follow the route northward. Often, due to the effect of a northern wind, they go south, but generally not farther than Holsteinborg. Iceberg density in offshore Disko has been estimated at 0.08 in May. The maximum density is encountered in February, March, June, July and August, and it is about 1.5 times the density in the other months. Yearly variations from 0.75 to 1.50 can be expected.

In considering exploration drilling in this area, it is again important to know the probability of collision between an iceberg and the drilling platform. Calculations of this danger based on available data (Meteorologiske Institute, 1957-1962) during the ice-free season indicates a possible 8 collisions south of 62°N, less than 1 between 64° and 66°N and less than 0.1 at 67°N. Offshore Disko Island, the number of collisions has been estimated at between 5 and 15 depending on the exact area. Thus, to drill with safety in these areas, it would appear that new techniques will have to be developed.

FEASIBILITY OF OFFSHORE OIL OPERATIONS

When drilling offshore, several hours must elapse between the moment a decision is made to stop the operations and the moment it becomes possible to move the drilling vessel. Depending upon the operations, the time to secure the well and disconnect the riser varies from less than 1 hour to more than 1 day. The dynamic-positioning ship PELICAN, currently being built for SOMASER, which will be operational early 1972, will require a shorter time to move offsite than is usually necessary because the ship is not conventionally anchored. This ship, with some modifications, would be able to drill in cold weather (Fig. 6). Semisubmersible platforms are not suitable in ice-infested waters, because they are not self-propelled and towing in heavy seas may be hazardous. In summary, it would appear that the best concept for drilling in iceberg waters would be a self-propelled semisubmersible (Fig. 7) with a dynamic-positioning capability.

Concept of Approach Zones of Icebergs

When an iceberg is determined to have entered the danger zone of a drillship (Fig. 8), a decision is made to secure the well and disconnect the riser. When the berg reaches the avoidance limit, the ship must move. By decreasing the time necessary to move offsite, the area of the avoidance zone can be reduced in order to save drilling time. The avoidance distance, D_1, is calculated by multiplying the speed of icebergs by the time to move. The danger distance, D_3, is the speed of the icebergs multiplied by the time to secure the well and disconnect the riser plus the avoidance distance. Thus it is necessary to know at any time the speed of the iceberg and the time needed to secure the well.

Detection and Travel Path of Icebergs

The first problem is to detect all the icebergs reaching the danger zone and in the majority of cases a detection range of 6 miles in all directions should be sufficient. Radar is not a completely adequate device for iceberg detection, because iceberg ice has a poor reflector coefficient and waves over 4 feet might obscure a dangerous growler. A secondary form of detection is the use of a small ship equipped with radar and sonar patrolling around the

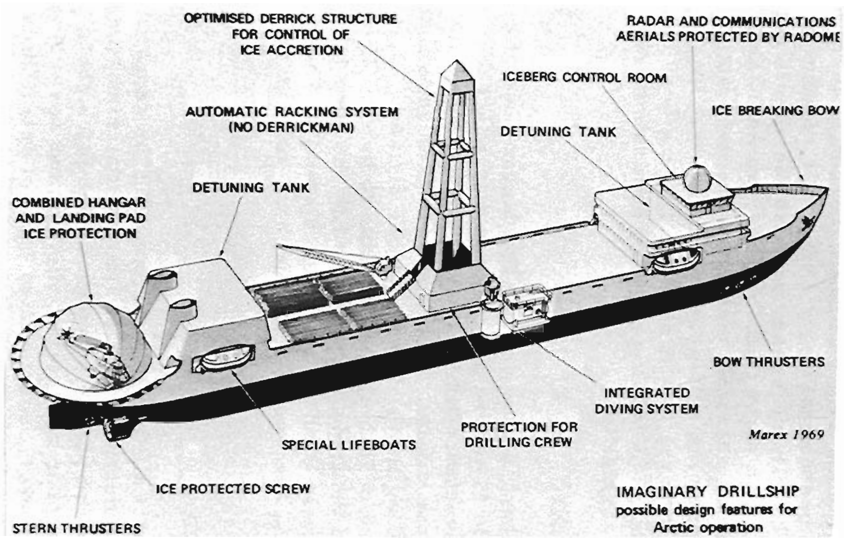


Figure 6 (above)

Proposed drillship – desirable design features for Arctic operation.

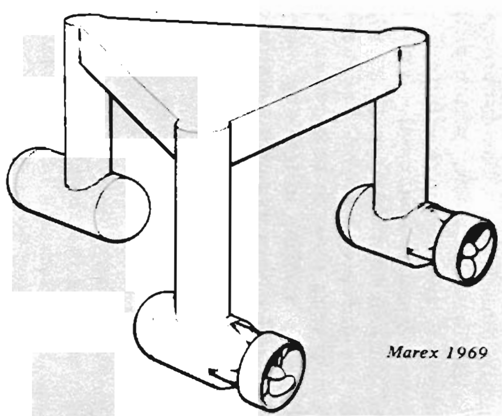


Figure 7 (right)

Self-propelled semisubmersible drilling platform.

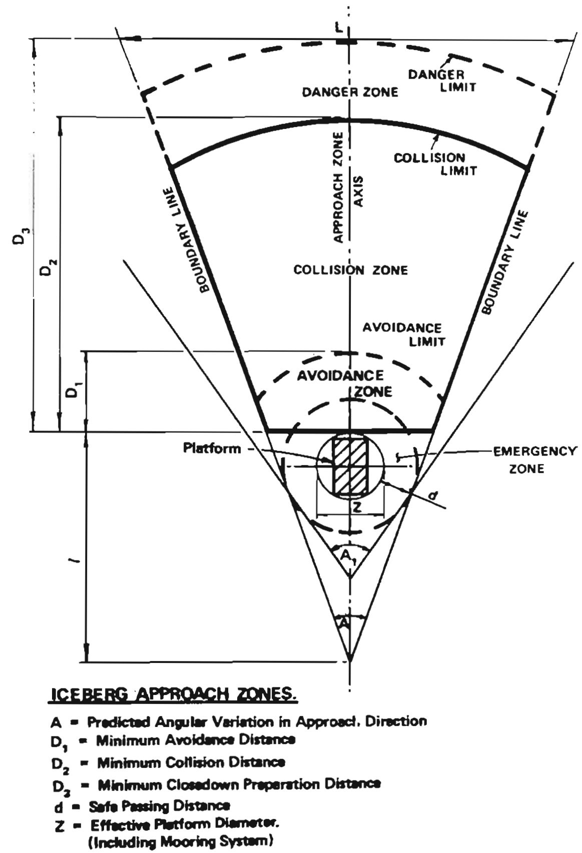


Figure 8. Iceberg approach zones.

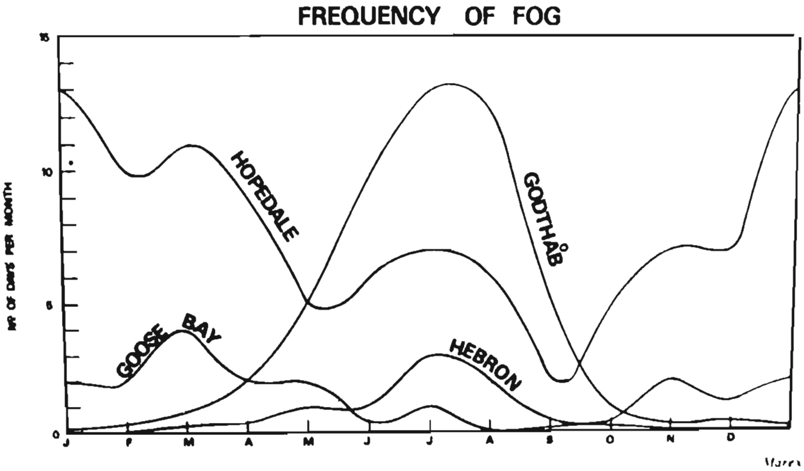


Figure 9. Frequency of fog at Goose Bay, Hopedale and Hebron, Labrador and Godthab, west Greenland.

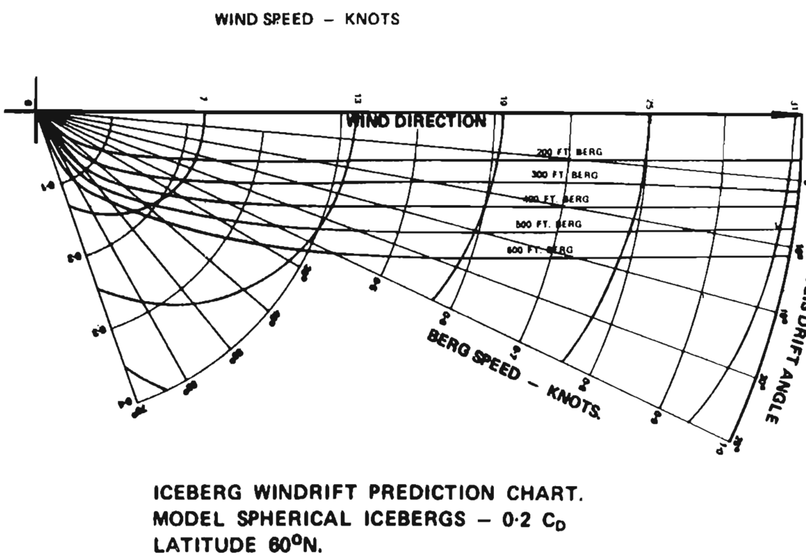


Figure 10. Iceberg windrift prediction chart.

drilling site particularly in the direction of the source of the icebergs. This ship could mark the icebergs with radar reflectors to permit their tracking from the drilling ship. If the danger distance is more than 10 miles, the use of a helicopter patrol is suggested. Efficiency of the helicopter and the patrol ship will be greatly reduced when fog or stormy weather occurs (Fig. 9).

When an iceberg is detected, one of the essential requirements it to be able to accurately forecast the short-term drift in order to estimate the danger corridor. Routes of icebergs are generally erratic, but a tentative method has been developed to calculate the drift. It requires information on speed and direction of wind, currents, waves, submarine topography, shape and size of the iceberg and observed drift behaviour. The different factors have been examined and their effects have been estimated (Fig. 10). A time factor has been considered for introducing the parameters for the last 24 hours and their forecast for the next 24 hours. This method should be tested in order to know the range of error of the calculated route and to define a corridor of danger. If the tests prove that the method is of value, a computer program can be prepared to estimate the corridor. This program will allow a definition of a new corridor of danger every half hour and permit taking a more accurate decision as to whether the well is secured or the operations continued.

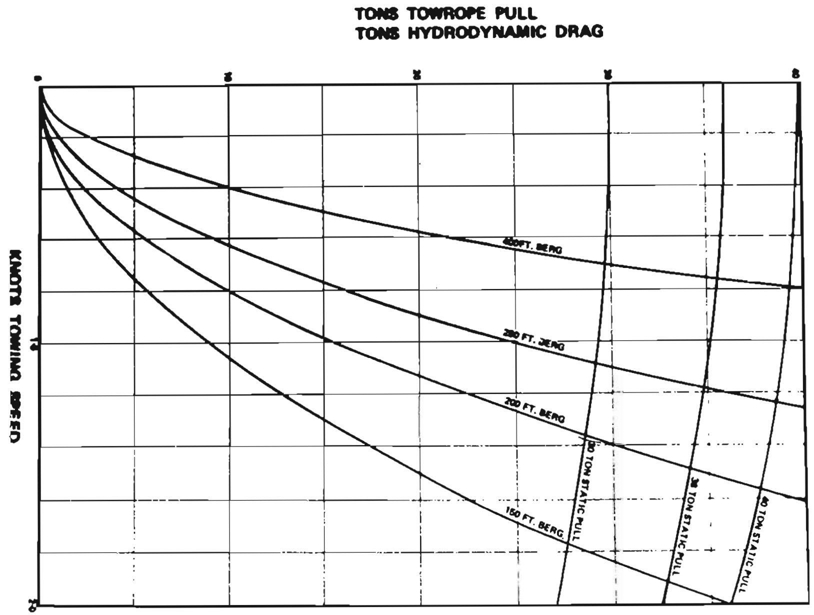
Icebergs will ground when their draft is greater than the water depth. Grounded icebergs can form trenches and scours in the sea bottom and could therefore cause damage to submarine equipment especially to the submarine wellhead. In most of the offshore areas of Labrador and Southwest Greenland, it appears that no protection for the wellhead in exploration drilling is necessary because the risk of an impact is very low. Should an iceberg ground inside the danger zone it would be inadvisable to continue to drill as the iceberg could overturn and sail again at any time. In such a case, it would be necessary to clear the bank of the berg by possibly towing, or using explosives to calve some pieces of ice and permit the iceberg to refloat.

Feasibility of Towing Icebergs

If the probability of occurrence of an iceberg is low, it would be possible to drill with no protection other than leaving the site when necessary. If the probability is high, it would be impossible to drill due to continuous warning. Due to the tremendous mass of icebergs, protective structures are not suitable for resisting the force of their impact, and so the possibility of towing bergs was investigated by TOTAL. Towing speed versus water-line length of icebergs has been estimated (Fig. 11). For safety considerations the tug will not be able to approach close to the iceberg. Different towing systems (Fig. 12) have been investigated but a surface cable seems the more suitable. The effect of a 30-ton force will generally not affect the stability of an iceberg but we should be prepared for this eventually. A successful towing trial was effected during a Greenland iceberg reconnaissance by TOTAL in April 1970 (Fig. 13). An iceberg, 80 metres long, 15 metres high, and having an estimated mass of 100,000 tons, was towed at a speed of 0.06 metres per second using a force of 1.5 tons. The towing of icebergs is feasible, but it should be conducted outside the danger zone in order to permit movement of the drillship offsite should there be a failure due to heavy sea, breakage of equipment or overturning of the iceberg.

Development of an Oilfield

If there is a discovery in any of the areas under discussion a fixed platform is eliminated because it is impossible to move an iceberg in the pack ice. Since the sea ice-free season lasts more than 5 months per year, there is enough time to drill the development wells during the ice-free season. However, a submarine completion will be necessary. Currently



Plot of DRAG of representative Spherical Model icebergs of Medium size, against TOWROPE PULL of representative general purpose tugs.

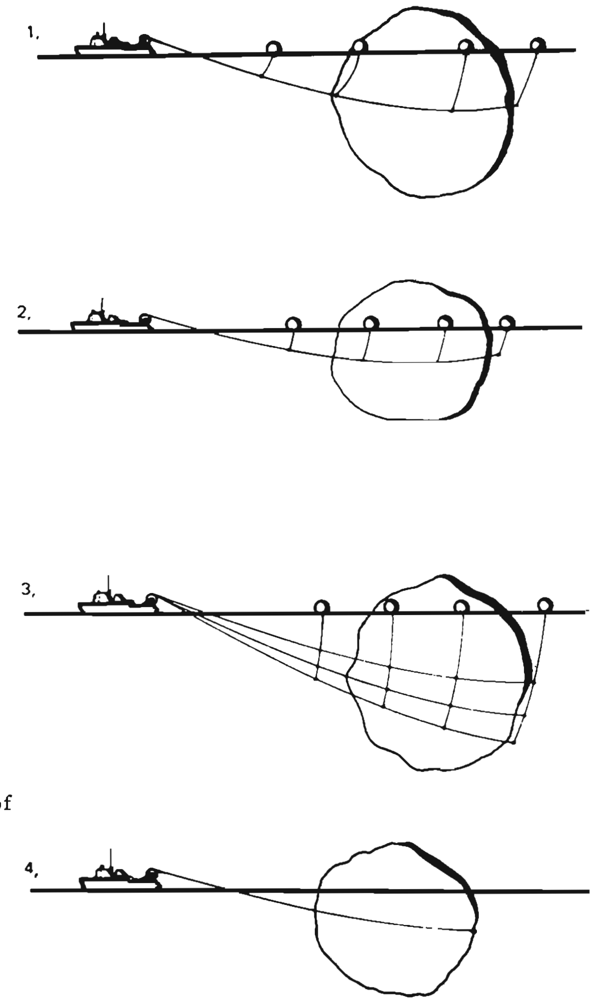


Figure 11 (above). Plot of drag of representative spherical model icebergs of medium size against towrope pull of representative general purpose tugs.

Figure 12 (right). Towing techniques for icebergs.

Figure 13. Towing an iceberg off Greenland.

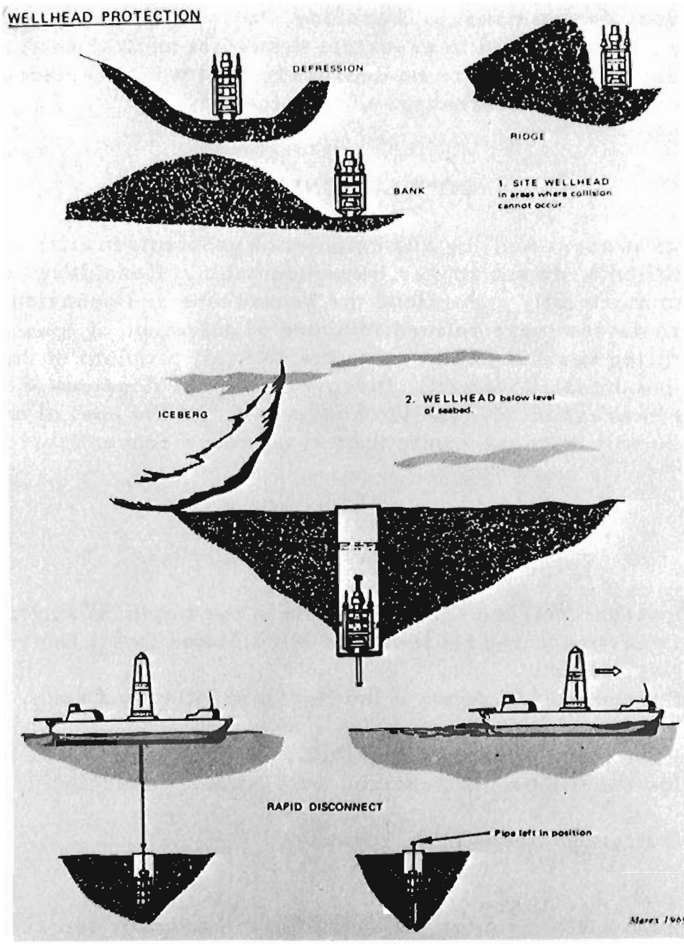


Figure 14. Wellhead protection techniques.

some companies are designing submarine completions which might be operational in two or three years. The biggest problem is to protect the bottom equipment against the effects of the grounded icebergs. The survival time of a submarine Christmas tree has been estimated in offshore Disko as being about two years, and the survival time varies from 4 years at 64°N to 200 years at 67°N. As the normal life of an oilfield is about 20 years, it will be generally necessary to protect the bottom equipment. Due to the tremendous forces which icebergs exert on impact, protective structures do not appear feasible. One solution is to bury the Christmas tree in order that its top is below the estimated scouring depth (Fig. 14). Another possible solution would be to break all icebergs which could ground on the field and thus allow them to float over the area.

The flowlines should also be protected. For offshore Cape Farewell, on an average, two submarine cables break every year because of the grounding of bergs (pers. comm. of Christoffersen, Det Store Nordiske Telegraf Selskab, Copenhagen) and these cables are laid along a sinuous route which follows the natural depressions. Therefore, to lay pipelines, a detailed bathymetry map should be used to ascertain the safest natural route. However, if in certain areas there are no depressions, it would be necessary to bury lines or take the risk of breakage of the pipe.

CONCLUSIONS

Icebergs present drilling and completion problems in offshore waters, which, though difficult, do not appear insurmountable. Considerably more data is needed to more fully understand the occurrence and behaviour of icebergs in order to devise more refined methods of detection of icebergs, the protection of drilling vessels and to solve the difficult problem of the completion of wells in ice-infested waters. Discovery and development of hydrocarbons is feasible in areas of adverse ice conditions, but the cost of exploitation of these reserves will be more costly than in the more conventional offshore areas.

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