Meddelelser om Grønland

Nares Strait and the drift of Greenland: a conflict in plate tectonics

edited by

P. R. Dawes & J. W. Kerr



Geoscience
8 · 1982

Nares Strait and the drift of Greenland: a conflict in plate tectonics

edited by

P. R. Dawes & J. W. Kerr



MEDDELELSER OM GRØNLAND, GEOSCIENCE 8 · 1982

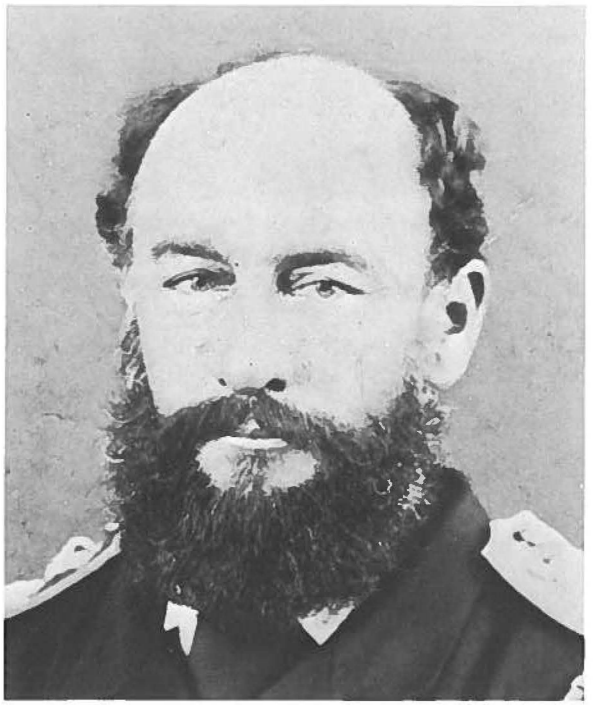
Frontispiece 1

The Kennedy Channel, the central part of Nares Strait. View is approximately north from Washington Land, Greenland, towards Judge Daly Promontory, Ellesmere Island, Canada. Hans Ø, at the territorial boundary between Canada and Greenland, is to the right of the large island (Franklin Ø). Greenland: Silurian limestones, carbonate buildups and shales of the Arctic Platform; Ellesmere Island: folded and faulted strata of the Ellesmere–Greenland fold belt. Photograph 545 H–N, no. 11803, July 1953; copyright Geodætisk Institut, Denmark.

Frontispiece 2

Personalities of Nares Strait. Admiral Sir George Strong Nares commanded the first successful return navigation of the narrow Strait in 1875–6 in H. M. Ships *Alert* and *Discovery;* Captain Henry Wemyss Feilden was the first to compile a geological map of the Strait from own observations (published in 1878); Frank Bursley Taylor initially drew attention to Nares Strait as an important strike-slip lineament separating the continent of North America from Greenland; and Alfred Lothar Wegener, who through his widely publicised writings on continental drift, focused attention on Nares Strait as a horizontal large-scale dislocation along which Greenland had drifted past Ellesmere Island, such that in geological discussion, Nares Strait has become synonomous with the 'Wegener Fault'.

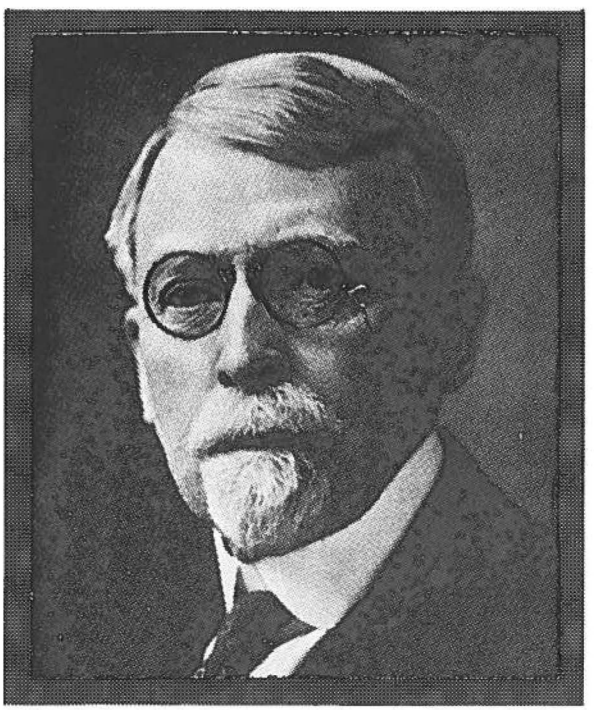
Copenhagen 1982-10-06
ISSN 0106-1046
ISBN 87-17-02969-4
Printed in Denmark by
AiO Print as, Odense, Denmark



Sir George S. Nares
1831–1915



Henry W. Feilden
1838–1921



Frank B. Taylor
1860–1937



Alfred L. Wegener
1880–1930

List of Contributors

R. L. Christie
Geological Survey of Canada

J. D. Collinson
University of Keele, U.K.

P. R. Dawes
Geological Survey of Greenland, Denmark

C. D. S. de Vries
Deceased

K. Ellitsgaard-Rasmussen
Geological Survey of Greenland, Denmark

J. England
University of Alberta, Canada

R. K. H. Falconer
Geological Survey of Canada

D. A. Forsyth
Earth Physics Branch, Canada

T. Frisch
Geological Survey of Canada

J. G. Fyles
Geological Survey of Canada

A. C. Grant
Geological Survey of Canada

A. K. Higgins
Geological Survey of Greenland, Denmark

J. M. Hurst
Geological Survey of Greenland, Denmark

H. F. Jepsen
Geological Survey of Greenland, Denmark

G. L. Johnson
Office of Naval Research, U.S.A.

C. E. Keen
Geological Survey of Canada

J. W. Kerr
Geological Survey of Canada

L. C. Kovacs
Naval Research Laboratory, U.S.A.

R. A. Langel
National Aeronautics and Space Administration, U.S.A.

N. J. McMillan
Canterra Energy Ltd., Canada

U. Mayr
Geological Survey of Canada

A. W. Menzies
Esso Resources Canada Ltd.

D. Monahan
Canadian Hydrographic Service

P. H. Newman
Dalhousie University, Canada

J. S. Peel
Geological Survey of Greenland, Denmark

J. W. Peirce
Petro-Canada Exploration Inc.

L. W. Sobczak
Earth Physics Branch, Canada

N. J. Soper
University of Sheffield, U.K.

S. P. Srivastava
Geological Survey of Canada

F. Surlyk
Geological Survey of Greenland, Denmark

L. Thorning
Geological Survey of Greenland, Denmark

R. J. Wetmiller
Earth Physics Branch, Canada

J. T. Wilson
Ontario Science Centre, Canada

Contents

Foreword 11
J. Tuzo Wilson

Preface 13
K. Ellitsgaard-Rasmussen and J. G. Fyles

Introduction 15
P. R. Dawes and J. W. Kerr

HISTORY

History of exploration and geology in the Nares Strait region 19
P. R. Dawes and R. L. Christie

History and implications of the Nares Strait conflict 37
J. W. Kerr

PHYSIOGRAPHY

Physiography of Nares Strait: importance to the origin of the Wegener Fault 53
D. Monahan and G. L. Johnson

Postglacial emergence along northern Nares Strait 65
J. England

STRATIGRAPHY

The Precambrian Shield of northernmost Baffin Bay: correlation across Nares Strait 79
T. Frisch and P. R. Dawes

The Proterozoic Thule Basin of Greenland and Ellesmere Island: importance to the Nares Strait debate 89
P. R. Dawes, T. Frisch and R. L. Christie

Proterozoic – basal Cambrian stratigraphy across Nares Strait: correlation between Inglefield Land and Bache Peninsula 105
J. S. Peel, P. R. Dawes, J. D. Collinson and R. L. Christie

Cambrian–Ordovician platform stratigraphy: correlations around Kane Basin 117
J. S. Peel and R. L. Christie

Upper Ordovician to Silurian facies patterns in eastern Ellesmere Island and western North Greenland and their bearing on the Nares Strait lineament 137
J. M. Hurst and J. W. Kerr

Nares Strait and the down-current termination of the Silurian turbidite basin of North Greenland 147
F. Surlyk

The Bjørnehiet Formation: a faulted preglacial conglomerate, Washington Land, North Greenland 151
H. F. Jepsen

STRUCTURE

Fold belts and metamorphic zones of northern Ellesmere Island and North Greenland 159
A. K. Higgins, U. Mayr and N. J. Soper

Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago 167
U. Mayr and C. D. S. de Vries

The Nyeboe Land fault zone: a major dislocation on the Greenland coast along northern Nares Strait 177
P. R. Dawes

Brecciated lineaments in Washington Land, North Greenland, and their relation to Nares Strait 193
J. M. Hurst and J. W. Kerr

A geological case for movement between Canada and Greenland along Nares Strait 199
P. H. Newman

Cretaceous–Tertiary magmatic and tectonic events in North Greenland and the history of adjacent ocean basins 205
N. J. Soper, P. R. Dawes and A. K. Higgins

Fragmentation of the Canadian Arctic Archipelago, Greenland and surrounding oceans 221
L. W. Sobczak

The evolution of the Nares Strait lineament and its relation to the Eurekan orogeny 237
J. W. Peirce

GEOPHYSICS

Marine geophysical study of southern Nares Strait 255
P. H. Newman

Review of seismicity and other geophysical data near Nares Strait 261
R. J. Wetmiller and D. A. Forsyth

Motion along Nares Strait recorded in the Lincoln Sea: aeromagnetic evidence 275
L. C. Kovacs

Satellite magnetic field over the Nares Strait region 291
R. A. Langel and L. Thorning

Crustal history and basin development of Baffin Bay 295
A. W. Menzies

Problems with plate tectonic models for Baffin Bay – Nares Strait: evidence from the Labrador Sea 313
A. C. Grant

The geophysical implications of minimal Tertiary motion along Nares Strait 327
C. E. Keen and J. W. Peirce

Nares Strait: a conflict between plate tectonic predictions and geological interpretation 339
S. P. Srivastava and R. K. H. Falconer

ECONOMIC GEOLOGY

Nares Strait and the petroleum explorer 355
N. J. McMillan

SUMMARY

The case for major displacement along Nares Strait 365
G. L. Johnson and S. P. Srivastava

The case against major displacement along Nares Strait 369
P. R. Dawes and J. W. Kerr

APPENDIX

Note on the magnetic polarity time scales used in the Nares Strait volume 389
P. R. Dawes

Foreword

The editors of this excellent symposium have done me the honour of inviting me to write an introduction to the proceedings. Although flattered I am surprised, for I have not worked nearer to the area than the outcrops of Tertiary lavas on Baffin Island and I have long been engaged in administrative work.

I shall therefore leave to others discussion of the detailed evidence around Nares Strait and draw attention to two broad aspects which must influence thinking on the subject.

One is the revolution in the earth sciences which culminated less than 15 years ago. Plate tectonics is the theory which resulted, but the cause of the revolution was the success of new methods in interpreting the ocean floors, the Earth's interior, palaeomagnetism and radiometric ages. These observations are quite secure and demand explanation as much as do observations by traditional geological methods.

Fifty-five years ago, when Noel Odell first introduced me to geology, little was known of geophysics or geochemistry. As one distinguished geologist stated geological mapping was both the method and the sole aim of the earth sciences. Theories were scoffed at as 'armchair geology'. Those acquainted with other sciences were puzzled because these had excellent theories.

One explanation of this lack in geology which was never mentioned or even considered was that Wegener might be right. If so it was indeed futile to try to use a fixed, unchanging model to theorize about a mobile Earth.

Nevertheless the revolution did come and like other revolutions was not introduced nor welcomed by the establishment. The geological surveys, university departments of geology, the mining industry and the petroleum companies all were opposed to change. The new methods were introduced by physicists with no formal training in geology who included Wegener, Vening Meinesz, Rothe, Gutenberg, Aston, Nier, Ewing, Bullard, Runcorn, Mackenzie and many oceanographers, while the new interpretations were for the most part suggested by geologists who had more physical training or insight than most, men such as Daly, Holmes, Carey, Irving, Hess, Vine, Morgan and Le Pichon or those few like Argand and du Toit whose field work had shown them particularly clear illustrations of continental motions.

The new ideas were not popular. All of us who had to change our minds realize how difficult it was and many older geologists still wish for a miracle that would disprove the new ideas.

I clashed with some of them early when mapping near Great Slave Lake where, following C. H. Stockwell and A. W. Jolliffe, I found evidence of great faults and superb foliation. Both suggestions were unpopular with geologists who believed that contraction was the sole source of motion in the Earth and that all the basement rocks had crystallized from molten magmas. Later I was involved in debates with those who opposed large motions on the Great Glen, San Andreas, Alpine and Jordan Valley faults. That these motions are now accepted does nothing to prove that there has also been extensive motion along Nares Strait, but shows that there has been a history of disbelief in large fault displacements.

This revolution has gone so far that two sea-going physicists recently published a long paper giving a precise account of the opening of the whole South Atlantic Ocean with scarcely a reference to the coastal geology and none to that of the islands along the mid-ocean ridge.

This brings me to the second broad concept which I wish to discuss. Some authors take the parochial view that the amount of displacement along Nares Strait provides a test of the theory of plate tectonics. I doubt this.

Everyone seems to admit that field geology, air photographs and an argument concerning the origin of linear features all suggest at least a small displacement.

Those who do not accept a large motion suggest that the geology on the two sides matches too closely to admit of it, but if, instead of sea, Nares Strait were land, how many field geologists would accept such a correlation in mountainous and Precambrian terrain without mapping the gap of 24 kilometres between the sides? Unfortunately, there is as yet a shortage of clear geophysical data from the sea floors in the Strait and in all the adjacent region.

In any case alternatives are not confined to a simple choice between great displacement along both Nares Strait and Baffin Bay or slight motions in both cases. Account must be taken of the origin of Lancaster and Jones Sounds. K. Burke has suggested that they may be spreading aulacogens and motions in other places could be hidden by ice or in complex geology.

The whole Atlantic Ocean appears to be an entity whose different parts spread by similar but not contemporaneous movements.

First about 190 million years ago following uplift, rifting and intrusion between the Mediterranean and Caribbean areas, North America separated from Africa. The Azores and a ridge to Gibraltar still indicate one boundary.

About 130 million years ago spreading began be-

tween South America and Africa. Principal areas of outpouring left lavas of that age on opposite coasts and formed such trails as the Cape, Walvis, Rio Grande, Guinea and Sierra Leone rises. The hot spots which formed these are still evident as the volcanic or uplift islands named Bouvet, Tristan da Cunha, Gough, Ascension and St. Paul's Rocks.

The separation of North America, Greenland and Europe involved two spreading ridges. One extends from a triple junction south of Greenland through the active hot spots of Iceland and the island of Jan Mayen, to the De Geer Line which cuts across from south of Spitsbergen along the northeast coast of Greenland, to the Yermak hot spot north of Greenland. The other ridge extended from the same triple junction to the Yermak hot spot through an extinct hot spot presumed to have lain under Davis Strait. Several of these hot spots left trails of which the earliest evidence is the outcrops of early Tertiary lavas on Cape Dyer, Baffin Island, on the east and west coasts of Greenland and on the Faeroe Islands.

The concept that the Atlantic Ocean, Norwegian Sea and Baffin Bay all opened along lines joining hot spots is derived from H. Cloos' suggestion for the origin of the Rhine Graben, the Red Sea and the Gulf of Aden.

According to this view, which many geophysical data and Tertiary geology described in papers in this symposium supplements, there is much that is interesting but nothing unusual about the geology of Nares Strait. The only problem is to obtain sufficient data to correctly apportion the spreading motion of Baffin Bay between Nares Strait, Lancaster and Jones Sounds and other possible places.

May I again congratulate the editors and authors of this symposium on clarifying how to tackle this problem.

J. Tuzo Wilson
Director General
Ontario Science Centre
770 Don Mills Road
Don Mills, Ontario
Canada M3C 1T3

Preface

This book outlines the reasons why Nares Strait is considered to be one of the most important structural elements in Arctic geology. The origin of the Strait is central to plate tectonic models concerned with the origin of the North Atlantic and Arctic Oceans. The basic conflict elucidated in this volume has had a long and complicated history and the ideas and theories presented by F. B. Taylor, A. Wegener, S. W. Carey and J. Tuzo Wilson involving Nares Strait have been foremost in influencing geological thought on global tectonics.

It is clear that understanding of a geological structure of the magnitude and importance of Nares Strait, forming as it does a national boundary between Canada and Greenland, requires close international cooperation. This was achieved by the Nares Strait symposium held at Halifax in May 1980, of which this book is the outcome. All those who have debated the nature of Nares Strait through the years have been united in their wish for more geological information from this isolated and poorly known region. In recent years, however, the problem of remoteness has been overcome by increased exploration with more efficient logistic support and by international cooperation. Thus, the present volume presents a wealth of new information from Greenland, Canada and Nares Strait itself, and these data contribute greatly to the primary objective of examining the history and origin of the Strait. We stress, however, that much remains to be done to achieve a full understanding and uniform map presentation of the onshore and offshore geology of the Nares Strait region. One of the most serious deficits, the lack of offshore geophysical work in the central and northern parts of Nares Strait, is an important target for future activity.

It is not necessary to emphasise the significance of Nares Strait and plate tectonic models based upon it in terms of hydrocarbon exploration, at a time when petroleum exploration is advancing further north, both onshore and on the continental shelves of the Arctic and North Atlantic Oceans. In particular, solving the Nares Strait controversy has a crucial bearing on our understanding of the origin and evolution of Baffin Bay and the Labrador Sea. These two seas contain large sedimentary basins in Canadian and Danish-Greenlandic territory, where petroleum assessment and exploration is currently being actively undertaken.

We are pleased to see the part played in this venture by geological and geophysical operations initiated by our respective organisations, although we also note the major contributions by specialists from other institutions who share a common interest in the Arctic. The many papers jointly authored by scientists based in Canada and Denmark are a most welcome indication of increased international cooperation across our mutual frontier.

K. Ellitsgaard-Rasmussen
Director
Geological Survey of Greenland
Øster Voldgade 10
DK–1350 Copenhagen K
Denmark

J. G. Fyles
Chief Geologist
Acting Director-General
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario
Canada K1A OE8

Introduction

This volume has arisen out of the symposium "Did Greenland drift along Nares Strait?", which was a special session of "Halifax '80" — the annual meeting of the Geological Association of Canada and the Mineralogical Association of Canada, held at Dalhousie University, Halifax, Nova Scotia, May 19–21, 1980. The symposium was organised jointly by the Geological Survey of Canada (GSC) and the Geological Survey of Greenland (GGU) and co-chaired by the undersigned and R. K. H. Falconer.

Planning for the Nares Strait symposium began in early 1979 and by summer circulars inviting contributions had been sent to persons and institutions known to have a direct geological interest in the Nares Strait region. The main aim of the symposium was to resolve the question "Did Greenland drift along Nares Strait?". Papers from all branches of geology and geophysics which had a bearing on the structure and history of Nares Strait were solicited. We particularly encouraged scientists based in North America and in Denmark to collaborate in writing papers that dealt with both sides of the Strait. We were confident that the meeting would provide the most up-to-date information from the region, hopefully inspire new discussion and understanding, and reveal where more work needs to be carried out. It was hoped that the symposium might reach a meaningful consensus on the question of displacement in Nares Strait.

The symposium programme consisted of 18 papers which were delivered as lectures, and a concluding address by 'referee' M. J. Keen of the Atlantic Geoscience Centre. Those 18 papers are presented in this volume, with some titles and authorships differing from the abstracts published in the "Halifax '80" programme. During the planning of the symposium it became clear that in addition to the papers which could be presented at Halifax, many other relevant data were available and as a result this volume contains 30 papers.

The papers are grouped into seven sections, although there is naturally subject overlap between the sections.

1) *History.* The two papers of this section set the scene for the discussion of the main topic of the Nares Strait symposium; a history of exploration and geology leads to the specific question of the Nares Strait conflict and its regional implications.
2) *Physiography.* Two papers evaluate the Holocene physiographical aspects of the Nares Strait region; present-day physiography and postglacial emergence are viewed in terms of fault activity along Nares Strait.
3) *Stratigraphy.* This section of seven papers describes the main elements of the onshore Precambrian to Cenozoic stratigraphy on both sides of Nares Strait. The papers are arranged chronologically and they examine the stratigraphy in terms of the relative positions of Ellesmere Island and Greenland.
4) *Structure.* The eight papers of this section describe the structural, and some stratigraphic, features of the Nares Strait region. The first four papers discuss the onland Palaeozoic and Tertiary structural pattern; the remaining papers formulate a more regional tectonic appraisal of Nares Strait and its role in the North Atlantic.
5) *Geophysics.* The eight papers grouped here discuss seismic, gravity and magnetic data from both the marine and onshore areas of Nares Strait and the adjacent seas. The majority of papers assess the information in terms of plate tectonic models for the region.
6) *Economic geology.* This paper discusses some implications of the Nares Strait debate on petroleum exploration of the region.
7) *Summary.* Two sequel papers that summarise the arguments presented for and against Nares Strait being the site of major strike-slip displacement between Greenland and North America.

Many papers in the volume treat Nares Strait in the context of sea-floor spreading and plate tectonics. In some of these a magnetic polarity time scale is used as a framework with which to describe the timing of geological events. An attempt to adopt but one of the many geomagnetic time scales as a standard for the volume was deemed 'impracticable'. Instead, to effect an easier correlation of the time frameworks presented in the different contributions, a comparison of the various geomagnetic time scales is given in an appendix.

Editorial policy of Meddelelser om Grønland has been followed in this volume, and all contributions have been critically reviewed by referees. However, since the volume is essentially the proceedings of a symposium, the manuscripts were accepted as resubmitted by the authors, without additional comments on the scientific content. Thus, the authors alone are responsible for this aspect of their papers. Presentation in Meddelelser om Grønland style led to standardisation of certain spellings, as well as changes to authorised place names. However, because authors' intentions when describing geological time or time units were not always apparent, no attempt was made to standardise stratigraphical terminology, such as "early", "Late", "Upper" and

"lower". References have been consistently 'spot-checked' and inspected for general clarity but many have not been scrutinised for accuracy — this remains the responsibility of the authors.

Acknowledgements

We are pleased to acknowledge the contribution to the symposium and the volume made by GGU and GSC and by members of their scientific and technical staff. We thank R. K. H. Falconer and R. T. Haworth of the Atlantic Geoscience Centre, Dartmouth, for the parts they played in organising and arranging the symposium as a special session of the "Halifax '80". A. Sutherland-Brown, President of the Geological Association of Canada kindly encouraged us to include the symposium as part of the annual meeting of the Association. We are also very obliged to 25 persons from both sides of the Atlantic who agreed to act as scientific referees to the volume. R. F. B. de Caen of Union Oil Company of Canada Ltd. compiled and drafted a base map of the Nares Strait region for which we are grateful.

All the 'technical' editing of the material has been carried out at GGU in Copenhagen. We owe special thanks to Esben Glendal for his scrupulous preparation of the manuscripts and his many suggestions for their improvement. His contribution also includes typing and retyping of more than half of the papers, and assistance in proof checking. Jakob Lautrup and assistants carried out much of the photographic work including reduction of figures to format size, while Gurli E. Hansen, Grethe Haugaard, Ulla Lundblad, Annette T. Rasmussen and Bodil Sikker-Hansen are thanked for drafting work. The aerial photographs which appear as figures are the copyright of the Geodetic Institute, Denmark, and the National Air Photo Library, Canada. We are grateful to the following institutions for providing the photographs included in the frontispiece: The Allen County Public Library, Fort Wayne, Indiana, U.S.A. (F. B. Taylor); H. Hirschsprungs Forlag, Copenhagen, Denmark (A. L. Wegener); The Museum of the King's Own Royal (Lancaster) Regiment, England (H. W. Feilden) and The Royal Geographical Society, London (G. S. Nares).

In conclusion we wish to express our thanks to the Commission for Scientific Research in Greenland for sponsorship of this volume and for financial assistance for travel in the final editorial stages, as well as to T. C. R. Pulvertaft, scientific editor of Meddelelser om Grønland, Geoscience Series, and Gregers E. Andersen and Dagny Rosing for their cooperation.

P. R. Dawes
J. W. Kerr

History

History of exploration and geology in the Nares Strait region

PETER R. DAWES and ROBERT L. CHRISTIE

Dawes, P. R. & Christie, R. L. 1982. History of exploration and geology in the Nares Strait region. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 19–36.

Exploration in the Nares Strait region in the late 19th and early 20th centuries was connected with the seaway's position as a principal route of geographic discovery. Few of the early expeditions were directed towards obtaining data for the young science of geology. At the turn of the century, with the passing of the main era of geographical discovery, including the race for the North Pole and the establishment of Greenland's insularity, geological understanding of the region advanced rapidly and geologists more or less became 'standard' members of expeditions to this part of the Arctic.

Systematic geological studies in the Nares Strait region began when Lauge Koch mapped the Greenland side of the Strait in the period 1916–23; such investigations on the Canadian side by the Geological Survey of Canada took place in the 1950s and later. Early private expeditions and later work by university and petroleum and mineral enterprises have also contributed many geological data, as have several 'military operations' centred on Thule Air Base in Greenland. Regional geological mapping in Greenland was renewed by the Geological Survey of Greenland in the 1970s and continues today.

Cooperative Danish–Canadian projects, initiated by "Operation Grant Land" in 1965–66 in northern Nares Strait, have aimed at coordinating field studies in order to better assess correlation of stratigraphy and structure across the Strait.

Greenland and Nares Strait held important positions in the early ideas about the horizontal mass movements of the continents, and both Frank B. Taylor and Alfred Wegener featured the narrow linear channel between Ellesmere Island and Greenland in their respective theories of continental drift. These two creative theorists built on the immense global knowledge assembled by Eduard Suess, who published one of the earliest appraisals of the region. Early geological maps of the Nares Strait region by Bailey Willis and Lauge Koch were used by Wegener in support of his theory of continental drift.

P. R. Dawes, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK–1350 Copenhagen K, Denmark and R. L. Christie, Department of Energy, Mines and Resources, Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303–33rd St. N.W., Calgary, Alberta, Canada, T2L 2A7.

The southern extremity of Nares Strait was discovered in 1616 by William Baffin and Robert Bylot, sailing in the *Discovery*. Guided by Baffin, the navigator, they negotiated the ice-filled seas of Melville Bugt and named a cape — "Sir Dudley Digges' Cape" (Fig. 1). They reached a wide sound with an island at its entrance which were duly named "Sir John Wolstenholme's Sound and Ile". Further north they entered an even wider inlet and named it "Whale Sound" after the great number of whales they found there. The ship's northerly progress was halted by ice at a small island — "Hakluit's Ile". Turning west, the explorers saw to the north another great entrance, "Sir Thomas Smith's Sound". This sound, they remarked "runneth to the North of seventy eight degrees ..." (S. Purchas in Taylor 1955: 16). Baffin and Bylot had sailed past the entrance of a linear series of basins and channels that is now named Nares Strait and which "runs to the north" beyond latitude 82°N, or some 500 km.

After discovering the "Careyes Islands" in the middle of the seaway, Baffin sailed southward and found the entrance of what he called "Alderman Jones Sound". A brief landing in this sound is the first recorded European landing in the vicinity of Nares Strait; which of the islands Ellesmere, Coburg or Devon was touched on, however, cannot be told from Baffin's description.

Nares Strait is named after Admiral Sir George S. Nares of the Royal Navy, who commanded the first successful navigation of the Strait to and from the shores of the Polar Sea in 1875–6 (Fig. 2, Nares 1878). Up to about the turn of the 19th century, expeditions in the Nares Strait region were primarily aimed at geographical discovery and were often fired by national ambition, one result of which was the "race to the North



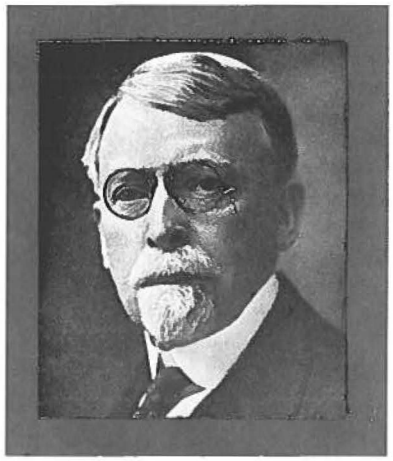
Fig. 1. Index map of the Nares Strait region.



Sir George S. Nares
1831–1915



Henry W. Feilden
1838–1921



Frank B. Taylor
1860–1937



Per Schei
1875–1905



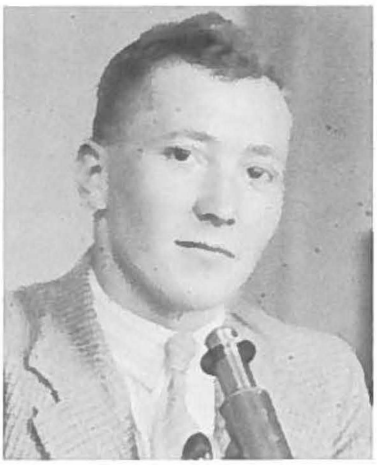
Alfred L. Wegener
1880–1930



W. Elmer Ebklaw
1882–1949



Lauge Koch
1892–1964



Robert Bentham
1913–1968



Johannes C. Troelsen
1913–

Fig. 2. Personalities in the history of exploration and geology of the Nares Strait region. For sources see acknowledgements.

Pole". By 1900, when R. E. Peary reached Kap Morris Jesup and proved the insularity of Greenland, the position of Nares Strait as a linear channel linking "Baffin's Bay" to the "Open Polar Sea" was established.

Geological knowledge of the Nares Strait region has been accumulating since early in the 19th century. Although geologist-naturalists had visited the Nares Strait region as members of the early exploration expeditions, the type of geological data obtained was often cursory and few expeditions included teams specifically directed to geology. Systematic geological mapping in the Nares Strait region began in 1916 with the Danish 2nd Thule Expedition on which Lauge Koch acted as geologist and cartographer, but these investigations were restricted to the Greenland side of the Strait. Regional studies on the Canadian side began somewhat later, in the 1950s, through the efforts of the Geological Survey of Canada. The systematic collection of data has continued on both sides of the Strait, albeit at various paces, until the present day.

The work style and type of data obtained have, of course, changed considerably over the years. Thus, the early practise of wintering in the Arctic and relying mainly on spring and autumn sledging has been superseded by aircraft-supported operations allowing a wide diversity of geological studies to be carried out where and when demanded. The introduction of airborne operations in the late 1940s is an important landmark and is here taken as a convenient point of dividing the chronological description (see Christie & Kerr 1981).

Early exploration (1818–1946)

Two centuries passed after Baffin's and Bylot's exploration in 1616 before others sailed in northern Baffin Bay: John Ross in 1818, following Baffin's course; William E. Parry, in 1819, entering Lancaster Sound; and Parry, again entering the sound in 1825. Few landings were made and beyond John Ross' meeting with the "Arctic Highlanders" — the "Eskimaux" of the Thule region — geographical knowledge of Nares Strait was little advanced. Thus, while Ross had named the opposite capes of Smith Sound after the two ships of his expedition, HMS *Isabella* and *Alexander,* he regarded "Smiths Sound" as closed and indicated this on his map (Ross 1819).

The first step in reporting on the geology of the region had been made: the country forming northern Baffin Bay was described as being composed of "primary formation" (Precambrian) granite and gneiss with some greenstone. Samples were collected along the Greenland coast at "Bushnan's Isle" and "the coast between Cape York and Cape Dudley Digges" and these were described by Dr. J. M'Culloch (Ross 1819). On Parry's later voyage (1825), R. Jamieson reported and collected Precambrian rock at Cape Warrender on Devon Island.

Sir John Franklin of the Royal Navy began a search for the Northwest Passage in 1845. Franklin's entire expedition was lost, but this disaster led to the launching, over the next decades, of numerous search expeditions to the Arctic Islands. Two of these, one led by Commander Edward A. Inglefield in 1852 and a second by Dr. Elisha K. Kane, 1853–55, contributed to the geological knowledge of Nares Strait.

The earliest description of the geology of Nares Strait is that by Peter C. Sutherland, surgeon and naturalist in 1852 on the *Isabel,* commanded by Cdr. Inglefield. Inglefield achieved a new northing (78°28′21″), just entering Kane Basin before being driven southward by storm winds and wind-driven ice. The *Isabel* then



Fig. 3. The Crystal Palace Cliffs, Greenland, seen from the north. "At Cape Alexander, the eastern boundary of the entrance of Sir Thomas Smith's Sound, in latitude 78°15′N, we again find the strata a little curved; but only a few miles farther north they are so regular in parallelism and succession, as to vie with a distant view of the Crystal Palace ..." (Sutherland 1853b: 156). Precambrian crystalline basement overlain by sediments and basic sills of the Proterozoic Thule Group. The height of the cliffs is about 400 m above sea level.



Fig. 4. Kap Constitution, Greenland, viewed from the south — the northernmost point reached by the Kane expedition 1853–55. "The cliffs ... were perpendicular, and nearly two thousand feet high" (W. Morton in Kane 1856, vol. 2: 378). A Silurian carbonate mound about 500 m high. Photograph Lauge Koch (from Dawes & Haller 1979: fig. 9).

coasted south-eastern Ellesmere Island and entered Jones Sound, without landing. Sutherland's description of the "physical geography", presented to the Geological Society of London, is extremely illuminating, made so by his many sketches of the coastline and by his attention to structural detail, so that inclinations of strata for several localities were reported. Sutherland (1853a, b) recognised the gneiss and granite basement and the overlying homoclinal cover of unmetamorphosed strata with associated trap rocks (Thule Group, Fig. 3). He compared these formations with the Tertiary igneous and sedimentary rocks of the Disko region of West Greenland, with which he was familiar.

The American explorers, Kane (in 1853–55 as noted) and Isaac I. Hayes (in 1860–61) pushed into Smith Sound and wintered there, at the southern entrance of Nares Strait. Exploring parties sledged with great difficulty northward along the shores of the now named Kane Basin and viewed the immense reaches of Kennedy Channel. The Kane and Hayes expeditions, however, contributed mainly geographical rather than geological knowledge, although several references to the granitic and gneissic rocks, trap, basalt and greenstone, and the overlying sediments can be found in the expedition reports (Kane 1856, Hayes 1867).

Kane's expedition wintered in Rensselaer Bugt, Greenland and travelled widely in the Inglefield Land – Humboldt Gletscher region about which Kane remarks: "Its geological structure is of the older red sandstones and silurian limestones, overlying a primary basis of massive syenites. The sandstones to the south ... are in series, with intercalated greenstones ..." (Kane 1856, vol. 2: 308); a pertinent description of the general geology. Kane's expedition must be credited with the discovery of the homoclinal platform strata north of the Humboldt Gletscher (Fig. 8); these were reached by William Morton en route to a new farthest north just south of Kap Constitution (80°34'N), where the perpendicular cliffs were scaled with much difficulty (W. Morton in Kane 1856, vol. 2: 378) (Fig. 4).

Isaac Hayes, who had been surgeon on the Kane expedition, chose the Ellesmere Island shore of Kennedy Channel for his push northward. He briefly described the "geological features of coast" that included "cliffs of silurian rocks – sandstone and limestone" (Hayes 1867: 340). Fossils collected from the shore of Kane Basin were identified by Prof. F. B. Meek as Silurian in age and this represents the first palaeontological description from the Nares Strait region (Meek 1865).

Geological understanding of the southern part of Nares Strait had advanced but slowly during the mid-1800s. Thus, a map compiled by Samuel Haughton in 1858 showed only "granitic and granitoid rocks" between "Cape York" and "Granville Bay" in Greenland, on the "Carey Isles" and on eastern "North Devon" Island (Fig. 5). In the text, however, Haughton (1858, 1859) does mention the sandstone and conglomerate of the region "Wolstenholme and Whale Sounds" that overlie the crystalline rocks.

In 1871, Charles Francis Hall, an American, took the ship USS *Polaris* to 82°11'N, actually entering the long-sought-for Polar Sea at the northern entrance of Robeson Channel. The unfortunate Hall died during the winter (probably poisoned by one of his divided crew (Loomis 1971)) and the party retreated southward next spring. Some geological observations were made, particularly by Emil Bessels, surgeon and chief of the scientific staff, and these are included in his and the official account of the expedition (Davis 1876, Bessels 1879). Of particular interest is the recognition in the Hall Basin region of an extensive southern area of flat-lying, mainly limestone rocks that included fossiliferous Silurian strata and, to the north, a series of folded sandstone rocks. This is the first record of the southern

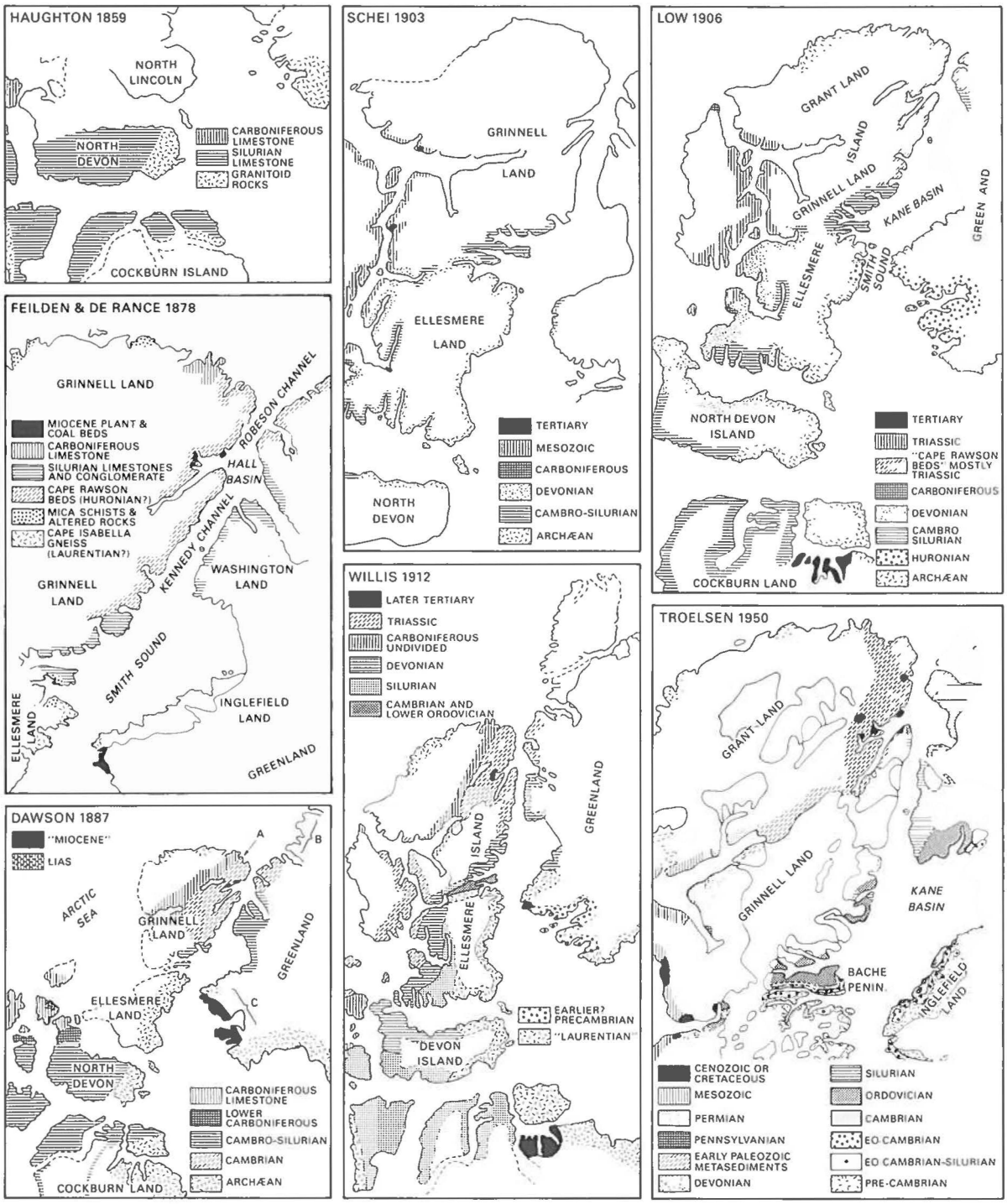


Fig. 5. Seven bedrock geological maps of the Nares Strait region from the early exploration period (1818–1946). Maps redrawn from the originals with some modifications. *Feilden & De Rance 1878,* Silurian symbol along the coast of southern Washington Land is included from the description in text (1878: 558); *Dawson 1887,* "Cambro-Silurian" of legend "probably including Devonian in Arctic Islands", A = "Devonian fossils", B = "coast probably Cambrian or Archæan", C = "coast described generally as consisting of trap rocks, sandstone etc. probably Tertiary. Archæan rocks are, however, also present"; *Schei 1903,* geology compiled from Schei (1903, 1904) and Holtedahl (1917), "Cambro-Silurian" of legend includes Schei's "Series B"; *Low 1906,* "Cambro-Silurian" of legend refers to "Silurian and Cambro-Silurian"; *Willis 1912,* "Laurentian" of legend includes "gneisses and associated plutonic rocks", "Earlier? Precambrian" refers to "unclassified" strata of "Huronian" or "Algonkian" age, "Later Tertiary" refers to "Miocene and Pliocene"; *Troelsen 1950,* legend has been abbreviated from the original map at scale 1:1 660 000 and the two-tone colouring scheme into "formations of known age" and "formations whose age is not known with certainty" has been omitted. "Pre-Cambrian" of legend refers to "metamorphic and igneous rocks", "Eo-Cambrian–Silurian" refers to "Eo-Cambrian, Cambrian, Ordovician and Silurian, undifferentiated", "Early Paleozoic metasediments" refers to "metamorphosed sediments, probably of Early Paleozoic age", "Mesozoic" refers to "Triassic and Lower Jurassic?".

margin of the Franklinian geosyncline — carbonate platform flanked on the north by the Ellesmere –Greenland fold belt.

In 1875, the British government sent a naval expedition northward under the command of Capt. George S. Nares. The two ships, HMS *Alert* and *Discovery,* followed the route pioneered by Hall and prepared for wintering at two localities on Ellesmere Island: the *Discovery* in Discovery Harbour of Lady Franklin Bay and the *Alert* at Cape Sheridan, latitude 80°28′N. Spring journeys with man-hauled sledges were made northward toward the Pole and westward and eastward along the north coasts of Ellesmere Island and Greenland. Considerable geographical and geological exploration and mapping was accomplished by these parties in spite of their short period in the field: Nares withdrew and returned to England in 1876 because of an outbreak of scurvy. The northern coasts, as well as the channels, basins and tributary fjords of Nares Strait had been charted by these skilled navigators and their maps formed the bases for atlases until well into the 20th century.

The senior naturalist of the expedition, Henry W. Feilden (Fig. 2), prepared a geological map and report (with C. E. De Rance) founded on geological data obtained through his own and others' field studies and rock collections. Feilden's map (Fig. 5) of the region between Discovery Harbour and Feilden Peninsula, Ellesmere Island, was the first on Arctic geology prepared by a geologist from his own field work and his report included important observations that had been made farther south along Nares Strait during the initial sail northwards, for example on Bache Peninsula and the shore of Kane Basin (Fig. 6). The fossils were described by Etheridge (1878). Feilden and his co-author commented on the similarity of the Carboniferous limestone of Feilden Peninsula to that of Svalbard and suggested that "a continuation of the direction of the known strike of the limestones of Feilden Peninsula, carried over the Polar area, passes through the neighbourhood of Spitzbergen ..." (Feilden & De Rance 1878: 560).

The first formal stratigraphic name given to rocks in the Nares Strait region was introduced by Feilden, viz. the Cape Rawson Beds, to describe a series of folded slates and sandstones; strata that were suggested to be "the equivalents in time" of the "Huronians of Nova Scotia". Of particular importance was the recognition of the Cape Rawson Beds on both sides of the Robeson Channel and of the unfolded Silurian limestone terrain to the south. In Greenland this limestone terrain was traced from the southern Kennedy Channel at 80°N to the Hall Basin at 81°30′N. Concerning the junction between the two terrains we find "the boundary, of whatever character it may be, traverses Kennedy Channel, and reappears in Hall Land, traversing the country from Polaris Bay to the southern end of Newman Bay, its situation being determined within narrow limits by the occurrence of Cape-Rawson beds at Thank-God Harbour to the north, and of Silurian limestones at Cape Tyson and Offley Island to the south, whence the limestones extend to both sides of Petermann Fiord and Bessel's Bay, and southward by way of Franklin and Crozier Islands, Capes Constitution and Andrew Jackson to the great Humboldt glacier, and they doubtless underlie the whole of the ice-cap covering Washington Land" (Feilden & De Rance 1878: 558 — see Fig. 8 for "Cape Andrew Jackson"). This is an amazingly accurate description of the geology of northern Nares Strait. It also outlines the geological boundary that has been most discussed in the question of lateral displacement along the Strait since Wegener (1915) first referred to it in that context.

Geology matured rapidly as a science during the late 1800s and the northern expeditions returned with many new collections and observations. Lieut. Adolphus W. Greely of the U.S. Army led a scientific expedition to Discovery Harbour in 1881–83; Greely's station, Fort



Fig. 6. The flat-lying Lower Palaeozoic carbonates at Victoria Head, eastern Bache Peninsula, seen from the south. The "syenitic and granitoid rocks" of the southern coast of Bache Peninsula are overlain by coarse basement rocks that "are overlain by mural cliffs of limestone, rising to a height of more than 1000 feet, dipping gently to N.N.W. as far as Victoria Head, where a landing was effected and a small collection of fossils obtained" (Feilden & De Rance 1878: 557. From Christie (1967: pl. XV).



Fig. 7. The Robeson Channel, the northernmost part of Nares Strait, with Hall Land, Greenland in the foreground. "Whether this channel was carved in the solid land by the force of pre-Adamite glaciers, or whether it is a Titanic cleft formed by the breaking off of Greenland from Grant Land, is a question still undetermined by geologists" (Peary 1910: 92).

Conger, was one of several occupied simultaneously during the "International Polar Year" of that time. The "Polar Year" was aimed mainly at meteorological, auroral and magnetic observations, but despite the lack of a geologist, the Greely party made important geological discoveries during their travels across Ellesmere Island, on Judge Daly Promontory and along the north coast of Greenland. Incidentally, the Greenland sledge party led by Lieut. J. B. Lockwood achieved a new farthest north (83°24'N) at Lockwood Ø, where volcanic rocks were recognised and collected by Sgt. D. L. Brainard, the first report of the Kap Washington volcanics (Greely 1886, vol. 1: 338). The Greely expedition began its disastrous retreat southward from Fort Conger in 1883 and was obliged to abandon the rock collections. During the same summer, Alfred G. Nathorst on the *Sophia* examined the geology in southern Nares Strait, although little information from this was published (Nathorst 1884, Nordenskiöld 1885).

George Mercer Dawson (1887), one of the pioneers of the Geological Survey of Canada, was able to include data from the Greely expedition in his geological map of the northern part of the Dominion of Canada and Greenland that was compiled in 1886 (Fig. 5). His report contains a full geological description of Nares Strait in a section "Ellesmere Land, Grinnell Land and neighbouring coast of North Greenland" (Dawson 1887: 51–58).

Lieut., later Commander, Robert E. Peary of the U.S. Navy travelled the length of Nares Strait by sledge and by ship several times between 1898 and 1909. His objectives were mainly in the exploration of northern Greenland and the attainment of the North Pole. Thus, while earth scientists were part of, or auxilliary to, the Peary expeditions (e.g., Thomas C. Chamberlin, C. C. Craft, L. L. Dyche, Angelo Heilprin, Rollin D. Salisbury and John M. Verhoeff), geological work was concentrated in the southern Nares Strait area and few geological results were released. Perhaps the most important is Chamberlin's (1895) description of the Thule region in which he makes a subdivision of the flat-lying strata seen by Sutherland 40 years before. Little attention was paid by Peary to the geology of the northerly channels and we are left with but sporadic references in Peary's accounts — see, for example, Peary's (1898) account of his early expeditions between 1891 and 1897.

In view of this it is perhaps surprising to find in Peary's (1910) book "The North Pole" one of the earliest references to the origin of Nares Strait and to the possibility of the tectonic displacement of Greenland (Fig. 7).

Following Fridtjof Nansen's epic crossing of the Arctic Ocean, the polar ship *Fram,* again commanded by Capt. Otto Sverdrup, was sailed into the Canadian Arctic Islands in 1898; when the party at last returned to Norway in 1902 they had advanced geographical and geological understanding of the islands more than any expedition to that time. The first winter in the ice was spent at Fram Haven, inside Pim Island, not far from a ship on which Peary was also wintering, on the the west shore of Kane Basin. The geologist, Per Schei, explored the surrounding region. The geology of both the Precambrian metamorphic complex and the overlying Palaeozoic and Mesozoic sedimentary rocks was firmly established through Schei's preliminary reports and sketch maps (Schei 1903, 1904, Fig. 5) and, following his death soon after his return, by amplified reports on his work written by others (Bugge 1910, Holtedahl 1913, 1917).

A Canadian government expedition in the DGS *Neptune,* with the geologist Albert P. Low in charge,

navigated as far north as Cape Sabine in 1904; the ship had sailed northward along the Greenland coast before crossing to Ellesmere Island. The expedition was in part scientific, but its main objectives were the establishment of Canadian sovereignty in the islands and the collection of whaling dues. Low (1906) included in his report a summary of the geology then known of the region traversed by his ship; his geological map included the eastern and northern islands and the northern part of the west coast of Greenland to the Robeson Channel (Fig. 5).

Schei and Low had occasion for only a brief examination of rocks in Greenland but remarked on the similarity of the geology on the two sides of Smith Sound. Schei noted in both lands that the crystalline basement was composed of a peculiar complex of igneous and metamorphic rocks overlain unconformably by a comparable sequence of homoclinal strata containing basic sills. Low recognised on both coasts steeply-dipping metasedimentary rock series containing crystalline limestone which were intruded by granites (Low 1906: 208). The state of geological knowledge of Nares Strait at that time is illustrated by Low's map, and also by Bailey Willis' (1912) map compilation included in his "Stratigraphy of North America" (Fig. 5). Naturally, both Low and Willis were influenced by Schei's work; in fact Schei's discoveries of deformed Devonian and Mesozoic deposits in central west Ellesmere Island radically affected geological interpretation of the Nares Strait region. Thus, the Cape Rawson Beds, considered of possible Huronian age by Feilden & De Rance (1878) and of Cambrian age by Dawson (1887), were interpreted as Triassic by Low and Willis. In addition, the flat-lying fossiliferous Silurian limestones of the Hall Basin – Kennedy Channel area, the age of which had been established decades before by Meek (1865) and Etheridge (1878), were indicated as Devonian. This age designation is interesting in the context of this symposium, since it was Willis' map that formed the basis of Wegener's (1915) initial comments about the geological evidence for the displacement of Greenland along Nares Strait (Fig. 13).

Geological exploration of the Nares Strait region was carried out by numerous 'explorer-geologists' using dog sledge and ship during the decades between 1910 and the late 1940s when aircraft began to be used in geological reconnaissance. The main studies were concentrated in the southern part of Nares Strait and it was during this phase that the similarity in the geology across Smith Sound, recognised by Schei and Low, was confirmed and placed in a regional context. The geologists involved in this phase were: W. Elmer Ebklaw of the Donald B. MacMillan Crocker Land Expedition 1913–17 (MacMillan 1918), who visited both sides of the Strait; Lauge Koch with Knud Rasmussen's 2nd Thule Expedition 1916–18 (Rasmussen 1921) and later on his own Bicentenary Jubilee Expedition 1920–23 (Koch 1926a); L. J. Weeks of the Geological Survey of Canada, who visited the Royal Canadian Mounted Police posts on Devon and Ellesmere Islands in 1925 (Weeks 1927); Robert Bentham of the Oxford University Ellesmere Land Expedition 1934–35 and indepen-

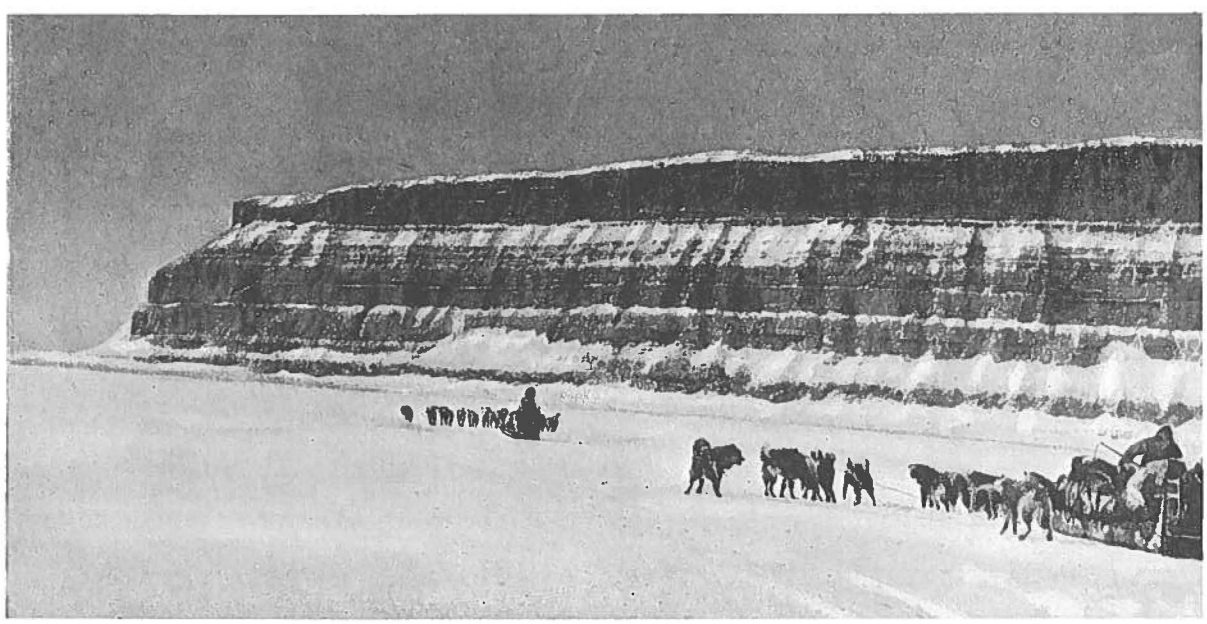


Fig. 8. Flat-lying strata of the Arctic platform — part of Lauge Koch's "great sediment plain" (see Fig. 14). These spectacular coastal cliffs of Ordovician limestones about 500 m high are of Kap Jackson, Washington Land, seen from the south-east. The homoclinal Lower Palaeozoic platform rocks of the Inglefield Land – Washington Land region of Greenland were first recorded by the Kane Expedition 1853–55. Photograph is from Koch (1920: fig. 7).

dently in 1936–38 and who wintered in both Greenland and Ellesmere Island (Bentham 1936, 1941); Sole Munck on the Danish Natural History Expedition to Northwest Greenland in 1936 (Munck 1941); James M. Wordie in 1937, who visited Bache Peninsula, southern Ellesmere Island and North-West Greenland (Wordie 1938) and finally Johannes C. Troelsen of the Danish Thule and Ellesmere Land Expedition 1939–1941, who travelled widely in Greenland and the Canadian Arctic Islands (Troelsen 1950).

Knowledge of the stratigraphy, structure and geological history steadily accumulated during this period and geological syntheses of the region were made by Koch (1920, 1925, 1929a, 1935) and Troelsen (1950, 1956). Koch travelled the entire Greenland coast from Melville Bugt to Peary Land carrying out cartographical and glaciological work, in addition to his geological tasks. He recognised the main stratigraphical and structural elements of the Nares Strait region (Fig. 14A), helped in his dog sledge reconnaissance methods by the beautifully exposed sea-cliff sections (Fig. 8). Koch's geological map contribution, particularly noteworthy, takes the form of five map sheets that cover the entire south-eastern coast of Nares Strait from Melville Bugt to eastern Peary Land (Fig. 9). Troelsen's work represents the first comprehensive study of parts of both sides of the Strait and his travels took him along much of the Greenland side between 77° and 81°N and across central Ellesmere Island and into Axel Heiberg Island. The geological map accompanying his 1950 report covers the region 78° to 82°N, with only a few blanks on the western side of Kennedy Channel marked as "unexplored" (Fig. 5).

Bentham (1936) had been impressed by the peneplained surface of the Precambrian basement in south-western Inglefield Land and Bache Peninsula and also by the similarity of the overlying strata. He remarks that parts of the cover series in Bache Peninsula "are indistinguishable from the Thule Formation of Inglefield Land. This similarity is emphasised by the occurrence in these beds, at Cape Camperdown, of two sills of dolerite" (Bentham 1936: 336). Troelsen, building on the observations of Bentham and Koch, established a common stratigraphic framework for the unmetamorphosed cover rocks of the Smith Sound region. The period of "early exploration" thus closes with identical formational names in use for both the Greenland and the Ellesmere Island successions.

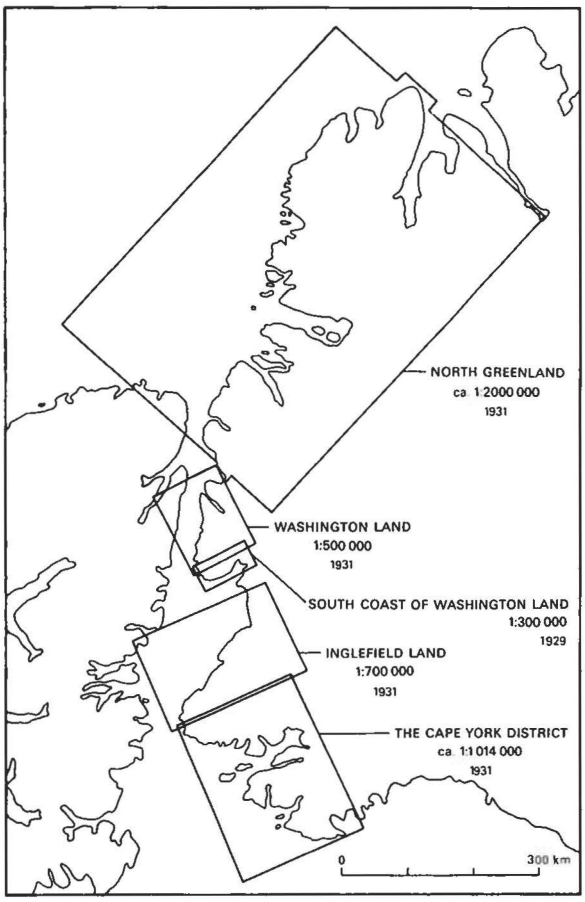


Fig. 9. Sketch showing the coverage of Lauge Koch's five coloured geological maps. The "south coast of Washington Land" and the "Inglefield Land" sheets were published by Koch (1929b, 1933); the remaining three maps, although originally printed in 1931, were published nearly half a century later (Dawes & Haller 1979).

Modern exploration (1946–1980)

The modern period of geological reconnaissance, in which aircraft support of field work became increasingly prominent, comprises the late 1940s and later decades. During this period, two strategic airports were established at the two ends of Nares Strait: the jointly operated Canadian–United States weather station at Alert in north-eastern Ellesmere Island, and the U.S. Thule Air Force Base at Dundas in Greenland (Fig. 1). Geological and other scientific studies were stimulated by the logistic services available during both the establishment of the airfields in the late 1940s and early 1950s and their later operation.

Many of the early helicopter operations in Greenland were restricted to the immediate area of Thule Air Base (Kurtz & Wales 1951, Fernald & Horowitz 1954, Davies et al. 1963) although some geological work (for example that of Vincent E. Kurtz and others (1952) in 1949 from the icebreaker USS *Edisto*) extended to the Canadian side of the Strait. The military projects "Operation Ice Cap" (1953–54), "Operation Defrost" (1956) and "Operation Groundhog" (1957–60) covered large areas of northern Greenland and during this period William E. Davies of the U.S. Geological Survey and others carried out a variety of bedrock and superficial studies on the Greenland side of Nares Strait



Fig. 10. View from Judge Daly Promontory across Hall Basin to Hall Land, Greenland. The Judge Daly fault — commonly regarded as the onland segment of the Wegener Fault joining Baffin Bay to the Arctic Ocean — is the conspicuous linear valley parallel to the outer coast. National Air Photo Library, Canada, T397C–148.

(Clebsch 1954, Davies et al. 1959, Nichols 1961, Davies 1972). John Cowie (1961) studied the Proterozoic and Lower Palaeozoic strata of Inglefield Land in 1957 and, having also visited equivalent units on Ellesmere and Devon Islands, later suggested correlation of certain formations across the Strait (Cowie 1971).

The Geological Survey of Canada began airborne geological reconnaissance in the Arctic Islands in 1947 (Fortier 1948). In the 1950s and following years with the development of 'bush flying' in the Far North, a number of operations used the logistic network of the "Joint Arctic Weather Stations" (or JAWS) and carried out work in the Nares Strait region. V. K. Prest, accompanying an icebreaker resupply mission in 1950 to the northernmost of the JAWS stations at Alert, made some geological observations in northern Baffin Bay and Nares Strait. Prest (1952) recorded coast-parallel faults in eastern Ellesmere Island along the Kennedy Channel and one of these — the Judge Daly fault (Fig. 10, see also Mayr & de Vries, this volume: fig. 3) — is a particularly conspicuous linear feature which has a significant place in the context of this symposium (Wilson 1963). Other investigations followed: G. Hattersley-Smith and R. G. Blackadar at Alert in 1953; Blackadar and R. R. H. Lemon in the Admiralty Inlet area of Baffin Island in 1954; R. Thorsteinsson at Copes Bay, Kane Basin, during the GSC's "Operation Franklin" in 1955 (Fortier et al. 1963); R. L. Christie, travelling the Hazen Plateau and Judge Daly Promontory during the Defence Research Board of Canada's "Operation Hazen" in 1957–58; Christie, in 1960 and 1961, carrying out a reconnaissance of south-eastern Ellesmere Island; J. W. Kerr and H. P. Trettin in central and eastern Ellesmere Island in 1961–62; and Blackadar, W. L. Davison and Trettin during the Survey's "Operation Admiralty" on northern Baffin Island in 1963.

The year 1965 marked a new venture in Nares Strait geological exploration: a two-year cooperative project of the Geological Surveys of Canada and Greenland — Operation Grant Land — was launched to examine the geology on both sides of the Kennedy Channel – Lincoln Sea region (Christie 1966). Christie, Trettin, J. H. Allaart, P. R. Dawes, T. Frisch, W. W. Nassichuk, B. S. Norford and A. A. Petryk took part in this operation. In 1968–69, Christie continued the reconnaissance of south-eastern Ellesmere Island and eastern Devon Island, in 1968 in the company of J. W. Cowie.

Special mention in this period must be made of B. R.

Pelletier, who carried out reconnaissance of the submarine geology of parts of Nares Strait and northern Baffin Bay in 1963, and of H. P. Trettin, who studied the central part of the Franklinian geosyncline of northern Ellesmere Island during much of the 1960s and 1970s. In addition, the 1960s saw the beginning of systematic aeromagnetic flights over the High Arctic and some of these programmes reached northern Baffin Bay and Nares Strait (e.g. Hood & Bower 1975).

The past decade, from 1970, has seen considerable work by the Geological Surveys of Canada and Greenland, some of it cooperative, on both sides of Nares Strait and carried out by geologists too numerous to list in this paper. The field programmes have been generally undertaken with helicopter and fixed-wing support and a wealth of more detailed information has been placed in its regional context, helped by the scientific cooperation between the two Surveys. The results of these recent field projects, of course, are but recently published and many appear in this volume.

In addition, field parties of mineral exploration groups have collected valuable structural and stratigraphic data. In this field no history would be complete without mention of the pioneer work of the Canadian J. C. Sproule whose confidence in the economic potential of the Far North led to geological studies that stretched from the Beaufort Sea across the Canadian Arctic Islands to north-eastern Greenland. A notable contribution was the regional assessment of northern Greenland from Melville Bugt to Peary Land carried out from 1969 to 1973. This work made available important new information, including, incidentally, the initial discovery of Upper Cambrian strata in the High Arctic (1969, in Washington Land). Geologists of the J. C. Sproule company covered much of the Canadian side of Nares Strait and of note in the context of this symposium is the discovery in 1972–73 of deformed Tertiary conglomerates on the shore of the Kane Basin — information released by the company for this symposium (Mayr & de Vries, this volume).

Geological syntheses

The papers that follow in this volume will consider today's understanding as well as the problems of the Nares Strait region. First, however, it should be interesting to take note of the ideas expressed in some published syntheses arising from the early geological reports: Eduard Suess, Frank B. Taylor, Alfred L. Wegener and Charles Schuchert each considered various geological aspects of Greenland, the Canadian Arctic Islands or the intervening seaways, while Lauge Koch as a geologist involved in the actual collection of the early data, also expressed ideas pertaining to the regional geology of Nares Strait.

Eduard Suess, Viennese professor of geology, is justly famous for his monumental treatise, *Das Antlitz der Erde* (The face of the Earth), published in several volumes and translations between 1885 and 1924. In volume 2, Suess, writing about the regional geology of the North American continent north of the great Archaean Shield, states that "... the several sedimentary formations of this border are so arranged that they strike to the east or north-east, and become progressively more recent as they are traced towards the pole" (Suess 1906: 40) — an early description of the passage from the Precambrian craton and platform into the Franklinian geosyncline. Chapter VII of volume 4 (translation published in 1909) of the monograph concerns "Laurentia and the Northern Islands". Suess described the fold mountains of Kennedy Channel as due to south-directed orogenic forces with the Canadian Shield acting as a foreland; he further compared the folded range with the "Asiatic arcs" and regarded the Ellesmerian folds as a "terminal part of the Asiatic structure extending across the North Pole" (Suess 1909: 251). Greenland was considered to be a fragment of Laurentia (the ancient continental nucleus of North America) that stands as a horst "between subsidences of different age" (Suess 1909: 258). In this early and remarkably complete synthesis of Nares Strait geology it is clear that Suess recognised the faulted character of the Strait and considered the "fragments" of Laurentia, Greenland and the Canadian Shield to be more or less undisturbed in a horizontal sense along Nares Strait.

Two creative theorists soon used Suess' compendious information on the geology of the Earth to advance new global theories: F. B. Taylor in the United States and Alfred Wegener in Germany both proposed that the continents had been horizontally displaced or had "drifted". Greenland was an important element in each of the new theories and both authors regarded the narrow channel between Ellesmere Island and Greenland as a rift along which there has been appreciable sinistral transcurrent motion (Fig. 11A and B), although they interpreted the movement roles of the two crustal blocks — North America and Greenland — differently. Their respective theories, earlier often referred to as "The Taylor–Wegener hypothesis", differed also in other important respects.

Frank Bursley Taylor, a Quaternary geologist by training, had deliberated for many years on "the structure and disposition of continents"; according to Totten (1981) at least 20 years. Finally, he read a paper in 1908 to the Geological Society of America on "continental movement" and its relationship to the Earth's Tertiary mountain belts. Taylor (1910) described a "mighty creeping" movement of the continental crust away from the Earth's poles. He regarded the Tertiary mountain belts to have formed by horizontal collision of crustal "sheets", i.e. the "plates" of Suess, and he described the mid-Atlantic ridge as "a residual ridge along a line of parting or rifting" so that "the two continents on opposite sides of it have crept away in nearly parallel and opposite directions" (Taylor 1910: 217–218). He

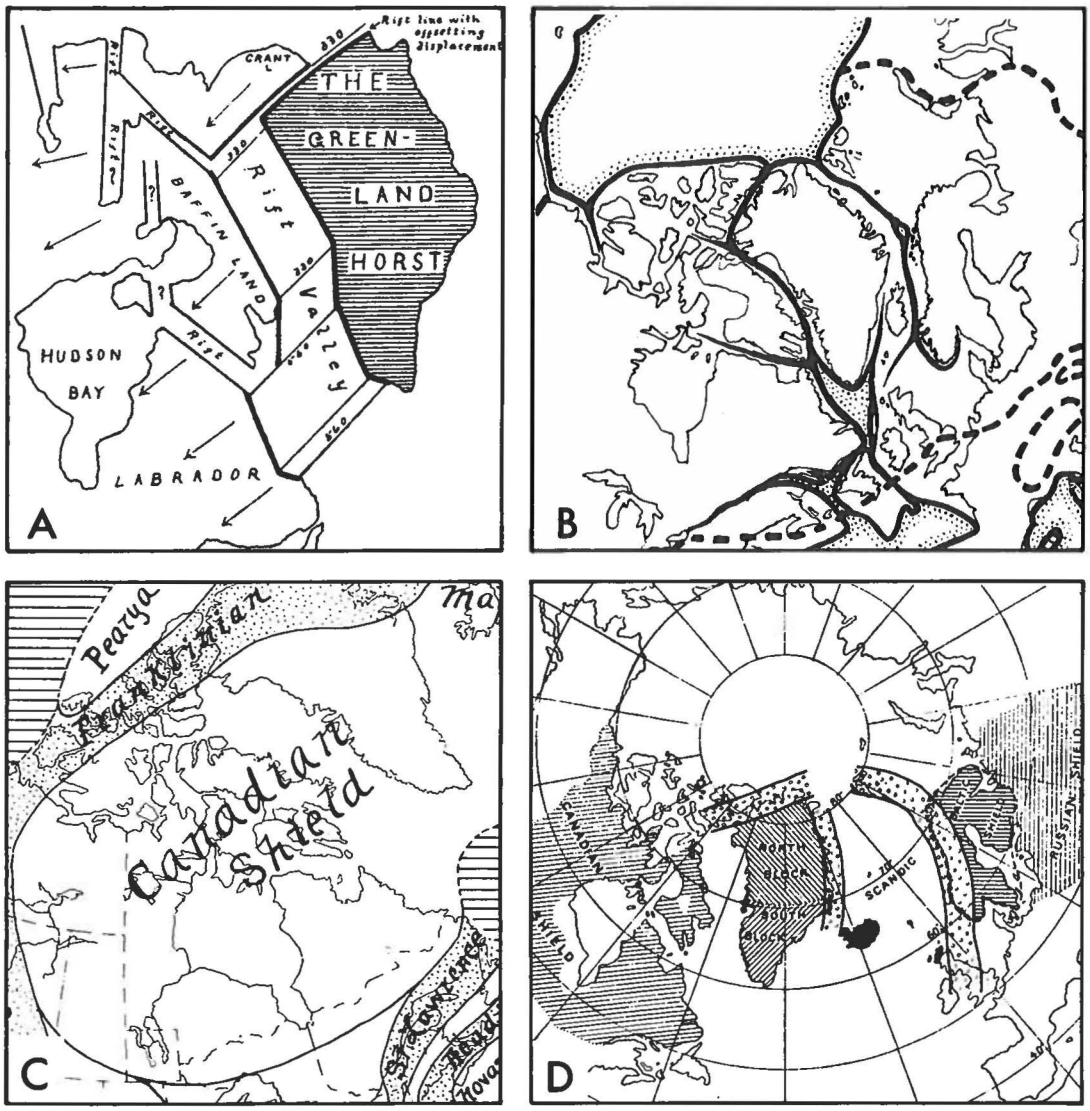


Fig. 11. Maps illustrating some early regional geological syntheses of the Nares Strait region. A = Taylor (1910), B = Wegener (1915), C = Schuchert (1923), D = Koch (1929a). Maps reproduced at approximately the same scale as the originals; B, C & D are taken from larger maps.

matched the coastlines of the Atlantic Ocean and concluded that Greenland was once united with NW Europe, noting the reconstructed shape of the mountain belts. The west coast of Greenland he matched with the eastern edge of the Canadian Arctic Islands: Greenland, according to Taylor, was alone an immovable region of the northern hemisphere. He called Greenland "the great northern horst" and concluded that North America had moved some 330 miles away from this horst (Fig. 11A).

Alfred Lothar Wegener held a doctorate in astronomy and had a keen interest in meteorology and geophysics as well as polar exploration. At about the time that Taylor's first paper appeared, he became inspired by an English collier, W. H. Pickering, who suggested that the Earth's moon was former continental material that had been flung out into orbit (Wood 1980). That part of the crust remaining on Earth "was torn in two to form the eastern and western continents. These then floated on the liquid surface like two large ice-floes" (Pickering 1907: 30). Pickering, like Taylor, also matched the coastlines of the Atlantic Ocean. Wegener also quoted Taylor's paper and accepted Greenland's stationary "keystone" position in the North Atlantic Ocean. Wegener's first papers appeared in 1912 and his book, in which "Pangaea" is described, was published in 1915. Revised 2nd, 3rd and 4th editions were published in 1920, 1922 and 1929, and his work was translated into



Fig. 12. Alfred Wegener at his desk in the expedition station "Borg", at Storstrømmen, East Greenland during the winter 1912–13. Wegener gave his first public lectures on continental drift early in 1912 and his initial short papers on the subject were written on his way by ship to Greenland later the same year (Wegener 1912a, b). His first book on the subject, including the appraisal of the geology of the Nares Strait region, was published on his return from the 1912–13 expedition after the crossing of the Inland Ice. Photograph copyright, Arktisk Institut, Denmark.

several languages between 1924 and 1926. Wegener believed that geodetic measurements between Greenland and Europe could confirm the (supposedly rapid) active separation of the continents. By 1913 Wegener had participated in two Greenland expeditions (Fig. 12) and in 1929 he organised an expedition to the subcontinent, one goal of which was the accurate measurement of longitude using radio. Wegener died in 1930 during an early winter trek, and he was buried in the firn of the Inland Ice. The subcontinent Greenland and the theory of continental drift have since become immortally associated with the memory of Alfred Wegener. The geology of the Nares Strait region featured prominently in Wegener's arguments for the horizontal displacement of the continents (Fig. 13) and Nares Strait has become synonymous with "The Wegener Fault".

Charles Schuchert, Yale professor of palaeontology and historical geology, is the last of the early writers considered here, who synthesised Arctic geology into 'continental' models. Schuchert (1923) assigned the name Franklinian geosyncline to the sedimentary belt of northern Ellesmere Island and Greenland, and he also named Pearya, a hypothetical landmass that was to have lain 'outboard' of the geosyncline (Fig. 11C). Metamorphic rocks of north-coastal Ellesmere Island and Peary Land were taken to be remnants of the ancient borderland, much of which had "fractured into ocean depths". The fractures that transect the continent, however, were not considered, so that the Canadian Shield or "Laurentia" appears, in Schuchert's figures, as an undisturbed landmass. The Franklinian geosyncline was depicted as a sedimentary belt unaffected by major displacement along Nares Strait; incidentally, later, Schuchert (1928) did consider the implications of the reconstruction of Wegener's Pangaea on the disposition of the northern geosynclinal belts and he argued against Wegener's treatment of the geological data.

Lauge Koch was a cartographer, geologist and explorer. His field work in the Nares Strait region was specifically directed to Greenland territory, although in 1921 he visited Ellesmere Island. Koch recognised Schuchert's Franklinian geosyncline, but in his doctorate thesis "Stratigraphy of Greenland" (1929a) referred to it as the Smith Sound geosyncline. The 1929 paper outlines Koch's early ideas about the "Shields and geosynclines of the North Atlantic Ocean" and he divided the Smith Sound geosyncline into a number of structural units (Fig. 14A). Koch regarded this geosyncline, deformed by "Caledonian Folding", to strike uninterruptedly across Nares Strait, to join up with the northern parts of the East Greenland and Scandinavian–Svalbard geosynclines in the Barents Shelf region (Fig. 11D).

Koch was impressed by the fault tectonics in the southern Nares Strait region and he described "a new fault zone in Northwest Greenland" (Koch 1926b) as a major Tertiary feature and suggested its counterpart in the Arctic Islands (Fig. 14). He concluded that "future investigations will reveal strong faulting north of Baffin Bay, which faulting has possibly some connection with the development of Davis Strait – Baffin Bay" (Koch 1929a: 319). However, the "new fault zone" traverses southern Nares Strait and the Strait itself was not considered an associated lineament.

Koch's (1920) early geological map of the Nares Strait region was used by Wegener in later editions of his thesis on continental drift, replacing the map by the U.S. Geological Survey, i.e. compiled by Bailey Willis

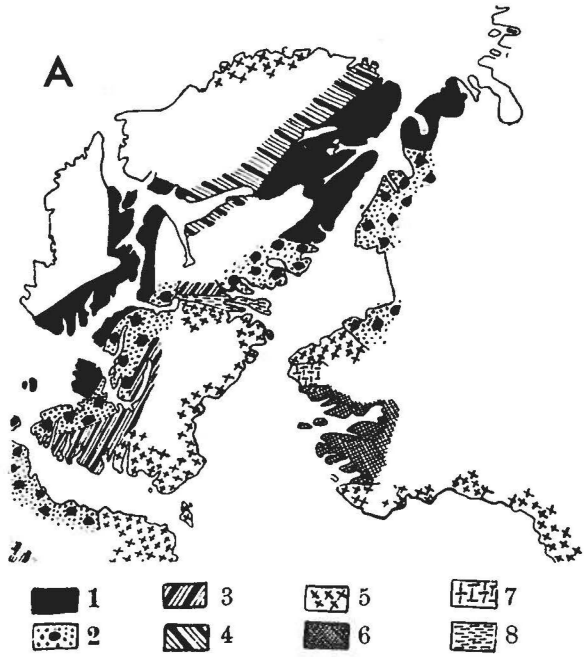
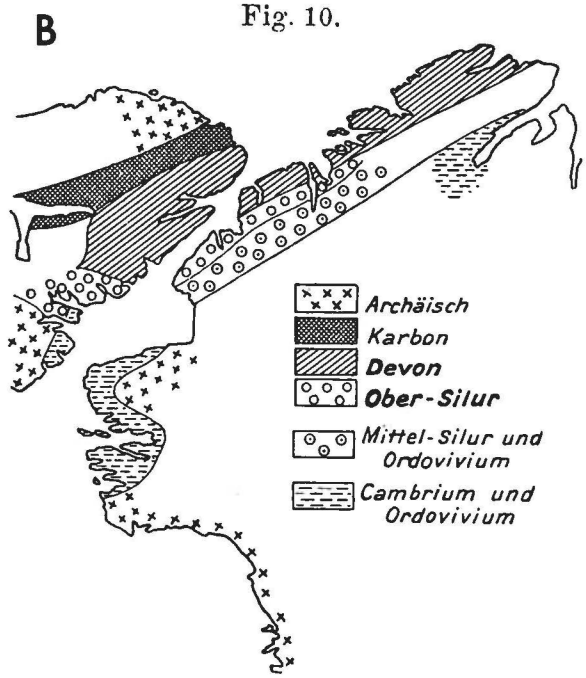


Fig. 15. Geologische Karte des Smith-Sundes und des Robeson-Kanals, nach der Geologic Map of North America.

1 Trias, 2 Devon. 3 Silur, 4 Karbon, 5 Gneis, 6 Vorkambrium, 7 Spättertiär, 8. Kambrium und Unter-Ordovicium.

Fig. 10.



Geologische Karte von Nordwestgrönland, nach Lauge-Koch.

Fig. 13. The geological maps used by Wegener in his account of the horizontal displacement of Greenland relative to North America. A: U.S. Geological Survey map compiled by Willis (1912) (from Wegener 1915: fig. 15). B: Lauge Koch's (1920) map (from Wegener 1922: fig. 10).

(Figs 5 and 13). Wegener drew attention to the position of two geological boundaries which he regarded as indicative of major strike-slip displacement of Greenland past Ellesmere Island. "If one looks for the boundary line between the Devonian and the Silurian periods, this lies at 80°10′ in Grinnell Land and 81°30′ in Greenland. Also, in the Caledonian fold system discovered by this author [Koch], which stretches across from Greenland to Grinnell Land, one can detect the same displacement" (Wegener 1966: 79). However, Koch did not agree with this interpretation and remarked "that detailed investigations, especially on Grinnell Land, are required before a decided standpoint can be taken for or against Wegener's view" (Koch 1929a: 312). Incidentally, Koch (1924) had noted the absence of faults on the Greenland shore of the "Smith Sound–Robeson Channel" and concluded that the seaway was mainly a product of river erosion, although later (1929a: 313) he raises the possibility that the channel could be a "Grabensenkung".

Wegman (1948) included Koch's viewpoint in his account of the scientific planning of geological tests of continental drift, while Willis (1928), although not directly referring to Wegener's use of his map compilation of the Nares Strait, argued strongly from geological evidence against the theory of continental drift.

Interestingly enough, the early dispute about the interpretation of the geology of the Nares Strait region illustrated above has continued, albeit in various nuances, until the present day — hence this symposium. Thus Taylor's and Wegener's ideas about the mass movements of continental blocks and more specifically the relationship between Greenland and North America have gained recognition through the modern concept of plate tectonics and sea-floor spreading which predicts Nares Strait to be a site of appreciable transcurrent motion. Likewise, Schuchert's and Koch's ideas about continental models, cratons and geosynclinal belts have been expounded upon by the systematic study of the continental blocks and sedimentary basins which has strengthened the view that the geology of the Nares Strait region does not indicate that such major strike-slip displacement has occurred.

Present status

From the preceding paragraphs it is evident that the geological understanding in the Nares Strait region proceeded at a slow pace during early exploration, but quickened with time and the development of the science. In addition, progress has 'see-sawed' on the two

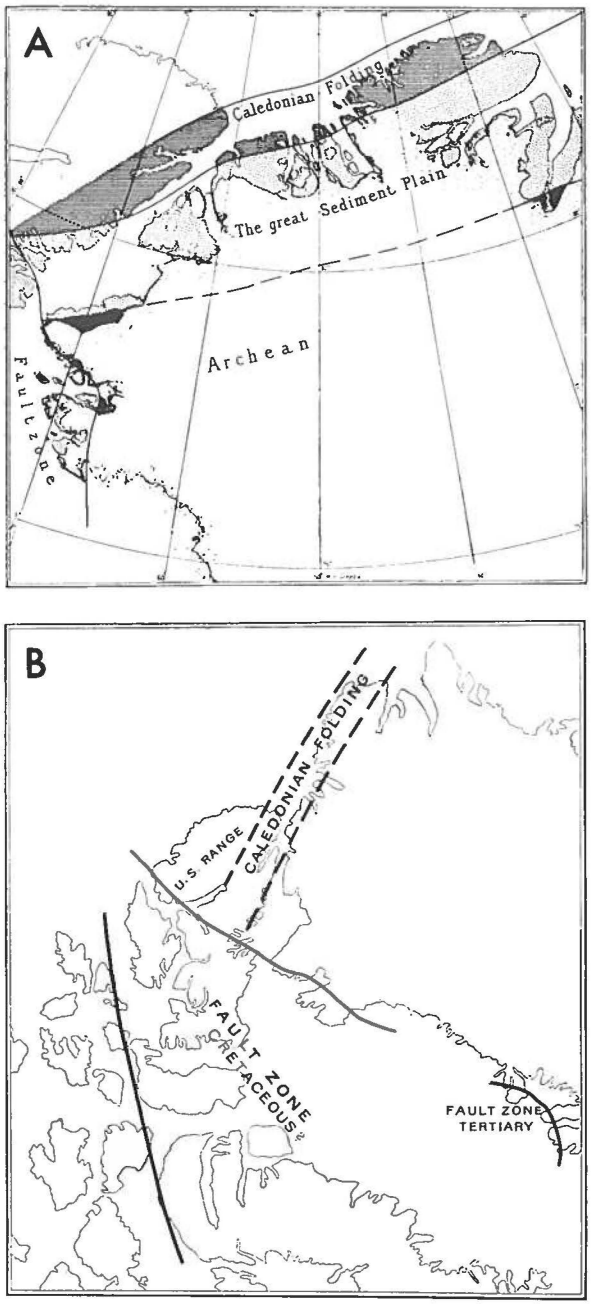


Fig. 14. Koch's interpretation of the structural framework of the Nares Strait region. A = main structural elements of eastern Ellesmere Island and northern Greenland (from Koch 1925: fig. 2). B = relationship of the late Phanerozoic faults of Baffin Bay and the Caledonian folding (from Koch 1929a: fig. 61).

sides of the Strait: thus, following Lauge Koch's reconnaissance mapping of the Greenland side between 1916 and 1923, systematic work on the Canadian side did not begin until the 1950s when aircraft were used in the field work. The Geological Survey of Canada's work has continued through a series of airborne 'operations' until today. The Geological Survey of Greenland began its systematic mapping of the Greenland side of the Strait in the Thule region in 1971 and studies have continued along the Strait up to the present day.

In terms of modern onshore geological map coverage, the whole of the Nares Strait region (Baffin Bay to Lincoln Sea) has been mapped at 1:250 000 and in many areas at 1:100 000 or larger scales. Many of the maps of the Canadian side have been published; maps of the Greenland side are in compilation or in drafting stage. The cumulative knowledge of the region is now at a level which allows reasonably confident assessment and correlation of the stratigraphy and structure of these parts of Greenland and adjacent Canada.

Acknowledgements

We are grateful to the following persons and institutions for providing or allowing publication of the photographs reproduced in Fig. 2: The Royal Geographical Society, London (G. S. Nares), The Museum of the King's Own Royal (Lancaster) Regiment, Lancaster (H. W. Feilden), The Allen County Public Library, Fort Wayne, Indiana (F. B. Taylor), Det Norske Videnskaps-Akademi, Oslo (P. Schei), H. Hirschsprungs Forlag, Copenhagen (A. L. Wegener), Gyldendalske Boghandel, Nordiske Forlag, Copenhagen (W. E. Ebklaw), Ove Koch and family, Copenhagen (L. Koch), Mrs Patricia O. Bentham, Sussex, England (R. Bentham) and Eigil Knuth, The Danish Peary Land Expeditions, Copenhagen (J. C. Troelsen).

We also thank the following persons for information and help in tracking down the photographs and biographical data used in Fig. 2: Lord Shackleton and Geoffrey Hattersley-Smith (London), George White (Univ. of Illinois), Knut Heggelund (Norsk Polarinstitutt, Oslo), Terence Armstrong and H. G. R. King (Scott Polar Research Institute, Cambridge), Mrs. Edith Tyson (Lancaster City Council, U.K.), Brig. R. J. Lewendon (Royal Artillery Institution, London), Mrs. Susan Pallone (Fort Wayne, Indiana), Ole G. Evensen (Norges Geologiske Undersøkelse, Trondheim) and D. S. Stonham (National Maritime Museum, London). Figures 10 and 12 are reproduced with permission of the National Air Photo Library, Ottawa and the Arktisk Institut, Denmark, respectively.

The paper was critically read by Anthony K. Higgins and for his comments and those of Esben Glendal and W. Stuart Watt, we are most grateful. The paper is published with the permission of the Director of the Geological Survey of Greenland.

References

Bentham, R. 1936. Appendix IV, Geology. – In: Shackleton, E., Arctic journeys. The story of the Oxford University Ellesmere Land Expedition 1934–5: 333–337. – Hodder & Stoughton Ltd., London.

Bentham, R. 1941. Structure and glaciers of southern Ellesmere Island. – Geogr. J. 97: 36–45.

Bessels, E. 1879. Die amerikanische Nordpol-Expedition. – Von Wilhelm Engelmann, Leipzig: 647 pp.

Bugge, C. 1910. Petrographische Resultate der 2ten Fram-Expedition. – Rep. 2nd Norwegian Arctic Exped. "Fram" 1898–1902, 22: 38 pp. – Videnskabs-Selskabet, Kristiania.

Chamberlin, T. C. 1895. Appendix A. Geology. – In: Bryant, H. G., The Peary Auxiliary Expedition of 1894. – Bull. geogr. Club Philadelphia 1(5): 166–174.

Christie, R. L. 1966. Operation Grant Land. – Pap. geol. Surv. Can. 66–1: 3–4.

Christie, R. L. 1967. Bache Peninsula, Ellesmere Island, Arctic Archipelago. – Mem. geol. Surv. Can. 347: 63 pp.

Christie, R. L. & Kerr, J. W. 1981. Geological exploration of the Canadian Arctic Islands. – In: Zaslow, M. (ed.), A century of Canada's Arctic Islands, 1880–1980: 187–202. – Royal Soc. Can., Ottawa.

Clebsch, A. 1954. The bedrock geology of Inglefield Land. – Prelim. rep., Program Operation Ice Cap 1953, SRI Project 828. – Stanford Res. Inst., California.

Cowie, J. W. 1961. Contributions to the geology of North Greenland. – Meddr Grønland 164(3): 47 pp.

Cowie, J. W. 1971. The Cambrian of the North American Arctic regions. – In: Holland, C. H. (ed.), Cambrian of the New World: 325–383. – Wiley-Interscience, New York.

Davies, W. E. 1972. Landscape of northern Greenland. – Spec. Pap. Cold Reg. Res. Engng Lab. 164: 54 pp.

Davies, W. E., Needleman, S. M. & Klick, D. W. 1959. Report on Operation Groundhog (1958) North Greenland. Investigation of ice-free sites for aircraft landings, Polaris Promontory, North Greenland. – U.S. Air Force Cambridge Res. Center, Bedford: 45 pp.

Davies, W. E., Krinsley, D. B. & Nicol, A. H. 1963. Geology of the North Star Bugt area, Northwest Greenland. – Meddr Grønland 162(12): 68 pp.

Davis, C. H. 1876. Narrative of the North Polar Expedition U.S. Ship Polaris, Captain Charles Francis Hall commanding. – Gov. Print. Off., Washington: 696 pp.

Dawes, P. R. & Haller, J. 1979. Historical aspects in the geological investigation of northern Greenland. Part 1: New maps and photographs from the 2nd Thule Expedition 1916–1918 and the Bicentenary Jubilee Expedition 1920–1923. – Meddr Grønland 200(4): 38 pp.

Dawson, G. M. 1887. Notes to accompany a geological map of the northern portion of the Dominion of Canada, east of the Rocky Mountains. – Geol. Nat. Hist. Surv. Can. Ann. Rep. II (R): 56 pp.

Etheridge, R. 1878. Palæontology of the coasts of the Arctic lands visited by the late British Expedition under Captain Sir George Nares, R. N., K. C. B., F. R. S. – Q. Jl geol. Soc. Lond. 34: 568–636.

Feilden, H. W. & De Rance, C. E. 1878. Geology of the coasts of the Arctic lands visited by the late British Expedition under Captain Sir George Nares, R. N., K. C. B., F. R. S. – Q. Jl geol. Soc. Lond. 34: 556–567.

Fernald, A. T. & Horowitz, A. S. 1964. Bedrock geology of the Nunatarssuaq area, Northwest Greenland. – Meddr Grønland 172(6): 44 pp.

Fortier, Y. O. 1948. Flights in 1947 over the region of the North Magnetic Pole and the mainland between the Arctic coast, Great Slave Lake, and Hudson Bay, Northwest Territories. – Pap. geol. Surv. Can. 48–23: 11 pp.

Fortier, Y. O., Blackadar, R. G., Glenister, B. F., Greiner, H. R., McLaren, D. J., McMillan, N. J., Norris, A. W., Roots, E. F., Souther, J. G., Thorsteinsson, R. & Tozer, E. T. 1963. Geology of the north-central part of the Arctic Archipelago, Northwest Territories (Operation Franklin). – Mem. geol. Surv. Can. 320: 671 pp.

Greely, A. W. 1886. Three years of Arctic service. An account of the Lady Franklin Bay Expedition of 1881–84 and the attainment of the farthest north. – Richard Bentley & Son, London, vol. I: 428 pp., vol. II: 444 pp.

Haughton, S. 1858. Geological notes and illustrations. – In: M'Clintock, F. L., Reminiscences of Arctic ice-travel in search of Sir John Franklin and his companions. – J. Roy. Dublin Soc. 1: 183–250.

Haughton, S. 1859. Appendix No. IV. Geological account of the Arctic Archipelago, drawn up principally from the specimens collected by Captain F. L. M'Clintock, R. N., from 1849 to 1859. – In: M'Clintock, Capt. [F.L.], A narrative of discovery of the fate of Sir John Franklin and his companions: 372–399. – John Murray, London.

Hayes, I. I. 1867. The Open Polar Sea. A narrative of a voyage of discovery towards the North Pole in the schooner "United States". – Hurd and Houghton, New York: 454 pp.

Holtedahl, O. 1913. The Cambro-Ordovician beds of Bache Peninsula and the neighbouring regions of Ellesmere Land. – Rep. 2nd Norwegian Arctic Exped. "Fram" 1898–1902, 28: 14 pp. – Videnskabs-Selskabet, Kristiania.

Holtedahl, O. 1917. Summary of geological results. – Rep. 2nd Norwegian Arctic Exped. "Fram" 1898–1902, 36: 27 pp. – Videnskabs-Selskabet, Kristiania.

Hood, P.[J.] & Bower, M.[E.] 1975. Aeromagnetic reconnaissance of Davis Strait and adjacent areas. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 433–451.

Kane, E. K. 1856. Arctic explorations: The Second Grinnell Expedition in search of Sir John Franklin, 1853, '54, '55. – Childs & Petersen, Philadelphia, vol. 1: 464 pp., vol. 2: 467 pp.

Koch, L. 1920. Stratigraphy of Northwest Greenland. – Meddr dansk geol. Foren. 5(17): 78 pp.

Koch, L. 1924. De videnskabelige resultater af Jubilæumsekspeditionen Nord om Grønland. Rapport 1: Kartografi og geologi. – Geogr. Tidsskr. 27: 208–218.

Koch, L. 1925. The geology of North Greenland. – Am. J. Sci. 9: 271–285.

Koch, L. 1926a. Report on the Danish Bicentenary Jubilee Expedition north of Greenland 1920–23. – Meddr Grønland 70,1(1): 1–232.

Koch, L. 1926b. A new fault zone in Northwest Greenland. – Am. J. Sci. 12: 301–310.

Koch, L. 1929a. Stratigraphy of Greenland. – Meddr Grønland 73,2(2): 205–320.

Koch, L. 1929b. The geology of the south coast of Washington Land. – Meddr Grønland 73,1(1): 39 pp.

Koch, L. 1933. The geology of Inglefield Land. – Meddr Grønland 73,1(2): 38 pp.

Koch, L. 1935. Geologie von Grönland. – In: Krenkel, E. (ed.), Geologie der Erde: 159 pp. – Gebrüder Borntraeger, Berlin.

Kurtz, V. E. & Wales, D. B. 1951. Geology of the Thule area, Greenland. – Proc. Oklahoma Acad. Sci. 31: 83–89.

Kurtz, V. E., McNair, A. H. & Wales, D. B. 1952. Stratigraphy of the Dundas Harbour area, Devon Island, Arctic Archipelago. – Am. J. Sci. 250: 636–655.

Loomis, C. C. 1971. Weird and tragic shores. The story of Charles Francis Hall, explorer. – Alfred A. Knopf, New York: 367 pp.

Low, A. P. 1906. Report on the Dominion Government Expedition to Hudson Bay and the Arctic Islands on board the D. G. S. Neptune 1903–1904. – Gov. Print. Bureau, Ottawa: 355 pp.

MacMillan, D. B. 1918. Four years in the white north. – Harper & Brothers Publishers, New York: 426 pp.

Mayr, U. & de Vries, C. D. S. 1982. Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago. – This volume.

Meek, F. B. 1865. Preliminary notice of a small collection of fossils found by Dr. Hays, on the west shore of Kennedy Channel, at the highest northern localities ever explored. – Am. J. Sci. 40: 31–34.

Munck, S. 1941. Geological observations from the Thule District in the summer of 1936. – Meddr Grønland 124(4): 38 pp.

Nares, G. S. 1878. Narrative of a voyage to the Polar Sea during 1875–6 in H. M. Ships 'Alert' and 'Discovery', with notes on the natural history edited by H. W. Feilden, F. G. S., C. M. Z. S. naturalist to the expedition. – Sampson Low, Marston, Searle and Rivington, London, vol. 1: 395 pp, vol. 2: 378 pp.

Nathorst, A. G. 1884. Den Svenska expeditionen till Grønland år 1883. – Ymer 4(1): 15–38.

Nichols, R. L. 1961. Geomorphology of Inglefield Land, North Greenland. – Meddr Grønland 188(1): 109 pp.

Nordenskiöld, A. E. 1885. Den andra Dicksonska expeditionen till Grønland dess inre Isöken och dess Ostkust itförd år 1883. – F. & G. Beijer, Stockholm: 328–334.

Peary, R. E. 1898. Northward over the "Great Ice". A narrative of life and work along the shores and upon the interior ice-cap of northern Greenland in the years 1886 and 1891–1897. – Frederick A. Stokes, New York, vol. 1: 521 pp., vol. 2: 618 pp.

Peary, R. E. 1910. The North Pole. – Hodder and Stoughton, New York: 326 pp.

Pickering, W. H. 1907. The place of origin of the moon – the volcanic problem. – J. Geol. 15: 23–28.

Prest, V. K. 1952. Notes on the geology of parts of Ellesmere and Devon Islands, Northwest Territories. – Pap. geol. Surv. Can. 52–32: 15 pp.

Rasmussen, K. 1921. Greenland by the Polar Sea. The story of the Thule Expedition from Melville Bay to Cape Morris Jesup. – William Heinemann, London: 326 pp.

Ross, J. 1819. A voyage of discovery, made under the orders of the Admiralty in His Majesty's Ships Isabella and Alexander, for the purpose of exploring Baffin's Bay, and inquiring into the probability of a North-West Passage. – John Murray, London: 252 pp. + appendices: cxliv pp.

Schei, P. 1903. Summary of geological results. – In: Sverdrup, O., The Second Norwegian Polar Expedition in the "Fram", 1898–1902. – Geogr. J. 22: 56–65.

Schei, P. 1904. Appendix 1. Preliminary account of the geological investigations made during the Second Norwegian Polar Expedition in the Fram. – In: Sverdrup, O., New Land. Four years in the Arctic regions, vol. 2: 455–466. – Longmans, Green, and Co., London.

Schuchert, C. 1923. Sites and nature of the North American geosynclines. – Bull. geol. Soc. Am. 34: 151–230.

Schuchert, C. 1928. The hypothesis of continental displacement. – In: van Waterschoot van der Gracht, W. A. J. M. et al., Theory of continental drift. A symposium on the origin and movement of land masses both inter-continental and intra-continental, as proposed by Alfred Wegener: 104–144. – Am. Ass. Petrol. Geol., Tulsa, Oklahoma.

Suess, E. 1906. The face of the Earth (Das Antlitz der Erde). Vol. II. – Clarendon Press, Oxford: 556 pp. (Transl. by H. B. C. Sollas).

Suess, E. 1909. The face of the Earth (Das Antlitz der Erde). Vol. IV. – Clarendon Press, Oxford: 673 pp. (Transl. by H. B. C. Sollas).

Sutherland, P. C. 1853a. On the geological and glacial phænomena of the coasts of Davis' Strait and Baffin's Bay. – Q. Jl geol. Soc. Lond. 9: 296–312.

Sutherland, P. C. 1853b. Appendix II. A few remarks on the physical geography, & c., of Davis Strait, and its east and west shores. – In: Inglefield, E. A., A summer search for Sir John Franklin; with a peep into the Polar Basin: 145–192. – Thomas Harrison, London.

Taylor, A. 1955. Geographical discovery and exploration in the Queen Elizabeth Islands. – Mem. Geogr. Branch, Dept. Mines Tech. Surv. Can. 3: 172 pp.

Taylor, F. B. 1910. Bearing of the Tertiary mountain belt on the origin of the earth's plan. – Bull. geol. Soc. Am. 21: 179–226.

Totten, S. M. 1981. Frank B. Taylor, plate tectonics, and continental drift. – J. geol. Education 29: 212–219.

Troelsen, J. C. 1950. Contributions to the geology of Northwest Greenland, Ellesmere Island and Axel Heiberg Island. – Meddr Grønland 149(7): 85 pp.

Troelsen, J. C. 1956. The Cambrian of North Greenland and Ellesmere Island. – In: El sistema Cámbrico, su paleogeografía y el problema de su base. – 20 Intern. geol. Congr. México, Symp. 3(1): 71–90.

Weeks, L. J. 1927. The geology of parts of eastern Arctic Canada. – Sum. Rep. geol. Surv. Can. 1925, pt. C: 136–141.

Wegener, A 1912a. Die Entstehung der Kontinente. – Peterm. Mitt. 58(1): 185–195, 253–256, 305–309.

Wegener, A. 1912b. Die Entstehung der Kontinente. – Geol. Rund. 3: 276–292.

Wegener, A. 1915. Die Entstehung der Kontinente und Ozeane. – Friedr. Vieweg & Sohn, Braunschweig: 94 pp.

Wegener, A. 1922. Die Entstehung der Kontinente und Ozeane. – Friedr. Vieweg & Sohn, Braunschweig: 144 pp. (3rd revised edit.).

Wegener, A. 1924. The origin of continents and oceans. – Methuen & Co., London: 212 pp. (Transl. 3rd edit. by J. G. A. Skerl).

Wegener, A. 1966. The origin of continents and oceans. – Dover Publications, Inc., New York: 246 pp. (Transl. 4th edit. by J. Biram).

Wegmann, C. E. 1948. Geological tests of the hypothesis of continental drift in the Arctic regions. Scientific planning. – Meddr Grønland 144(7): 48 pp.

Willis, B. 1912. Index to the stratigraphy of North America. – Prof. Pap. U.S. geol. Surv. 71: 894 pp.

Willis, B. 1928. Continental drift. – In: van Waterschoot van der Gracht, W. A. J. M. et al., Theory of continental drift. A symposium on the origin and movement of land masses both inter-continental and intra-continental, as proposed by Alfred Wegener: 76–82. – Am. Ass. Petrol. Geol., Tulsa, Oklahoma.

Wilson, J. T. 1963. Hypothesis of Earth's behaviour. – Nature, Lond. 198: 925–929.

Wordie, J. M. 1938. An expedition to Northwest Greenland and the Arctic Islands in 1937. – Geogr. J. 92: 385–421.

Wood, R. M. 1980. Coming apart at the seams [on the centennial of Alfred Wegener's birth]. – New Scientist 85: 252–254.

History and implications of the Nares Strait conflict

J. WILLIAM KERR

Kerr, J. W. 1982. History and implications of the Nares Strait conflict. - In: Dawes, P.R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. - Meddr Grnnland, Geosci. 8: 37-49.

The question "Did Greenland drift along Nares Strait?" has been debated for more than 70 years. There have been five periods, each represented by a consensus on one or the other extreme, or by a conflict.

At times opinions on the Strait have changed rapidly when a new global theory came in or went out. There is a possibility that at these times opinion on the Strait has been determined more by theory than by evidence.

At present there is a conflict. One group of earth scientists have worked on the shores of the Strait and have correlated stratigraphic and structural features across it. They conclude that a slight movement of some kilometres, but not exceeding 25 km, may have occurred. Another group have worked in the oceanic basins surrounding Greenland and their conclusion is that there has been 220 km or more of sinistral movement along the Strait.

The solution to the Nares Strait conflict appears to be clear-cut. We must determine whether displacement was a) less than 25 km or b) more than 220 km. There have been no serious suggestions about sinistral motion between these two extremes.

The origin of Nares Strait has enormous implications for tectonics. The Strait is an important element in tectonics of the northern hemisphere, and it is a key to the pre-drift position of Greenland. It is thereby an important feature in the reconstruction of the North Atlantic Ocean as well as the Arctic Ocean. If there has been hundreds of kilometres of sinistral displacement of Greenland along the Strait, then much support is given to the conventional plate tectonic theory which is widely accepted today. If displacement was minor, or non-existent, then plate tectonic theory is faced with a major problem: what are the origins of Baffin Bay and Labrador Sea?

J. W. Kerr, Department of Energy, Mines and Resourc(!s. Geological Survey of Canada, Institute of Sedimelllary and Petroleum Geology, 3303-33rd St. N. W., Calgary, Alberta, Canada, T2L 2A7. Present address: J. William Kerr & Associates Ltd., Consulting Geologists, Suite 407, 630-Bth Ave. S. W., Calgary, Alberta, Canada, T2P IG6.

"Science should always try to follow the path of least astonishment" - Manik Talwani, 1981.

History of the conflict

Nares Strait (Fig. 1) has been a battleground for tectonic theories for more than 70 years. During this time the pendulum of scientific opinion concerning displacement along the Strait has swung back and forth. The history of discussion on the Strait is divided into five periods, each represented by a consensus on one or the other extreme, or by a conflict. It is interesting to note that the swings in opinion on Nares Strait have coincided with swings in popularity of global theories about horizontal displacements generally.

1. Original consensus

The earliest geological work along Nares Strait was done by the 19th century explorers, who were concerned primarily with geographical exploration (see Dawes & Christie, this volume). The Strait was shown on early geological maps (e.g. Feilden & De Rance 1878), but n mention was made of horizontal displacements. Early authors presumably took it for granted that the lands adjacent to the Strait were fixed geographically. This was part of an original consensus about the Earth, as yet unchallenged.

2. Continental drift emerges

The first published suggestion of great displacement along Nares Strait was made by F. B. Taylor (1910). He considered that Greenland had been fixed in position and that the surrounding lands had moved away from it. Alfred Wegener expanded Taylor's ideas in his theory of continental drift and publicised the theory widely (Fig. 2b) (Wegener 1915, 1922, 1924). Movement of Greenland along Nares Strait was a very important element of Wegener's theory.

The theory of continental drift was not accepted by the majority of earth scientists. Nevertheless, it was a serious challenge to conventional theories and had such

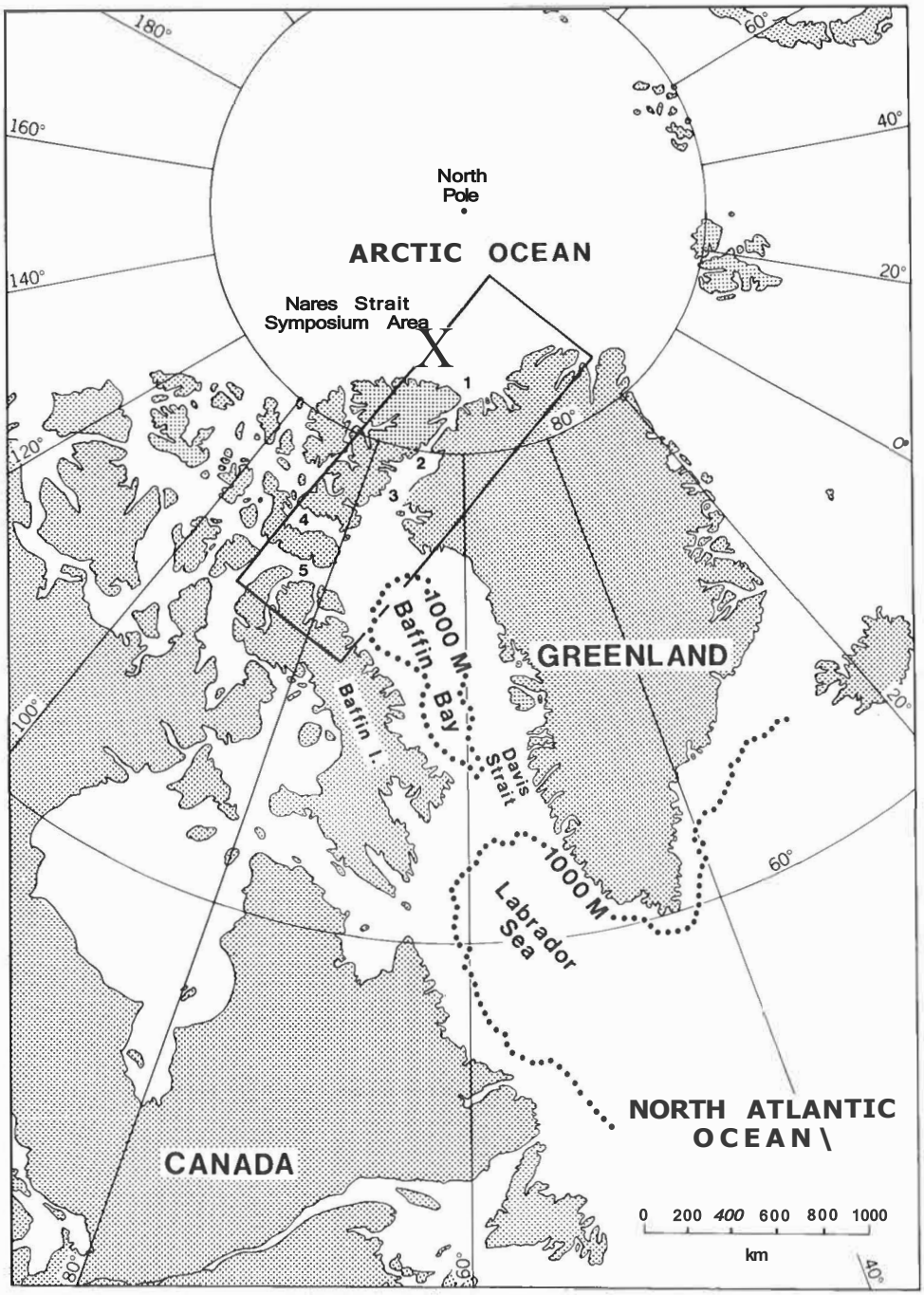


Fig. I. Present geography of the Nares Strait area. The amount of horizontal displacement that occurred in the Strait or nearby has a major bearing on the movement of Greenland and thereby on the origin of the North Atlantic and Arctic Oceans. I = Lincoln Sea, 2 = Kane Basin, 3 = Smith Sound, 4 = Jones Sound, 5 = Lancaster Sound. Modified from Kerr (1980a).

a large following that two symposia were held to discuss it. The British Association for the Advancement of Science discussed the theory in London (Wright 1923) and the American Association of Petroleum Geologists discussed it in Tulsa (van Waterschoot van der Gracht et al. 1928). At both symposia the theory was strongly criticised.

3. Drift discredited and ignored

Continental drift received considerable acceptance in the southern hemisphere, where du Toit (1937) supported it. Most geologists and geophysicists in the northern hemisphere, however, regarded the theory as outlandish and attacked it (Jeffreys 1929). Those papers and symposia in the 1920s that attacked it convinced the scientific communities in North America and Europe that Wegener's theory had been discredited and could be regarded as dead. For the next 30 years, geological papers rarely mentioned the subject. It is presumed that their authors considered that drift had not occurred. However, there were exceptions to this: for example Hilgenberg (1948) and Wegmann (1948).

While Wegener's global theory of continental drift was being discredited at international symposia, his more local theory of displacement along Nares Strait was being discredited on a more local level. Lauge Koch had worked on the entire coast of Greenland adjacent to Nares Strait, and his early work (Koch 1920) had been quoted by Wegener as supporting strike-slip displacement there. Koch in fact was opposed to drift along Nares Strait and considered that Wegener had misused his observations (Koch 1929). Koch suggested that Nares Strait could have formed by the erosion of a "Grabensenkung", related to Cretaceous-Tertiary faulting farther south, around Baffin Bay (see Dawes & Christie, this volume: fig. 14). Critics of Wegener curtailed discussion of continental drift globally. In the same way Koch's opposition curtailed discussion of drift along Nares Strait for an equal period of time.

After Koch's (1929) paper opposing it, the suggestion of drift along Nares Strait lay dormant for about 30 years. Many geologists visited Nares Strait and published on the adjacent coasts of Greenland and Canada (Wordie 1938, Bentham 1941, Munck 1941, Troelsen 1950, Kurtz & Wales 1951, Kurtz et al. 1952, Prest 1952, Davies et al. 1959, Cowie 1961, Nichols 1961, Christie 1962a, b, c, Davies et al. 1963, Thorsteinsson 1963, Fernald & Horowitz 1964). None of these papers mentioned Wegener or continental drift. One can only assume that the authors adhered to the unwritten assumption that there had not been strike-slip displacement along Nares Strait.

It is truly remarkable that continental drift generally and drift along Nares Strait in particular could have received so little mention for about three decades. It appears that the two related phenomena were regarded as so obviously impossible that they were not seriously considered.

4. Continental drift revived

The idea of major displacement along Nares Strait was rejuvenated by S. W. Carey (1958, see Fig. 2c). This was part of his major effort to rejuvenate the theory of continental drift world-wide. Again, as with Wegener, drift of Greenland along Nares Strait was an important element of the global theory. Initial reaction to Carey's idea was unfavourable, but within a few years others modified his theory and the opposition was overcome.

The suggestion of great drift along Nares Strait suddenly became widely accepted. This occurred at about the same time and rate that continental drift became accepted globally. Wilson's papers (1963a, b, 1965a, b) were a major turning point for both the global and the local ideas, and after those papers acceptance of both ideas increased rapidly. By the mid-1970s, large displacements of continents were almost universally accepted as part of the newly named theory of plate tectonics.

The change in scientific opinion accepting great displacements globally was so rapid and complete that it has been called a scientific revolution (Hallam 1973). Acceptance of great displacement along Nares Strait occurred with equal suddenness. It was a small revolution, just as plate tectonics had been.

Since the revival of continental drift many different map reconstructions have been published, showing Greenland moved back hundreds of kilometres and Baffin Bay closed (e.g. Carey 1958, King 1958, Bullard et al. 1965, Holmes 1965, Wilson 1965a, Hilgenberg 1966, Harland 1969, Tarling & Tarling 1971, Dewey 1972, Keen et al. 1972, Bridgwater et al. 1973, Churkin 1973, Hall 1973, Harland 1973, Hyndman 1973, Martin 1973, Roy 1973, Tailleur 1973, Herron et al. 1974, Lambert 1974, Vogt & Avery 1974, Beh 1975, Hallam 1975, Irving 1977, Newman 1977, Sclater et al. 1977, Windley 1977, Johnson 1978, Johnson et al. 1978, Srivastava 1978, Sweeney et al. 1978, Feden et al. 1979, Irving 1979, Churkin & Trexler 1981, Kent 1981, McWhae 1981, Miall 1981, Srivastava et al. 1981). A few of these reconstructions are shown in Fig. 2. In addition to the above papers in which actual map reconstructions were made, other papers, too numerous to mention here, specifically stated or assumed in passing that Greenland has drifted hundreds of kilometres along Nares Strait.

It is interesting to note that during the period of revival and acceptance of great displacement along Nares Strait there was only one new piece of evidence after Wegener's paper (1924) from the area of the actual Strait that was quoted to support strike-slip displacement. This was a linear fault of unspecified displacement but trending parallel to the Strait on Ellesmere Island, reported by Prest (1952) and Christie (1962a), and quoted first by Wilson (1963a, see this volume, Dawes & Christie: fig. 10, Mayr & de Vries: fig. 3). Although this evidence was supportive of transcurrent movement, Christie (1964) as the only geologist to have mapped the fault and its environs at that time, concluded that large-scale strike-slip displacement along that fault had not occurred (Christie 1967). The acceptance of great displacement along Nares Strait in the many reconstructions listed above was mainly because it

fit with plate tectonic theory, and indeed was predicted by such theory.

When Wilson {1963a) supported the idea of great displacement along Nares Strait there was no compelling evidence to deny it, and in fact the general geology of the two coasts superficially appeared to support it, as initially noted by Wegener (1915). Soon thereafter strong contrary evidence was published, as discussed later (Kerr 1967a); however, that evidence was at first disputed and later disregarded. Few of the papers that made reconstructions of Nares Strait quoted evidence either for or against displacement there. Those that did quote evidence from the Strait itself normally quoted Wilson {1963a), or one another. In essence, then, the near universal acceptance of great strike-slip along Nares Strait occurred mainly because of the momentum of plate tectonic theory, which predicted that there ***must have been*** such displacement.

5. A challenge and a conflict

An interpretation of Nares Strait arguing that there was only minor displacement was proposed by this author **(Kerr** 1967a), using geological evidence from the actual Strait region. **It** suggested that there had been little strike-slip along the Strait, several kilometres at most. For a few years authors regarded that interpretation as incorrect, and many failed even to refer to it. Kerr's view {1967a) and apparently also the evidence were overshadowed by the great swing in opinion which was part of the momentum of the plate tectonic revolution.

Kerr's {1967a) paper was based mainly on his first-hand knowledge of the Canadian side of Nares Strait and comparison with the geology of the Greenland side. Some areas of the Greenland side had not been investigated since Koch's work in the early 1920s, and Kerr reinterpreted some of that work. However, most geologists concluded that while it was certainly not necessary to invoke any strike-slip motion to explain the regional geology of Nares Strait, detailed geological knowledge at that time, particularly of the Greenland side, was not sufficient to conclusively rule out appreciable transcurrent displacement of Greenland {Trettin et al. 1972, Dawes 1973).

Geological investigation on the Greenland side of Nares Strait gathered momentum in the 1970s and Koch's work was superseded. Gradually, the view that there was little or no strike-slip displacement along the Strait was confirmed by geological studies in the region (e.g. Christie et al. 1978, Frisch et al. 1978, Dawes 1979, Christie et al. 1981, Dawes & Peel 1981) as well as by some geophysical studies (e.g. Riddihough et al. 1973, Sobczak & Stephens 1974, Wetmiller 1974, Coles et al. 1976, Grant 1980). In contrast to this trend some authors still maintained that the geological data from the Strait (e.g. Newman 1977, Newman & Falconer 1978, Srivastava 1978) and from the general region (e.g. McWhae 1981, Miall 1981) indicated major strike-slip displacement.

The suggestion that drift along Nares Strait had not been great was a major challenge to conventional plate tectonic theories of that region. **It** has resulted in a conflict fundamental to plate tectonic studies of the Arctic.

Interpretations of Nares Strait

Interpretations of displacement along Nares Strait fall into three groups. The popularity of each interpretation has risen and fallen at various times. Since the conflict began {Taylor 1910), none of these has ever been universally accepted nor has any been universally rejected.

In recent years, since plate tectonic theory and associated major horizontal displacements have become widely accepted, many scientists have speculated on displacement in Nares Strait. A major conflict has developed, with opinions on the nature of the Strait being of three types. Those who adhere to conventional plate tectonic theories have required a great left-lateral fault along Nares Strait. Those with fixist views consider that there have been no plate movements in the Arctic, and implicitly suggest there was no lateral displacement along the Strait. The third group used evidence from the shores of the Strait to show that there has been little or no lateral displacement. These three interpretations arc outlined below.

Fixist theories

Fixists have largely stayed out of the Nares Strait conflict, for no one who has actually studied or written about Nares Strait recently appears to conclude that displacement there was nil. The debate has largely polarised between the other two camps, both of which accept plate tectonics in one form or another. Two notable opponents of hypotheses concerning horizontal displacements of continental blocks, and adherents of fixist theories, are A. A. Meyerhoff and V. V. Bcloussov. Meyerhoff (1970, 1973) favoured a fixist model for the Arctic and North Atlantic Oceans and suggested that continents there have occupied their present locations since Proterozoic time (Fig. 2a). A school of Russian geologists headed by V. V. Beloussov (1970, 1979) maintain that continents have not moved relative to each other. These fixist authors have not scrutinied Nares Strait particularly nor emphasised it in their papers, but fixist views implicitly suggest that plate movements have not occurred and therefore there has been no displacement along the Strait.

Conventional plate tectonic theories

Plate tectonics is now very widely accepted by earth scientists. This theory is the successor of a rejuvenated

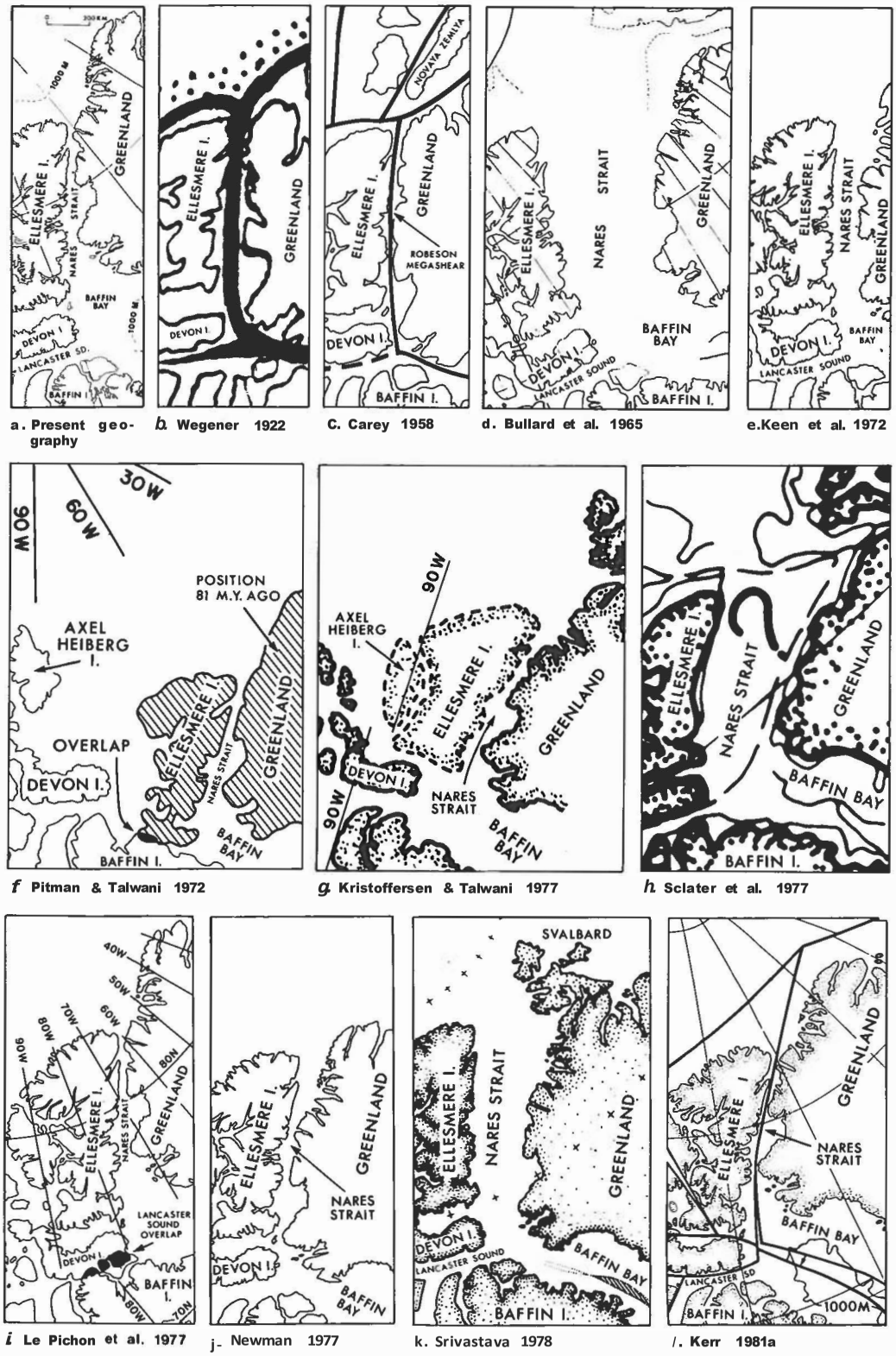


Fig. 2. The Nares Strait region and some reconstructions that have been suggested: a, present geography, representing fixist theories; b, strike-slip displacement of about 350 km (Wegener 1915, 1922); c, strike-slip displacement of about 400 km (Carey 1958); d, oblique displacement (Bullard et al. 1965); e, strike-slip displacement of about 220 km (Keen e, al. 1972); f, no displacement, but great oblique displacement farther west (Pitman & Talwani 1972); g, about 100 km of transverse displacement taken up farther west, combined with about 90 km of strike-slip oblique to Nares Strait (Kristoffersen & Talwani 1977); h, great oblique displacement (Sclater et al. 1977); i, no displacement along the Strait, but movement northward and westward of a larger block that includes Ellesmere and Devon Islands (Le Pichon et al. 1977); j, left-lateral displacement of 250 km (Newman 1977); k, an oblique approach of Greenland toward Ellesmere Island (Srivastava 1978, Srivastava et al. 1981), involving two phases, the first compressive (oblique, 50 to 220 km), and the second left-lateral (250 km);/, minor strike-slip displacement, minor rotational opening and major foundering (Kerr 1967a, b, 1981a). Slightly modified from Kerr (1980a).

theory of continental drift which burst upon the scene in the late 1950s, went through variations called the new global tectonics, sea-floor spreading, and then finally reached its present understanding as plate tectonics. In all, this was a very revolutionary process and has been called "a revolution in the earth sciences" (Hallam 1973). Now, however, the theory of plate tectonics is so widely accepted that the revolution is over and the theory has become quite conventional. *Its opponents are now the revolutionaries.* Certainly, there are differences between the many interpretations of plate motions governed by conventional plate tectonic theory, but these are minor compared to the overwhelming agreement and uniformity amongst them. The conventional group includes nearly all those writing on plate tectonics, and many of those who made reconstructions involving Nares Strait have been quoted earlier.

Conventional plate tectonic theory requires that there has been a great left-lateral fault along Nares Strait in order to explain the oceanic crust in Baffin Bay by a mechanism that is acceptable to that theory. The adherents of this conventional scenario looked at the global situation first and reached a conclusion on how ocean basins formed. They looked at Nares Strait later and attempted to explain the Strait in terms that were compatible with the conclusions reached earlier about the ocean basins.

Nares Strait is an important element of conventional plate tectonic theories (Dewey 1972). The usual reconstruction of Nares Strait is that of Wilson (1965a) which restores about 350 km of left-lateral offset. Wilson (1965b) introduced the concept of transform faults and suggested that one existed along Nares Strait.

Integrated plate tectonic theory

This theory, put forth by Kerr (1967a, b) and elaborated upon later (Kerr 1980b, c, 1981 a, b) began as an attempt to find a solution to the Nares Strait conflict. Kerr (1967a) looked first at Nares Strait itself and reached a conclusion on how it formed, after which he studied the nearby oceanic basins and attempted to explain them in terms that were compatible with that conclusion. His theory concluded that some displacement has occurred along Nares Strait; it probably was only several kilometres, and almost certainly less than 25 km (Kerr 1981a), which is the amount shown in Fig. 2/.

This theory is called an integrated theory because it integrates and is a compromise between two opposing views - fixism and conventional plate tectonics. The suggestion in this theory that there was only small displacement along Nares Strait is incompatible with the origins of Baffin Bay and Labrador Sea that are generally accepted by conventional plate tectonic theory. One of these theories on Nares Strait, Baffin Bay and the Labrador Sea must be wrong.

The integrated plate tectonic theory suggested that Baffin Bay and Labrador Sea had much less lateral spread than conventional plate tectonic theories have suggested. Nevertheless, these basins still formed by a mechanism which is a form of plate tectonics. The new theory does not in any way intend to overthrow plate tectonics, which it regards as in many ways correct. Instead it considers that plate tectonics and fixist theories can be integrated and reconciled, and Nares Strait is the place that shows this is possible.

Implications of drift along Nares Strait

The implications of drift along Nares Strait have been discussed earlier (Kerr 1980a) and are briefly reviewed here. Nares Strait is the key to the pre-drift position of Greenland, and thereby is an important feature in reconstruction of the North Atlantic Ocean, as well as the Arctic Ocean (Fig. 1). If indeed there have been hundreds of kilometres of displacement of Greenland along Nares Strait or in that region, as many authors contend (Fig. 2b-k), then a pre-drift reconstruction brings Greenland back by juxtaposing the continental shelf edges represented approximately by the 1000 m lines (Fig. 1). If Nares Strait or the nearby region had such great displacement and the conventional restoration of Greenland is possible, then further support is given to the conventional plate tectonic theory that is widely accepted today. Thus Baffin Bay and Labrador Sea can he readily explained as they have been in the past (Pitman & Talwani 1972, Srivastava 1978, Srivastava et al. 1981), viz. as oceanic basins lying between continental blocks that moved apart and were formerly joined at the continental shelf edges or approximate 1000 m lines.

If Greenland has not moved along Nares Strait at all as fixists maintain (Fig. 2a), then conventional plate tectonic theory is faced with a major problem: what arc the origins of Baffin Bay and Labrador Sea?

If Greenland moved only a short distance along Nares Strait, however, and there has been slight rotational separation only, as suggested by Kerr (Fig. 2/, 1967a, b, 1981a, b), plate tectonic theory is still faced with a major problem concerning Baffin Bay and Labrador Sea. Baffin Bay is quite wide, with the 1000 m lines in most places about 250 km apart (Fig. I). It contains oceanic crust with magnetic striping (Keen et al. 1972. 1974, Pelletier et al. 1974, Srivastava 1978, Jackson et al. 1979, Srivastava et al. 1981). Yet, without great strike-slip in Nares Strait, Baffin Bay cannot be closed in a reconstruction in which the opposing 1000 m lines are brought adjacent to each other. A small rotation back in Nares Strait produces only minor closing of Baffin Bay. Labrador Sea is even wider, with the 1000 m lines in most places about 600 km apart. Labrador Sea also contains a very wide region of oceanic crust with magnetic striping (Hyndman 1973, Hood & Bower 1975, Kristoffersen & Talwani 1977, Srivastava 1978,

Srivastava et al. 1981). A small rotation of Nares Strait back to a former position produces moderate closure of Labrador Sea, but still leaves the shelf edges or 1000 m lines several hundred kilometres apart.

If Baffin Bay and Labrador Sea did not form through sea-floor spreading, the continental shelf edges cannot be brought back together in the manner suggested by the conventional plate tectonic theory (Wilson 1963a, 1965a, b, Dewey 1972, Sclater et al. 1977). Thus, how *did* these seaways form? The alternatives are 1) they are original seaways that existed before the suggested plate tectonic episode, in accordance with fixist model of Meyerhoff (1970, 1973); 2) they formed by foundering and oceanisation of continental crust with no lateral separation, in accordance with fixist model of Belousov (1970, 1979); or 3) they formed by widespread stretching, thinning, and foundering of continental crust, with only slight to moderate lateral separation of plates by rotation, as in the model put forth by Kerr (1967b, 1981a).

Grant (1975) suggested that in the Davis Strait area the sea-floor may be underlain mainly by continental crust. In the case of Labrador Sea, subsided continental material may extend far from shore as suggested by van der Linden (1975, 1977), Umpleby (1979), and Grant (1980). In addition, there may be a remnant of continental crust beneath the dormant mid-Labrador Sea ridge structure as originally suggested by Kerr (1967b). Kerr's model involves a combination of plate tectonics and oceanisation; a small lateral separation (plate tectonics) combined with widespread thinning and foundering (and oceanisation) of a large segment of intervening continental crust that now lies beneath the continental shelves and axis of Labrador Sea. Such originally continental material, now thinned and altered, may also underlie much of Baffin Bay.

If it can be proven that Greenland did drift 220 to 400 km along Nares Strait, then conventional plate tectonic theory will have successfully overcome another major challenge. If Greenland moved an order of magnitude less or not at all, then one of the three alternatives outlined above is correct in this case and conventional plate tectonic theory needs major modification because of problems related to the formation of Baffin Bay and Labrador Sea.

Labrador Sea, a branch of the North Atlantic Ocean, is a deep oceanic basin with depths exceeding 3500 m over a wide area (Fig. 1). It has a well expressed axial zone that may be a dormant mid-ocean ridge (Wilson 1963b, Kerr 1967b, Hyndman 1973, Kristoffersen & Talwani 1977, Srivastava 1978, Srivastava et al. 1981). If the conventional plate tectonic origin of Labrador Sea is incorrect then the widely accepted origin of much of the North Atlantic Ocean is also questionable. Thus, the Nares Strait conflict has far-reaching implications for tectonics, particularly for reconstructions of the North Atlantic Ocean north of lat. 60°N, where that ocean is rather narrow (Fig. 1).

Labrador Sea and Baffin Bay and associated parts of the Nares Strait region (Lancaster Sound, Jones Sound, Smith Sound, Kane Basin and Lincoln Sea) contain thick sedimentary successions. These are of promising hydrocarbon potential and many areas already are being actively explored. Thus, the Nares Strait conflict also has very important economic implications (McMillan 1973, this volume).

Summary of the Nares Strait conflict

The Nares Strait conflict can be summarised very briefly and simply. It is epitomised by Figs 3 and 4, and the conflict can be visualised by examining them together.

One group of scientists have examined the stratigraphy and structures on the shores of the Strait in detail. They conclude that the evidence from correlations across it indicates that displacement along it has been minor, probably close to zero or less than several kilometres and certainly less than 25 km (Kerr 1967a, 1981a, Christie et al. 1981). The main proponent of this view has then looked at the adjacent oceanic basins (Kerr 1967b, 1981a, b) and interpreted them in a way that does not conflict with his ideas on Nares Strait.

Several geological features which continue across the Strait without apparent offset were described by Kerr (1967a), the most accurately defined of which at that time were Silurian facies belts (Fig. 3). As geological knowledge of the region increased in the last decade, many additional features have been recognised, which range in age from Precambrian to Cenozoic. These so-called markers are described in several papers in the volume and they are considered to confirm that any displacement was minor. The case for minor displacement along Nares Strait relies mainly on evidence such as the correlation illustrated by Kerr (1967a) and reproduced here as Fig. 3. If there were great displacement along the Strait, then correlations such as these must have a different explanation. There have been no serious attempts by proponents of great displacement along the Strait to present alternative correlations.

Another group interpreted from studies of the oceanic areas surrounding Greenland that there has been great displacement along Nares Strait. The strongest datum to support this conclusion is the existence of oceanic crust with magnetic striping in Baffin Bay and Labrador Sea. The striping has been shown by Srivastava et al. (1981) and Srivastava & Falconer (this volume) and is reproduced here as Fig. 4.

Magnetic anomalies are youngest along the axes of mid-oceanic ridges and they become increasingly older towards the coasts. The anomalies in Baffin Bay and Labrador Sea were presumed to have formed successively as Greenland drifted physically north-eastward, and as new oceanic crust formed in its wake (Srivastava 1978, Srivastava et al. 1981). Before the onset of plate

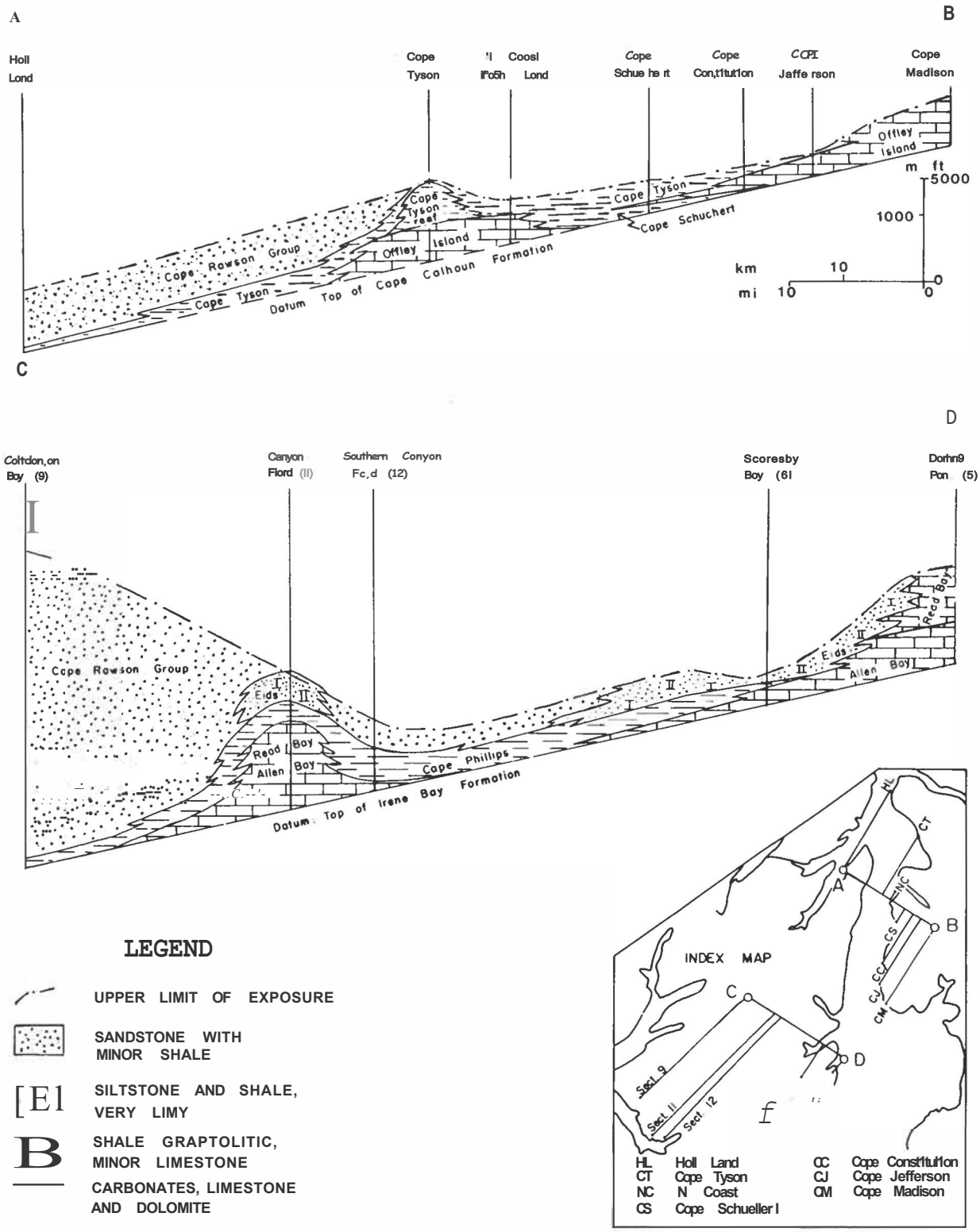


Fig. 3. On-strike correlation of Lower Palaeozoic facies of the Ellesmere Island side of Nares Strait with rocks of the same age on Greenland. Correlations such as these are the main type of evidence in the case against major displacement along Nares Strait. From Kerr (1967a).

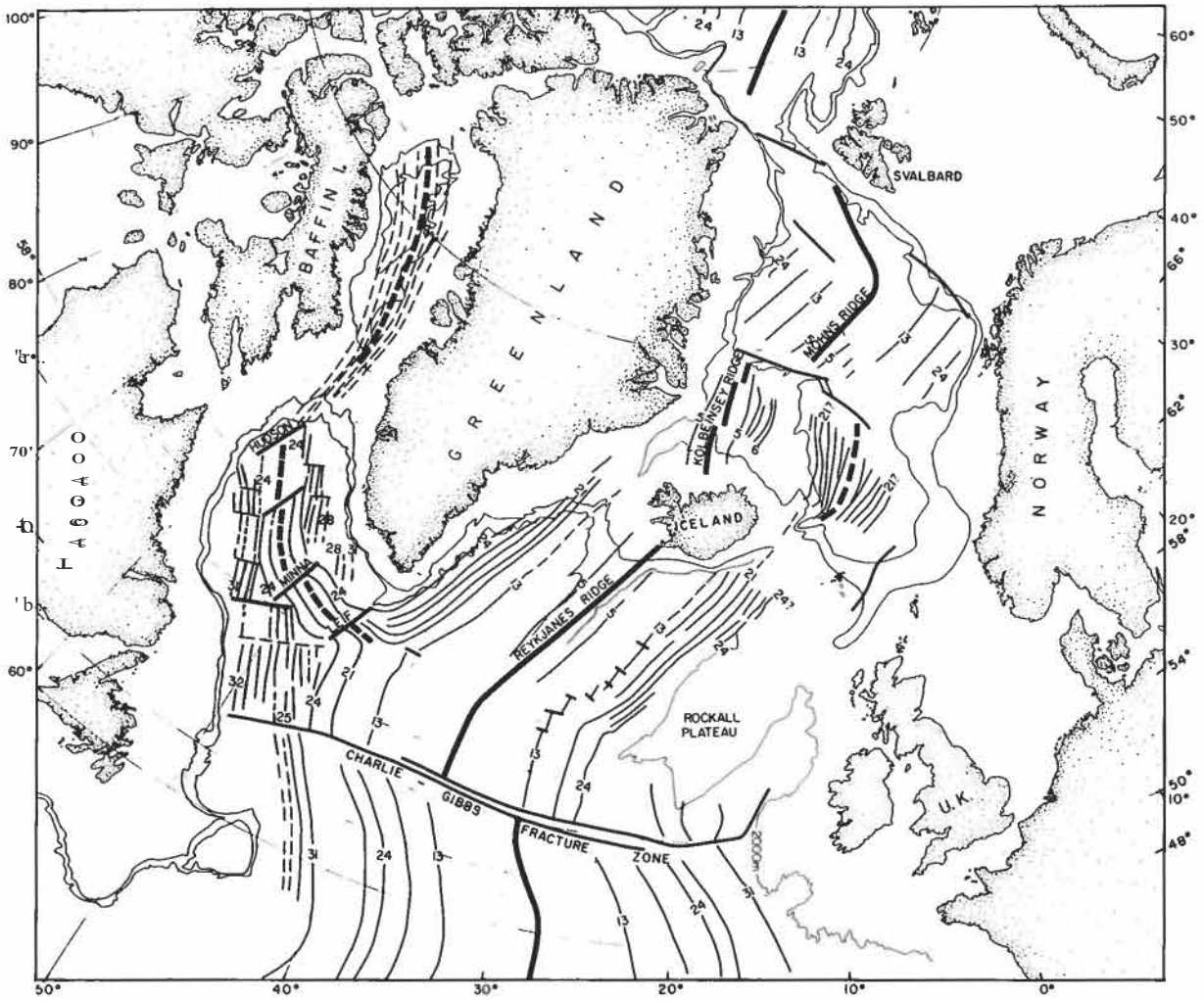


Fig. 4. Magnetic patterns in the oceanic basins surrounding Greenland. The presence of oceanic crust and the arrangement of magnetic striping in Baffin Bay and the Labrador Sea have been the main evidence in the case for major displacement along Nares Strait. From Srivastava et al. (1981). The interpretations that have been made of this Fig. and Fig. 3 are in conflict with each other, and they epitomise the Nard Strait debate.

movement, Greenland was farther south-west, such that the oldest anomaly on the Greenland side of Baffin Bay lay immediately adjacent to an anomaly of similar age near Baffin Island. Likewise, the oldest anomalies in the Labrador Sea (anomaly 31) are near Labrador and Greenland, and these also were originally adjacent. After reaching the conclusion that Greenland moved hundreds of kilometres away from Baffin Island and Labrador, it was logical to interpret that this motion was accommodated by strike-slip along Nares Strait (Keen et al. 1972, Srivastava 1978, Srivastava et al. 1981).

These workers, who use conventional plate tectonic theory, have studied the ocean basins and reached a conclusion on their origin. Nares Strait has been a secondary topic and has been interpreted by them in a way that does not conflict with their ideas on the oceans.

Both approaches outlined above are similar. The authors first made interpretations of their primary study area, after which they proceeded to make an interpretation of the area they had studied less well. A conflict between the two groups resulted. Solution of the conflict was the purpose of the Nares Strait symposium and of this volume.

There is an onus on proponents of minor displacement along Nares Strait to give a reasonable interpretation of the crustal structure and magnetic striping of Baffin Bay and Labrador Sea which is compatible with the observations reported there. This has been attempted by the integrated plate tectonic theory (Kerr 196 7b, 1981 a, b). There is an equal onus on proponents of great displacement along Nares Strait to give a reasonable alternative explanation of the stratigraphic and

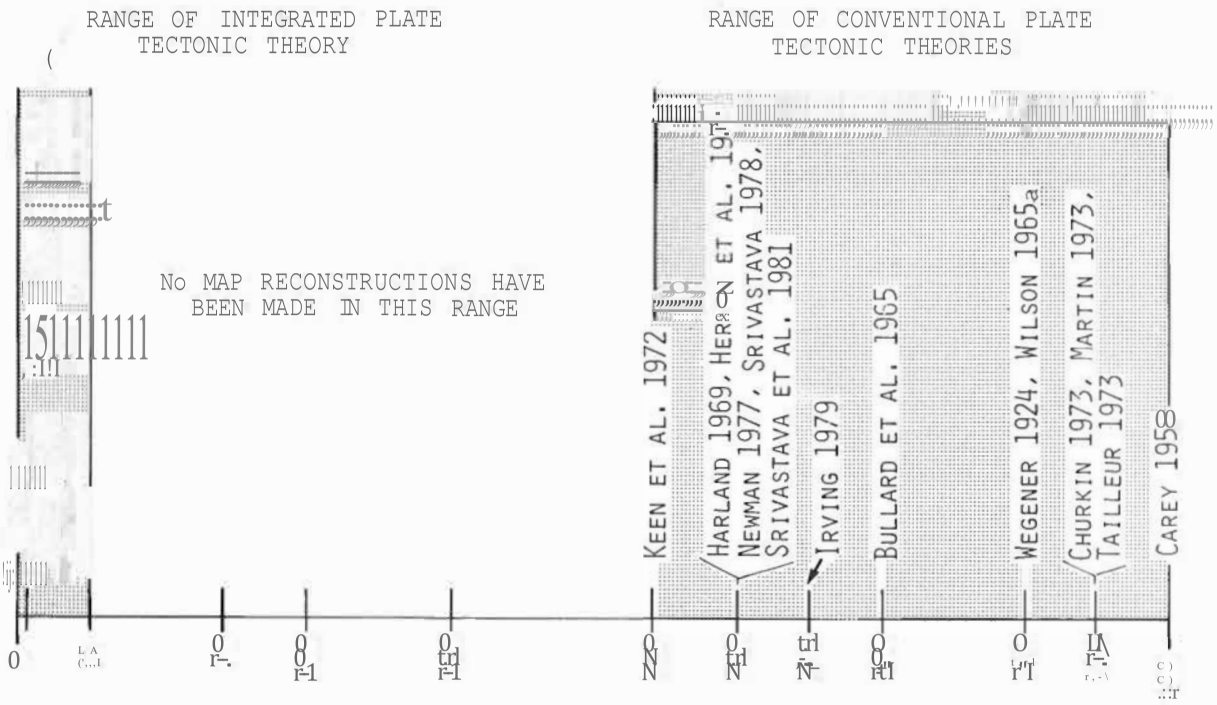


Fig. 5. The sinistral lateral displacements that have been suggested along Nares Strait in map reconstructions. The Nares Strait conflict may be solved by determining in which of the two shaded ranges the actual strike-slip displacement can be placed.

structural correlations that have been suggested across Nares Strait.

Solution to the conflict

It is likely that movement along Nares Strait can never be measured precisely. Even without a precise measure of movement, however, the conflict can still be resolved. This will be done if earth scientists can simply agree upon which of two distance ranges sinistral strike-slip displacement falls in. The displacements that have been suggested along Nares Strait are shown in Fig. 5.

The range on the right, 220 to 400 km, includes nearly all conventional plate tectonic reconstructions. The range on the left, less than 25 km, includes the integrated plate. tectonic theory. There is an enormous gap between the two ranges, extending from 25 to 220 km.

Only a few papers have suggested that displacement occurred in the intermediate range. Keen et al. (1972) suggested in their text that displacement along the Strait was 150 km; however, their map reconstruction showed about 220 km of restoration. Pitman & Talwani (1972) showed no displacement on their map reconstruction. They moved Ellesmere Island with Greenland and obtained an otherwise conventional reconstruction of Baffin Bay and Labrador Sea. Le Pichon et al. (1977) suggested in their text that displacement might have been a maximum of 100 km; however, they rekm.:cl to Kerr's (1967a) paper and showed no displacement on their map reconstruction. Kristoffersen & Tai\vani (1977) also considered that Greenland and Ellesmere Island had travelled together, with the displacement being largely taken up farther west. They considered that those lands reached their present positions in two stages, first travelling about 100 km in a direction normal to the Strait, and later about 90 km oblique to the Strait with some left-lateral displacement. However, the latter three papers suggesting intermediate displacement, introduced displacements farther west in the Arctic which are considered unreasonable. When these three papers and that of Keen et al. (1972) arc eliminated because the displacements discussed were nnt actually used on the maps, or because they rely on unreasonable displacement elsewhere, the range 25 to 220 km is empty (Fig. 5). The reason for this is that mov..:-ment in that particular range does not provide an explanation which is satisfactory to either side of the conflict.

Because there are no serious suggestions of displacement between 25 and 220 km, the views on the Strait are polarised and the solution to the Nares Strait conflict is greatly simplified. The debate should concentrate only on determining in which of the two ranges the displacement along Nares Strait falls, less than 25 km or more than 220 km. The papers in this volume should focus on that objective.

Summary

The Nares Strait conflict should not be regarded as a negative thing in any way. Regardless of the outcome it will have a constructive impact on science. It has forced all of us to look at our own data far more critically than we otherwise would have done. It has inspired communication between colleagues widely separated by geography or discipline in an attempt to work together to solve a fascinating problem. Surely the final outcome, regardless of who is right, will be a better understanding of the history of the Earth.

Acknowledgements

I thank the Geological Survey of Canada for many years of support of my work on Nares Strait and related problems when I was a member of that organisation. Much assistance with the preparation of this paper was provided by P. R. Dawes and E. Glenda(for which I am grateful.

References

Beh, R. L. 1975. Evolution and geology of western Baffin Bay and Davis Strait, Canada. - In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. - Mem. Can. Soc. Petrol. Geol. 4: 453-476.

Beloussov, V. V. 1970. Against the hypothesis of ocean floor spreading. - Tectonophysics 9: 489-511.

Beloussov, V. V. 1979. Why I do not accept plate tectonics. - Trans. Am. geophys. Un. 60: 207-21 I.

Bentham, 1941. Structure and glaciers of southern Ellesmere Island. - Geogr. J. 97: 36-45.

Bridgwater, D., Escher, A., Jackson, G. D., Taylor, F. C. & Windley, B. F. 1973. Development of the Precambrian Shield in West Greenland, Labrador, and Baffin Island. - In: Pitcher, M. G. (ed.), Arctic geology. - Mem. Am. Ass. Petrol. Geol. 19: 99-116.

Bullard, E., Everett, J. E. & Smith, A. G. 1965. The fit of the continents around the Atlantic. - In: Blackett, P. M. S., Bullard, E. & Runcorn, S. K., A symposium on continental drift. - Phil. Trans. Roy. Soc. Lond. 258A: 41-51.

Carey, S. W. 1958. The tectonic approach to continental drift. - In: Carey, S. W. (convener), Continental drift. A symposium: 177-355. - Univ. Tasmania, Hobart.

Christie, R. L. 1962a. Northeastern Ellesmere Island, District of Franklin. - Pap. geol. Surv. Can. 62-10: 15 pp.

Christie, R. L. 1962b. Geology, Alexandra Fiord, Ellesmere Island, District of Franklin (map with marginal notes) 1" to 4 miles. - Map. geol. Surv. Can. 9-1962.

Christie, R. L. 1962c. Geology, southeast Ellesmere Island, District of Franklin (map with marginal notes) I" to 4 miles. - Map. geol. Surv. Can. 12-1962.

Christie, R. L. 1964. Geological reconnaissance of northeastern Ellesmere Island, District of Franklin. - Mem. geol. Surv. Can. 331: 79pp.

Christie, R. L. 1967. Operation Grant Land (1966), northern Ellesmere Island: - Pap. geol. Surv. Can. 67-1: 2-3.

Christie, R. L., Dawes, P. R. & Frisch, T. (O.] 1978. Basal strata of the Proterozoic Thule Group on Ellesmere Island and Greenland. - Geol. Soc. Am. Abs. with Prog. 10: 380 only.

Christie, R. L., Dawes, P. R., Frisch, T. (O.], Higgins, A. K., Hurst, J. M., Kerr, J. W. & Peel, J. S. 1981. Geological evidence against major displacement in the Nares Strait. - Nature, Lond. 291: 478-480.

Churkin, M. 1973. Geologic concepts of Arctic Ocean basin. - In: Pitcher, M. G. (ed.), Arctic geology. - Mem. Am. Ass. Petrol. Geol. 19: 485-499.

Churkin, M. & Trexler, J. H. 1981. Continental plates and accreted oceanic terranes in the Arctic. - In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 1-20. - Plenum Press, New York & London.

Coles, R. L., Haines, G. V. & Hannaford, W. 1976. Large scale magnetic anomalies over western Canada and the Arctic: a discussion. - Can. J. Earth Sci. 13: 790-802.

Cowie, J. W. 1961. Contributions to the geology of North Greenland. - Meddr Grnnland 164(3): 47 pp.

Davies, W. E., Needleman, S. M. & Klick, D. W. 1959. Report on Operation Groundhog (1958) North Greenland. Investigations of ice-free sites for aircraft landings, Polaris Promontory, North Greenland. - U.S. Air Force Cambridge Res. Center, Bedford: 45 pp.

Davies, W. E., Krinsley, D. B. & Nicol, A. H. 1963. Geology of the North Star Bugt area, Northwest Greenland. - Meddr Grnnland 162(12): 68 pp.

Dawes, P. R. 1973. The North Greenland fold belt: a clue to the history of the Arctic Ocean basin and the Nares Strait lineament. - In: Tarling, D. H. & Runcorn, S. K. (eds), Implications of continental drift to the earth sciences 2: 925-947. - Academic Press, London & New York.

Dawes, P. R. 1979. Precambrian and Palaeozoic development of northern Greenland. - Norsk Polarinst. Skr. 167: 321-324.

Dawes, P. R. & Christie, R. L. 1982. History of exploration and geology in the Nares Strait region. - This volume.

Dawes, P. R. & Peel, J. S. 1981. The northern margin of Greenland from Baffin Bay to the Greenland Sea. - In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 201-264. - Plenum Press, New York & London.

Dewey, J. F. 1972. Plate tectonics. - Scient. Am. 226(5): 56-68.

Du Toit, A. L. 1937. Our wandering continents: hypothesis of continental drifting. - Oliver and Boyd, Edinburgh: 366 pp.

Feden, R. H., Vogt, P. R. & Fleming, H. S. 1979. Magnetic and bathymetric evidence for the "Yermak hot spot" northwest of Svalbard in the Arctic Basin. - Earth planet Sci. Lett. 44: 18-38.

Feilden, H. W. & De Rance, C. E. 1878. Geology of the coasts of the Arctic lands visited by the late British Expedition under Captain Sir George Nares, R. N., K. C. B., F. R. S. - Q. JI geol. Soc. Lond. 34: 556-567.

Fernald, A. T. & Horowitz, A. S. 1964. Bedrock geology of the Nunatarssuaq area, Northwest Greenland. - Meddr Grnnland 172(6): 44 pp.

Frisch, T., (O.], Morgan, W. C. & Dunning, G. R. 1978. Reconnaissance geology of the Precambrian Shield of Ellesmere and Coburg islands, Canadian Arctic Archipelago. - Pap. geol. Surv. Can. 78-1A: 135-138.

Grant, A. C. 1975. Geophysical results from the continental margin off southern Baffin Island. - In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. - Mem. Can. Soc. Petrol. Geol. 4: 411-431.

Grant, A. C. 1980. Problems with plate tectonics: the Labrador Sea. - Bull. Can. Petrol. Geol. 28: 252-278.

Hall, J. K. 1973. Geophysical evidence for ancient sea-floor spreading from Alpha Cordillera and Mendeleyev Ridge. - In: Pitcher, M. G. (ed.), Arctic geology. - Mem. Am. Ass. Petrol. Geol. 19: 542-561.

Hallam, A. 1973. A revolution in the earth sciences from continental drift to plate tectonics. - Clarendon Press, Oxford: 127 pp.

Hallam, A. 1975. Alfred Wegener and the hypothesis of continental drift. - Scient. Am. 232(2): 88-97.

Harland, W. B. 1969. Contribution of Spitsbergen to understanding of tectonic evolution of North Atlantic region. - In: Kay, M. (ed.), North Atlantic - geology and continental drift, a symposium. - Mem. Am. Ass. Petrol. Geol. 12: 817-851.

Harland, W. B. 1973. Tectonic evolution of the Barents Shelf and related plates. - In: Pitcher, M. G. (ed.), Arctic geology. - Mem. Am. Ass. Petrol. Geol. 19: 599-608.

Herron, E. M., Dewey, J. F. & Pitman, W. C. 1974. Plate tectonic model for the evolution of the Arctic. - Geology 2: 377-380.

Hilgenberg, 0. C. 1948. Die Bruchstruktur der Erdrinde, insbesondere von Gronland, verglichen mit dem Schwachenetz von sproden Prilfkorpern. - Diss. Berlin Tech. Univ.: 106 pp.

Hilgenberg, 0. C. 1966. Bestatigung der Kennedy-Channel-Scherung durch die Bruchstruktur von Gronland und Nordost-Kanada. - Geotekt. Forsch. 22: 1-74.

Holmes, A. 1965. Principles of physical geology. - Thomas Nelson and Sons Ltd., London: 1288 pp. (Fully revised edit.).

Hood, P. [J.] & Bower, M. [E.] 1975. Aeromagnetic reconnaissance of Davis Strait and adjacent areas. - In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. - Mem. Can. Soc. Petrol. Geol. 4: 433-451.

Hyndman, R. D. 1973. Evolution of the Labrador Sea. - Can. J. Earth Sci. 10: 636-644.

Irving, E. 1977. Drift of major continental blocks since the Devonian. - Nature, Lond. 270: 304-309.

Irving, E. 1979. Paleopoles and paleolatitudes of North America and speculations about displaced terrains. - Can. J. Earth Sci. 16: 669-694.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor spreading in central Baffin Bay. - Can. J. Earth Sci. 16: 2122-2135.

Jeffreys, H. 1929. The Earth: its origin, history and physical constitution. - Cambridge Univ. Press, U.K.: 525 pp.

Johnson, G. L., Taylor, P. T., Vogt, P. R. & Sweeney, J. F. 1978. Arctic Basin morphology. - Polarforschung 48(1/2): 20-30.

Johnson, J. A. L. 1978. European plate movement during the Carboniferous. - In: Tarling, D. H. (ed.), Evolution of the Earth's crust: 343-360. - Academic Press, London, New York & San Francisco.

Keen, C. E., Barrett, D. L., Manchester, K. S. & Ross, D. I. 1972. Geophysical studies in Baffin Bay and some tectonic implications. - Can. J. Earth Sci. 9: 239-256.

Keen, C. E., Keen, M. J., Ross, D. I. & Lack, M. 1974. Baffin Bay: small ocean basin formed by sea-floor spreading. - Bull. Am. Ass. Petrol. Geol. 58: 1089-1108.

Kent, P. E. 1981. The history of the northeast Atlantic margin in a world setting. - In: Kerr, J. W. & Fergusson, A. J. (eds), Geology of the North Atlantic borderlands. - Mem. Can. Soc. Petrol. Geol. 7: 1-10.

Kerr, J. W. 1967a. Nares submarine rift valley and the relative rotation of north Greenland. - Bull. Can. Petrol. Geol. 15: 483-520.

Kerr, J. W. 1967b. A submerged continental remnant beneath the Labrador Sea. - Earth planet. Sci. Lett. 2: 283-289.

Kerr, J. W. 1980a. Did Greenland drift along Nares Strait? - Bull. Can. Petrol. Geol. 24: 279-289.

Kerr, J. W. 1980b. A plate tectonic contest in Arctic Canada. - In: Strangway, D. W. (ed.), Crust of the Earth and its mineral deposits. - Spee. Pap. geol. Ass. Can. 20: 457-486.

Kerr, J. W. 1980c. Structural framework of Lancaster aulacogen, Arctic Canada. - Bull. geol. Surv. Can. 319: 24 pp.

Kerr, J. W. 1981 a. Evolution of the Canadian Arctic Islands: a transition between the Atlantic and Arctic Oceans. - In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 105-199. - Plenum Press, New York & London.

Kerr, J. W. 1981b. Stretching of the North American plate by a now dormant Atlantic spreading centre. - In: Kerr, J. W. & Fergusson, A. J. (eds), Geology of the North Atlantic borderlands. - Mem. Can. Soc. Petrol. Geol. 7: 245-278.

King, L. 1958. A new reconstruction of Laurasia. - In: Carey, S. W. (convener), Continental drift. A symposium: 13-23. - Univ. Tasmania, Hobart.

Koch, L. 1920. Stratigraphy of Northwest Greenland. - Meddr dansk geol. Foren. 5(17): 78 pp.

Koch, L. 1929. Stratigraphy of Greenland. - Mcddr Grnnland 73,2(2): 205-320.

Kristoffersen, Y. & Talwani, M. 1977. Extinct triple junction south of Greenland and the Tertiary motion of Greenland relative to North America. - Bull. geol. Soc. Am. 88: 1037-1049.

Kurtz, V. E. & Wales, D. B. 1951. Geology of the Thule area, Greenland. - Proc. Oklahoma Acad. Sci. 31 : 83-89.

Kurtz, V. E., McNair, A. H. & Wales, D. B. 1952. Stratigraphy of the Dundas Harbour area, Devon Island, Arctic Archipelago. - Am. J. Sci. 250: 636-655.

Lambert, R. St. J. 1974. Global tectonics and the Canadian Arctic continental shelf. - In: Aitken, J. D. & Glass, D. J. (eds), Geology of the Canadian Arctic: 5-22. - Genl. Ass. Can. & Can. Soc. Petrol. Geol.

Le Pichon, X., Sibuet, J.-C. & Francheteau, J. 1977. The fit of the continents around the North Atlantic Ocean. - Tectonophysics 38: 169-209.

Martin, R. 1973. Cretaceous-Early Tertiary rift basin of Baffin Bay - continental drift without sea-floor spreading. - In: Pitcher, M. G. (ed.), Arctic geology. - Mem. Am. Ass. Petrol. Geol. 19: 500-505.

Mayr, U. & de Vries, C. D. S. 1982. Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago. - This volume.

McMillan, N. J. 1973. Shelves of Labrador Sea, Baffin Bay, Canada. - In: McCrossan, R. G. (ed.), Future petroleum provinces of Canada, their geology and potential. - Mcm. Can. Soc. Petrol. Geol. I: 473-517.

McMillan, N. J. 1982. Nares Strait and the petroleum explorer. - This volume.

McWhae, J. R.H. 1981. Structure and spreading history of the northwestern Atlantic region from the Scotian Shelf to Baffin Bay. - In: Kerr, J. W. & Fergusson, A. J. (ed.), Geology of the North Atlantic borderlands. - Mem. Can. Soc. Petrol. Geol. 7: 299-332.

Meyerhoff, A. A. 1970. Continental drift II: high latitude evaporite deposits and geologic history of Arctic and North Atlantic Ocean. - J. Geol. 78: 406-444.

Meyerhoff, A. A. 1973. Origin of Arctic and North Atlantic Oceans. - In: Pitcher, M. G. (ed.), Arctic geology. - Mcm. Am. Ass. Petrol. Geol. 19: 562-582.

Miall, A. D. 1981. Late Cretaceous and Paleogcne sedimentation and tectonics in the Canadian Arctic Islands. - In: Miall, A. D. (ed.), Sedimentation and tectonics in alluvial basins. - Spee. Pap. geol. Ass. Can. 23: 221-272.

Munck, S. 1941. Geological observations from the Thule District in the summer of 1936. - Meddr Grnnland 124(4): 38 pp.

Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. - Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Newman, P. H. & Falconer, R. K. H. 1978. Evidence for movement between Greenland and Canada along Nares Strait. - Geol. Soc. Am. Abs. with Prog. 10: 463 only.

Nichols, R. L. 1961. Geomorphology of Inglefield Land, North Greenland. - Meddr Grnnland 188(1): 109 pp.

Pelletier, B. R., Ross, D. I., Keen, C. E. & Keen, M. J. 1974. Geology and geophysics of Baffin Bay. - Pap. geol. Surv. Can. 74-30(2): 247-258.
Pitman, W. C. & Talwani, M. 1972. Sea-floor spreading in the North Atlantic. - Bull. geol. Soc. Am. 83: 619-646.
Prest, V. K. 1952. Notes on the geology of parts of Ellesmere and Devon Islands, Northwest Territories. - Pap. geol. Surv. Can. 52-32: 15 pp.
Riddihough, R. P., Haines, C. V. & Hannaford, W. 1973. Regional magnetic anomalies of the Canadian Arctic. - Can. J. Earth Sci. 10: 157-163.
Roy, J. L. 1973. Latitude maps of the eastern North American - western European paleoblock. - In: Hood, P. J. (ed.), Earth science symposium on offshore eastern Canada. - Pap. geol. Surv. Can. 71-23: 3-22.
Sclater, J. G., Hellinger, S. & Tapscott, C. 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. - J. Geol. 85: 509-552.
Sobczak, L. W. & Stephens, L. E. 1974. The gravity field of northeastern Ellesmere Island, part of northern Greenland and Lincoln Sea with map. Lincoln Sea 1:500 000. - Earth Physics Branch, Gravity Map Series 114: 9 pp.
Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. - Geophys. J. Roy. astr. Soc. 52: 313-357.
Srivastava, S. P. & Falconer, R. K. H. 1982. Nares Strait: a conflict between plate tectonic predictions and geological interpretation. - This volume.
Srivastava, S. P., Falconer, R. K. H. & Maclean, B. 1981. Labrador Sea, Davis Strait, Baffin Bay: geology and geophysics - a review. - In: Kerr, J. W. & Fergusson, A. J. (eds), Geology of the North Atlantic borderlands. - Mem. Can. Soc. Petrol. Geol. 7: 333-398.
Sweeney, J. F., Irving, E. & Geuer, J. W. 1978. Evolution of the Arctic Basin. - In: Sweeney, J. F. (ed.}, Arctic geophysical review. - Puhl. Earth Physics Branch 45(4): 91-100.
Tailleur, I. L. 1973. Probable rift origin of Canada Basin, Arctic Ocean. - In: Pitcher, M. G. (ed.}, Arctic geology. - Mem. Am. Ass. Petrol. Geol. 19: 526-535.
Tarling, D. H. & Tarling, M. P. 1971. Continental drift: a study of the Earth's moving surface. - G. Bell and Sons Ltd., London: 112 pp.
Taylor, F. B. 1910. Bearing of the Tertiary mountain belt on the origin of the earth's plan. - Bull. geol. Soc. Am. 21: 179-226.
Thorsteinsson, R. 1963. Copes Bay. - In: Fortier, Y. 0. et al., Geology of the north-central part of the Arctic Archipelago, Northwest Territories (Operation Franklin). - Mem. geol. Surv. Can. 320: 386-395.
Trettin, H. P., Frisch, T. 0., Sobczak, L. W., Weber, J. R., Niblett, E. R., Law, L. K., DeLaurier, J. M. & Whitham, K. 1972. The Innuitian Province. - In: Price, R. A. & Douglas, R. J. W. (eds), Variations in tectonic styles in Canada. - Spec. Pap. geol. Ass. Can. 11: 83-179.
Troelsen, J. C. 1950. Contributions to the geology of Northwest Greenland, Ellesmere Island and Axel Heiberg Island. - Meddr Grnnland 149(7): 85 pp.
Umpleby, D. C. 1979. Geology of the Labrador Shelf. - Pap. geol. Surv. Can. 79-13: 34 pp.
van der Linden, W. J. M. 1975. Crustal attenuation and sea-floor spreading in the Labrador Sea. - Earth planet. Sci. Lett. 27: 409-423.
van der Linden, W. J.M. 1977. How much continent under the oceans? - J. Mar. geophys. Res. 3: 209-224.
van Waterschoot van der Gracht, W. A. J. M. et al. 1928. Theory of continental drift. A symposiJm on the origin and movement of land masses, both inter-continental and intra-continental, as proposed by Alfred Wegener. - Am. Ass. Petrol. Geol., Tulsa: 240 pp.
Vogt, P.R. & Avery, 0. E. 1974. Detailed magnetic surveys in the northeast Atlantic and Labrador Sea. - J. geophys. Res. 79: 363-389.
Wegener, A. 1915. Die Entstehung der Kontinente und Ozeane. - Friedr. Vieweg & Sohn, Braunschweig: 94 pp.
Wegener, A. 1922. Die Entstehung der Kontinente und Ozeane. - Friedr. Vieweg & Sohn, Braunschweig: 144 pp (3rd revised edit.).
Wegener, A. 1924. The origin of continents and oceans. - Methuen & Co., London: 212 pp. (Transl. 3rd edit. by J. G. A. Skerl}.
Wegmann, C. E. 1948. Geological tests of the hypothesis of continental drift in the Arctic regions. Scientific planning. - Meddr Grnnland 144(7): 48 pp.
Wetmiller, R. G. 1974. Crustal structure of Baffin Bay from earthquake-generated Lg phase. - Can. J. Earth Sci. 45: 15-24.
Wilson, J. T. 1963a. Hypothesis of Earth's behaviour. - Nature, Lond. 198: 925-929.
Wilson, J. T. 1963b. Continental drift. - Scient. Am. 208(4): 86-100.
Wilson, J. T. 1965a. Evidence from ocean islands suggesting movement in the Earth. - In: Blackett, P. M. S., Bullard, E. & Runcorn, S. K., A symposium on continental drift. - Phil. Trans. Roy. Soc. Lond. 258A: 145-167.
Wilson, J. T. 1965b. A new class of faults and their bearing on continental drift. - Nature, Lond. 207: 343-347.
Windley, B. F. 1977. The evolving continents. - John Wiley & Sons, London: 385 pp.
Wordie, J. M. 1938. An expedition to Northwest Greenland and the Arctic Islands in 1937. - Geogr. J. 92: 385-421.
Wright, W. B. 1923. The Wegener hypothesis. - Nature, Lond. 111: 30-31.

Physiography

Physiography of Nares Strait: importance to the origin of the Wegener Fault

DAVID MONAHAN and G. LEONARD JOHNSON

Monahan, D. & Johnson, G. L. 1982. Physiography of Nares Strait: importance to the origin of the Wegener Fault. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 53–64.

Nares Strait is a long, deep trough which, at its shoreline at least, is straight. Most theories of its origin are deduced from observations remote from the Strait and even those observations adjacent to the Strait were seldom if ever taken for the express purpose of elucidating its origin. Direct observations of the floor of the Strait are very scarce, yet without them the dilemma of movement or non-movement seems unresolvable. We have been able to map the gross physiography of the Strait and can deduce from it that the area has undergone compression along an approximately northwest–southeast axis.

In order to resolve the dilemma of motion or no motion along Nares Strait, we suggest that a post-Palaeozoic northward movement (50–100 km) of the Canadian Arctic Islands by the process of crustal thinning in Lancaster and Jones Sounds and in the Sverdrup Basin can, when combined with geologic/geometric uncertainties, account for the present-day relatively small observed net offset.

D. Monahan, Canadian Hydrographic Service, Department of Fisheries & Oceans, 615 Booth Street, Ottawa, Ontario, Canada, K1A OE6 and G. L. Johnson, Department of the Navy, Office of Naval Research, Arlington, Virginia 22217, U.S.A.

The Nares Strait dilemma is a mapping problem; participants in the Nares Strait symposium have mapped and interpreted some phenomena which have a bearing on the nature of the feature itself or of earth processes and motions that may or may not have taken place along it. Carefully reasoned, internally consistent arguments starting from different measurements and based on different premises lead to conflicting conclusions, indicating that either the measurements are erroneous or irrelevant or that the premises on which the arguments rest are invalid. In this contribution we are concerned with the morphology of the Strait and its spatial relationships with the adjacent land and oceanic areas, and of course our measurements and premises are as fraught with dangers as are everyone else's.

The facts, absolute and unarguable, of this case are extremely few; there is a big trough between Greenland and Ellesmere Island (Fig. 1). We do not know what it is, how it got there or when it was formed. Those workers who have examined the geology of one or both sides, and we should emphasize that they are extremely few, have largely ignored the necessity for explaining the presence of the trough or have fallen into the trap they implicitly ascribe to others in that they have made assumptions as to the origin of Nares Strait based on scant or non-existent information. Alternatively those who ascribe an origin to the trough because conclusions they reach regarding the history of adjacent oceanic areas are otherwise untenable, seem to believe the trough offers no restrictions to their conjecture; the physical reality of Nares Strait enters their calculations not at all.

What does the trough look like? As stated earlier, we are dealing with a mapping problem and consequently our answer depends on the scale and area of investigation. Looked at from far enough away, Nares Strait would not exist at all, that is, there would be no break between Greenland and Ellesmere Island. But as we move closer, it would become apparent that Greenland and Ellesmere are separated by water; through this we may be tempted to draw a straight line, but closer inspection still would reveal that the straight line would only be a very rough approximation to the complex shape that now appears. To the man in the field, the trough looks very complicated indeed. The most important of these complications is that he finds that rather than looking down at the trough as we do when looking at a map, he is looking up at land surrounding him. The trough has an impressive vertical component. In cross-section we have two very dissected plateaux of about 1500 m elevation separated by a trough as deep as 800 m below sea level and from 25 to 100 kilometres wide (Fig. 2) Did material ever fill this void? If so, where is it now? Any possible mode of origin must answer these questions as well as satisfy all the relevant conditions outlined by other papers in this symposium and in the literature.

Assuming the trough to approximate a long, linear depression we must ask how such a physiographic fea-

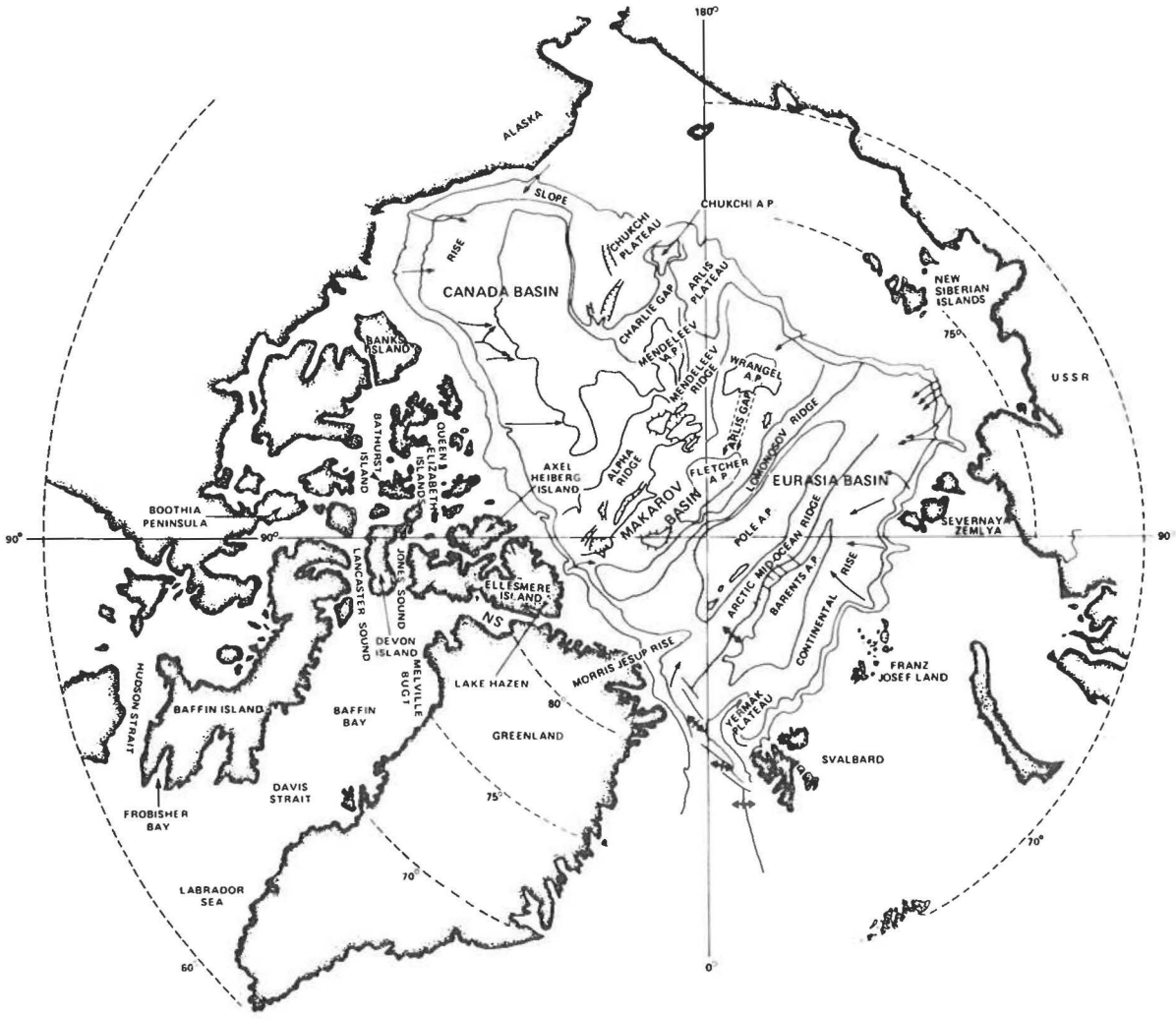


Fig. 1. Index map of the oceanic and continental regions adjacent to Nares Strait (NS).

ture could form. Firstly, assume it was formed in a continuous landmass, that is, no drift occurred, then, (Fig. 3b) erosion could have preferentially removed material to form the trough; compression, extension or upwarp of the crust could cause a graben to form (Fig. 3c). Compression could have caused the crust to buckle or down warp. Secondly, assume that the trough formed in a continuous landmass but allow the landmass to be fractured by a plate tectonic event, then (Fig. 3e) a transform fault (wrench fault) could have cut the crust; this would have to include either a component of drift transverse to the fault to open up the trough (Fig. 3f, 1) or, failing that, be followed by lateral erosion away from the fault trace (Fig. 3f, 2). Rifting apart of the two sides of the trough due to rotation of Greenland counterclockwise relative to Ellesmere Island (Fig. 3g) or rifting without rotation, that is, a straight pull apart between Greenland and Ellesmere Island (Fig. 3d), could also have occurred. Thirdly, we could suppose that Greenland and Ellesmere Island currently rest on plates that are drifting towards one another, their proximity at present forming a trough. We thus have eight possibilities, some of which have had adherents in the literature (see Kerr 1980), some of which apparently have been ignored, but all of which warrant some attention especially in view of our current state of ignorance of Nares Strait.

It is sobering to put our ignorance into perspective by considering approaches taken to similar problems in the past. Consider, for example, the San Andreas Fault. There is a strong resemblance between the objectives of this volume and that addressed by two symposia on a similar and related theme, namely the magnitude of movements along the San Andreas Fault. Forty papers (Dickinson & Grantz 1968) followed by 50 papers (Kovach & Nur 1973), addressed the problems of

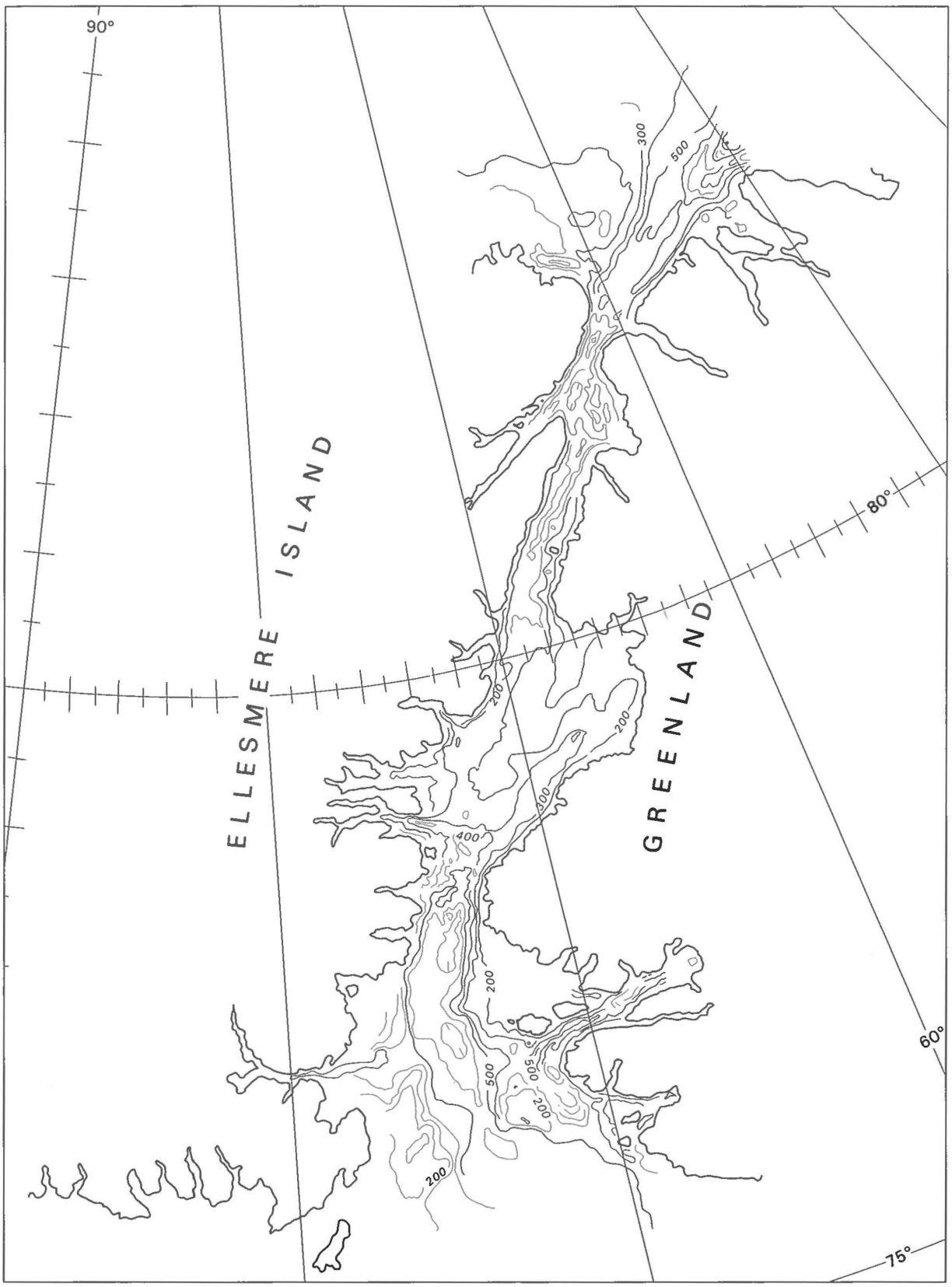


Fig. 2. Bathymetry of Nares Strait (after Johnson et al. 1979). Depths are in metres.

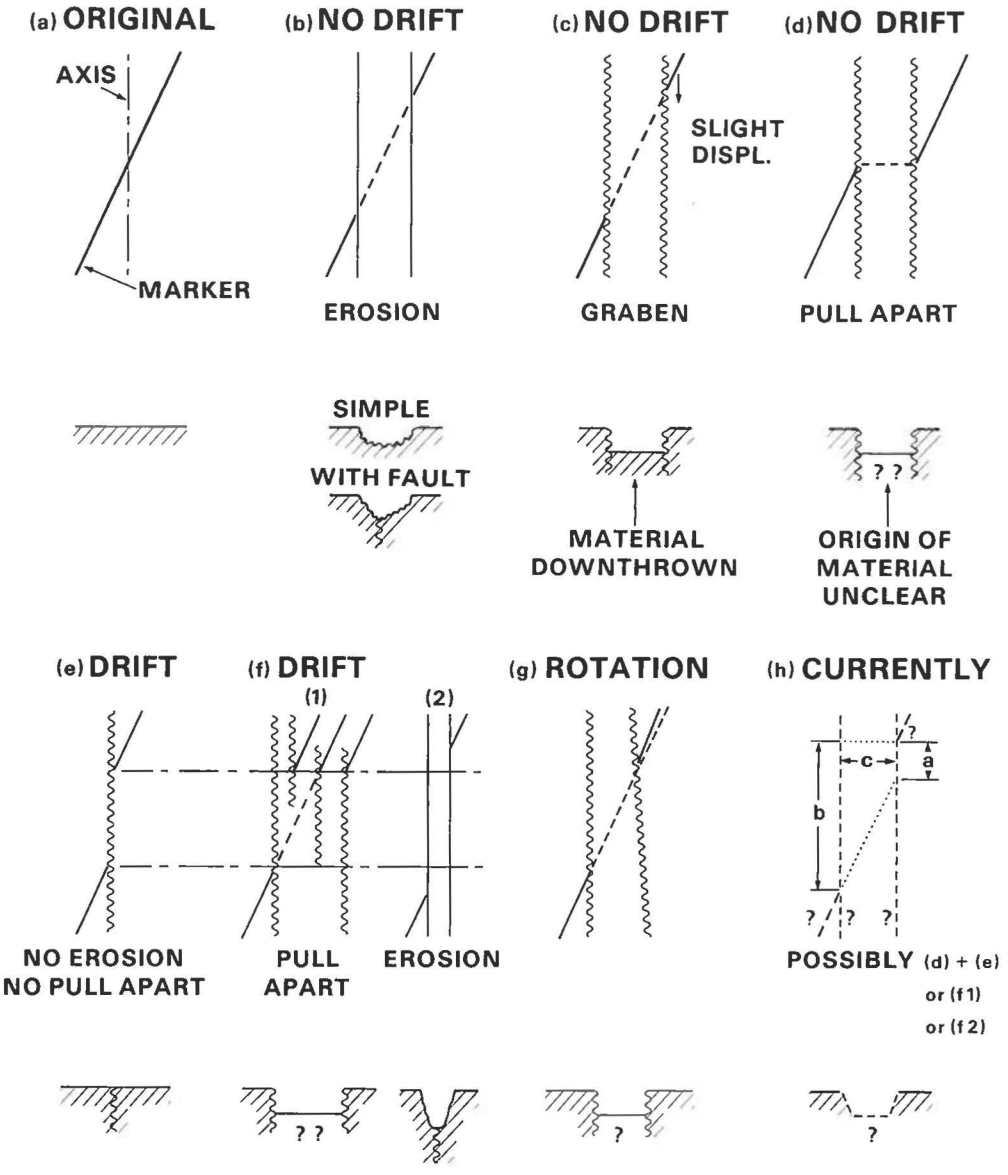


Fig. 3. Diagrammatic representation of several modes of origin of Nares Strait. Upper portion of each diagram is a plan view, lower portion a cross-section. The marker shown intersects Nares Strait at 30° which is approximately representative of many markers mapped along the Strait. (a) shows one possibility for the original configuration before the Strait was formed. This assumes that the Strait was formed in a single continental plate although the possibility that Greenland and Ellesmere Island have never been part of the same plate cannot be ignored, (b) shows a simple case of preferential erosion, (c) shows a simple graben, (d) shows faulting followed by separation between each side, (e) shows simple drift along a fault axis, (f) shows drift followed by separation (1) or by erosion (2). Note that in (1), depending on the amount of separation, geometry similar to that in (b), (c) and (d) could be produced, (g) shows the case for slight rotational separation and (h) shows the situation as currently mapped by some workers, for some markers. Assuming no vertical movements, this could have arisen from some combination of (d) and (e) or from (f). In such a case, minimum drift is 'a', maximum 'b' and uncertainty 'a–b'.

movement along the San Andreas Fault, a feature that is: a) clearly a fault; b) clearly has had strike-slip movement along it; c) conveniently outcrops on land for over 800 km; d) does not form an international boundary; e) is readily accessible and f) has a strong, fault trace derived topographic expression. These produced only a plethora of contradictory estimates of movement ranging from 45 yards to 450 km. We, on the other hand, are attempting to elucidate a feature: a) whose nature is not clearly established, that is, we do not know

that it is a fault; b) may or may not be the product of strike-slip movement; c) is inconveniently covered with sea-water or ice; d) forms an international boundary adding to the already logistically difficult location; e) is not only inaccessible but hides some of the relevant geology under permanent ice caps or yearly snowfall and f) has a most enigmatic topographic expression. Should the San Andreas Fault not seem an apt enough model, consider the Great Glen Fault in Scotland which has a mere 160 kilometres of outcrop. A summary by Harris et al. (1978) indicates that this fault has reportedly suffered sinistral shifts of between 104 and 224 kilometres, or dextral shifts of between 29 and 160 kilometres, depending on whose field evidence is accepted. These examples are cited to point out the extreme difficulty in determining or refuting strike-slip movement of any sort on the ground. We are faced with a far more difficult situation in that we do not even know whether Nares Strait represents the surface expression of strike-slip movement.

This evidence of the extreme difficulty in dealing with major earth movements in the field is again symptomatic of the problems of scale in relating field observations to major crustal deformation. The high number of hypotheses and magnitudes of movements advocated on other similar features indicates their extreme complexity. A hypothesis that explains one portion of a feature of this nature may not be adequate or appropriate in other sectors of its length. In Nares Strait, very, very few people have traversed its entire length looking for a unifying hypothesis that would explain the many manifestations of this complicated feature.

One could then consider the measurements made in the oceans surrounding the Strait. Here again we are extremely ignorant. The best way of illustrating this is to refer to the papers by Grant (1980, this volume) in which he presents a composite diagram showing 16 different positions for Greenland as reported in the literature mainly based on measurements made at sea pertaining to plate tectonics. Not all of the 16 positions can be correct. None of the 16 reconstructions he cited nor any of the others that could have also been included that we have been able to find in the literature deal with the problem of showing what the Strait itself is, although it is presumed to be a transform fault by most marine geologists and geophysicists.

We thus find two polarizations, almost ideologies of approach, neither of which has had the direct objective of studying the origin of the trough. The efforts of land based geologists have been directed towards elucidating the formations which occur on one side or on the other side of the Strait; they have largely ignored, and continued to ignore throughout the symposium, modes of origin for the Strait. Several of them have discussed their work with their colleagues working on the opposite side of the Strait and decided on the basis of these discussions that they are mapping equivalent formations. This is still not conclusive evidence that drift did or did not occur, since equivalent formations can be mapped on either side of the Atlantic Ocean to name one example. Conclusions of no drift seem to be a manifestation of the Law of Parsimony; since there is no reason to invoke drift in explaining the geology, it is not invoked.

The alternative philosophical approach is based on marine geophysical observations some distance removed from the Strait. These observations, primarily magnetic lineations and oceanic crustal velocities, were first interpreted in the Atlantic Ocean as manifestations of sea-floor spreading. Subsequently, similar observations were interpreted in the Labrador Sea and in Baffin Bay to mean that Greenland and the North American landmass separated from one another by such a process. This interpretation, and it is not a necessary interpretation since the Atlantic could have opened without displacing Greenland, requires that there be relative movement between Greenland and some part of the Canadian Arctic; this movement, on maps at least, is most easily accommodated along Nares Strait. The literature pertaining to the geology surrounding Nares Strait has not been abundant, and certainly not been conclusive, and to marine geophysicists it has not offered sufficient constraints to deter them from moving Greenland relative to Ellesmere Island. This too is an application of the Law of Parsimony; since there is no reason not to move Greenland, move it.

This situation of having two contrary schools of thought is a natural, healthy one, from which will come a new third viewpoint that will be closer to explaining how Nares Strait was actually formed. Such a third view will only come through attempting to synthesize what each school has to contribute.

Geometric considerations

Let us examine the geometry of the situation starting from the totally unsubstantiated assumption that Greenland and Ellesmere Island have contributed through mutual interaction to form Nares Strait. How would drift along Nares Strait manifest itself under the many possible circumstances that could have taken place? Fig. 3a shows diagrammatically the axis of an incipient Nares Strait together with a geological marker or boundary intersecting the axis at 30°, which approximates the situation in at least the northern portion of the Strait. Not all markers intersect the Strait at 30°, but it seems that those best mapped do so. Note the totally unfounded assumption that the marker is straight. Fig. 3b shows one possible origin for Nares Strait in which simple erosion focused on the Strait for some unknown reason, possibly along some type of fault, created a trough with no relative movement between Greenland and Ellesmere Island. The marker remains linear. Should Nares Strait be a graben, then as

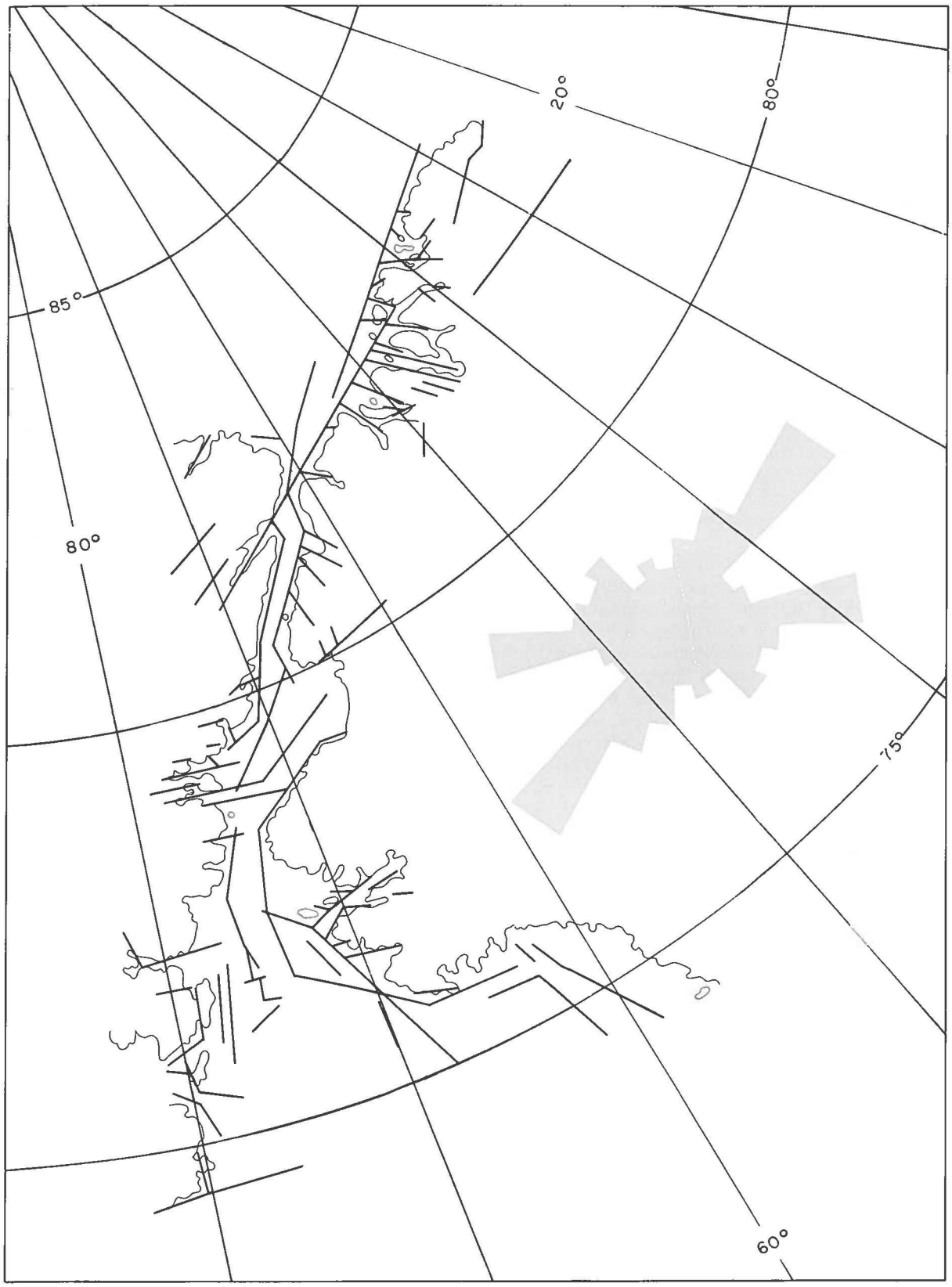
Fig. 3c indicates, there would be a very slight displacement of the marker and the material flooring the trough should be similar to that on the adjacent land. Next, as in Fig. 3d, we assume that there has been no drift but that Greenland and Ellesmere Island have pulled apart from one another. In this case the origin of the material flooring the trough would be unclear and the marker would be displaced significantly from its otherwise linear configuration. Turning to situations in which drift did occur, Fig. 3e shows displacement of the marker which would occur with merely strike-slip movements along the axis. This is meant to be read in conjunction with Fig. 3f in which drift accompanied by pull apart or by erosion has taken place. Note that with pull apart it is entirely possible to arrive at marker configurations similar to those in Figs 3b or 3c in which no drift occurred. Pulling apart even further would produce confusing situations somewhere between the configuration shown in Figs 3c and 3d. Again the nature of the material forming the floor of the trough would be unknown. Not to be ignored is the possibility that Greenland rotated counter-clockwise relative to Ellesmere Island as shown in Fig. 3g. It is possible from the geometry of the Strait itself to postulate rotations of about $3^1/_2°$ and such rotation would be virtually undetectable with the present configuration of markers.

The situation shown in Fig. 3h is that currently mapped by many workers. It shows a strait of unknown origin whose sides were formed in some as yet unresolved fashion and with geological markers displaced away from an extension of one another. If this configuration is correct, and if the situation has not been confused by vertical movements, then this situation could have only been caused by some combination of events which has to include those portrayed in Figs 3e and 3f. That is, some drift must have occurred for the mapped marker to be displaced by some amount. It is most useful to attach some approximate numbers to this diagram to determine the limits that any one marker can provide. We know from Fig. 3h that at least the amount of drift shown by 'a', can have taken place and possibly the amount shown by 'b'. The difference between 'a' and 'b' is therefore the amount of uncertainty in any measurements or estimates of drift based on geologic markers on either side of the Strait. If the Strait is 30 kilometres wide and the angle of intersection between the marker and the Strait is 25°, then the zone of uncertainty is about 65 km long: a 30° intersection gives 52 km and a 35° intersection gives 42 km. This slop could worsen to about 107 km if the width of the Strait were 50 km and the angle of intersection 25° and diminish to only 4 km if the Strait were 25 km wide and the angle of intersection 50°. It seems safe therefore to consider that there is approximately 50 km of slop inherent in the geometry of any measurement. This is a limit on the resolution of any movement that may or may not have taken place along the Strait. Unless evidence is presented that will substantiate or refute opening perpendicular to the axis of the Strait, drift less than approximately 50 km cannot be detected.

In the above it was assumed that the geological marker was not only straight, but clearly agreed upon. This is certainly not the case. Dawes (1973) discusses the many problems of deciding which boundaries should be safely correlated across the Strait and it is those very problems that attracted some participants to this conference. Part of the problem arises from the different interpretation of subdivisions used by those workers who advocate the geosynclinal model, but a more basic problem arises from the actual tectonic differences on each side of the Strait. Dawes (1973) reviews the fact that folding differs on either side of the Strait in that axial planes of folds dip north in Ellesmere Island and south in Greenland. This is not at all unexpected provided that the Strait is a major tectonic feature; textbooks on structural geology written twenty or thirty years ago are replete with descriptions of what were then called wrench faults, on either side of which folding directions changed in a manner very similar to that described in the region of Nares Strait. This is usually attributed to compression and is consistent with the pattern of linear scarps summarized in Fig. 4. Fig. 4 shows the trace of the scarps, together with a rose diagram in which length and sense of scarps have been weighted and assembled using a method suggested by Scheidegger (1979). This pattern clearly supports compression along a NW–SE axis.

The continuity of geological features across the Strait, which leads some workers to the conclusion that little or no drift has occurred, paradoxically led others in the nineteen-sixties to believe in drift between, say, North America and Europe. The apparent continuity of similar features across a body of water argues that they were once contiguous and since, as we have shown above, in Nares Strait neither continuity nor geometry offers any degree of rigidity, we feel that the case of non-drift is definitely not proven. Nor, however, is the case for drift. Although there are several models of the Arctic Ocean, Atlantic Ocean and Baffin Bay opening, none has any direct evidence from within the Strait itself. Until extensive geophysical work examines the floor of the Strait neither case can be definitively proven.

Fig. 4. 'Scarp' lineations in Nares Strait. Straight lines are scarp traces. Rose diagram represents an assemblage of lineations as percentage of total length of lineations compiled after a method suggested by Scheidegger (1979). The pattern shown could be interpreted as manifestation of compression along a NW–SE axis.


20°
80°
85°
80°
75°
60°

Geologic history of the Northwest Territories

Next we will go on to discuss briefly the history of the adjacent regions and present a hypothesis which might help to resolve the Nares Strait dilemma.

Kerr (1981) has related the Canadian Northwest Territories (NWT) to oceanic spreading in the Arctic and sub-Arctic. This study, based on a somewhat more refined history of sea-floor spreading in the Arctic is built upon his pioneering study. Our summary is contained in Table 1 with appropriate bibliographic references.

Kerr notes that initial fragmentation of the Precambrian and Palaeozoic core commenced in latest Devonian or early Mississippian time when the Sverdrup Basin first commenced receiving sediment. We consider this, the earliest event, as an abortive spreading centre which remained active until rifting was replaced by general subsidence as sediment continued to pour in. Faunal evidence (presence of circum-Arctic Upper Palaeozoic fauna) suggests that at least shallow waterways were present and indeed perhaps one occupied the suture from which Alaska was rifted away from the Canadian Arctic.

In the early Cretaceous a tensional regime fractured the crust, causing subsidence of the Sverdrup Basin into which the thick Isachsen Formation sandstone was deposited. This coincided with active sea-floor spreading in the Canada Basin which might be assumed to be the cause of the regional tension. Concomitant volcanism occurred with local basalt flows in the Isachsen Formation of northwest Axel Heiberg Island, while faulting and folding occurred in the Wandel Sea Basin of northern Greenland and in Svalbard (Dawes & Peel 1981).

The Eurekan rifting episode commenced in the late Cretaceous (Campanian–Maastrichtian time) as evidenced by sea-floor spreading in Labrador Sea/Baffin Bay as Greenland separated from North America. At this time, the faults and fault-controlled channels in the southeastern rift system formed. The Lancaster aulacogen is the most spectacular of these features. To the northwest, the Boreal orogeny resulted in regional uplift and erosion of the broad intra-basin arches in latest Cretaceous and early Tertiary. In the late Cretaceous there apparently was rifting in the southeast and northwest with little or none in the region between. The Canadian Arctic Islands region was subjected to two rifting events. The Boreal rifting episode caused uplift of the Pearya geanticline and Sverdrup rim. In the Paleocene and Eocene (65–45 m.y.), the crust of the Archipelago was extended by creation of large grabens and uplift of arches (Balkwill & Bustin 1978). Minor volcanism is recorded by the single basalt flow on Bathurst Island and farther southeast by the well-known early Tertiary volcanism on Baffin Island (Trettin et al. 1972). We suggest the uplift related to the doming and initiation of sea-floor spreading in the Fletcher Abyssal Plain region which commenced in the Campanian and ended in the mid-Eocene (Taylor et al. 1981). The main volcanic episode occurred in eastern Axel Heiberg Island at this time (75 m.y.) and comprises over 250 m of basalt flows — the Strand Fiord Formation (Trettin et al. 1972).

The Eurekan deformation reached its climatic phase in mid-Tertiary time (middle Eocene to early Miocene). During this stage, sea-floor spreading slowed and stopped in Baffin Bay, and the rifts were propagated farther northwest. The advance of the rifts northwestward into the Queen Elizabeth Islands was impeded by the pre-existing structural trends, some of which were transverse to their paths and deflected their propagation. The Lancaster aulacogen was then further deflected southwestward by the Boothia Uplift.

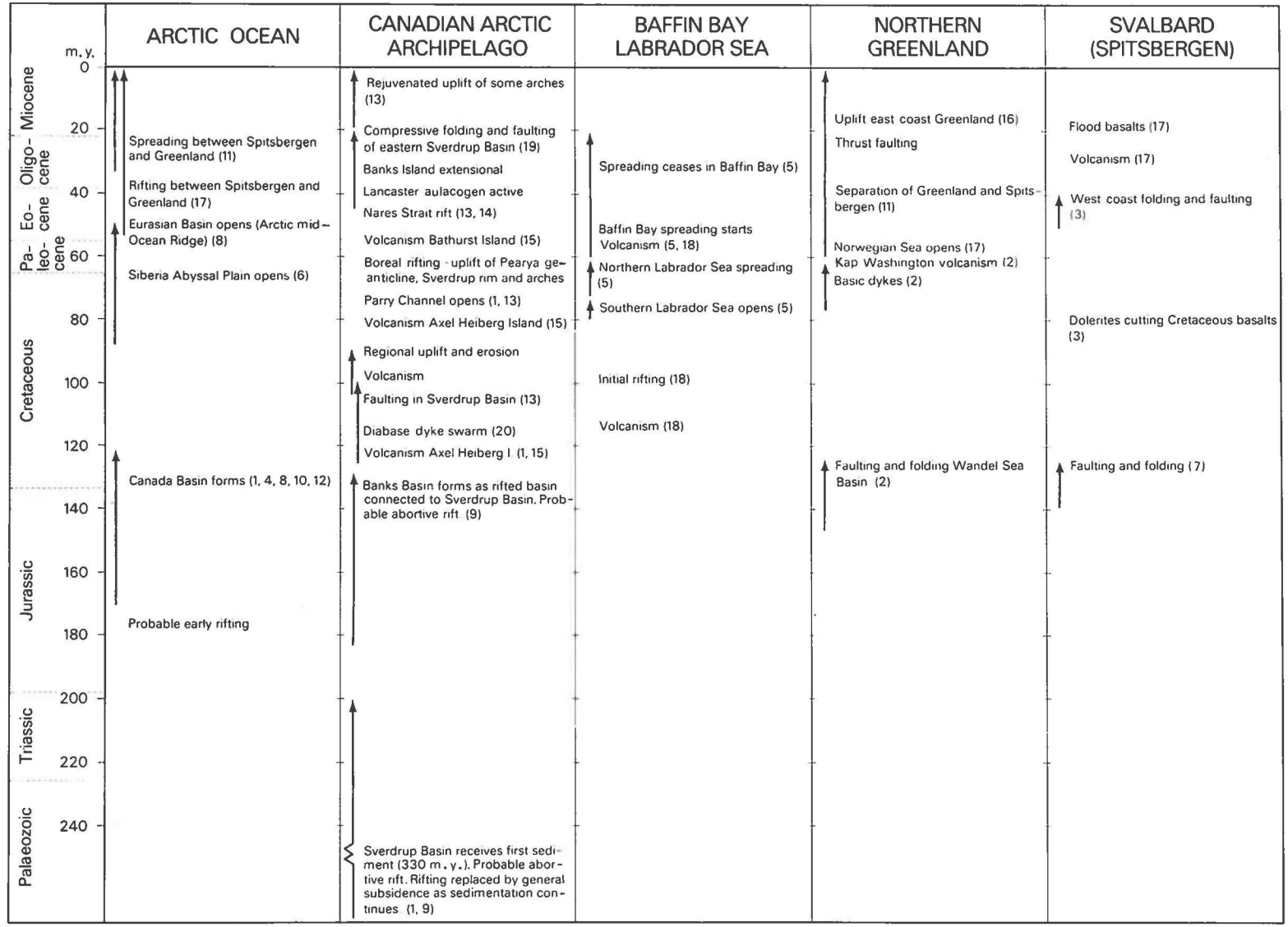
Mid-Tertiary compressive deformation of the northeast NWT may be related to cessation of Baffin Bay spreading together with pressure from Greenland exerted by the Norwegian–Greenland spreading centre as Greenland affixed itself to the American plate. Pressure was perhaps applied northwestwards by Greenland across Nares Strait. Some support for this is found in the pattern of linear scarps, both onshore and underwater, in the Nares Strait region. Fig. 4 shows the trace of the scarps, which clearly supports compression along a NW–SE axis. In the Miocene there was some rejuvenated uplift of the NWT which might relate to an increase in the spreading rate of the Nansen Ridge (Vogt et al. 1979a).

Triple junctions

A common world-wide characteristic of graben systems is the occurrence of trilete patterns where three potential spreading ridges diverge to form rift, rift, rift or RRR junctions (McKenzie & Morgan 1969). Non-oceanic examples include the trilete systems of the North Sea, Gulf of Suez and Benue Valley (Whiteman et al. 1975). The RRR junctions are postulated to be generated by thermal expansion of the crust as a result of a "plume" or hot spot in the mantle (Morgan 1971, 1972). The hypothesis holds that updoming with con-

Table 1.
Numbers in parenthesis refer to references cited as follows: (1) Balkwill (1978), (2) Dawes (1976), (3) Harland (1973), (4) Sweeney et al. (1978a), (5) Srivastava (1978), (6) Taylor (1978), (7) Harland (1969), (8) Vogt et al. (1979a), (9) Bally (1976), (10) Johnson et al. (1979), (11) Feden et al. (1979), (12) Churkin et al. (1979), (13) Kerr (1980), (14) Kerr (1981), (15) Trettin et al. (1972), (16) Brooks (1979), (17) Vogt et al. (1979b), (18) Hinz et al. (1979), (19) Newman et al. (1977).

Table 1. Summary of relevant events in the formation of the Arctic region (after Monahan & Johnson 1980).



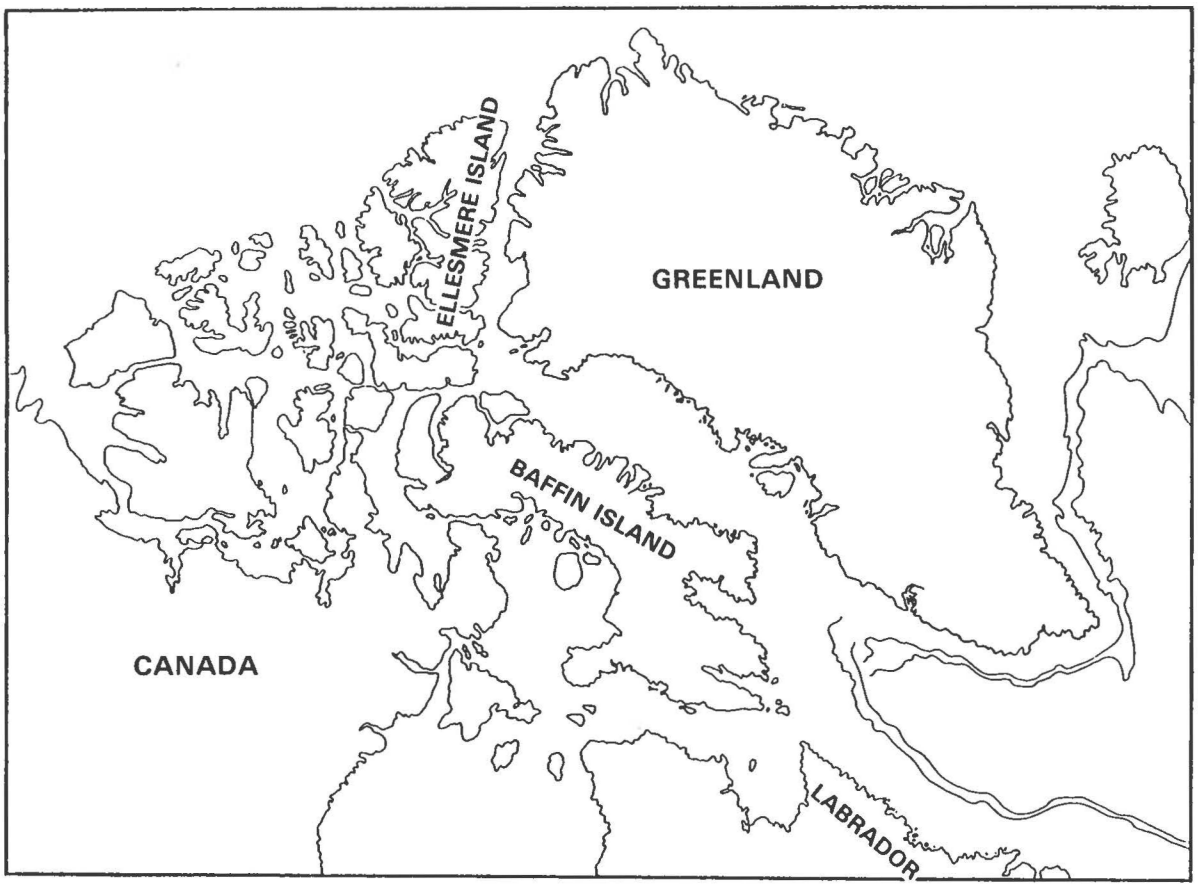


Fig. 5. Suggested late Palaeozoic reconstruction of the landmasses around Nares Strait showing Lancaster Sound and Jones Sound closed.

sequent thinning of the Earth's crust by subcrustal thermal erosion occurs over the thermal anomaly. Subsequent to the doming, crustal separation may occur with sea-floor spreading and the generation of new ocean crust. The trilete rifting may represent a least work configuration. Morgan (1972) has postulated that plume centres are responsible for the initial split of the Atlantic and other oceans. In the case of the North Sea and other continental examples, extensional faulting may occur; however there is no physical separation of the lithospheric plate. These have been referred to as "failed arms" by Brooks (1973) in describing a trilete pattern in East Greenland.

It seems possible that Baffin Bay, Lancaster Sound and the Nares Strait lineament are trilete rifts consequent to a "plume" or thermal anomaly which became active in northern Baffin Bay during the late Cretaceous–Paleocene time. This idea is not unique and several authors have briefly alluded to the possibility (Burke 1976).

In the late Cretaceous, doming affected extensive areas of the eastern Canadian Arctic Islands and northern Greenland (Trettin et al. 1972). It is postulated that some of the uplift resulted from a plume centred northeast of Lancaster Sound. This uplift led to rifting of the crust at several places (e.g. Frobisher Bay). It seems likely that the initial pattern was a rift or graben along the continental margin of Greenland (the Melville Bugt graben). Fig. 5 shows our postulated late Palaeozoic arrangement of the landmasses with Nares Strait, Lancaster Sound and Baffin Bay forming the trilete pattern. The latter contained at least a limited spreading axis while the two former are "failed arms".

No basic igneous rocks of proven Cretaceous–Tertiary age are known in the northern Baffin Bay region, but Paleocene basalts have been reported near Lake Hazen, northeastern Ellesmere Island (Christie 1964). In this regard northern Baffin Bay may be similar to the North Sea where basaltic rocks are obscured by sediment cover (Ziegler 1978).

Arctic Ocean

The Eurasia Basin is well demarcated by magnetic anomalies which date the separation of the Yermak Plateau and Morris Jesup Rise at anomaly 13 (38 m.y., Heirtzler et al. 1968, Feden et al. 1979) and the separation of the Lomonosov Ridge from Eurasia at or before anomaly 24 time (57 m.y., LaBrecque et al. 1977, Vogt et al. 1979b). However, following the suggestion of Taylor et al. (1981), that there are spreading anomalies between the Lomonosov and Alpha Ridges, we favour a continental origin for the Alpha Ridge (Johnson et al. 1978).

The date of the presumed rifting of the Alpha Ridge from the Lomonosov Ridge is uncertain. It might coincide with the initial rifting between Greenland and North America which according to Srivastava (1978) started during late Cretaceous and continued through anomaly 32 (75 m.y., Heirtzler et al. 1968). Anomaly 32 is the oldest anomaly that can be identified in the Labrador Sea (Srivastava 1978), and therefore is a favorable time to rift Alpha Ridge from the Eurasian continental margin. Srivastava's pole of opening at anomaly 32 time was located at 70.80°N, 150.93°E which allows the Arctic to be under a condition of tensional tectonic stress. Taylor et al. (1981) suggest that this spreading episode commenced in the late Cretaceous (anomaly 34/80 m.y.) and continued until mid-Eocene (anomaly 19/47 m.y., Heirtzler et al. 1968).

Palaeomagnetic data and evidence from regional structural and stratigraphic relationships in northern Alaska presented by Newman et al. (1977) suggest that the rotation of the Arctic Alaskan plate away from the Canadian Arctic Archipelago to create the Canada Basin began during the late Jurassic or earliest opening of the North Atlantic Ocean (Sweeney et al. 1978a, b). Rotation may have been followed by southward translation of the entire rotated block away from the Alpha–Mendeleev Ridge. The inception of rotation may have been as early as the Triassic (Newman et al. 1977, Sweeney et al. 1978 a, b), but no direct evidence of faulting of this age is seen in northern Alaska. Based on aeromagnetic data Taylor et al. (1981) have dated this spreading episode as occurring between 153 m.y. and 127 m.y. (Upper Jurassic – Lower Cretaceous). The spreading history thus indicates that with concurrent spreading in the Nansen Basin, Makarov Basin and Baffin Bay, Nares Strait would serve as a transform fault with a triple junction in northern Baffin Bay as well as at the northern end of Nares Strait.

Conclusion

Did Nares Strait act as a transform fault? Present reconstructions (Srivastava 1978, Srivastava & Falconer, this volume) show a northward movement of Greenland relative to Ellesmere Island by a left-lateral motion along Nares Strait of 250 km. We have previously noted that there can be a geometric ambiguity of 60–70 km. Some land geologists generally allow for 25–100 km of offset (Christie et al. 1981). Assuming the maximum, this still leaves a minimum of 80 km unaccounted for. We suggest this amount and possibly more can be accounted for by post-Palaeozoic extension or continental thinning in the Canadian Arctic. This would be similar to the continental "thinning or stretching" suggested for the North Sea grabens by Sclater & Christie (1980). This would have the effect of shifting the original Palaeozoic marker beds of Ellesmere Island to the north, therefore this motion can be subtracted from the unresolved 80 km. The width of Lancaster Sound is 75 km, Jones Sound is 40 km and the Sverdrup Basin has apparently been under tension since the late Palaeozoic with perhaps another 50 km of extension (Table 1). Therefore our hypothesis to resolve the dilemma of motion or no motion along Nares Strait is as follows: During the late Palaeozoic the continental fragments were arranged as shown in Fig. 5. As indicated by Srivastava 1978) Greenland generally moved to the NE some 250 km. The geometry of the situation indicates a 60–100 km ambiguity (Fig. 3). Some land geologists can tolerate a 50–100 km offset (Christie et al. 1981). We suggest the remaining kilometres were accommodated by the expansion of the NWT platelet by the process of crustal thinning in Lancaster and Jones Sounds and in the Sverdrup Basin.

References

Balkwill, H. R. 1978. Evolution of Sverdrup Basin, Arctic Canada. – Bull. Am. Ass. Petrol. Geol. 62: 1004–1028.

Balkwill, H. R. & Bustin, R. M. 1978. Late Cretaceous and Tertiary structures, Queen Elizabeth Islands, Arctic Canada, and plate motions of Arctic North America. – Trans. Am. geophys. Un. 59: 362 only.

Bally, A. W. 1976. Canada's passive continental margins – a review. – J. Mar. geophys. Res. 2: 327–340.

Brooks, C. K. 1973. Rifting and doming in southern East Greenland. – Nature, Lond. 244: 23–24.

Brooks, C. K. 1979. Geomorphological observations at Kangerdlugssuaq, East Greenland. – Meddr Grønland, Geosci. 1: 24 pp.

Burke, K. 1976. Development of graben associated with the initial ruptures of the Atlantic Ocean. – Tectonophysics 36: 93–112.

Christie, R. L. 1964. Geological reconnaissance of northeastern Ellesmere Island, District of Franklin. – Mem. geol. Surv. Can. 331: 79 pp.

Christie, R. L., Dawes, P. R., Frisch, T. [O.], Higgins, A. K., Hurst, J. M., Kerr, J. W. & Peel, J. S. 1981. Geological evidence against major displacement in the Nares Strait. – Nature, Lond. 291: 478–480.

Churkin, M., Nokleberg, W. J. & Hull, C. 1979. Collision-deformed Paleozoic continental margin, western Brooks Range, Alaska. – Geology 7: 379–383.

Dawes, P. R. 1973. The North Greenland fold belt: a clue to

the history of the Arctic Ocean basin and the Nares Strait lineament. – In: Tarling, D. H. & Runcorn, S. K. (eds), Implications of continental drift to the earth sciences 2: 925–947. – Academic Press, London & New York.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R. & Peel, J. S. 1981. The northern margin of Greenland from Baffin Bay to the Greenland Sea. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 201–264. – Plenum Press, New York & London.

Dickinson, W. R. & Grantz, A. 1968. Proceedings of conference on geologic problems of San Andreas Fault System. – Stanford Univ. Pub. Geol. Sci. XI: 650 pp.

Feden, R. H., Vogt, P. R. & Fleming, H. S. 1979. Magnetic and bathymetric evidence for the "Yermak hot spot" northwest of Svalbard in the Arctic Basin. – Earth planet. Sci. Lett. 44: 18–38.

Grant, A. C. 1980. Problems with plate tectonics: the Labrador Sea. – Bull. Can. Petrol. Geol. 28: 252–278.

Grant, A. C. 1982. Problems with plate tectonic models for Baffin Bay – Nares Strait: evidence from the Labrador Sea. – This volume.

Harland, W. B. 1969. Contribution of Spitsbergen to understanding of tectonic evolution of North Atlantic region. – In: Kay, M. (ed.), North Atlantic – geology and continental drift. A symposium. – Mem. Am. Ass. Petrol. Geol. 12: 817–851.

Harland, W. B. 1973. Tectonic evolution of the Barents Shelf and related plates. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 599–608.

Harris, A. L., Johnson, M. R. W. & Powell, D. 1978. The orthotectonic Caledonides (Moines and Dalradians) of Scotland. – Pap. geol. Surv. Can. 78–13: 78–85.

Heirtzler, J. R., Dickson, G. O., Herron, E. M., Pitman, W. C. & Le Pichon, X. 1968. Marine magnetic anomalies, geomagnetic field reversals, and motions of the ocean floor and continents. – J. geophys. Res. 73: 2119–2136.

Hinz, K., Schlüter, H.-U., Grant, A. C., Srivastava, S. P., Umpleby, D. & Woodside, J. 1979. Geophysical transects of the Labrador Sea: Labrador to southwest Greenland. – Tectonophysics 59: 151–183.

Johnson, G. L., Taylor, P. T., Vogt, P. R. & Sweeney, J. F. 1978. Arctic Basin morphology. – Polarforschung 48: 20–30.

Johnson, G. L., Monahan, D., Grønlie, G. & Sobczak, L. [W.] 1979. General bathymetric chart of the oceans (GEBCO), 1:6 000 000, Sheet 5.17, The Arctic Ocean. – Can. Hydrogr. Serv., Ottawa.

Kerr, J. W. 1980. Did Greenland drift along Nares Strait? – Bull. Can. Petrol. Geol. 28: 279–289.

Kerr, J. W. 1981. Evolution of the Canadian Arctic Islands: a transition between the Atlantic and the Arctic Oceans. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 105–199. – Plenum Press, New York & London.

Kovach, R. L. & Nur, A. 1973. Proceedings of the conference on tectonic problems of the San Andreas Fault System. – Stanford Univ. Pub. Geol. Sci. XIII: 650 pp.

LaBrecque, J. L., Kent, D. V. & Cande, S. C. 1977. Revised magnetic polarity time scale for Late Cretaceous and Cenozoic time. – Geology 5: 330–335.

McKenzie, D. P. & Morgan, W. J. 1969. Evolution of triple junctions. – Nature, Lond. 224: 125–133.

Monahan, D. & Johnson, G. L. 1980. Geomorphology of the inland waterways of the Northwest Territories, Canada, and adjacent oceanic basins. – Abs. Intern. Alfred Wegener Symp.: 154 only. – Dietrich Reimer, Berlin.

Morgan, W. J. 1971. Convection plumes in the lower mantle. – Nature, Lond. 230: 42–43.

Morgan, W. J. 1972. Deep mantle convection plumes and plate motions. – Bull. Am. Ass. Petrol. Geol. 56: 203–213.

Newman, G. W., Mull, C. G. & Watkins, N. D. 1977. Northern Alaska paleomagnetism, plate rotation and tectonics. – Alaska geol. Soc. Symp., Anchorage, Alaska: 16–19.

Scheidegger, A. E. 1979. Orientationsstruktur der Talonlagen in der Schweiz. – Geographica Helvetica 1: 9–15.

Sclater, J. G. & Christie, P. A. F. 1980. Continental stretching: an explanation of the post-mid-Cretaceous subsidence of the central North Sea Basin. – J. geophys. Res. 85: 3711–3739.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Srivastava, S. P. & Falconer, R. K. H. 1982. Nares Strait: a conflict between plate tectonic predictions and geological interpretation. – This volume.

Sweeney, J. F., Coles, R. L., DeLaurier, J. M., Forsyth, D. A., Irving, E., Judge, A. S., Sobczak, L. W. & Wetmiller, R. J. 1978a. Arctic geophysical review – a summary. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 101–108.

Sweeney, J. F., Irving, E. & Geuer, J. W. 1978b. Evolution of the Arctic Basin. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 91–100.

Taylor, P. T. 1978. Low-level aeromagnetic data across the western Arctic Basin. – Trans. Am. geophys. Un. 59: 268–269.

Taylor, P. T., Kovacs, L. C., Vogt, P. R. & Johnson, G. L. 1981. Detailed aeromagnetic investigation of the Arctic Basin, 2. – J. geophys. Res. 86: 6323–6333.

Trettin, H. P., Frisch, T. O., Sobczak, L. W., Weber, J. R., Niblett, E. R., Law, L. K., DeLaurier, J. M. & Whitham, K. 1972. The Innuitian Province. – In: Price, R. A. & Douglas, R. J. W. (eds), Variations in tectonic styles in Canada. – Spec. Pap. geol. Ass. Can. 11: 83–179.

Vogt, P. R., Kovacs, L. C., Johnson, G. L. & Feden, R. H. 1979a. The Eurasia Basin. – Proc. Norwegian Sea symposium, 1979, NSS–3: 1–29. – Norsk Petroleum Forening, Oslo.

Vogt, P. R., Taylor, P. T., Kovacs, L. C. & Johnson, G. L. 1979b. Detailed aeromagnetic investigation of the Arctic Basin. – J. geophys. Res. 84: 1071–1089.

Whiteman, A., Naylor, D., Pegrum, R. & Rees, G. 1975. North Sea troughs and plate tectonics. – Tectonophysics 26: 39–54.

Ziegler, P. A. 1978. North Sea rift and basin development. – In: Ramberg, I. B. & Neumann, E. R. (eds), Tectonics and geophysics of continental rifts: 249–277. – Reidel Publishing Co., Boston.

Postglacial emergence along northern Nares Strait

JOHN ENGLAND

England, J. 1982. Postglacial emergence along northern Nares Strait. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 65–75.

During the last glaciation much of northern Nares Strait remained an ice-free corridor separating the northeast Ellesmere Island and northwest Greenland ice sheets. The disproportionate size of these ice sheets resulted in the lithosphere being differentially loaded on either side of this prominent rift valley. Postglacial emergence in this area is analyzed in order to determine whether glacio-isostatic unloading engendered any abnormal displacement along the Nares Strait fault zone. Present data suggest that synchronous shorelines dated 6000, 7000 and 8000 BP rise from north to south across northeast Ellesmere Island and northern Nares Strait towards the Greenland ice sheet. This is considered to represent the glacio-isostatic dominance of the Greenland ice sheet during the last glaciation together with earlier postglacial emergence towards northernmost Ellesmere Island which lay beyond the influence of the Greenland ice sheet. This Greenland dominance indicates that northeast Ellesmere Island lay in the depression marginal to the Greenland ice sheet. This, in turn, requires a lithospheric flexural parameter extending in an undisrupted manner across the Nares Strait rift valley. Hence, on a regional scale, it appears that the lithosphere in this area has integrated the depressions from these separated ice sheets without any observable unconformities along Nares Strait. Although postglacial faulting has followed initial glacio-isostatic unloading in other areas the present data base does not have the resolution to document similar events along Nares Strait.

J. England, Department of Geography, University of Alberta, Edmonton, Alberta, Canada, T6G 2H4.

Regional reconstructions of postglacial isobases have long been used for determining the distribution, magnitude and subsequent unloading of ice sheets from the last glaciation (cf. Andrews 1970, 1974). The relationship between former ice sheets and the vertical displacement of the lithosphere and asthenosphere is seen today in ongoing relative sea level adjustments and in negative free air gravity anomalies, both of which show that many areas are not in isostatic equilibrium (Walcott 1972a). In addition to postglacial isobases, postglacial emergence curves document the vertical displacement of specific sites through time following glacial unloading. Collectively this information has been increasingly used to develop geophysical models which attempt to duplicate and predict the rheological responses to different ice sheet histories (Broecker 1966, Brotchie & Silvester 1969, Walcott 1970, 1972b, Farrell & Clark 1976, Peltier 1976, Peltier & Andrews 1976, Clark et al. 1978, Quinlan in press). In most areas glacial unloading is thought to have produced a relatively gentle and continuous response in the lithosphere, resulting in a smooth net emergence of the land within and along the former ice sheet margin (Clark et al. 1978). In a few areas, however, glacial unloading was accompanied by faulting, most notably documented in Sweden (Lundqvist & Lagerbäck 1976, Lagerlund 1977, Mörner 1978). Postglacial faulting along Nares Strait has not yet

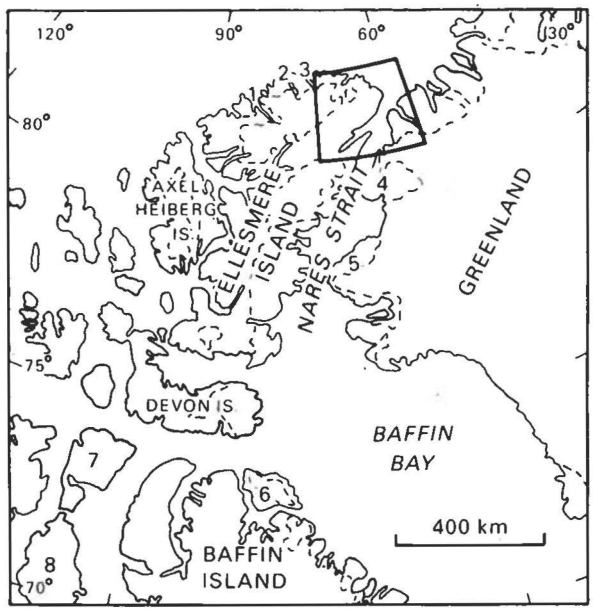


Fig. 1. Map of eastern Queen Elizabeth Islands and northwestern Greenland showing the location of the study area (inset) in the Nares Strait region. Main ice limits shown. 1 = Yelverton Inlet, 2 = M'Clintock Inlet, 3 = Ward Hunt Island, 4 = Washington Land, 5 = Inglefield Land, 6 = Bylot Island, 7 = Somerset Island, 8 = Boothia Peninsula.

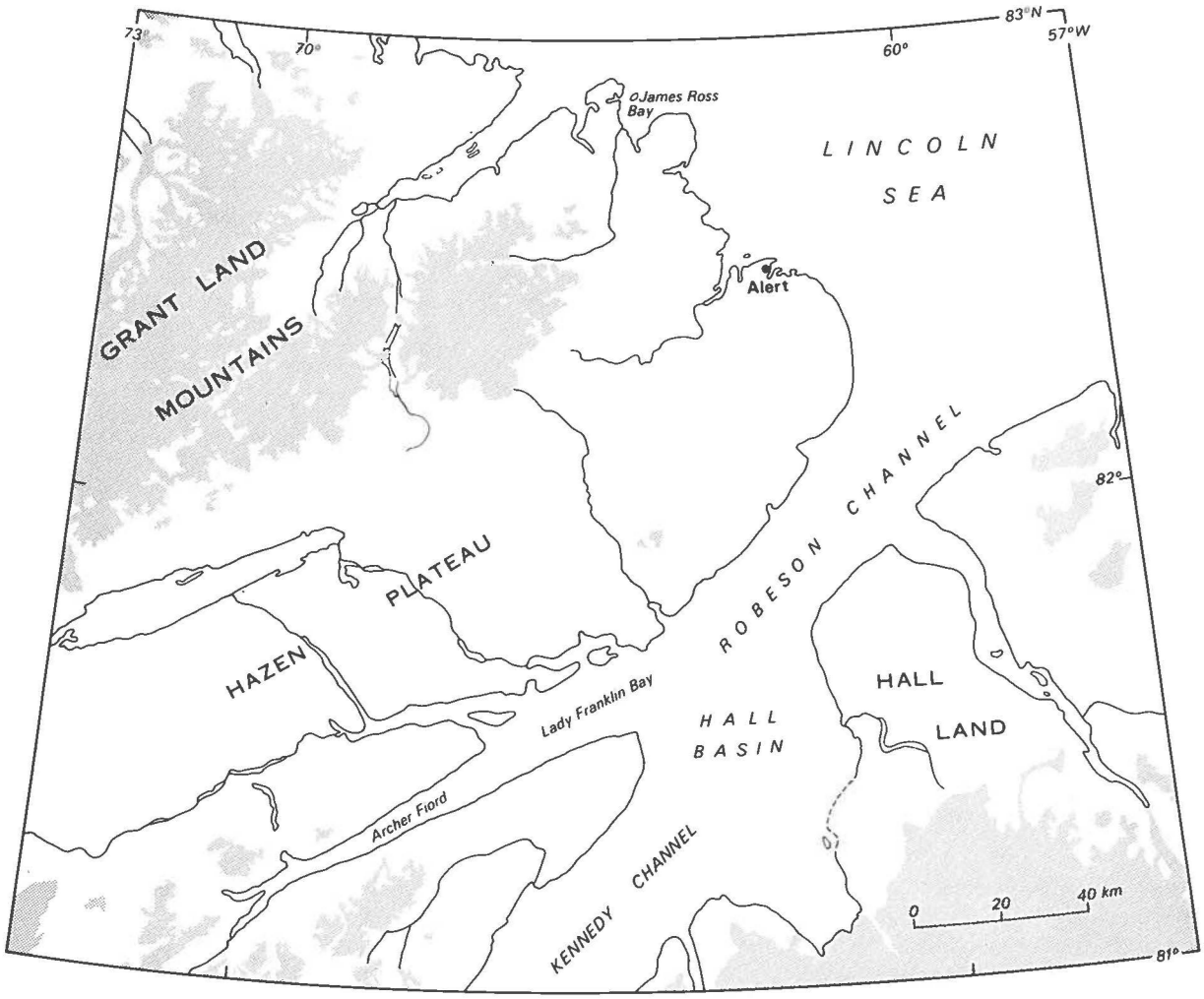


Fig. 2. Map of the northern Nares Strait study area showing contemporary ice sheets and outlet glaciers (stipled areas).

been reported although it is a major tectonic structure which some scientists consider has undergone hundreds of kilometres of displacement during its evolution (cf. summary by Kerr 1980). This paper discusses postglacial emergence along northern Nares Strait with particular emphasis on the problems and potential of the present data base as it pertains to the general tectonics of the area (Figs 1 and 2).

Along northern Nares Strait, Ellesmere Island and Greenland attain their greatest proximity separated by only 24 to 40 km. In this area their contemporary ice sheet margins are only ~150 km apart (Fig. 3). Under normal lithospheric conditions ice sheets generate ice-marginal depressions of ~180 km beyond which is a low amplitude forebulge where crustal displacement is positive (Brotchie & Silvester 1969, Walcott 1970). Assuming a normal 'crust' under the Lincoln Sea (Trettin et al. 1972 and Fig. 2), the present marginal depressions caused by the Ellesmere Island and Greenland ice sheets intersect in the area of Hazen Plateau and Lady Franklin Bay (England 1976 and Fig. 2). During the late Wisconsin/Würm glaciation the Ellesmere Island and Greenland ice sheets did not coalesce along northern Nares Strait, and hence an ice-free corridor existed along this seaway (England 1976, 1978, England & Bradley 1978, England et al. 1978, 1981). Consequently, northern Nares Strait is an ideal area for investigating the glacio-isostatic interactions of two ice sheet systems which differentially loaded the lithosphere on either side of this prominent rift valley (Kerr 1967). Deglaciation of the areas bordering the ice-free corridor began ~8000 to 8400 BP based on evidence from the Ellesmere Island side. The problem of concern here is whether glacio-isostatic flexing of the crust engendered any abnormal movement along the northern Nares Strait rift valley following the last glaciation.

Previous results

Several authors have drawn regional isobases for the High Arctic that are shown to extend uniformly from northwest Greenland into the eastern Queen Elizabeth Islands (Andrews 1970, Walcott 1972a, England 1976, Weidick 1976). Although the absolute values and orientations of the isobases differ from study to study, no abrupt break, hinge line, or saddle has been suggested for the northern Nares Strait area. In addition, England (1976) drew isobases on the 7500 and 6000 BP shorelines along northeast Ellesmere Island that indicated a rise in these shorelines from northwest to southeast towards the present margin of the Greenland ice sheet. The gradients of these shorelines indicate: 1) the glacio-isostatic dominance during the last glaciation of the Greenland ice sheet to the southeast and 2) the lack of abrupt differential movement along either the Alert geomagnetic anomaly (Praus et al. 1971) or the Nares Strait rift valley (Kerr 1967). Finally, because the Greenland ice sheet did not cover northeast Ellesmere Island during the last glaciation, its glacio-isostatic dominance requires that northeast Ellesmere Island lay in the zone of depression marginal to the Greenland ice sheet. This in turn requires some type of lithospheric "flexural parameter" (Walcott 1970) which extended in an undisrupted manner across the Nares Strait rift valley.

England (1974) used a static geophysical model (modified from Brotchie & Silvester 1969) that simulated the isostatic effects of possible ice loads during the last glaciation as suggested by the geological data bordering northern Nares Strait. The model assumes a thin, elastic, spherical lithosphere overlying a viscous asthenosphere. The geological parameters which could be varied include the number of ice sheets, their locations, thicknesses, radii and rates of retreat (all of which are little known in detail). The geophysical parameters which could be varied include the lithosphere's radius of relative stiffness (similar to Walcott's flexural parameter) and the density of the asthenosphere. In the minimum ice load model, which assumed the aforementioned ice-free corridor, the northwest Greenland ice sheet was assigned a radius of ~500 km and a maximum thickness of 3200 m (compared to a radius of 400 km and a maximum thickness of 2500 m today, Weidick 1971). On the other hand, the Ellesmere Island ice caps were assigned a combined radius of ~180 km and a maximum thickness of 1500 m (compared to a combined radius of 120 km and a maximum thickness of 900 m today, Hattersley-Smith et al. 1969). Using a 'normal' flexural parameter of ~180 km and the restricted ice margins on northwest Greenland (Davies 1972, Weidick 1972) and northeast Ellesmere Island (England 1978) it was found that the predicted emergence from this model was in good agreement with the pattern and magnitude of the observed postglacial emergence. Although more sophisticated geophysical models are now available for predicting postglacial emergence (Clark 1977, Clark et al. 1978) the static depressions estimated by the Brotchie & Silvester model are still considered reasonable and they simply suggest that solutions based on a normal lithosphere in this area are presently consistent with the glacio-isostatic evidence. More recent evidence suggesting that northern Nares Strait is aseismic (Praus et al. 1971, Basham et al. 1977, Wetmiller & Forsyth, this volume) may also indicate a stable lithosphere whose large-scale activity is relatively uninfluenced by its numerous surface faults (cf. Kerr 1967, Trettin 1971).

Existing problems

The most outstanding problems for the analysis of glacio-isostatic adjustments along northern Nares Strait are those of: 1) limited control points; 2) the establishment of marine limits in the marginal depression and 3) the comparability of some radiocarbon age determinations. These problems must first be resolved before postglacial isobases can be accurately drawn and interpreted in terms of recent tectonic activity in the area. The present data base contains some inconsistencies which could be attributable to any of these problems and hence the question of whether or not there has been



Fig. 3. Visual satellite image (NOAA–6) of the Nares Strait region taken in late July, 1979. The disproportionate glacial loading presently bordering Nares Strait is clearly exhibited by the northwest Greenland ice sheet to the right and the smaller Ellesmere Island ice caps to the left. Top left ice field occurs on Axel Heiberg Island and lower left ice field occurs on Bylot Island bordering the 'North Water' polynya of northern Baffin Bay.

Holocene faulting in northern Nares Strait cannot be definitely answered at this time.

Scarcity of data

The most obvious limitation to analyzing postglacial emergence along northern Nares Strait (north of 81° latitude) is the scarcity of control points, particularly on the Greenland side where only five ^{14}C dates are available, four of which come from Hall Land (Fig. 2; Weidick 1972, 1978). By comparison over one hundred ^{14}C dates are now available from the adjacent coast of Ellesmere Island. Due to this uneven distribution in the data base, a number of different isobase patterns can be drawn (Weidick 1976) and these can only be resolved by further field work.

Recognition of marine limits in zone of marginal depression

Relatively little attention has been given to the problems of marine sedimentation in the marginal depression of former ice sheets (England 1978). As suggested above, considerable areas along the northeast coast of Ellesmere Island remained ice-free during the last glaciation. Today, this polar desert landscape is characterized by numerous, small streams with little ability to erode and transport sediments since the annual water balance is close to zero (Bovis & Barry 1974). During the last glaciation these same streams provided so little sediment to the sea that the marine limit was not well recorded on the landscape in most places. In addition, many of these streams discharge into deep marine basins, such as Archer Fiord and Lady Franklin Bay (Fig. 2), that effectively dispersed these small sediment inputs. Consequently, many of these watersheds beyond the glacial limit produced small, submarine fan deltas that never aggraded to the marine limit. Marine fossils from these deltas, therefore, often relate to sea level positions considerably above these delta surfaces. An example of this occurs at site A, Fig. 4, where a ^{14}C date of 7910±145 BP (DIC–545) was obtained on marine shells from a non-graded fan delta which rises to 60 m a.s.l. in a small, ice-free watershed. In a larger watershed, twenty kilometres to the southwest (site B, Fig. 4), marine shells from a 90 m terrace provided a similar date (8200±260 BP, DIC–549). This suggests that the date of 7910±145 BP likely relates to a sea level position considerably higher than 60 m. Other examples could be cited where anomalously old dates were obtained from the foreset beds of relatively low deltas and these in turn are considered to represent inadequate sedimentation in certain sectors of the marginal depression rather than differential emergence or local tectonics. Finally, sedimentation in the marginal depression may not necessarily have coincided with initial glacial unloading but rather with climatically induced fluvial responses later in the Holocene. Hence strongly developed delta fans below the marine limit may be mistaken for the uppermost extent of the sea.

Radiometric problems

Another problem with the existing data base pertains to inconsistencies in radiocarbon age determinations provided by different laboratories. An initial date collected from the marine limit at site C, Fig. 4, was 7500±140 BP (St 4089, uncorrected date) and the ^{13}C correction increased the age determination to 7710±140 BP. Subsequent assessments, uncorrected for ^{13}C, from the same sample are 6845±70 BP (SI–4029) and 6790±410 BP (UQ–134, the large standard error is due to the small sample analyzed). A marine limit at the same elevation 12 km to the northwest (site D, Fig. 4) also dated 6860±70 BP (SI–3300) and these consistent dates on the same relative sea level are given precedence over the original determination (St 4089). Also an uncorrected date on marine shells from the surface of a 110 m terrace at Cape Baird (site E, Fig. 4) was 7025±150 BP (St 4099, 7565±150 BP, corrected for ^{13}C) and this contrasts with marine shells from the topset beds of the same terrace that dated 8380±150 BP (DIC–737, uncorrected) which is far more consistent with other dates on initial emergence in this area (England & Bradley 1978). For the most part these radiometric problems have now been eliminated by the recollection and dating of critical samples and they can likely be avoided by adequate pre-treatment of future analyses.

Recent data

Additional control points obtained since the initial construction of the local postglacial isobase maps (England 1976) warrant a re-evaluation of these data. However, these results remain provisional as many new field collections are presently submitted for ^{14}C analyses and a more detailed presentation is planned.

Initial emergence: Alert and James Ross Bay

Along the north coast of Ellesmere Island there are only a few ^{14}C dates available on shorelines at or near the marine limits. Most of these occur to the west of the study area (Fig. 4, see Fig. 1) such as Ward Hunt Island (Christie 1967, Lyons & Mielke 1973); M'Clintock Inlet (Christie 1967) and Yelverton Inlet (Blake 1972). Near Alert, which borders northernmost Nares Strait (Fig. 2), a marine limit at ~120 m a.s.l. dated 10100±210 BP (GSC–1815). Although this date is still an exception amongst >100 ^{14}C age determinations along the southern Hazen Plateau, ~110 km to the southwest (Fig. 2), it may indicate earlier emergence in the north as was originally suggested by Christie (1967) to explain the apparently higher marine limits there. In

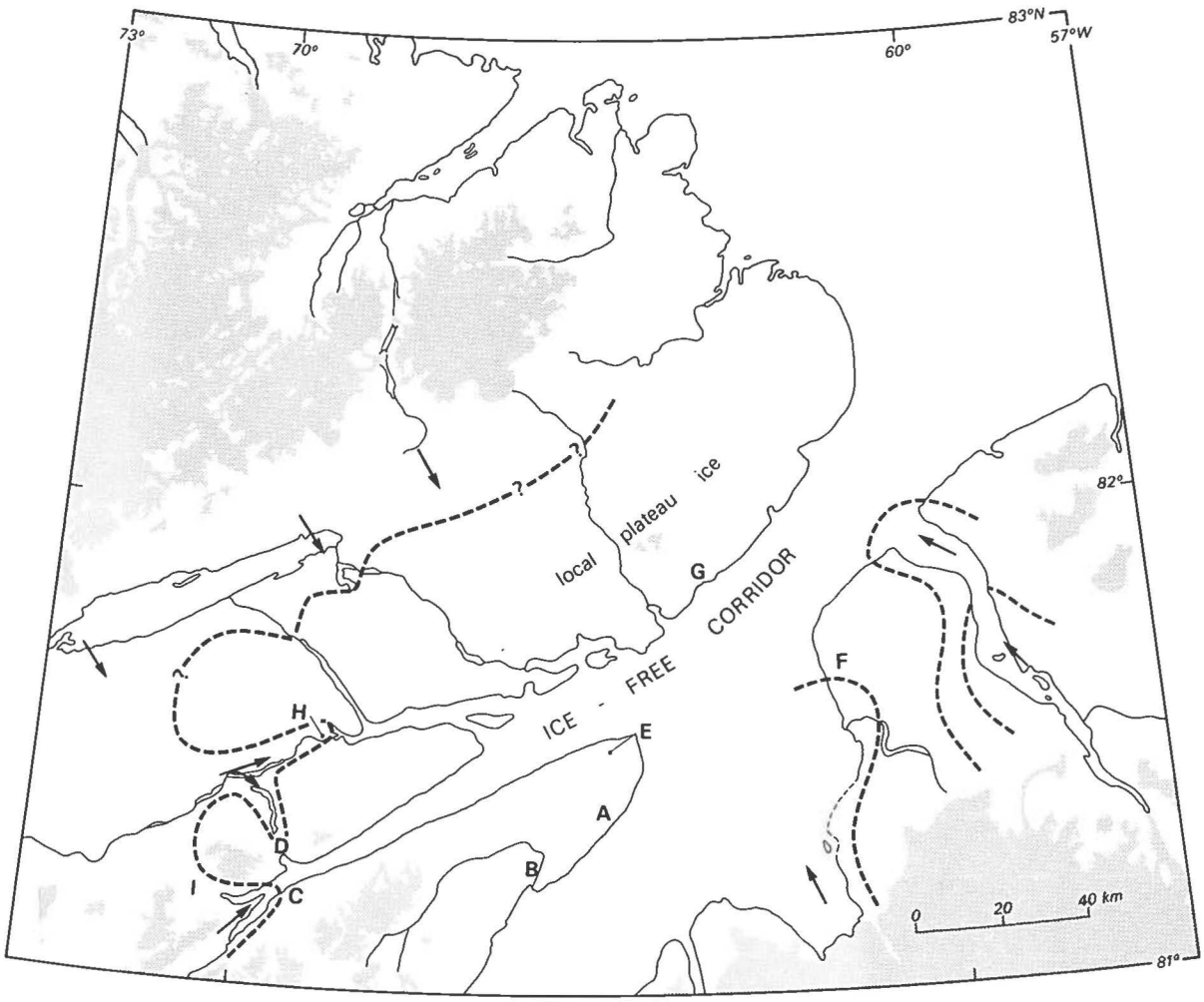


Fig. 4. Provisional limit of the last glaciation (dashed line) on northeast Ellesmere Island and northwest Greenland. These ice limits suggest that much of northern Nares Strait remained ice-free at this time. Site locations discussed in text are also shown. Arrows indicate former ice flow directions.

addition, a shell sample from the foreset beds of a 78 m delta inland from James Ross Bay (Fig. 2) recently dated 8280±300 BP (GSC–3002) and this occurs topographically below undated marine deposits at 90 m a.s.l. which must be older still.

Archer Fiord and Lady Franklin Bay

Compared to the aforementioned dates from the north coast of Ellesmere Island, the oldest available dates on initial postglacial emergence throughout Archer Fiord and Lady Franklin Bay, as well as along western Kennedy and Robeson Channels (Fig. 2), are ~8000 to 8400 BP (England 1978, England & Bradley 1978). Since similar dates on initial emergence extend inland from the northeast Ellesmere Island coast to the approximate margin of the last glaciation (a distance of ~75 km, Fig. 4), this zone of synchronous emergence (~8000 to 8400 BP) was interpreted to mark a former ice-marginal depression unloaded between the separated northeast Ellesmere Island and northwest Greenland ice sheets (England 1978). Any similar ice-free areas along the margin of the northwest Greenland ice sheet (which was presumably controlling this depression) also should have started emerging at the same time. This suggestion is indirectly supported by a date on marine shells from a terrace at 82 m a.s.l. from western Hall Land (site F, Fig. 4; 6100±300 BP, W–816; W. E. Davies in Rubin & Alexander 1960). This date indicates that the local marine limit, which is ~110 m a.s.l. (Weidick 1972), could be at least as young as ~8000 BP (England 1976).

Western Hall Land

More recently, Weidick (1978) has reported three additional ^{14}C dates from western Hall Land, two of which conflict with age assessment of W–816 (6100±300 BP). One of these dated samples (Weidick 1978: 121) was collected from "silt on top of a marine terrace at 80 m a.s.l." (9000±145 BP, I–9690) and the other came from "clayey silt under a veneer of shingles" at 85 m a.s.l. (9180±150 BP, I–9687). Weidick interprets these dates as part of a sea level regression following deglaciation that he estimates to have occurred at or before 9500 BP which would also date the local marine limit at 110 m a.s.l. If this interpretation is correct it seriously contradicts the isobases of this and previous studies (England 1976, Weidick 1976) as the estimated postglacial emergence since 6000 BP on western Hall Land is reduced to ~25 m (Weidick 1978) compared to values of 60–80 m (to be discussed). However, given this interpretation the 6000 BP shoreline on western Hall Land (provisionally 25 m a.s.l.) does not tilt up towards Washington Land (Weidick 1978) but rather towards the coast of Ellesmere Island where shorelines of the same age range from 60–80 m a.s.l. It is considered unlikely that the combined ice loads bordering northern Nares Strait, which are far larger on the Greenland side (Fig. 3), could produce such an abrupt reversal in the isobase gradients shown in Figs 5–7 unless postglacial faulting occurred.

An alternative interpretation of these ~9000 BP dates on western Hall Land, which would not contradict the present isobases, is also possible. I suggest that these older samples may record the transgression to the marine limit which in turn experienced an initial

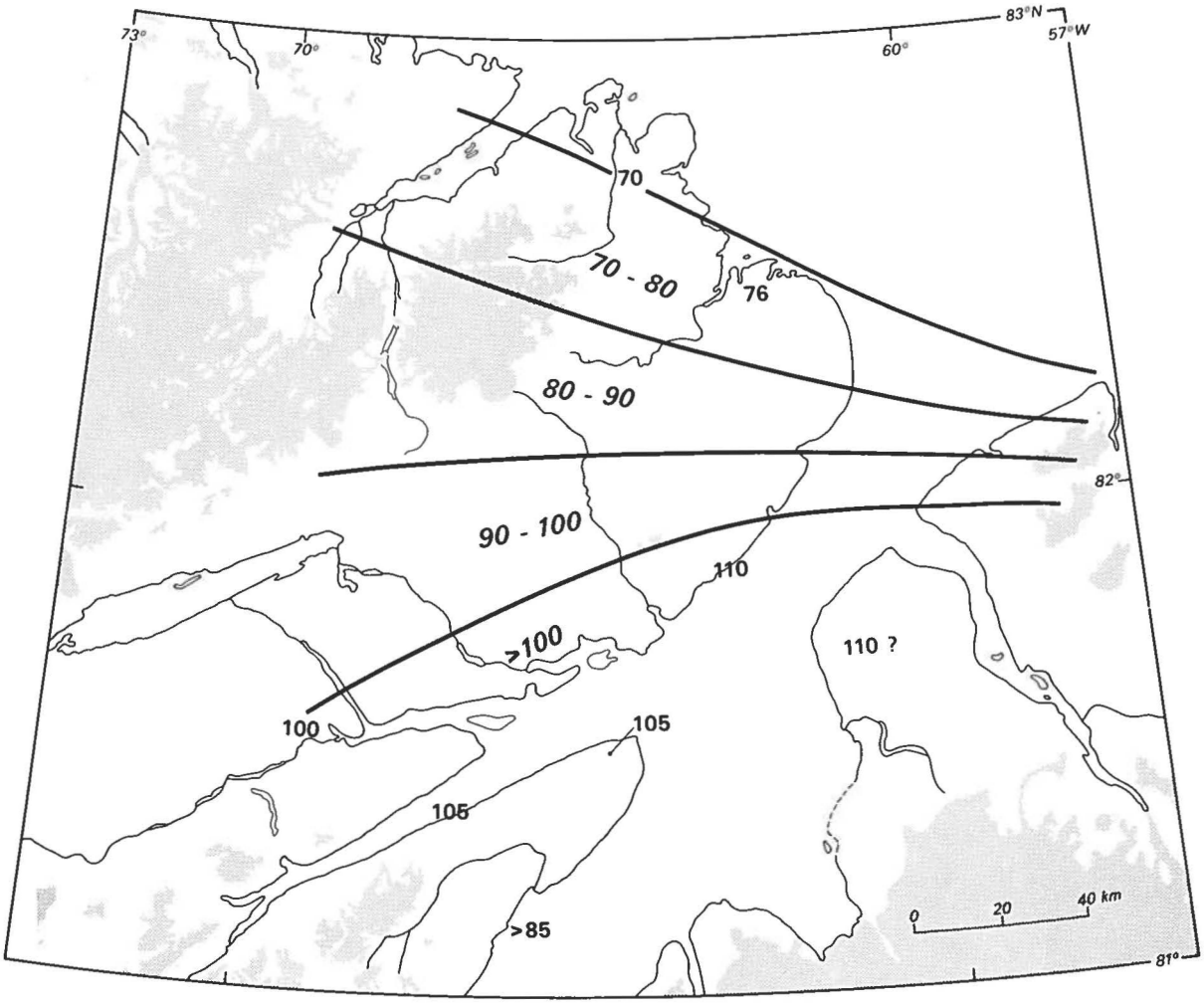


Fig. 5. Provisional zones of postglacial emergence (in metres above sea level) drawn for 8000 BP. Zones of emergence rather than precise isobases are shown in order to avoid over-representing the accuracy of the data. Control points are also shown in metres a.s.l.

emergence ~8000 to 8400 BP. This would also be compatible with a subsequent marine regression, following deglaciation, that deposited the 82 m shells dated at 6100±300 BP (W–816) in the same area. In this case it is of interest that directly across from this locality on western Hall Land, on the Ellesmere Island side of Robeson Channel, a marine limit at a similar elevation (115 m a.s.l.) is recently dated at 8050±120 BP (GSC–3041, site G, Fig. 4). Until the details of the raised marine stratigraphy on western Hall Land are determined (i.e. transgressive vs. regressive deposition) alternative interpretations of the postglacial isobases cannot be excluded (Weidick 1976). It is apparent, however, that within the ice-free corridor, which separated the northern Ellesmere Island and Greenland ice sheets during the last glaciation, there must be older marine deposits (ca. >9000 BP) which record the transgression to the marine limits dated 8000–8400 BP.

Postglacial emergence

Figs 5–7 show zones of postglacial emergence drawn along northern Nares Strait for 8000, 7000 and 6000 BP. All three diagrams show a progressive rise in synchronous shorelines from north to south across northeast Ellesmere Island towards northwest Greenland. If only the northern Ellesmere Island ice load is considered in relation to this differential emergence it is apparent that a radial expansion of the present day Grant Land Mountains ice cap (Figs 2 and 3) would result in a roughly uniform glacio-isostatic loading around its former margin, i.e. a similar amount of displacement both to the north and south of its centre. Consequently, the increasing postglacial emergence to the southeast (Figs 5–7) cannot be explained by a differential outflow from this central ice cap towards Nares Strait since the emergence on the southern Hazen Plateau would still

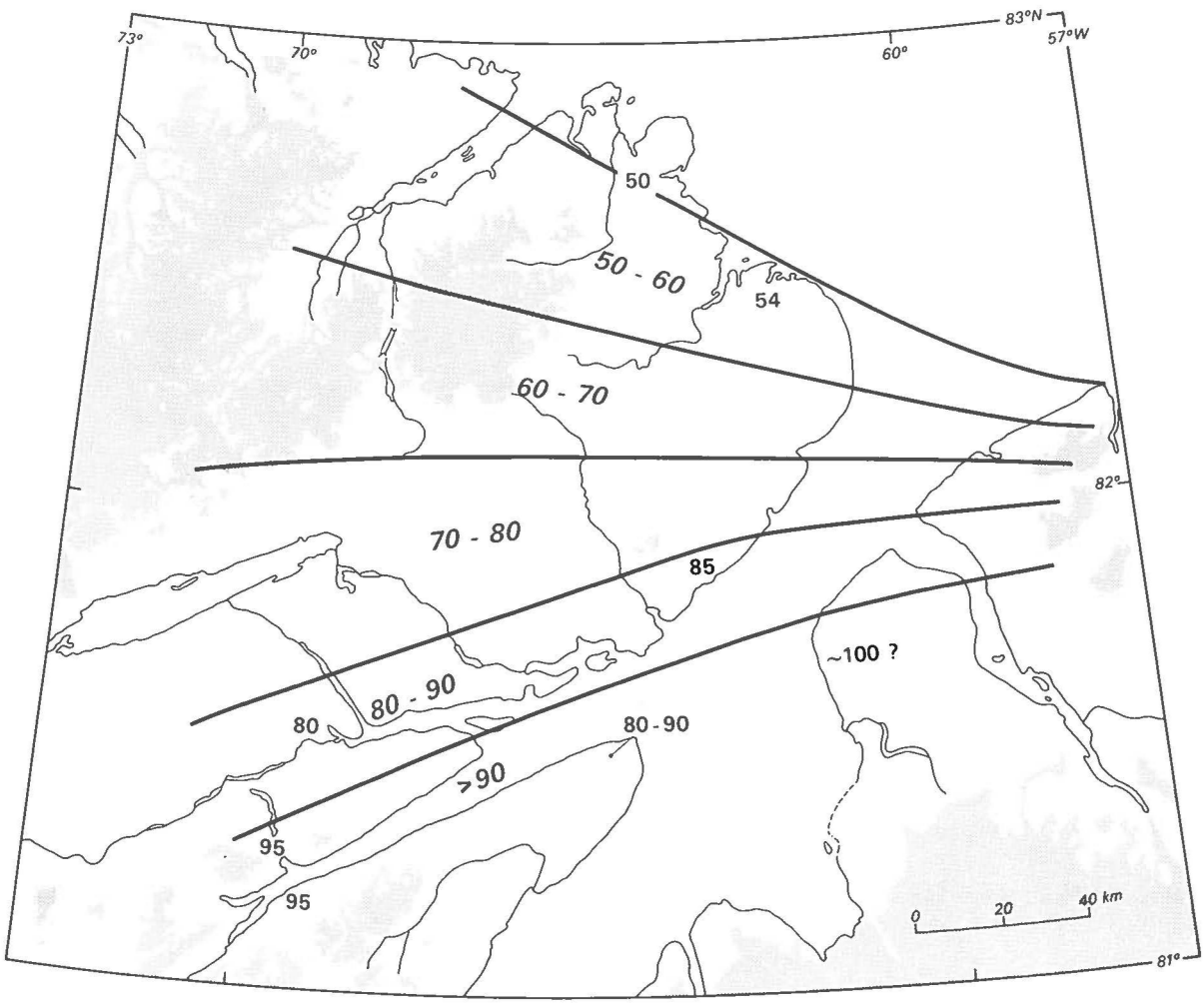


Fig. 6. Provisional zones of postglacial emergence drawn for 7000 BP. All elevations are in metres a.s.l.

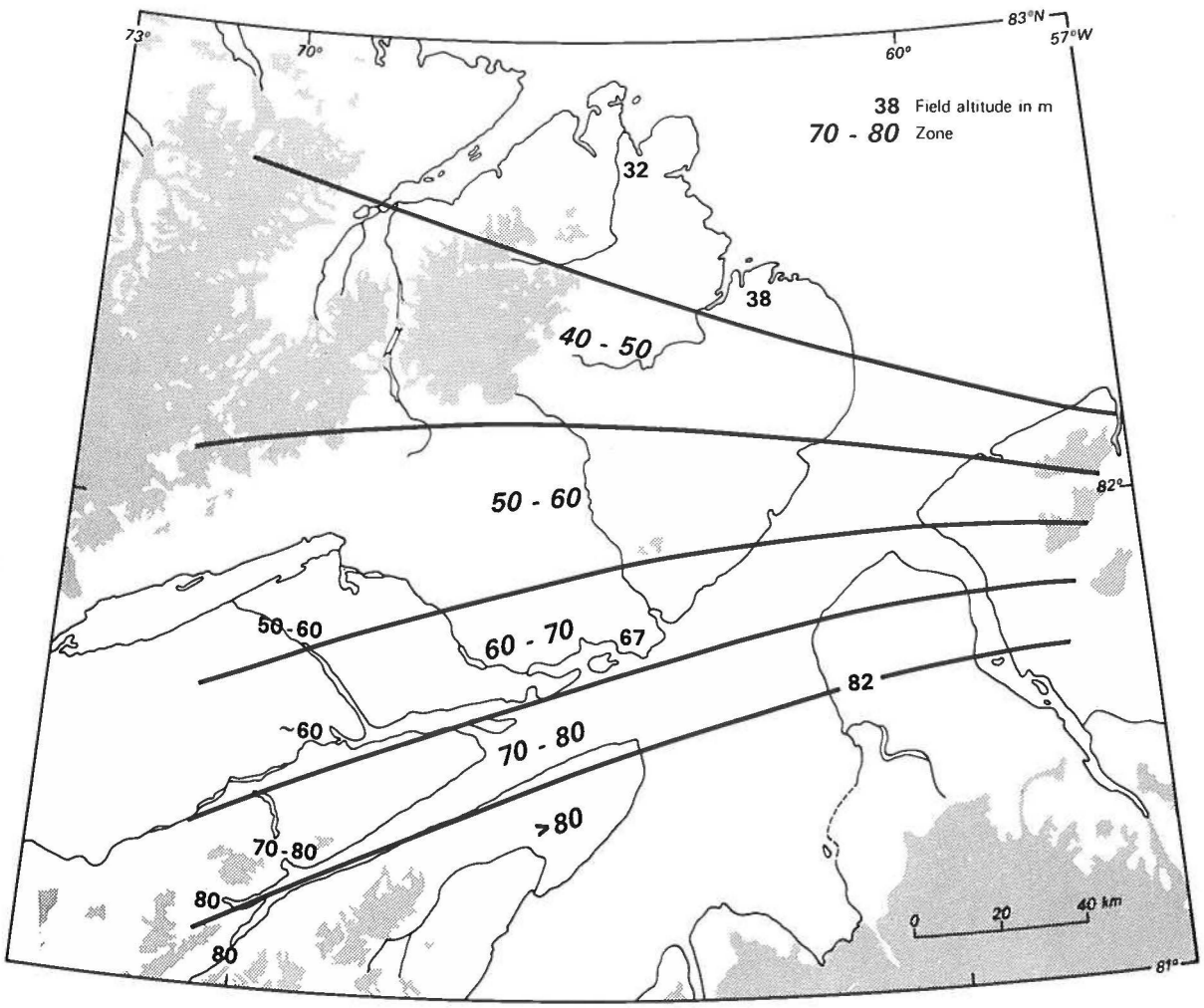


Fig. 7. Provisional zones of postglacial emergence drawn for 6000 BP. All elevations are in metres a.s.l.

have to be inclined toward the former ice source (i.e. to the north). On the other hand ice recession earlier along the northern margin than along the southern margin of this former ice cap might partly explain this differential emergence (cf. Christie 1967 and the ~10,000 BP Alert date, this paper). However, since postglacial emergence attains its maximum values along the approximate southeast margin of the last ice sheet on northeast Ellesmere Island (Figs 4 and 5), and since these sites also occur within a few tens of kilometres of the proposed Greenland ice sheet terminus in Hall Basin (Weidick 1972), it is proposed that the marginal depression from the Greenland ice also contributed to this southward rise of synchronous shorelines. Given the much greater size of the Greenland ice sheet, it is logical that the postglacial emergence on northeast Ellesmere Island is dominated by it (England 1976). The similarity in the generalized emergence in all three diagrams (Figs 5–7) reinforces this interpretation. It is apparent that the gradient of the isobases (Figs 5–7) becomes much less steep on northeast Ellesmere Island (away from northern Nares Strait) and this may suggest a more uniform glacio-isostatic loading in this area during the last glaciation.

The above hypothesis would argue that although there was a common ice sheet over a large part of northern Ellesmere Island, initial postglacial emergence along its southeastern flank (i.e. along the southern Hazen Plateau) was controlled by the recession of the northwest Greenland ice sheet (England 1976, England & Bradley 1978). On the other hand, the fjord systems along the northernmost coast of Ellesmere Island are presently >200 km from the Greenland ice sheet and hence their initial emergence would more likely record the independent recession of this local ice load which may not correspond to the initial emergence reported in the Archer Fiord–Lady Franklin Bay area to the southeast (Greenland induced, ~8000 to 8400 BP). Con-

sequently, the view of earlier ice recession on northernmost Ellesmere Island (Christie 1967) combined with the glacio-isostatic dominance of the Greenland ice sheet along northern Nares Strait (England 1974, 1976), may best explain the differential emergence shown in the Figs 5–7. This pattern of differential emergence is similar, both in its orientation and slope, to the previously constructed isobase maps for northeast Ellesmere Island (England 1976).

Discussion

On a regional scale it is presently maintained that the glacio-isostatic deformation of the lithosphere and asthenosphere by the northern Ellesmere Island and Greenland ice sheets has been integrated coherently without recognizable unconformities produced by structural weaknesses along the Nares Strait rift valley. Also on a regional scale, epicentral solutions for Arctic earthquakes generally suggest that Nares Strait is noticeably aseismic although the most detailed records only post-date 1962 (Praus et al. 1971, Basham et al. 1977, Wetmiller & Forsyth 1978). In addition, Niblett & Whitham (1970) found no heat flow anomalies on northern Ellesmere Island which also suggests a normal lithospheric thickness. Finally, although large-scale stability for this area is suggested by the geophysical and geological evidence (Kerr 1967, Dawes 1979, Wetmiller & Forsyth, this volume) it is also apparent that numerous faults and graben cross-cut the Phanerozoic sediments throughout the field area under discussion (Christie 1964, 1974, Trettin 1971). Along the continental shelf of eastern Canada, from western Baffin Bay to Nova Scotia, faults that have been mapped by seismic studies generally extend to depths of ~10 km (Beh 1975, Smith 1975, Keen & Hyndman 1979). If the faults adjacent to and including northern Nares Strait are similar in depth (~10 km) then they may be exerting only minor effects on the large-scale lithospheric response to former ice loads in the area.

In comparison to this perspective of regional stability, however, some small-scale anomalies in postglacial emergence could still occur along such faults where elastic stresses might be differentially released, particularly during the rapid phase of initial emergence (Mörner 1978). Basham et al. (1977) and Wetmiller & Forsyth (1978) have suggested that contemporary seismicity in the Baffin Island – Foxe Basin area and along the Boothia Uplift, respectively, may be the product of stresses triggered by glacio-isostatic unloading. In eastern Canada, Stein et al. (1979) have estimated that postglacial crustal flexure could attain stresses of 100 to 150 bars which they suggest is adequate to reactivate pre-existing faults. In southwestern Sweden, a vertical displacement of up to 100 m has been reported along some faults since the Late Weichselian based on till-stratigraphic studies (Lagerlund 1977). Most dramatically, in northern Sweden, the 150 km long Pärve Fault has been demonstrated to be of late-glacial age having a vertical displacement of ~10 m (Lundqvist & Lagerbäck 1976). Adams (1981) has discussed the effects of postglacial faulting in eastern Canada, however all the offset movements noted so far are small (<10 m and generally <2 m). Finally, postglacial offset has also been discussed for portions of the Denali Fault system in south-central Alaska (Lanphere 1978) and the possibility of postglacial tectonics has been raised for the Puget Sound Lowland, Washington State (Thorson 1979).

In Arctic Canada no direct evidence has been presented demonstrating the effects of recent tectonics on the pattern of postglacial emergence although several areas are seismically active (Basham et al. 1977). However, recent work by Dyke (1979) on northern Boothia Peninsula, mainland Canada, has suggested very rapid rates of initial emergence (>30 m/100 years compared with generally reported rates of <10 m/100 years; Andrews 1970, Blake 1975). Although instantaneous effects such as ice-water gravitational attraction and elastic rebound of the crust might partly explain this initial emergence (>30 m/100 years) it is also possible that the reactivation of faults also took place along the seismically active Boothia Uplift (A. S. Dyke, pers. comm. 1980).

On northeast Ellesmere Island the present data base does not have the resolution to decipher small-scale, fault-related anomalies in the local isobases. The one possible exception occurs between Simmonds Bay and the Dodge River (sites D and H, respectively, Fig. 4) along the southwestern margin of the Hazen Plateau. These two sites are only 25 km apart yet the 95 m marine limit at Simmonds Bay dated 6860±70 BP (SI–3300) whereas the 105 m marine limit in Dodge River dated 8130±200 BP (GSC–1775). In both cases significant errors in either the stratigraphy or the ^{14}C dating can be excluded due to repeated and consistent results in all of these analyses. Since the average rate of initial emergence previously estimated in this area is ~3 m/100 years (England 1976) it is apparent that the dated shoreline at Simmonds Bay (~1200 radiocarbon years younger than the Dodge River shoreline) is abnormally high for its age. This Simmonds Bay anomaly does not exist in isolation but rather it is consistent with similarly high, young shorelines at the fjord head, 15 km to the southwest (site I, Fig. 4). There, the marine limit at ~85 m a.s.l. dated 6430±150 BP (GSC–1614). This contrast between Simmonds Bay and Dodge River may suggest that some postglacial faulting has contributed to this differential emergence since both sites were similarly occupied by late Wisconsin ice. Alternatively, initial postglacial emergence in this intervening period (ca. 8200 to 6900 BP) was abnormally slow (<1 m/100 years). In conclusion, although the reactivation of faults by glacio-isostatic stresses cannot be excluded these ef-

fects are presently considered to be minimal and beyond detection by the available field data bordering northern Nares Strait.

Finally, if the postglacial isobases do cross from northwestern Greenland to northeast Ellesmere Island (Figs 5–7) then they must eventually close themselves and return to Greenland across southern Nares Strait (cf. England 1976). The precise configuration of these isobases is not presently known although Blake (1975) does show a northeast–southwest orientation for his 5000 BP isolines which extend across central Devon and southeast Ellesmere Islands in the direction of southern Nares Strait and Inglefield Land, northwest Greenland. England (1976) indicated that the highest isobase values (for 7500 and 6000 BP) on northeast Ellesmere Island also cross back to Greenland in the area of Inglefield Land. Although these two isobase patterns (Blake 1975, England 1976) show somewhat different orientations they are not necessarily mutually exclusive since the absolute values (for any given sea level) differ considerably between northeast and southwest Ellesmere Island. For example, the 6000 BP shoreline is displaced up to 60–80 m a.s.l. on northeast Ellesmere Island whereas on southwestern Ellesmere Island it declines to ca. 30–35 m a.s.l. (Blake 1975). Hence, although the actual isolines for 6000 BP vary in magnitude and orientation between these two areas they may still represent the overall isobase pattern which converges upon and crosses southern Nares Strait.

One complicating factor in reconstructing the postglacial isobases in southern Nares Strait pertains to the differential buildup of ice during the late Holocene (Koerner 1977). Southern Nares Strait borders the important moisture source of the North Water polynya (in northern Baffin Bay) and as a result adjacent glaciers on both northwest Greenland and southeast Ellesmere Island are favoured by greater accumulation (Mock 1968, Koerner 1977). Hence, the buildup of greater ice thicknesses in this area during the late Holocene may have suppressed postglacial emergence such that the isobases in northern Nares Strait have higher maximum values than their counterparts to the south. England (1976) alluded to this problem in southern Nares Strait since the maximum values for postglacial emergence on Inglefield Land appeared too low compared to those on northeast Ellesmere Island. Nichols (1969) has also suggested a late postglacial advance of the Greenland ice sheet on Inglefield Land which may account for this lower emergence. Therefore, if some asymmetry occurs in the postglacial emergence between northern and southern Nares Strait it is more likely the result of the glacio-climatic regime than postglacial faulting (which has not yet been documented).

Acknowledgements

Logistical support for this study was provided by the Polar Continental Shelf Project, Department of Energy, Mines and Resources, Ottawa. The research was also supported by grants from the Boreal Institute for Northern Studies, University of Alberta, and from the National Science and Engineering Research Council of Canada. Radiocarbon dates were provided by Dr. Robert Stuckenrath, Smithsonian Institute, Maryland, and by Dr. Weston Blake Jr., Geological Survey of Canada. Dr. A. S. Dyke, Geological Survey of Canada, provided a conscientious review on an earlier draft of the manuscript and Dr. Richard Lambert, Department of Geology, University of Alberta, helpfully discussed a number of points on the tectonics of Nares Strait.

References

Adams, J. 1981. Postglacial faulting in eastern Canada. – Atomic Energy of Canada Ltd., Technical Record Series, Ottawa.

Andrews, J. T. 1970. A geomorphic study of post-glacial uplift with particular reference to Arctic Canada. – Inst. Br. Geogr. Spec. Publ. 2: 157 pp.

Andrews, J. T. 1974. Glacio-isostasy. – Benchmark papers in geology 10: 491 pp. – Dowden, Hutchinson and Ross Inc., Stroudsburg, Pennsylvania.

Basham, P. W., Forsyth, D. A. & Wetmiller, R. J. 1977. The seismicity of northern Canada. – Can. J. Earth Sci. 14: 1646–1667.

Beh, R. L. 1975. Evolution and geology of western Baffin Bay and Davis Strait, Canada. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 453–476.

Blake, W. 1972. Climatic implications of radiocarbon-dated driftwood in the Queen Elizabeth Islands, Arctic Canada. – In: Vasari, Y., Hyvärinen, H. & Hicks, S. (eds), Climatic changes in Arctic areas during the last ten-thousand years. – Acta Univ. Oulu, Ser. A(3), Geol. 1: 77–104.

Blake, W. 1975. Radiocarbon age determinations and postglacial emergence at Cape Storm, southern Ellesmere Island, Arctic Canada. – Geogr. Annlr. Ser. A: 1–71.

Bovis, M. J. & Barry, R. G. 1974. A climatological analysis of north polar desert areas. – In: Smiley, T. L. & Zumberge, J. H. (eds), Polar deserts and modern man: 23–31. – The University of Arizona Press, Tucson, Arizona.

Broecker, W. S. 1966. Glacial rebound and the deformation of the shorelines of proglacial lakes. – J. geophys. Res. 71: 4777–4783.

Brotchie, J. F. & Silvester, R. 1969. On crustal flexure. – J. geophys. Res. 74: 5240–5252.

Christie, R. L. 1964. Geological reconnaissance of northeastern Ellesmere Island, District of Franklin. – Mem. geol. Surv. Can. 331: 79 pp.

Christie, R. L. 1967. Reconnaissance of the surficial geology of northeastern Ellesmere Island, Arctic Archipelago. – Bull. geol. Surv. Can. 138: 50 pp.

Christie, R. L. 1974. Northeastern Ellesmere Island: Lake Hazen and Judge Daly Promontory. – Pap. geol. Surv. Can. 74–1A: 297–299.

Clark, J. A. 1977. An inverse problem in glacial geology: the reconstruction of glacial thinning in Glacier Bay, Alaska between A. D. 1910 and 1960 from relative sea-level data. – J. Glaciol. 18: 481–503.

Clark, J. A., Farrell, W. E. & Peltier, W. R. 1978. Global changes in postglacial sea level: a numerical calculation. – Quaternary Res. 9: 265–287.

Davies, W. E. 1972. Landscape of northern Greenland. – Spec. Rep. Cold Reg. Res. & Engng. Lab. 164: 67 pp. – Hanover, New Hampshire.

Dawes, P. R. 1979. Precambrian and Palaeozoic development of northern Greenland. – Norsk Polarinst. Skr. 167: 321–324.

Dyke, A. S. 1979. Radiocarbon-dated Holocene emergence of Somerset Island, central Canadian Arctic. – Pap. geol. Surv. Can. 79–1B: 307–318.

England, J. 1974. The glacial geology of the Archer Fiord/Lady Franklin Bay area, northeastern Ellesmere Island, N. W. T., Canada. – Unpubl. Ph. D. thesis, Univ. of Colorado, Boulder, U.S.A.: 234 pp.

England, J. 1976. Postglacial isobases and uplift curves from the Canadian and Greenland High Arctic. – Arct. Alp. Res. 8: 61–78.

England, J. 1978. The glacial geology of northeastern Ellesmere Island, N. W. T., Canada. – Can. J. Earth Sci. 15: 603–617.

England, J. & Bradley, R. S. 1978. Past glacial activity in the Canadian High Arctic. – Science 200: 265–270.

England, J., Bradley, R. S. & Miller, G. H. 1978. Former ice shelves in the Canadian High Arctic. – J. Glaciol. 20: 393–404.

England, J., Bradley, R. S. & Stuckenrath, R. 1981. Multiple glaciations and marine transgressions along western Kennedy Channel, N. W. T., Canada. – Boreas 10: 71–89.

Farrell, W. E. & Clark, J. A. 1976. On postglacial sea level. – Geophys. J. Roy. astr. Soc. 46: 647–667.

Hattersley-Smith, G., Fuzesy, A. & Evans, S. 1969. Glacial depths in northern Ellesmere Island: airborne radio echo sounding in 1966. – Defence Res. Board, Canada, DREO Tech. Note 69–6: 23 pp.

Keen, C. E. & Hyndman, R. D. 1979. Geophysical review of the continental margins of eastern and western Canada. – Can. J. Earth Sci. 16: 712–747.

Kerr, J. W. 1967. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Kerr, J. W. 1980. Did Greenland drift along Nares Strait? – Bull. Can. Petrol. Geol. 28: 279–289.

Koerner, R. M. 1977. Ice thickness measurements and their implications with respect to past and present ice volumes in the Canadian High Arctic ice caps. – Can. J. Earth Sci. 14: 2697–2705.

Lagerlund, E. 1977. Till studies and neotectonics in northwest Skåne, south Sweden. – Boreas 6: 159–166.

Lanphere, M. A. 1978. Displacement history of the Denali fault system, Alaska and Canada. – Can. J. Earth Sci. 15: 817–822.

Lundqvist, J. & Lagerbäck, R. 1976. The Pärve Fault: a late-glacial fault in the Precambrian of Swedish Lapland. Geol. Fören. Stockh. Förh. 98: 45–51.

Lyons, J. B. & Mielke, J. E. 1973. Holocene history of a part of northernmost Ellesmere Island. – Arctic 26: 314–323.

Mock, S. J. 1968. Snow accumulation studies on the Thule peninsula, Greenland. – J. Glaciol. 7: 59–76.

Mörner, N.-A. 1978. Faulting, fracturing, and seismicity as functions of glacio-isostasy in Fennoscandia. – Geology 6: 41–45.

Niblett, E. R. & Whitham, K. 1970. Multidisciplinary studies in the Canadian Arctic. – J. Geomag. and Geoelect. 22: 91–111.

Nichols, R. L. 1969. Geomorphology of Inglefield Land, North Greenland. – Meddr Grønland 188(1): 109 pp.

Peltier, W. R. 1976. Glacial isostatic adjustment II: the inverse problem. – Geophys. J. Roy. astr. Soc. 46: 669–706.

Peltier, W. R. & Andrews, J. T. 1976. Glacio-isostatic adjustments – I. The forward problem. – Geophys. J. Roy. astr. Soc. 46: 605–646.

Praus, O., DeLaurier, J. M. & Law, L. K. 1971. The extension of the Alert geomagnetic anomaly through northern Ellesmere Island, Canada. – Can. J. Earth Sci. 8: 50–64.

Quinlan, G. in press. A comparison of observed and theoretical Holocene relative sea level in Atlantic Canada. – Can. J. Earth Sci.

Rubin, M. & Alexander, C. 1960. U. S. Geological Survey radiocarbon dates V. – Am. J. Sci. Radiocarbon Suppl. 2: 129–185.

Smith, H. A. 1975. Geology of the West Sable structure. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 133–153.

Stein, S., Sleep, N. H., Geller, R. J., Wang, S.-C. & Kroeger, G. C. 1979. Earthquakes along the passive margin of eastern Canada. – Geophys. Res. Lett. 6: 537–540.

Thorson, R. M. 1979. Isostatic effects of the last glaciation in the Puget Sound Lowland, Washington. – Unpubl. Ph.D. thesis, Univ. of Washington, Seattle, U.S.A.: 173 pp.

Trettin, H. P. 1971. Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, Arctic Archipelago. – Bull. geol. Surv. Can 203: 134 pp.

Trettin, H. P., Frisch, T. O., Sobczak, L. W., Weber, J. R., Niblett, E. R., Law, L. K., DeLaurier, J. M. & Whitham, K. 1972. The Innuitian Province. – In: Price, R. A. & Douglas, R. J. W. (eds), Variations in tectonic styles in Canada. – Spec. Pap. geol. Ass. Can. 11: 83–179.

Walcott, R. I. 1970. Isostatic response to loading of the crust in Canada. – Can. J. Earth Sci. 7: 716–727.

Walcott, R. I. 1972a. Late Quaternary vertical movements in eastern North America: quantitative evidence of glacio-isostatic rebound. – Rev. Geophys. Space Phys. 10: 849–884.

Walcott, R. I. 1972b. Past sea levels, eustasy and deformation of the earth. – Quaternary Res. 2: 1–14.

Weidick, A. 1971. Quaternary map of Greenland, 1:2 500 000. – Geol. Surv. Greenland, Copenhagen.

Weidick, A. 1972. Holocene shore-lines and glacial stages in Greenland – an attempt at correlation. – Rapp. Grønlands geol. Unders. 41: 39 pp.

Weidick, A. 1976. Glaciations of northern Greenland – new evidence. – Polarforschung 46: 26–33.

Weidick, A. 1978. Comments on radiocarbon dates from northern Greenland made during 1977. – Rapp. Grønlands geol. Unders. 90: 124–128.

Wetmiller, R. J. & Forsyth, D. A. 1978. Seismicity of the Arctic, 1908–1975. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 15–24.

Wetmiller, R. J. & Forsyth, D. A. 1982. Review of seismicity and other geophysical data near Nares Strait. – This volume.

Stratigraphy

The Precambrian Shield of northernmost Baffin Bay: correlation across Nares Strait

THOMAS FRISCH and PETER R. DAWES

Frisch, T. & Dawes, P. R. 1982. The Precambrian Shield of northernmost Baffin Bay: correlation across Nares Strait. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 79–88.

The northern rim of the North American craton outcrops in the Baffin Bay – Nares Strait region where it forms large parts of southeastern Ellesmere Island and northwestern Greenland. Between 76°N and 78°N the Precambrian Shield comprises high-grade gneiss and metasediments which show a comparable chronological history in Ellesmere Island and Greenland. Two major features that emphasise the close relationship between the Shield on either side of Nares Strait are the presence of marble-rich metasedimentary tracts and a distinctive suite of late Archaean granitic to basic plutonic rocks which have locally been transformed by intense Proterozoic (probably Hudsonian) deformation into complexly folded orthogneisses.

Marble-rich tracts are common throughout southeastern Ellesmere Island: in northwestern Greenland they are restricted to Inglefield Land north of 78°N. The regional structure suggests that the marble-rich tracts constitute a single marble province and that those on opposite sides of Smith Sound at Cape Isabella (Canada) and Sunrise Pynt (Greenland) may well represent on-strike parts of the same belt of supracrustal rocks.

Correlation of the distinctive features of the Shield in southeastern Ellesmere Island and northwestern Greenland is entirely consistent with there having been no major strike-slip movement along Nares Strait.

T. Frisch, Department of Energy, Mines and Resources, Geological Survey of Canada, Precambrian Division, 588 Booth Street, Ottawa, Ontario, Canada, K1A OE4 and P. R. Dawes, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark.

The oldest rocks of the Nares Strait region are those of the Precambrian Shield that outcrop between 76°N and 79°N in northernmost Baffin Bay up to the Kane Basin. The Shield is overlain extensively in parts of northwestern Greenland, less so in southeastern Ellesmere Island, by unmetamorphosed Neohelikian strata of the Thule Basin (Fig. 1). The Precambrian rocks are bordered to the west, in Ellesmere Island, and north, in the region of 79°N, by Palaeozoic strata of the Arctic platform and Franklinian geosyncline.

The aim of this paper is to describe some distinctive features of the Precambrian Shield between 76°N and 79°N which bear on geological correlations between Canada and Greenland.

Field work and scope of the paper

There is little doubt that the Shield terrains of Greenland and Canada are parts of the same Archaean–Proterozoic crustal block. Several workers, e.g. Allaart et al. (1969) and Bridgwater et al. (1973), have emphasised the close similarities between the Precambrian chronologies of the Labrador – Baffin Island coast and West Greenland, in some cases to the extent that a much closer pre-drift geographical proximity is predicted. This close similarity is also evident in the present study area where, as early as the turn of the century, Per Schei and Albert P. Low, geologists to the 2nd Fram Expedition 1898–1902 (Schei 1904, Holtedahl 1917) and the Dominion Government Expedition 1903–1904 (Low 1906) respectively, recognised similar rock suites on opposite sides of Smith Sound — the narrowest part of the Baffin Bay – Nares Strait region. These early discoveries in the Smith Sound region have been confirmed by our field work which has enlarged on the geological detail and placed it in its regional setting.

In this context a brief review of our investigations in the region may be in order. Dawes began work in the Thule district of Greenland in 1971 and has spent six seasons studying the entire Precambrian terrain from Inglefield Land to southern Melville Bugt (75°N). Frisch mapped in the Shield terrains of Ellesmere Island and Devon Island in 1977, 1978 and 1980 and worked in Greenland with Dawes in 1978 and 1980.

These studies have established the general similarity in the evolution of the Shield in northwestern Greenland and adjacent Canada on a par with that recorded

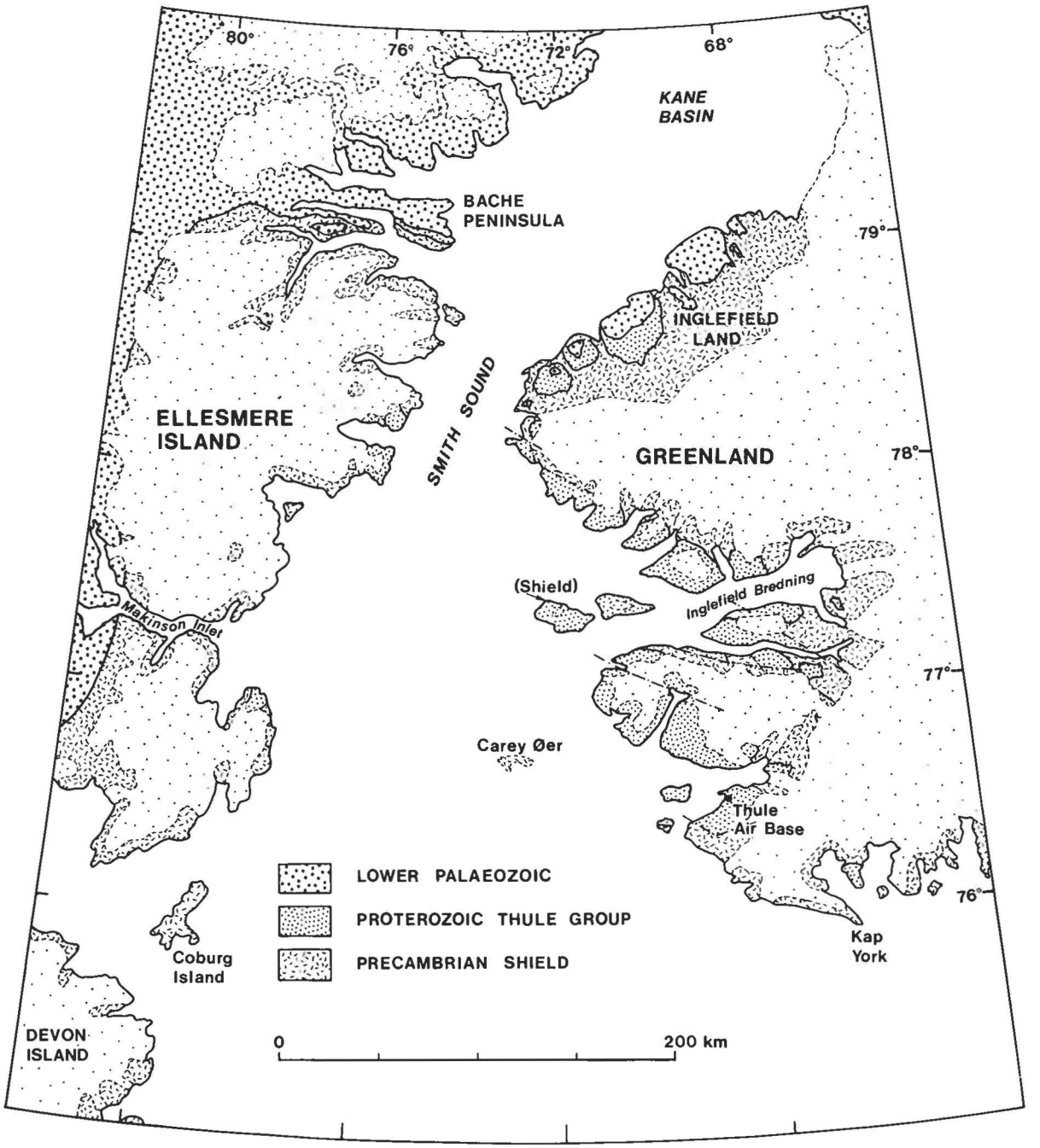


Fig. 1. Simplified geological map of northernmost Baffin Bay and Smith Sound showing outcrop of the Precambrian Shield. The location in the Nares Strait region is shown in Fig. 6. Main areas of ice indicated by widely spaced dot symbol.

between Baffin Island – Labrador and the West Greenland coasts mentioned above. Only a brief summary of the regional geology and evolution is given in this paper, enough to establish that the Shield in Ellesmere Island and adjacent Greenland shows a comparable plutonic and tectonic history. For the purpose of this symposium we prefer to focus on Smith Sound (Fig. 2) which, as the narrowest part of the seaway, is the region most likely to provide data meaningful to the question of displacement along Nares Strait.

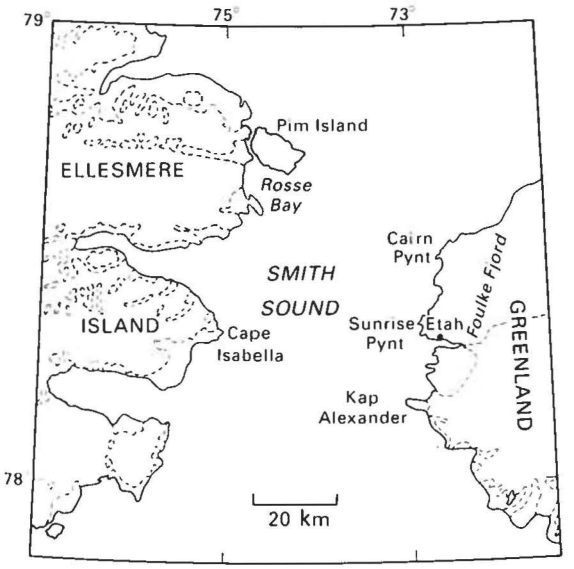


Fig. 2. Index map of the Smith Sound area showing the main ice limits.

Regional chronostratigraphic history

In both southeastern Ellesmere Island and northwestern Greenland the Precambrian Shield is composed of a variety of highly metamorphosed and deformed gneisses, metasediments and meta-igneous rocks which represent both Archaean and Proterozoic crustal material. The metamorphic grade of the Shield varies from amphibolite to granulite facies and in both areas the amphibolite facies, more common in Greenland, is regarded as having been downgraded from granulite facies.

The crystalline rocks have been described both in Greenland and Ellesmere Island in terms of a number of rock associations. Dawes (1976) mentioned six main lithostratigraphic units in northwestern Greenland (gneiss-schist and granite, anorthosite, psammite and amphibolite, pelite and marble, gabbroic to granitoid rocks, and variable melanocratic to quartz-rich gneiss) while Frisch et al. (1978), building on the early reconnaissance work of Christie (1962a, b), divided the adjacent terrain of Ellesmere and Coburg Islands into five broad groups (massive to foliated quartzo-feldspathic granulites, metasediments, amphibolites, granulite gneisses and granitic rocks). Frisch et al. (1978: 136) concluded that "in lithology, grade of metamorphism and tectonic style, the crystalline basement of Ellesmere Island bears many similarities to that of northwestern Greenland". In addition, these authors emphasise that all the granites present in Ellesmere Island are at least in part foliated, being pre- or syn-tectonic; indeed, the basement is characterised by the absence of post-tectonic intrusions. This is equally true of the crystalline terrain of northwestern Greenland.

Much of the Shield has undergone polyphase deformation and metamorphism, and age relationships between many rock units are still uncertain. Ice and snow form a significant cover, particularly in Ellesmere Island, where the regional structure and rock relationships have had to be pieced together in nunatak terrain. However, a fairly detailed chronostratigraphic history can be erected for both Ellesmere Island and Greenland

Table 1. Chronostratigraphic history of NW Greenland and SE Ellesmere Island (75°N–79°N).

Eon	Era	NW GREENLAND	Age m.y.	SE ELLESMERE	Age m.y.
PROTEROZOIC	HADRYNIAN	Basic dykes	600	Basic dykes	600–800
		THULE GROUP, sediments extrusives, intrusives	1100–1200	THULE GROUP, sediments extrusives, intrusives	1000–1200
	HELIKIAN	Basic dykes	1550	?Basic dykes	
	APHEBIAN	(5) HUDSONIAN OROGENY – high-grade mixed gneiss complex	1610–1960	(5) HUDSONIAN OROGENY – high-grade mixed gneiss complex	1760
		(4) Metasedimentary and meta-igneous rocks		(4) ?	
ARCHAEAN		(3) Meta-igneous complex basic to acid suite	2700	(3) Meta-igneous complex basic to acid suite	
		(2) Metasediments, marble, pelite, psammite, garnet schist, amphibolite		(2) High-grade multiphase & gneiss and metasedimentary	
		(1) High-grade multiphase gneiss complex with amphibolite, anorthosite		(1) complex. Marble, pelite, psammite, amphibolite	

The isotopic ages quoted are both K/Ar and Rb/Sr ages reported in Christie (1962b), Larsen & Dawes (1974), Dawes (1976), Kalsbeek & Dawes (1980) and Frisch & Christie (1982).

(Table 1). In both regions a number of correlatable rock complexes and events are apparent.

In Greenland a basement–cover relationship is preserved in the Inglefield Bredning area which establishes the age relationship between the high-grade gneiss complex (1) and at least some of the metasedimentary and meta-igneous rocks (2). In Inglefield Land, about 100 km to the northwest, a thick metasedimentary sequence characterised by abundant marble (Etah Group of Dawes 1972) is regarded as a possible equivalent of this cover sequence (2). No basement–cover relationship has been recognised in Ellesmere Island and supracrustal rocks are everywhere intimately associated with the high-grade gneiss complex (1).

The meta-igneous rocks of complex (3) form large areas in both Ellesmere Island and Greenland and they intrude the supracrustal rocks referred to complex (2). Rb/Sr isotope work on igneous rocks of the Etah and Kap York areas in Greenland indicates that the intrusions are of Archaean age (c. 2700 m.y.) although the rocks are severely overprinted by Proterozoic metamorphism (Larsen & Dawes 1974, Kalsbeek & Dawes 1980).

Complex (4) is only known from the Melville Bugt area of Greenland where certain pelitic schist sequences post-date igneous rocks assigned to complex (3) (Dawes 1979, Dawes & Frisch 1981). However, supracrustal rocks of this complex may also be present elsewhere in NW Greenland and Ellesmere Island possibly represented by some of the supracrustal rocks that are associated in the high-grade gneisses of complex (1).

A mixed gneiss complex (5) derived from rocks of complexes (2) and (3) during the Hudsonian orogeny is well documented in both Ellesmere Island and Greenland. This major tectonometamorphic period is dated as Aphebian by K/Ar and Rb/Sr isotopic ages which fall between about 1600 and 1900 m.y. (Christie 1962b, Larsen & Dawes 1974).

Cross-cutting, undeformed dolerite dykes intrude the crystalline terrain in both Ellesmere Island and Greenland, but many of these dykes post-date the Thule Group and are of late Proterozoic (Hadrynian) age. However, at least some dykes in Greenland are of middle Proterozoic (Helikian) age; it is uncertain whether Helikian basic dykes are represented in southeastern Ellesmere Island (Table 1).

Smith Sound

The Shield on either side of Smith Sound (78° to 79°N) is composed of rocks belonging to complexes (2), (3) and (5) as defined in Table 1. The rocks of the three complexes are intimately associated and severely deformed; isoclinal folds are common. The regional foliation attitude varies, but steep to vertical gneissic and lithostratigraphic dips predominate. In coastal Ellesmere Island the regional gneissic trend varies from northerly to northeasterly, while in Greenland a conspicuous NE to ENE trend is formed by steeply-dipping linear belts of marble-rich metasediments flanked by gneisses and granitic rocks (Figs 3 and 4). Two typical metasedimentary-rich successions in Ellesmere Island and Greenland are shown in Fig. 7, illustrating the steep dips, lithological heterogeneity and presence of marble.

Two rock groups, one of sedimentary, the other of igneous origin, are described and their importance in correlation between Greenland and Ellesmere Island is outlined.

Marble-rich metasedimentary belts

Coarse-grained, white marble with diopside, forsterite, wollastonite and spinel is a characteristic component of supracrustal belts in Ellesmere Island and Inglefield Land. The major outcrops of marble and marble-rich supracrustal units are plotted in Fig. 5.

The presence of marbles in the Smith Sound area has

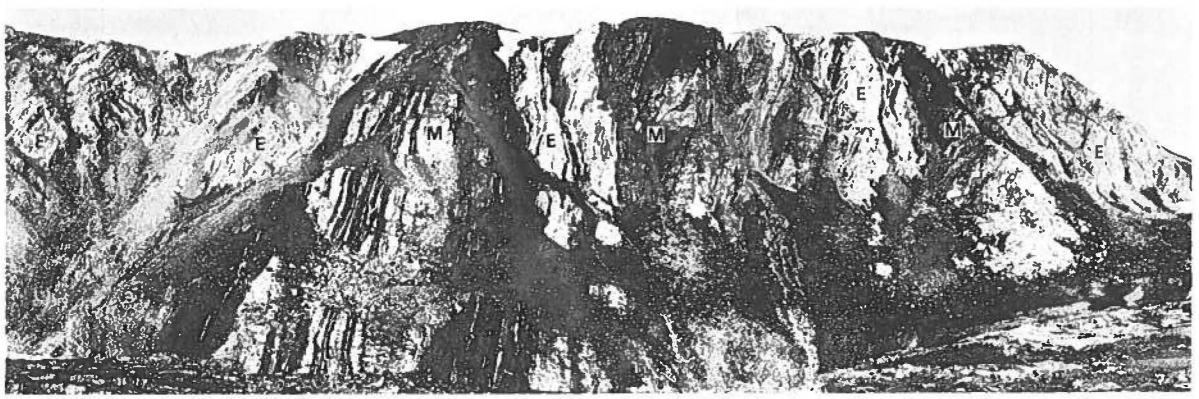


Fig. 3. Southern part of the Sunrise marble belt, Greenland, viewed inland along the regional strike. Vertical marble (M) and darker pelitic schists of the Etah Group are invaded by granitic and quartz dioritic rocks (E) of the Etah meta-igneous complex. Height of the cliff is about 350 m. From Dawes (1976: fig. 228).

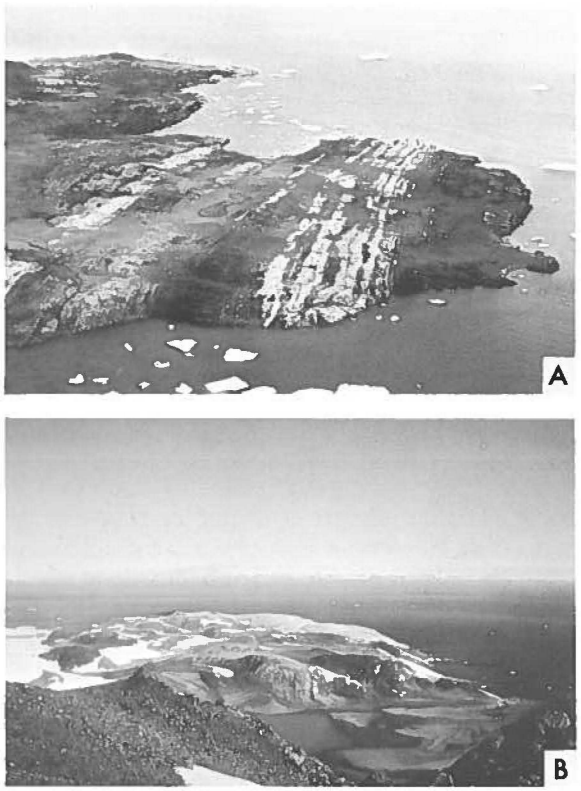


Fig. 4. A: The Sunrise Pynt marble belt, viewed from the east, at the low-lying outer coast striking towards Ellesmere Island. The white units are marble, darker units pelitic schists invaded by igneous rocks; in the far left granitic gneiss. B: View of Cape Isabella with the Greenland coast in the background. White marble at the actual cape is bordered by pelitic gneiss and pink hypersthene-bearing granite; hypersthene gneiss in the foreground. This locality is the most easterly marble occurrence on Ellesmere Island — less than 50 km from the Sunrise Pynt marbles on Greenland (see Fig. 5).

been known for some time. Low (1906: 208) recorded crystalline limestone at both Cape Isabella in Canada and in the region of Foulke Fjord (Sunrise Pynt) in Greenland. He correlated the two sequences and astutely noted that in both regions "great masses of granite" had invaded the metasediments. Koch (1929, 1933) referred the crystalline limestones of Inglefield Land, as well as those on the Ellesmere Island side of Smith Sound, to the Etah Formation (now the Etah Group of Dawes 1972).

The Cape Isabella and Sunrise Pynt marbles remain the best known occurrences in the region (Figs 3 and 4). At Cape Isabella, marble forms a substantial part of a meta-igneous/metasedimentary assemblage bordered by granite-veined hypersthene gneiss. The supracrustal rocks comprise interlayered belts, trending northeast, of marble, calc-silicate rocks, (cordierite-sillimanite-) garnet gneiss, and two-pyroxene quartzo-feldspathic rock (meta-igneous? granulite), all isoclinally folded and severely faulted. Red granite and pegmatite have intruded marble and other metasediments and, despite later intense deformation, transgressive contacts are preserved. Subsequent flowage of marble has disrupted much of the pegmatite and trains of fragments mark the sites of former pegmatite sheets.

Sunrise Pynt offers an extremely well-exposed cross-section through the thickest of the numerous supracrustal belts of Inglefield Land. Vertically-dipping marble and pelitic (sillimanite-) garnet schist are intercalated with meta-igneous pink granite, pegmatite and hypersthene-bearing quartz dioritic to granitic rocks in a cliff section about 1 km long (Fig. 3). This supracrustal belt strikes northeast, parallel to the regional trend, and is bordered on both sides by hypersthene gneisses veined by granite. The southern contact of the marble belt is subconcordant to the lithostratigraphic layering of the metasediments and the discordant intrusive contact with the bordering pink granitic rocks is preserved on a regional scale. Isoclinal folding — as seen by, for example, tightly-folded inclusions in marble — has not destroyed original intrusive contacts but boudinaged

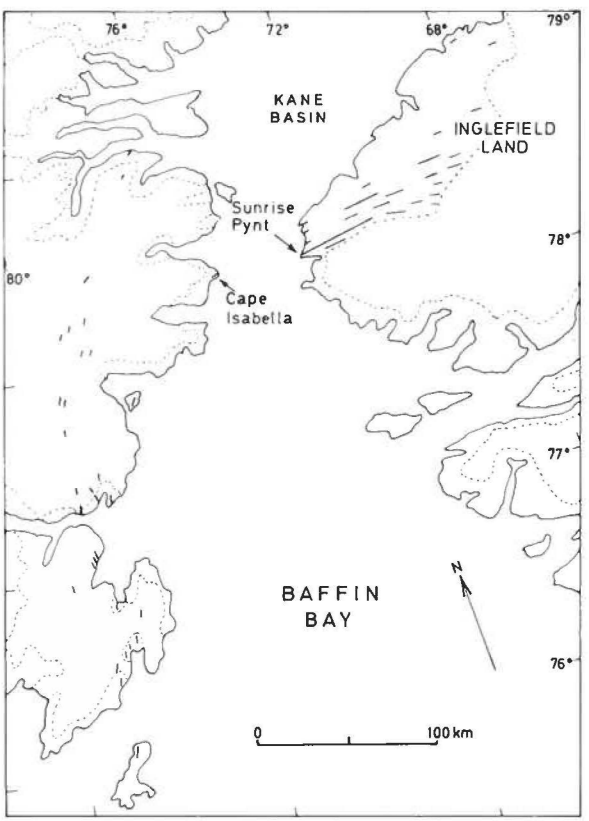


Fig. 5. Distribution of known marble outcrops of the Ellesmere Island – Inglefield Land marble province. The dashes represent single outcrops. In Inglefield Land the dashes are generally longer since the ice-free terrain allows each marble unit to be traced along strike; in Ellesmere Island marble units are less continuous due to the nunatak terrain.

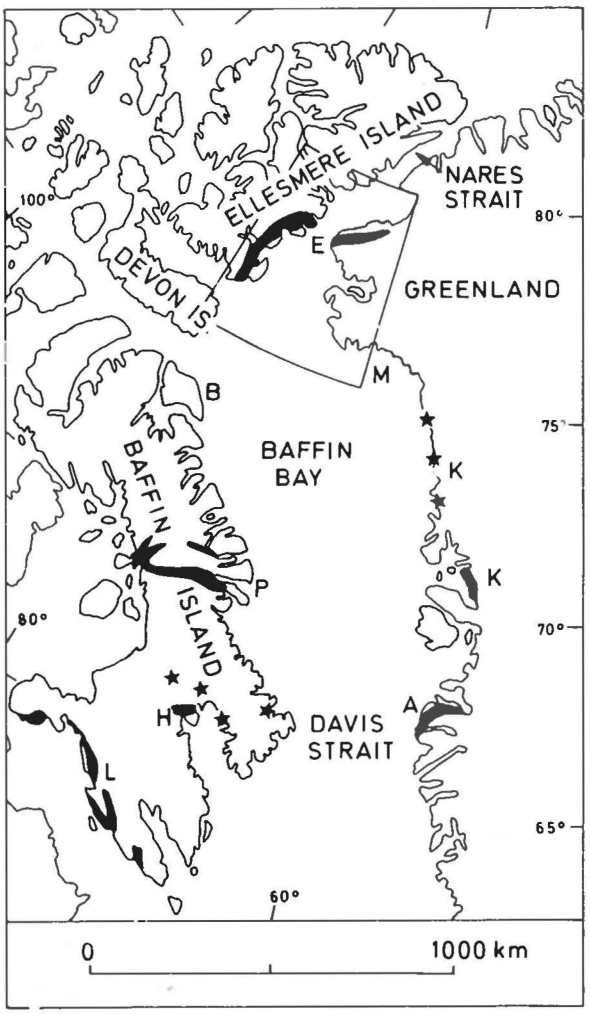


Fig. 6. The regional distribution of marble in the Precambrian Shield of the Baffin Bay – Davis Strait region. Black areas indicate important marble provinces, stars minor occurrences. E = Etah Group; P = Piling Group, Foxe fold belt; K = Karrat Group, Rinkian mobile belt; L = Lake Harbour Group, Dorset fold belt; H = Hoare Bay Group; A = Agto supracrustals, Nagssugtoqidian mobile belt; B = Bylot Island; M = Melville Bugt. Data for Baffin Island and West Greenland from Jackson & Taylor (1972), Escher & Pulvertaft (1976), Escher et al. (1976), Escher (in press). Frame locates the present study area as shown in Fig. 1.

remnants and other fragments of meta-igneous rock are common in marble.

The similarities of lithology, structure and chronology of geologic events at Cape Isabella and Sunrise Pynt are striking. Further, the two localities are in close proximity, roughly on strike, and the marble belts in Greenland and Ellesmere Island are known to have regional persistency. The Sunrise Pynt belt can be traced inland for at least 90 km towards the Inland Ice margin, and although the Cape Isabella outcrop cannot be mapped continuously inland because of extensive ice cover, it forms part of a marble-rich metasedimentary zone that can be traced for 300 km throughout the nunatak terrain of southeastern Ellesmere Island (Fig. 5). The evidence thus points to the two coastal supracrustal complexes having once been joined in a single belt or, at least, belonging to the same family of supracrustal belts. To the north, between Sunrise Pynt and Cairn Pynt, several thinner linear belts of marble reach the coast and strike towards the Cape Isabella area. Each of these displays a similar lithology and has the same age relationships to the surrounding high-grade granitic rocks as described for the Sunrise Pynt section. We regard the presence of on-strike marble belts invaded by comparable igneous material, on opposite sides of the 50 km wide seaway, as particularly noteworthy, and strongly suggestive of a direct correlation of the two regions.

Further evidence for this correlation is provided by the regional distribution of marble sequences in the Baffin Bay – Davis Strait region, the main occurrences of which are indicated in Fig. 6. The Sunrise Pynt marble belt represents the southernmost marble occurrence in northwestern Greenland; in the entire Thule – Melville Bugt region (75° to 78°N) no marble has been reported (Dawes 1976, Dawes & Frisch 1981). The next marble outcrops to the south belong to the Rinkian mobile belt (Escher & Pulvertaft 1976): between latitudes 75° and 73°N, thin horizons of calc-silicate rock form part of the Proterozoic Karrat Group (Escher & Stecher 1978, 1980, Escher in press), while to the south, in the Mârmorilik region of central West Greenland (71°N), marble constitutes an important rock type of the early Proterozoic metasedimentary sequences (Garde 1978).

In Ellesmere Island the 300 km belt of marble-rich metasediments peters out south of 76°N, i.e. marble is present on Coburg Island but is practically absent from the metasedimentary belts of Devon Island (Frisch 1981a). No marble has been reported from northern Baffin Island or Bylot Island (Blackadar 1970, Jackson & Davidson 1975) and the closest marbles to the south occur in central Baffin Island at about 70°N. These marbles occur as extensive units in the Proterozoic (Aphebian) Piling Group of the Foxe fold belt (Jackson & Taylor 1972) that in lithology, geological setting and age has much in common with the Karrat Group of central West Greenland, suggesting direct correlation (J. C. Escher, J. R. Henderson, G. D. Jackson, T. C. R. Pulvertaft, pers. comms). One can suggest correlation of the marble and metasedimentary successions farther south in Baffin Island and West Greenland (Fig. 6). Whatever the value of such long-range correlations, we consider the grouping of marble and metasedimentary provinces on either side of Baffin Bay – Davis Strait highly significant, and regard the Etah Group marbles as constituting a discrete lithologic province, hereby termed the Ellesmere Island – Inglefield Land marble province, an entity separate and distinct from adjacent areas of Precambrian Shield.

Quartzo-feldspathic plutonic rocks

Close correlation can also be made between a distinctive suite of magmatic rocks on both sides of Smith Sound. Eighty years ago Per Schei (Schei 1904, Bugge 1910, Holtedahl 1917) recognised the similarity of the hypersthene-bearing quartzo-feldspathic rocks in the Foulke Fjord – Etah area and in Rosse Bay and Pim Island (Fig. 2). Recent work by us has confirmed and extended Schei's conclusion (Frisch 1981b). In both areas the main rock types are orthopyroxene-bearing quartz diorite and granodiorite (termed granulite by Frisch et al. 1978) and intrusive granite, which together far outweigh in abundance the Etah Group supracrustal rocks described above. This magmatic suite has been termed the Etah meta-igneous complex in Greenland (Dawes 1972), but we suggest extension of the term to cover the corresponding terrain of southeastern Ellesmere Island.

The orthopyroxene-bearing rocks are medium- to coarse-grained, dark, often brownish weathering, generally massive or weakly gneissic. Fresh surfaces show typically the greenish colour and waxy sheen, characteristic of quartzo-feldspathic rocks of the granulite facies. The rocks are composed mainly of oligoclase, alkali feldspar, quartz, orthopyroxene, olive-green hornblende and biotite, with local garnet. The feldspar is typically green and some blue quartz occurs.

Whereas the orthopyroxene-bearing rocks appear to be primarily meta-igneous and can be seen at Sunrise Pynt to intrude the Etah Group supracrustals, they are themselves invaded by pink to red granite in veins a few centimetres thick to sheets or irregular bodies many tens of metres thick, as well as more extensive masses. Although commonly discordant in detail the granitic intrusions are broadly concordant with the regional structure; indeed, granulite-granite layering in places constitutes the main structure. Where deformation was particularly intense, granulite injected lit-par-lit by granite has been transformed into well-banded gneiss and various stages of the process are preserved in outcrop. Similarly, gradations from massive granulite to granulite gneiss are commonplace in both Inglefield Land and Ellesmere Island and attest to major deformation after intrusive activity.

Another notable feature of the quartzo-feldspathic

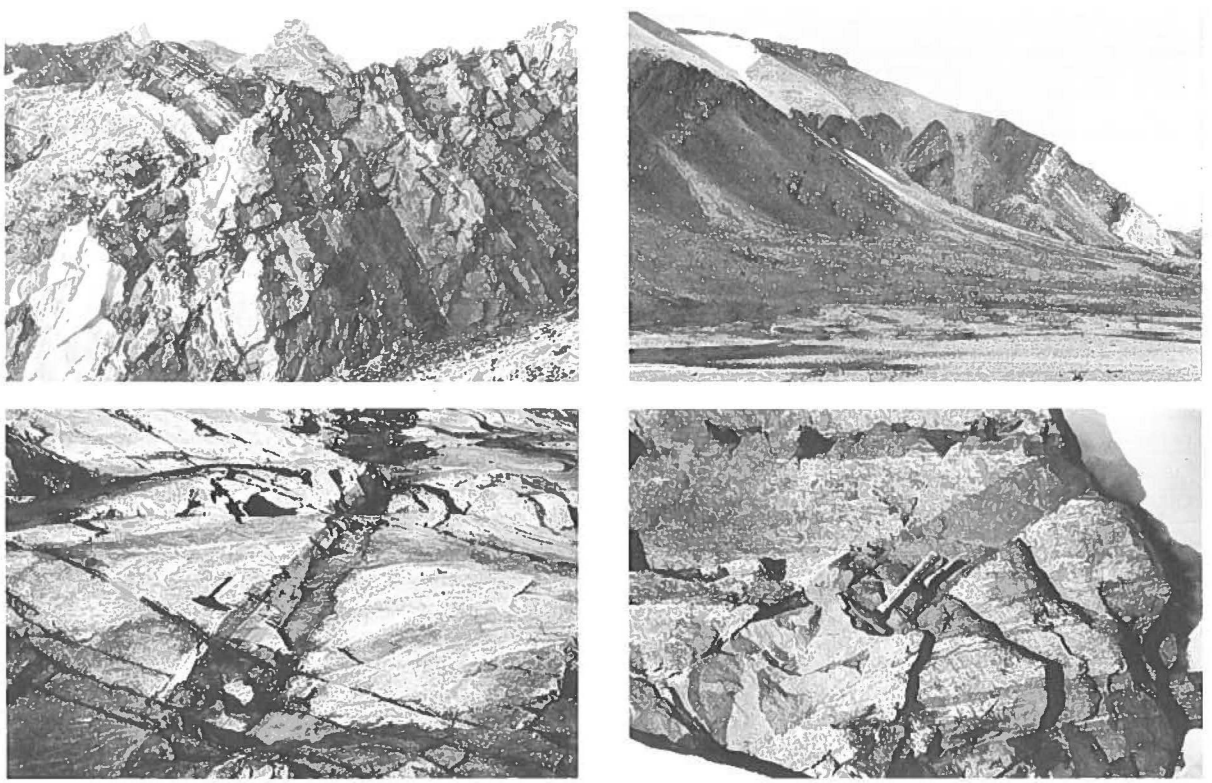


Fig. 7. Comparison of common rock associations of the Shield on opposite sides of Smith Sound. Ellesmere Island examples on the left, Greenland on the right. Upper: typical steeply-dipping metasedimentary/ meta-igneous succession including conspicuous marble (white) and quartzo-feldspathic plutonic rocks. In the Greenland outcrop, north of Sunrise Pynt, the Shield is unconformably overlain by the Proterozoic Thule Group. Ellesmere Island locality is from the south shore of Makinson Inlet (photograph by G. R. Dunning). Lower: typical outcrop of hypersthene-bearing granulite (dark) veined by reddish granite, cut by basic metadykes. The examples are from Kjöthaugen, Pim Island (see Bugge 1910: fig. 5, p. 11) and the coast north of Cairn Pynt.

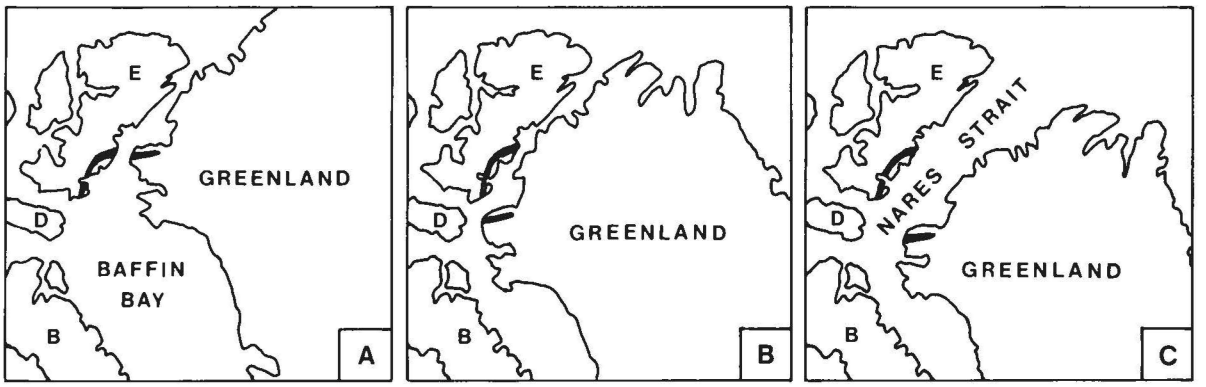


Fig. 8. Locations of the Precambrian Ellesmere Island – Inglefield Land marble province (black) in present-day geography (A) compared with two pre-drift reconstructions, early Tertiary (B) and late Cretaceous (C). Position of lands in (B) and (C) are redrawn from Srivastava (1978). E = Ellesmere Island, D = Devon Island, B = Baffin Island.

rock terrains on both sides of Smith Sound is the sporadic occurrence of thin tabular bodies of metabasite composed of primarily pyroxene-hornblende-biotite-plagioclase. The contacts of these bodies vary from sharp to diffuse and some are discordant, suggesting that the bodies are metadykes or remnants thereof (Fig. 7). The metadykes cut both the dark hypersthene-bearing and later red granitic rocks and they could represent several ages of intrusion. The metadykes are everywhere thoroughly metamorphosed and are also veined by granitic material.

Isotopic age determinations from Greenland provide evidence that the Etah meta-igneous rocks are of late Archaean age and underwent major deformation in Hudsonian time (c. 1800 m.y.) (Larsen & Dawes 1974, Kalsbeek & Dawes 1980). Further geochronological work is in progress but, whatever the precise age of the events, the evidence of intrusion of the same distinctive suite of plutonic rocks followed by major deformation can be easily documented in the field on each side of Smith Sound.

Inferences on continental drift

The opening of Labrador Sea and Baffin Bay by sea-floor spreading, the resultant lateral separation of Greenland and Baffin Island as separate plates and the consequent strike-slip movement along Nares Strait are ideas that have gained wide support (e.g. Keen et al. 1974, Sclater et al. 1977, Srivastava 1978, Jackson et al. 1979) and they form a central theme in this symposium. With regard to the question of the movement of Greenland relative to Ellesmere Island, we wish to emphasise two main aspects of correlation related to the Precambrian Shield:

1) The Precambrian crystalline rocks in southeastern Ellesmere Island and northwestern Greenland (76°–79°N) show clear similarity in chronostratigraphy and magmatic-tectonic history which conclusively establishes that the two areas form parts of the same Shield terrain.

2) At Smith Sound, where the crystalline rocks of Ellesmere Island and Greenland are in closest geographical proximity (50 km), *directly* correlatable metasedimentary and meta-igneous rock associations characterise the Shield. The evidence of this correlation is summarised as follows.

   a. The marble-rich metasedimentary belts (Etah Group) in Inglefield Land and Ellesmere Island form a lithologic province separate and distinct from the Shield areas farther south.
   b. In both Inglefield Land and Ellesmere Island the marble belts are regionally persistent.
   c. The marble belts, including the coastal outcrops at Cape Isabella and Sunrise Pynt, have similar lithology and mineralogy and are roughly on strike.
   d. The marble belts on both sides of Smith Sound have the same relationship to a suite of younger plutonic rocks (Etah meta-igneous complex).
   e. The Etah meta-igneous complex is a distinctive suite of granitic to basic rocks that, on both sides of Smith Sound, is composed of identical rock types and the same sequence of intrusion phases.
   f. The Etah meta-igneous rocks have been transformed locally into highly folded orthogneisses in the granulite facies; rock types of the granite or quartz diorite to gneiss transition on either side of Smith Sound are indistinguishable.

These two aspects of correlation can be considered as two categories of correlation refinement. Hence, data of category 1, although essential in establishing the broad temporal and regional setting of Ellesmere Island and adjacent Greenland, can contribute little to a leading

question of this symposium, i.e. has there been displacement along Nares Strait and if so, how much? On the other hand, evidence from category 2 is of direct significance to this question and we regard the marble-rich metasedimentary belts intruded by distinctive magmatic rocks as a geological marker across Nares Strait.

The present-day distribution pattern of marble-rich metasedimentary belts and associated magmatic rocks in Ellesmere Island and Inglefield Land is entirely consistent with there having been no strike-slip movement along Nares Strait. However, the data cannot be used for a precise definition of the amount of net displacement permissible, although a dextral displacement of Greenland relative to Ellesmere Island of as little as 50 km disrupts the continuation of the Cape Isabella/Sunrise Pynt – Cairn Pynt marble belts. Nevertheless, such is the nature of Shield terrain, developed through complex structural histories, that exclusion of such a 50 km offset reconstruction solely on the basis of data from the Precambrian Shield would be dangerous.

At the other end of the movement scale, a dextral offset of 300 km (envisaged by, for example, Newman 1977, Newman & Falconer 1978 and Srivastava 1978, to accommodate a corresponding sinistral movement in the Tertiary) places the Inglefield Land marbles south of the southern termination of the Ellesmere Island metasedimentary zone at about 75°N (Fig. 8B and C). The Inglefield Land linear marble belts thus strike towards Devon Island, an area devoid of marble-rich metasedimentary belts. However, a convenient postulate might be that the Greenland marbles, rather than being a northeasterly continuation of the Ellesmere Island metasedimentary zone, as present-day geography suggests, form an across-strike correlation, perhaps with the Inglefield Land marbles depicting one limb, and the Ellesmere Island marbles the other, of a large-scale, southerly-closed fold structure, the axial trace of which is in Nares Strait.

We have considered various reconstructions of the region and the realms of geological improbability which each involves, but we cannot favour complex and speculative explanations over fundamental and straightforward correlation of metasedimentary and plutonic complexes directly across Nares Strait. We are aware that the types of geological structures and boundaries commonly preserved in Precambrian crystalline rocks — often the products of repeated orogenic disturbances and transformations — generally are less likely to be of use as geological markers in precise correlation between landmasses than, for example, unmetamorphosed, undeformed and fossiliferous sequences, the age, character and structure of which can be unequivocally determined. We thus regard it as very significant that the conclusion on strike-slip displacement based on data from the Archaean and early Proterozoic crystalline rocks is entirely consistent with that based on a variety of stratigraphical and structural aspects from the late Proterozoic and Palaeozoic sediments, reported elsewhere in this volume.

Conclusion

Field and laboratory studies of the onshore region between 76° and 79°N in northernmost Baffin Bay have extended the geological correlation across the Labrador Sea – Baffin Bay seaway to the northern extremity of the North American craton. Our data demonstrate that correlation at Smith Sound, the narrowest part of the seaway, can be based on the same supracrustal group and the same plutonic complex. It is thus considerably more refined than the general and regional correlation based on similar rocks, structures and events of comparable ages that can only be made farther south where the landmasses are widely separated. This correlation across Smith Sound militates against any large-scale net strike-slip displacement along Nares Strait.

Acknowledgements

T. F. is greatly indebted to the Polar Continental Shelf Project for indispensable logistic support and to his colleagues, W. C. Morgan and G. R. Dunning, who shared in the geological mapping on Ellesmere Island. P. R. D. thanks Kurt Thomsen for able field assistance. We are grateful to A. K. Higgins for critically reading the manuscript. The paper is published with the permission of the Director of the Geological Survey of Greenland.

References

Allaart, J. H., Bridgwater, D. & Henriksen, N. 1969. Pre-Quaternary geology of southwestern Greenland and its bearing on North Atlantic correlation problems. – In: Kay, M. (ed.), North Atlantic – geology and continental drift. A symposium. – Mem. Am. Ass. Petrol. Geol. 12: 859–882.

Blackadar, R. G. 1970. Precambrian geology of northwestern Baffin Island. – Bull. geol. Surv. Can. 191: 89 pp.

Bridgwater, D., Escher, A., Jackson, G. D., Taylor, F. C. & Windley, B. F. 1973. Development of the Precambrian Shield in West Greenland, Labrador, and Baffin Island. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 99–116.

Bugge, C. 1910. Petrographische Resultate der 2ten Fram-Expedition. – Rep. 2nd Norwegian Arctic Exped. "Fram" 1898–1902, 22: 38 pp. – Videnskabs-Selskabet, Kristiania.

Christie, R. L. 1962a. Geology, Alexandra Fiord, Ellesmere Island, District of Franklin (map with marginal notes) 1" to 4 miles. – Map geol. Surv. Can. 9–1962.

Christie, R. L. 1962b. Geology, southeast Ellesmere Island, District of Franklin (map with marginal notes) 1" to 4 miles. – Map geol. Surv. Can. 12–1962.

Dawes, P. R. 1972. Precambrian crystalline rocks and younger sediments of the Thule district, North Greenland. – Rapp. Grønlands geol. Unders. 45: 10–15.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R. 1979. Field investigations in the Precambrian

terrain of the Thule district, North-West Greenland. – Rapp. Grønlands geol. Unders. 95: 14–22.

Dawes, P. R. & Frisch, T. 1981. Geological reconnaissance of the Greenland Shield in Melville Bugt, North-West Greenland. – Rapp. Grønlands geol. Unders. 105: 18–26.

Escher, A. & Pulvertaft, T. C. R. 1976. Rinkian mobile belt of West Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 102–119. – Geol. Surv. Greenland, Copenhagen.

Escher, A., Sørensen, K. & Zeck, H. P. 1976. Nagssugtoqidian mobile belt in West Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 76–95. – Geol. Surv. Greenland, Copenhagen.

Escher, J. C. in press: Geologiske kort over Grønland 1:500 000 map series. Sheet 4. Grønlands geol. Unders.

Escher, J. C. & Stecher, O. 1978. Precambrian geology of the Upernavik–Red Head region (72°15′–75°15′N), northern West Greenland. – Rapp. Grønlands geol. Unders. 90: 23–26.

Escher, J. C. & Stecher, O. 1980. Field work on Precambrian granites and metasediments in the Søndre Upernavik–Kuvdlorssuaq region (72°00′–74°40′N), northern West Greenland. – Rapp. Grønlands geol. Unders. 100: 38–41.

Frisch, T. 1981a. Further reconnaissance mapping of the Precambrian Shield on Devon Island, District of Franklin. – Pap. geol. Surv. Can. 81–1A: 31–32.

Frisch, T. 1981b. Comparative note on the Precambrian basement of southern Inglefield Land and eastern Ellesmere Island. – Rapp. Grønlands geol. Unders. 105: 14–18.

Frisch, T. & Christie, R. L. 1982. Stratigraphy of the Proterozoic Thule Group, southeastern Ellesmere Island, Arctic Archipelago. – Pap. geol. Surv. Can. 81–19: 13 pp.

Frisch, T., Morgan, W. C. & Dunning, G. R. 1978. Reconnaissance geology of the Precambrian Shield on Ellesmere and Coburg islands, Canadian Arctic Archipelago. – Pap. geol. Surv. Can. 78–1A: 135–138.

Garde, A. A. 1978. The Lower Proterozoic Marmorilik Formation, east of Mârmorilik, West Greenland. – Meddr Grønland 200(3): 71 pp.

Holtedahl, O. 1917. Summary of geological results. – Rep. 2nd Norwegian Arctic Exped. "Fram" 1898–1902, 36: 32 pp. – Videnskabs-Selskabet, Kristiania.

Jackson, G. D. & Davidson, A. 1975. Bylot Island map-area, District of Franklin. – Pap. geol. Surv. Can. 74–29: 12 pp.

Jackson, G. D. & Taylor, F. C. 1972. Correlation of major Aphebian rock units in the northeastern Canadian Shield. – Can. J. Earth Sci. 9: 1650–1669.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor spreading in central Baffin Bay. – Can. J. Earth. Sci. 16: 2122–2135.

Kalsbeek, F. & Dawes, P. R. 1980. Rb–Sr whole-rock measurements of the Kap York meta-igneous complex, Thule district, North-West Greenland. – Rapp. Grønlands geol. Unders. 100: 30–33.

Keen, C. E., Keen, M. J., Ross, D. I. & Lack, M. 1974. Baffin Bay: small ocean basin formed by sea-floor spreading. – Bull. Am. Ass. Petrol. Geol. 58: 1089–1108.

Koch, L. 1929. Stratigraphy of Greenland. – Meddr Grønland 73,2(2): 205–320.

Koch, L. 1933. The geology of Inglefield Land. – Meddr Grønland 73,1(2): 38 pp.

Larsen, O. & Dawes, P. R. 1974. K/Ar and Rb/Sr age determinations on Precambrian crystalline rocks in the Inglefield Land-Inglefield Bredning region, Thule district, western North Greenland. – Rapp. Grønlands geol. Unders. 66: 4–8.

Low, A. P. 1906. Report on the Dominion Government Expedition to Hudson Bay and the Arctic Islands on board the D.G.S. Neptune 1903–1904. – Gov. Print. Bureau, Ottawa: 355 pp.

Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Newman, P. H. & Falconer, R. K. H. 1978. Evidence for movement between Greenland and Canada along Nares Strait. – Geol. Soc. Am. Abs with Prog. 10: 463 only.

Schei, P. 1904. Preliminary account of the geological investigations made during the Second Norwegian Polar Expedition in the "Fram". – In: Sverdrup, O., New Land. Four Years in the Arctic Regions 2: 455–466. – Longmans, Green & Co. London.

Sclater, J. G., Hellinger, S. & Tapscott, C. 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. – J. Geol. 85: 509–552.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

The Proterozoic Thule Basin of Greenland and Ellesmere Island: importance to the Nares Strait debate

PETER R. DAWES, THOMAS FRISCH and ROBERT L. CHRISTIE

Dawes, P. R., Frisch, T. and Christie, R. L. 1982. The Proterozoic Thule Basin of Greenland and Ellesmere Island: importance to the Nares Strait debate. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 89–104.

Onshore geological investigations in the Smith Sound region are now so advanced as to allow correlation between Canada and Greenland to be made with confidence.

The Precambrian Shield is unconformably overlain by unmetamorphosed Proterozoic strata (Thule Group) that are best preserved in Greenland, where they attain a thickness of at least 4.5 km. Less than 1100 m are present in south-eastern Ellesmere Island, but the succession is so similar to the lower part of the Greenland succession that unit to unit correlation of both sedimentary and volcanic rocks is possible. This correlation strongly supports the concept of a single intracratonic basin (Thule Basin) spanning the southernmost part of Nares Strait. In Greenland the basin is well defined and its northern margin is at about 78°15'N. In Ellesmere Island paucity of outcrop provides less definition but the northern margin lies between Baird Inlet (78°30'N) and Bache Peninsula (79°N).

The lithological and thickness correlation of the Proterozoic successions on the opposite shores of the Smith Sound region strongly suggests that any tectonic movement along the Nares Strait lineament has not resulted in major net transcurrent displacement of Greenland and Ellesmere Island.

P. R. Dawes, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark; T. Frisch, Department of Energy, Mines and Resources, Geological Survey of Canada, Precambrian Division, 588 Booth Street, Ottawa, Ontario, Canada, K1A OE4 and R. L. Christie, Department of Energy, Mines and Resources, Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303–33rd St. N.W., Calgary, Alberta, Canada, T2L 2A7.

Few students of continental drift and plate tectonics of the Arctic have resisted the temptation to accommodate movement between Greenland and North America by transcurrent displacement along the linear channel between Greenland and Ellesmere Island — Nares Strait (Taylor 1910, Wegener 1922, Carey 1958, Wilson 1963, Srivastava 1978). Modern reconstructions, many of which have received wide approval, consider the Labrador Sea and Baffin Bay as oceans formed by sea-floor spreading, with the consequent separation of Greenland and Baffin Island resulting in substantial movement along Nares Strait; in some recent models as much as 300 km or more of sinistral motion has been involved (e.g. Sclater et al. 1977, Srivastava 1978).

In the northernmost Baffin Bay – southern Nares Strait region (herein referred to as the Smith Sound region) the Precambrian Shield forming the northern part of the North American craton is overlain by unmetamorphosed sedimentary and volcanic strata of Proterozoic age (Figs 1 and 2). These strata are extensively developed in coastal North-West Greenland where they have been named the Thule Group. More restricted outcrops in coastal south-eastern Ellesmere Island on the opposite shore of the Smith Sound region correlate with the lower part of the Greenland succession. The Thule Group rocks in Greenland and Ellesmere Island are interpreted as parts of a single sedimentary basin — the Thule Basin. This basin spans the southernmost part of the Nares Strait and as such is of particular relevance in any deliberations about the postulated palaeopositions of the two landmasses on which it outcrops.

The aim of this paper is to summarise the main features of stratigraphical correlation between the Proterozoic sequences of Greenland and Ellesmere Island and to examine the implications of this for the problem of displacement along Nares Strait. The paper is based on field investigations by all three authors; in Greenland by Dawes (1972, 1975, 1976a, 1979) and in Ellesmere Island by Christie (1962a, b, 1967) and Frisch (Frisch et al. 1978). Frisch also carried out comparative studies in Greenland in 1978 and 1980. The most recent regional descriptions of the Thule Group are given in Dawes (1976b) and Frisch & Christie (1982).

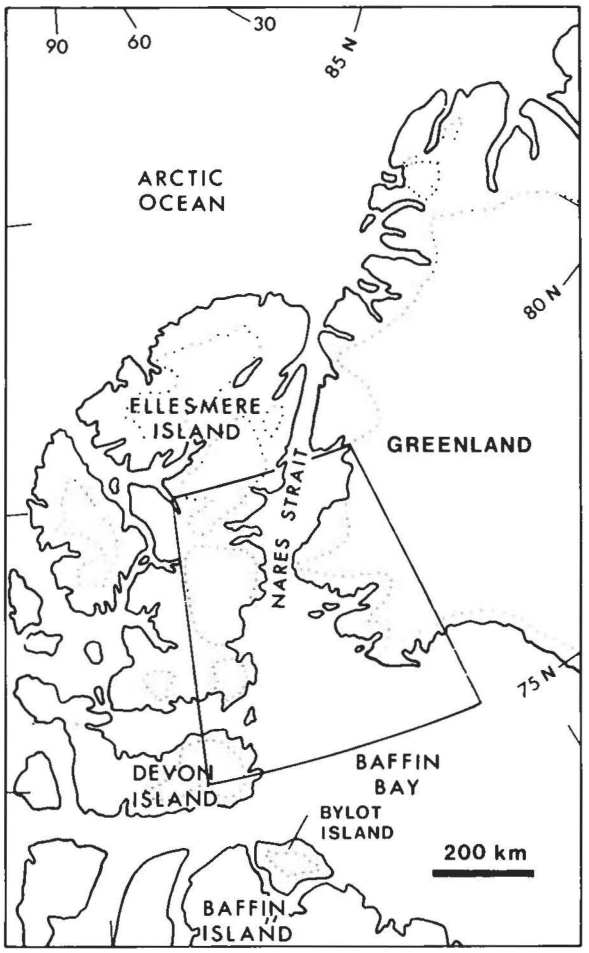


Fig. 1. Index map of the Nares Strait region with frame of the Smith Sound region as shown in Fig. 3. Main ice fields are indicated.

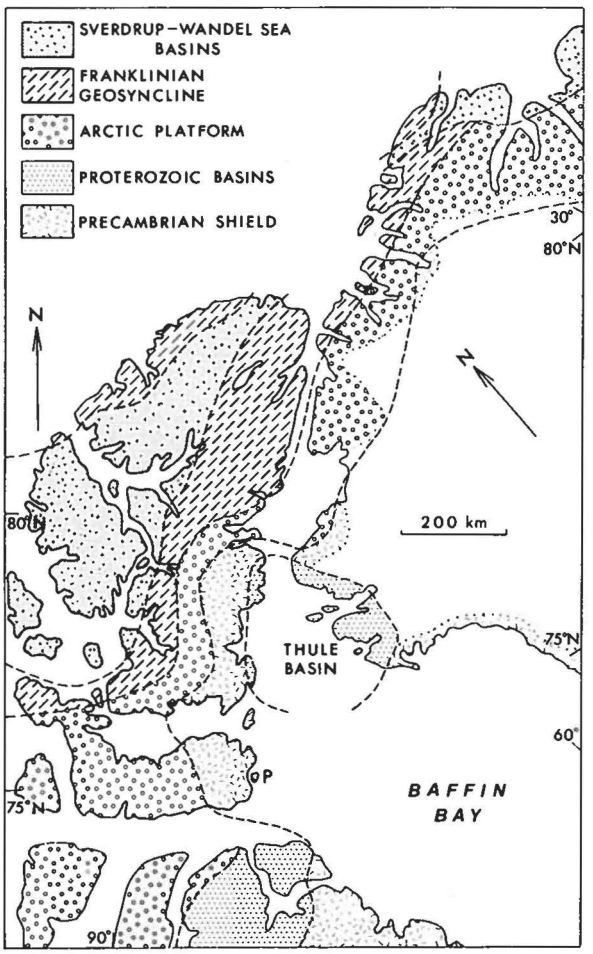


Fig. 2. Main geological provinces of the Nares Strait region showing the setting of the Thule Basin. P = Philpots Island.

The Thule Group – Thule Basin: concept and nomenclature

The unmetamorphosed Proterozoic rocks of the Smith Sound region consist of two geographically separated sequences: a thick southern sequence overlain only by Quaternary accumulations and a conspicuously thinner northern sequence overlain in both Greenland (Inglefield Land) and Ellesmere Island (Bache Peninsula) by Lower Palaeozoic strata. The thick southern sequence forms the Thule Basin (Kerr 1967, Christie 1972), while the thinner sequence (Rensselaer Bay Formation) forms part of the stable platform, south of the Franklinian geosyncline (Fig. 2). The structural high separating the Thule Basin from the platform has been termed the Bache Peninsula arch (Kerr 1967).

In Greenland there is relatively good exposure between the Thule Basin and the thinner platform sequence to the north, and Koch (1929, 1933) used the name Thule Formation to cover both sequences. However, in Ellesmere Island the two sequences are separated and no Thule Group strata are preserved over the Bache Peninsula arch (Fig. 3). This has led to some inconsistencies in stratigraphic nomenclature. Thus, some authors (e.g. Christie 1967, 1972) have restricted the name Thule Group to the thick succession of the Thule Basin, while others (e.g. Troelsen 1950) have included the thin Rensselaer Bay Formation of the platform as part of the Thule Group.

A nomenclatorial revision of the Thule Group based on the type area in Greenland is in preparation by one of us (PRD), but in the present paper the name Thule Group is retained and used to cover all the Proterozoic strata (Thule Basin and adjacent platform deposits) that overlie the Precambrian Shield in the Smith Sound region. The reader is referred to the paper by Peel et al. (this volume) for a description of the platform sequences north of the Thule Basin including a redefinition of the Rensselaer Bay Formation to which the Pro-

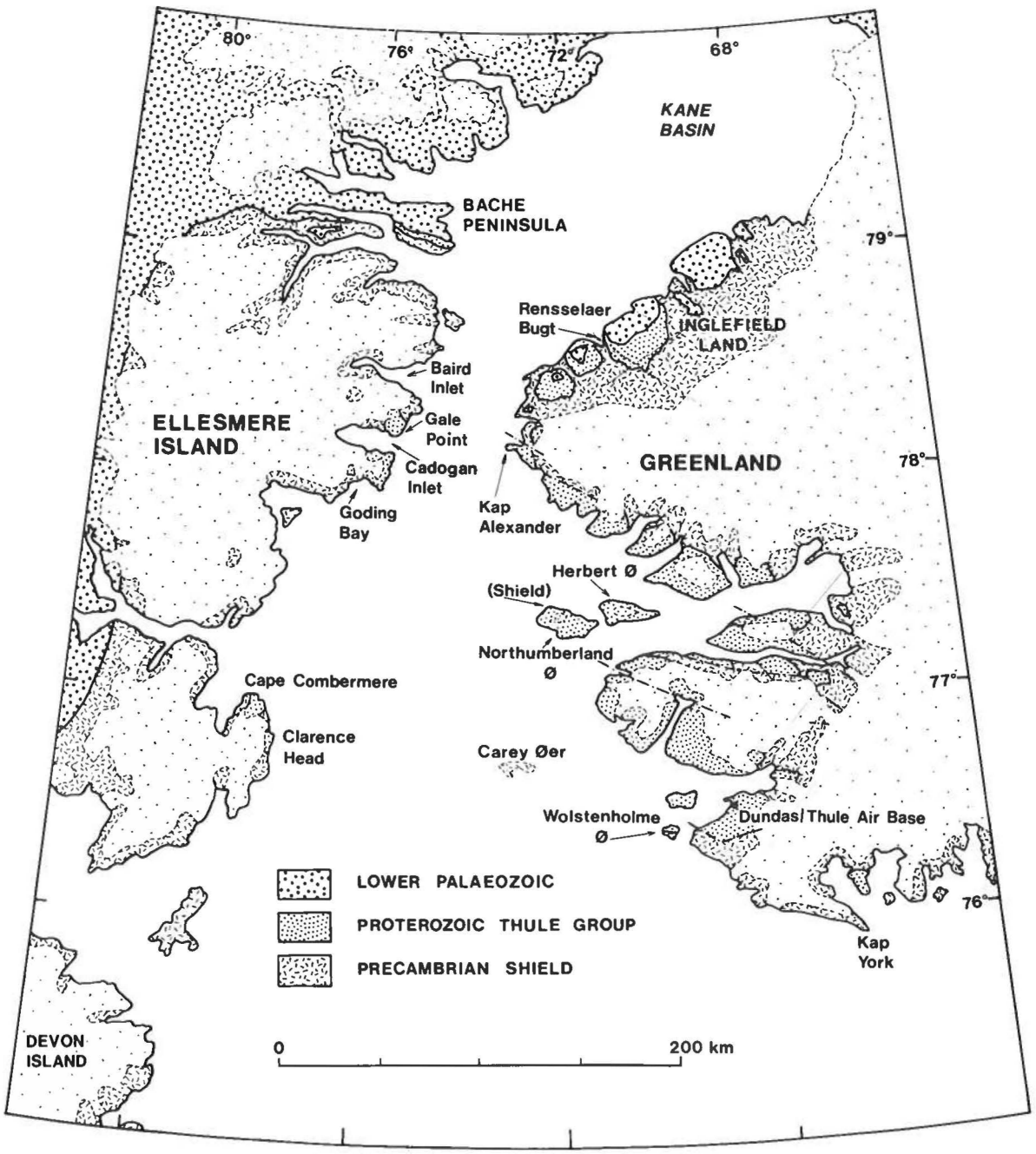


Fig. 3. Simplified geological map of the Smith Sound region.

terozoic strata of Bache Peninsula and Inglefield Land are referable.

Stratigraphy

The main outcrops of the Thule Basin are preserved in Greenland between Wolstenholme Ø (76°30′N) and Kap Alexander (78°10′N), and in Ellesmere Island between Clarence Head (76°45′N) and Cadogan Inlet (78°10′N) (Fig. 3). Lower Thule Group strata occur as an isolated outlier north of Kap York (75°55′N) and as a small patch on Ellesmere Island, north of Baird Inlet (78°30′N).

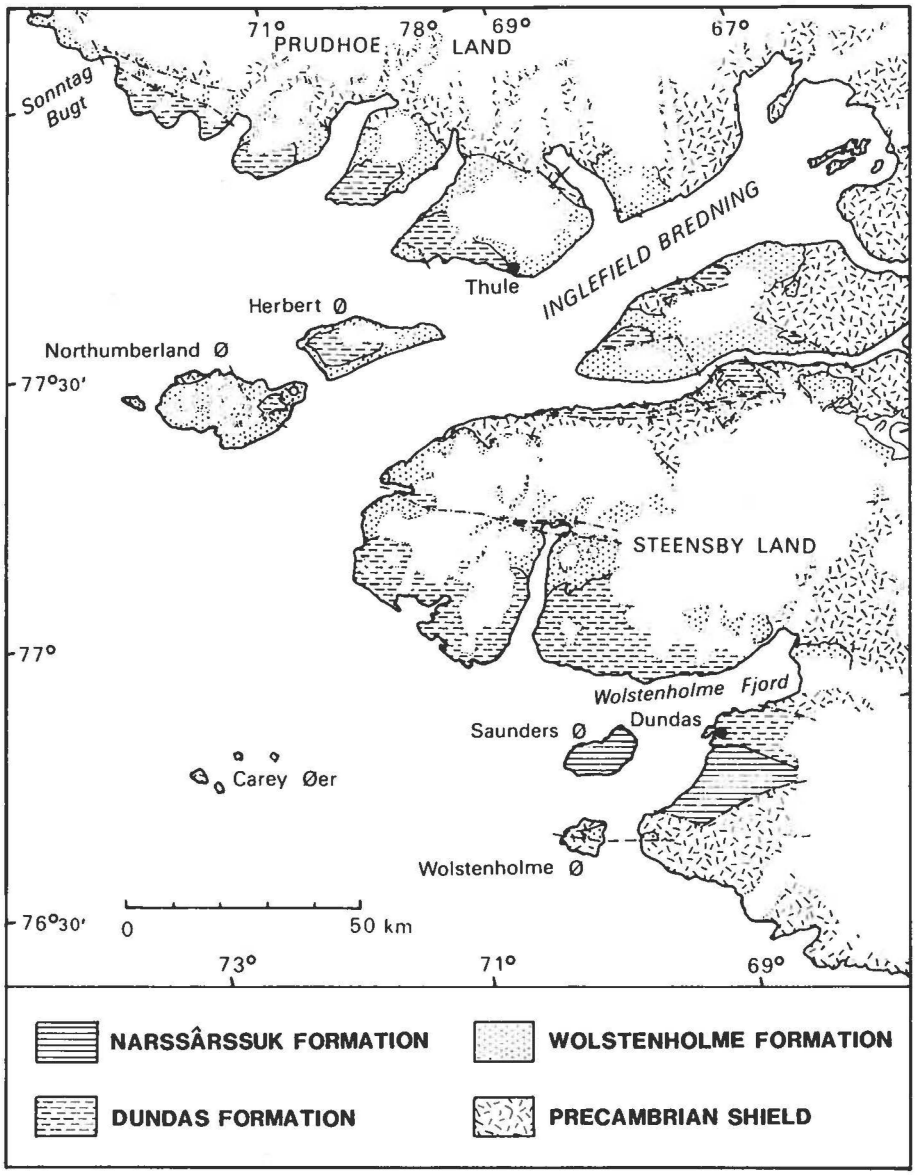


Fig. 4. Geological map of the Thule Basin in North-West Greenland.

The Thule Group has its thickest development in Greenland where the composite section reaches 4500 m. This is presently subdivided into three formations (Davies et al. 1963), i.e. from bottom to top, the *Wolstenholme Formation* (lower Thule Group) of sandstone and conglomerate with basaltic lavas and sills, the *Dundas Formation* of mainly shale and sandstone and the *Narssârssuk Formation* of cyclic carbonate-clastic deposits. The Dundas and Narssârssuk Formations (upper Thule Group) are confined to Greenland where they have restricted occurrences; in particular the Narssârssuk Formation only occurs in the region south of Dundas and on Saunders Ø (Fig. 4).

In south-eastern Ellesmere Island a little less than 1100 m of the Proterozoic sequence is preserved and all strata have been referred to the lower Thule Group or Wolstenholme Formation (Frisch et al. 1978). Hence, the definition of the Thule Basin as a depositional entity spanning southern Nares Strait is based solely on the lower Thule Group and thus only that part of the succession is relevant to the present discussion. However, since some workers (e.g. Newman & Falconer 1978, Newman, this volume) prefer to correlate the Thule Group in Greenland with the Proterozoic succession in northern Baffin Island, 500 km further to the south, a brief summary of the two formations of the upper Thule Group is given below to complete the description of the stratigraphy.

Upper Thule Group

The Dundas Formation of the upper Thule Group conformably overlies the Wolstenholme Formation with a gradational contact. The relationship between the Dundas and Narssârssuk Formations is uncertain although lithological and structural considerations suggest that the latter formation is all, or in part, of younger age (Dawes 1979).

The *Dundas Formation,* at least 1000 m thick, is a dark weathering, thin-bedded sequence of mainly alternating shale, siltstone and sandstone with some thin beds of calcareous shale, dolomite and, in the upper part, minor evaporite, chert and siliceous siltstone. The dolomites are stromatolitic in places and some small carbonate mounds occur. The formation is interpreted as a shallow-water, deltaic to coastal plain deposit.

The *Narssârssuk Formation* is a multicoloured, conspicuously banded, red bed sequence at least 2500 m thick. It is characterised by the cyclic alternation of dolomite, limestone, fine-grained sandstone, siltstone and shale with some thin evaporite, chert and cherty dolomite beds. A variety of stromatolite and algal-mat associations are present. The formation is a shallow-water deposit; its red bed, evaporitic and cyclic nature suggests rather stable lacustrine or lagoonal conditions, perhaps an environment analogous to modern sabkhas.

Both formations are intruded by basic igneous material. K/Ar whole-rock dates on dykes and sills show a range from 532 ± 20 m.y. to 727 ± 30 m.y. (Dawes et al. 1973). Acritarchs suggest a late Riphean age for the main thickness of the Dundas Formation, ranging to Vendian in its uppermost part, while the Narssârssuk Formation is regarded as Vendian in age throughout (Vidal & Dawes 1980).

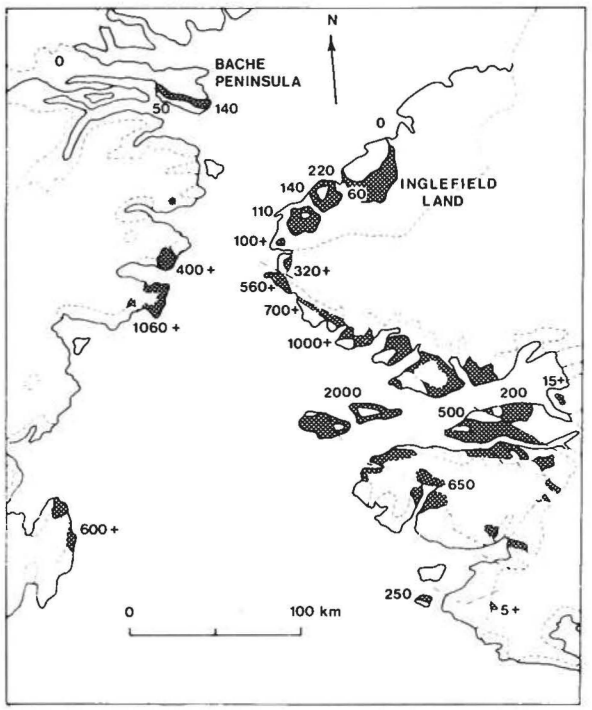


Fig. 5. Map showing the outcrops of the lower Thule Group (including the Rensselaer Bay Formation of Inglefield Land and Bache Peninsula). Asterix north of Baird Inlet represents a small outlier of the Thule Group. Thicknesses are in metres. A positive sign indicates a minimum thickness; either a minimum estimate or a detailed measured section, the upper limit of which is an erosional surface. Thicknesses include basic sills.

Lower Thule Group

The lower Thule Group has a composite thickness of over 2000 m, and its distribution is shown in Fig. 5. In both Greenland and Ellesmere Island it includes a wide range of rock types, both sedimentary and igneous. The lithological diversity clearly warrants higher stratigraphical status than that of the single formation accorded it in the literature, i.e. Wolstenholme Formation.

The succession consists essentially of a basal sandstone sequence overlain by basaltic extrusive and hypabyssal rocks which are interbedded with red beds. These are overlain by a much thicker clastic sequence of sandstone and quartz-pebble conglomerate with siltstone and shale. The succession displays lateral lithological and thickness variations which can best be studied in Greenland, where the largest part of the Thule Basin is preserved. However, a regional description of the lower Thule Group is outside the scope of this paper and for convenience the Northumberland Ø sequence is taken as a reference section for comparison with Ellesmere Island. This section is particularly suitable since its lower and upper contacts are preserved and it represents that part of the Greenland succession closest to the area in which the main Ellesmere Island outcrops occur.

Greenland – Northumberland Ø. – A complete sequence of the lower Thule Group is present on Northumberland Ø (Fig. 6). The succession has a tectonically undisturbed, unconformable contact with the crystalline basement on the north-western side of the island (Fig. 7) and it is overlain by the Dundas Formation in a downthrown fault block that forms the eastern end of the island. The succession has been divided into six main units (Dawes 1976a: fig. 8, and Fig. 6).

Unit 1 – Basal sandstone unit. This is composed of a light-weathering, generally medium-bedded, and in places cross-bedded, succession comprising red, purple, orange, brown and buff sandstones that are commonly variegated and mottled. Although generally recessive, some sandstone beds are characteristically cliff-forming (Fig. 7). Some purple and green shales with mud cracks and thin-bedded, fine-grained sandstones occur in the middle part. Basic sills are present in the lower part and

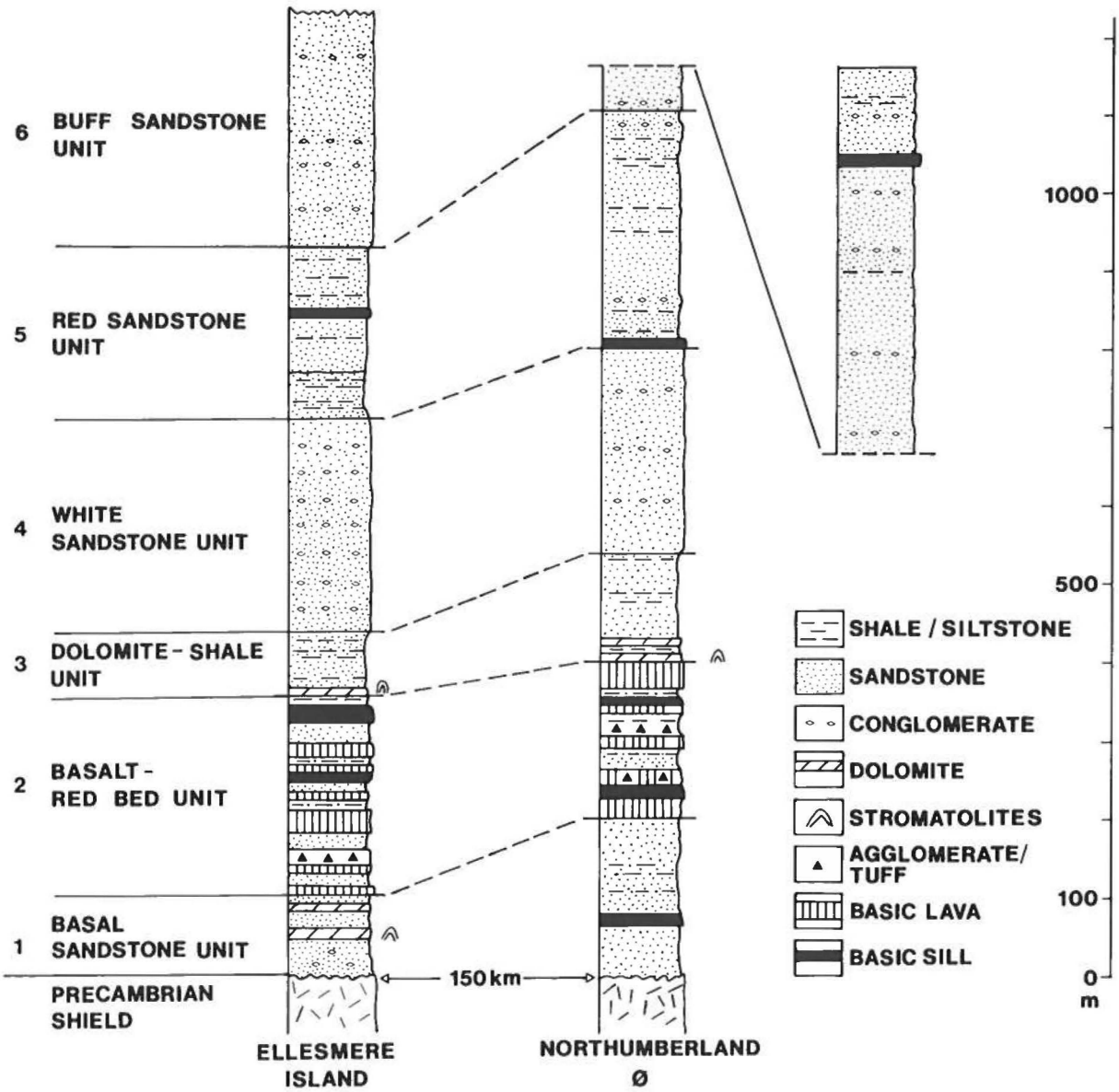


Fig. 6. Simplified stratigraphical correlation chart of the lower Thule Group on Ellesmere Island and Northumberland Ø, Greenland. The Canadian section is a composite of the three main outcrop areas (Clarence Head, Goding Bay and Gale Point) redrawn from Frisch et al. (1978); the Greenland section is modified from Dawes 1976a.

the thickness of the unit, including the sills, reaches 200 m.

Unit 2 – Basalt-Red bed unit. This is a dark-weathering, rather complex sequence of extrusive, hypabyssal, volcaniclastic and red bed rocks which is at least 200 m thick. It shows some lateral variation, but is characterised by having resistant lower and upper parts of basaltic flows and sills, between which is a dominantly recessive sedimentary sequence (Fig. 8). Several of the basaltic members have amygdaloidal tops and a rubbly volcanic appearance, but it is not everywhere apparent whether the igneous members are extrusive flows or intrusive sills.

The sedimentary rocks are thin-bedded and consist of reddish brown to purple tuffs, agglomerates and volcanic breccias which in places are interbedded with red to brown shales, siltstones and fine-grained sandstones.

Unit 3 – Dolomite-Shale unit is a dark-weathering, thin-bedded, recessive unit reaching a thickness of about 140 m (Fig. 8). In the lower part it consists of a variable sequence up to 25 m thick of resistant dolomite and dolomitic limestone, which is partly stromatolitic,



Fig. 7. Contact of the Precambrian Shield and Thule Group on the western side of Robins Gletscher, north-western Northumberland Ø, Greenland. Basal sandstone unit 1 (about 180 m) is overlain by the basalt-red bed unit 2. The cliff-forming layer forming the top of the section is a lava flow or basic sill.

particularly at the base, with maroon and red shales. These rocks are overlain by multicoloured (red, purple, green, buff), medium- to fine-grained sandstones with some grey green and red shales. The sandstones are cross-bedded and ripple marked. In one locality the middle and upper parts of the unit contain some fine-grained volcaniclastic and extrusive rocks in association with black shales characterised by abundant mud cracks.

Unit 4 – White sandstone unit. This consists of a light-weathering, resistant and cliff-forming sequence up to 260 m thick composed of mainly thick-bedded quartz sandstones or orthoquartzite with frequent quartz-pebble conglomerate beds. The overall colour of

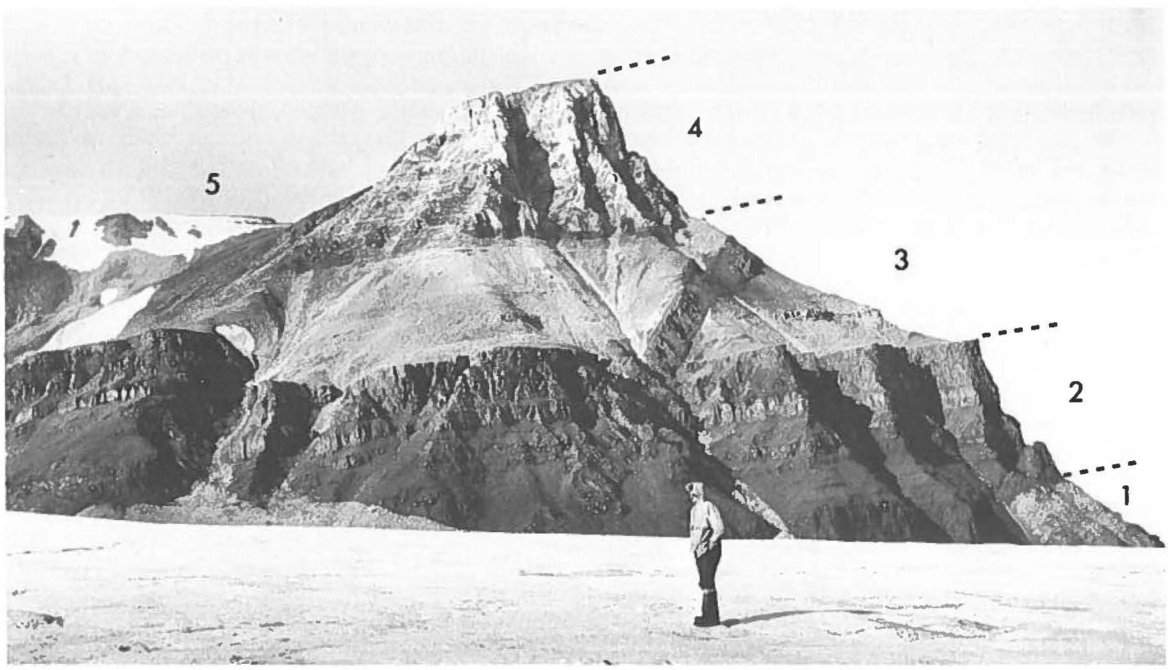


Fig. 8. The lower part of the Thule Group in a cliff section on the west side of Kissel Gletscher, north-western Northumberland Ø, Greenland. The Precambrian Shield is just out of view to the right at sea level. The summit of the headland at approximately 700 m above sea level is at the boundary with unit 5, part of which can be seen at the top of the ridge in the background. The sequence is cut by a late Proterozoic dolerite dyke.

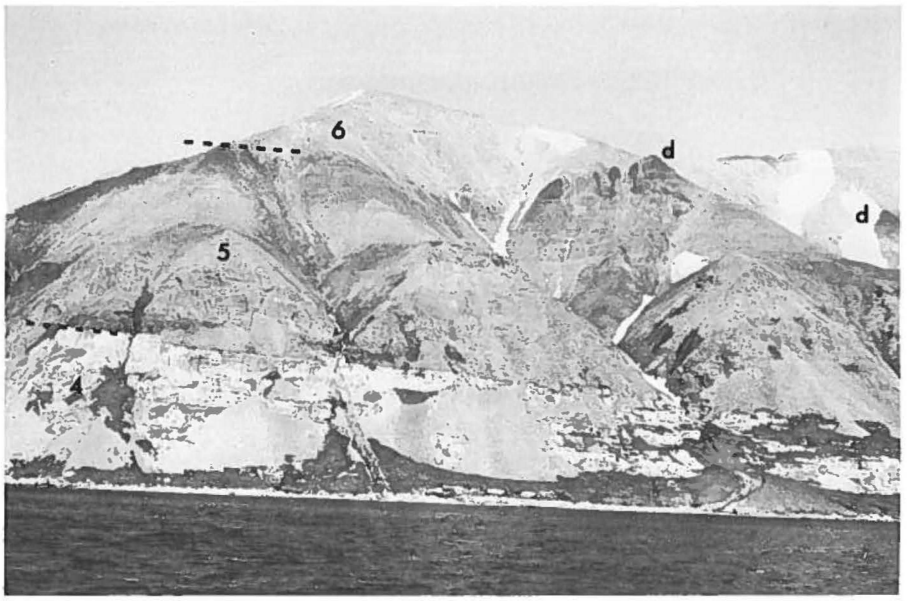


Fig. 9. Part of the south-western coast of Northumberland Ø in Greenland showing the upper units of the lower Thule Group. Note the sharp boundary between the white ortho-quartzites of unit 4 and the red beds of unit 5 and the more transitional boundary between units 5 and 6 — features that also characterise the Clarence Head section on Ellesmere Island (Fig. 10). The dark patches marked (d) are dolerite intrusions. The summit of the mountain is about 650 m above sea level.

the unit is light grey to white (Fig. 8). The sandstones are medium- to coarse-grained, white to pale pink on fresh surfaces and they are characterised throughout by cross-bedding and ripple marks. Some thin purple sandstones occur in the lower part of the unit where a purple and white banding, rather irregular and discordant to bedding, is a common feature.

Unit 5 – Red sandstone-shale unit. This is a variable clastic red bed sequence probably exceeding a thickness of 300 m and having a sharp contact with the underlying unit (Fig. 9). It is composed mainly of multicoloured (purple, red, brown, orange), thin- to medium-bedded and fine- to medium-grained, ferruginous sandstones which commonly show iron staining and liesegang rings. Some pale buff to lilac sandstones are present and pebbly sandstone and quartz-pebble conglomerate beds occur. The sandstones show frequent cross-bedding and ripple marks. At several levels, particularly near the base and the top, siltstones and grey, green and dark shales occur, either in discrete beds or as shaly partings in the sandstones.

Unit 6 – Buff sandstone unit. This is a thick and rather uniform, buff, yellow to pale lilac and pink weathering sequence of quartz sandstones with interbedded quartz-pebble conglomerates. The unit is commonly gradational with the underlying strata of unit 5 (Fig. 9). The sandstones are generally medium-grained and thin-bedded. However, towards the top they are finer-grained and become interbedded with darker siltstones and shales. Cross-bedding and ripple marks in the sandstones and mud cracks in the shales are common.

Unit 6 on Northumberland Ø is several hundred metres thick; the exact thickness is uncertain because of faulting. On neighbouring Herbert Ø the unit is estimated to be about 900 m thick.

Ellesmere Island. – Proterozoic strata comparable to the lower Thule Group of Greenland outcrop in coastal Ellesmere Island in three principal areas: Clarence Head, north of Goding Bay and at Gale Point (Fig. 3).

The thickest section is found on Goding Bay, but the lowermost beds are developed best near Gale Point. In terms of number of units exposed, the Clarence Head section is perhaps the most instructive (Fig. 6). The section given in Fig. 6, being a composite of the three main outcrop areas, is somewhat schematic but is readily applicable to any of the Ellesmere Island sections. As in Greenland, the section is composed of basal sandstones overlain by interbedded basalt and red beds, followed by variegated sandstone, siltstone and shale. All six units of the Northumberland Ø section are clearly recognisable.

The similarities to the strata in Greenland were pointed out by Christie (1962a, b), who subsequently assigned the Ellesmere Island strata to the Thule Basin sequence (Christie 1972). A full description of the Thule Group rocks on Ellesmere Island is given in a separate paper (Frisch & Christie 1982), but the salient features of the succession are enumerated below, with particular reference to the Clarence Head section.

Just south of Clarence Head, 660 m of strata overlie crystalline basement (Fig. 10). The contact appears to be a faulted unconformity and only a few metres of buff sandstone and conglomerate of the basal sandstone unit (unit 1) intervene between basement and a probable lava flow. A series of basalt sills and lavas interbedded

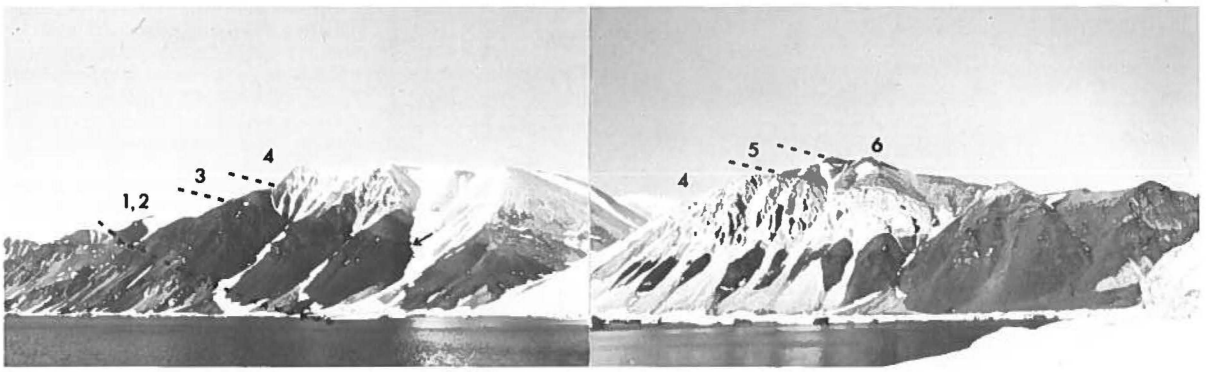


Fig. 10. View of the Clarence Head section, Ellesmere Island, seen from the east. The numbers refer to the six units into which the Thule Group has been divided and which correlate with units in Greenland. In this particular section only small thicknesses of units 1 and 6 are preserved and a basaltic sill has intruded between units 5 and 6. The sill lies about 600 m above sea level. The arrow points to the resistant stromatolitic carbonate bed at the base of unit 3.

with red siltstone or shale and sandstone, and at least one volcaniclastic bed of agglomerate with minor tuff, follow. The entire basalt-red bed unit (unit 2) is estimated to be a little more than 200 m thick.

The basal 13 m of the overlying strata (unit 3) at Clarence Head are stromatolitic dolomite and minor limestone, and these rocks form a resistant marker bed (Fig. 10). Above this lies a 66 m thick, mainly recessive sequence of thin-bedded, varicoloured, quartz-rich sandstone and red shale, interbedded. Lenticular bedding, herringbone cross-bedding, ripple marks and mud cracks are abundantly developed.

Strikingly white orthoquartzite, commonly conglomeratic, forms the next unit (unit 4) which is wedge-shaped, 100–200 m thick at Clarence Head and characteristically cliff forming. The lower beds are banded purple and white and cross-bedding and ripple marks occur throughout.

The overlying unit (unit 5) marks a return to red bed deposition that is marked by a sharp boundary — as it is on Northumberland Ø (Fig. 9). The unit consists of thin-bedded pink quartz sandstone with red micaceous siltstone partings and red shaly siltstone. Bedding surfaces are commonly veneered by specularite. At Clarence Head a basalt sill caps the unit.

The uppermost unit (unit 6) comprises 24 m of very pale pink quartz sandstone with an eroded upper surface.

Each of the units making up the Clarence Head section may be found in the other major exposures farther north on the Ellesmere Island coast, although there are variations. Thus, west of Gale Point (Fig. 11), nearly 50 m of quartz sandstone, conglomerate and several stromatolitic dolomite beds unconformably overlie crystalline basement and underlie the basaltic unit 2 in which sills predominate over lavas, and white orthoquartzite as well as red and green silty rocks occur. Also at this locality a lava flow occurs, capping unit 3. At Goding Bay, 12 m of black shale and thin basalt sills lie at the top of the basal unit 2 and the stromatolitic dolomite of unit 3. At this locality the uppermost buff sandstone unit is more than 300 m thick (top is an erosion surface) representing the unit's thickest preservation in Ellesmere Island.

Age and correlation: Northumberland Ø and Ellesmere Island

The Greenland and Ellesmere Island successions show an obvious similarity both in general lithostratigraphy and thickness, and the disposition of the lithological units and the corresponding red bed sedimentation levels makes a unit to unit correlation possible (Fig. 6). In addition, several specific points of comparison are notable in the Northumberland Ø and Clarence Head sections, e.g. the presence of pyroclastic beds (agglomerate, pisolitic tuffs) in the basalt-red bed unit 2; the stromatolite dolomites overlying the basalt-red bed unit; the intermittent occurrence of volcanic rocks within the dolomite-shale unit 3; and the purple banding in the lower part of the white sandstone unit 4.

Acritarchs from unit 4 on Northumberland Ø suggest a late Riphean age for the upper part of the sequence (Vidal & Dawes 1980). K/Ar ages obtained from the basaltic lavas and sills in Ellesmere Island and from basic sills in equivalent strata in Greenland cluster around 1100–1200 m.y. It is thus concluded that the lower part of the sequence, i.e. units 1 and 2 and probably unit 3, is about 1200 m.y. old, i.e. Helikian in North American terms, while the upper part of the sequence may be as young as late Riphean, i.e. Hadrynian in North American terms.

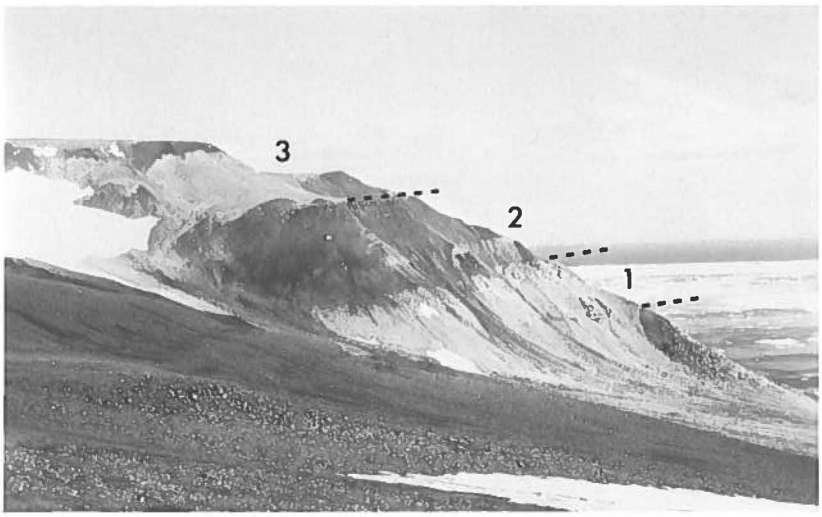


Fig. 11. The lower part of the Thule Group overlying basement rocks at Gale Point, Ellesmere Island, viewed eastwards. The units are designated as in the Clarence Head section (Fig. 10). A lava flow (dark) caps the recessive unit 3 and this is at about 550 m above sea level.

Thule Basin: sedimentary environment, limits and tectonics

The present outline of the Thule Basin is defined by the outcrops of the lower Thule Group in that no overlapping of the Dundas and Narssârssuk Formations (upper Thule Group) on to the Precambrian basement occurs, neither at the borders of the basin nor in uplifted fault blocks within it. The Thule Group strata form flat-lying to gently dipping homoclinal sections and dips greater than 30° are rare. Broad flexures and in some places folds due to fault tectonics occur.

The sedimentary structures of the lower Thule Group in both Greenland and Ellesmere Island indicate it to be a shallow-water, near-shore deposit representing fluvial to tidal environments. The presence of large units of nearly pure sandstone suggests substantial reworking and sorting, while the shale-silt-sand sequences may indicate intermittent more rapid basinal accumulation and migration seaward of the near-shore sites of deposition of the sandy facies. The red colouration could indicate *bona fide* continental oxidising conditions, but it is perhaps more likely to represent diagenetic adjustment of iron contained in the 'normal' clastic sedimentary material. Iron would be derived in abundance from the type of granitoid Archaean and Aphebian crystalline terrain that surrounded and underlay the Thule Basin.

Palaeocurrent directions have not been adequately determined on a systematic regional scale. In Greenland the main transport directions are to the south and south-west, away from the present outcrops of Precambrian basement (B. O'Connor, pers. comm.). However, some variation is present due to syndepositional faulting producing local source areas within the basin and fault blocks with contrasting sequences. A major westward trend characterises the transport direction of the Ellesmere Island sections indicating land somewhere to the east.

The lower Thule Group shows appreciable variation in thickness and lithology, but a correlation has been established throughout the Greenland and Ellesmere Island parts of the Thule Basin (Dawes 1976b: fig. 232, Frisch & Christie 1982: fig. 2). In addition, correlation has also been effected between the Thule Basin deposits in Greenland and the much thinner Rensselaer Bay Formation of Inglefield Land (Dawes 1979, Peel et al., this volume: fig. 3). These correlations establish that on a regional scale the succession in Greenland thins to the north, east and south. Hence, the thickest and also most complete succession of the lower Thule Group occurs in the west, on Northumberland Ø and adjacent Herbert Ø. The lowest units of the Northumberland Ø succession (the basal sandstone, basalt-red bed and the dolomite-shale units) peter out to the east and south where the younger sandstone units overlap on to the crystalline Shield.

In Ellesmere Island the limited extent of the outcrops of the Thule Group isolated in large areas of permanent snow and ice makes regional stratigraphical comparisons difficult. In addition, all the sections end in an erosion surface and the absolute thickness of the Thule Group is unknown. However, judged purely from total thicknesses of sections preserved, there is a stratigraphical thinning to the north and south away from Goding Bay (Fig. 5). Furthermore, contacts with the Precambrian basement in the west are generally covered. There is no evidence of westward thinning and it appears most likely that the Thule Basin is fault bounded to the west. Faults cutting the succession, although numerous, show no preferred orientation.

The Thule Basin in Greenland is cut by prominent faults first reported by Koch (1926). Major sets trend WNW and NW and cut the Thule Group into a number

of fault blocks of which several are southerly tilted. A major fault in the south juxtaposes the youngest strata of the Thule Group against crystalline basement and involves a northerly downthrow of several kilometres (Fig. 4). Undisturbed stratigraphic contacts between the lower Thule Group and the Precambrian Shield are preserved south of this fault. Similar contacts are preserved around the entire basin, except in the extreme north, to the south-east of Sonntag Bugt in northern Prudhoe Land (78°N) (Fig. 4). This area is traversed by coast-parallel faults characterised by south-westerly downthrow juxtaposing the lower Thule Group on the seaward side against crystalline rocks.

Fig. 5 shows the measured or estimated thicknesses of the lower Thule Group in Greenland and Ellesmere Island. The southern limit of the Thule Basin is undefined although on Wolstenholme Ø, Greenland, where the lower part of the lower Thule Group is missing, the total thickness of the succession (limited upwards by the Dundas Formation of the upper Thule Group) decreases to a couple of hundred metres. In addition, thin veneers of lower Thule Group occur as scattered outliers on the crystalline basement terrain, perhaps suggesting the southern termination of the basin not far to the south. This parallels the situation along the eastern margin of the basin where the upper part of the lower Thule Group oversteps on to the crystalline basement and is preserved in thin scattered outcrops.

On the Canadian side, the southern limit of the Thule Basin probably lies between Clarence Head (76°48′N), where the southernmost outcrops of Thule Group are preserved, and Philpots Island, off the eastern coast of

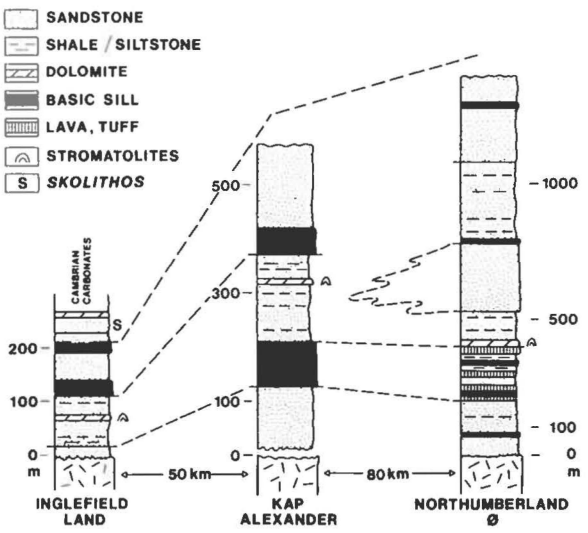


Fig. 12. Stratigraphical correlation scheme of the Proterozoic successions of the Thule Basin (Northumberland Ø and Kap Alexander) and the platform (Inglefield Land). Note the smaller metre scale for the Northumberland Ø section.

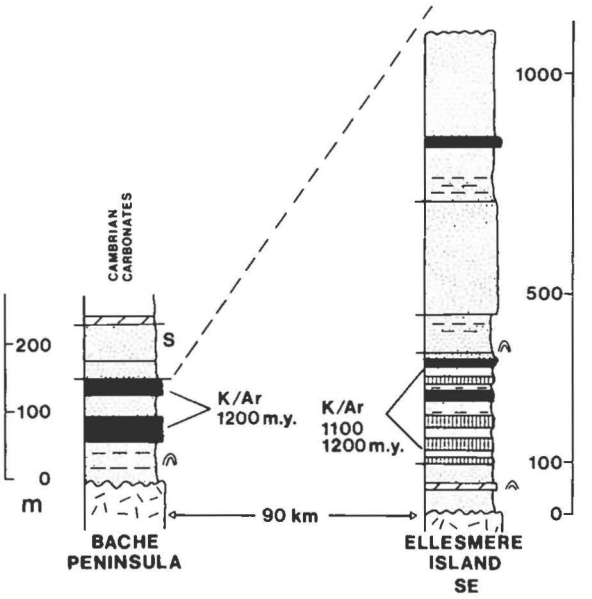


Fig. 13. Stratigraphical correlation scheme between the Proterozoic successions of south-eastern Ellesmere Island and Bache Peninsula. Symbols are explained in Fig. 12.

Devon Island at 75°N (Fig. 2), where Lower Palaeozoic rocks lie directly on crystalline basement (Frisch 1981).

The margin of the Thule Basin is best preserved in the north, although the paucity of outcrop on the Ellesmere Island side prevents precise overall definition. Immediately north of Kap Alexander in Greenland, the lower Thule Group strata thin markedly over a few kilometres from a section approaching a thickness of 600 m, of which the base and top are not exposed, to a sequence of about 300 m. The location of this significant thickness change is now marked by a major fault with downthrow to the south and it is taken to represent the northern margin of the Thule Basin. A general correlation scheme between the Thule Basin succession and the much thinner platform section of Inglefield Land is presented in Fig. 12.

The northernmost occurrence of Thule Basin rocks known in Ellesmere Island lies north of Baird Inlet at 78°32′N where a few tens of metres of sandstone and stromatolitic carbonate rock, very similar to the basal rocks of the Gale Point section (40 km to the south), rest on crystalline basement in a downdropped fault block (Fig. 5). This isolated outcrop is near mid-way between the Gale Point section and Bache Peninsula, a distance of about 90 km. This occurrence, suggesting northward thinning of the lower Thule Group, and the same lithological contrast of the succession to the thinner correlatable strata of Bache Peninsula as seen in Greenland (Figs 5, 12 and 13; see below) indicates that the northern limit of the Thule Basin in Ellesmere Island lies between 78°30′N and 79°N.

Inglefield Land – Bache Peninsula correlation

A full description and correlation of the thin platform sequences to the north of the Thule Basin are dealt with elsewhere in this volume (Peel et al.). The close similarities between the Inglefield Land and the Bache Peninsula sections have long been established and Troelsen (1950), who examined both sequences, had no hesitation in correlating and using Greenland formational names for the Ellesmere Island section; a correlation confirmed by all subsequent workers (e.g. Cowie 1961, Christie 1967).

The Precambrian crystalline basement in both regions is overlain with marked angular unconformity by unmetamorphosed red bed clastic rocks with minor shale and dolomite — the redefined Rensselaer Bay Formation (Fig. 14). The Proterozoic age assignment of this 0 to 220 m thick sequence is based on its presence unconformably below early Cambrian sandstone and dolomites and its penetration by basic sills that have yielded Precambrian K/Ar whole-rock ages. In Greenland the available K/Ar dates fall in the range 1073 ± 40 to 1190 ± 40 m.y. (Dawes et al. 1973), strongly suggesting that the sills are products of the same magmatism that affected the Thule Basin. In Bache Peninsula two basalt sills intrude the lower part of the sequence (Christie 1967: plate II). Weathering and frost-shattering have been intense at the igneous contacts and sampling for K/Ar dating was possible only near, but not at, the sill margins. All samples show alteration of one or more of the mafic minerals (olivine and clinopyroxene), but those from the lower sill are fresher. One sample from the latter has given 1197 ± 33 m.y.; determinations on two samples from the upper sill gave 1463 ± 77 m.y. and 1369 ± 66 m.y. The 1197 m.y. date is preferred as being nearest to the time of basic intrusion on Bache Peninsula.

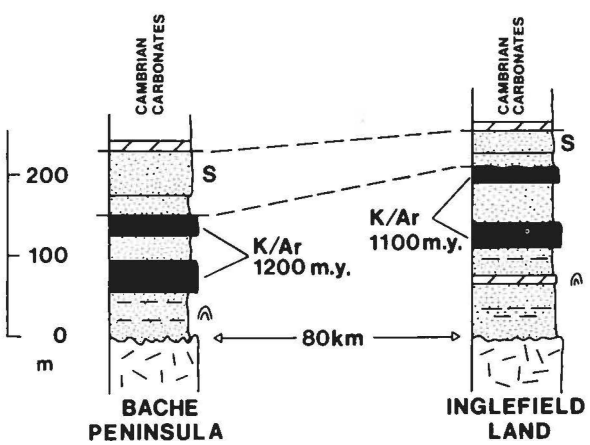


Fig. 14. Stratigraphical correlation scheme of the platform successions of Bache Peninsula and Inglefield Land, showing the Rensselaer Bay Formation overlain by Cambrian clastic and carbonate rocks. Symbols are explained in Fig. 12.

Inferences on Nares Strait

Summary of data

The detailed comparison that can now be drawn between the Proterozoic successions of the Smith Sound region brings out the following main features.

1. The late Proterozoic Thule Group section at Clarence Head in Ellesmere Island can be correlated unit by unit with the lower part of the Thule Group on Northumberland Ø, Greenland.
2. Similarity in physical characteristics between the Greenland and Ellesmere Island sections extends to features, measurable in terms of metres, which occur at the same stratigraphical level in each section.
3. The Northumberland Ø and Clarence Head sections correlate respectively with others to the north, south and east in Greenland and to the north in Ellesmere Island (no Thule Group rocks outcrop to the south of Clarence Head).
4. The lower Thule Group in Greenland is thickest in the Northumberland Ø – Herbert Ø area (77°30'N); in Ellesmere Island around Goding Bay (78°N). In Greenland, based on true thicknesses, and in Ellesmere Island based purely on total thickness of strata preserved, there is stratigraphical thinning north and south of these maxima.
5. K/Ar ages of extrusive and hypabyssal basalts in the lower Thule Group on both sides of northern Baffin Bay are closely comparable (1000 to 1200 m.y.).
6. The late Proterozoic sequences of Inglefield Land (78°–79°N) find their close counterparts on Bache Peninsula in Ellesmere Island (79°N) in lithology, thickness and the presence of 1100–1200 m.y. old basaltic sills.
7. The entire Inglefield Land section intruded by the basic sills can be traced, despite fault complications, over the Bache Peninsula arch to correlate with the lower Thule Group of Northumberland Ø.

The above features strongly suggest that the Proterozoic sequences of the Smith Sound region define a single, sedimentary basin (the Thule Basin) developed in an overall platform environment. This depocentre is intracratonic, and lies on the northern part of the North American craton. To the north it is bordered by the stable sedimentary platform and by the pericratonic orogenic belt of the Franklinian geosyncline (Figs 2 and 15). The northern margin of the Thule Basin lies at about 78°15'N in Greenland and between 78°30'N and 79°N in Ellesmere Island. This margin can be traced with little displacement across Smith Sound and there is

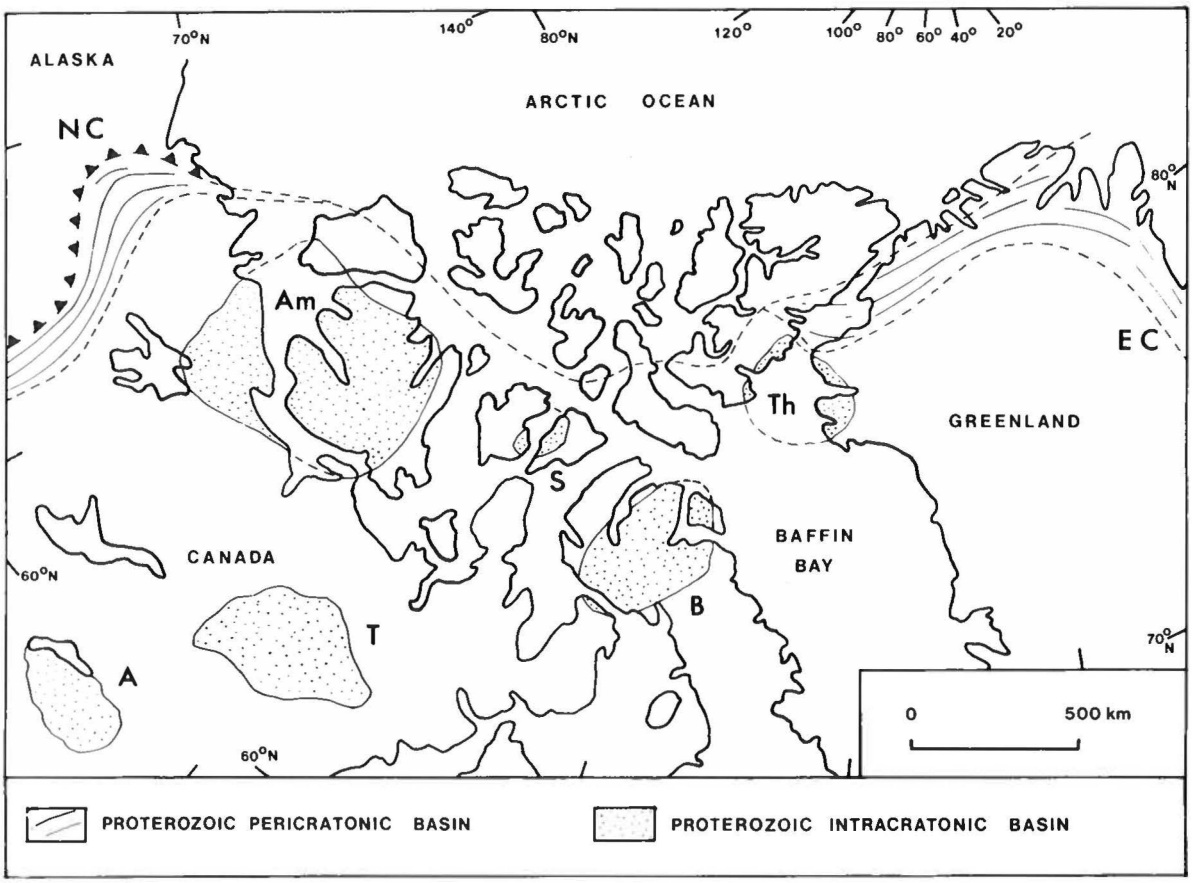


Fig. 15. Map of the northern part of the North American craton showing the distribution of the Proterozoic intracratonic basins and the linear pericratonic orogenic belts. Th = Thule Basin, B = Borden Basin, S = Somerset Basin, Am = Amundsen Basin, T = Thelon Basin, A = Athabasca Basin, NC = Northern Cordillera, EC = East Greenland Caledonides.

clearly no need to envisage any movement along Nares Strait to satisfactorily explain the feature or the thickness and general disposition of the Proterozoic sequences (Figs 5 and 16).

Thule Basin – Borden Basin correlation

A common criticism raised to those studies favouring direct geological continuity from Greenland to Ellesmere Island, as opposed to major offset along Nares Strait, is the value placed on the correlation of homoclinal platform strata. A characteristic of platform successions is that individual units can be traced over large areas and thus there are obvious limitations of correlations based on such units in cases for or against lateral movement between land masses on which the strata are found. However, in the Smith Sound region it is not simply a case of general lithological correlation of layer-cake platform successions. The correlation includes a thick intracratonic basin sequence which is flanked on the north by a thin platform section, both of which can be correlated in detail directly across Nares Strait.

On the other hand, it is surprising that generalised correlations of homoclinal strata have been made in support of the hypotheses that Greenland has moved some 300 km northwards relative to Canada along Nares Strait. Thus, the ~1200 m.y. old sedimentary and igneous rocks of the Borden Basin in northern Baffin Island and Bylot Island (Jackson et al. 1978, see Fig. 2) have been so correlated with the Thule Group of North-West Greenland (Newman 1977, this volume). While it is clear that the Borden Basin succession shows some general similarity to the Thule Group (including strata of similar facies to the upper Thule Group), there is a comparable similarity between both basin successions and the other intracratonic basins along the northern rim of the North American craton (Christie 1972, Young 1979, Fig. 15). In any case, the similarities between the Thule and Borden successions are in no way so close as to permit the unit to unit correlation that exists between Ellesmere Island and Greenland. Fur-

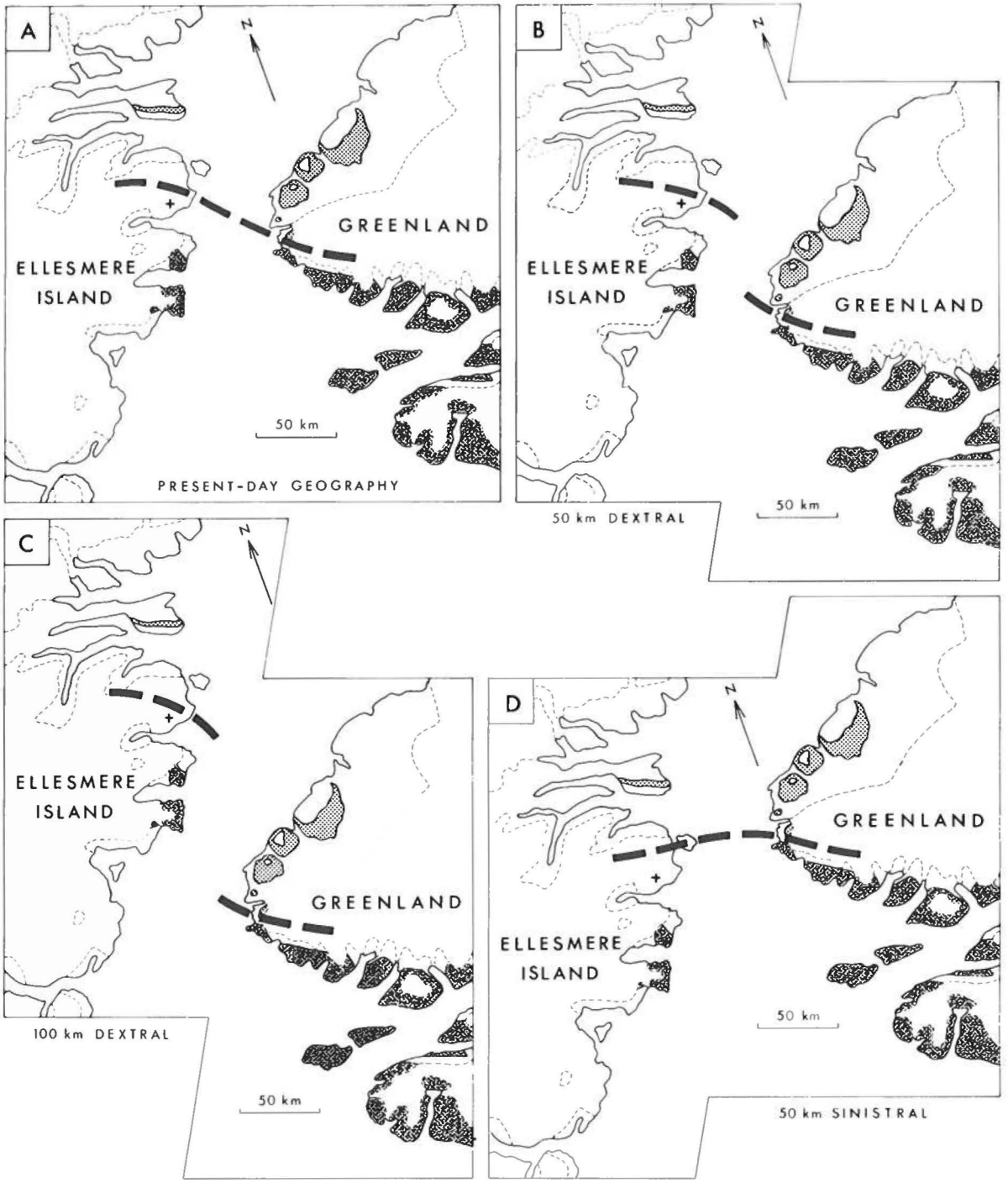


Fig. 16. The present-day position of Greenland and Ellesmere Island compared to reconstructions assuming strike-slip movement along Nares Strait. Shading indicates the outcrop of the Thule Group in the Thule Basin (dark) and the thinner Rensselaer Bay Formation of the platform (light). Cross, north of Baird Inlet, indicates northernmost outcrop of Thule Basin deposits in Ellesmere Island. The heavy broken line represents the northern margin of the Thule Basin. The present-day position (A) or that showing a sinistral movement of 50 km (D) appear to be the best fits with regard to the Proterozoic geology.

thermore, Jackson & Iannelli (1981) regard the Borden and Thule Basins as separate and distinct, albeit coeval, depocentres — an interpretation with which we concur.

The absence of upper Thule Group strata in Ellesmere Island (perhaps seen at first to be a significant point by which to dissociate the Ellesmere Island and Greenland successions) is of little relevance to the discussion, particularly in view of the restricted Proterozoic outcrop preserved in Ellesmere Island and the rather local development of the upper Thule Group in Greenland. However, the southern margin of the Thule Basin is poorly defined and it is to be expected that the Thule Group sedimentation covered a larger area than the southernmost known onshore outcrops, viz. 76°45'N in Ellesmere Island and 76°10'N in Greenland (Fig. 3). Whether the Thule Basin deposits extended so far south as to constitute what would be an extensive Borden–Thule depositional site is unknown.

The long-range correlation of the Proterozoic sequences of the Thule Group and Borden Basin that leads to a 300 km displacement of Greenland to the south relative to Canada is highly speculative and it violates the known Proterozoic geology of the Smith Sound region. Such a correlation indirectly defines the Ellesmere Island Proterozoic sequences as an isolated intracratonic basin–platform system. In such a reconstruction the Precambrian Shield and the small intracratonic basin of Ellesmere Island are placed in juxtaposition against the Silurian shelf and slope limestone and shale deposits of Greenland. Attempts to reconcile this unacceptable arrangement of unassociated geological provinces by suggestions of vertical displacement along Nares Strait, leading to substantial differential erosion of Greenland and Ellesmere Island, are unconvincing. The peneplain surface of the Precambrian crystalline basement on which the Thule Group rests is well developed and in south-west Inglefield Land (Koch 1933, Troelsen 1950, Cowie 1961) and Bache Peninsula (Christie 1967) it has a variable but comparable altitude. Hence, appreciable differential vertical movement of Ellesmere Island and Greenland in the Smith Sound area can be ruled out.

Constraints on strike-slip displacement

We suggest that the correlation of Proterozoic sequences of the Smith Sound region refutes pre-drift reconstructions showing major net strike-slip displacement along Nares Strait. However, an accurate estimate in kilometres of the offset permissible is not possible with the reliability level of the data available. The incomplete sections in Ellesmere Island make it impossible to construct other than crude isopachs of the lower Thule Group across Nares Strait. Thus we choose the northern margin of the Thule Basin as a geological marker and represent this in Fig. 16. A dextral movement of as little as 50 km along Nares Strait produces a significant displacement of this marker (Fig. 16B) and, without evoking an unreasonable shape for the Thule Basin, such reconstructions for the Proterozoic as shown in Fig. 16C can be excluded. Carried one stage further this exercise shows that less stringent controls are present on possible strike-slip displacement when Greenland is offset sinistrally from its present-day position. Thus, a relative position of the two landmasses as shown in Fig. 16D can be accommodated without appreciable disruption of the Proterozoic geological pattern.

Conclusion

The proliferation of plate tectonic models which envisage appreciable amounts of strike-slip displacement along Nares Strait has doubtless been facilitated by the scarcity of published information concerning the geology of Nares Strait and northernmost Baffin Bay. This deficit has been corrected by recent field work in the Smith Sound region, where onshore Proterozoic sequences indicate that the present-day relative positions of Greenland and Ellesmere Island cannot be far removed (if at all) from those present at the time of deposition. In consequence, it is the main conclusion of this paper that there has been little net displacement along Nares Strait — a conclusion supported by other geological studies reported in this volume of the Archaean and Palaeozoic geology of Greenland and Ellesmere Island.

Acknowledgements

We are grateful to J. S. Peel for critical reading of the manuscript. T. F. thanks the Polar Continental Shelf Project for logistic support and G. R. Dunning and W. C. Morgan for collaboration in the field. P. R. D. thanks K. Thomsen for able field assistance. The paper is published with the permission of the Director of the Geological Survey of Greenland.

References

Carey, S. W. 1958. The tectonic approach to continental drift. – In: Carey, S. W. (convener), Continental drift. A symposium: 177–355. – Univ. Tasmania, Hobart.

Christie, R. L. 1962a. Geology, Alexandra Fiord, Ellesmere Island, District of Franklin (map with marginal notes) 1'' to 4 miles. – Map geol. Surv. Can. 9–1962.

Christie, R. L. 1962b. Geology, southeast Ellesmere Island, District of Franklin (map with marginal notes) 1'' to 4 miles. – Map geol. Surv. Can. 12–1962.

Christie, R. L. 1967. Bache Peninsula, Ellesmere Island, Arctic Archipelago. – Mem. geol. Surv. Can. 347: 63 pp.

Christie, R. L. 1972. Central stable region. – In: Christie, R. L., Cook, D. G., Nassichuk, W. W., Trettin, H. P. & Yorath, C. J. (eds), The Canadian Arctic Islands and the Mackenzie region. – 24 Intern. geol. Congr. Montreal, Excurs. guide A66: 40–87.

Cowie, J. W. 1961. Contributions to the geology of North Greenland. – Meddr Grønland 164(3): 47 pp.

Davies, W. E., Krinsley, D. B. & Nicol, A. H. 1963. Geology

of the North Star Bugt area, Northwest Greenland. – Meddr Grønland 162(12): 68 pp.

Dawes, P. R. 1972. Precambrian crystalline rocks and younger sediments of the Thule district, North Greenland. – Rapp. Grønlands geol. Unders. 45: 10–15.

Dawes, P. R. 1975. Reconnaissance of the Thule Group and underlying basement rocks between Inglefield Bredning and Melville Bugt, western North Greenland. – Rapp. Grønlands geol. Unders. 75: 34–38.

Dawes, P. R. 1976a. 1: 500 000 mapping of the Thule district, North-West Greenland. – Rapp. Grønlands geol. Unders. 80: 23–28.

Dawes, P. R. 1976b. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R. 1979. Field investigations in the Precambrian terrain of the Thule district, North-West Greenland. – Rapp. Grønlands geol. Unders. 95: 14–22.

Dawes, P. R., Rex, D. C. & Jepsen, H. F. 1973. K/Ar whole rock ages of dolerites from the Thule district, western North Greenland. – Rapp. Grønlands geol. Unders. 55: 61–66.

Frisch, T. 1981. Further reconnaissance mapping of the Precambrian Shield on Devon Island, District of Franklin. – Pap. geol. Surv. Can. 81–1A: 31–32.

Frisch, T. & Christie, R. L. 1982. Stratigraphy of the Proterozoic Thule Group, southeastern Ellesmere Island, Arctic Archipelago. – Pap. geol. Surv. Can. 81–19: 13 pp.

Frisch, T., Morgan, W. C. & Dunning, G. R. 1978. Reconnaissance geology of the Precambrian Shield on Ellesmere and Coburg islands, Canadian Arctic Archipelago. – Pap. geol. Surv. Can. 78–1A: 135–138.

Jackson, G. D. & Iannelli, T. R. 1981. Rift-related cyclic sedimentation in the Neohelikian Borden Basin, northern Baffin Island. – In: Campbell, F. H. A. (ed.), Proterozoic basins of Canada. – Pap. geol. Surv. Can. 81–10: 269–302.

Jackson, G. D., Iannelli, T. R., Narbonne, G. M. & Wallace, P. J. 1978. Upper Proterozoic sedimentary and volcanic rocks of northwestern Baffin Island. – Pap. geol. Surv. Can. 78–14: 15 pp.

Kerr, J. W. 1967. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Koch, L. 1926. A new fault zone in northwest Greenland. – Am. J. Sci. 12: 301–310.

Koch, L. 1929. Stratigraphy of Greenland. – Meddr Grønland 73,2(2): 205–320.

Koch, L. 1933. The geology of Inglefield Land. – Meddr Grønland 73,1(2): 38 pp.

Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Newman, P. H. 1982. A geological case for movement between Canada and Greenland along Nares Strait. – This volume.

Newman, P. H. & Falconer, R. K. H. 1978. Evidence for movement between Greenland and Canada along Nares Strait. – Geol. Soc. Am. Abs. with Prog. 10: 463 only.

Peel, J. S., Dawes, P. R., Collinson, J. D. & Christie, R. L. 1982. Proterozoic – basal Cambrian stratigraphy across Nares Strait: correlation between Inglefield Land and Bache Peninsula. – This volume.

Sclater, J. G., Hellinger, S. & Tapscott, C. 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. – J. Geol. 85: 509–552.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Taylor, F. B. 1910. Bearing of the Tertiary mountain belt on the origin of the earth's plan. – Bull. geol. Soc. Am. 21: 179–226.

Troelsen, J. C. 1950. Contributions to the geology of Northwest Greenland, Ellesmere Island and Axel Heiberg Island. – Meddr Grønland 149(7): 85 pp.

Vidal, G. & Dawes, P. R. 1980. Acritarchs from the Proterozoic Thule Group, North-West Greenland. – Rapp. Grønlands geol. Unders. 100: 24–29.

Wegener, A. 1922. Die Entstehung der Kontinente und Ozeane. – Friedr. Vieweg & Sohn, Braunschweig: 144 pp. (3rd revised edit.).

Wilson, J. T. 1963. Continental drift. – Scient. Am. 208(4): 86–100.

Young, G. M. 1979. Correlation of middle and upper Proterozoic strata of the northern rim of the North Atlantic craton. – Trans. Roy. Soc. Edin. 70: 323–336.

Proterozoic – basal Cambrian stratigraphy across Nares Strait: correlation between Inglefield Land and Bache Peninsula

JOHN S. PEEL, PETER R. DAWES, JOHN D. COLLINSON and ROBERT L. CHRISTIE

Peel, J. S., Dawes, P. R., Collinson, J. D. & Christie, R. L. 1982. Proterozoic – basal Cambrian stratigraphy across Nares Strait: correlation between Inglefield Land and Bache Peninsula. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 105–115.

In the Kane Basin region of Nares Strait (78° to 79°N) unmetamorphosed Proterozoic and basal Cambrian platform strata overlie the Precambrian Shield both in Inglefield Land in Greenland, and on Bache Peninsula in Ellesmere Island. The platform succession is composed of two shallow-water clastic sequences separated by an erosional disconformity. The lower formation, of multicoloured sandstone and siltstone with minor shale and stromatolitic dolomite, is assigned to the redefined Rensselaer Bay Formation; the upper, composed of red and overlying white sandstone with dolomitic sandstone and siltstone at the top, is assigned to the Dallas Bugt Formation. Basic sills from the Rensselaer Bay Formation have yielded isotopic ages as old as 1200 m.y., while the trace fossils *Rusophycus* and *Skolithos* suggest a basal Cambrian age for the Dallas Bugt Formation.

The platform sequences of Inglefield Land and Bache Peninsula show striking similarity in lithology and thickness and the detail correlation is supported by the evidence from isotopic ages of basic sills and from trace fossils. In both areas the bipartite succession is overlain with apparent conformity by dolomites that pass upwards into limestones containing Early Cambrian macrofauna.

The Rensselaer Bay Formation thickens southwards and passes into the much thicker sedimentary sequence of the Thule Basin, while the Dallas Bugt Formation and the overlying dolomites thicken northwards, indicating that the two formations were laid down in separate sedimentary basins. The close correlation of the platform successions and the corresponding configuration of the depositional slopes on both sides of the Kane Basin clearly indicate that the Proterozoic – Early Cambrian sedimentary pattern has not been radically disturbed by tectonism or by any appreciable transcurrent movement along Nares Strait.

J. S. Peel and P. R. Dawes, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK–1350 Copenhagen K, Denmark, J. D. Collinson, Department of Geology, University of Keele, Keele, Staffordshire ST5 5BG, England and R. L. Christie, Department of Energy, Mines and Resources, Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303–33rd St. N.W., Calgary, Alberta, Canada, T2L 2A7.

A striking feature of the unmetamorphosed Proterozoic and Cambrian sequences around Kane Basin (Fig. 1) is the close similarity between clastic and carbonate strata overlying the crystalline basement in Inglefield Land, Greenland, and on Bache Peninsula, Ellesmere Island. The similarity was noted by Schei (Holtedahl 1917), Bentham (1936a, b, 1941) and Wordie (1938), but is best known through observations in both areas made by J. C. Troelsen as a member of The Danish Thule–Ellesmere Land Expedition, 1939–41. Troelsen's preliminary results (Poulsen 1940) were expanded by Poulsen (1946: 304) whose correlations indicated "that the lands round the Kane Basin belong to one and the same morphological unit." Later stratigraphic descriptions by Troelsen (1950) emphasised the similarities, with several formations being recognised as common to both areas. Subsequent field work by Christie (1967) on Bache Peninsula and by J. D. C., P. R. D. and J. S. P. in Inglefield Land (Dawes 1972, 1979, Peel 1978, Collinson et al. in press) has maintained and strengthened the comparison.

This paper summarises the stratigraphy of the predominantly clastic strata which occur on Bache Peninsula and in Inglefield Land between the Precambrian crystalline basement and carbonates of Early Cambrian age. The early Cambrian carbonates and the overlying Cambro-Ordovician part of the sequence are described by Peel & Christie (this volume). Formal description of the units in question is beyond the scope of the present paper and has been undertaken elsewhere (Collinson et

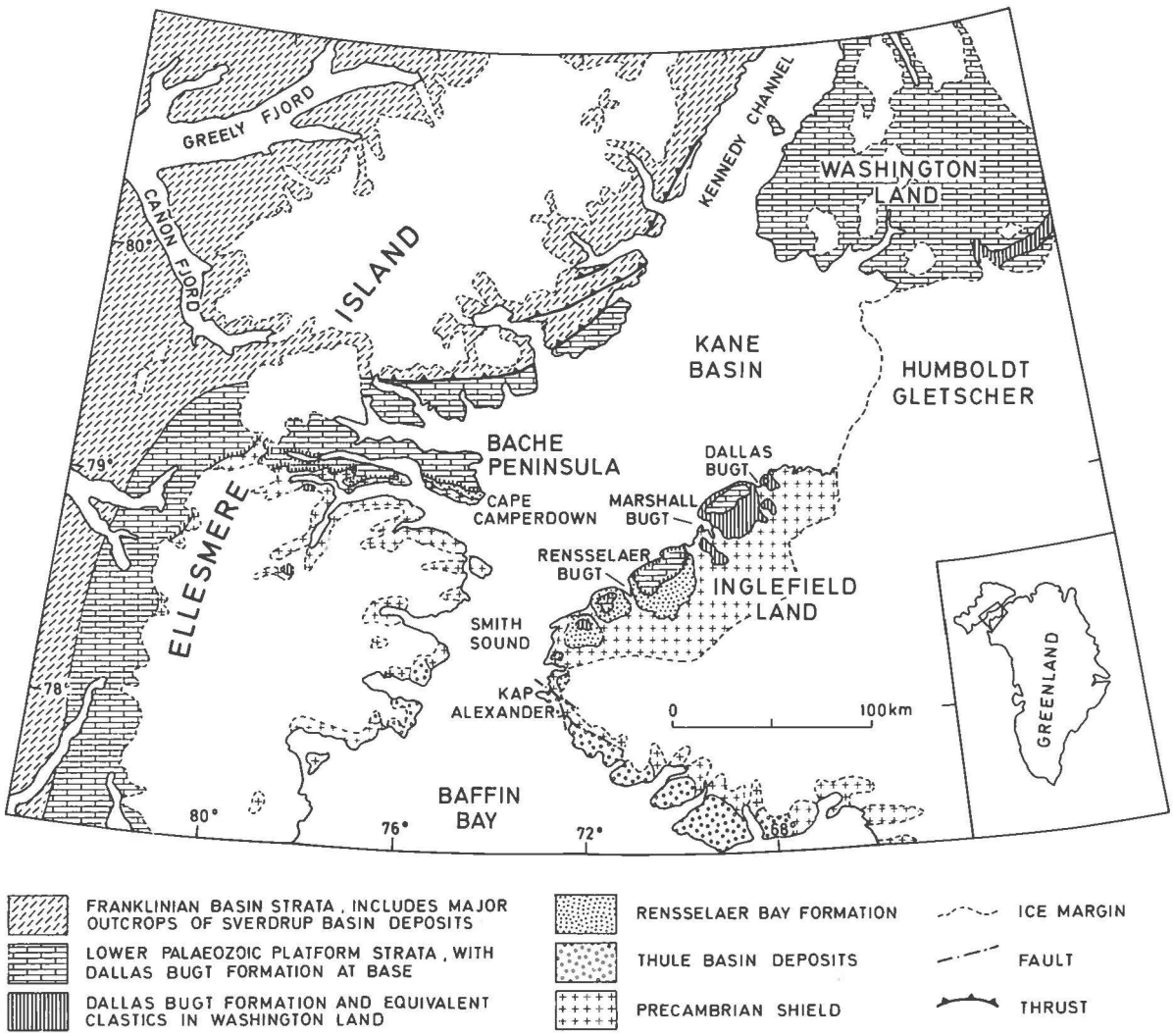


Fig. 1. Geological sketch map of land areas around Kane Basin. The thrusts shown mark the southern margin of the Ellesmere–Greenland fold belt.

al. in press). These clastic strata have previously been considered to form a uniform sequence of either late Precambrian or Early Cambrian age. However, our data indicate that the clastic sequence is composed of two separate formations of widely different ages, which were deposited in two independent basins. The two basins are separated by a regional basement high, the so-called Bache Peninsula arch of Kerr (1967b), although deposits in both basins overlap across the crest of the arch.

This significant revision serves to strengthen the stratigraphic comparison between Bache Peninsula and Inglefield Land. In addition, the line of junction between the two sedimentary basins in the two areas provides a datum which can be used in the present context for evaluation of net displacement along Nares Strait.

On the basis of stratigraphic similarity and the juxtaposition of basin margins across Nares Strait, it is concluded that large-scale net displacement along Nares Strait, i.e. 250 km, suggested on geological and geophysical evidence by authors such as Newman (1977, this volume), Sclater et al. (1977), Newman & Falconer (1978) and Srivastava (1978) has not taken place. These relationships are best explained by the present-day configuration of landmasses.

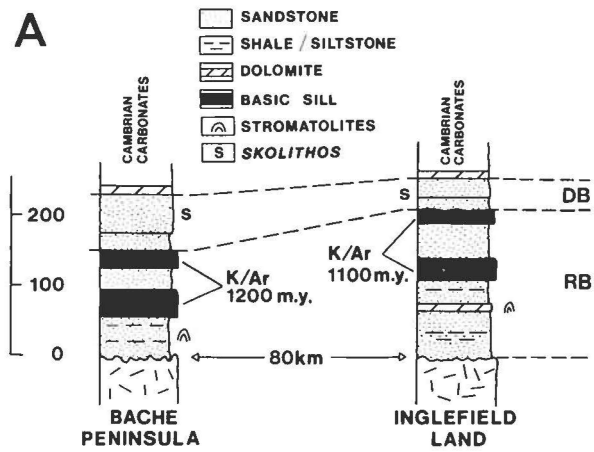
History

The strata described in this paper have previously been assigned to the Rensselaer Bay sandstone or Formation

of Troelsen (1950) and Christie (1967). They overlie the Precambrian Shield with profound unconformity and have been regarded as Proterozoic (Koch 1929, 1933, Bentham 1941, Cowie 1961), Eocambrian (Troelsen 1950, 1956a, b) or Cambrian (Wordie 1938, Kerr 1967a, b, Christie 1967, Cowie 1971) in age. Apparent conformity below proven Lower Cambrian strata, and suggested lateral correlation in Canada with the Lower Cambrian clastics and carbonates of the Franklinian shelf (Kerr 1967a) led to the conclusion that the Rensselaer Bay Formation was solely of Cambrian age, although Christie (1967, 1972) did not exclude the possibility that the lower part of the formation was of Proterozoic age.

Dolerite sills, present in the lower part of the formation in both Canada and Greenland, were isotopically dated on Greenland material as Proterozoic (Dawes et al. 1973). This put in doubt the Palaeozoic age assignment for at least that part of the formation in Greenland and, by inference, also in Bache Peninsula, although Christie's (1967) record of "scolithid-like" structures in the upper part of the Bache Peninsula sequence suggested a basal Cambrian age for part of the formation. Field work in Inglefield Land in 1977, 1978 and 1980 (Peel 1978, Dawes 1979, Collinson et al. in press) demonstrated that a major erosional unconformity exists between a lower sandstone sequence containing the basic sills and an overlying sandstone containing the trace fossils *Rusophycus* and *Skolithos*. Thus, it became clear that a bipartite division of the Rensselaer Bay sandstone/Formation of Troelsen (1950) and Christie (1967) could be achieved in both Greenland and Canada (Fig. 2).

Accordingly, the Rensselaer Bay Formation was restricted by Collinson et al. (in press) to the lower, Proterozoic part of the sequence which contains the isotopically dated intrusions. The upper part was assigned, in the same publication, to a new formation, the Dallas Bugt Formation (Figs 2 and 3). The geographical distribution of the two formations is shown in Fig. 1, from which it can be seen that the Dallas Bugt Formation oversteps the Rensselaer Bay Formation in both Greenland and Canada to lie in stratigraphic contact with the Precambrian Shield.



B

	Bache Peninsula	Inglefield Land	
DALLAS BUGT FORMATION	Sverdrup Member	Marshall Bugt Member	Early Cambrian
		Qáqaitsut Member	
	Bache Peninsula Member	Kap Scott Member	
RENSSELAER BAY FORMATION	Camperdown Member	(not formally subdivided)	Late Proterozoic

Fig. 2. Late Proterozoic and basal Cambrian stratigraphic correlation schemes between Bache Peninsula, Ellesmere Island and Inglefield Land, Greenland. A – generalised composite stratigraphy, RB = Rensselaer Bay Formation, DB = Dallas Bugt Formation; B – current stratigraphic nomenclature.

Stratigraphy

Rensselaer Bay Formation of Inglefield Land

In the type area, around Rensselaer Bugt, the Rensselaer Bay Formation, as redefined by Collinson et al. (in press), is an approximately 220 m thick complex of banded, multicoloured, often dark red and green sandstones and siltstones with subordinate shales which contain beds and lenses of stromatolitic dolomite (Fig. 4). Three principal sedimentary associations are recognised. An association of interbedded mudstone, siltstone and sandstone is characterised by dark red mudstone, and by units of interbedded siltstone and sandstone in beds a few centimetres thick. Dessication cracks and ripple marks are common. A medium-scale cross-bedded sandstone association contains poorly sorted, medium- to coarse-grained sandstones in beds up to 6 m thick, showing cross-bedding in troughs up to 50 cm. Quite well sorted, medium-grained sandstones showing large-scale cross-bedding in sets up to 3 m thick form the third association.

In south-westernmost Inglefield Land, the Rensselaer Bay Formation can be divided into three members, with light coloured, medium- to coarse-grained, cross-bedded sandstones forming the upper and lower members (Fig. 3). In the type area, these latter members are not developed.

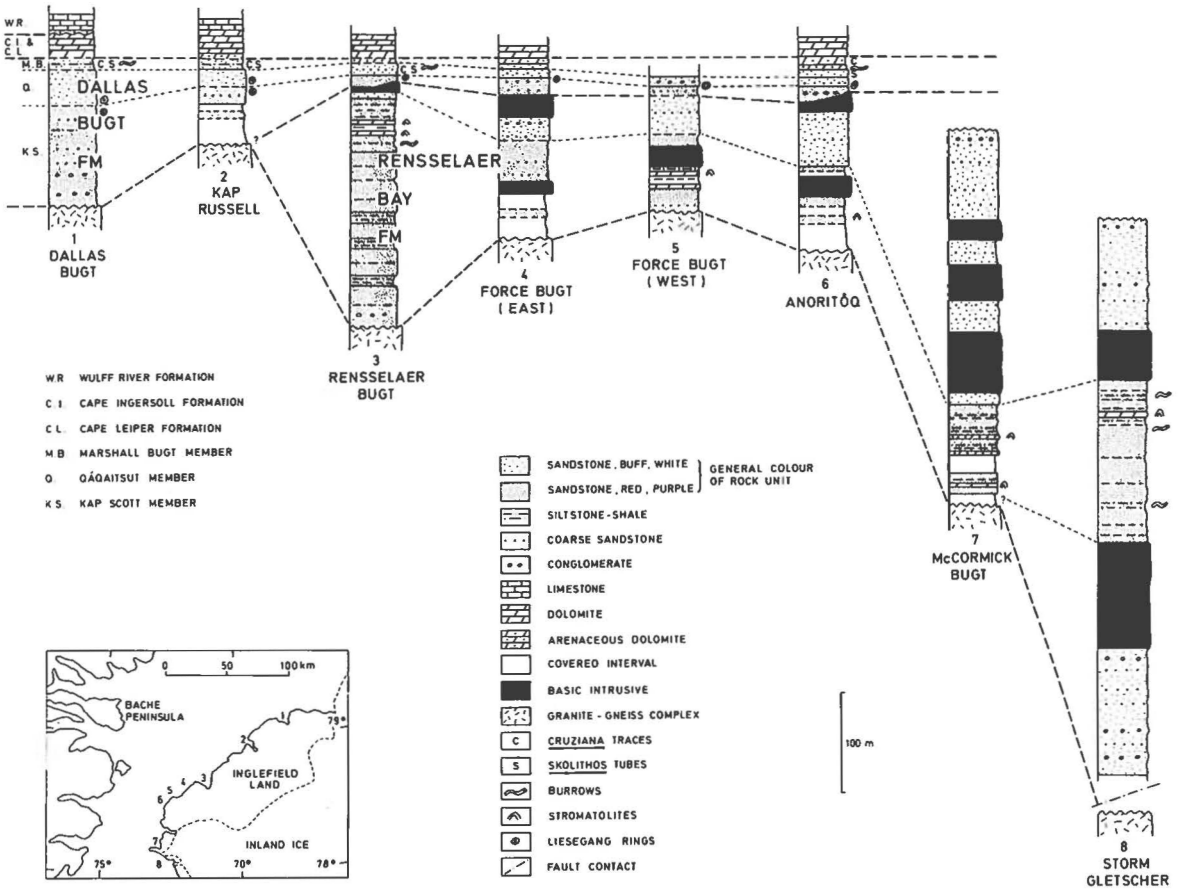


Fig. 3. Stratigraphical relationship between the Rensselaer Bay Formation and the Dallas Bugt Formation in North-West Greenland.

The Rensselaer Bay Formation rests with profound unconformity on crystalline basement (Fig. 5). Around Rensselaer Bugt, in Inglefield Land, the basement shows a local palaeorelief in excess of 150 m within about a kilometre, but elsewhere in Inglefield Land, to the south-west, the unconformity is planar. The upper boundary of the formation with the overlying Dallas Bugt Formation is apparently planar (Fig. 6).

The Rensselaer Bay Formation is intruded by dolerite sills (Figs 5 and 7B), which have yielded K/Ar whole-rock ages in the range 1070 ± 40 m.y. to 1190 ± 40 m.y. (Dawes et al. 1973). Crystalline basement overlain by the formation has yielded dates as young as 1520 m.y. (Larsen & Dawes 1974), suggesting that the Rensselaer Bay Formation is of late Proterozoic (Helikian) age.

Rensselaer Bay Formation of Bache Peninsula

Collinson et al. (in press) retained only the basal Camperdown Member of Christie's (1967) Rensselaer Bay Formation (*sensu lato*) within the redefined Rensselaer Bay Formation (Fig. 2B). The overlying Bache Peninsula and Sverdrup Members were assigned to the Dallas Bugt Formation, as discussed below. The Camperdown Member consists mainly of thin-bedded, red and green shales and white, greenish and brownish sandstones. Thin calcareous beds, often stromatolitic, sporadically occur. The Camperdown Member has a maximum thickness at Cape Camperdown of about 85 m, but it gradually thins out and eventually disappears to the west (Fig. 1). The member is intruded by two dolerite sills (Fig. 7A). The freshest sample available, from the lower sill, has given a K/Ar whole-rock age of 1197 ± 33 m.y. (Dawes et al., this volume). Where the two dolerite sills are present the member is increased to a maximum of about 140 m in thickness (Fig. 2A).

A correlation in terms of members is not attempted at this time between the section recorded by Christie (1967) in the Camperdown Member at Bache Peninsula and the undivided Rensselaer Bay Formation of Inglefield Land. However, the similarity in lithology, particularly the distinctive red and green colouration, stratigraphic position between crystalline basement and

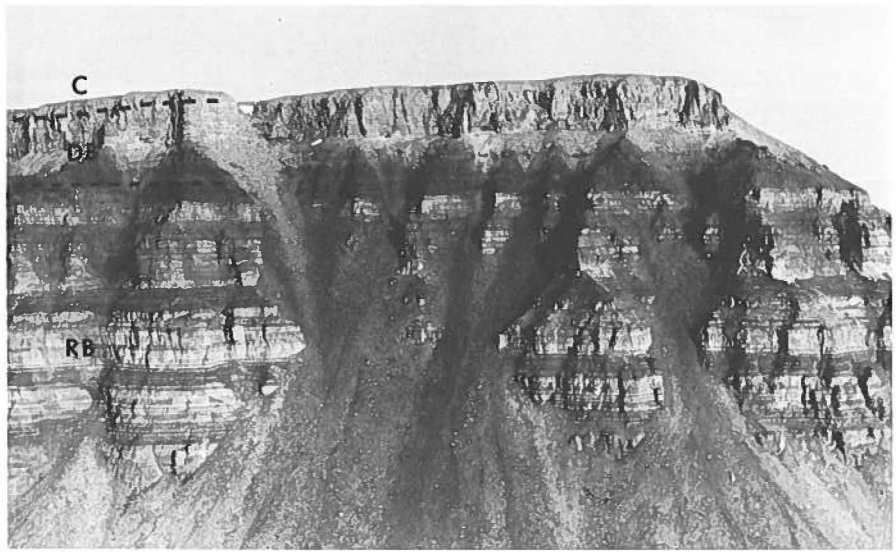


Fig. 4. Cliff section, west of Rensselaer Bugt showing the Proterozoic Rensselaer Bay Formation (RB) overlain by the basal Cambrian Dallas Bugt Formation (DB) and Cambrian carbonates (C). A basic sill forms the uppermost stratum of the Proterozoic sequence (see Fig. 7). Height of section about 200 m.

the Dallas Bugt Formation, and the presence of the dolerite intrusions clearly demonstrate that the Camperdown Member is an integral part of the redefined Rensselaer Bay Formation.

Dallas Bugt Formation of Inglefield Land

This formation has been defined and fully described by Collinson et al. (in press). At the type area around Dallas Bugt (Fig. 1) it attains a maximum thickness of about 150 m. This thickness is gradually reduced to only 25 m at Rensselaer Bugt, but increases slightly again to the south-west (Fig. 3). The formation is subdivided into three members (Figs 2B and 3). A basal Kap Scott Member is mainly composed of cross-bedded, coarse feldspathic sandstones and fine conglomerates which are very heavily stained red-purple by hematite. The member is about 100 m thick at Dallas Bugt, but is reduced to less than 10 m at Force Bugt. The boundary with the overlying Qáqaitsut Member (35 m, thinning to c. 10 m to the south-west) is gradational. This latter member consists predominantly of white, medium to coarse, cross-bedded, quartz-rich sandstones which contain little feldspar. The Marshall Bugt Member (5–10 m) consists of medium- to coarse-grained sandstones and interbedded green siltstones. It is gradational with the dolomites of the overlying Cape Leiper and Cape Ingersoll Formations.

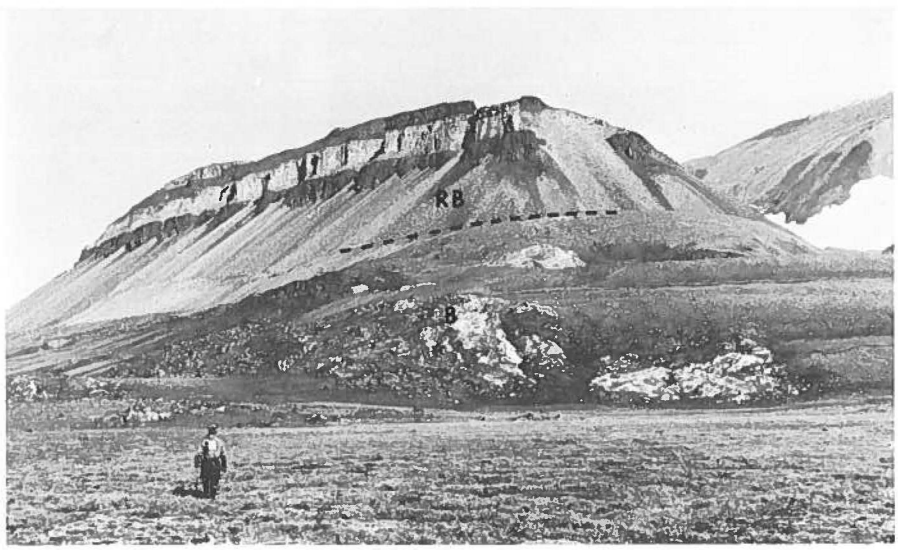


Fig. 5. The Rensselaer Bay Formation (RB), cut by two dolerite sills, overlying the Precambrian crystalline basement (B) at Hatherton Bugt, Greenland. The upper cliff-forming, pale sandstone unit has been referred by Cowie (1971) to the basal Cambrian Sverdrup Member of Bache Peninsula. However, the two basic sills are of Proterozoic age and the use of the name Sverdrup in Greenland is discontinued.

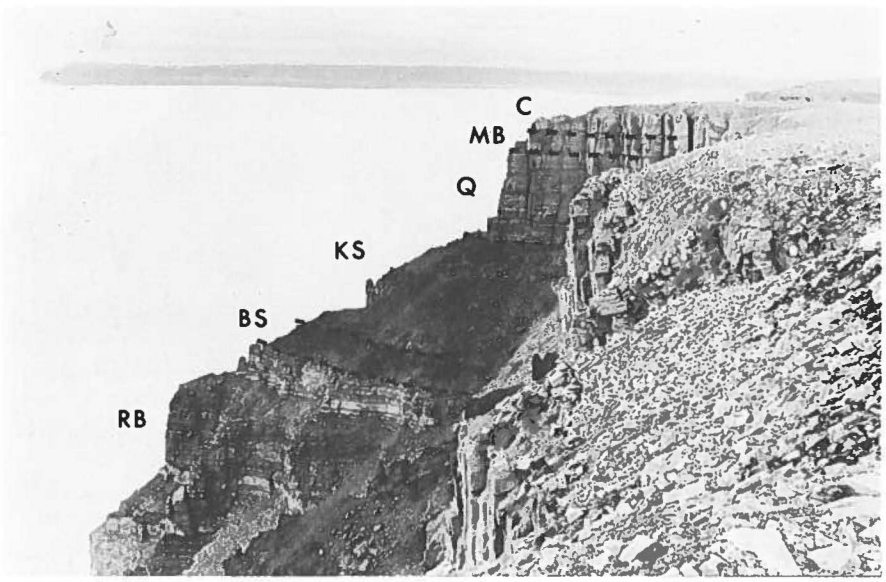


Fig. 6. The disconformity between the Rensselaer Bay Formation (RB) and the Kap Scott Member (KS) of the Dallas Bugt Formation. A basic sill (BS) with an eroded top is present below the unconformity. The boundaries between the Qâqaitsut Member (Q) and the Marshall Bugt Member (MB) of the Dallas Bugt Formation and the overlying Cambrian carbonates (C) are gradational. View is east across Rensselaer Bugt.

The Dallas Bugt Formation lies unconformably on the underlying Rensselaer Bay Formation throughout south-western Inglefield Land (Fig. 6), although the precise contact is often obscured in the field. The unconformity is seen to truncate the dolerite sills within the Rensselaer Bay Formation at Force Bugt. In north-eastern Inglefield Land, the Dallas Bugt Formation lies directly on crystalline basement (Fig. 8).

The Dallas Bugt Formation is considered to be of Early Cambrian age on the basis of its stratigraphic position below carbonates of the Wulff River Formation, which contain Early Cambrian trilobites (C.

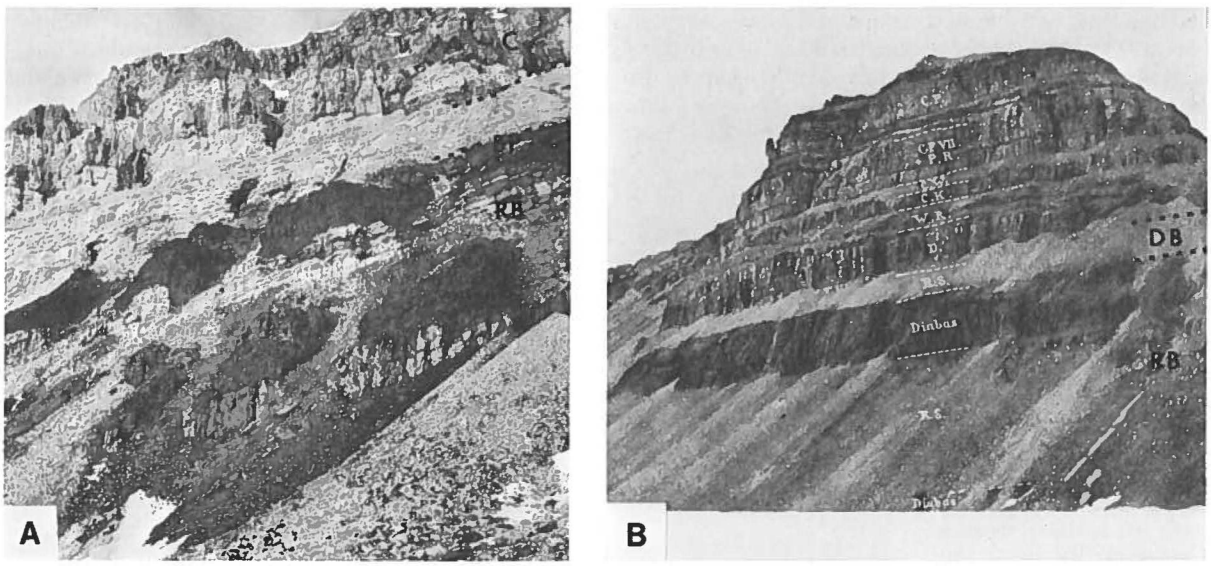


Fig. 7. Comparison of the Proterozoic–Cambrian sequences at Bache Peninsula, Ellesmere Island (A) and in Inglefield Land, Greenland (B). A: Cliff section west of Cape Camperdown, showing the Rensselaer Bay Formation (RB) with two basic sills (dark) overlain by the Dallas Bugt Formation — the Bache Peninsula (BP) and Sverdrup (S) Members — and the Cambrian carbonates (C). From Christie (1967: pl. II). B: The cliffs west of Rensselaer Bugt, showing the Rensselaer Bay Formation (RB) with two basic (Diabas) sills, overlain by the Cambrian succession. From Koch (1933: fig. 13) with original annotations. The Dallas Bugt Formation (DB) includes the upper part of Koch's red sandstones (RS) and the basal part of his dolomite unit (D). The bulk of this unit, and the other strata labelled by Koch, are carbonates of Cambrian age (C). Current designations lie to the right.



Fig. 8. Cambrian clastics of the Dallas Bugt Formation (DB) and succeeding Cambrian carbonates (C), overlying the Precambrian crystalline basement (B) at Marshall Bugt, Inglefield Land. View northwards across Nares Strait to Ellesmere Island. Aerial photograph 544 C–N, No. 11829; copyright Geodætisk Institut, Denmark.

Poulsen 1927, 1958, V. Poulsen 1964) and on the occurrence of *Skolithos* and *Rusophycus* in the Qáqaitsut and Marshall Bugt Members.

Dallas Bugt Formation of Bache Peninsula

Collinson et al. (in press) formally transferred the Bache Peninsula Member and the Sverdrup Member of the Rensselaer Bay Formation (*sensu lato*) of Christie (1967) to the Dallas Bugt Formation, assigning the basal Camperdown Member to the redefined Rensselaer Bay Formation (Fig. 2B). The Dallas Bugt Formation is about 60 m thick at Bache Peninsula. The lower member, the Bache Peninsula Member, consists of 9 m of coarse, purple-brown conglomeratic arkose and thins rapidly to the west. The overlying Sverdrup Member, white or pale yellow weathering, medium- to coarse-grained, cross-bedded sandstone, increases to 95 m in thickness to the west and rests directly on crystalline basement (Fig. 9). In places, the sandstone is characterised by regular, sub-columnar tubes or pipes, probably referable to *Skolithos*.

The Bache Peninsula Member is equivalent to the Kap Scott Member of Inglefield Land (Fig. 2B). The overlying Sverdrup Member is correlated principally with the Qáqaitsut Member, but fine-grained greenish sandstones at the top of the Sverdrup Member (Christie 1967) are probably correlatives of the Marshall Bugt Member.



Fig. 9. Precambrian crystalline basement (B) overlain by recessive, pale sandstones of the Sverdrup Member (S) of the Dallas Bugt Formation and succeeding Cambrian carbonates (C), west of Bache Peninsula, Ellesmere Island. From Christie (1967: pl. V).

Basic intrusions

A well-known feature of the Proterozoic sequences of North-West Greenland is the presence of basic intrusions in the form of dykes and sills (Figs 5 and 7B). Koch (1933) noted that the diabase intrusions, present throughout the Thule district to the south, continued into Inglefield Land and that the igneous province ended around Rensselaer Bugt (Fig. 1).

The age of the intrusions in Inglefield Land and their relationship to the sediments, particularly the dolomites above the clastics now referred to the Rensselaer Bay and Dallas Bugt Formations, have been long debated (Koch 1933: 21, Troelsen 1950: 19, Cowie 1961: 20, Dawes et al. 1973: 63, Christie 1967: 55, Christie 1972: 58, Dawes 1976: 262). Our data indicate that the basic intrusives are restricted to the redefined Rensselaer Bay Formation and that the stratigraphically highest sill has an eroded upper surface when in contact with the overlying Dallas Bugt Formation (Fig. 6). On a large scale, for example in Force Bugt, one sill is cut out at the unconformity. In detail, the upper surface of the sill is irregular, has a rubbly weathering appearance and lacks the chilled margin that characterises the lower contact.

The recognition of the two sandstone sequences reported on in this paper — a lower, redefined Rensselaer Bay Formation of Proterozoic age with intrusions and an upper intrusion-free Dallas Bugt Formation of Early Cambrian age — also explains the restriction of the sills to south-western Inglefield Land, i.e. the outcrop area of the Rensselaer Bay Formation. Basic sills are also restricted to the redefined Rensselaer Bay Formation on Bache Peninsula (Fig. 7A) and again serve to stress the similarity with the Inglefield Land sequence.

Bache Peninsula arch

The name Bache Peninsula arch was given by Kerr (1967b) to a structurally high area of Precambrian crystalline rocks which trends approximately west-north-west between south-western Inglefield Land and the Bache Peninsula area. The arch was defined as a "positive basement feature that influenced the depositional patterns of overlying sediments of Early Cambrian to Late Devonian age" (Kerr 1967b: 485), although it must also have been structurally high in the late Proterozoic, since it forms the northern margin of the Thule Basin (Fig. 10).

The Bache Peninsula arch separates two sedimentary basins: the Thule Basin to the south and the Franklinian Basin to the north and north-west. In essence, the Bache Peninsula arch is a shoulder of the Precambrian Shield projecting out into the Franklinian Basin. While structurally high, it is not a tectonically active feature, although intracratonic faulting accompanied the form-

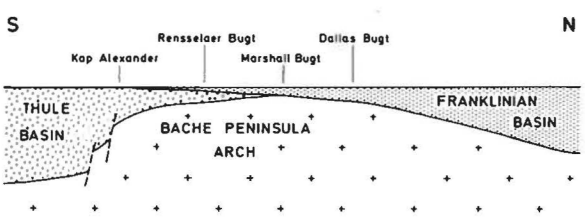


Fig. 10. Schematic section through the Bache Peninsula arch in North-West Greenland, showing the relationship between the intracratonic Thule Basin to the south, and the Franklinian Basin to the north. These relationships are also present in adjacent Ellesmere Island.

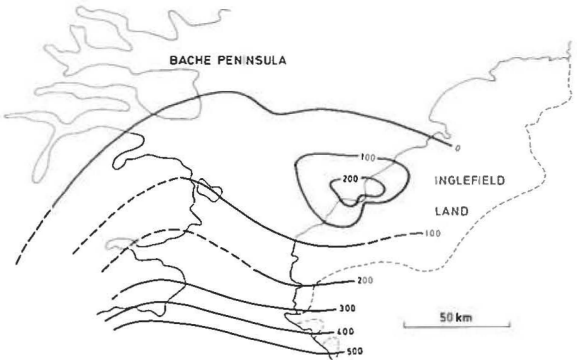


Fig. 11. Isopach map showing the preserved thickness of sediments of the Rensselaer Bay Formation and equivalent Proterozoic deposits at the northern margin of the Thule Basin. Intrusive dolerites are excluded from the thickness estimates.

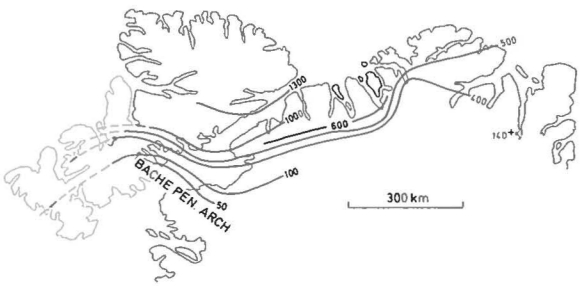


Fig. 12. Isopach map of the Dallas Bugt Formation and equivalent Cambrian clastics in Ellesmere Island and northern Greenland.

ation of the Thule Basin to the south. Thus, in Greenland, the 4.5 km thick Proterozoic sequence of the intracratonic Thule Basin thins northward against the Bache Peninsula arch, although a thin sheet of the Rensselaer Bay Formation persists over the arch (Figs 10 and 11). In Ellesmere Island, there is a paucity of outcrop over the arch and no Proterozoic strata are known between Bache Peninsula (79°N) and the northernmost outcrops of the Thule Basin at about 78°30′N (Dawes et al., this volume). However, a contrast in thickness similar to that in Greenland is evident in Ellesmere Island, with only 85 m of sediment at Bache Peninsula and a sequence at least 1 km thick to the south.

The Dallas Bugt Formation and younger Lower Palaeozoic formations thicken away from the Bache Peninsula arch and the rest of the Precambrian Shield into the Franklinian Basin (Fig. 12). In north-western Greenland, thickening is generally to the north or north-west but, in Ellesmere Island, thickening also takes place to the west as the sedimentary basin swings around the Bache Peninsula arch. The effect of the Bache Peninsula arch decreases westward as it plunges under the Lower Palaeozoic cover sequence that forms central Ellesmere Island (Figs 1 and 12).

Inferences concerning displacement along Nares Strait

All authors who have studied the geology of the land areas around Kane Basin and Smith Sound, from Per Schei (Holtedahl 1917) until the present day, have correlated sections on Bache Peninsula and in Inglefield Land without identifying any geological features which would require a configuration of Greenland relative to Ellesmere Island other than that found at the present day. The geological similarity between the two areas has been stated forcibly by Bentham (1936a, b, 1941) Poulsen (1946), Troelsen (1950, 1956a), Christie (1967), Dawes (1976), Collinson et al. (in press) and Dawes et al. (this volume). These papers contain no reference to geological features which require major strike-slip movement between Ellesmere Island and Greenland.

General correlations of lithology, thickness and age of the late Proterozoic Rensselaer Bay Formation and its contained dolerite intrusions, and the Early Cambrian Dallas Bugt Formation of Inglefield Land and Bache Peninsula indicate conclusively that Greenland and Ellesmere Island developed as part of the same plate at that time. No geological phenomenon exists which demands substantial strike-slip movement along Nares Strait to explain it; neither is there geological evidence to suggest that net displacement of this magnitude has taken place.

In addition to the close similarity in lithology, thickness and age between sequences on either side of Kane Basin, discussed above, four geological features are of particular relevance to any discussion of major displacement along the Strait:

1) Sedimentary sequences on both sides of the Strait show the same disposition in terms of height above sea level and topography to the peneplained surface of the crystalline rocks of the Precambrian Shield. Even though regional structural dips of the sediments above the basement are low (a couple of degrees to the north in the case of the Dallas Bugt Formation in Inglefield Land), any major displacement oblique to regional strike trends along Nares Strait, of the type proposed by Newman (1977), Newman & Falconer (1978), Srivastava (1978) and several papers in this volume, would inevitably produce substantial anomalies in these sediment-basement topographic relationships across the displacement. Discrepancies of this type are not present.
2) A demonstration of the regional continuity of Early Cambrian clastic sedimentation across northern Greenland and Ellesmere Island is provided by an isopach map of the Dallas Bugt Formation and its stratigraphic equivalents (Fig. 12). Of particular note is the effect of the protruding shoulder of Precambrian Shield, the Bache Peninsula arch, on the

configuration of the southern margin of the Franklinian Basin.

3) The boundary between the thick Proterozoic sequence of the Thule Basin and the thin Rensselaer Bay Formation in south-western Inglefield Land delineates a relatively well-defined datum suitable for correlation across Nares Strait (Dawes et al., this volume). While paucity of exposure prevents *precise* location of this datum in Ellesmere Island, it is clear from regional isopach patterns that displacement of Ellesmere Island relative to Greenland in terms of, say, a hundred kilometres is extremely doubtful (Fig. 11).
4) Overlap between the Thule Basin sequence, in the form of the Rensselaer Bay Formation, and the Dallas Bugt Formation of the Franklinian Basin provides a possible marker across Nares Strait at the point at which the Rensselaer Bay Formation disappears. Sections in Inglefield Land are oriented south-west to north-east, while on Bache Peninsula an east-west orientation is present. However, an isopach map drawn on the residual thickness of the Rensselaer Bay Formation, exclusive of dolerite intrusions, emphasises the continuity of this marker (Fig. 11). The anomalous thicknesses around Rensselaer Bugt in Inglefield Land reflect sediment fill in local depressions or valleys in the crystalline basement. In general, the upper, eroded surface of the basement lacks substantial palaeorelief.

Conclusions

Lithological and stratigraphical similarities between the Rensselaer Bay Formation and the Dallas Bugt Formation in Inglefield Land and on Bache Peninsula, together with evaluation of the four geological features discussed in the previous section, are entirely consistent with there having been little or no net displacement along Nares Strait since early Palaeozoic time. It is not possible to refute movements in the range of 0–50 km, in either dextral or sinistral sense. However, movements greater than this produce increasingly unlikely geological relationships. Offsets in the range 50–100 km are considered improbable, while the postulated sinistral movements of 100 km, 200 km, 300 km or even more can be ruled out.

Acknowledgements

The paper is published with the permission of the Director of the Geological Survey of Greenland. The aerial photograph of Marshall Bugt is published with the permission (A. 495/79) of the Geodætisk Institut, Copenhagen, Denmark.

References

Bentham, R. 1936a. Appendix 1: Geology. – In: Humphreys, N., Shackleton, E. & Moore, A. W., Oxford University Ellesmere Land Expedition. – Geogr. J. 87: 427–431.

Bentham, R. 1936b. Appendix IV, Geology. – In: Shackleton, E., Arctic journeys. The story of the Oxford University Ellesmere Land Expedition 1934–5: 333–337. – Hodder & Stoughton Ltd., London.

Bentham, R. 1941. Structure and glaciers of southern Ellesmere Island. – Geogr. J. 97: 36–45.

Christie, R. L. 1967. Bache Peninsula, Ellesmere Island, Arctic Archipelago. – Mem. geol. Surv. Can. 347: 63 pp.

Christie, R. L. 1972. Central stable region. – In: Christie, R. L., Cook, D. G., Nassichuk, W. W., Trettin, H. P. & Yorath, C. J. (eds), The Canadian Arctic Islands and the Mackenzie region. – 24 Intern. geol. Congr. Montreal, Excurs. guide A66: 40–87.

Collinson, J. D., Dawes, P. R., Peel, J. S. & Christie, R. L. in press. Late Proterozoic and basal Cambrian stratigraphy of Inglefield Land, North-West Greenland: redefinition and correlation with Bache Peninsula, Ellesmere Island. – Rapp. Grønlands geol. Unders.

Cowie, J. W. 1961. Contributions to the geology of North Greenland. – Meddr Grønland 164(3): 47 pp.

Cowie, J. W. 1971. The Cambrian of the North American Arctic regions. – In: Holland, C. H. (ed.), Cambrian of the New World: 325–383. – Wiley-Interscience, New York.

Dawes, P. R. 1972. Precambrian crystalline rocks and younger sediments of the Thule district, North Greenland. – Rapp. Grønlands geol. Unders. 45: 10–15.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R. 1979. Field investigations in the Precambrian terrain of the Thule district, North-West Greenland. – Rapp. Grønlands geol. Unders 95: 14–22.

Dawes, P. R., Rex, D. C. & Jepsen, H. F. 1973. K/Ar whole rock ages of dolerites from the Thule district, western North Greenland. – Rapp. Grønlands geol. Unders. 55: 61–66.

Dawes, P. R., Frisch, T. & Christie, R. L. 1982. The Proterozoic Thule Basin of Greenland and Ellesmere Island: importance to the Nares Strait debate. – This volume.

Holtedahl, O. 1917. Summary of geological results. – Rep. 2nd Norwegian Arctic Exped. "Fram" 1898–1902, 36: 27 pp. – Videnskabs-Selskabet, Kristiania.

Kerr, J. W. 1967a. Stratigraphy of central and eastern Ellesmere Island, Arctic Canada. Part 1. Proterozoic and Cambrian. – Pap. geol. Surv. Can. 67–27(1): 63 pp.

Kerr, J. W. 1967b. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Koch, L. 1929. Stratigraphy of Greenland. – Meddr Grønland 73,2(2): 205–320.

Koch, L. 1933. The geology of Inglefield Land. – Meddr Grønland 73,1(2): 38 pp.

Larsen, O. & Dawes, P. R. 1974. K/Ar and Rb/Sr age determinations on Precambrian crystalline rocks in the Inglefield Land–Inglefield Bredning region, Thule district, western North Greenland. – Rapp. Grønlands geol. Unders. 66: 4–8.

Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Newman, P. H. 1982. A geological case for movement be-

tween Canada and Greenland along Nares Strait. – This volume.

Newman, P. H. & Falconer, R. K. H. 1978. Evidence for movement between Greenland and Canada along the Nares Strait. – Geol. Soc. Am. Abs. with Prog. 10: 463 only.

Peel, J. S. 1978. Geological investigations in Lower Palaeozoic terrain of northern Greenland between 78°30'N and 81°30'N. – Rapp. Grønlands geol. Unders. 90: 14–16.

Peel, J. S. & Christie, R. L. 1982. Cambrian–Ordovician platform stratigraphy: correlations around Kane Basin. – This volume.

Poulsen, C. 1927. The Cambrian, Ozarkian and Canadian faunas of Northwest Greenland. – Meddr Grønland 70,1(2): 233–343.

Poulsen, C. 1940. Foreløbig meddelelse om resultater af Mag. scient. J. Troelsens geologiske undersøgelser i Inglefield Land, Grinnell Land og Ellesmere Land. – Meddr dansk geol. Foren. 9: 638–641.

Poulsen, C. 1946. Notes on Cambro-Ordovician fossils collected by the Oxford University Ellesmere Land Expedition 1934–5. – Q. Jl geol. Soc. Lond. 102: 299–335.

Poulsen, C. 1958. Contribution to the palaeontology of the Lower Cambrian Wulff River Formation. – Meddr Grønland 162(2): 25 pp.

Poulsen, V. 1964. Contribution to the Lower and Middle Cambrian paleontology and stratigraphy of Northwest Greenland. – Meddr Grønland 164(6): 105 pp.

Sclater, J. G., Hellinger, S. & Tapscott, C. 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. – J. Geol. 85: 509–552.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Troelsen, J. C. 1950. Contributions to the geology of Northwest Greenland, Ellesmere Island and Axel Heiberg Island. – Meddr Grønland 149(7): 86 pp.

Troelsen, J. C. 1956a. The Cambrian of North Greenland and Ellesmere Island. – In: El sistema Cámbrico, su paleogeografía y el problema de su base. – 20 Intern. geol. Congr. México, Symp. 3(1): 71–90.

Troelsen, J. C. 1956b. Groenland. – In: Lexique Stratigraphique International 1(1a): 116 pp. – Centre Nat. de la Recherche Scientifique, Paris.

Wordie, J. M. 1938. An expedition to Northwest Greenland and the Canadian Arctic in 1937. – Geogr. J. 92: 385–421.

Cambrian–Ordovician platform stratigraphy: correlations around Kane Basin

JOHN S. PEEL and ROBERT L. CHRISTIE

Peel, J. S. & Christie, R. L. 1982. Cambrian–Ordovician platform stratigraphy: correlations around Kane Basin. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 117–135.

Cambrian and Ordovician platform strata in western North Greenland and adjacent Ellesmere Island were deposited on an elongate platform lying to the south of the Franklinian Trough. The sequence is dominated by carbonates with prominent evaporite horizons and subsidiary shales and sandstones. A transect across the platform shows changes in facies and increase in thickness from inner platform sequences in Inglefield Land and Bache Peninsula, to the south, to outer platform sequences in the Judge Daly Promontory area; clastic trough deposition characterises sections in northern Ellesmere Island. This paper briefly describes and correlates these sequences occurring around Kane Basin. The regional framework and close similarity of Cambrian and Ordovician sequences in Greenland and Canada indicate that substantial transcurrent net displacement has not taken place along Nares Strait. Of special note are narrow belts of Ordovician evaporites which cross Nares Strait from Ellesmere Island into Greenland without displacement at the present day. The continuity of these environmentally specialised belts of deposition demonstrates that net transcurrent displacement along Nares Strait since the Early Ordovician has not exceeded 50 km.

J. S. Peel, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK–1350 Copenhagen K, Denmark and R. L. Christie, Department of Energy, Mines and Resources, Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303–33rd St. N.W., Calgary, Alberta, Canada, T2L 2A7.

The Cambrian and Ordovician strata discussed here were deposited on a broad marine platform, the Arctic Platform, located between the Precambrian Shield to the south-east, and the Franklinian Trough of Ellesmere Island and northern Greenland to the north and north-west (Fig. 1). This platform regime corresponds to the Central Stable Region (the inner platform) and the Franklinian miogeosyncline (the outer platform) of Kerr (1967a: fig. 2). In general, the Cambrian and Ordovician sequence of the platform is rich in carbonates and thickens away from the Precambrian Shield towards the clastic trough sequence to the north. Basal Cambrian sandstones and conglomerates and thick Ordovician evaporites are present with the carbonates, and there is considerable variation in thickness and facies.

A transect through the Cambrian and Ordovician of the platform is best demonstrated in Ellesmere Island (Figs 2 and 3) where inner platform sequences on Bache Peninsula (Christie 1967) pass northward into outer platform sequences on Judge Daly Promontory (Kerr 1967a, b, c, 1968) and, finally, into Franklinian Trough sequences in the Hazen Plateau region (Trettin 1971, Trettin & Balkwill 1979, Trettin et al. 1979). Inner platform deposition in adjacent Greenland is well represented by sequences in Inglefield Land and southern Washington Land, but deposits of the outer platform are covered by overlying Silurian strata (see Hurst & Kerr, this volume). At the northern extremity of Nares Strait, Cambrian and Ordovician strata outcrop in the northern parts of Hall Land, Nyeboe Land and Wulff Land (Dawes, this volume, Hurst & Kerr, this volume) and in the latter two areas these rocks have been interpreted as indicative of the transition from the outer platform to the trough.

Cambrian and Ordovician platform stratigraphy in northern Greenland

The oldest Cambrian sediments occurring in platform sequences in Greenland adjacent to Kane Basin are sandstones with subsidiary conglomerates and siltstones of the Dallas Bugt Formation in Inglefield Land, and the Humboldt Formation of Washington Land (Figs 3–5). Older Cambrian strata may occur in the Nyeboe Land – Wulff Land area within the Nyeboe Land fault zone (Dawes, this volume). To the east in Wulff Land and Peary Land (Fig. 1) equivalent sandstones are given the name Buen Formation (Fig. 5) and are underlain by Lower Cambrian dolomites of the Portfjeld Formation

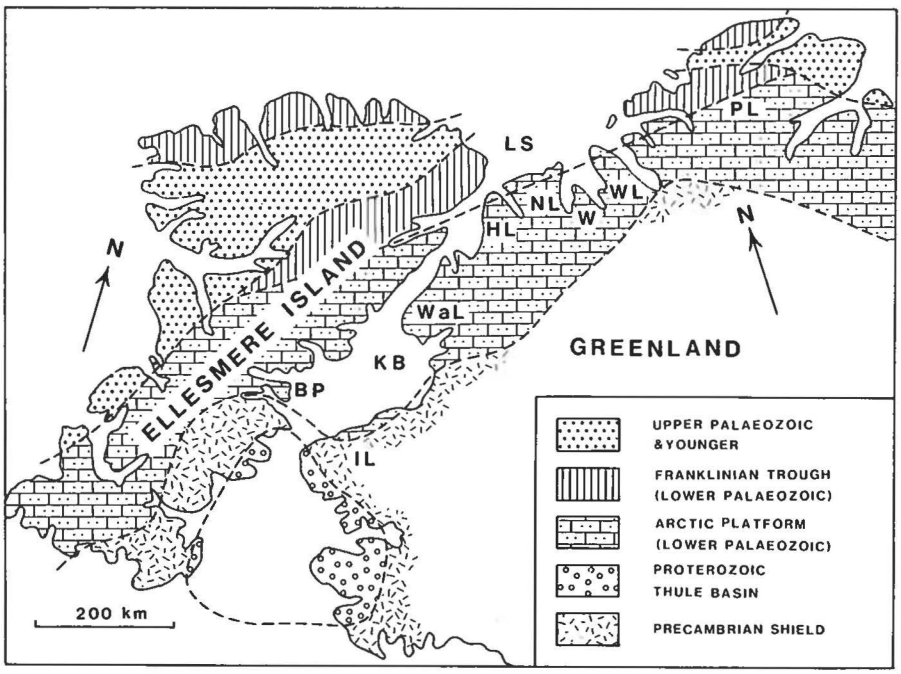


Fig. 1. Main structural units of North-West and North Greenland and Ellesmere Island. BP, Bache Peninsula; IL, Inglefield Land; KB, Kane Basin; WaL, Washington Land (including Daugaard-Jensen Land); HL, Hall Land; NL, Nyeboe Land; W, Warming Land; WL, Wulff Land; PL, Peary Land; LS, Lincoln Sea.

(Jepsen 1971, Hurst & Peel 1979, Ineson & Peel 1980). The absence of the Portfjeld Formation to the west of Wulff Land is attributed to overstep by the overlying clastics as the Cambrian sea transgressed on to the Precambrian Shield.

The Dallas Bugt Formation (maximum thickness 145 m) overlies late Proterozoic clastics and Precambrian crystalline rocks in Inglefield Land, Greenland (Fig. 6), and on Bache Peninsula, Ellesmere Island. It is more fully discussed elsewhere in this volume (Peel et al.). The equivalent Humboldt Formation of Washington Land (Fig. 4) is slightly thicker, but is also reported to overlie crystalline basement immediately north of the Humboldt Gletscher (B. R. Pelletier in Palmer & Peel 1981: 7). Both these formations contain *Skolithos* and *Cruziana/Rusophycus,* although the uppermost beds of the Humboldt Formation have also yielded an olenellid trilobite indicative of an Early Cambrian age.

Well-bedded dolomites of the Kastrup Elv Formation, 140 m thick and of Early–Middle Cambrian age, overlie the Humboldt Formation in Washington Land (Figs 4 and 5) (Henriksen & Peel 1976). The overlying Telt Bugt Formation (Middle Cambrian) consists of 5–100 m of thin-bedded lime mudstones with silty mottling and laminations. To the south-west (Figs 3 and 4), these two formations are correlated with the Cape Leiper, Cape Ingersoll, Wulff River, Cape Kent and Cape Wood Formations of Inglefield Land (C. Poulsen 1927, Koch 1929b, 1933, Troelsen 1950, V. Poulsen 1964). The Cape Kent Formation forms a distinctive marker unit of oolitic limestone in both Inglefield Land and Bache Peninsula (Fig. 6). Limestones of the underlying Wulff River Formation (Early Cambrian) and the overlying Cape Wood Formation (Middle Cambrian) are lithologically indistinguishable from those of the Telt Bugt Formation, demonstrating the diachronous boundary between a lower dolomite facies and an upper limestone facies in the Lower and Middle Cambrian around Kane Basin (Fig. 4).

The Cass Fjord Formation (Poulsen 1927) has recently been redescribed in the type area of southern Washington Land (Figs 1, 3 and 7) where it attains a thickness of up to 470 m (Henriksen & Peel 1976, Peel & Cowie 1979, Palmer & Peel 1981). Grey-green, lenticular lime mudstones with thin beds of intraformational conglomerate are characteristic, but a variety of thin-bedded carbonates are present, together with evaporites and thin quartzites. The Cass Fjord Formation, which also occurs in Inglefield Land (Troelsen 1950) and on Bache Peninsula (Christie 1967), has previously been considered to be entirely of Ordovician age. However, Late Cambrian (Dresbachian, Franconian and Trempealeauan) trilobites occur through most of the thickness in Greenland (Henriksen & Peel 1976, Palmer & Peel 1981); a few tens of metres at the base and top of the formation are of Middle Cambrian and Early Ordovician age.

The Cambrian sequence in Peary Land to the east (Figs 1 and 5) bears little similarity to the Washington Land and Inglefield Land sequence described above. The Buen Formation can be correlated with the Humboldt and Dallas Bugt Formations, but rocks equivalent to the Portfjeld Formation of Peary Land have not been recognised adjacent to Nares Strait. Overlying Lower,

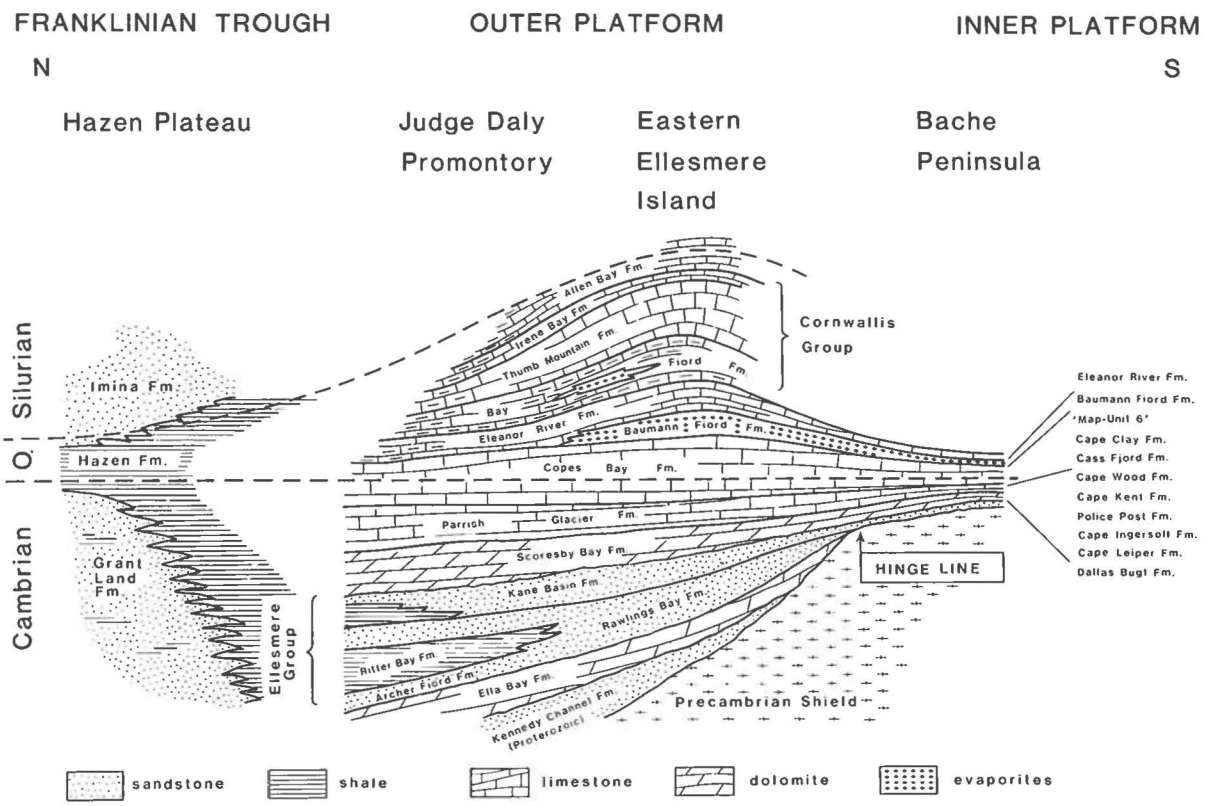


Fig. 2. Schematic geological section through the Cambrian and Ordovician of Ellesmere Island showing the transition from the Arctic Platform into the Franklinian Trough. Thin carbonate sequences of the inner platform (Bache Peninsula) pass through thick carbonate-dominated sequences of the outer platform (eastern Ellesmere Island and Judge Daly Promontory) into clastic sequences of the trough (Hazen Plateau).

Middle and Upper Cambrian carbonates and subsidiary clastics in Peary Land are referred to the Brønlund Fjord and Tavsens Iskappe Groups, which have a combined thickness of almost 1 km. Early Middle Cambrian faunas (*Glossopleura, Clavaspidella*), present in western North Greenland, are absent in Peary Land, apparently as a result of the widespread regional unconformity of the Hawke Bay Event (Palmer & James 1980). Only Dresbachian strata are as yet demonstrated within the Upper Cambrian sequence of Peary Land, although higher mainly unfossiliferous strata within the Tavsens Iskappe Group may extend into the Franconian or Trempealeauan. The Brønlund Fjord and Tavsens Iskappe Groups essentially form a south–north prograding sequence on the southern margin of the Greenland segment of the Franklinian Trough (Ineson 1980, Surlyk et al. 1980). The two groups are subdivided into 13 formations, reflecting pronounced lateral variation and diachronism (Ineson & Peel 1980).

The Ordovician sequence in Washington Land, more than 1 km thick, is divided into 12 formations, in sharp contrast with that in Peary Land where the platform Ordovician is referred to only three formations with a total thickness of about 850 m (Fig. 5). The uppermost Early Ordovician beds of the Cass Fjord Formation are succeeded by pale, massive weathering limestone of the Cape Clay Formation (50 m) (Fig. 7) yielding a rich fauna of Early Ordovician age. The Christian Elv Formation (Fig. 8) is a banded unit some 140 m in thickness, consisting of grey limestones, intraformational conglomerates, dolomites and cross-bedded sandstones (Henriksen & Peel 1976). Early Ordovician fossils are common in the middle part.

The Poulsen Cliff Formation (Figs 8 and 9), a recessive unit of shales, dolomites and thin sandstones with conspicuous beds of evaporites, is about 100 m thick in central southern Washington Land, but appears to increase in thickness by about one third in more western outcrops. The Nygaard Bay Formation (Fig. 9) is about 40 m thick on the south coast of Washington Land and can be sub-divided into a lower resistant unit (20 m) of thin intraformational limestone conglomerates and limestones, and an upper evaporitic unit (20 m). The Canyon Elv Formation (Fig. 9), the former Cape Weber Formation of Troelsen (1950, see Peel & Cowie 1979), is a 50 m thick unit of purple-tinged, grey limestones

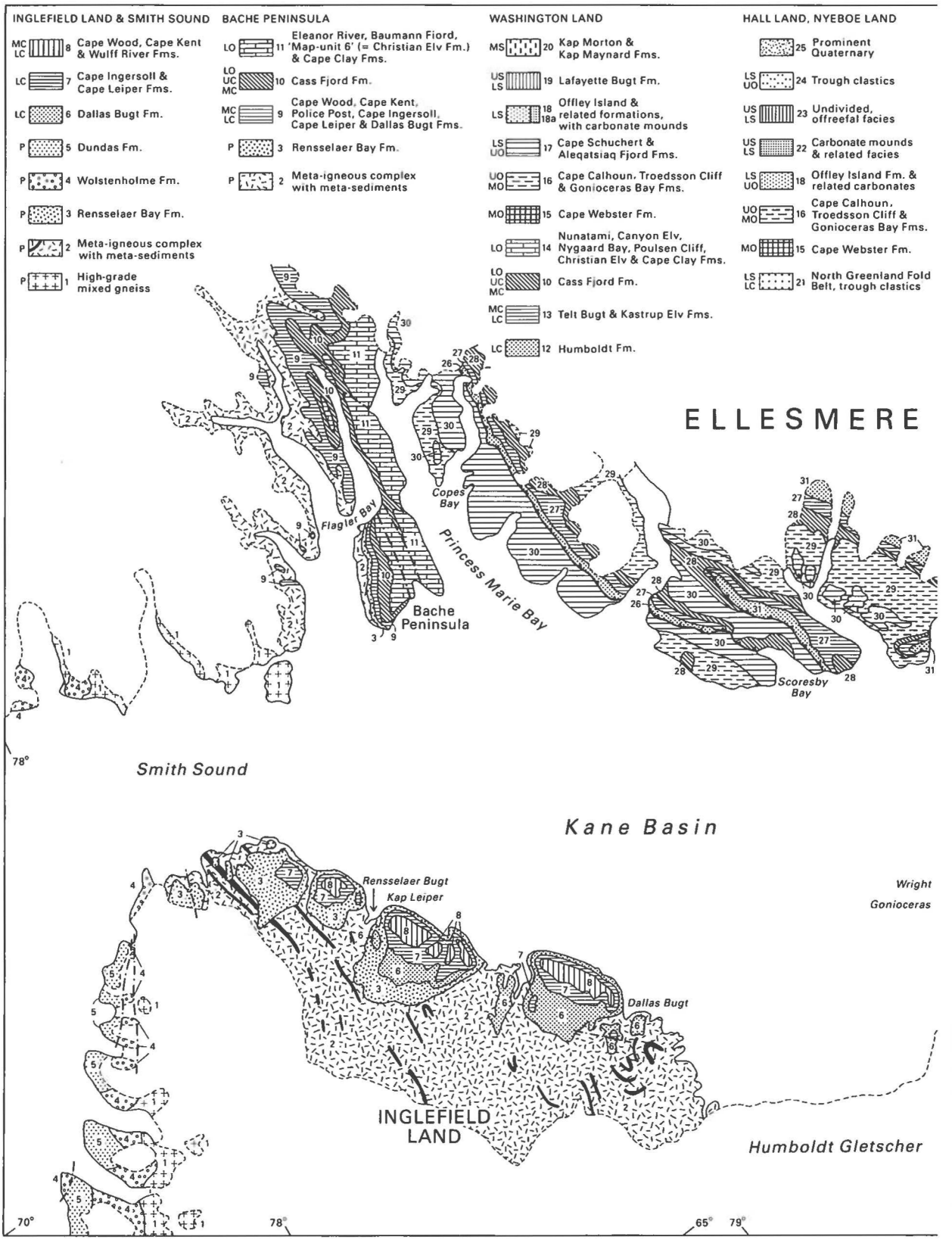
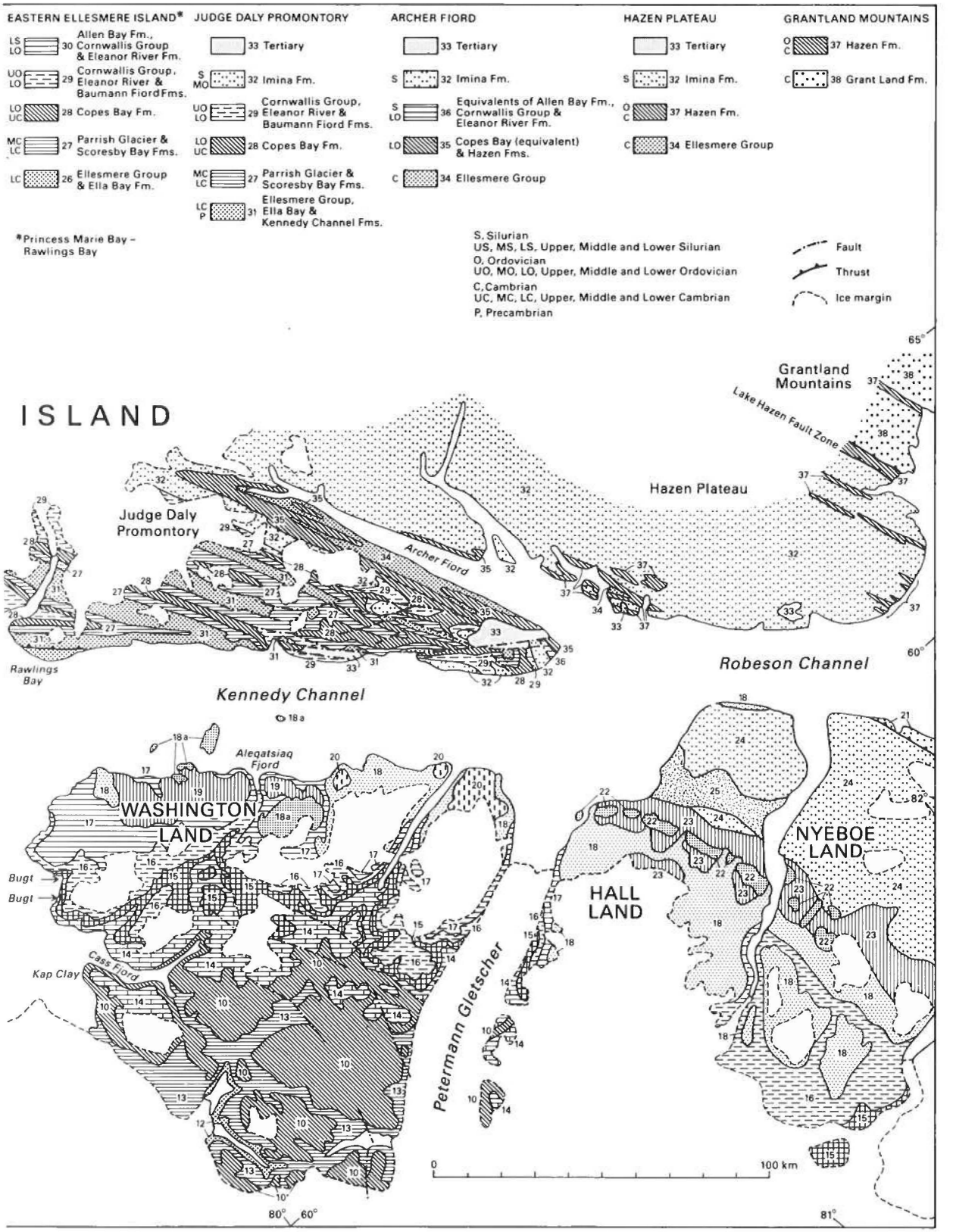


Fig. 3. Geological map of the Nares Strait region drawn from published and unpublished GGU and GSC sources. Main unpub-Sea, map at 1:250 000 by H. P. Trettin, 1981.



lished sources: Greenland, map at 1:1 000 000 by P. R. Dawes, 1980; Canada, north-west Judge Daly Promontory to Lincoln

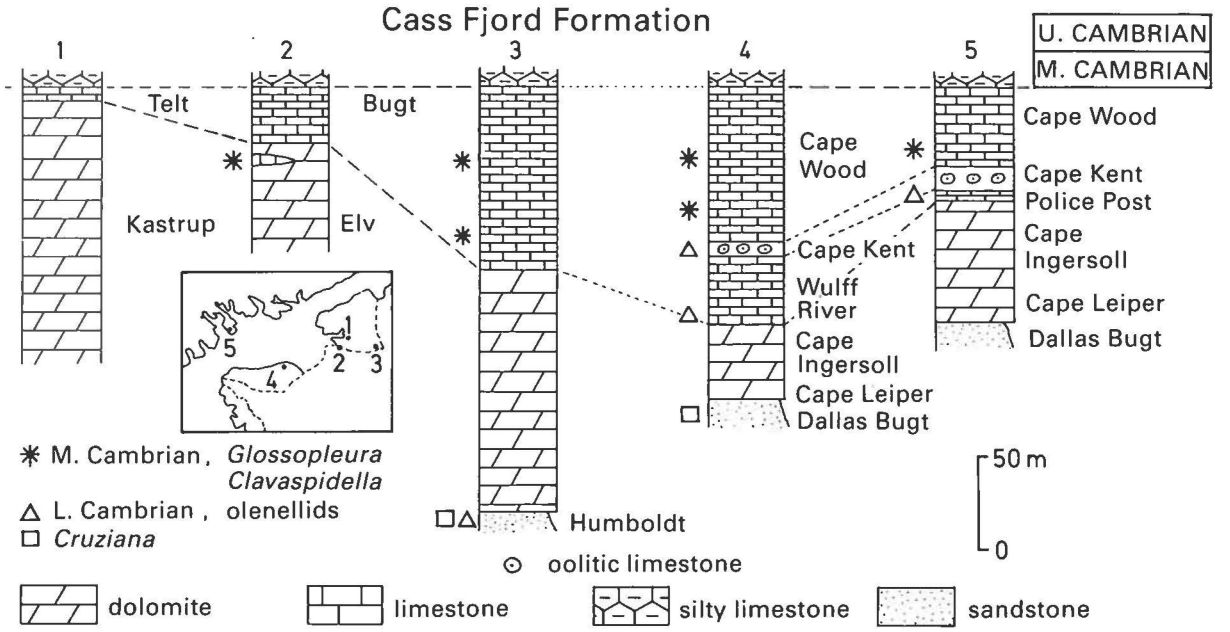


Fig. 4. Lower and Middle Cambrian stratigraphy around Kane Basin showing the diachronous relationship between dolomite and overlying limestone facies. Localities 1–3, Washington Land; 4, Inglefield Land; 5, Bache Peninsula.

which outcrop in the foot of the prominent cliff formed by the overlying Nunatami Formation (160 m). The latter formation is a richly fossiliferous (Poulsen 1927), well-bedded unit of thin- to medium-bedded, grey weathering limestones and intraformational limestone conglomerates of late Early Ordovician age.

The Cape Webster Formation (c. 280 m) is a poorly exposed and generally recessive unit of pale, thin-bedded evaporites, shales, dolomite and greenish siltstones of early Middle Ordovician age (Fig. 10), overlain by cliff-forming, dark weathering limestones of the Morris Bugt Group (Peel & Hurst 1980). The latter (Fig. 11A, B) contains the revised Gonioceras Bay, Troedsson Cliff and Cape Calhoun Formations, and the Aleqatsiaq Fjord Formation (Hurst 1980, see also Hurst & Kerr, this volume). The Gonioceras Bay Formation (c. 75 m) and the Troedsson Cliff Formation (c. 250 m) are brownish weathering, brown or grey lime mudstones. The former tends to be massive weathering, while the latter typically consists of alternating 1 m thick beds of nodular, rubbly weathering limestones and more massive, prominent, nodular limestones. The overlying Cape Calhoun Formation (c. 65 m) is a distinctive greenish weathering unit of pale grey lime mudstone with green shaly partings (Fig. 11B). These three formations have yielded rich Middle and Upper Ordovician faunas (Troedsson 1926, 1928). The Aleqatsiaq Fjord Formation (250 to 360 m thick) ranges in age from Late Ordovician to earliest Silurian (Hurst 1980). The formation is cliff-forming and composed mainly of grey-brown weathering lime mudstones and wackestones.

The Ordovician platform sequence of Peary Land lacks strata of early and middle Early Ordovician age (Fig. 5). The Wandel Valley Formation rests unconformably on the Late Cambrian Tavsens Iskappe Group or older units, and its lower beds are correlated on the basis of the occurrence of *Ceratopea unguis* with the Nunatami Formation of Washington Land of late Early Ordovician age (Peel & Yochelson 1979). The Cape Webster Formation of Washington Land can be traced eastwards to Warming Land (Peel 1980) where it is equivalent to the upper two thirds of the Wandel Valley Formation (350 m), although the latter lacks evaporites. The transition from recessive, dolomitic strata below to cliff-forming limestones, witnessed by the transition from the Cape Webster Formation to the Morris Bugt Group in Washington Land, can similarly be traced to Peary Land where the Wandel Valley Formation is succeeded by the Børglum River Formation (450 m). The latter cannot be subdivided in a similar manner to the Morris Bugt Group, but it has a generally similar lime mudstone lithology. The Ordovician–Silurian boundary in Peary Land lies within a dolomite unit, the un-named Silurian(?) dolomite formation of Christie & Peel (1977) and Armstrong & Lane (1981).

Cambrian and Ordovician platform stratigraphy of Bache Peninsula

The stratigraphic nomenclature of the Cambrian and Ordovician sequence on Bache Peninsula employs a number of names originally defined in adjacent Greenland, reflecting field work undertaken in both areas by Troelsen (1950). Subsequent description by Christie (1967) forms the basis of the present discussion.

Cambrian sandstones and conglomerates, the Sverdrup and Bache Peninsula Members of Christie (1967) now referred to the Dallas Bugt Formation (Collinson et al. in press, Peel et al., this volume), overlie late Proterozoic sandstones with basic intrusions of the Rensselaer Bay Formation and, ultimately, the crystalline basement. Dolomites of the Cape Leiper and Cape Ingersoll Formations (Figs 2, 4 and 12), formally named by Troelsen (1950) in Inglefield Land, also occur on Bache Peninsula. The overlying Police Post Formation, a grey-green arenaceous limestone (4–5 m), was not recognised in Greenland by Troelsen (1950: 38) but can be readily correlated with the Wulff River Formation on account of similar late Early Cambrian age and stratigraphic position between the Cape Ingersoll Formation and the overlying Cape Kent Formation. The distinctive marker unit of dolomite, sometimes oolitic, formed by the Cape Kent Formation in Inglefield Land (Fig. 6), was also noted by Troelsen (1950) on Bache Peninsula (Fig. 12) where Christie (1967) noted a thickness of 10–15 m.

		EASTERN ELLESMERE ISL.	WASHINGTON LAND	PEARY LAND
Sil.		Allen Bay	Aleqatsiaq Fjord	dolomite
ORDOVICIAN	U.	Irene Bay	Cape Calhoun	Børglum River
	M.	Thumb Mountain	Troedsson Cliff / Gonioceras Bay	
		Bay Fiord	Cape Webster	Wandel Valley
		Eleanor River	Nunatami / Canyon Elv	
	L.	Baumann Fiord	Nygaard Bay / Poulsen Cliff	
		Copes Bay	Christian Elv / Cape Clay	
CAMBRIAN	U.		Cass Fjord	Tavsens Iskappe Group
	M.	Parrish Glacier	Telt Bugt	Brønlund Fjord Group
		Scoresby Bay	Kastrup Elv	
	L.	Ellesmere Group	Humboldt	Buen
		Ella Bay		Portfjeld

Fig. 5. Correlation of Cambrian–Ordovician platform sequences in eastern Ellesmere Island, Washington Land (western North Greenland) and Peary Land (eastern North Greenland).

The Cape Wood Formation (Middle Cambrian) is at least 40 m thick on Bache Peninsula (Fig. 12) and consists of a grey-brown, brown weathering dolomite and limestone. Faunas of this and other units have been described and compared to Greenland faunas by Poulsen (1946). Troelsen (1950: 48) commented that



Fig. 6. Cambrian carbonates overlying Lower Cambrian sandstones of the Dallas Bugt Formation (DB) and Proterozoic sandstones of the Rensselaer Bay Formation (RB), view looking south-west toward Kap Leiper, with Rensselaer Bugt beyond, Inglefield Land. CL, Cape Leiper and Cape Ingersoll Formations; WR, Wulff River Formation; CK, Cape Kent Formation; CW, Cape Wood Formation. The closely similar sequence on Bache Peninsula is illustrated in Fig. 12, where the Police Post Formation is equivalent to the Wulff River Formation.



Fig. 7. Cambrian and Lower Ordovician carbonates, innermost Cass Fjord, Washington Land. Resistant Ordovician limestones of the Cape Clay Formation and uppermost Cass Fjord Formation (CC) overlie the Middle–Upper Cambrian Cass Fjord Formation (CF) and dolomites of the Lower–Middle Cambrian Kastrup Elv Formation (KE). The Cass Fjord Formation is about 450 m thick in this section. Horizons a and b in the Cass Fjord Formation can be recognised in Fig. 13, Flagler Bay, Ellesmere Island.

"there can be no reasonable doubt about the identity of the limestone and conglomerates [above the Cape Wood Formation on Bache Peninsula] with the Cass Fjord formation of southern Washington Land ...". The formation attains a thickness of about 450 m at Flagler Bay (Fig. 13), similar to the thickness in Washington Land (Fig. 7). However, the upper two thirds of this unit, together with younger Cambrian and Ordovician formations, are not preserved in Inglefield Land.

The Cape Clay Formation (Early Ordovician) consists of about 90 m of pale, cliff-forming dolomite (Fig. 13). This is almost twice the thickness recorded from near the type section in southern Washington Land where original limestone persists (Fig. 7). However, dolomitised Cape Clay Formation is also present in Washington Land, east of the Cass Fjord area, and Poulsen (1946) has stressed the close faunal similarity between outcrops of this formation on both sides of Nares Strait.



Fig. 8. Lower Ordovician strata near the head of Cass Fjord, Washington Land. Pale weathering, evaporite-rich Poulsen Cliff Formation (PC) overlying banded limestones, dolomites and sandstones of the Christian Elv Formation (CE, 140 m thick) and pale limestones of the Cape Clay Formation (CC).

Christie (1967) referred about 90 m of limestones conformably overlying the Cape Clay Formation to 'Map-unit 6' (Fig. 2) and suggested a correlation with some or all of the Poulsen Cliff Formation to the Nunatami Formation sequence recorded by Troelsen (1950) from southern Washington Land. It is now evident that 'Map-unit 6' can be referred to the Christian Elv Formation (Fig. 8) of Henriksen & Peel (1976), the existence of which was unknown to Troelsen (1950) and Christie (1967).

Christie (1967) referred gypsum-rich strata overlying 'Map-unit 6' (Fig. 14) to the Cornwallis Group but in a footnote (p. 49) commented that this correlation on the basis of the gypsum horizon was even then known to be



Fig. 9. Lower Ordovician strata, north side of Cass Fjord, Washington Land. Cliff-forming limestones of the Nunatami Formation (N) and the Canyon Elv Formation (CE) overlying evaporite-rich carbonates and shales of the Nygaard Bay Formation (NB) and the Poulsen Cliff Formation (PC). The evaporites of the Poulsen Cliff Formation and the upper Nygaard Bay Formation are separated by dark, cliff-forming limestones of the lower Nygaard Bay Formation (c. 20 m in thickness). An equivalent section on Bache Peninsula is shown in Fig. 14, where evaporite-bearing strata are referred to the Baumann Fiord Formation.



Fig. 10. Lower–Middle Ordovician sequence at Gonioceras Bugt (left of picture), southern Washington Land. Cliff-forming, dark limestones of the Gonioceras Bay Formation (GB) overlie the pale, recessive, evaporite-bearing Cape Webster Formation (CW). Equivalent evaporites in Ellesmere Island are referred to the Bay Fiord Formation. Cliffs in the foreground (c. 200 m) capped by limestones of the Nunatami Formation (N) show the sequence illustrated in Fig. 9. Aerial photograph $545K_1$–SØ, no. 2301; copyright Geodætisk Institut, Denmark.

incorrect. Member A of the Cornwallis Formation as used by Christie (1967) is referred to the Baumann Fiord Formation, while Member B is referred to the Eleanor River Formation; both units are discussed below.

Cambrian and Ordovician platform stratigraphy of eastern Ellesmere Island and Judge Daly Promontory

The following account of Cambrian and Ordovician stratigraphy in eastern Ellesmere Island and Judge Daly Promontory draws heavily on the studies of Kerr (1967a, b, c, 1968).

Cambrian strata include the Ellesmere Group (the Archer Fiord, Ritter Bay, Rawlings Bay and Kane Basin Formations), the Scoresby Bay and Parrish Glacier Formations and the lower part of the Copes Bay Formation (Figs 2 and 5). These units have been correlated with some confidence with distinctly thinner units on Bache Peninsula (Kerr 1967b, c), discussed in the previous section (Fig. 2).

The Lower Cambrian Ellesmere Group (140–1400 m) rests with regional angular unconformity upon the predominantly carbonate Ella Bay Formation. Kerr (1967b) considered this latter unit to be of late Proterozoic age, but a correlation with the Early Cambrian Portfjeld Formation of Greenland is possible (Fig. 5). The Ellesmere Group is everywhere marked at its base by a basal pebble conglomerate or by sandstone. A gradation occurs from rather pure quartzose sandstone in the south-east to shalier sandstones and ultimately to shaly sandstones and black phyllite toward the north-west, nearer to the axis of the trough (Fig. 2).

Near the former hinge line north of Bache Peninsula (Fig. 2) the Ellesmere Group consists of only two formations (the Rawlings Bay and Kane Basin Formations). The lower, Rawlings Bay Formation, contains coarse pebble conglomerates in this region; to the west it thickens and the lower part grades into black phyllites of the Ritter Bay Formation. The Archer Fiord Formation, quartz sandstone with pebbly beds at the base, underlies the Ritter Bay Formation to form a basal, north-westerly protruding tongue of the Rawlings Bay clastic unit. The Kane Basin Formation is a fine-grained, quartzose sandstone in the thin sections near Bache Peninsula and grades through shaly siltstone to black shales and phyllites in thicker sections to the north-west (Fig. 2).

Overlying the predominantly clastic Ellesmere Group

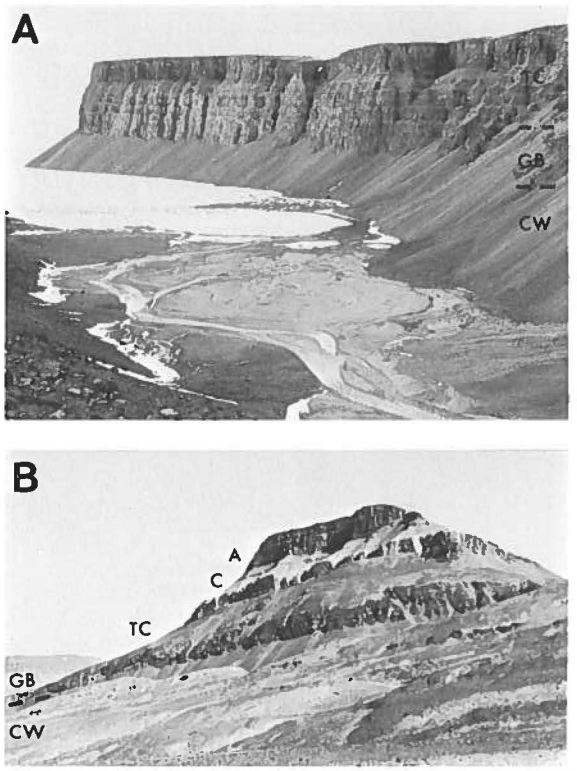


Fig. 11. Middle Ordovician – Silurian limestones of the Morris Bugt Group (GB, TC, C, A) overlying the Cape Webster Formation (CW), Washington Land. A: Gonioceras Bay (GB) and Troedsson Cliff (TC) Formations, Middle–Upper Ordovician, overlying the Cape Webster Formation (CW) at Wright Bugt, Washington Land. Height of the cliffs is in excess of 300 m. B: Gonioceras Bay (GB) and Troedsson Cliff (TC) Formations overlain by the recessive Cape Calhoun Formation (C) and the cliff-forming Aleqatsiaq Fjord Formation (A), Aleqatsiaq Fjord, Washington Land. Uppermost beds of the Aleqatsiaq Fjord Formation are of Early Silurian age. In Ellesmere Island, limestones of the Gonioceras Bay and Troedsson Cliff Formations are referred to as the Thumb Mountain Formation, while the recessive Cape Calhoun Formation is known as the Irene Bay Formation. The Aleqatsiaq Fjord Formation of Greenland is termed the Allen Bay Formation in Canada. Height of the section is c. 600 m.

is the Lower Cambrian Scoresby Bay Formation (260–860 m), composed mainly of limy, sugary, mottled dolomite (Figs 2 and 5). The Scoresby Bay Formation rests conformably on the Ellesmere Group except in northern exposures where it is locally unconformable. The formation thickens gradually and rather uniformly to the north-west across the platform.

The Parrish Glacier Formation (265–647 m) consists of a variety of brightly coloured, mainly carbonate lithologies. It conformably overlies the Scoresby Bay Formation and similarly thickens gradually and uniformly towards the north-west.

The Cambro-Ordovician boundary lies within the Copes Bay Formation (Figs 2 and 5) which was tentatively considered to be entirely Early Ordovician in age (Kerr 1968). Recently, however, faunas of Late Cambrian age have been discovered in the Copes Bay Formation and the equivalent Cass Fjord Formation of Greenland (Henriksen & Peel 1976, Mayr 1978, Peel & Cowie 1979, Palmer & Peel 1981). The Copes Bay Formation is a thick shallow water succession, mainly composed of limestone, argillaceous limestone, flat-pebble conglomerate and minor amounts of evaporites. The unit is thickest (1400 m) at the type section, about 30 km north of Bache Peninsula; it thins markedly southward, toward the craton, and more gradually northward, toward the trough. Its lithology is fairly uniform across the basin, but evaporites are most prominent (i.e. more and thicker evaporite beds) at the type section, which evidently lay near a basinal axis. Thin evaporites are also present in the equivalent Cass Fjord Formation of Washington Land (Henriksen & Peel 1976).

Ordovician to Early Silurian platform sedimentation west of Nares Strait was dominated by carbonate and evaporite deposition, with the Baumann Fiord and Eleanor River Formations, the Cornwallis Group, and the Allen Bay Formation overlying the Copes Bay Formation (Figs 2, 3 and 5).

The Baumann Fiord Formation, named in central Ellesmere Island (Fig. 16), is mainly evaporites (Fig. 14). The formation attains a maximum thickness of 475 m but thins to the north-west, across the platform and south-eastward toward the craton. The evaporites disappear north-eastward on Judge Daly Promontory, where they perhaps are in part replaced by a disconformity (Fig. 2). The Baumann Fiord Formation thus appears to be an elongated lens in form, the trend of the former basin paralleling the north-easterly regional structural direction (Fig. 15).

The nature of the north-western termination of the Baumann Fiord Formation is not well known at the present stage of knowledge. The possibility of a regional disconformity around the edge of the basin has been noted (Kerr 1968: 25), but it is clear that the Baumann Fiord Formation is equivalent to part of the (mainly) overlying Eleanor River Formation. An upper evaporitic member of the Baumann Fiord Formation (member C in Fig. 14) grades into the Eleanor River Formation, and the middle limestone member B is indistinguishable from, and passes into, the Eleanor River Formation. The very much thicker, lower evaporite member A passes into enclosing formations at the margin of the Baumann Fiord basin, at least in part by gradation. The north-eastern boundary of the evaporite facies crosses Nares Strait north of Rawlings Bay (see Fig. 15); it is represented in Washington Land by the Poulsen Cliff and Nygaard Bay Formations.

The Eleanor River Formation, a Lower and Middle Ordovician limestone unit that forms numerous prominent bluffs on Ellesmere Island (Fig. 14), overlies the Baumann Fiord Formation or, where evaporites are missing, the Copes Bay Formation. The Eleanor River



Fig. 12. The Lower–Middle Cambrian sequence, ravine behind 'Police Post', southern coast of Bache Peninsula. DB, Dallas Bugt Formation; CL, Cape Leiper Formation; CI, Cape Ingersoll Formation; PP, Police Post Formation; CK, Cape Kent Formation; CW, Cape Wood Formation. The similar sequence of Inglefield Land is illustrated in Fig. 6.

Formation is lensoid or trough-formed, similar to and parallel with the Baumann Fiord Formation and reaches a maximum thickness of about 1 km. The Eleanor River Formation thins north-westward both across the platform and south-eastward toward the craton in similar fashion to the underlying Baumann Fiord Formation. Whether or not the Eleanor River Formation is separated from the Copes Bay Formation by a disconformity on the outer platform is unclear: the units are structurally conformable and the Baumann Fiord Formation evaporites may be represented by basal beds of the Eleanor River Formation (Fig. 2).

The lithological characteristics of the Eleanor River (and other Ordovician units) change north-westward across Judge Daly Promontory toward the Franklinian Trough. However, the changes are gradual and the nature of the facies boundary is not yet fully known.

The youngest group of rocks considered here is the Cornwallis Group, which comprises three distinctive carbonate and evaporite units that conformably overlie the Eleanor River Formation. The succession of formations, from the lowest, is: the Bay Fiord Formation, recessive shaly limestone, siltstone, evaporites and dolomite; the Thumb Mountain Formation, cliff-forming dark grey limestone; and the Irene Bay Formation, recessive greenish weathering shaly limestone (Fig. 5). The Cornwallis Group ranges in age from Middle to early Late Ordovician. The equivalent formations in

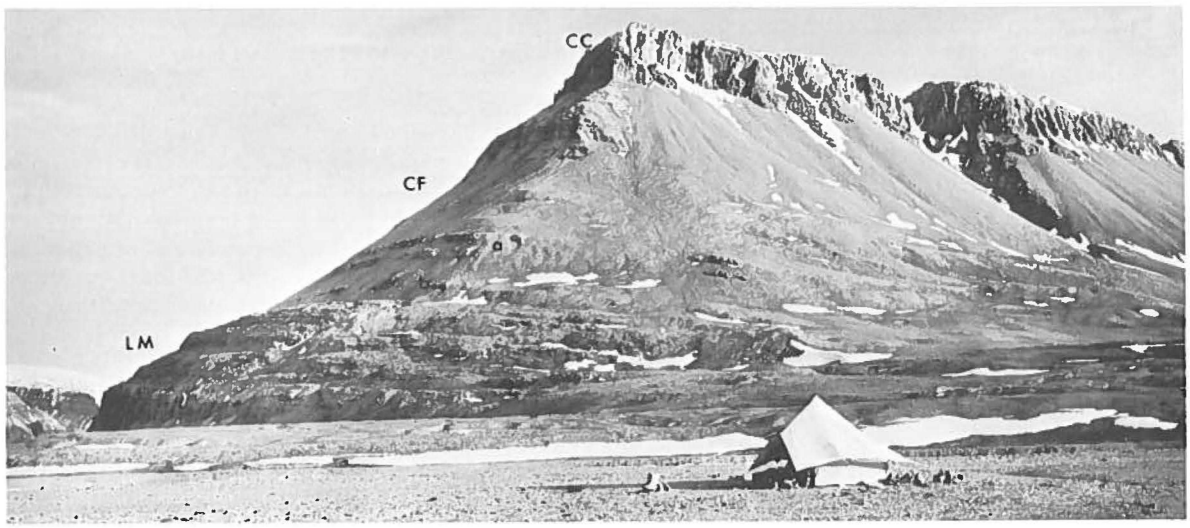


Fig. 13. Cambrian–Ordovician carbonates at Flagler Bay, Bache Peninsula. The resistant Cape Clay Formation (CC) and the uppermost Cass Fjord Formation form a cap-rock to pale, recessive outcrops of the Cass Fjord Formation (CF) and the Lower–Middle Cambrian carbonate sequence (LM) illustrated in Fig. 12. Horizons a and b can be recognised in Fig. 7 in Washington Land.



Fig. 14. Lower Ordovician evaporites of the Baumann Fiord Formation (BF), Sanddöla Creek, western Bache Peninsula. The recessive evaporite members A and C are separated by the cliff-forming member B. Cliff-forming carbonates of 'Map-unit 6' (M, equivalent to the Christian Elv Formation of Washington Land) and the Eleanor River Formation (ER) occur below and above, respectively, the 265 m thick evaporite formation. An equivalent section in Greenland is shown in Fig. 9, where the Eleanor River Formation is known as the Canyon Elv and Nunatami Formations and the Baumann Fiord Formation as the Poulsen Cliff and Nygaard Bay Formations.

Greenland showing characteristic weathering patterns are illustrated in Figs 10 and 11.

The Bay Fiord Formation (330–1100 m) exhibits marked lateral changes in lithology. A conspicuous evaporite component occurs in a narrow, north-east trending belt near the south-eastern margin of the platform (Fig. 2); limestone grades north-westward into shaly limestone and argillite in the north-western part of the platform, toward the trough. The Bay Fiord evaporite belt of Ellesmere Island is a narrowing north-eastern extension of a broad evaporite basin on Bathurst, Cornwallis and Devon Islands; in Greenland the evaporite basin is represented within the Cape Webster Formation (Figs 10 and 11).

The upper formations of the Cornwallis Group, the Thumb Mountain and Irene Bay Formations, form a somewhat lensoid body with an axis of maximum thickness trending north-east, parallel to earlier depositional trends. The depositional axis of the combined Thumb Mountain and Irene Bay Formations roughly coincides with that of the combined Eleanor River Formation and the Cornwallis Group. The Thumb Mountain Formation (150–600 m) becomes somewhat more shaly and less resistant in outcrop to the north-west, repeating in a less distinct way the 'shaling-out' feature of the underlying Bay Fiord Formation. The overlying Irene Bay Formation (36–83 m) similarly is replaced northward and westward by beds of dark grey to black shale and in places is indistinguishable as a separate formation of the Cornwallis Group. The Thumb Mountain Formation is equivalent to the revised Gonioceras Bay and Troedsson Cliff Formations of Washington Land (Peel & Hurst 1980); the Irene Bay Formation correlates with the Cape Calhoun Formation (Figs 5, 10 and 11).

Cambrian and Ordovician of the Franklinian Trough

Description of Cambrian and Ordovician trough sequences is beyond the scope of this paper and the brief discussion given here is only intended to supplement Figs 1–3. Recent summaries of the trough sequence in northern Ellesmere Island are given by Trettin & Balkwill (1979) and Trettin et al. (1979), and in Greenland by Dawes & Soper (1979), Soper et al. (1980), Surlyk et al. (1980), Friderichsen et al. (1982) and Higgins et al. (in press).

The Grant Land Formation of northern Ellesmere Island (Figs 2 and 3) is a thick sequence of alternating sandstones and phyllites, much of which was deposited by braided rivers. The boundary with the overlying Hazen Formation is strongly diachronous, from Early to Late Cambrian age, and the Grant Land Formation can be correlated with much of the Ellesmere Group of Judge Daly Promontory, to the south.

The Hazen Formation is a condensed sequence of carbonates, fine clastics and cherts which ranges in age from Early Cambrian to Early Silurian (Trettin et al. 1979). Its upper and lower boundaries are diachronous, the contact with the overlying turbidites of the Imina Formation being near the Ordovician–Silurian boundary.

Rocks of these three formations can be traced to the shores of Nares Strait. The same sequences, although referred to different formations, make up the trough succession across northern Greenland (Dawes & Soper 1979, Hurst & Surlyk 1980, Surlyk et al. 1980, Dawes, this volume, Surlyk, this volume).

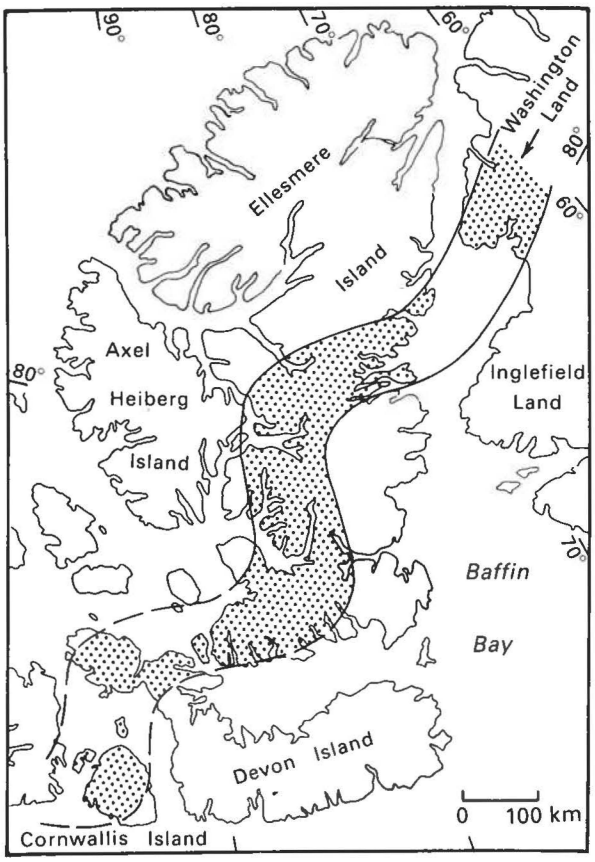


Fig. 15. Lower Ordovician evaporites of the Baumann Fiord Formation (Arctic Canada) and the Poulsen Cliff and Nygaard Bay Formations (North Greenland). Adapted from Mossop (1979).

Geological correlation across Nares Strait

The considerable similarity between Cambrian and Ordovician platform sequences on either side of Nares Strait is evident from even cursory examination of published accounts, and a ready correlation is possible (Fig. 5). Where geological field work has been undertaken across the Strait, the same stratigraphic terminology has been employed in both Greenland and Ellesmere Island (Troelsen 1950). If new information from the 'classic' sequences in Greenland, originally described by Koch (1925, 1929a, b, 1933), Troedsson (1926, 1928), Poulsen (1927) and Troelsen (1950), had been available to Kerr (1967a, b, c, 1968) it would undoubtedly have been possible to employ many of the older stratigraphic terms from Greenland in Ellesmere Island instead of formulating new names. The dual system of nomenclature also reflects the presence of a national frontier between Greenland and Ellesmere Island.

It is emphasised that much of the geological information available is of a regional character. However, such detailed studies as are currently on hand strengthen the correlation between Ellesmere Island and Greenland; a convincing example is the recent study by Mossop (1979) of Lower Ordovician evaporites forming the Baumann Fiord Formation in Ellesmere Island.

The Baumann Fiord Formation is subdivided into three members: member A is a thick, recessive, anhydrite-bearing unit; member B is a resistant limestone unit; member C is a thin, recessive anhydrite-bearing unit (Fig. 14). The formation is overlain by resistant limestones of the Eleanor River Formation. Member A of the Baumann Fiord Formation is correlated conclusively with the Poulsen Cliff Formation of Washington Land, while members B and C equate with the lower and upper members of the Nygaard Bay Formation (Fig. 9).

Mossop (1979) noted that the evaporites of the Baumann Fiord Formation extended in a narrow, sinuous belt some 1100 km long, from Cornwallis Island in the south-west to Ellesmere Island in the north-east (Fig. 15). Following the recognition of evaporites in the Poulsen Cliff Formation (Henriksen & Peel 1976), this

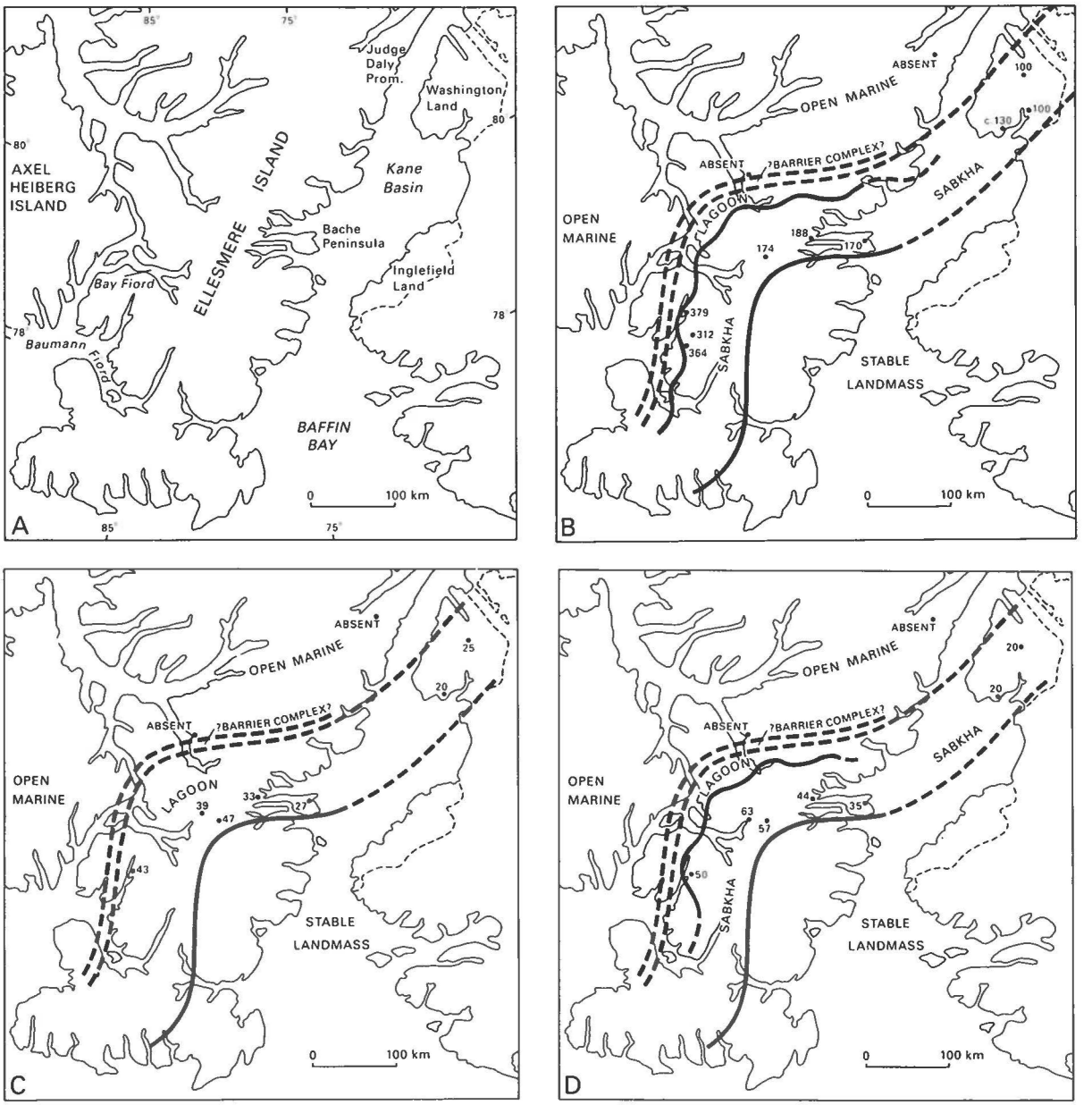


Fig. 16. Palaeogeography of Lower Ordovician evaporites in Ellesmere Island and adjacent Greenland. A: locality map; B: member A of the Baumann Fiord Formation of Ellesmere Island and the Poulsen Cliff Formation of Washington Land; C: member B of the Baumann Fiord Formation and the lower member of the Nygaard Bay Formation of Washington Land; D: member C of the Baumann Fiord Formation and the upper member of the Nygaard Bay Formation. In B and D evaporites were deposited on a sabkha plain passing seawards into a lagoon which formed behind a postulated barrier complex. In C lagoonal carbonates were deposited throughout the area with no accumulation of evaporites. Thicknesses in metres; the diagram is extended into Greenland from figs 24–26 of Mossop (1976).

distribution can now be extended some 250 km across Washington Land to Petermann Gletscher. Ice covers Ordovician outcrops to the east of Washington Land until the Warming Land and Wulff Land area of central North Greenland (Fig. 1); here evaporites of Early Ordovician age have not been recognised (Peel 1980).

Mossop (1979: figs 24–26) drew three palaeogeographic maps, one for each of the members of the Baumann Fiord Formation. These form the basis of Fig. 16 where recent, previously unpublished, information from the Poulsen Cliff and the Nygaard Bay Formations of Washington Land is included. Of particular note is

the tendency for Washington Land units to be thinner, supporting the inference from Peel's (1980) observations that the evaporite basin is dying out toward the east. The continuity of the strata in question across Nares Strait is self-evident but it is unfortunate that Silurian cover prevents establishment of a north-western boundary to the evaporite-bearing sequence in Greenland of the type recognised in Ellesmere Island.

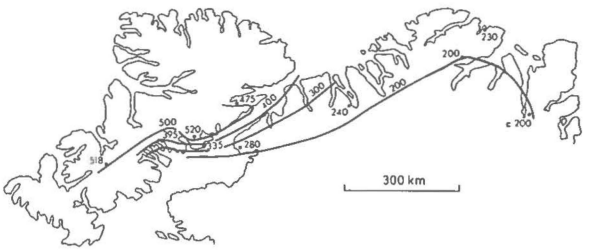


Fig. 17. Isopach map for the Bay Fiord Formation of Ellesmere Island and the Cape Webster Formation (and equivalents) of North Greenland showing the regional continuity of the Ordovician deposition across Nares Strait. Thicknesses in metres; sections in Ellesmere Island from Kerr (1967a, 1968).

The question of displacement along Nares Strait

The comparison and correlation of Cambrian and Ordovician platform sediments around Kane Basin contribute several parameters to the discussion of how much post-depositional, net transcurrent displacement, if any, has taken place between Ellesmere Island and Greenland.

1. Regional depositional framework

The close similarity between sections in Ellesmere Island and adjacent Greenland, discussed in the previous section, indicates that both regions formed part of the same sedimentary platform in Cambrian and Ordovician times. This platform has a wide regional extent but is relatively narrow, with a well-developed trough (the Franklinian Trough) forming its northern margin and crystalline basement rocks lying to the south. The continuity at the present day of these three components of the regional geology, the Franklinian Trough, the Arctic Platform and the Precambrian Shield, indicates that net sinistral displacement along Nares Strait of the order of 200 km or more has not occurred. This argues against conventional plate tectonic, sea-floor spreading models of the region (e.g. Keen et al. 1972, Herron et al. 1974, Sclater et al. 1977, Srivastava 1978, Srivastava et al. 1981, Srivastava & Falconer, this volume) as well as those interpretations of the geology of the Nares Strait region put forward in support of such models (e.g. Newman 1977, this volume, Newman & Falconer 1978, Miall 1981, Peirce, this volume).

2. Geological correlation

A feature of Cambrian and Ordovician sequences around Kane Basin is the close correlation between sections on the Canadian and Greenlandic sides of the basin (compare, for example, Figs 9 and 14). Thus, Cambrian sections on Bache Peninsula correlate most closely with those in Inglefield Land (Figs 4, 6 and 12). Correlation from Bache Peninsula south across Ellesmere Island is less satisfactory, as is correlation from Inglefield Land eastward to Washington Land and across North Greenland. Sections are readily correlated across the platform toward the Franklinian Trough, i.e. from the inner platform sequences of Bache Peninsula to the outer platform sequences of Judge Daly Promontory, but considerable changes in facies and thickness are evident (Fig. 2). This close similarity of Cambrian sequences on Bache Peninsula and in Inglefield Land, and decreasing ease of correlation away from Kane Basin, strongly suggests that these two areas were closely juxtaposed during Cambrian times and that they have not subsequently suffered large-scale net displacement.

It is clear that not all Cambrian and Ordovician stratigraphic units permit meaningful comment concerning transcurrent displacement along Nares Strait. In the Middle Ordovician, the incoming of limestones of the Thumb Mountain, Gonioceras Bay and Børglum River Formations (Fig. 5) is a regional event which can be recognised throughout much of the Canadian Arctic and Greenland. The limestones are uniform in character and can be traced unchanged over large distances, although they are absent from the Franklinian Trough. In the absence of detailed stratigraphic and sedimentological studies, comparison of these limestones across Kane Basin can neither confirm nor deny displacement along Nares Strait.

3. Isopach maps

Kerr (1967a) used isopach maps to compare sequences of Cambrian and Ordovician sediments around Kane Basin, although much information concerning general stratigraphy and thicknesses from Greenland at that time was of necessity derived from older imprecise literature. These maps can now be replotted, using new information from Greenland combined with Kerr's (1967b, c, 1968) data from Ellesmere Island, and a typical example is given here (Fig. 17, see also Peel et al., this volume: fig. 12). Few major departures from Kerr's maps were noted, but it soon became evident that comparison of isopach maps across Kane Basin at this regional scale could provide little *precise* evidence concerning displacement along Nares Strait. However, very large-scale displacements (200 km or greater) in a post-depositional sinistral sense are judged extremely unlikely. The lack of precise control concerning possible

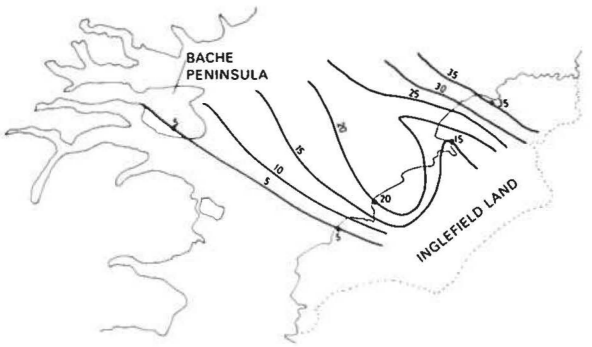


Fig. 18. Isopach for the Police Post Formation of Bache Peninsula and the Wulff River Formation of Inglefield Land. Thicknesses in metres.

displacement is partly due to the lack of 'down-dip' exposure of the Cambrian and Ordovician in Greenland where most exposures tend to follow the regional strike, and partly due to the large distances over which extrapolation is necessary.

More detailed isopach maps can play a role in assessing relationships across Nares Strait (Fig. 18). Again, large-scale movements can be ruled out, but the scarcity of control points at the present time can neither confirm nor deny smaller movements.

A relatively constant feature of the regional isopach maps (Fig. 17, Peel et al., this volume: fig. 12) is an embayment over western North Greenland. If it is rather unrealistically assumed that the basin of deposition originally had a straight margin, the embayment can be cited as evidence for substantial post-depositional dextral displacement of Greenland relative to Ellesmere Island. This sense of movement is opposite to that which could be inferred, equally unrealistically, from the northward swing in tectonic belts described by Higgins et al. (this volume). It is also opposite to the direction of net displacement invoked by proponents of large-scale displacement between Greenland and Ellesmere Island from the early ideas of Taylor (1910) and Wegener (1915) to modern plate tectonic pre-drift fits, for example of Srivastava (1978), Irving (1979), Miall (1981) and Srivastava & Falconer (this volume). The change in trend of the isopachs is clearly an original feature, as is the swing in tectonic trends discussed by Higgins et al. (this volume), and in part reflects the influence of the Bache Peninsula area as a tectonic high.

4. Lower Cambrian hinge line

A transect from Bache Peninsula northward through eastern Ellesmere Island to Judge Daly Promontory (Fig. 2) shows a rapid increase in thickness of Lower Cambrian sediments just north of Bache Peninsula. The Dallas Bugt Formation of the former area, deposited on the inner platform, passes into the outer platform Ellesmere Group, which is underlain by carbonates of the Ella Bay Formation, across a relatively well-developed hinge line. This hinge line is not so obvious in adjacent Greenland, due to cover by stratigraphically younger sediments. However, it is evident that the thin clastic sequences of the Dallas Bugt and Humboldt Formations in Inglefield Land and Washington Land lie to the south-east of the hinge line; carbonates equivalent to the Ella Bay Formation are also absent from Inglefield Land and Washington Land, as well as from Bache Peninsula.

The presence of inner platform Lower Cambrian sediments in southern Washington Land indicates that large-scale (200 km or more) post-depositional net dextral movement of Greenland relative to Ellesmere Island has not taken place. If such a movement had occurred it would be expected that thick clastic sequences of the outer platform would be preserved in southern Washington Land, as is patently not the case. Post-depositional sinistral displacement cannot be ruled out on this criterion alone. However, the inner platform clastics in Greenland thin rapidly from Washington Land to the south-west across Inglefield Land (see Peel et al., this volume: figs 3 and 12) in comparable fashion to thinning seen toward Bache Peninsula in Ellesmere Island. Displacement of 150 km or more would produce an unacceptable dislocation between the successions on either side of Nares Strait.

5. Ordovician evaporites

Evaporites are a conspicuous feature of the Lower and Middle Ordovician of Ellesmere Island and western North Greenland. Two units are present, represented by the Baumann Fiord, Poulsen Cliff and Nygaard Bay Formations (Figs 5, 8, 9, 14, 15 and 16) and by the Bay Fiord and Cape Webster Formations, respectively (Figs 5, 10, 11 and 17). Evaporites are the product of a specialised sedimentological environment and the stratigraphic continuity of these deposits from Ellesmere Island into North Greenland once again stresses the considerable geological similarity between the two regions.

The recent detailed study of the *Lower Ordovician* evaporites (Baumann Fiord Formation) in Ellesmere Island by Mossop (1979) is readily extended into Washington Land. Mossop considered that the evaporites were deposited on a sinuous sabkha plain approximately 100 km wide which can now be traced some 1350 km from Cornwallis Island in the south-west to Washington Land in the north-east (Figs 15 and 16). Equivalent evaporites occur throughout Washington Land from Kane Basin to Petermann Gletscher, within the Poulsen Cliff and Nygaard Bay Formations (Figs 3, 8 and 9).

The undoubted continuity of this relatively narrow sabkha plain from Ellesmere Island to Greenland places severe constraints on the magnitude of possible transcurrent movements along Nares Strait and suggestions

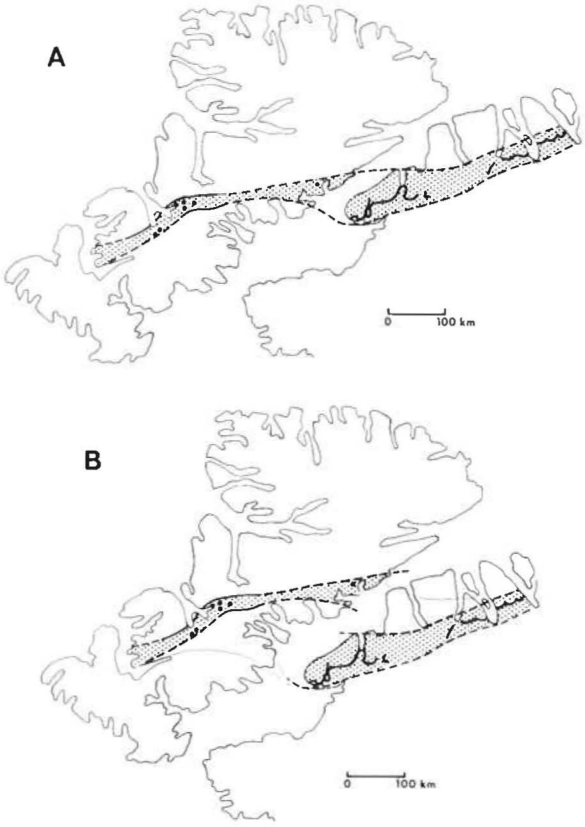


Fig. 19. Middle Ordovician evaporites of the Bay Fiord Formation (Ellesmere Island) and the Cape Webster Formation (Greenland). Outcrops in Greenland are indicated by solid black; sections with evaporites noted by Kerr (1968) in Ellesmere Island are indicated by black dots. A: present distribution; B: distribution with Greenland displaced 200 km dextrally. This totally unacceptable dislocation of the evaporite belt corresponds to the pre-drift position of Greenland relative to Ellesmere Island required by proponents of a 200 km sinistral displacement along Nares Strait. Conventional plate tectonic reconstructions of the region commonly invoke sinistral displacement of 200–300 km or more. Similar diagrams can be constructed for the Baumann Fiord Formation evaporite belt (Fig. 16), that also demonstrate the impossibility of these large-scale displacements.

of displacement of 200 km or more can be refuted. Natural reconstruction of the Lower Ordovician sabkha plain, taking full account of regional structural and environmental trends, requires no net displacement between Ellesmere Island and Greenland (Fig. 16). Minor sinistral displacement of the evaporite belt may have taken place, but it is judged that this movement, if present, could not have exceeded 50 km. This value is without positive geological support and merely reflects the inevitable limitations of any correlation across an area of this size.

The evaporites in the lower part of the *Middle Ordovician* Bay Fiord Formation of Ellesmere Island have not been subjected to the same detailed study as their Lower Ordovician counterparts. Kerr (1968) speculated that the Bay Fiord Formation could be correlated with the Cape Webster Formation of Greenland, although evaporites were not known at that time from this formation. Evaporites are now known to occur within the Cape Webster Formation from Washington Land to Wulff Land (Peel 1977, 1980), and the correlation is now secure. Kerr (1968: fig. 9) considered the Bay Fiord Formation evaporites to occur in a very narrow belt across Ellesmere Island, but the general recessive character of the formation, and consequent poor exposure, makes precise delimitation of such a belt difficult. In Washington Land white weathering evaporites are prominent at the base of the Cape Webster Formation as in Ellesmere Island, but darker coloured evaporites are diffused throughout the formation in strata which appear to be poorly exposed shales to a cursory glance. In the absence of more detailed observations it is not considered possible to delimit the evaporite belt as narrowly as Kerr (1968) would suggest, since data from the Canadian and Greenlandic sides are not strictly homogenous. However, the continuity of this specialised sedimentary environment from Ellesmere Island to Greenland refutes large-scale (200 km or more) net post-depositional sinistral displacement along Nares Strait (Fig. 19). Indeed, any sinistral displacement at all would be difficult to accommodate.

Conclusions

Correlation of Cambrian and Ordovician platform sequences around Kane Basin provides no evidence to suggest that net transcurrent displacement has occurred between Ellesmere Island and Greenland. The continuity of regional depositional environments evident from the close similarity of geological units on either side of Nares Strait refutes plate tectonic models invoking large-scale lateral displacements (200 km or more). The reconstruction of the Lower Ordovician evaporite basin permits a maximum net strike-slip displacement in either a sinistral or dextral sense of up to 50 km along Nares Strait, but even this value is a measure of tolerance within the model itself, rather than an indication that any movement has taken place.

Acknowledgements

P. R. Dawes is thanked for critically reviewing the manuscript and, together with H. P. Trettin, for providing unpublished information in the form of recent geological map compilations that have been used in Fig. 3. The aerial photograph in Fig. 10 is published with the permission (A. 495/79) of the Geodætisk Institut, Copenhagen, Denmark. The paper is published with the permission of the Director of the Geological Survey of Greenland.

References

Armstrong, H. A. & Lane, P. D. 1981. The un-named Silurian(?) dolomite formation, Børglum Elv, central Peary Land. – Rapp. Grønlands geol. Unders. 106: 29–34.

Christie, R. L. 1967. Bache Peninsula, Ellesmere Island, Arctic Archipelago. – Mem. geol. Surv. Can. 347: 63 pp.

Christie, R. L. & Peel, J. S. 1977. Cambrian–Silurian stratigraphy of Børglum Elv, Peary Land, eastern North Greenland. – Rapp. Grønlands geol. Unders. 82: 48 pp.

Collinson, J. D., Dawes, P. R., Peel, J. S. & Christie, R. L. (in press). Late Proterozoic and basal Cambrian stratigraphy of Inglefield Land, North-West Greenland: redefinition and correlation with Bache Peninsula, Ellesmere Island. – Rapp. Grønlands geol. Unders.

Dawes, P. R. 1982. The Nyeboe Land fault zone: a major dislocation on the Greenland coast along northern Nares Strait. – This volume.

Dawes, P. R. & Soper, N. J. 1979: Structural and stratigraphic framework of the North Greenland fold belt in Johannes V. Jensen Land, Peary Land. – Rapp. Grønlands geol. Unders. 93: 40 pp.

Friderichsen, J. D., Higgins, A. K., Hurst, J. M., Pedersen, S. A. S., Soper, N. J. & Surlyk, F. 1982. Lithostratigraphic framework of the Upper Proterozoic and Lower Palaeozoic deep water clastic deposits of North Greenland. – Rapp. Grønlands geol. Unders. 107: 20 pp.

Henriksen, N. & Peel, J. S. 1976: Cambrian–Early Ordovician stratigraphy in south-western Washington Land, western North Greenland. – Rapp. Grønlands geol. Unders. 80: 17–23.

Herron, E. M., Dewey, J. F. & Pitman, W. C. 1974. Plate tectonic model for the evolution of the Arctic. – Geology 2: 377–380.

Higgins, A. K., Mayr, U. & Soper, N. J. 1982. Fold belts and metamorphic zones of northern Ellesmere Island and North Greenland. – This volume.

Higgins, A. K., Soper, N. J. & Friderichsen, J. D. (in press). The North Greenland fold belt in eastern North Greenland. – In: Gee, D. G. & Sturt, B. A. (eds), The Caledonide Orogen – Scandinavia and related areas. – John Wiley & Sons, New York.

Hurst, J. M. 1980. Silurian stratigraphy and facies distribution in Washington Land and western Hall Land, North Greenland. – Bull. Grønlands geol. Unders. 138: 95 pp.

Hurst, J. M. & Kerr, J. W. 1982. Upper Ordovician to Silurian facies patterns in eastern Ellesmere Island and western North Greenland and their bearing on the Nares Strait lineament. – This volume.

Hurst, J. M. & Peel, J. S. 1979. Late Proterozoic(?) to Silurian stratigraphy of southern Wulff Land, North Greenland. – Rapp. Grønlands geol. Unders. 91: 37–56.

Hurst, J. M. & Surlyk, F. 1980: Notes on the Lower Palaeozoic clastic sediments of Peary Land, North Greenland. – Rapp. Grønlands geol. Unders. 99: 73–78.

Ineson, J. R. 1980. Carbonate debris flows in the Cambrian of south-west Peary Land, eastern North Greenland. – Rapp. Grønlands geol. Unders. 99: 43–49.

Ineson, J. R. & Peel, J. S. 1980. Cambrian stratigraphy in Peary Land, eastern North Greenland. – Rapp. Grønlands geol. Unders. 99: 33–42.

Irving, E. 1979. Paleopoles and paleolatitudes of North America and speculations about displaced terrains. – Can. J. Earth Sci. 16: 669–694.

Jepsen, H. F. 1971. The Precambrian, Eocambrian and early Palaeozoic stratigraphy of the Jørgen Brønlund Fjord area, Peary Land, North Greenland. – Bull. Grønlands geol. Unders. 96: 42 pp. (also Meddr Grønland 192(2)).

Keen, C. E., Barrett, D. L., Manchester, K. S. & Ross, D. I. 1972. Geophysical studies in Baffin Bay and some tectonic implications. – Can. J. Earth Sci. 9: 239–256.

Kerr, J. W. 1967a. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 16: 483–520.

Kerr, J. W. 1967b. Stratigraphy of central and eastern Ellesmere Island, Arctic Canada. Part I. Proterozoic and Cambrian. – Pap. geol. Surv. Can. 67–27(1): 63 pp.

Kerr, J. W. 1967c. New nomenclature for Ordovician rock units of the eastern and southern Queen Elizabeth Islands, Arctic Canada. – Bull. Can. Petrol. Geol. 15: 91–113.

Kerr, J. W. 1968. Stratigraphy of central and eastern Ellesmere Island, Arctic Canada. Part II. Ordovician. – Pap. geol. Surv. Can. 67–27(2): 92 pp.

Koch, L. 1925. The geology of North Greenland. – Am. J. Sci. 9: 271–285.

Koch, L. 1929a. Stratigraphy of Greenland. – Meddr Grønland 73,2(2): 205–320.

Koch, L. 1929b. The geology of the south coast of Washington Land. – Meddr Grønland 73,1(1): 39 pp.

Koch, L. 1933. The geology of Inglefield Land. – Meddr Grønland 73,1(2): 38 pp.

Mayr, U. 1978. Stratigraphy and correlation of Lower Paleozoic formations, subsurface of Cornwallis, Devon, Somerset and Russell Islands, Canadian Arctic Archipelago. – Bull. geol. Surv. Can. 276: 55 pp.

Miall, A. D. 1981. Late Cretaceous and Paleogene sedimentation and tectonics in the Canadian Arctic Islands. – In: Miall, A. D. (ed.), Sedimentation and tectonics in alluvial basins. – Spec. Pap. geol. Ass. Can. 23: 221–272.

Mossop, G. D. 1979. The evaporites of the Ordovician Baumann Fiord Formation, Ellesmere Island, Arctic Canada. – Bull. geol. Surv. Can. 298: 52 pp.

Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Newman, P. H. 1982. A geological case for movement between Canada and Greenland along Nares Strait. – This volume.

Newman, P. H. & Falconer, R. K. H. 1978. Evidence for movement between Greenland and Canada along Nares Strait. – Geol. Soc. Am. Abs. with Prog. 10: 463 only.

Palmer, A. R. & James, N. P. 1980. The Hawke Bay Event: a circum-Iapetus regression near the Lower Middle Cambrian boundary. – In: Wones, D. R. (ed.), The Caledonides in the U.S.A. – Mem. 2: 15–18. – Virginia Polytechnic Instit. & State Univ., Blacksburg.

Palmer, A. R. & Peel, J. S. 1981. Dresbachian trilobites and stratigraphy of the Cass Fjord Formation, western North Greenland. – Bull. Grønlands geol. Unders. 141: 46 pp.

Peel, J. S. 1977. Cambrian–Silurian studies in Washington Land, western North Greenland. – Rapp. Grønlands geol. Unders. 85: 30–33.

Peel, J. S. 1980. Cambrian and Ordovician geology of Warming Land and southern Wulff Land, central North Greenland. – Rapp. Grønlands geol. Unders. 101: 55–60.

Peel, J. S. & Cowie, J. W. 1979. New names for Ordovician formations in Greenland. – Rapp. Grønlands geol. Unders. 91: 117–124.

Peel, J. S. & Hurst, J. M. 1980. Late Ordovician and early Silurian stratigraphy of Washington Land, western North Greenland. – Rapp. Grønlands geol. Unders. 100: 18–24.

Peel, J. S. & Yochelson, E. L. 1979. *Ceratopea* (Gastropoda) from Washington Land, western North Greenland. – Rapp. Grønlands geol. Unders. 91: 87–91.

Peel, J. S., Dawes, P. R., Collinson, J. D. & Christie, R. L. 1982. Proterozoic – basal Cambrian stratigraphy across Nares Strait: correlation between Inglefield Land and Bache Peninsula. – This volume.

Peirce, J. W. 1982. The evolution of the Nares Strait lineament and its relation to the Eurekan orogeny. – This volume.

Poulsen, C. 1927. The Cambrian, Ozarkian and Canadian faunas of Northwest Greenland. – Meddr Grønland 70,1(2): 233–343.

Poulsen, C. 1946. Notes on the Cambro-Ordovician fossils collected by the Oxford University Ellesmere Island Expedition 1934–5. – Q. Jl geol. Soc. Lond. 102: 299–337.

Poulsen, V. 1964. Contribution to the Lower and Middle Cambrian palaeontology and stratigraphy of Northwest Greenland. – Meddr Grønland 164(6): 105 pp.

Sclater, J. G., Hellinger, S. & Tapscott, C. 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. – J. Geol. 85: 509–552.

Soper, N. J., Higgins, A. K. & Friderichsen, J. D. 1980. The North Greenland fold belt in eastern Johannes V. Jensen Land. – Rapp. Grønlands geol. Unders. 99: 89–98.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Srivastava, S. P. & Falconer, R. K. H. 1982. Nares Strait: a conflict between plate tectonic predictions and geological interpretation. – This volume.

Srivastava, S. P., Falconer, R. K. H. & MacLean, B. 1981. Labrador Sea, Davis Strait, Baffin Bay: geology and geophysics – a review. – In: Kerr, J. W. & Fergusson, A. J. (eds), Geology of the North Atlantic borderlands. – Mem. Can. Soc. Petrol. Geol. 7: 333–398.

Surlyk, F. 1982. Nares Strait and the down-current termination of the Silurian turbidite basin of North Greenland. – This volume.

Surlyk, F., Hurst, J. M. & Bjerreskov, M. 1980. First age-diagnostic fossils from the central part of the North Greenland foldbelt. – Nature, Lond. 286: 800–803.

Taylor, F. B. 1910. Bearing of the Tertiary mountain belt on the origin of the earth's plan. – Bull. geol. Soc. Am. 21: 179–226.

Trettin, H. P. 1971. Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, Arctic Archipelago. – Bull. geol. Surv. Can. 203: 134 pp.

Trettin, H. P. & Balkwill, H. R. 1979. Contributions to the tectonic history of the Innuitian Province, Arctic Canada. – Can. J. Earth Sci. 16: 748–769.

Trettin, H. P., Barnes, C. R., Kerr, J. W., Norford, B. S., Pedder, A. E. H., Riva, J., Tipnis, R. S. & Uyeno, T. T. 1979. Progress in lower Paleozoic stratigraphy, northern Ellesmere Island, District of Franklin. – Pap. geol. Surv. Can. 79–1B: 269–279.

Troedsson, G. T. 1926. On the Middle and Upper Ordovician faunas of northern Greenland. I. Cephalopods. – Meddr Grønland 71(1): 1–157.

Troedsson, G. T. 1928. On the Middle and Upper Ordovician faunas of northern Greenland. II. – Meddr Grønland 72,1(1): 197 pp.

Troelsen, J. C. 1950. Contributions to the geology of Northwest Greenland, Ellesmere Island and Axel Heiberg Island. – Meddr Grønland 149(7): 86 pp.

Wegener, A. 1915. Die Entstehung der Kontinente und Ozeane. – Friedr. Vieweg & Sohn, Braunschweig: 94 pp.

Upper Ordovician to Silurian facies patterns in eastern Ellesmere Island and western North Greenland and their bearing on the Nares Strait lineament

JOHN M. HURST and J. WILLIAM KERR

Hurst, J. M. & Kerr, J. W. 1982. Upper Ordovician to Silurian facies patterns in eastern Ellesmere Island and western North Greenland and their bearing on the Nares Strait lineament. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 137–145.

The upper Ordovician and lower Silurian sediments in the northern Nares Strait region formed during a series of platform margin collapses and platform subsidences. Facies changes from platform carbonates to slope and trough clastics are intricate, but two almost vertical facies fronts can be correlated across Nares Strait.

The northernmost facies front is the carbonate platform margin (associated with a horst) which apparently formed at the fault-bounded edge of the stable craton. At this margin there is a facies boundary (interdigitation) between platform carbonates and clastics of the trough and this feature had a stationary location during the upper Ordovician and lower Silurian. On Ellesmere Island it is traceable from Cañon Fiord to Judge Daly Promontory; on Greenland, directly along strike, available data indicate that the linear coastline of northern Hall Land, and probably Nyeboe Land, coincides with the platform margin. There appears to have been little or no post-depositional offset of this margin along Nares Strait.

Mudstones progressively migrated southward in Ellesmere Island during the upper Ordovician and the same facies reached Washington Land and Hall Land, Greenland, in the middle Llandovery (lower Silurian). The line of interdigitation (facies front) between carbonate and mudstone facies was a static, near vertical, feature from the middle Llandovery to upper Silurian in Washington Land and into the lower Devonian in Ellesmere Island. This facies boundary has a slightly curved form when projected across Nares Strait; a shape which is entirely consistent with the meandering nature of facies boundaries.

The present distribution and relationship of upper Ordovician to lower Silurian sediments indicate that they were part of a single laterally continuous sedimentary regime. Both vertical facies fronts are entirely consistent with no post-depositional movement along Nares Strait; however, they cannot rule out left-lateral displacement of as much as 25 km. Displacements greater than this introduce far more geological problems than they solve.

J. M. Hurst, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark and J. W. Kerr, Department of Energy, Mines and Resources, Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303–33rd St. N. W., Calgary, Alberta, Canada, T2L 2A7. J. W. K. present address: J. William Kerr & Associates Ltd., Consulting Geologists, Suite 407, 630–8th Ave. S. W., Calgary, Alberta, Canada, T2P 1G6.

Nares Strait is a strikingly linear waterway that connects Baffin Bay in the south with the Arctic Ocean in the north, and separates Greenland from Ellesmere Island, Arctic Canada (Fig. 1). This prominent feature has become the focus of a debate concerning whether the Strait is the locus of great sinistral strike-slip faulting during the Cretaceous and Tertiary (e.g. Wegener 1924, Carey 1958, Wilson 1963, 1965, Bullard et al. 1965, Keen et al. 1972, Kristoffersen & Talwani 1977, Newman 1977, Sclater et al. 1977, Newman & Falconer 1978) or not (e.g. Kerr 1967a, b, 1980a, b, 1981, Grant 1975, Le Pichon et al. 1977).

Proponents of great displacement along Nares Strait have pointed out some of the difficulties of assessing geological features across it. The main difficulty is that geological units tend to be deposited as near horizontal sheets and only near vertical changes can be used for *precise* measurement of offset. The situation is complicated where stratigraphic and structural trends cross Nares Strait at a very oblique angle (Trettin et al. 1972). Features which are most suited to successfully show displacement or lack of it are vertical features, but these are rare. However, recent increase in activity in high arctic regions now makes it possible to confidently cor-

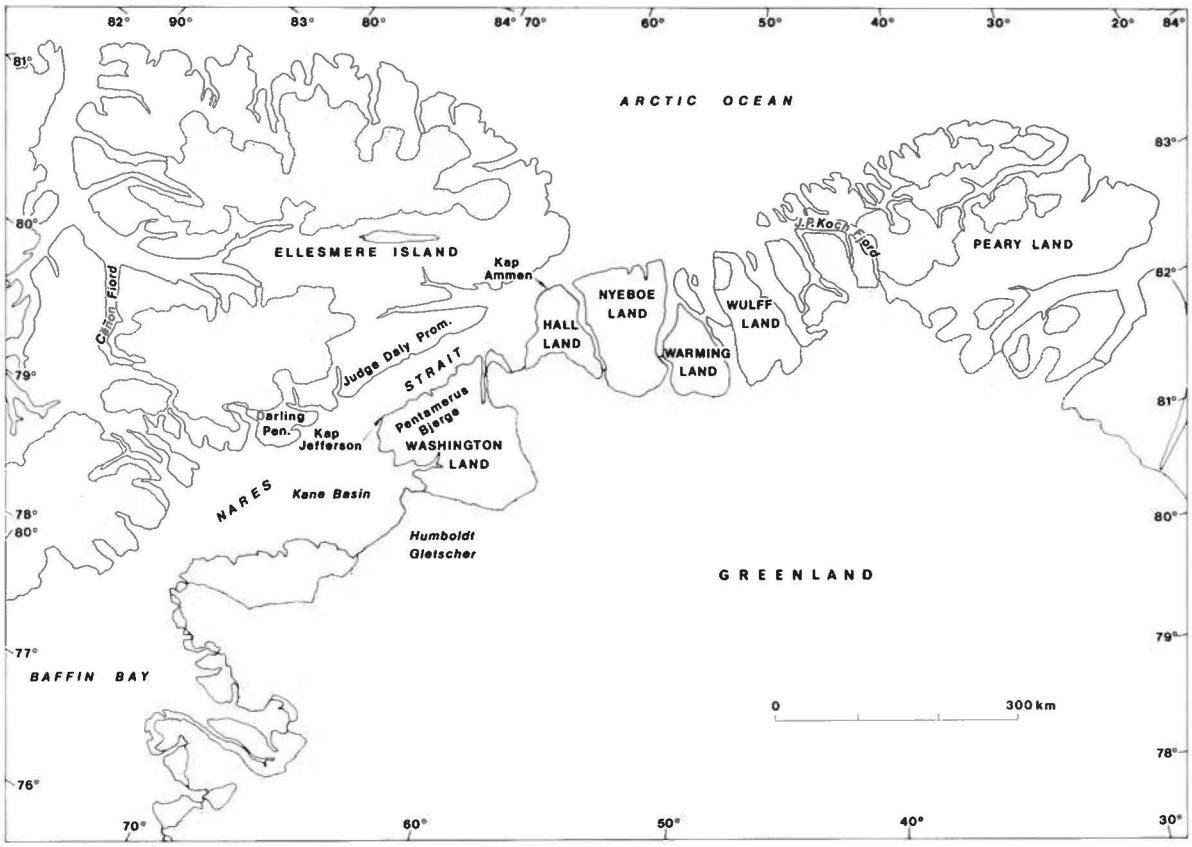


Fig. 1. The Nares Strait region showing the main geographic features referred to in the text. The main ice limits are illustrated.

relate Palaeozoic facies and stratigraphies on both sides of Nares Strait and to assess their bearing on later Cretaceous–Tertiary movement along the Strait.

This paper is concerned specifically with upper Ordovician to Silurian facies patterns and stratigraphic histories in the Nares Strait region. This time interval is of particular importance because linear facies fronts, resulting from rapid facies changes can demonstrate offset or lack of offset along Nares Strait. This paper documents and analyses what these facies changes are and shows how they relate to solving the dilemma. Facies changes between platform carbonates and trough clastics produce marker boundaries (often vertical) that are traceable with varying degrees of accuracy from one side of Nares Strait to the other. The situation in the time interval treated here differs from that in many other intervals, where the sediments are more uniform over large areas and with few internal vertical markers that can demonstrate precisely offset or lack of offset along Nares Strait.

Regional setting

The lower Palaeozoic histories of North Greenland and Ellesmere Island are very similar and closely related. In general terms this similarity is exemplified by the upper Ordovician facies patterns (Fig. 2). Throughout the Palaeozoic a sequence of limestones, dolomites and evaporites developed on a platform which bounded the northern edge of the North American craton (Kerr 1967a, 1976, Dawes 1971, 1976). In western North Greenland the platform carbonates terminated in the upper Llandovery to Ludlow interval (Hurst 1979, 1980a, b, Surlyk et al. 1980, Fig. 3), whilst in eastern Ellesmere Island they continue through into the lower Devonian (Kerr 1976, Trettin 1979, column 4 in Fig. 3). To the north and west of the platform (Fig. 2) the carbonates passed abruptly into mudstones, resedimented limestone and chert conglomerates and turbidites of a slope and trough sequence (Fig. 3).

The platform margin was not a static feature. During the Palaeozoic the area of trough sediments expanded in Ellesmere Island and Greenland (Dawes 1976, Trettin & Balkwill 1979) but the tectonic mechanisms promoting this were not explained. In Greenland Hurst

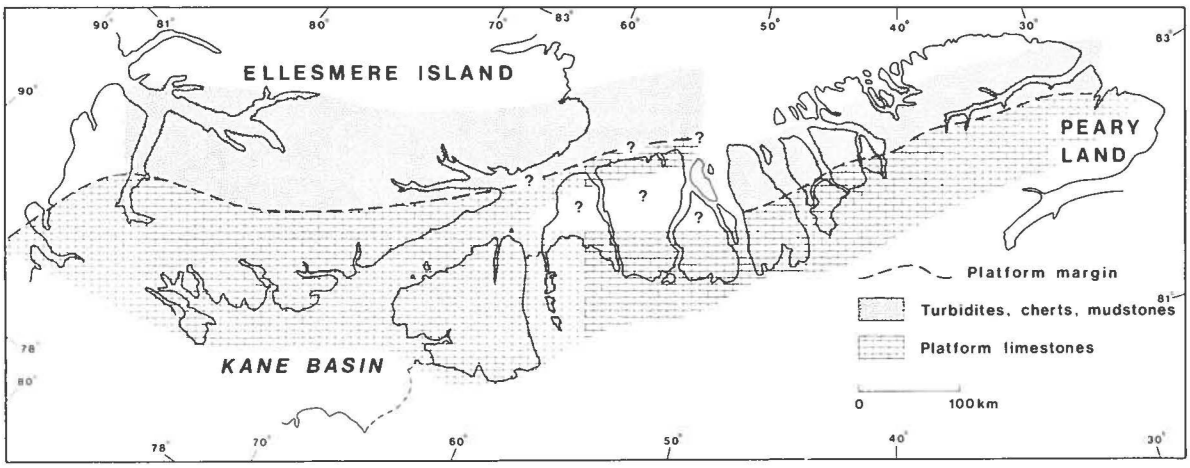


Fig. 2. Regional setting of upper Ordovician (Eden–Richmond) sediments in eastern Ellesmere Island and North Greenland, immediately prior to situation in Fig. 4.

(1980a), Hurst & Surlyk (1980) and Surlyk et al. (1980) relate it to repeated platform margin collapse with resultant drowning of platform carbonates by trough clastics.

During the upper Ordovician the platform margin trended roughly northeast–southwest in both Ellesmere Island and North Greenland (Fig. 2). Repeated collapsing and drowning of the platform in the upper Ordovician to middle Silurian in western North Greenland (Hurst 1980a, b) and eastern Ellesmere Island produced some geological features which can be traced accurately across Nares Strait.

Geological features across Nares Strait

During the upper Ordovician and lower Silurian a marked change occurred in facies patterns adjacent to Nares Strait (Figs 3 and 4, Kerr 1967a, b, Trettin & Balkwill 1979, Hurst 1980a) due to platform margin collapse and platform subsidence. The mechanisms responsible for the partial and differential subsidence of the platform and platform margin are not yet clearly understood. Nevertheless, during the upper Ordovician and lower Silurian the southern platform of Ellesmere Island, bordering the turbidite trough, broke down into three sub-parallel and interfingering facies belts, which proceeding from north to south onto the North American craton (Fig. 4) are:

1) linear intrabasinal horst; at approximately the site of the edge of platform carbonate deposition.
2) mudstones, cherts and resedimented conglomerates.
3) platform carbonates.

These facies are distinctive and traceable with varying degrees of accuracy across Nares Strait into western North Greenland (Figs 3 and 4). The linear intrabasinal horst (the site of the pre-Silurian platform margin, see Dawes, this volume) and the facies front between mudstone and platform carbonates are important vertical markers traceable across Nares Strait.

Linear intrabasinal horst

The abrupt change from platform carbonates along the horst to trough clastic facies is well documented in eastern Ellesmere Island (Kerr 1976, Trettin & Balkwill 1979, Figs 3 and 4). The horst forms a linear feature from Cañon Fiord to northern Judge Daly Promontory, across Nares Strait to the northern coast of Hall Land and possibly to western Nyeboe Land, North Greenland.

Limited graptolite evidence from northern Warming Land (Fig. 1) indicates that mudstones were deposited contemporaneously with and south of the northern Hall Land (and Nyeboe Land?) carbonates. These limestones were apparently isolated from the southerly platform sequences as in Ellesmere Island, and represent the easterly extension of the linear horst (Fig. 4). Thus, the spatial distribution of the lower Silurian facies in western North Greenland is extremely similar to that in eastern Ellesmere Island.

The isolated belt of platform carbonates in Ellesmere Island and Hall Land is a narrow and remarkably linear feature. It is very similar to the platform margin in Peary Land, which is the site of a fault zone (Hurst & Surlyk 1980, Surlyk et al. 1980). The northwest edge of this carbonate belt of Ellesmere Island is directly in line with the northern coast of Hall Land and Nyeboe Land

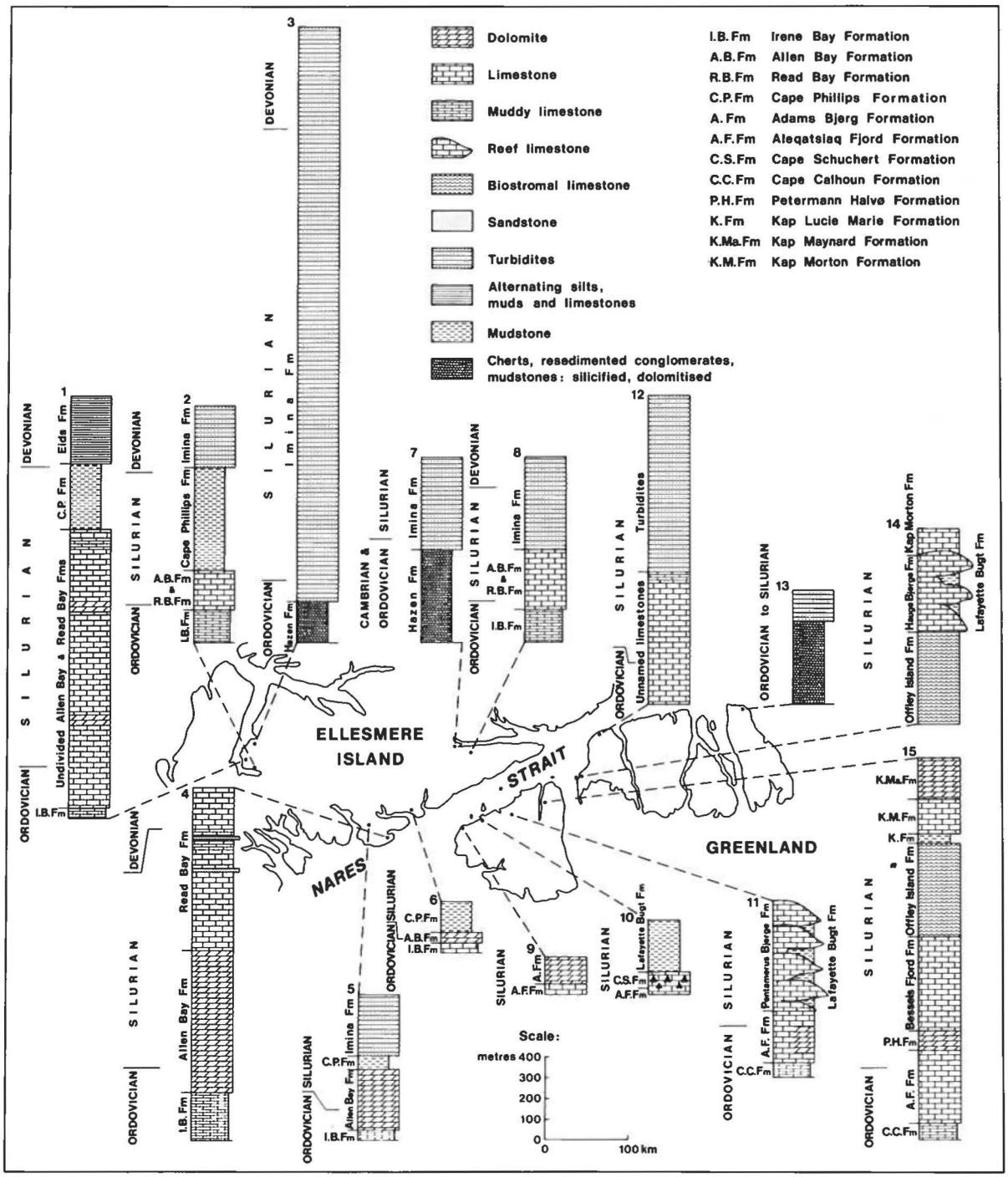


Fig. 3. Upper Ordovician to Devonian stratigraphic columns, adjacent to the northern part of Nares Strait. Sections compiled from Kerr (1976), Trettin (1979), Trettin et al. (1979), Hurst (1980b), personal observations and U. Mayr (pers. comm.).

(Figs 1 and 4). The straightness implies some form of fault control (see also Dawes, this volume).

Further evidence suggesting that the coastline of northern Hall Land is the extension of the Judge Daly Promontory platform margin and horst can be gained from the nature of the margin itself. From the aerial photographic mapping of Dawes (1976) some carbonate buildups (reefs) were thought to be associated with

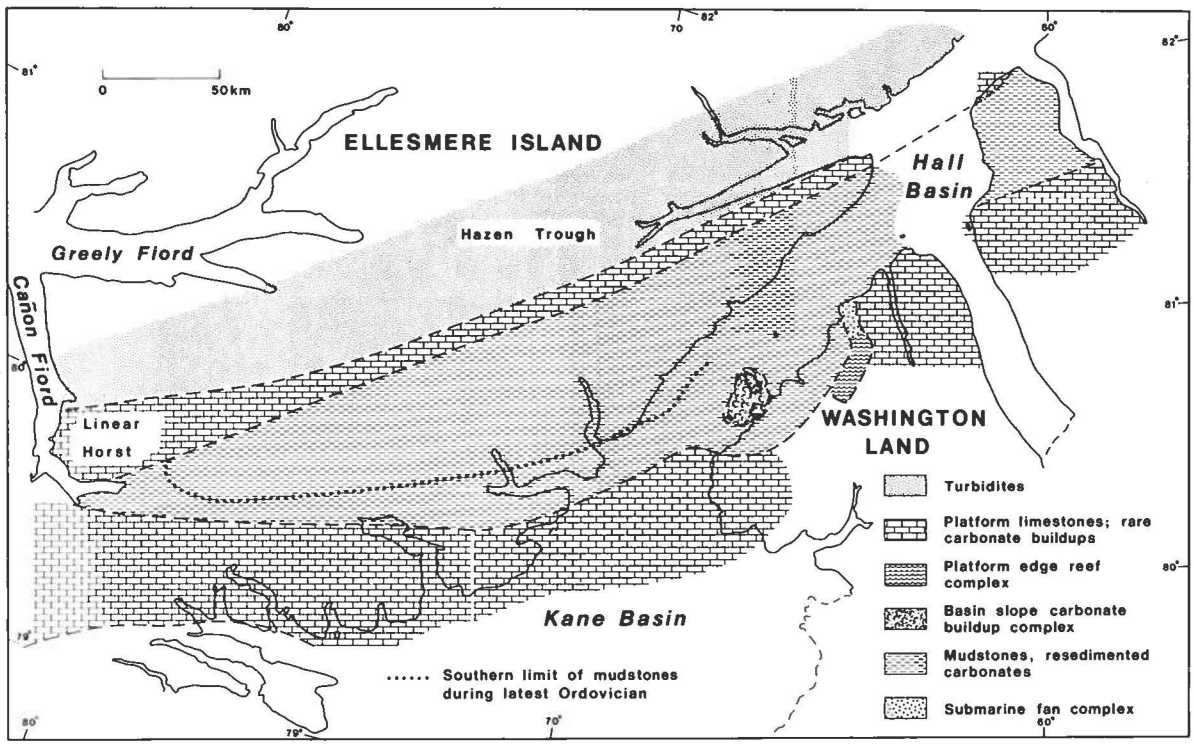


Fig. 4. Facies belts adjacent to the northern part of Nares Strait during the lower Silurian (middle Llandovery and later). The southern boundary between carbonates and mudstones is precisely located in Washington Land. In Ellesmere Island the facies front is not as precise and may be up to 10 km north or south of the line shown. Either way does not disturb the general facies trend across Nares Strait.

the transition from platform to trough. From the work of Hurst (1980a) some of the carbonate buildups at the platform margin are now known to be faulted, sheared and slumped carbonate masses which weather into smooth glaciated banks, simulating true carbonate buildups. These are restricted to the platform margin (Fig. 5A, D). In northern Hall Land a similar carbonate mass is known at the coastline (Fig. 5A) suggesting the edge of the platform margin.

In summary, the northwest edge of the platform margin with the linear horst is an important marker across Nares Strait and it appears to have little or no offset.

Middle Llandovery facies boundaries

During latest Ordovician time (Richmondian) mudstones succeeded platform carbonates between Darling Peninsula and central to eastern Judge Daly Promontory (Fig. 4). No mudstones of Ordovician age occur in Washington Land or Hall Land (Figs 3 and 4). By middle Llandovery time (lower Silurian) mudstones interdigitate with platform carbonates along a line from Cañon Fiord to Darling Peninsula in eastern Ellesmere Island (Fig. 4, Kerr 1976). Similarly in western Washington Land cherty lime mudstones, cherts and calcarenites followed by a thick mudstone, chert and resedimented conglomerate unit, all of middle Llandovery and younger age, interdigitate with platform carbonates and platform edge reefs (Figs 3, 4 and 5B, C, Hurst 1980a, b).

The progressive diachronous spread of mudstones southeastwards in eastern Ellesmere Island and eventually into western Washington Land indicates that both areas were part of the same depositional regime (Fig. 6). Subsequently, turbidites spread south in eastern Ellesmere Island (Kerr 1976). A thick turbidite succession is not known from Washington Land but according to Koch (1920, 1929) some sands do occur in the highest part of the mudstone sequence. This again emphasises the facies continuity between eastern Ellesmere Island and western North Greenland (Fig. 6).

The facies front (Figs 5B, C) between mudstones and platform carbonates had a stable geographic position from middle Llandovery to Devonian time in Ellesmere Island (Kerr 1976). The present-day alignment of the facies boundary across Nares Strait involves only a slight kink which is entirely consistent with primary depositional characteristics (Fig. 4).

Isopachs are further evidence that the mudstone formed in a single discrete basin (Fig. 7). The true sedimentary thickness of the mudstones in Washington Land is not possible to determine because the upper

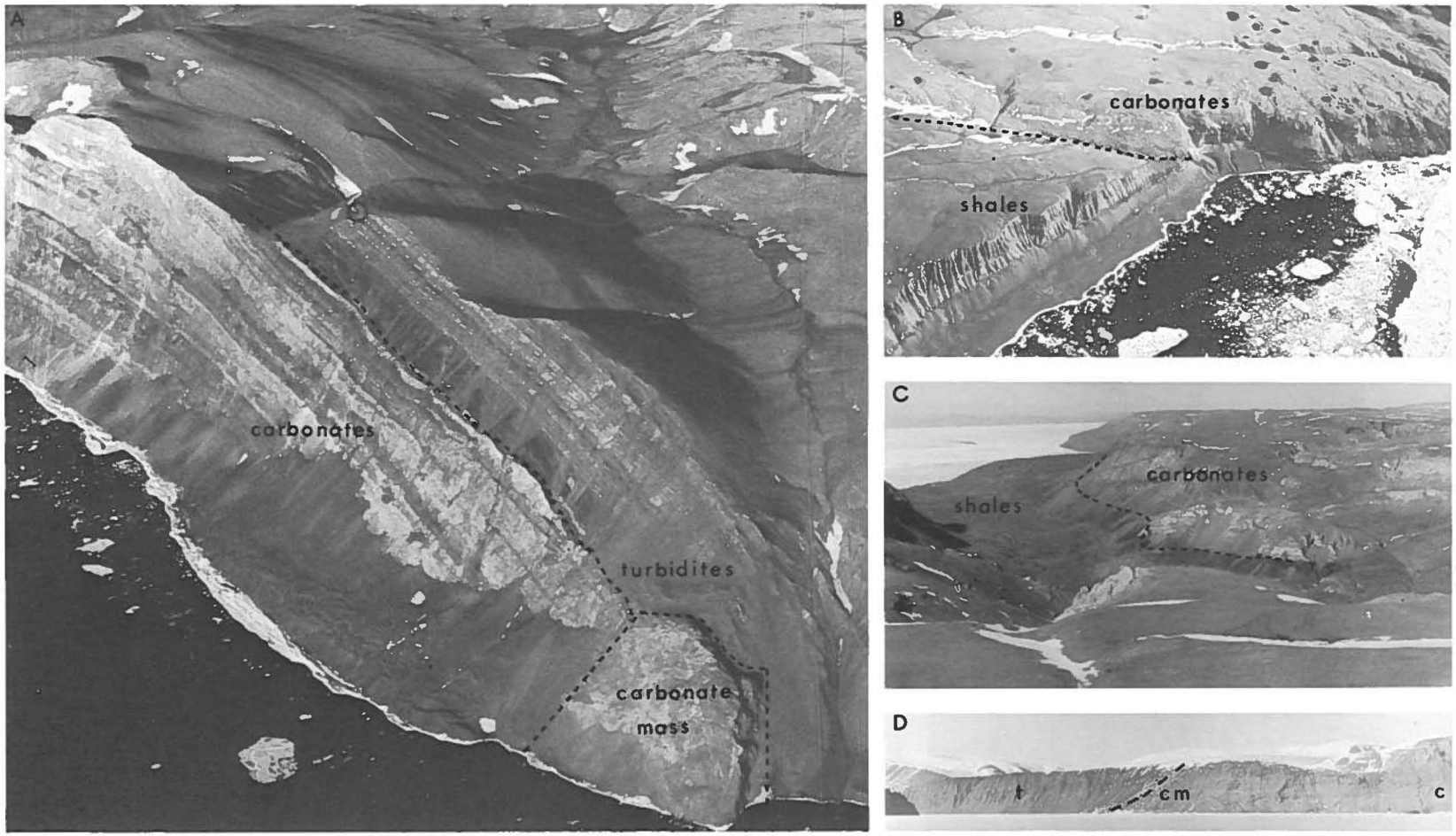


Fig. 5. A. Northern coastline of Hall Land west of Kap Ammen, Greenland, showing upper Ordovician to lower Silurian carbonates overlain by turbidites. Note the large carbonate mass which simulates a platform margin carbonate buildup and resembles that in Fig. 5D. Aerial photograph 546 K–SØ, no. 2199; copyright Geodætisk Institut, Denmark.
B. Interdigitation of platform carbonates with shales of Llandovery age at Kap Jefferson, Washington Land, Greenland. Aerial photograph 545 H–S, no. 11781; copyright Geodætisk Institut, Denmark.
C. Interdigitation of platform carbonates with marginal carbonate buildups with shales in the Pentamerus Bjerge of Washington Land, Greenland. Llandovery to Wenlock age.
D. Cambrian to Silurian platform margin in J. P. Koch Fjord, Peary Land, Greenland. Note development of a simulated carbonate buildup (cm) at the edge of the carbonate platform (c) and turbidite (t) trough. Note that the platform margin and facies interdigitation are vertical or near vertical features.

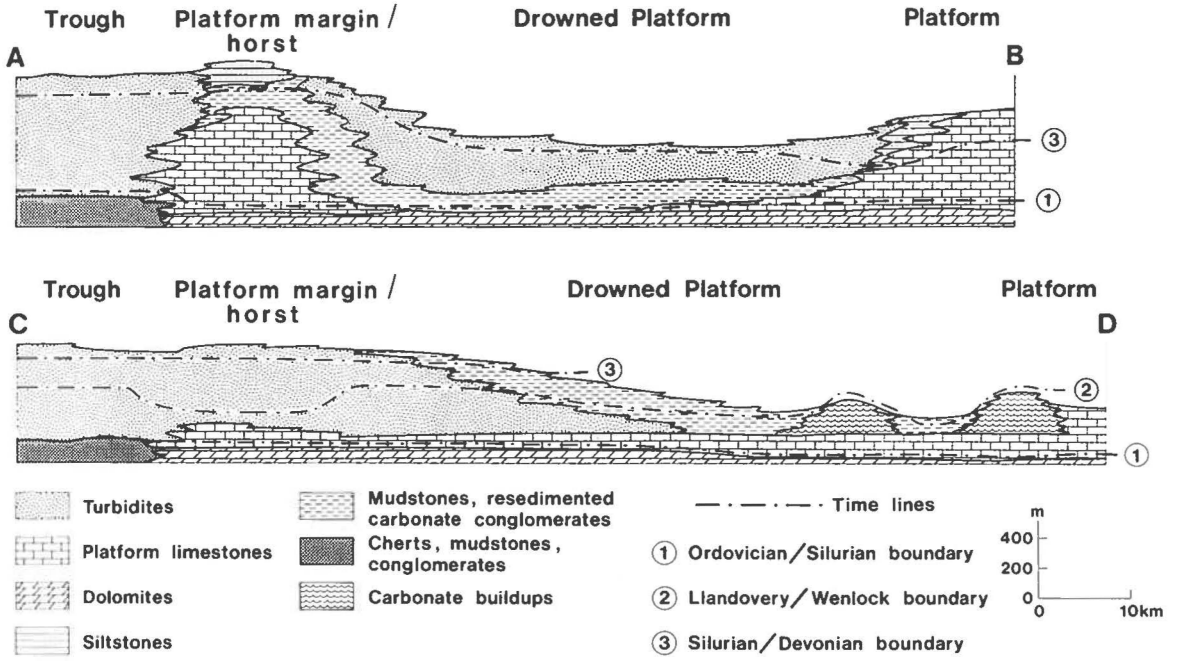


Fig. 6. Schematic cross-section of upper Ordovician to Silurian platform trough facies. Line of sections shown on Fig. 7. Note similarity of sections emphasising continuity of sedimentary regimes across Nares Strait. The line A–B is composite with the northerly parts based on the sections in Cañon Fiord (see Trettin 1979).

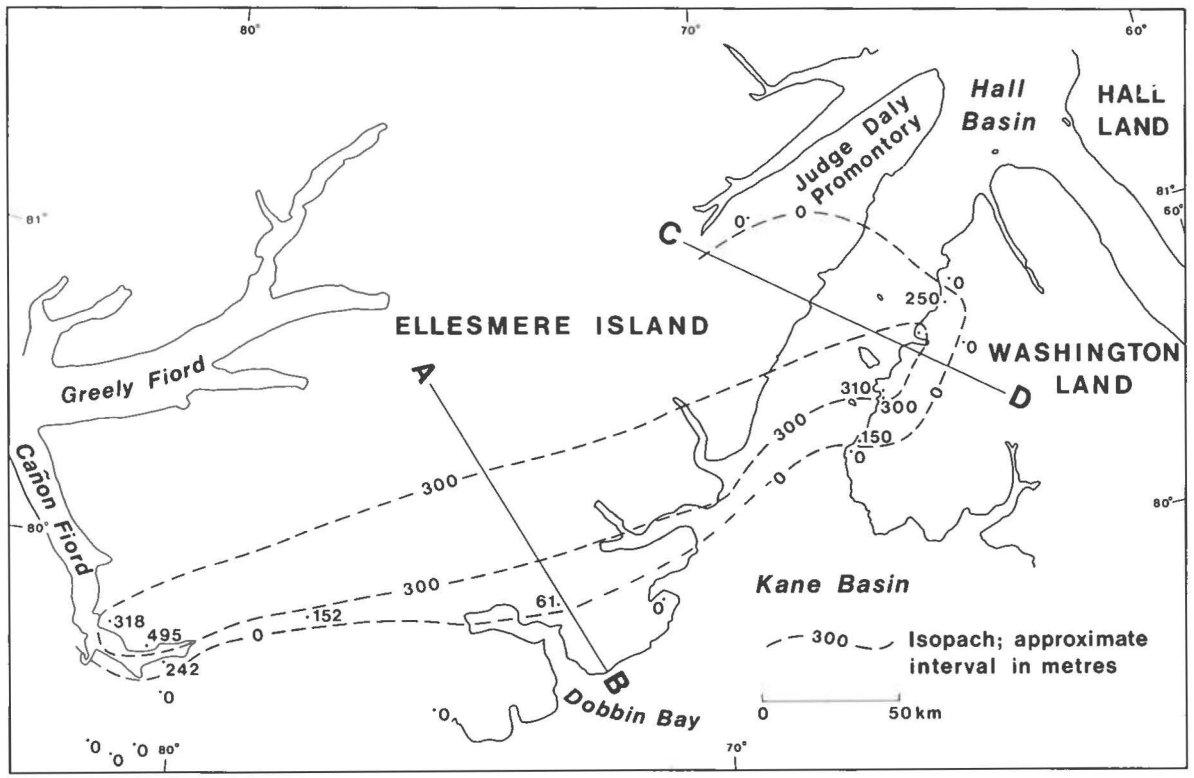


Fig. 7. Isopach map indicating thicknesses (in metres) of the upper Ordovician to Devonian mudstone (shale) facies of eastern Ellesmere Island and Washington Land, Greenland. Lines A–B, C–D locate the cross-sections shown in Fig. 6.

surface is eroded. Nevertheless, the minimum thicknesses indicate that the axis of mudstone deposition trends from Cañon Fiord in Ellesmere Island into Washington Land.

Conclusions and constraints

The upper Ordovician and Silurian successions of eastern Ellesmere Island and western North Greenland show two geological features which can be used with precision to correlate across Nares Strait. These are 1) the long persisting platform margin with linear horst that persisted during the upper Ordovician and lower Silurian, and which bounded the southern margin of the trough turbidites, and 2) the facies interdigitation between the platform carbonates (with marginal carbonate buildups in Washington Land) and mudstones from the middle Llandovery (lower Silurian) to Devonian.

These features together with the present-day distribution of upper Ordovician and Silurian sediments demonstrate clearly that suggestions of 220 km or more of left-lateral displacement along Nares Strait are highly improbable. Assuming left-lateral displacement between 100 and 220 km there is a great mismatch of the geological features, and displacements of this amount can be discounted. Because of the uncertainty of the precise original shape of facies boundaries it is not possible to be absolutely accurate about displacement. The two markers discussed here argue strongly that displacement was no more than 25 km.

In conclusion the present-day distribution of upper Ordovician and lower Silurian sediments and marker horizons in eastern Ellesmere Island and western North Greenland is consistent with no strike-slip displacement along Nares Strait since Silurian time.

Acknowledgements

We are grateful to F. Surlyk (GGU) and H. P. Trettin (GSC) for critically reading the manuscript. JMH publishes with the permission of the Director of the Geological Survey of Greenland. The aerial photographs are published with the permission (A.495/79) of the Geodætisk Institut, Copenhagen, Denmark.

References

Bullard, E., Everett, J. E. & Smith, A. G. 1965. The fit of the continents around the Atlantic. – In: Blackett, P. M. S., Bullard, E. & Runcorn, S. K., A symposium on continental drift. – Phil. Trans. Roy. Soc. Lond. 258A: 41–51.

Carey, S. W. 1958. The tectonic approach to continental drift. – In: Carey, S. W. (convener), Continental drift. A symposium: 177–355. – Univ. Tasmania, Hobart.

Dawes, P. R. 1971. The North Greenland fold belt and environs. – Bull. geol. Soc. Denmark 20: 197–239.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R. 1982. The Nyeboe Land fault zone: a major dislocation on the Greenland coast along northern Nares Strait. – This volume.

Grant, A. C. 1975. Geophysical results from the continental margin off southern Baffin Island. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 7: 411–431.

Hurst, J. M. 1979. Uppermost Ordovician and Silurian geology of north-west Peary Land, North Greenland. – Rapp. Grønlands geol. Unders. 88: 41–49.

Hurst, J. M. 1980a. Paleogeographic and stratigraphic differentiation of Silurian carbonate buildups and biostromes of North Greenland. – Bull. Am. Ass. Petrol. Geol. 64: 527–548.

Hurst, J. M. 1980b. Silurian stratigraphy and facies distribution in Washington Land and western Hall Land, North Greenland. – Bull. Grønlands geol. Unders. 134: 95 pp.

Hurst, J. M. & Surlyk, F. 1980. Notes on the Lower Palaeozoic clastic sediments of Peary Land, North Greenland. – Rapp. Grønlands geol. Unders. 99: 73–78.

Keen, C. E., Barrett, D. L., Manchester, K. S. & Ross, D. I. 1972. Geophysical studies in Baffin Bay and some tectonic implications. – Can. J. Earth Sci. 9: 239–256.

Kerr, J. W. 1967a. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Kerr, J. W. 1967b. A submerged continental remnant beneath the Labrador Sea. – Earth planet. Sci. Lett. 2: 283–289.

Kerr, J. W. 1976. Stratigraphy of central and eastern Ellesmere Island, Arctic Canada. Part III. Upper Ordovician (Richmondian), Silurian and Devonian. – Bull. geol. Surv. Can. 260: 55 pp.

Kerr, J. W. 1980a. Did Greenland drift along Nares Strait? – Bull. Can. Petrol. Geol. 28: 279–289.

Kerr, J. W. 1980b. A plate tectonic contest in Arctic Canada. – In: Strangway, D. W. (ed.), The continental crust and its mineral deposits. – Spec. Pap. geol. Ass. Can. 20: 457–486.

Kerr, J. W. 1981. Evolution of the Canadian Arctic Islands: a transition between the Atlantic and Arctic Oceans. – In: Nairn, A. E. M., Churkin, M. & Stehli F. G. (eds.), The ocean basins and margins 5, The Arctic Ocean: 105–199. Plenum Press, New York & London.

Koch, L. 1920. Stratigraphy of Northwest Greenland. – Meddr dansk geol. Foren. 5(17): 78 pp.

Koch, L. 1929. Stratigraphy of Greenland. – Meddr Grønland 73,2(2): 205–320.

Kristoffersen, Y. & Talwani, M. 1977. Extinct triple junction south of Greenland and the Tertiary motion of Greenland relative to North America. – Bull. geol. Soc. Am. 88: 1037–1049.

Le Pichon, X., Sibuet, J.-C. & Francheteau, J. 1977. The fit of the continents around the North Atlantic Ocean. – Tectonophysics 38: 169–209.

Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Newman P. H. & Falconer, R. K. H. 1978. Evidence for movement between Greenland and Canada along Nares Strait. – Geol. Soc. Am. Abs with Prog. 10: 463 only.

Sclater, J. G., Hellinger, S. & Tapscott, C. 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. – J. Geol. 85: 509–552.

Surlyk, F., Hurst, J. M. & Bjerreskov, M. 1980. First age-diagnostic fossils from the central part of the North Greenland foldbelt. – Nature, Lond. 286: 800–803.

Trettin, H. P. 1979. Middle Ordovician to Lower Devonian deep-water succession at southeastern margin of Hazen Trough, Cañon Fiord, Ellesmere Island. – Bull. geol. Surv. Can. 272: 84 pp.

Trettin, H. P. & Balkwill, H. R. 1979. Contributions to the tectonic history of the Innuitian Province, Arctic Canada. – Can. J. Earth. Sci. 16: 748–769.

Trettin, H. P., Frisch, T. O., Sobczak, L. W., Weber, J. R., Niblett, E. R., Law, L. K., DeLaurier, J. M. & Whitlam, K. 1972. The Innuitian Province. – In: Price, R. A. & Douglas, R. J. W. (eds.), Variations in tectonic styles in Canada. – Spec. Pap. geol. Ass. Can. 11: 83–179.

Trettin, H. P., Barnes, C. R., Kerr, J. W., Norford, B. S., Pedder, A. E. H., Riva, J., Tipnis, R. S. & Uyeno, T. T. 1979. Progress in lower Paleozoic stratigraphy, northern Ellesmere Island, District of Franklin. – Pap. geol. Surv. Can. 79–1B: 269–279.

Wegener, A. 1924. The origin of continents and oceans. Methuen & Co., London: 212 pp. (Transl. 3rd edit. by J. G. A. Skerl).

Wilson, J. T. 1963. Continental drift. – Scient. Am. 208(4): 86–100.

Wilson, J. T. 1965. A new class of faults and their bearing on continental drift. – Nature, Lond. 207: 343–347.

Nares Strait and the down-current termination of the Silurian turbidite basin of North Greenland

FINN SURLYK

Surlyk, F. 1982. Nares Strait and the down-current termination of the Silurian turbidite basin of North Greenland. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 147–150.

Combined palaeoenvironmental and palaeocurrent evidence demonstrates that a significant change took place in large-scale Silurian palaeogeographic configuration along the North Greenland – Canadian continental margin. The source area of the deep-water turbidite basin of North Greenland was the rising Caledonides to the east. The currents flowed towards the west parallel to the southern carbonate platform. There is no evidence of a northern landmass and the abyssal plains off the present-day east coast of the United States are suggested as analogues. The axial flows seem to have continued westwards along the axis of the Ellesmere Island turbidite basin. West of Nares Strait a significant inflow of turbidity currents from a northerly shelf area enters the basin where they were deflected down the axis, and the existence of a northern landmass, possibly of volcanic arc origin, has been demonstrated in Ellesmere Island. The Ellesmere Island turbidite basin thus represents a deeply submerged back-arc rift or perhaps an intracontinental aulacogen. There is no evidence from the Silurian turbidites of any contemporaneous or later wrench faulting along Nares Strait. Rather, Nares Strait represents the transitional area in the Silurian between two different types of continental margins or even greater geotectonic differences. Future studies of this difference may be instrumental in unravelling the later history of this enigmatic segment of the Earth's crust.

F. Surlyk, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark.

The Silurian turbidite basin of North Greenland is one of the world's largest ancient 'longitudinal flysch basins'. It has a preserved length of about 800 km, a width of about 200 km and a thickness of up to 4 km. Until GGU's expeditions in 1979 and 1980 it was, however, relatively little known due to its remote and inaccessible position, scarcity of biostratigraphic data, and tectonic disturbances, especially in the northern part. In 1979 a large number of representative sections covering the whole basin were measured in detail by the author and J. M. Hurst. The palaeocurrent pattern appeared to be extremely uniform from east to west, following the basinal axis (Fig. 1). The basin continues south-westwards into Ellesmere Island, Arctic Canada, where a SE-directed palaeocurrent direction was introduced in addition to the axial SW direction. This indication of a north-westerly provenance is supported by the presence of a Silurian shelf area in northernmost Ellesmere Island. The change in provenance and palaeocurrent pattern took place in the Nares Strait region. The present paper aims to demonstrate this in more detail and to discuss the possible geotectonic implications.

Geological setting

The Lower Palaeozoic rocks of North Greenland can be divided into a southern carbonate platform and a northern deep-water clastic basin. This division is particularly well displayed by the Silurian section. The platform to basin transition seems to be controlled by major east–west-trending fault systems. To the east it is developed like a scarp (Surlyk et al. 1980) and to the west like a ramp (Hurst 1980). The outer northern part of the platform underwent a major collapse phase in the early Silurian and the turbidite basin expanded several hundred kilometres to the south (Hurst & Surlyk 1980). Thus, mid- and late Silurian turbiditic sediments overlie shallow marine carbonates in a broad belt along the southern margin of the basin. The northern limit of the basin is unknown since the turbiditic facies extend to the coast.

The distribution of the main facies belts is roughly the same in Ellesmere Island, although the configuration of the basin margin is less clear as a result of much stronger tectonic disturbances. In contrast to Greenland a north-

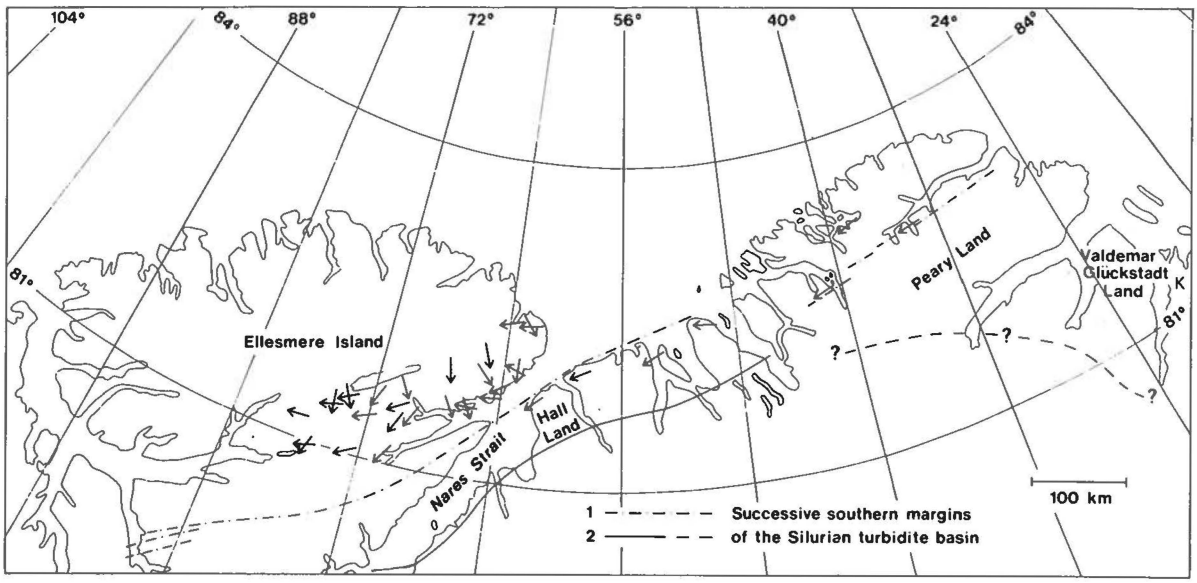


Fig. 1. Palaeocurrent map of the Silurian turbidite basin of Ellesmere Island (data from Trettin 1979) and North Greenland (based on 340 measurements made by the author and J. M. Hurst in 1979). K = western coast of Kronprins Christian Land. The two successive margin positions of the Silurian turbidite basin correspond to the two geological markers discussed in this volume by Hurst & Kerr.

ern platform and shelf sequence is preserved in northernmost Ellesmere Island.

Sedimentary environments of the Silurian turbidite basin

Silurian turbidite deposition followed a long period of basin starvation with deposition of mainly black mudstones and cherts, which lasted through most of the Ordovician. Three stages of Silurian turbidite sedimentation can be recognised. The basal sandy unit is of mainly Llandovery age. It is followed by a muddy unit of latest Llandovery – early Wenlock age, which is topped with marked diachroneity by the final sandy latest Llandovery – Ludlow turbidite unit.

Llandovery turbidites

The initial phase of turbidite sedimentation can only be demonstrated to the east in the Peary Land area. Farther to the west the Silurian platform–basin transition lay north of the present coastline (Fig. 1). The sedimentary faices are mainly medium-bedded rather fine-grained sandy turbidites, which conform to the Bouma model (Bouma 1962). The sand–shale ratio is high and bed geometry is sheet-like apart from scours and channels in the upper part of the unit. Facies analysis by this author and J. M. Hurst (in prep.) demonstrates the presence of basin plain, outer fan fringe and lobe and, towards the top, braided midfan. Transport is towards the west throughout the unit and the overall setting is one of an extremely elongate sand-rich deep-sea fan, imperceptibly passing into a basin plain.

Latest Llandovery – Wenlock mudstones

The catastrophic collapse of the platform margin led to basin expansion, transgression, and starvation of the fan and turbidite basin system. As a result, a sequence of mudstones was deposited on top of the lower turbidite sequence and on the collapsed platform carbonates along the southern basin margin. The mudstone unit is rather thin towards the east but thickens drastically to the west where muddy deposition continued for a long time even when turbidite sedimentation was resumed towards the east in late Llandovery times. Along the axial parts of the basin thin-bedded turbidites occur within the mudstones, while resedimented limestone and chert conglomerates fringe the southern carbonate platform. The depositional environments can be interpreted as a muddy basin plain passing gradually into slope muds along the southern basin margin. Transport data are few, but the quite drastic westward decrease in grain size and increase in thickness of the mudstone unit again point towards a longitudinal east–west fill of the elongate basin.

Late Llandovery – Ludlow turbidites

Turbidite deposition was resumed in late Llandovery time and the basin reached its widest extent from Kron-

prins Christian Land in the east to Hall Land in the west and from the north coast to Valdemar Glückstadt Land in the south. The southern limit is, however, purely erosional and the basinal facies may have had an even wider extent.

The sedimentary facies include mainly medium- and thin-bedded turbidites, which can be described by the Bouma model. To the west the proportion of muddy units increases, while resedimented chert conglomerates dominate the upper parts of the sequence to the east.

The sedimentary environments include outer fan fringe and lobe, and braided midfan towards the top in the eastern part of the basin. The main western part of the unit was, however, deposited in a system of huge deep-sea channels, levées and intervening basin plains. The transport was as in the two preceding units towards the west along the axis of the basin.

Environments and transport patterns in Ellesmere Island

Knowledge of the depositional environments of the Silurian turbidites in north-eastern Ellesmere Island is mainly based on provenance and transport direction studies (Trettin 1970, 1971, 1973, 1979). Trettin demonstrated that the turbidite palaeocurrents could be divided into a south-easterly and a south-westerly population. The first was interpreted as reflecting deposition in a canyon deep-sea fan system developed south-east of the north-western shelf. The turbidity currents which reached the fan toe were deflected towards the south-west where they flowed down the basin axis parallel to the southern platform margin. This interpretation is accepted here with the modification that parts of the axial component probably represent the direct down-current westward continuation of the axial flows of the North Greenland basin.

Discussion

Combined palaeocurrent and palaeoenvironmental evidence demonstrates that a significant change in large-scale palaeogeographic configuration of the Silurian North Greenland – Canadian continental margin took place in the Nares Strait area.

The source area of the North Greenland part of the basin was located in the northern end of the Atlantic Caledonian mountain belt. The Silurian orogenic phases obviously resulted in uplift of enormous landmasses which became the subject of erosion ultimately resulting in deposition of an up to 4 km thick, 800 km long and several hundred kilometres wide basin fill. There is no direct evidence of a northern borderland. The very clearly defined east to west palaeocurrent pattern indicates, however, the existence of a northern barrier, which assisted in funnelling the turbidity currents down an east–west axis. The absence of any southward directed palaeocurrents suggests that the North Greenland basin was a normal continental margin deep-sea plain, perhaps directly analogous to the present-day abyssal plains occurring off the east coast of the United States (Pilkey et al. 1980). These deep basins are fed by turbidity currents entering via submarine canyons at one end. Some of the currents are deflected when they bypass the fan toes and flow along much of the length of the abyssal plains parallel to the continental margin.

In contrast, the Ellesmere Island part of the basin is limited by shelf areas towards the north-west and the south-east. The geotectonic nature of the northern landmass and source area is not known, although it is illustrated as a volcanic arc by Trettin & Balkwill (1979: fig. 8). The basin, as illustrated, was probably formed by back-arc spreading. It could alternatively be interpreted as a linear intra-continental feature formed by rifting conforming to the aulacogen model.

The Silurian North Greenland – Canadian turbidite basin thus seems to change from a simple continental margin deep-sea plain to a deeply submerged back-arc rift or aulacogen in a down-current direction in the Nares Strait region. The axial transport appears to have continued uninterruptedly, while a very important northern source area came in west of Nares Strait. Thus, Nares Strait seems not to have been the locus of any significant wrench faulting during or after the Silurian. Rather, it represents the transitional area between two different types of continental margins or possibly an even greater geotectonic difference in the Silurian. The description of the precise nature of this difference may be instrumental in unravelling the later history of this segment of the Earth's crust.

Acknowledgements

The field work that forms the basis of the present study was carried out in company with J. M. Hurst. E. Glendal and B. Thomasen are thanked for technical assistance in the preparation of the manuscript. The paper is published with permission of the Director of the Geological Survey of Greenland.

References

Bouma, A. H. 1962. Sedimentology of some flysch deposits. A graphic approach to facies interpretation. – Elsevier Publ. Co., Amsterdam: 168 pp.

Hurst, J. M. 1980. Paleogeographic and stratigraphic differentiation of Silurian carbonate buildups and biostromes of North Greenland. – Bull. Am. Ass. Petrol. Geol. 64: 527–548.

Hurst, J. M. & Kerr, J. W. 1982. Upper Ordovician to Silurian facies patterns in eastern Ellesmere Island and western North Greenland and their bearing on the Nares Strait lineament. – This volume.

Hurst, J. M. & Surlyk, F. 1980. Notes on the Lower Palaeozoic

clastic sediments of Peary Land, North Greenland. – Rapp. Grønlands geol. Unders. 99: 73–78.
Pilkey, O. H., Locker, S. D. & Cleary, W. J. 1980. Comparison of sand-layer geometry on flat floors of 10 modern depositional basins. – Bull. Am. Ass. Petrol. Geol. 64: 841–856.
Surlyk, F., Hurst, J. M. & Bjerreskov, M. 1980. First age-diagnostic fossils from the central part of the North Greenland foldbelt. – Nature, Lond. 286: 800–803.
Trettin, H. P. 1970. Ordovician–Silurian flysch sedimentation in the axial trough of the Franklinian geosyncline, northeastern Ellesmere Island, Arctic Canada. – In: Lajoie, J. (ed.), Flysch sedimentology in North America. – Spec. Pap. geol. Ass. Can. 7: 13–35.
Trettin, H. P. 1971. Geology of the lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, Arctic Archipelago. – Bull. geol. Surv. Can. 203: 134 pp.
Trettin, H. P. 1973. Early Paleozoic evolution of northern parts of Canadian Arctic Archipelago. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 57–75.
Trettin, H. P. 1979. Middle Ordovician to Lower Devonian deep-water succession at southeastern margin of Hazen Trough, Cañon Fiord, Ellesmere Island. – Bull. geol. Surv. Can. 272: 84 pp.
Trettin, H. P. & Balkwill, H. R. 1979. Contributions to the tectonic history of the Innuitian Province, Arctic Canada. – Can. J. Earth Sci. 16: 748–769.

The Bjørnehiet Formation: a faulted preglacial conglomerate, Washington Land, North Greenland

HANS F. JEPSEN

Jepsen, H. F. 1982. The Bjørnehiet Formation: a faulted preglacial conglomerate, Washington Land, North Greenland. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 151–155.

A new occurrence of a Tertiary or Quaternary fluvial conglomerate from North Greenland is described, and named the Bjørnehiet Formation. The main lithology is a carbonate-clast, matrix-supported, oligomictic conglomerate with some coarse sandstone beds. The composite thickness of the formation is about 45 m.

The conglomerate is known only from one locality in Washington Land, about 40 km from the shore of Nares Strait. The deposit has been affected by post-depositional faulting and is considered to be preglacial in age. By comparisons with Arctic Canada, the conglomerate is regarded as an equivalent of either the Palaeogene Eureka Sound Formation or the Neogene Beaufort Formation, both of which contain important fluvial conglomerate units.

The conglomerate represents the only evidence of post-Palaeozoic, preglacial deposits in this part of Greenland. The occurrence is thus particularly important and provides a way of recording Cenozoic fault movements. If the deposit can be accurately dated, and when its full distribution is known, the formation and the structures affecting it will provide significant data about the Cenozoic palaeogeography and tectonics of the Nares Strait region.

H. F. Jepsen, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark.

In 1977, during the regional geological mapping programme of the Lower Palaeozoic rocks of Washington Land, North Greenland (Fig. 1), a fluvial conglomerate was discovered on the steep slopes of a V-shaped valley eroded into the Ordovician bedrock. The locality was discovered late in the field season and it was unfortunately only briefly examined. Its discovery is mentioned in Peel (1978).

The aim of this paper is to describe the main features of the conglomerate occurrence and to formally designate the strata as the Bjørnehiet Formation. The deposit occurs in the Nares Strait region and it is affected by faulting; thus it is pertinent to the present symposium. However, this paper is written, not as a contribution to the question about strike-slip or other movement along the Strait, but rather to focus attention on a new rock occurrence that, with more field and laboratory work, will provide important data on the Cenozoic depositional and tectonic history of the Nares Strait region.

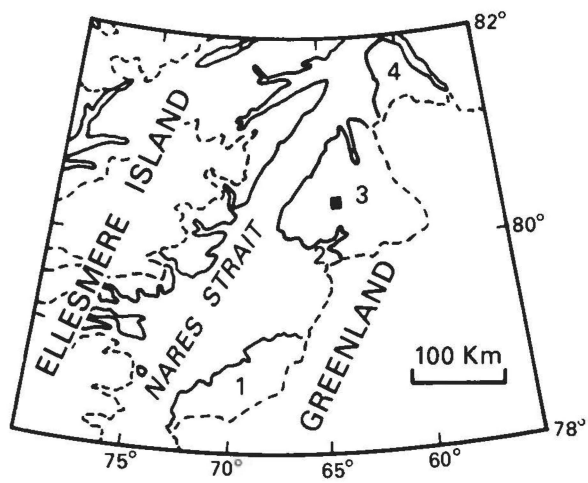


Fig. 1. Index map showing the position of Washington Land in the Nares Strait region. The black square illustrates the area shown in Fig. 2 and the location of the Bjørnehiet Formation. 1 = Inglefield Land, 2 = Cass Fjord, 3 = Washington Land, 4 = Hall Land.

Location and type locality

The Bjørnehiet Formation is located in central Washington Land about 10 km north of the hill called Bjørnehiet, at the head of Cass Fjord (Fig. 1). The formation has only been seen to outcrop along the banks of a westerly and southerly flowing river system, but it is thought to extend over at least 15 km² (Fig. 2). The strata form a flat fluvial plain at about 350 m above sea level in a watershed area between northerly and southerly flowing rivers. The outcrop is bounded by hills composed of Lower Palaeozoic strata that constitute the whole of Washington Land.

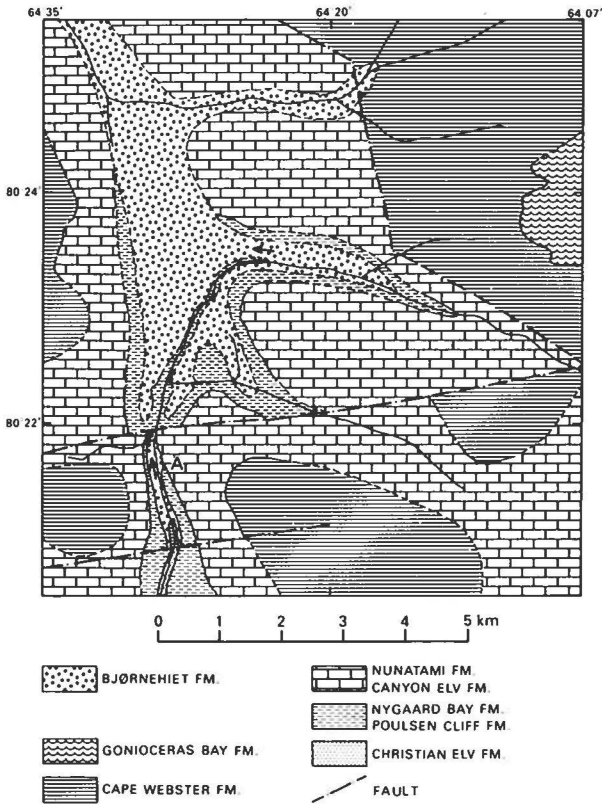


Fig. 2. Geological map showing the setting of the Bjørnehiet Formation. A: location of the type section; the arrows indicate the palaeotransport direction of the formation at two localities.

The type locality of the new formation is situated in a river canyon (marked A in Fig. 2) where the conglomerate outcrops as erosional remnants adjacent to the steep canyon walls (Fig. 3).

Stratigraphical position

The Bjørnehiet Formation rests with erosional unconformity on deeply excavated flat-lying strata belonging to the Ordovician Poulsen Cliff, Nygaard Bay, Canyon Elv and Nunatami Formations that are reported on elsewhere in this volume (Peel & Christie). The formation is covered sporadically by a few metres of non-consolidated glacial outwash and solifluction material of Holocene age.

Thickness and lithology

The Bjørnehiet Formation attains a composite thickness of 44 m in the type section where four members can be recognised (Table 1). Rapid lateral thickness and facies changes are common and it is uncertain if the formation ever reaches the composite thickness in any one outcrop



Fig. 3. Type section locality of the Bjørnehiet Formation seen from the south. The section has been measured along the west side (left) of the canyon. The height of the cliff left of centre is about 25 m.

along the river. The maximum thickness measured in a single outcrop is 25 m.

The formation is typically a brown-weathering, easily eroded, matrix-supported conglomerate with minor sandstone incursions (Figs 3 and 4). Apart from the basal part of the sequence, where metamorphic and granitic rocks (assumed Precambrian) form part of the clast content, the conglomerate is oligomictic, being composed of locally-derived carbonate pebbles, cobbles and boulders of the Lower Palaeozoic bedrock. Clast size varies appreciably and boulders up to 50 cm are common in the lower parts of the sequence. The clasts are predominantly rounded to well-rounded but some, particularly the locally-derived larger blocks, are angular. The matrix is a carbonate-cemented, coarse-grained sandstone and this lithology also forms thin lenses and beds intercalated with the rudaceous material, the largest of which is about 2 m thick in the middle of the sequence (Fig. 5).



Fig. 4. Cross-bedded fluvial conglomerate with some sandstone lenses. The direction of transport in this outcrop is towards the north (right). Part of the lower stratified conglomerate — member 2 of Table 1.

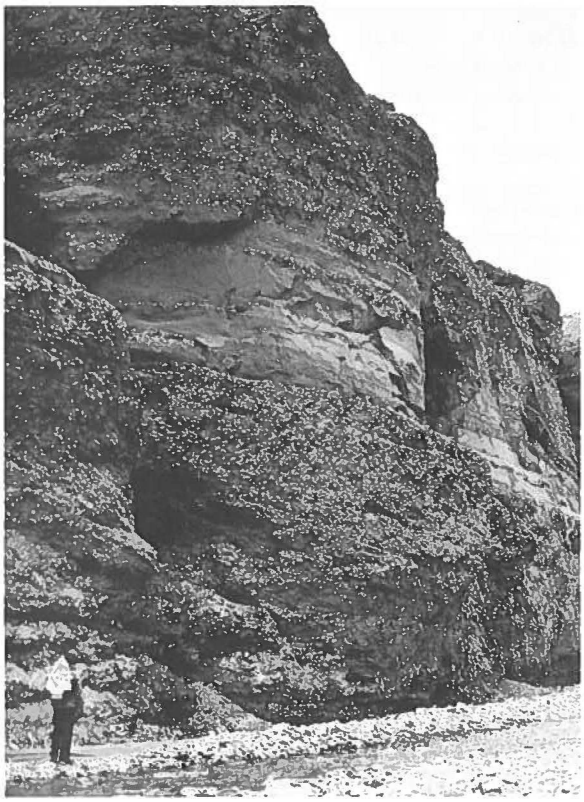


Fig. 5. The central *c.* 25 m of the Bjørnehiet Formation including the 2 m thick coarse-grained sandstone bed — member 3 of Table 1.

Table 1. Type section log of the Bjørnehiet Formation

Member 4, 21 m	Stratified matrix-supported conglomerate with angular and rounded carbonate blocks up to 30 cm in diameter. Large-scale cross-bedding and imbrication of tabular blocks. Carbonate-cemented coarse-grained sandstone matrix.
Member 3, 2 m	Coarse-grained carbonate-cemented sandstone with cross-bedding. Some pebbly sandstone and thin conglomerate horizons.
Member 2, 18 m	Stratified matrix-supported conglomerate with well-rounded carbonate clasts up to 20 cm in diameter. Includes a few coarse-grained dm-thick sandstone lenses. Large-scale tabular cross-bedding. Carbonate-cemented coarse-grained sandstone matrix.
Member 1, 3 m	Unstratified matrix-supported conglomerate with blocks and boulders up to 50 cm in diameter. Ten per cent of the clasts are well-rounded and derived from crystalline basement including quartzo-feldspathic gneiss and coarse-grained amphibolite. The rest of the clasts are angular or subrounded blocks derived from the local Palaeozoic carbonates. Carbonate-cemented coarse-grained sandstone matrix.

The lithology of the formation at the type locality is given in Table 1.

Frequent tabular cross-bedding in the sandstones and imbrication structure in the conglomerates indicate a consistent northward transport direction at the type locality (Fig. 2). This contrasts with the flow direction of the present-day river. About 3 km north-east of the type section the transport direction is towards the west. This suggests a southerly and easterly provenance for the deposit so that the main part of the formation lying to the west and north-west of the outcrops examined may well consist of a north-westerly outwash plain. The presence of crystalline rocks as clasts in the basal part of the formation is also readily explained by this provenance; the known outcrops of Precambrian Shield are to the south (Inglefield Land) and the regional structure suggests that extensive areas of the Shield outcrop under the Inland Ice to the east of Washington Land.

Regional distribution

The conglomerate occurrence in central Washington Land is the only known outcrop of the Bjørnehiet Formation. However, other outcrops may well be discovered in western North Greenland as systematic field work continues. Thus, it is of some interest that Koch (1929), who travelled extensively in this part of Greenland, described from the region immediately north of Washington Land (Hall Land and area to the east, Fig. 1), a loose, coarse clastic deposit, several hundred metres thick, as the youngest preglacial strata of western North Greenland. He referred the strata to the Polaris Harbour Formation of Lower Palaeozoic age.

The true nature and age of the coarse clastic deposits described by Koch (1929) have been a matter of debate for some time, mainly because Koch's lithological description of the deposit is difficult to reconcile with the Lower Palaeozoic bedrock in the distribution area, including the type locality, of the Polaris Harbour Formation. Thus Dawes & Haller (1979) suggested that the coarse clastic deposits recorded by Koch may well be stratigraphic equivalents of the Washington Land fluvial deposits referred in this paper to the Bjørnehiet Formation. The systematic mapping of western North Greenland planned for 1983 should clarify the question of the Polaris Harbour Formation and define the distribution limits of the Bjørnehiet Formation.

Faulting

The area where the Bjørnehiet Formation outcrops is cut by several faults. The most important is an east–west-trending normal fault just north of the type section (Fig. 2). This fault has displaced, with a vertical downthrow to the south, the Ordovician strata by about 140 m. Movement along this fault, and also along a parallel fault to the south, has affected the Bjørnehiet Formation; the northern fault block has been uplifted by at least 40 m.

Block faulting with minor tilting of the homoclinal Lower Palaeozoic strata is typical of the structure in parts of south-eastern Washington Land (Jepsen & Dueholm 1978). While the relationships of such fault tectonics to those affecting the Bjørnehiet Formation are uncertain, it is now clear that some faults mapped in the Lower Palaeozoic platform terrain have been reactivated by relatively late (Tertiary or Quaternary) fault movements.

Age

The age of the Bjørnehiet Formation is uncertain. No fossils have been collected from the conglomerate matrix. However, throughout the area where the formation is present branches and trunks of non-carbonised fossil conifer wood, up to 15 cm in diameter, are common. The wood has not been found in situ but from its distribution it seems likely that it has been washed out from the formation. Carbon-14 dating of wood samples from the outcrop area (Teledyne Isotope Inc.) has given ages older than 40 000 years B. P., suggesting that the Bjørnehiet Formation is of preglacial age.

Fossil wood of similar age is known about 50 km south-east of the type area of the Bjørnehiet Formation (Weidick 1978) where it has been found in a river valley about 245 m above sea level. No occurrence of deposits similar to the Bjørnehiet Formation has been reported at this locality.

Correlation

No comparable beds to the Bjørnehiet Formation of Washington Land are known anywhere else in western North Greenland. However, in coastal Ellesmere Island on the opposite shore of Nares Strait, a Tertiary (Paleocene) conglomerate with minor, thin sandstone beds of the Eureka Sound Formation form scattered outcrops indicating an extensive depositional area (see Mayr & de Vries, this volume). This conglomerate, like the Bjørnehiet Formation, is mainly oligomictic and characterised by locally-derived carbonate clasts. The Ellesmere Island conglomerate sequence reaches a thickness of up to 1000 m and it has been severely affected by thrusting and faulting of the Tertiary (Eurekan) orogeny.

Another possible correlative of the Bjørnehiet Formation is the rather unconsolidated, coarse alluvial deposits of the Beaufort Formation that post-date the Eurekan orogeny (Craig & Fyles 1960). This formation is widespread in Arctic Canada, but it has not been recorded from eastern Ellesmere Island (Thorsteinsson & Tozer 1970). The Beaufort Formation is characterised by its wood- and plant-bearing content by which it has been dated as late Tertiary (Miocene–Pliocene) or earliest Pleistocene. Thus, like the Bjørnehiet Formation, it is regarded essentially as a preglacial deposit.

At present it is unknown whether the Greenland fluvial deposit can be correlated with either of the two Cenozoic sequences from Canada; thus it remains uncertain whether the Bjørnehiet Formation is of Palaeogene or Neogene age. Its wood-bearing character perhaps strengthens the alliance with the Beaufort Formation, in which case Washington Land becomes particularly important for the presence of late Cenozoic faulting. On the other hand, if the Greenland deposit is a Eureka Sound Formation equivalent, the faulting affecting the deposit may represent the outer effect of the Eurekan orogeny that affected northern Ellesmere Island and large parts of northern Greenland. In addition, an early Tertiary age designation would be particularly significant for the present discussion of this symposium, since the degree of facies continuity so established across Nares Strait would provide comparative sedimentation and tectonic data which would lead to a much more complete Cenozoic palaeogeographic model of the Nares Strait region than is now possible, based solely on the eastern Ellesmere Island outcrops.

Conclusion

The Bjørnehiet Formation is a fluvial conglomerate of probable preglacial age which pre-dates at least some fault movements. Its present known distribution is restricted to central Washington Land, but other outcrop areas may exist farther north in western North Greenland.

The age of the formation, and thus its correlation with the Palaeogene or Neogene deposits of Arctic Canada, is uncertain. However, whatever its precise age, the conglomerate occurrence provides important Cenozoic, preglacial palaeogeographical data for a region of Greenland otherwise lacking this part of the stratigraphic record. If the formation can be accurately dated, its importance to the understanding of the palaeogeography and tectonic history of the Nares Strait region will be substantially increased.

Acknowledgements

The author thanks P. R. Dawes, A. K. Higgins and J. M. Hurst for critical reading of the paper which is published with the permission of the Director of the Geological Survey of Greenland.

References

Craig, B. G. & Fyles, J. G. 1960. Pleistocene geology of Arctic Canada. – Pap. geol. Surv. Can. 60–10: 21 pp.

Dawes, P. R. & Haller, J. 1979. Historical aspects in the geological investigation of northern Greenland. Part 1: New maps and photographs from the 2nd Thule Expedition 1916–1918 and the Bicentenary Jubilee Expedition 1920–1923. – Meddr Grønland 200(4): 38 pp.

Jepsen, H. F. & Dueholm, K. 1978. Computer supported geological photo-interpretation. – Rapp. Grønlands geol. Unders. 90: 146–150.

Koch, L. 1929. Stratigraphy of Greenland. – Meddr Grønland 73, 2(2): 205–320.

Mayr, U. & de Vries, C. D. S. 1982. Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago. – This volume.

Peel, J. S. 1978. Geological investigations in Lower Palaeozoic terrain of northern Greenland between 78°30'N and 81°30'N. – Rapp. Grønlands geol. Unders. 90: 14–16.

Peel, J. S. & Christie, R. L. 1982. Cambrian–Ordovician platform stratigraphy: correlations around Kane Basin. – This volume.

Thorsteinsson, R. & Tozer, E. T. 1970. Geology of the Arctic Archipelago. – In: Douglas, R. J. W. (ed.), Geology and economic minerals of Canada. – Econ. Geol. Rep. geol. Surv. Can. 1: 547–590.

Weidick, A. 1978. C^{14} dating of Survey material carried out in 1977. – Rapp. Grønlands geol. Unders. 90: 119–124.

Structure

Fold belts and metamorphic zones of northern Ellesmere Island and North Greenland

ANTHONY K. HIGGINS, ULRICH MAYR and NORMAN J. SOPER

Higgins, A. K., Mayr, U. & Soper, N. J. 1982. Fold belts and metamorphic zones of northern Ellesmere Island and North Greenland. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 159–166.

The Innuitian tectonic province extends in its eastern part through northern Ellesmere Island and across North Greenland. The mid- to late Palaeozoic Ellesmerian orogeny produced NE–SW to E–W trending fold belts and metamorphic zones in the strata of the Franklinian geosyncline. The late Cretaceous–Tertiary Eurekan orogeny deformed the Carboniferous–Tertiary rocks of the Sverdrup and Wandel Sea Basins, and overprinted or accentuated the older structures of the Franklinian geosyncline. The locations of the south limit of Ellesmerian folding, and of fold belts of comparable style, deformation intensity and metamorphic grade on the two sides of the Nares Strait lineament are consistent with models of left-lateral displacement in the range 0–50 km. Models of displacement greater than 50 km introduce increasingly improbable mismatches of fold belts and metamorphic zones.

A. K. Higgins, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark; U. Mayr, Department of Energy, Mines and Resources, Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303–33rd St. N.W., Calgary, Alberta, Canada, T2L 2A7 and N. J. Soper, Department of Geology, University of Sheffield, Mappin Street, Sheffield S1 3JD, England.

The Innuitian tectonic province extends in its eastern part through northern Ellesmere Island and into North Greenland. Proterozoic, Lower Palaeozoic and Devonian rocks of the Franklinian geosyncline were deformed in the Ellesmerian orogeny, and Carboniferous–Palaeogene sediments overlying the folded rocks were themselves deformed in the mid-Cenozoic Eurekan orogeny. The Nares Strait lineament lies oblique to the trend of the Ellesmerian fold belts and metamorphic zones, allowing a possibility of determining the displacement, if any, on the lineament. The most recent summaries of the province in Ellesmere Island and North Greenland are respectively those of Trettin & Balkwill (1979) and Dawes & Peel (1981).

Thick Proterozoic and Cambrian to Devonian sedimentary sequences accumulated in the Franklinian geosyncline which extended through Ellesmere Island and across northern Greenland. Shallow marine to non-marine conditions prevailed throughout the earliest Cambrian in Ellesmere Island and in North Greenland, with subsequent development of a marine basin. From the middle Cambrian to early Devonian three major depositional belts are apparent: an unstable south-eastern shelf which received mainly carbonate sediments; the Hazen Trough in which starved-basin sediments (cherts, graptolitic shales and redeposited carbonates) and thick flysch deposits accumulated; and a complex north-western belt including volcanic rocks and clastic and carbonate sediments. The third belt is restricted to northern Ellesmere Island. Clastic sedimentation continued until the Middle and Late Devonian in northern Ellesmere Island, and into the Upper Silurian and locally Devonian in North Greenland, being brought to a close by the Ellesmerian orogeny.

The Ellesmerian orogeny produced the most extensive and intense deformation of the Innuitian tectonic province. The folds developed in the mainly Cambrian–Devonian rocks show a strong alignment with NE–SW to E–W trends dominant, the folds being most intense in the north, and in the south passing into unfolded platform terrain. Significant metamorphic grades were only reached in northern Ellesmere Island and in northern Peary Land in Greenland. Plate-tectonic models have been suggested for the origin of the folding in Ellesmere Island (Trettin et al. 1972, Trettin & Balkwill 1979) and North Greenland (Dawes & Soper 1973), in both regions being interpreted in terms of southward underthrusting of ocean floor against and beneath the continental block and Franklinian geosyncline.

Folded Cambrian–Devonian sediments are overlain by great thicknesses of early Carboniferous to late Cretaceous, locally Eocene, sediments in Ellesmere Island (Sverdrup Basin: Balkwill 1978), and by Carboniferous to early Tertiary sediments in eastern North Greenland (Wandel Sea Basin: Dawes & Soper 1973, Håkansson 1979). In both the Sverdrup Basin and Wandel Sea Ba-

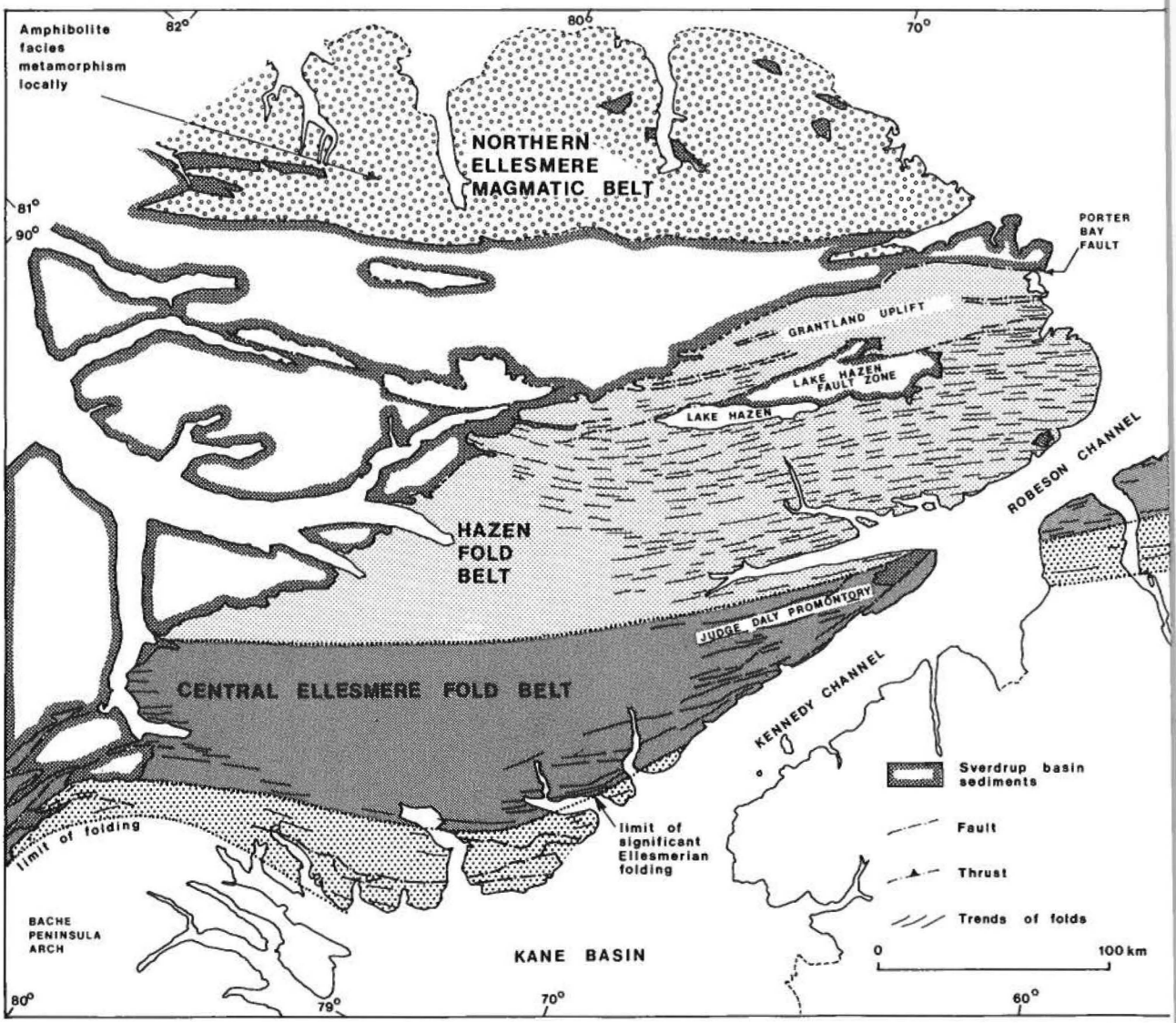


Fig. 1. Palaeozoic fold belts of northern Ellesmere Island. Intensity of deformation increases northwards, with amphibolite-facies metamorphism in parts of the northern Ellesmere magmatic belt. A narrow zone of very weak folding lies south of the central Ellesmere fold belt.

sin there was local deformation in the Permian, but the principal tectonism was in the late Cretaceous and Tertiary (Eurekan orogeny). Graben structures, folds and faults characterise different areas.

The idea of sinistral wrench displacement on the Nares Strait lineament has arisen from the necessity of accommodating the anti-clockwise rotation of Greenland relative to North America due to the spreading origin of the Labrador Sea. Spreading was confined to the period between anomalies 31 (68 m.y. late Cretaceous) and 13 (35 m.y. early Oligocene); note that the polarity time scale of Hailwood et al. (1979) is adopted. Since spreading as far north as Baffin Bay is debatable (Grant, this volume, Soper et al., this volume), motion in the region of the Canadian Arctic Archipelago and North Greenland must have been accommodated by within-plate deformations. The two obvious possibilities are sinistral displacement on the Nares Strait lineament and compression across the Eurekan orogen.

Any sinistral wrench movements along Nares Strait should be revealed by offset of trends and boundaries associated with the Ellesmerian (Palaeozoic) orogeny, whereas Eurekan structures should show little displacement. Such features have previously been used in arguments in support of little movement (Kerr 1967) or very extensive movement (Newman 1977, Newman & Falconer 1978). In this paper the deformation and metamorphic patterns due to the Ellesmerian orogeny

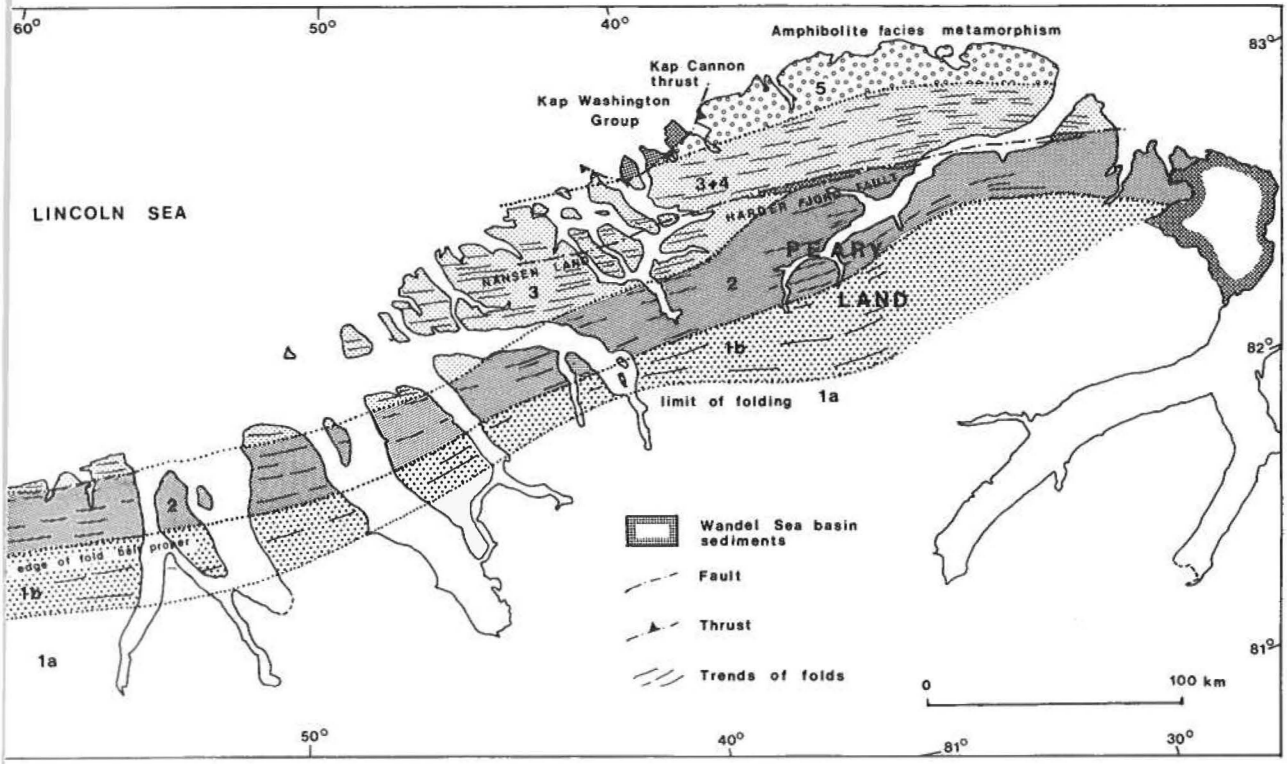


Fig. 2. Structural and metamorphic zones of the North Greenland fold belt (after Dawes 1976). Deformation intensity increases northwards, with amphibolite-facies metamorphism only in the northern part of zone 5. The ornament used for each zone indicates correlations with the equivalent fold belts of northern Ellesmere Island.

on the two sides of Nares Strait are reviewed, and the boundaries between comparable belts of deformation and zones of metamorphism are considered in a series of displacement models.

Ellesmere Island

The history of deformation within the Innuitian tectonic province of Ellesmere Island has most recently been reviewed by Trettin & Balkwill (1979). The Palaeozoic Ellesmerian orogeny progressed from north to south and produced three major fold belts with markedly different structural characteristics (Fig. 1): central Ellesmere fold belt, Hazen fold belt, northern Ellesmere magmatic belt (or fold belt). In the north deformation commenced not later than middle Devonian, and in the south probably began in latest Devonian or earliest Carboniferous time. The fold trends are generally ENE–WNW, but show locally pronounced variations, as for example in the south where structures swing around the Bache Peninsula arch. There is only a narrow zone of unfolded platform rocks at the present level of exposure.

The southernmost of the three fold belts, the *central Ellesmere fold belt,* is characterised by extensive concentric folds controlled by competent carbonate units. In the southern part of Judge Daly Promontory the folds have north-east trending, sub-horizontal fold axes, and individual folds can be traced for up to 50 km. Wide, open synclinoria alternate with tight, steep anticlines. A change in fold style occurs in the northern part of Judge Daly Promontory, where the Palaeozoic structures are strongly dissected by steep normal and reverse faults of probable Tertiary age. Typical recognisable Palaeozoic structures are monoclines and tight, steep anticlines. Some folds are overturned toward the north-west, indicating the beginning of a northerly vergence which is much more obvious in the folding in North Greenland. Southwards a zone of very weak folds separates the main central Ellesmere fold belt from unfolded platform carbonates.

The *Hazen fold belt* comprises the deformed strata of the north-east part of the Hazen Trough, and shows greater stratal shortening than the central Ellesmere fold belt (Trettin & Balkwill 1979). It is characterised by very numerous, closely-spaced tight and isoclinal major folds, of chevron type in the flysch, and concentric type where they deform thin chert units. Axial

planes dip generally north-west at angles between 75° and vertical, but south-east dips are also recorded (Trettin 1971). The folds display well-developed axial-plane cleavage, and they are associated with numerous faults, including south-eastward directed minor thrusts. The Grantland uplift, bordered to the south by the Lake Hazen fault zone and crossed by the Porter Bay fault zone, is now regarded as the northernmost part of the Hazen fold belt (H. P. Trettin, pers. comm. 1980). Parts of the Hazen fold belt are characterised by sub-greenschist and greenschist-facies metamorphism. It has not been greatly modified by Tertiary deformation, as outliers of Carboniferous and Tertiary strata, except where directly involved in the faulting, are only weakly deformed.

The *northern Ellesmere magmatic belt* exhibits a number of unique features: the presence of Grenvillian basement, Ordovician and Silurian volcanics, Devonian plutons, and amphibolite-facies grade early Palaeozoic rocks (Frisch 1974). It also differs from the Hazen fold belt by the presence of Ordovician and lower Silurian shelf carbonates. Fold trends are mainly NE–SW, but arcuate trends are also recorded.

Late Cretaceous to Tertiary tectonism (Eurekan orogeny) overprinted the Ellesmerian fold belts to a varying degree, in some cases merely accentuating the older structures. One of the principal results was the uplift of the southern Grantland mountains by thrust faulting on the Lake Hazen fault zone. The Grant Land Formation was thrust over early Ordovician to Tertiary rocks. Strike-slip faulting probably took place along the Porter Bay fault zone. The Sverdrup Basin sediments in the vicinity of the faults were warped into open folds. The compressional folds and faults are commonly cited as supporting the rotation of Greenland as a result of early Tertiary sea-floor spreading, but this is only possible if the revised polarity time scale of Hailwood et al. (1979) is employed. If the Heirtzler et al. (1968) time scale is used, the structures, generated between middle Eocene and early Miocene, are too young (Trettin & Balkwill 1979).

Along the south-east margin of that part of the central Ellesmere fold belt adjacent to Kane Basin several low-angle thrusts formed; Lower Palaeozoic formations are thrust on Paleocene conglomerate and sandstone. Along the Kennedy Channel coast the thrusts are transformed into strike-slip faults as far north as the Judge Daly fault zone on northern Judge Daly Promontory. Stresses associated with the strike-slip movement disrupted and overprinted the Ellesmerian structures (Mayr & de Vries, this volume). The earliest possible local age for thrusting and strike-slip movement is late Paleocene, but from regional considerations the deformation is more likely to be post-middle Eocene in age.

North Greenland

The North Greenland fold belt has been divided into a series of approximately E–W trending zones (Dawes 1976, Dawes & Soper 1973, 1979) which show increasing northwards intensity of deformation and metamorphism.

The hinterland of the fold belt is a broad zone of unfolded, flat-lying to gently northward-dipping Lower Palaeozoic strata, mainly platform carbonates (zone 1a, Fig. 2). The first indications of deformation are large-scale open warps and very gentle structures (zone 1b). The margin of the fold belt proper is abrupt and without regional dislocations; it is an easily traceable line across entire North Greenland. Zone 2 is marked by the appearance of E–W trending generally steep-limbed symmetrical folds. The folds vary from open to tight, and in several areas along the length of the fold belt they have a box-fold style. In Peary Land the folds at the south margin of the zone are southward-facing monoclines (Dawes & Soper 1979, Pedersen 1979), but the structures are more often overturned northwards with steep south-dipping axial planes. Parts of zone 2 studied by Pedersen (1980) in eastern North Greenland are characterised by westward-directed imbricate thrusts.

Zones 3, 4 and 5 represent increasingly complex deformation, a sequence only fully developed in eastern North Greenland. In zones 3 and 4 first folds are still conspicuous as chains of tight to isoclinal folds, upright or with steep south-dipping axial planes. Second phase folds are common, developed on south-dipping axial planes with a congruous cleavage, and become progressively more intense and overturned northwards in zone 5, where they are joined by similarly oriented third folds. It is in particular the second and third phase folds which give the strong northerly vergence to the fold belt.

Zones 1 and 2 are essentially non-metamorphic. Metamorphic grade increases gradually northwards in zones 3, 4 and 5 with the appearance of chlorite, biotite, garnet, and locally staurolite, andalusite and cordierite. Amphibolite-facies rocks are found only in the northern part of zone 5.

The three main deformation phases so far recognised and the progressive metamorphism are considered to be of Devonian and/or early Carboniferous age. However, cross-cutting basic dyke swarms, of presumed Cretaceous age, are locally affected by post-intrusion deformation and greenschist-facies metamorphism in the vicinity of the Tertiary Kap Cannon thrust. Late Phanerozoic K-Ar mineral ages have been obtained from Palaeozoic metasediments as well as from cross-cutting basic dykes (Dawes & Soper 1971).

The Kap Cannon thrust is a south-dipping major structure of probably mid-Tertiary age separating volcanic rocks known as the Kap Washington Group from overlying highly-deformed metasediments of the fold

belt. The Kap Washington Group is of late Cretaceous age and has given a Rb-Sr isochron age of 63 m.y. (Larsen et al. 1978, Batten et al. 1981). It overlies Carboniferous–Permian and Cretaceous sediments of the Wandel Sea Basin succession (Higgins et al. 1981, Håkansson et al. 1981).

Other Tertiary structures include a system of NW–SE trending faults and associated drag folds affecting the Wandel Sea Basin sediments (Håkansson 1979), and the Harder Fjord fault zone — a major E–W structure running through the centre of the fold belt. The Harder Fjord fault zone shows relative uplift of the northern block, while within graben-like structures in the fault zone there are preserved restricted sequences of Upper Permian, Cretaceous and Tertiary sediments (Croxton et al. 1980, Soper et al. 1980, Wagner et al. in press).

Discussion

It is evident that the mid-Palaeozoic fold belts of northern Ellesmere Island and North Greenland were formerly continuous. In both regions the effects of the Ellesmerian orogeny are seen in the northward increase in deformation intensity and metamorphic grade, and the 'fold belts' distinguished in northern Ellesmere Island can be closely compared and correlated with the 'structural-metamorphic zones' of North Greenland. The obvious correlations have been depicted in Figs 1 and 2 by use of the same ornamentation. However, the positioning of the boundary lines between the different 'belts' or 'zones' is subjective, and they cannot be used for other than generalised assessments of the displacement along the Nares Strait lineament. The northern Ellesmere magmatic belt is, for example, of different character from the northernmost parts of the North Greenland fold belt; their only common feature is the occurrence of amphibolite-facies Palaeozoic rocks.

A feature of interest is the general alignment of the Lake Hazen fault zone in Ellesmere Island and the Harder Fjord fault zone in North Greenland. Both structures show substantial uplift to the north and had their latest movements in the Tertiary; both structures may have had a long pre-Tertiary history, though the evidence for early movements is not well documented. The two faults are separated by about 200 km of the Arctic Ocean with the present configuration of Ellesmere Island and Greenland (Figs 1 and 2), but may represent a single fundamental fracture zone of potentially the same importance as the Nares Strait lineament. In addition, the important fault zones on Judge Daly Promontory and northern Nyeboe Land also show good alignment (see Dawes, this volume).

A striking feature of the Innuitian tectonic province in the region as a whole is its greater north–south width in Ellesmere Island compared to North Greenland (Figs 1 and 2). Complexly folded rocks in amphibolite facies are in the one case 350 km distant from the south limit of folding, and in the other 150 km. Another general difference between the folded rocks on the two sides of Nares Strait is the general southwards overturning of folds in Ellesmere Island compared to a general northwards sense of overturning in Greenland; there are exceptions in both regions. These features are no doubt original characteristics developed during the Ellesmerian orogeny, arising in part from the nature and distribution of the rocks undergoing deformation, and in part from the nature and location of the active deformation mechanism; discussion of their significance is beyond the scope of this paper.

In view of the above considerations, and since the Nares Strait lineament transects structural trends and boundaries at low angles, precise determination of wrench movements along the line of the lineament is not possible. Nevertheless there is sufficient information to place restraints on the likely range of displacement.

Constraints on displacement

A range of models ranging from 0 to 250 km left-lateral displacement are considered, with their consequences for correlation of structural and metamorphic features (Fig. 3).

No displacement (Fig. 3a). – The present-day configuration requires a pronounced swing of the south limit of folding parallel to the line of the Nares Strait lineament, but this swing in trend is clearly reflected by the trends of individual folds on the Canadian side of Kennedy Channel. There is a good line-up of the southern belts of more open folding, and also of the Hazen fold belt and the tightly folded rocks of Nansen Land. Fold trends show good alignment. The highest grade metamorphic areas lie on strike, but as they are 500 km apart this is not a conclusive line of argument.

50 km left-lateral displacement (Fig. 3b). – The swing of the south limit of folding is less pronounced in this model, but still reflects that of the folds on land. The two southern zones of more open folding line up well, the central Ellesmere fold belt showing a gradual, and realistic, attenuation eastwards into zone 2 on the Greenland side.

100 km left-lateral displacement (Fig. 3c). – The southern two fold belt boundaries can now be joined by sinuous curves, somewhat less pronounced than the swings of individual fold trends on the west side of Kennedy Channel. However, the south margin of the Hazen fold belt is now offset relative to the 2–3 zone boundary, such that open folds on Judge Daly Promontory lie on strike with tight folds in Greenland. Furthermore there is non-alignment of individual fold trends in this model from the central Ellesmere fold belt over to zone 2 on the Greenland side.

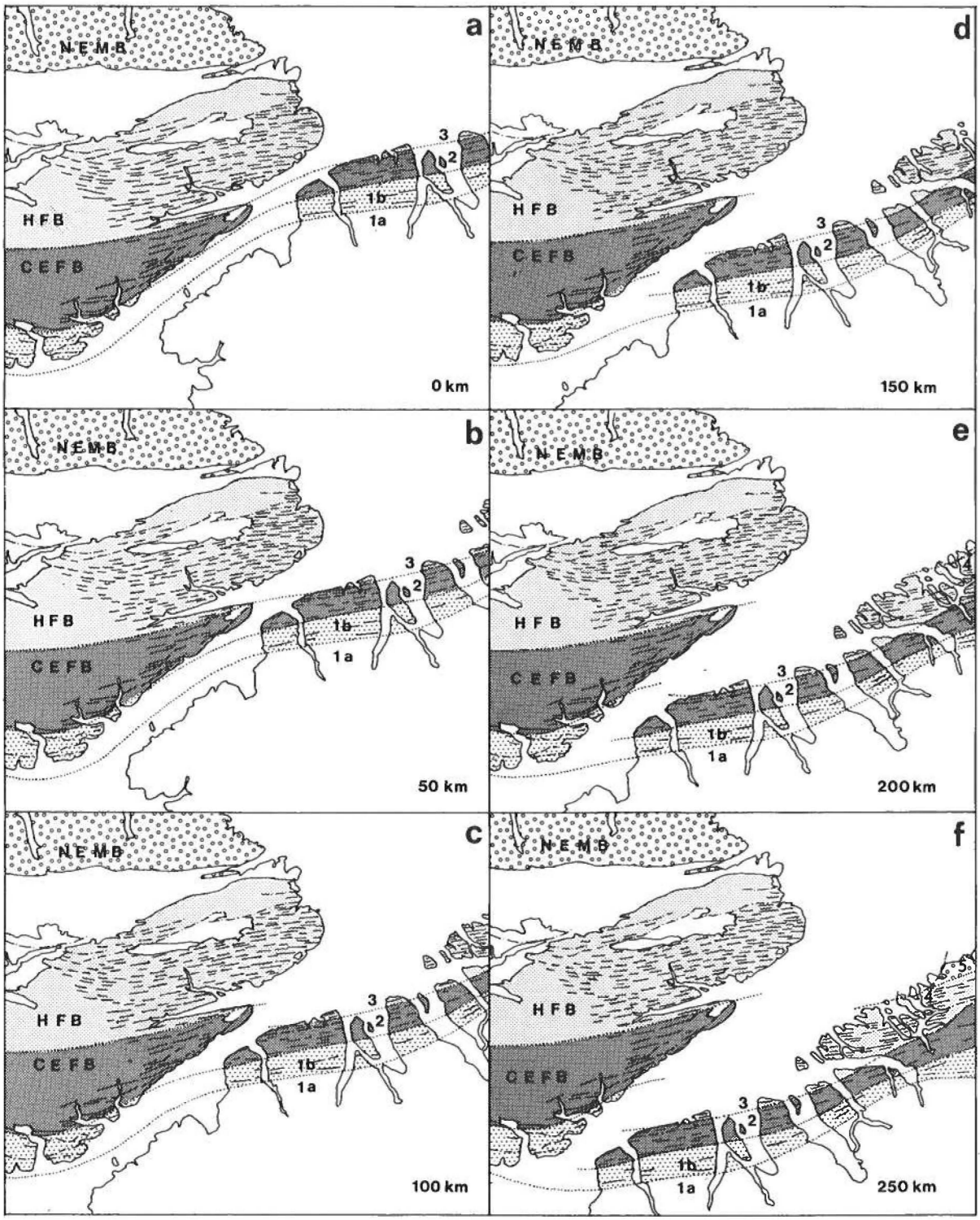


Fig. 3. Pre-displacement constructions of the relative positions of northern Ellesmere Island and North Greenland, assuming sinistral strike-slip displacements on Nares Strait of zero to 250 km. NEMB: northern Ellesmere magmatic belt. HFB: Hazen fold belt. CEFB: central Ellesmere fold belt. 1a, 1b, 2, 3, 4, 5: structural–metamorphic zones in North Greenland.

150, 200 and 250 km displacement (Fig. 3d, e, f). – With increasing displacement there is progressively greater mismatch of the fold belts, fold trends and metamorphic zones. In the worst case (250 km displacement), amphibolite-facies, intensely deformed metasediments in Greenland (zone 5) lie on strike with non-metamorphic gently folded terrain in Ellesmere Island.

Conclusions

The present distribution of fold belts and metamorphic zones in northern Ellesmere Island and North Greenland is consistent with no sinistral displacement on the Nares Strait lineament since mid-Palaeozoic time. However, we consider that displacements up to 50 km can be accommodated without causing unlikely geological relationships. In any case zero offset seems improbable in view of the scale and remarkable linearity of the structure. Displacement models greater than 50 km show increasing mismatch of structural and metamorphic elements and are considered most unlikely to have occurred.

The structural and metamorphic data thus favour those interpretations of Greenland – North America spreading geometry which require small wrench displacements on the Nares Strait line (see Soper et al., this volume).

Acknowledgements

We thank H. P. Trettin and P. R. Dawes for constructive criticism of the manuscript. A. K. H. and N. J. S. publish with permission of the Director of the Geological Survey of Greenland.

References

Balkwill, H. R. 1978. Evolution of Sverdrup Basin, Arctic Canada. – Bull. Am. Ass. Petrol. Geol. 62: 1004–1028.

Batten, D. J., Brown, P. E., Dawes, P. R., Higgins, A. K., Koch, B. E., Parsons, I. & Soper, N. J. 1981. Peralkaline volcanicity on the Eurasia Basin margin. – Nature, Lond. 294: 150–152.

Croxton, C. A., Dawes, P. R., Soper, N. J. & Thomsen, E. 1980. An occurrence of Tertiary shales from the Harder Fjord Fault, North Greenland fold belt, Peary Land. – Rapp. Grønlands geol. Unders. 101: 61–64.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R. 1982. The Nyeboe Land fault zone: a major dislocation on the Greenland coast along northern Nares Strait. – This volume.

Dawes, P. R. & Peel, J. S. 1981. The northern margin of Greenland from Baffin Bay to the Greenland Sea. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 201–264. – Plenum Press, New York & London.

Dawes, P. R. & Soper, N. J. 1971. Significance of K/Ar age determinations from northern Peary Land. – Rapp. Grønlands geol. Unders. 35: 60–62.

Dawes, P. R. & Soper, N. J. 1973. Pre-Quaternary history of North Greenland. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 117–134.

Dawes, P. R. & Soper, N. J. 1979. Structural and stratigraphic framework of the North Greenland fold belt in Johannes V. Jensen Land, Peary Land. – Rapp. Grønlands geol. Unders. 93: 40 pp.

Frisch, T. [O.] 1974. Metamorphic and plutonic rocks of northernmost Ellesmere Island, Canadian Arctic Archipelago. – Bull. geol. Surv. Can. 229: 87 pp.

Grant, A. C. 1982. Problems with plate tectonic models for Baffin Bay – Nares Strait: evidence from the Labrador Sea. – This volume.

Hailwood, E. A., Bock, W., Costa, L. [I]., Dupeuble, P. A., Müller, C. & Schnitker, D. 1979. Chronology and biostratigraphy of northeast Atlantic sediments, DSDP Leg. 48. – Initial Rep. Deep Sea Drilling Proj. 48: 1119–1141.– U.S. Gov. Print. Off., Washington, D.C.

Heirtzler, J. R., Dickson, G. O., Herron, E. M., Pitman, W. C. & Le Pichon, X. 1968. Marine magnetic anomalies, geomagnetic field reversals, and motions of the ocean floor and continents. – J. geophys. Res. 73: 2119–2136.

Higgins, A. K., Friderichsen, J. D. & Soper, N. J. 1981. The North Greenland fold belt between central Johannes V. Jensen Land and eastern Nansen Land. – Rapp. Grønlands geol. Unders. 106: 35–45.

Håkansson, E. 1979. Carboniferous to Tertiary development of the Wandel Sea Basin, eastern North Greenland. – Rapp. Grønlands geol. Unders. 88: 73–83.

Håkansson, E., Heinberg, C. & Stemmerik, L. 1981. The Wandel Sea Basin from Holm Land to Lockwood Ø, eastern North Greenland. – Rapp. Grønlands geol. Unders. 106: 47–63.

Kerr, J. W. 1967. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Larsen, O., Dawes, P. R. & Soper, N. J. 1978. Rb/Sr age of the Kap Washington Group, Peary Land, North Greenland, and its geotectonic implication. – Rapp. Grønlands geol. Unders. 90: 115–119.

Mayr, U. & de Vries, C. D. S. 1982. Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago. – This volume.

Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Newman, P. H. & Falconer, R. K. H. 1978. Evidence for movement between Greenland and Canada along Nares Strait. – Geol. Soc. Am. Abs with Prog. 10: 463 only.

Pedersen, S. A. S. 1979. Structural geology of central Peary Land, North Greenland. – Rapp. Grønlands geol. Unders. 88: 55–62.

Pedersen, S. A. S. 1980. Regional geology and thrust fault tectonics in the southern part of the North Greenland fold belt, North Peary Land. – Rapp. Grønlands geol. Unders. 99: 79–87.

Soper, N. J., Higgins, A. K. & Friderichsen, J. D. 1980. The North Greenland fold belt in eastern Johannes V. Jensen Land. – Rapp. Grønlands geol. Unders. 99: 89–98.

Soper, N. J., Dawes, P. R. & Higgins, A. K. 1982. Cretaceous–Tertiary magmatic and tectonic events in North Greenland and the history of adjacent ocean basins. – This volume.

Trettin, H. P. 1971. Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, Arctic Archipelago. – Bull. geol. Surv. Can. 203: 134 pp.

Trettin, H. P. & Balkwill, H. R. 1979. Contributions to the tectonic history of the Innuitian Province, Arctic Canada. – Can. J. Earth Sci. 16: 748–769.

Trettin, H. P., Frisch, T. O., Sobczak, L. W., Weber, J. R., Law, L. K., DeLaurier, J. M. & Whitham, K. 1972. The Innuitian Province. – In: Price, R. A. & Douglas, R. J. W. (eds), Variations in tectonic styles in Canada. – Spec. Pap. geol. Ass. Can. 11: 83–179.

Wagner, R. H., Soper, N. J. & Higgins, A. K. in press. A late Permian flora of Pechora affinity in North Greenland. – Rapp. Grønlands geol. Unders. 108.

Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago

ULRICH MAYR and CHRISTIAAN D.S. de VRIES

Mayr, U. & de Vries, C. D. S. 1982. Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 167–175.

In the coastal area of Ellesmere Island between Copes Bay in the south and Cape Baird in the north, at least three phases of Tertiary structural deformation can be distinguished. These are: 1) Paleocene uplifting, 2) Eocene thrusting and 3) strike-slip faulting, probably not younger than Miocene.

Uplifting is documented by remnants of a coarse, oligomictic, alluvial conglomerate of Tertiary age. In the southern and central part of the area the conglomerate is over-thrusted by Lower Palaeozoic rocks. Conjugate strike-slip faults are present on Darling Peninsula and northern Judge Daly Promontory. Sinistral displacement along the Judge Daly fault zone is estimated tentatively at 19 km.

Thrusting and strike-slip faulting in the Nares Strait area probably correspond to the main phase of the Eurekan orogeny.

U. Mayr, Department of Energy, Mines and Resources, Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303–33rd St. N.W., Calgary, Alberta, Canada, T2L 2A7 and C. D. S. de Vries, deceased, 3rd July 1979.

Nares Strait has long been considered a major fracture of the northern hemisphere and various amounts of sinistral displacement, ranging from a few tens to several hundreds of kilometres, have been proposed for this lineament (see Kerr 1980 and this volume for discussion of historical background). If Nares Strait is indeed a zone of strike-slip faulting, then structures which are the result of such strike-slip movement should be exposed along its shores. Published maps from Greenland (Koch 1929, Escher 1970, Dawes & Haller 1979) show that the platform rocks are, along the southeast coast of Nares Strait, without major deformation or features which can be attributed to strike-slip faulting. However along the northwest shore of Nares Strait, on Ellesmere Island between Copes Bay and Cape Baird at the northern tip of Judge Daly Promontory, thrusts, faults and folds originating from both the Ellesmerian (late Palaeozoic) and Eurekan (Tertiary) orogenies, are exposed. These structures form the southeastern margin of the Ellesmere fold belt.

It is the intent of the present investigation firstly to examine the Tertiary structures to see whether some are related to any strike-slip movement along Nares Strait, and secondly to see whether a sequence of separate structural events can be established and if such events can be correlated with the phases of the Eurekan orogeny as defined by Balkwill (1978) farther west in the Sverdrup Basin.

Regional geological setting

Fig. 1 shows the coastal region of eastern Ellesmere Island described in this paper. The area northwest of the plotted features contains various tectonic subdivisions, among them the Ellesmere fold belt. The plotted structures form the southeastern margin of the Ellesmere fold belt and for a discussion of the regional tectonic units and their relation to northern Greenland the reader is referred to the paper in this volume by Higgins et al.

Geological data

Several maps covering the area between Copes Bay and Cape Baird (Fig. 1) have been published by the Geological Survey of Canada. One of the earliest contained in a report by Christie (1964) covers Judge Daly Promontory as far south as Carl Ritter Bay. He suggested transcurrent movement along the Judge Daly fault zone between Cape Defosse and Cape Baird. On a later map and in the accompanying text Christie (1974) pointed out the extension of the Judge Daly fault zone as far south as Carl Ritter Bay.

The area between Carl Ritter Bay and Copes Bay is covered by three maps (Kerr 1973a, b, c) on which the major structural features were named. The same author

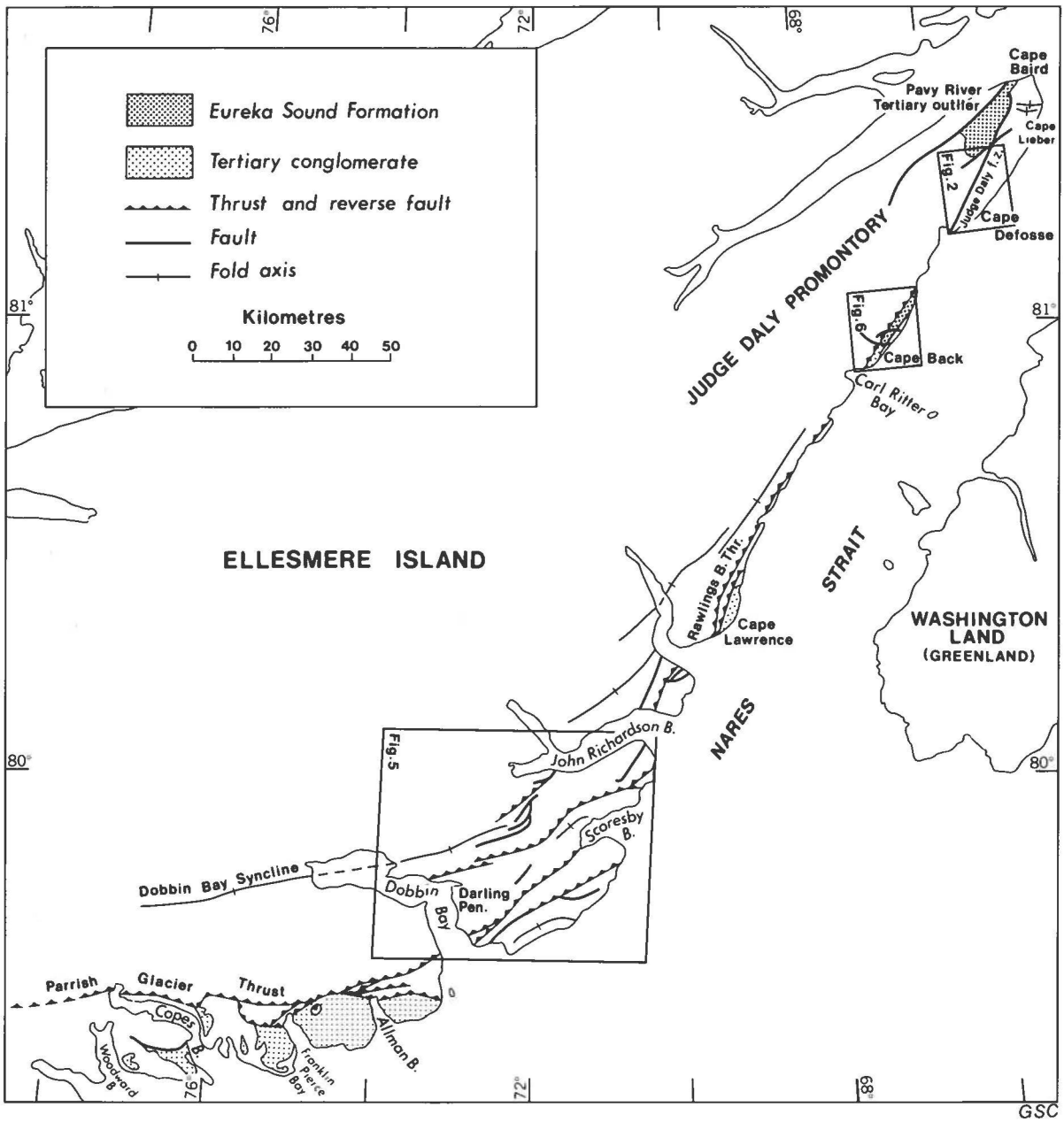


Fig. 1. Index map of central eastern Ellesmere Island, showing the main structural features along the Nares Strait coast and the locations of Figs 2, 5 and 6. Main Tertiary outcrop areas are shown; blank areas in Ellesmere Island are Proterozoic to Devonian of the Franklinian geosyncline.

(Kerr 1967) also named the Judge Daly fault zone. The published maps are generalized and not at sufficiently large scale for detailed structural interpretation of Nares Strait. The present interpretation is based on the re-examination of unpublished photogeological interpretations and field data obtained by the authors during 1972 and 1973 for Sproule Associates Limited, Calgary, Alberta, Canada.

Tertiary stratigraphy

Outliers of Tertiary sediments (Fig. 1) make it possible to distinguish Ellesmerian and Eurekan components in the margin area of the Ellesmere fold belt along Nares Strait. The lithology and distribution of these Tertiary rocks also furnishes important clues for describing a possible sequence of local structural events during the

Eurekan orogeny. Two Tertiary lithofacies are present: sandstone, siltstone, coal and mudstone of the Eureka Sound Formation and a boulder conglomerate. This conglomerate is described in this paper separately from the Eureka Sound Formation although Miall (1981) includes it as the uppermost unit of that formation.

Eureka Sound Formation

The Eureka Sound Formation is present as two significant outliers, one just north of Cape Back and the other one at Pavy River. In both locations the beds are moderately to gently folded and are preserved in graben structures. In the Pavy River outlier the Eureka Sound Formation unconformably overlies Lower Palaeozoic formations. The lithology has been described and interpreted in detail by Miall (1981). Estimated maximum thickness of the Eureka Sound Formation is in the order of 1000 m. The age of the exposed beds is Paleocene on the basis of palynology (R. L. Christie & W. S. Hopkins, pers. comm. 1980) and leaf impressions (Miall 1981).

Conglomerate

The conglomerate is exposed in three areas, at Cape Back, at Cape Lawrence and in isolated patches in the region between Dobbin Bay and Woodward Bay (Fig. 1). It unconformably overlies Lower Palaeozoic rocks. In the southern areas it is flat lying and the Rawlings Bay and Parrish Glacier Thrusts have carried Proterozoic and younger rocks over the conglomerate. At Cape Back a reverse fault has brought the conglomerate against Lower Palaeozoic rocks in the northwest (Fig. 6). The conglomerate and the apparently unconformably (Christie 1974) underlying Lower Palaeozoic formations are steeply dipping. The conglomerate at Cape Back has been described and interpreted as an alluvial fan deposit (Miall 1978, 1981). Generally, the conglomerate is oligomictic and consists of carbonate clasts up to 22 cm in diameter. Conspicuously rare or absent are quartzite, sandstone and noncalcareous siltstone, all lithologies of Lower Cambrian formations of the region. Rare, thin sandstone beds are present and at Cape Lawrence a sandstone bed in the upper part of the formation contains a large number of leaf impressions. Identified as *Crednaria spectabilis* (Heer) Koch, these leaves are of Paleocene age (G. E. Rouse, pers. comm. 1973). The conglomerate is up to 1000 m thick at Cape Lawrence, less elsewhere.

Two features are of importance when relating the conglomerate to the Eurekan structures:

1. The conglomerate is oligomictic. It does not contain any quartzite or sandstone of the Lower Cambrian Ellesmere Group, which structurally overlies the conglomerate along the Parrish Glacier and Rawlings Bay Thrusts. Thus the conglomerate is not derived from the lower parts of the thrust sheets that overlie it.
2. The conglomerate is not continuous beneath the Parrish Glacier Thrust. At Copes Bay the thrust sheet lies alternately on conglomerate or on Siluro-Devonian formations. Because the elevation of the conglomerate surface is higher than that of the Silurian and Devonian strata and no significant change in clast size was observed, the gap between the conglomerate patches is interpreted to have been caused by erosion prior to thrusting. Thus an interval of time must have passed after deposition of the conglomerate and prior to thrusting.

The following sequence of events is therefore suggested. The conglomerate was deposited in alluvial fans adjacent to an uplift, on which the oldest rocks exposed were Middle Cambrian to Silurian. Uplift slowed or ceased and the conglomerate was eroded. Next, thrust sheets were emplaced over the partially eroded conglomerate at Cape Lawrence and in the southern part of the area. Reverse faulting took place at Cape Back.

Eurekan structures

Two types of Tertiary structural regimes can be distinguished. In the north a strike-slip regime is present in the vicinity of the Judge Daly fault zone, in the south a compressive, thrust fault regime is revealed by the Parrish Glacier and Rawlings Bay Thrusts. The intervening areas are influenced by both regimes and strike-slip movement may be superimposed on thrust faults.

Thrusts

Major thrusts without strike-slip overprint are present between Copes Bay and Dobbin Bay (Parrish Glacier Thrust) and at Cape Lawrence (Rawlings Bay Thrust) (Fig. 1).

Parrish Glacier Thrust. – The Parrish Glacier Thrust is well exposed at the heads of Allman and Franklin Pierce Bays, at both sides of Dobbin Bay and it has also been mentioned by Thorsteinsson (1963) at Copes Bay. The thrust plane dips between 30° and 50° to the north and stratigraphic displacement is in the order of 4500 m. The hanging wall consists of Proterozoic to Silurian beds that dip northward, very slightly undulating, at 20° to 40°. In front of the main thrust are several imbricate thrust sheets with partial stratigraphical sequence and a klippe with an almost horizontal thrust plane. The foot wall consists of Tertiary conglomerate, or towards Copes Bay, of Silurian carbonates. The Tertiary and Silurian beds are flat lying and are very little disturbed except for minor drag folds near the thrust.

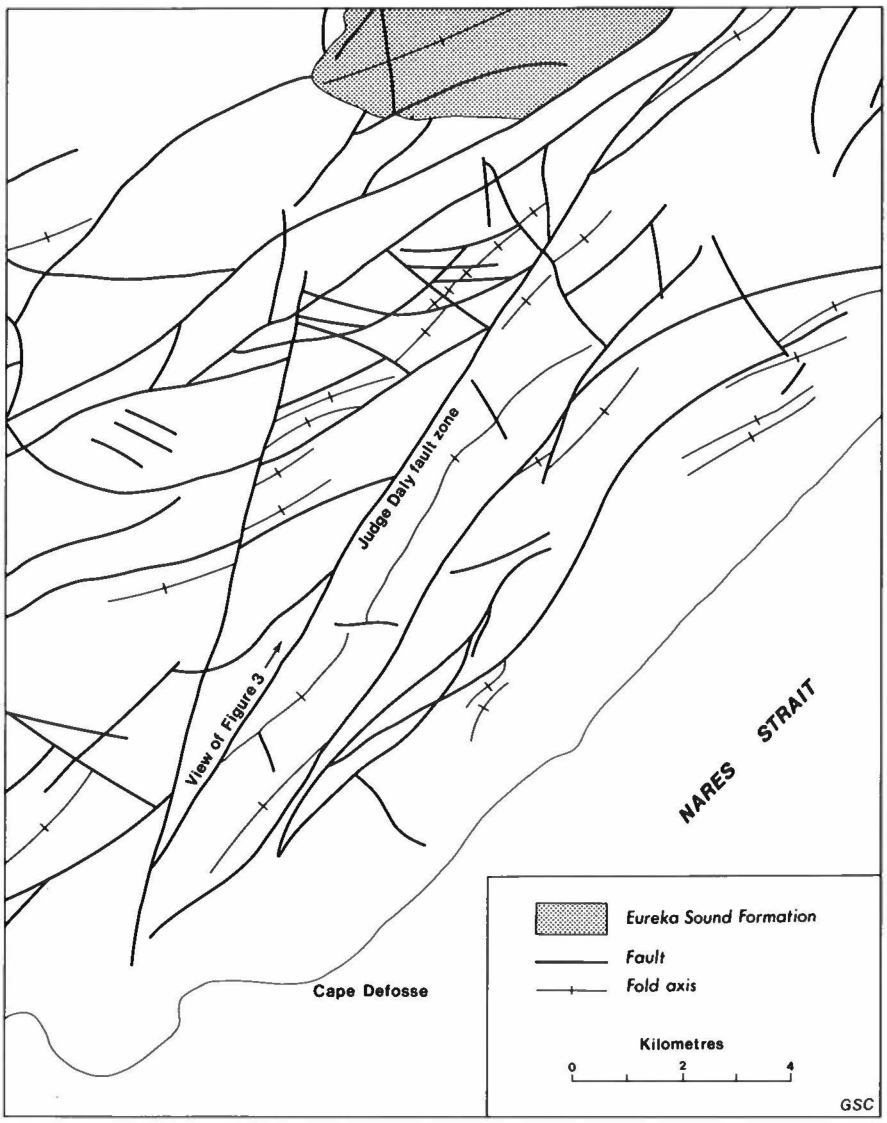


Fig. 2. Main structural features in the Cape Defosse area. For location, see Fig. 1.

Rawlings Bay Thrust. – The structural configuration of the Rawlings Bay Thrust in the Cape Lawrence area is very similar to that of the Parrish Glacier Thrust. The hanging wall dips slightly more steeply to the northwest and in front of the main thrust are an imbricate thrust with a partial stratigraphic sequence and a klippe with a horizontal base.

Strike-slip faults

Two types of strike-slip faulting can be distinguished, divergent strike-slip on northern Judge Daly Promontory and convergent wrenching in the vicinity of Darling Peninsula.

Judge Daly fault zone. – The Judge Daly fault zone is a prominent linear feature on northern Judge Daly Promontory with strong topographic expression (Fig. 2). It is almost 40 km long and is now marked by a linear trench (Fig. 3). Folded Cambrian and Ordovician strata, with fold axes at an acute angle to the fault zone, occupy both sides of the fault in the southern part of the area. In the north, the northwest side of the fault zone consists of Eureka Sound Formation folded into a gentle asymmetric syncline approximately parallel to the fault zone. Drag folds in the Eureka Sound Formation, parallel to the fault scarp, suggest that late vertical movement took place along the fault, downdropping the northwestern side. The main movement, however, is interpreted to be sinistral strike-slip, and this is supported by several lines of evidence.

First there is the presence of larger structures, which are compatible with wrench tectonics. Wilcox et al. (1973) described the attributes of wrench zones, and



Fig. 3. View from helicopter northeastwards along the trench which marks the Judge Daly fault zone (see Fig. 2 for approximate location of photograph). Bedrocks in foreground are folded Ordovician carbonates. The first set of mountains in the background consists of folded Silurian and Devonian carbonates and clastics, while the far mountains, forming the left part of the skyline, consist of folded Cambrian rocks.

singled out en échelon folds and conjugate strike-slip faults as the most important features. No en échelon folds can be identified with certainty. The Ellesmerian fold axes cut the Judge Daly fault zone with an angle of about 25° (Fig. 4) which is close to the most probable angle for left-handed en échelon folds, thus, in this case, the en échelon folds would be parallel to the older folds. At Cape Lieber E–W-striking folds are oblique to Ellesmerian folds. These folds cut the Judge Daly fault zone at an angle of about 60°, too large an angle for true en échelon folds. Such folds, however, could be of Tertiary origin and indirectly related to the strike-slip faulting because of their non-Ellesmerian alignment.

Conjugate strike-slip faults can be identified without difficulty on the northwest side of the southern part of Judge Daly fault zone. These fractures cut Ellesmerian folds and are thus younger, presumably Eurekan. Two sets are present: one is sinistral, striking N10°E and forming an angle of 25° with the Judge Daly fault zone; the second set shows dextral displacement and strikes about N60°W, forming an angle of 100° with the Judge Daly fault zone. The offset is revealed by colour banding of red bed units in the Parrish Glacier and Copes Bay Formations and can be easily observed on airphotos. Ground checking confirmed that the offsets are caused by horizontal, rather than vertical movement. Also fold axes are offset in this area (Fig. 2). The sinistral set of faults represents synthetic strike-slip faults, whereas the dextral set are antithetic strike-slip faults.

The second line of evidence for movement lies in the interpretation of small features. Hematite staining, large pods of brecciation and associated dolomitization of carbonates seem to be confined to the Judge Daly fault zone. Slickensides are very common in the brecciated rocks. Few in situ measurements of slickenside directions could be taken, but some near-horizontal attitudes were recorded. The axes of the Cape Lieber folds trend approximately east–west. In the vicinity of the Judge Daly fault zone the axes bend to the southwest. This deviation from the regional trend is interpreted as drag during left-lateral movement along the fault zone.

The graben containing Eureka Sound Formation to the northwest of the Judge Daly fault zone, caused by downdrop along the zone, is interpreted as a late extensional feature, indicative of divergent strike-slip movement. This interpretation is corroborated by the appar-

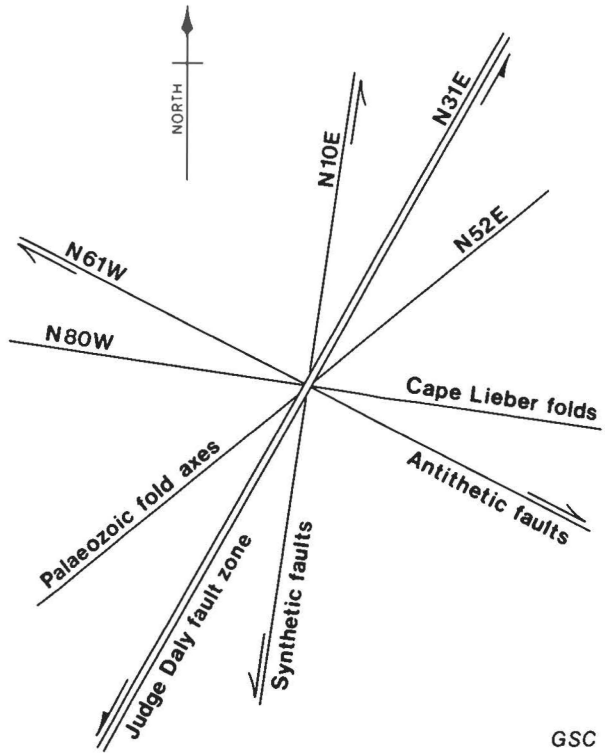


Fig. 4. Plot of directions of the main structural features on northern Judge Daly Promontory.

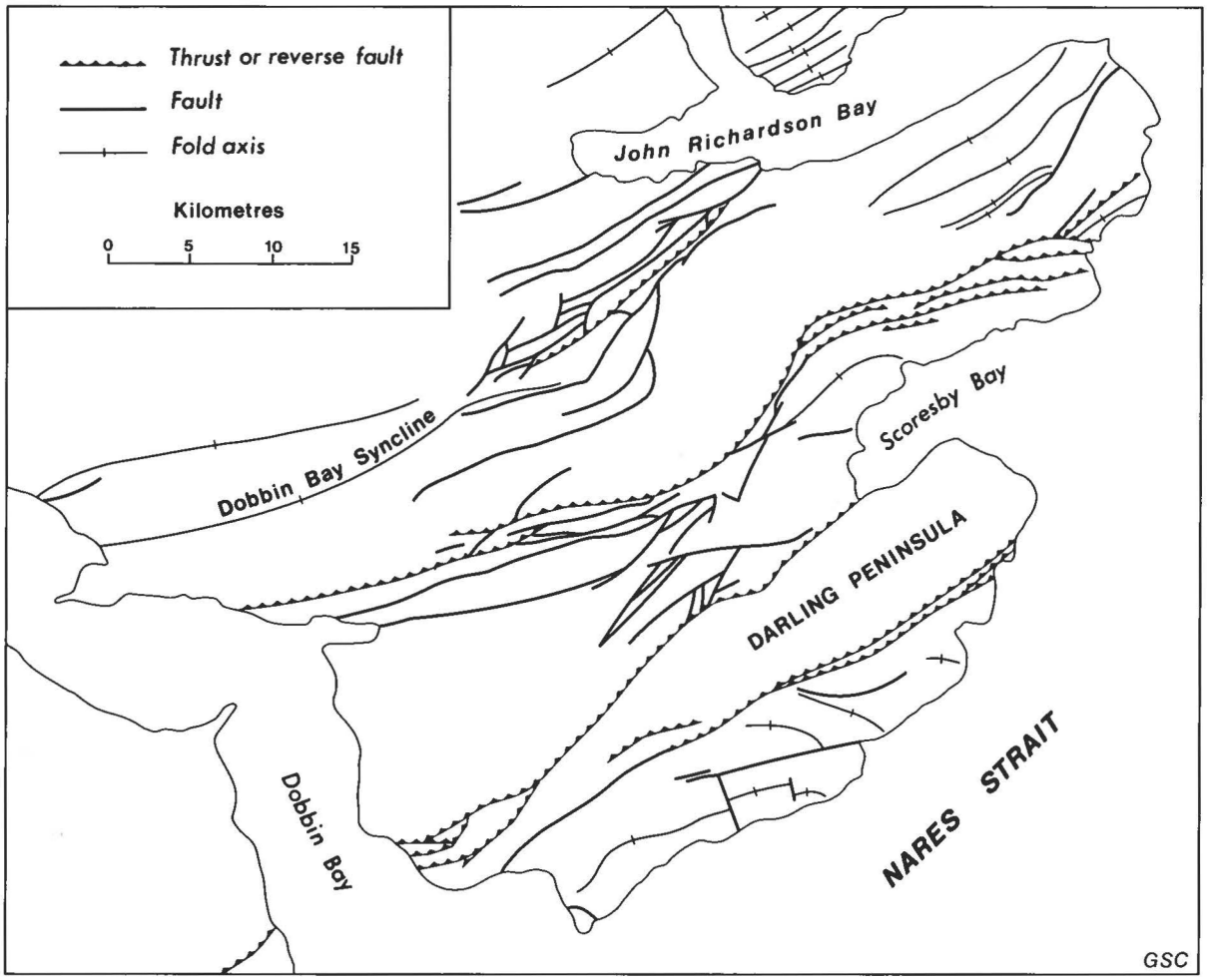


Fig. 5. Main structural features in the Darling Peninsula area. For location, see Fig. 1.

ent absence of thrusts or reverse faults in the vicinity of the Judge Daly fault zone.

Quantitative estimates of displacement along the fault zone are uncertain because of the absence of distinct geological markers and because of the structural fragmentation of the area. The only features found that could be matched tentatively across the fault zone are overturned panels of Lower Cambrian formations. At latitude 81°27′N, longitude 65°00′W, at the eastern side of the Judge Daly fault zone, a northeast plunging anticline (Christie 1974: fig. 1) exposes along its northern flank Rawlings Bay, Kane Basin, Scoresby Bay and Parrish Glacier Formations with an overturned dip of about 70°S. At latitude 81°17′N, longitude 65°45′W, on the western side of the fault zone, a triangular, about 3 km wide fault block consists of a southward (also about 70° and overturned) dipping panel of Rawlings Bay, Kane Basin, Scoresby Bay and Parrish Glacier Formations. If this southern, overturned panel was part of the northern, anticlinal flank, lateral displacement of about 19 km would be needed to realign the two panels. The match however, is uncertain because the anticlinal axis is not contained in the southern fault block. Furthermore no other corroborating alignments across the fault zone are apparent.

Darling Peninsula and adjacent area. – No remnants of Tertiary sediments are preserved in the region between Dobbin Bay and John Richardson Bay (Fig. 5); it is not possible to definitely separate Ellesmerian structures but some inferences can be made. The Ellesmerian structures consist of long, parallel, concentric anticlines and synclines. The Dobbin Bay Syncline appears to be the southeasternmost structure of this type, so that the area to the southeast should be dominated by Eurekan structures. The bedrock there consists of a large number of elongated, northeast-trending fault slices. The slices dip, in part steeply, to the northwest and contain several poorly defined folds. The slices are bounded by thrusts, reverse faults and by faults for which no information on

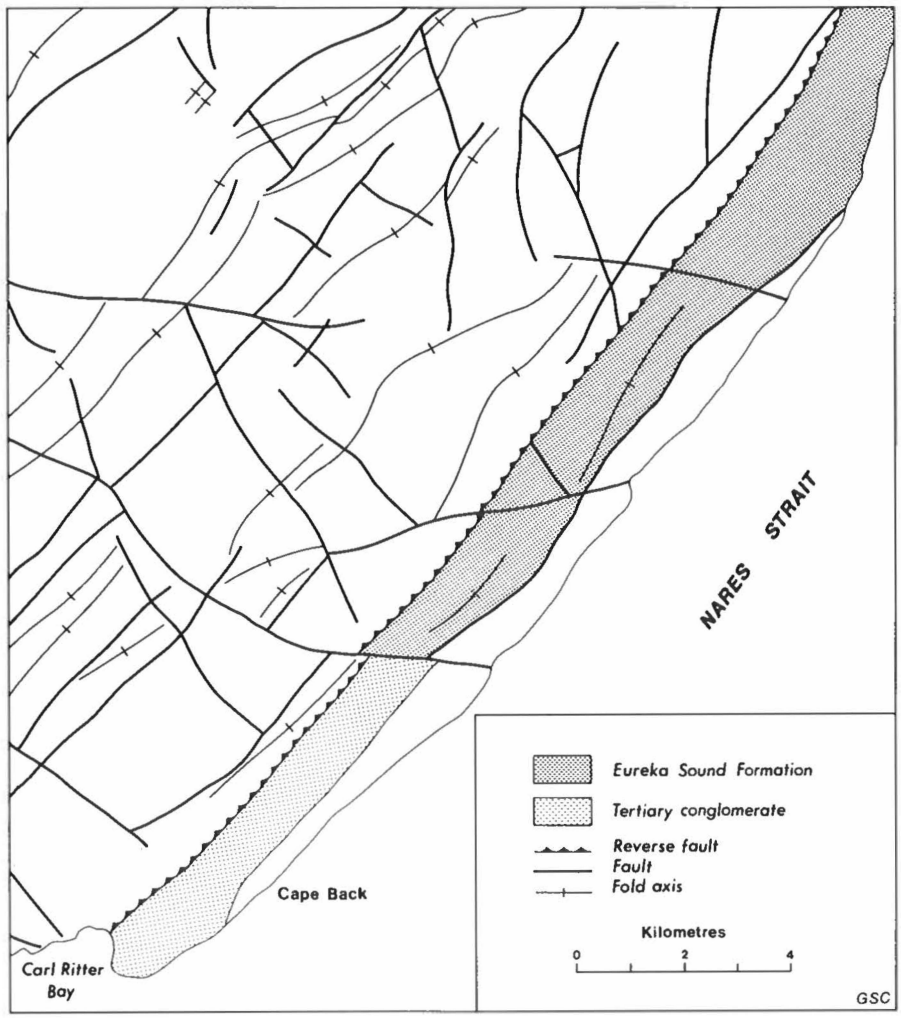


Fig. 6. Main structural features in the Cape Back area. The southeast boundary of the conglomerate is exposed in a steep cliff and has been mapped variously as a stratigraphic contact (Christie 1974) or as a fault (Miall 1981). The present authors tentatively interpret the contact as a folded unconformity. For location, see Fig. 1.

the dip direction of the fault planes is available. Many of the faults have an arcuate trace and terminate with an acute angle against other faults. They are arranged in a braided pattern of synthetic strike-slip faults typical of strike-slip zones world-wide and known from shears of any magnitude (Tchalenko 1970). No conclusions as to actual lateral movement along these faults can be drawn from the presently available information.

Combined thrusts and strike-slip faults

North of Carl Ritter Bay. – In this area folded Lower Palaeozoic rocks are in fault contact with folded Tertiary sediments (Fig. 6). The reverse fault, that forms the contact between the two rock suites, is an extension of the Rawlings Bay Thrust and also projects into the Judge Daly fault zone.

Evidence for thrusting is the reverse fault noted above and its southwestward extension into the Rawlings Bay Thrust. Indirect evidence for strike-slip faulting is found in the post-Ellesmerian deformation of the Palaeozoic rocks to the northwest of the reverse faults. There the Palaeozoic fold belt is segmented by a large number of normal faults. Two fault trends dominate: one is approximately perpendicular and the other about parallel to the main reverse fault. These faults can be interpreted as conjugate strike-slip faults caused by strike-slip movement along the main reverse fault. A few Ellesmerian fold axes appear to be offset along the faults but insufficient observations are available to distinguish between sinistral and dextral displacement along the synthetic and antithetic fault sets.

Discussion

A sequence of Tertiary events can be proposed for part of Ellesmere Island adjacent to Nares Strait, the area between Copes Bay and Cape Baird.

1. Uplift

Paleocene uplifting is indicated by the presence of the conglomerate. Nothing is known of the extent of the uplift which was the source for the conglomerate, but it must have extended at least from Carl Ritter Bay to Copes Bay. Beyond Copes Bay coarse sediments of this nature are not known from the area of the Central Ellesmere fold belt, but they may have existed and have since been eroded. Uplifting was followed by a structural interlude during which the conglomerate was partially eroded. The length of that erosional interlude is not known, but it had ended by late Eocene when the thrusting began.

2. Thrusting

The thrusting can be dated in the Irene Bay area, about 120 km southwest of Copes Bay. There the Parrish Glacier Thrust or extensions of it overlie the Eureka Sound Formation, the youngest strata of which are of middle Eocene age (West & Dawson 1977, 1980). In the Nares Strait area the thrust sheets containing Palaeozoic formations advanced over the conglomerate, but left it undisturbed. In the area north of Carl Ritter Bay, however, where some strike-slip movement also occurred, the conglomerates are folded. This implies a phase of strike-slip movement distinct from earlier thrusting.

3. Strike-slip faulting

The major effects of strike-slip faulting are seen in two areas: on Darling Peninsula and adjacent areas, and on Judge Daly Promontory north of Carl Ritter Bay. These areas are separated at Cape Lawrence by a structural block which was not affected by strike-slip movement. The presence of an intervening block corroborates the concept of separate thrust and strike-slip phases. If the thrusting and strike-slip movement had been caused simultaneously by the same stress-vector, the Cape Lawrence area would be expected to show strike-slip structures. The concept of separate phases is in agreement with some conclusions from geophysical observations. Kristoffersen & Talwani (1977) and Srivastava (1978) postulated earlier compressional and later strike-slip movement in studies of the relationship between Greenland and the Canadian Arctic Islands. These authors dated the phases earlier than in the present interpretation.

The time of the strike-slip movement cannot be fixed, but it is presumed later than the middle-late Eocene thrusting. The main phase of the Eurekan orogeny lasted from Eocene to early Miocene (Balkwill 1978), presumably the events of central eastern Ellesmere Island are related to this orogeny and thus a tentative maximum time range for the strike-slip movement would be late Eocene to early Miocene. The graben on northern Judge Daly Promontory is the last stage of divergent strike-slip movement in that area and should thus have formed during late Oligocene or early Miocene.

Acknowledgements

This account is based on unpublished reports by Sproule Associates Limited, Calgary, Alberta, Canada. The field work and photogeological interpretations for these reports were undertaken by the authors and other members of Sproule Associates Limited during 1972 and 1973. Present proprietor of the data is Norcen Energy Resources Limited, Calgary and the authors thank both companies for permission to use the reports.

C. D. S. de Vries was responsible for most of the structural field mapping and participated in the initial phases of this re-examination. Owing to de Vries' death in a helicopter accident in July 1979, Mayr is solely responsible for the interpretation and synthesis of the geological data.

References

Balkwill, H. R. 1978. Evolution of Sverdrup Basin, Arctic Canada. – Bull. Am. Ass. Petrol. Geol. 62: 1004–1028.

Christie, R. L. 1964. Geological reconnaissance of northeastern Ellesmere Island, District of Franklin. – Mem. geol. Surv. Can. 331: 79 pp.

Christie, R. L. 1974. Northeastern Ellesmere Island: Lake Hazen region and Judge Daly Promontory. – Pap. geol. Surv. Can. 74–1A: 297–299.

Dawes, P. R. & Haller, J. 1979. Historical aspects in the geological investigation of northern Greenland. Part 1: New maps and photographs from the 2nd Thule Expedition 1916–1918 and the Bicentenary Jubilee Expedition 1920–1923. – Meddr Grønland 200(4): 38 pp.

Escher, A. 1970. Tectonic/geological map of Greenland. 1: 2 500 000. – Geol. Surv. Greenland, Copenhagen.

Higgins, A. K., Mayr, U. & Soper, N. J. 1982. Fold belts and metamorphic zones of northern Ellesmere Island and North Greenland. – This volume.

Kerr, J. W. 1967. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Kerr, J. W. 1973a. Geology, Sawyer Bay, District of Franklin. 1: 250,000. – Map geol. Surv. Can. 1357A.

Kerr, J. W. 1973b. Geology, Dobbin Bay, District of Franklin. 1: 250,000. – Map geol. Surv. Can. 1358A.

Kerr, J. W. 1973c. Geology, Kennedy Channel and Lady Franklin Bay, District of Franklin. 1: 250,000. – Map geol. Surv. Can. 1359A.

Kerr, J. W. 1980. Did Greenland drift along Nares Strait? – Bull. Can. Petrol. Geol. 28: 279–289.

Kerr, J. W. 1982. History and implications of the Nares Strait conflict. – This volume.

Koch, L. 1929. The geology of the south coast of Washington Land. – Meddr Grønland 73,1(1): 39 pp.

Kristoffersen, Y. & Talwani, M. 1977. Extinct triple junction south of Greenland and the Tertiary motion of Greenland relative to North America. – Bull. geol. Soc. Am. 88: 1037–1049.

Miall, A. D. 1978. Tectonic setting and syndepositional deformation of molasse and other nonmarine–paralic sedimentary basins. – Can. J. Earth Sci. 15: 1613–1632.

Miall, A. D. 1981. Tertiary sedimentation and tectonics in the Judge Daly Basin, northeast Ellesmere Island, Arctic Canada. – Pap. geol. Surv. Can. 80–30: 47 pp.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its

bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Tchalenko, J. S. 1970. Similarities between shear zones of different magnitude. – Bull. geol. Soc. Am. 81: 1625–1640.

Thorsteinsson, R. 1963. Copes Bay. – In: Fortier, Y. O. et al., Geology of the north-central part of the Arctic Archipelago, Northwest Territories (Operation Franklin). – Mem. geol. Surv. Can. 320: 386–395.

West, R. M. & Dawson, M. R. 1977. Mammals from the Paleogene of the Eureka Sound Formation: Ellesmere Island, Arctic Canada. – Géobios, Mém. spécial 1: 107–124.

West, R. M. & Dawson, M. R. 1980. Paleogene paleontology, stratigraphy and environments, northern Arctic Canada. – Geol. Ass. Can., Min. Ass. Can. Prog. with Abs. 5: 87 only.

Wilcox, R. E., Harting, T. P. & Seely, D. R. 1973. Basic wrench tectonics. – Bull. Am. Ass. Petrol. Geol. 57: 75–96.

The Nyeboe Land fault zone: a major dislocation on the Greenland coast along northern Nares Strait

PETER R. DAWES

Dawes, P. R. 1982. The Nyeboe Land fault zone: a major dislocation on the Greenland coast along northern Nares Strait. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greeenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 177–192.

The name Nyeboe Land fault zone is proposed for a major dislocation zone in the western part of the North Greenland fold belt, traceable along the outer coast for 300 km. Two other names are introduced: 1) the Hand Bugt fault, a reverse fault that juxtaposes strata at least as old as Cambrian to the north, against Silurian rocks to the south, and 2) the Wulff Land anticline, a major fold structure that parallels the fault zone and the outer coast. The Wulff Land anticline is regarded as having formed during the main diastrophism of the Franklinian geosyncline in Devonian time; the Nyeboe Land fault zone is considered an expression of later Tertiary reactivation.

Stratigraphic data are presented that define more accurately than hitherto the Franklinian Cambro-Ordovician platform margin. This roughly coincides with the Nyeboe Land fault zone that probably had a history dating back at least to the early Palaeozoic. The location of a steep Bouguer gravity gradient in the coastal area of Nyeboe Land suggests that a deep-seated crustal structure controlled the margin of the Franklinian trough and the site of the Nyeboe Land fault zone.

The Nyeboe Land fault zone, the Wulff Land anticline and the platform margin can be correlated across Nares Strait with on-line features on Judge Daly Promontory within the central Ellesmere fold belt. This indicates that any strike-slip movement along Nares Strait has only resulted in minor net displacement between Greenland and Ellesmere Island.

The Nyeboe Land fault zone is correlated with the Judge Daly fault zone of Ellesmere Island to define an extensive fracture line that is oblique to Nares Strait. It is concluded that main crustal displacements in the late Phanerozoic probably took place along fracture systems oblique to Nares Strait rather than affecting the separation of Greenland and Ellesmere Island as separate plates.

P. R. Dawes, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark.

There is a general consensus that the Nares Strait lineament is fault controlled, but little as to the nature of the structure. Is the Strait the site of a transcurrent fault (Taylor 1910, Wegener 1924, du Toit 1937, Wilson 1963), a megashear (Carey 1958, Hilgenberg 1966), a transform fault (Wilson 1965), a rift valley (Kerr 1967a), a subduction zone (see Srivastava 1978), a spreading centre (Miall 1981), or even a combination of these structures?

In view of this discussion, it is unfortunate that little direct evidence of the existence of the Nares Strait fault can be found on land in Greenland. However, structural data are available which have a bearing on the tectonic origin of the Strait. It is with this in mind that a fault zone on the North Greenland coast on the shore of the Lincoln Sea is described in this paper (Figs 1 and 2). This important tectonic feature of the North Greenland fold belt is named here the Nyeboe Land fault zone. It strikes from the Robeson Channel over to Peary Land. Its eastern termination is unknown, but it has an onshore extent of at least 300 km.

Furthermore, the coastal area of Nyeboe Land, where the largest continuous stretch of the fault zone is exposed, has been suggested as the site of important Lower Palaeozoic facies changes that represent geological markers used in correlation across the Strait (Christie et al. 1981, Hurst & Kerr, this volume). Since no specific geological description of northern Nyeboe Land exists in the literature, a brief account of the main rock lithologies involved in the fault zone is also presented as a contribution to this discussion.

The Nyeboe Land fault zone coincides with a major anticlinal structure of the North Greenland fold belt. Correlation between Greenland and Ellesmere Island of the fault zone, the anticlinal structure, as well as rock lithologies is suggested and the implications of the correlation to the structure of Nares Strait and to the question of motion along it are explored.

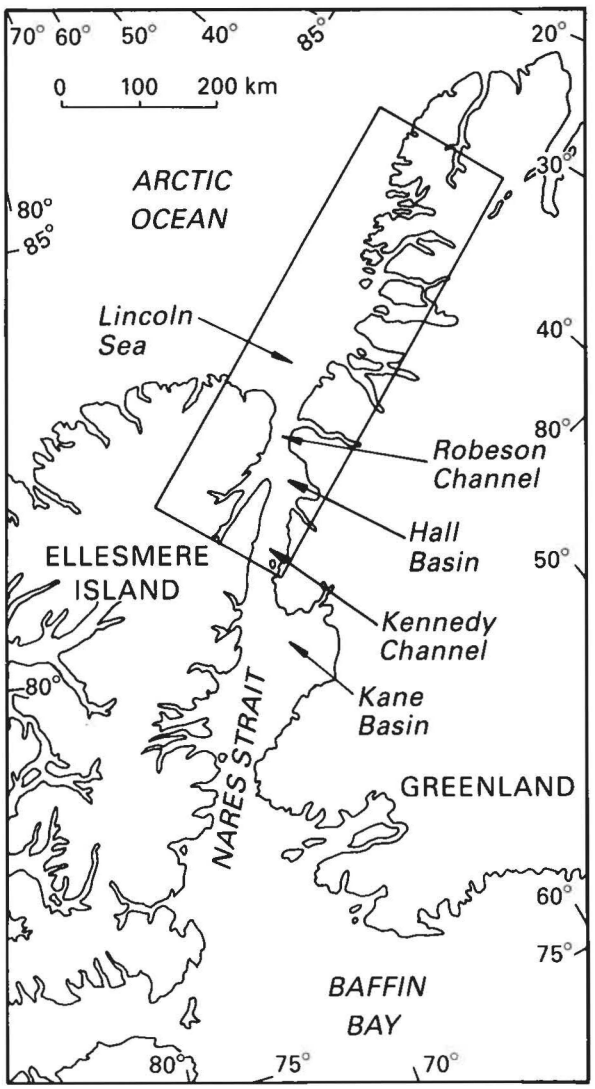


Fig. 1. Index map of the Nares Strait region. The frame marks the area covered by Fig. 7.

Investigations

Geological reconnaissance in the Hall Land – Wulff Land region (Fig. 2) was carried out by the present author in 1965 and 1966 (in 1965 with J. H. Allaart) during Operation Grant Land (Christie 1967). In 1966 several localities showing substantial fault displacements were examined and the fault zone was traced across northern Nyeboe Land. The fault zone was not examined in its entirety and it was mapped to the east by aerial reconnaissance and photogeological interpretation. Some geological data from the region of the fault zone have been included in several regional accounts (e.g. Dawes 1976).

Koch (1920) recognised several faults in north-western Nyeboe Land east of Kap Brevoort and on the basis of aerial photographs Haller (in Haller & Kulp 1962) suggested that important faults occurred in the region between Hall Land and J. P. Koch Fjord (see Figs 4 and 7 for location names). He interpreted the structures to form a Tertiary thrust-fold system, with a distinct southward drive.

Stratigraphy

Proterozoic(?) to Silurian strata occur in the fault zone and the rocks represent varying palaeoenvironments ranging from platform carbonates to trough clastics. Only an outline of the main strata is given here in so far as they help to interpret the structure of the fault zone and its relationship to the Franklinian platform margin. The general stratigraphy of the fault zone is summarised in Table 1.

Hall Land

In northern Hall Land a well-exposed cliff section exposes two successions of markedly contrasting character (Fig. 3); a lower carbonate group, interpreted as a platform facies, and a thick overlying clastic group of typical trough facies. This section has been examined to the east of Kap Ammen (Allaart 1965).

The *carbonate group* is about 550 m thick composed of four unnamed formations. Its base is not exposed. Two dark units, seen on Fig. 3, the lower at sea level, are composed of mainly fine-grained, dark grey, mottled dolomitic limestones and these alternate with units of rather massive light grey limestones. Thin-bedded dark limestones, shales and cherts occur in the uppermost part. In the western part of the cliff section at Kap Porter an isolated carbonate mass with irregular bedding resembles a reef (Dawes & Haller 1979); this is now interpreted as a platform margin slumped mass (Hurst & Kerr, this volume).

Faunal evidence indicates an Upper Ordovician to Lower Silurian (middle Llandovery or later) age for the carbonate group (J. S. Peel in Dawes 1976).

The *clastic group* has a total thickness that probably exceeds 1500 m; its top is not seen. The group is composed of a complex succession of turbidites: sandstones, siltstones and shales. Near the base there are some cherts, dark platy limestones and thin limestone beds. Towards the exposed top thin beds of limestone, often brecciated, come in. The sandstones are calcareous greywackes in the sense of Pettijohn (1954) and they have a characteristically high carbonate content; about 50 per cent carbonate in total rock mineral composition.

From field and photogeological work five map units have been erected and these have been mapped in Nyeboe Land within the main fault zone (Figs 3 and 4).

The age of the clastic group in Hall Land is Lower Silurian to latest Silurian (Pridoli) or ?early Devonian

Table 1. Geological history of the Nyeboe Land fault zone.

		MAIN SEDIMENTARY AND METAMORPHIC ROCK			TECTONISM AND METAMORPHIC REGIME
CENOZOIC	QUAT.	Glacial, fluvio-glacial and marine gravel, sands and silt			Uplift, erosion Isostatic movement
	TERTIARY	Quartz-calcite veins, breccias, mylonite			Normal faulting Thrust faulting, folding, mylonitisation, hydro-thermal activity
UPPER PALAEOZOIC –MESOZOIC	CARBONIFEROUS –CRETACEOUS	(Sverdrup Basin – Wandel Sea Basin deposition, major marine transgression. No rocks yet known in Nyeboe Land fault zone)			
	DEVONIAN	Quartz-calcite veins Chlorite-muscovite slate, psammite			Uplift, erosion Extensional faulting Polyphase folding, minor thrusting, low-grade metamorphism
		Hall Land	Nyeboe Land	Wulff Land	
LOWER PALAEOZOIC	SILURIAN	Calcareous greywacke, siltstone, shale Dark platy limestone, chert Carbonate buildup	Calcareous greywacke, siltstone, shale, chert-pebble conglomerate Calcareous slate	Calcareous greywacke, siltstone, shale, chert-pebble conglomerate	
	ORDOVICIAN	Dark and pale dolomitic limestone	Variable limestones and breccias ?	Dolomitised siltstone, shale, carbonate breccio-conglomerate	Fluctuating position of platform margin possibly associated with fault movements
	CAMBRIAN		Arkose, quartz sandstone, shale Variable limestones Dark bituminous limestone, breccia, chert, shale Dark limestone Pale limestone	Sandstone, arkose, conglomerate, siltstone, shale Pale dolomite	
PREC.	PROT.		?	Black shale Sandstone, conglomerate, siltstone	Subsidence – ? faulting

Bendix-Almgreen & Peel 1974, Berry et al. 1974, Dawes 1976, Surlyk et al. 1980).

Nyeboe Land – Hendrik Ø

The main fault of the Nyeboe Land fault zone (Hand Bugt fault) separates northern and southern areas of contrasting stratigraphy (Figs 4 and 5).

The *southern area* is composed of map units 3, 4 and 5 of the clastic group (Fig. 4). Lithologies are generally comparable with the Hall Land succession although coarser-grained rocks occur, and pebbly sandstone and chert-pebble conglomerates (Hendrik Conglomerate of Dawes 1966) form thick incursions in the uppermost part of the sequence. Some graptolite-bearing shales stratigraphically above the Hendrik Conglomerate are

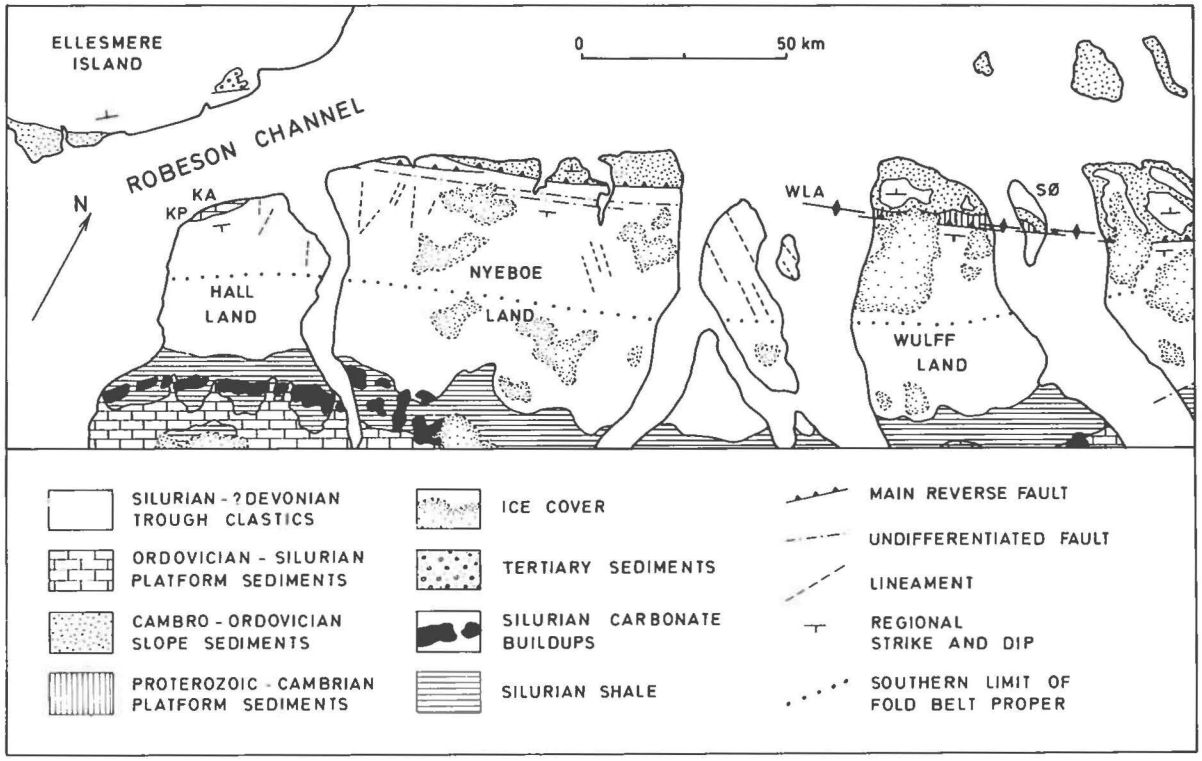


Fig. 2. Geological map of the Hall Land – Wulff Land region showing the Nyeboe Land fault zone. KP = Kap Porter, KA = Kap Ammen, SØ = Stephenson Ø, WLA = Wulff Land anticline.

of Upper Silurian (early Ludlow) age (Berry et al. 1974).

The fault block south of Repulse Havn (Fig. 4) is composed of map units 1 and 2 as well as older carbonate strata. The oldest rocks observed are a variable limestone succession containing dark, often bituminous, limestones, light limestones with some breccio-conglomerates, and calcareous slates. These limestones have yielded an indeterminate crinoid-coral fauna indicating a post-Cambrian, probably Ordovician age.

In the *northern area* a faulted and highly folded succession of unknown but undoubtedly large thickness is composed of two main rock sequences. These are a Cambrian mainly carbonate sequence and a clastic sequence of probably younger age.

The carbonate sequence, at least 600 m thick, is composed of pale limestone and breccio-conglomerates, dark, often bituminous, limestones, some of which are black, dense and dolomitic, banded typically yellow and grey 'tiger' limestones and a variety of mainly dark, thin-bedded limestones, calcareous shales and slates. Some thin chert and siliceous shale beds occur. The carbonate breccio-conglomerates are very common and a frequent type is intraformational breccias locally and directly derived from banded 'tiger' limestones. Dark bituminous limestones have yielded Lower and Middle Cambrian trilobite faunas (Poulsen 1969, Peel 1974, 1979).

The clastic sequence forms some large outcrop areas (Fig. 4) as well as tracts interfolded with the carbonate sequence. The clastic sequence is at least 300 m thick, possibly considerably more, and is composed of dark, in places rusty, weathering and sulphide-bearing rocks, shales, slates, quartz sandstones, arkoses with some quartzite, and sandy conglomerate beds containing shale clasts. Impure limestone and calcareous sandstones also occur. The rocks are of low metamorphic grade and commonly have a waxy sheen due to secondary mica and in places chlorite.

Only simple trace fossils are known from the clastic sequence and its age is uncertain. The strata in this region are intensely folded and faulted, but way-up criteria at several localities suggest that a substantial thickness of the clastic rocks is younger than the carbonates. In the Hand Bugt – Frankfield Bugt area the clastic sequence dips north and overlies the carbonate sequence in what appears to be an undisturbed stratigraphical relationship (Fig. 5). The clastic rocks show some graded bedding and bottom structures suggesting it to be a turbiditic suite, but the rocks contrast with the Silurian turbidites south of the Hand Bugt fault in several characters, notably in mineral composition.

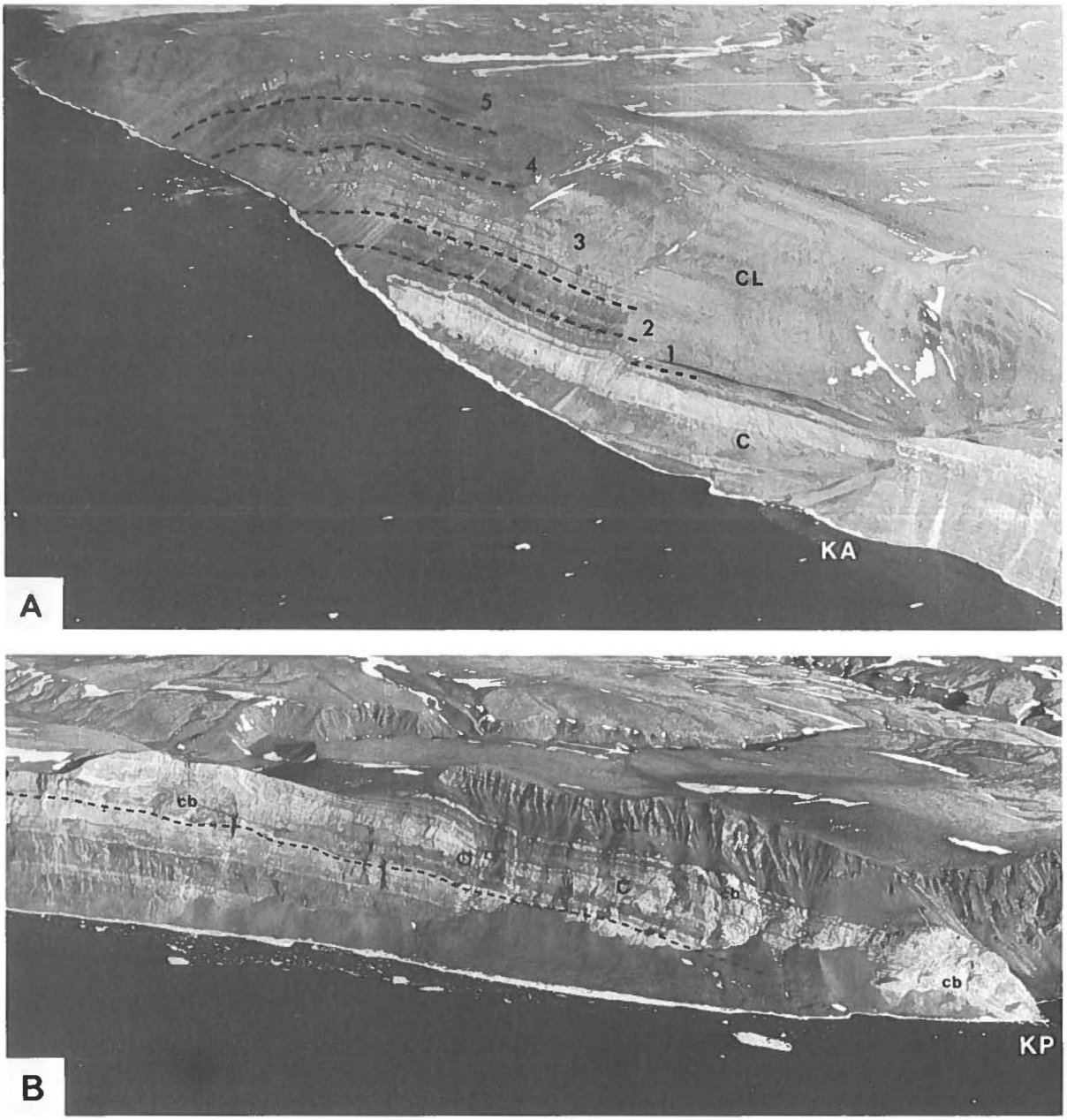


Fig. 3. Cliffed coast of northern Hall Land: east (A) and west (B) of Kap Ammen. Upper Ordovician – Lower Silurian carbonate group (C) overlain by Silurian clastic group (CL). A: the five map units of the clastic group are indicated, B: stratigraphic irregularities in the carbonate group due to tectonic dislocation (dashed line) and carbonate buildups (cb). Highest part of the cliffs in both views is above 700 m. KA = Kap Ammen, KP = Kap Porter. Aerial photographs 546 K–S, A: no. 2198, B: no. 2193; copyright Geodætisk Institut, Denmark.

The clastic rocks are correlated with the turbidites of Formation A or the older Polkorridoren Group of Peary Land (Dawes & Soper 1979). These rocks are now known to be of Cambrian age (Hurst & Surlyk 1980, J. M. Hurst, pers. comm.). Since the clastic rocks in Nyeboe Land overlie carbonates as young as Middle Cambrian, correlation with the younger Formation A is favoured.

Wulff Land

In Wulff Land on the north side of the Hand Bugt fault, a thick and variable succession contains Proterozoic(?)

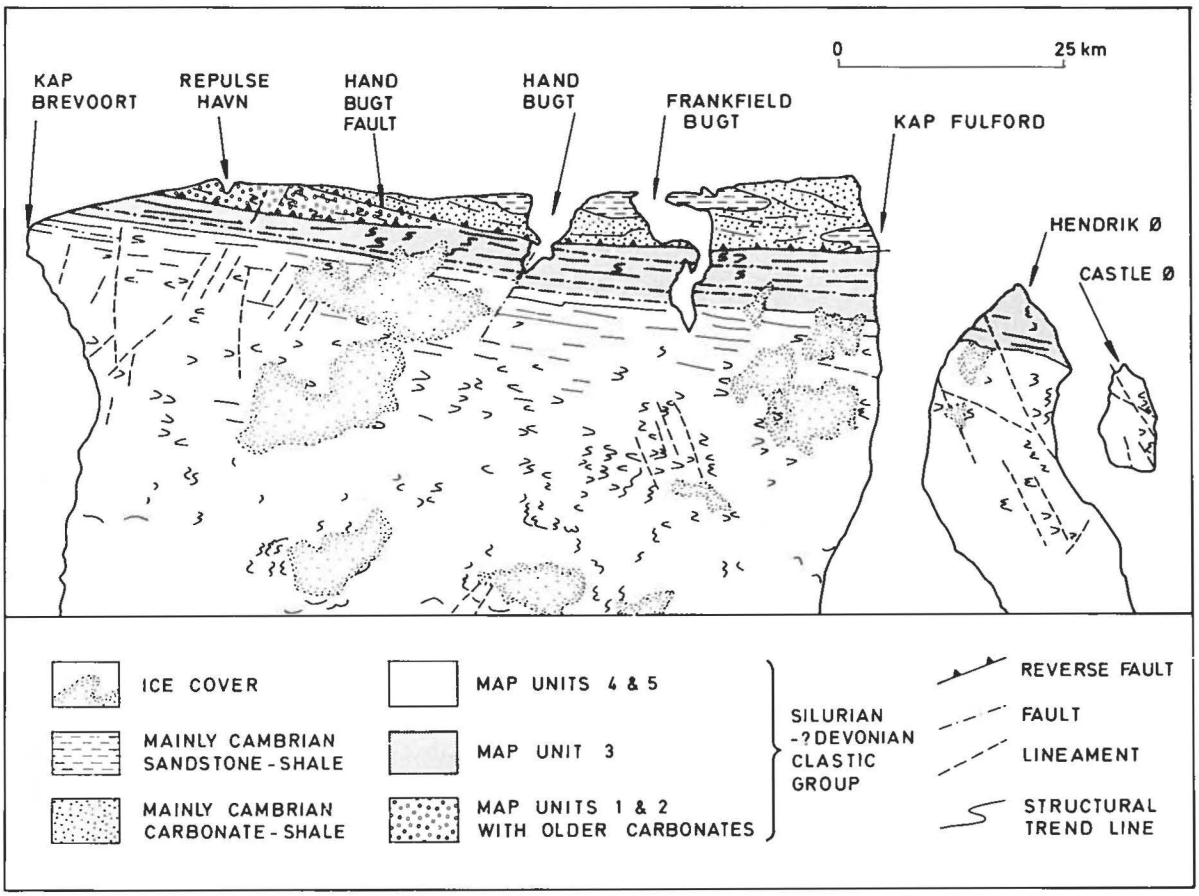


Fig. 4. Geological map of the Nyeboe Land fault zone, Nyeboe Land and Hendrik Ø.

to Silurian rocks. The base of the succession is not seen (cf. Koch 1920) and a sequence of at least 400 m of sandstone, siltstone and partly dolomitic calcareous and quartzitic conglomerate, with dark shale beds at the top, are the oldest rocks exposed. Overlying this is a pale massive dolomite at least 350 m thick that is regarded as a possible equivalent of the earliest Cambrian Portfjeld Formation known from the platform of Peary Land (J. S. Peel, pers. comm.). An overlying 450 m thick sequence of arkosic sandstone, quartz-pebble conglomerate, siltstone, bioturbated mudstone and black shale is reminiscent of the Buen Formation also described from Peary Land. These rocks are overlain by a sequence of dolomitised siltstones and shales and dolomitised breccio-conglomerates regarded by J. M. Hurst (pers. comm.) as equivalent of Formation B of Dawes & Soper (1979). The top of the succession is composed of Silurian turbidites which correlate with the clastic group of Hall Land, Nyeboe Land and Wulff Land, south of the Hand Bugt fault.

Sedimentary facies distribution

The main lithologies of the fault zone are summarised in Table 1 and an interpretation in terms of general depositional environment is given in Fig. 6. A relatively complete and stratigraphically undisturbed succession through the Franklinian geosyncline is preserved in the study region in northern Wulff Land. There a major transgressive cycle can be recognised involving the Proterozoic and Cambrian platform clastics and carbonates, through the Ordovician slope mudstones and carbonate breccio-conglomerates to the Silurian trough turbidites. In the Hall Land – Nyeboe Land region two transgressive cycles are indicated with the Lower and Middle Cambrian platform? and slope carbonates being followed by trough turbidite sedimentation. Platform sedimentation was again prevalent in Upper Ordovician time, or earlier, after which a second episode of trough turbidites reached the region and persisted throughout the Silurian.

The fluctuating position of the platform–trough margin illustrated by the Hall Land, Nyeboe Land and Wulff Land sections coincides in a general way with the



Fig. 5. View of the Nyeboe Land fault zone looking east across Nyeboe Land, showing the Hand Bugt fault (dashed line) and the steeply-dipping linear belt south of it. 1: Late Cambrian(?) clastic sequence, 2: Cambrian carbonate-shale sequence, 3: Silurian turbidites. Aerial photograph 546 E–Ø, no. 11577; copyright Geodætisk Institut, Denmark.

site of the Nyeboe Land fault zone. This suggests that the fault zone may have had a fundamental influence on the depositional history of the region. This conclusion gains some support from the sedimentary facies model suggested further to the east in Peary Land (Surlyk et al. 1980).

The southernmost known extent of the Cambro-Ordovician slope deposits is plotted in Fig. 7. The eastern extension of the fault zone is unknown but it is not apparent as a cross-cutting post-depositional fracture in the region between J. P. Koch Fjord and Frederick E. Hyde Fjord (A. K. Higgins, pers. comm.). However the fault zone does strike towards the platform edge – slope boundary as defined in the Frederick E. Hyde Fjord region of Peary Land (Dawes & Soper 1979, Hurst 1979). Hurst & Surlyk (1980) and Surlyk et al. (1980) have suggested that the Lower Palaeozoic platform–slope boundary in Peary Land is controlled by E–W-trending fault systems and the main syndepositional fault (Navarana Fjord fault) is placed in central Peary Land immediately south of Frederick E. Hyde Fjord (Surlyk et al. 1980: fig. 1). The Nyeboe Land fault zone is more or less on line with the proposed site of this master fault in central Peary Land. To the west, the Navarana Fjord fault strikes across North Greenland south of the Nyeboe Land fault zone. The relationship of the Nyeboe Land fault zone to the syndepositional faults envisaged by Hurst & Surlyk (1980) and Surlyk et al. (1980) is not yet clear but the Nyeboe Land fault zone could well represent a major fracture in the en echelon system of faults proposed by the above authors.

Of further interest is the fact that the Silurian turbidites outcropping across the whole of North Greenland for over 600 km have a persistent palaeocurrent transport direction from the east (see Surlyk, this volume) which has led Hurst & Surlyk (1980) to assume the presence of a northern barrier that influenced the direction of turbidite flow. Such a well-defined barrier may well have been a faulted uplift. The E–W-trending turbidite trough across western North Greenland is

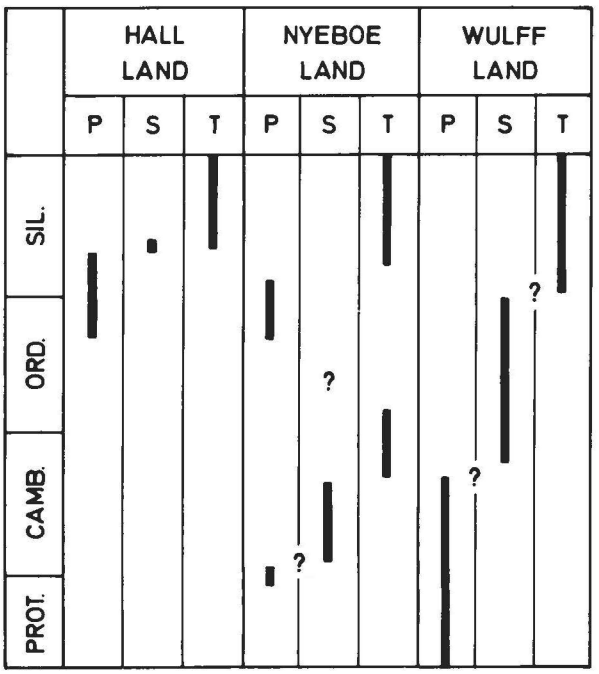


Fig. 6. Schematic representation of platform (P), slope (S) and trough (T) facies of the Nyeboe Land fault zone region.

preserved south of the Nyeboe Land fault zone, and this raises the question of whether the fault zone could also have had influence on the deposition of the Silurian turbidites. Thus the tectonic history of the Nyeboe Land fault zone should be studied in the light of possible influence on sedimentation throughout the whole of the Lower Palaeozoic.

Structure of the fault zone

The extent of the Nyeboe Land fault zone is seen in Fig. 7. It strikes approximately ENE, more or less parallel with the regional structural trend of the North Greenland fold belt and to the outer coast of Nyeboe Land. It consists of several high-angle, sub-parallel strike faults. Some WNW-trending lineaments crossing Hendrik Ø, Castle Ø and the coastal area of eastern Nyeboe Land may represent splay-out faults of the main fault zone. In addition to the main faults shown in Fig. 4, Haller (1971: fig. 157) recognises faults with a more easterly trend, i.e. oblique to the regional stratification. No field data are available at this time to confirm that this trend is a main fracture system.

The most conspicuous dislocation of the fault zone is a reverse fault that has a steeply-dipping (>75°) to vertical fault plane, the dip direction of which is mainly to the north, although steep dips to the south have been observed. This master fault is named the *Hand Bugt fault,* after the bay where it is well exposed (Figs 5 and 8). This fault separates rock successions of profoundly different ages: to the north an uplifted block contains strata at least as old as Cambrian while to the south Silurian clastic rocks occur. In the area to the west of Hand Bugt where Middle Cambrian rocks are in fault contact with Silurian strata, the regional trend of the strata in the uplifted northern block is discordant to the fault and, at the present level of exposure, strata are cut out towards the east.

The strata of the northern uplifted block are highly deformed and a variety of fold styles exist; the majority have steeply inclined limbs. In many places near the Hand Bugt fault the rocks are steeply-dipping and strata are commonly crushed and mylonitised at the actual

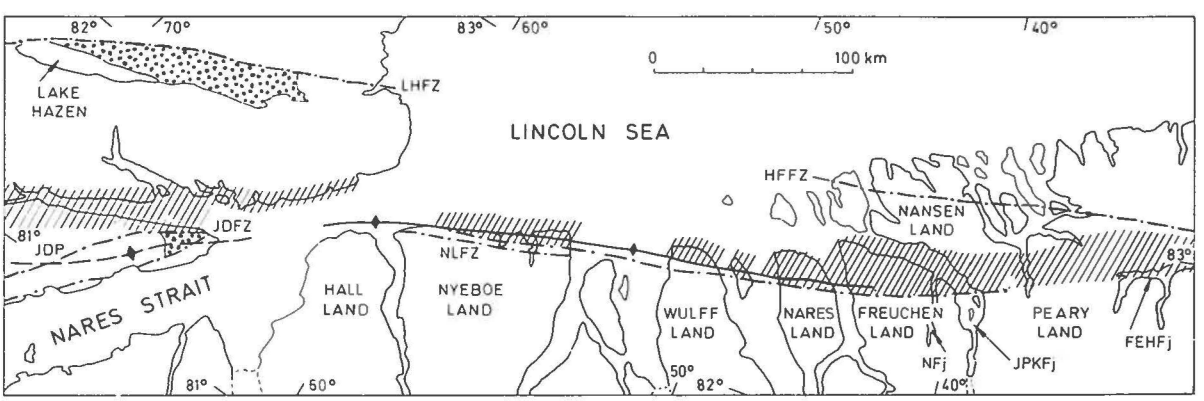


Fig. 7. Map of western North Greenland and north-eastern Ellesmere Island. Full black line marks the trend of the Wulff Land anticline; dashed line marks the position of the Judge Daly anticlinorium; cross-hatching represents the southern outcrops of the Cambro-Ordovician slope deposits, including Formation B of Greenland and Hazen Formation of Ellesmere Island. The dot ornament represents the Permian to Tertiary Sverdrup Basin deposits; only the main Tertiary outlier on Judge Daly Promontory (JDP) is shown. NLFZ = Nyeboe Land fault zone, JDFZ = Judge Daly fault zone, HFFZ = Harder Fjord fault zone, LHFZ = Lake Hazen fault zone, NFj = Navarana Fjord, JPKFj = J. P. Koch Fjord, FEHFj = Frederick E. Hyde Fjord.



Fig. 8. The Hand Bugt fault, Nyeboe Land viewed from the east. Steep, northerly-dipping fault plane with Silurian sandstone and shale (S) to the south, up-faulted Cambrian limestone, carbonate breccio-conglomerate and shale (L) to the north. Some folding in the up-faulted block is due to the fault movement. Height of the section above the river is about 300 m.

fault plane. The rocks are commonly brecciated and severely invaded by anastomosing calcite-quartz veins. Drag folds occur in the northern rocks at the fault plane (Fig. 8) and east of Frankfield Bugt these folds plunge to the east. Several other faults and crush zones occur cutting up the highly folded rock succession north of the Hand Bugt fault, but none of these have been traced for any distance.

The Hand Bugt fault is bounded to the south by an impressive steeply-dipping linear belt of upturned Silurian strata traceable throughout Nyeboe Land (Fig. 5) where it is up to 4 km wide, and across northern Hendrik Ø and onto Wulff Land. Within this linear belt of vertical to steeply-dipping strata several high-angle strike faults occur and, in some cases, recessive shale units appear to have acted as the dislocation surfaces. Many of the faults have steep southerly-dipping planes and it is difficult to identify such high-angle strike faults due to the concordance of fault planes and bedding. Apparently, these faults have not disturbed the essential stratigraphic order of the Silurian succession. It is uncertain if the faults represent fractures of the same type as the Hand Bugt fault or whether they represent southerly-dipping normal faults developed during or after the regional upturning.

In Hall Land no evidence of the Hand Bugt fault with its uplifted northern block is present, although the high, steep coastline suggests fault control (Fig. 3). However, in the carbonate sequence exposed in the sea cliffs near Kap Ammen and Kap Porter, there is thickening of the sequence which is regarded as repetition of stratigraphy due to fault tectonics. This structure is interpreted as the southerly expression of the Nyeboe Land fault zone.

In Wulff Land, the Hand Bugt fault truncates the southern limb of a major anticlinal structure and juxtaposes a northerly-dipping correct way-up succession of Proterozoic(?) to Silurian strata, in the uplifted northern block, against a southerly-dipping sequence of Silurian and (?) older rocks to the south. The sequence north of the reverse fault contains conspicuous black shale units, the outcrop pattern of which in northernmost Wulff Land suggests that several faults transect the area.

Between Nares Land and Peary Land (Fig. 7) the Hand Bugt fault is visible on aerial photographs as a topographical depression, although, due to widespread ice cover, particularly on Freuchen Land, little is known about the consistency of the faulting throughout the region. The reverse fault is well exposed on the east side of Navarana Fjord (J. M. Hurst, pers. comm.). In this area it juxtaposes, on the north side, a dark sequence of siltstones, mudstones and carbonate breccio-conglomerates (correlatable with Ordovician Formation B of Dawes & Soper 1979) against Silurian clastic rocks. The fault forms a distinct step in the coast on the east side of J. P. Koch Fjord and it probably coincides farther east with the conjectural extension of the syndepositional Navarana Fjord fault of Surlyk et al. (1980).

Wulff Land anticline

The major ENE-trending antiform which transects northern Wulff Land (Koch 1920) was visited in 1966 and platform-type Proterozoic(?) and Cambrian rocks were located in its core (Dawes 1976). The name Wulff Land anticline is proposed for this large structure which can be traced for over 200 km between Nyeboe Land and Nares Land (Figs 2 and 7). In Nyeboe Land the structure is highly dissected by the Nyeboe Land fault zone, but regional dips and orientation of small-scale folds indicate its general trend. In this region it is more of an anticlinorium with highly deformed strata in its central area. In the Wulff Land – Stephenson Ø area the anticline plunges to the east, but its continuation across Freuchen Land and towards Peary Land is unmapped although suggested on aerial photographs.

The anticline has normal way-up sequences on both northern and southern limbs, but the southern limb is truncated by the Hand Bugt fault. The anticline is considered an integral part of the mid-Palaeozoic structural pattern of the fold belt although the attitude of the limbs has undoubtedly been affected by the later fault movements.

Tectonic history of the fault zone

The relationship of the Nyeboe Land fault zone to both folding and regional metamorphic and hydrothermal mineral composition of the rocks is briefly outlined here. The main tectonic and metamorphic events recognised in the fault zone area are summarised in Table 1.

Folding and metamorphism

In the North Greenland fold belt the intensity of the mid-Palaeozoic deformation and metamorphism seen on a regional scale increases progressively northwards (Fränkl 1955, Dawes & Soper 1973).

The location of the fault zone relative to the fold intensity is illustrated by the bedding traces shown in Fig. 4. The southern part of the Nyeboe Land fault zone is marked by the sudden incoming of the linear belt of steeply-inclined to vertical strata in which tight to isoclinal folds occur (Fig. 5). This upturned linear belt is considered to have been formed during the fault movements that resulted in the uplift of the northern coast area along the Hand Bugt fault. The fault movements affected the attitude and style of the folds of this linear belt.

The folds within the northern uplifted block vary from rather tight to steeply-inclined structures to less frequent, more open folds with moderate dips. The cleavage-bedding relationships suggest that at least two main ages of folds occur. Minor thrusts and faults occur on fold limbs and quartz-calcite veins are associated with fractures and tension gashes. Although essentially a strike fault, the Hand Bugt fault is discordant in detail to the fold pattern and regional stratification of the Cambrian rocks on the north side of the fault (Fig. 5).

The fault zone coincides in a regional sense with the boundary of the incoming of metamorphic mica (Dawes 1976). The highest grade rocks in northern Nyeboe Land have both muscovite and chlorite in equilibrium with quartz and carbonate indicating a metamorphic grade corresponding to the low greenschist facies. Penetrative flow (slaty) cleavage dominates the argillaceous lithologies.

The northerly prograding metamorphic pattern seen elsewhere in the fold belt is disrupted by the Hand Bugt fault, where there is an abrupt increase in metamorphic grade. Hence, in the Kap Fulford area, Cambrian(?) turbidites with a distinct waxy sheen uplifted along the fault are in juxtaposition with relatively unmetamorphosed Silurian turbidites of similar lithology to the south. The Hand Bugt fault thus clearly post-dates the development of the regional thermal gradient that resulted in greenschist metamorphism.

In many places in the northern area north of the Hand Bugt fault, quartz-calcite veining is common in areas of crushing and brecciation. Veins have a variety of forms and thicknesses and they can be irregular and anastomosing. In places, some veins are deformed and show slickenside surfaces, and cross-cutting relationships between veins exist, providing evidence of repeated crushing and brecciation.

A summary of the above data, together with those given earlier in the description of the fault zone, indicates the following relationship of the faulting to the tectonic and metamorphic events.

1) The Nyeboe Land fault zone post-dates at least two ages of folds and associated cleavages, the main low-grade regional metamorphism and some fracturing and quartz-calcite veining.
2) The Nyeboe Land fault zone is associated with compressive uplift that influenced the style and regional attitude of the earlier folded rocks, and it was accompanied by brecciation, crushing and intense hydrothermal veining, as well as mylonitisation and folding.

Age and dimension of the fault movements

It has long been established that in both North Greenland and Ellesmere Island two major orogenies affected the Franklinian geosyncline, i.e. the mid-Palaeozoic Devonian–Carboniferous) Ellesmerian orogeny and the Tertiary Eurekan orogeny. In contrast to Ellesmere Island, where a thick late Palaeozoic, Mesozoic and Tertiary succession is preserved in the Sverdrup Basin (Balkwill 1978), no post-Devonian strata have yet been discovered in the Nyeboe Land region by which the Tertiary tectonic events can be readily differentiated.

However, to the east in Peary Land, where post-orogenic strata exist (Wandel Sea Basin), Tertiary tectonic and metamorphic events can be distinguished (Dawes & Soper 1973, Soper et al., this volume).

The Nyeboe Land fault zone post-dates the mid-Palaeozoic deformation and metamorphic pattern of the fold belt and the present author follows Haller (1971, Haller & Kulp 1962) in regarding the faulting as of Tertiary age. In view of the lack of Sverdrup Basin – Wandel Sea Basin deposits in this region, no direct evidence is yet available of late Phanerozoic tectonism. The age adopted here is based on regional chronological correlation with the adjacent areas both to the east (Peary Land) and west (Ellesmere Island). It would be a striking anomaly if the Nyeboe Land region represented the only large segment of the Innuitian Province that had escaped tectonic overprinting. However, the fault zone probably represents the site of a much older fracture associated with the southern margin of the Franklinian geosyncline as implied by the Lower Palaeozoic sedimentary facies distribution described in this paper. Thus, one explanation of the geometry of the Hand Bugt fault might be that an original syndepositional normal fault at the trough margin was later reactivated in the Tertiary and essentially transposed into a high-angle reverse fault. Evidence that the Nyeboe Land fault zone occupies the site of an ancient structure — probably related to a fundamental change in thickness of the crust — is discussed later in the section describing the gravity field of northern Nares Strait.

The main movement along the Nyeboe Land fault zone has been uplift of northern strata along steep fault planes. The net throw along the Hand Bugt fault is considerable since Cambrian and Silurian strata are in juxtaposition. Nevertheless, the strata were highly folded prior to the faulting and the reverse fault coincides with the Wulff Land anticline (Figs 2 and 7). Thus the actual net displacement is uncertain. However, since all, or perhaps the greater part of, the Ordovician succession (known to be present in areas to the west and east) is cut out, a movement of at least 1 km, and possibly considerably more, is indicated.

The linear belt marking the southern limit of the fault zone is regarded as the outer welt of the Tertiary tectonism that affected this part of the North Greenland fold belt. As mentioned earlier the precise nature of the steep strike faults present within this belt is uncertain, although it seems likely that normal fault movements (slides) may have occurred along the southerly-dipping fault planes in response to the regional upturning.

Correlation across Nares Strait

The Nyeboe Land fault zone is only 25 km distant from north-eastern Ellesmere Island across the Robeson Channel — the narrowest part of Nares Strait. The geological features described in this paper are major regional elements of the Franklinian geosyncline and thus they are to be expected in some form on the north-western side of Nares Strait. Corresponding Lower Palaeozoic facies, and structures comparable to the Nyeboe Land fault zone and Wulff Land anticline occur on Judge Daly Promontory of Ellesmere Island (Fig. 7).

Lower Palaeozoic facies distribution

The close geological similarity between Hall Land and north-eastern Ellesmere Island has been long established and the opposite coasts of Robeson Channel are composed of folded and on-strike Silurian trough turbidites characterised by calcareous greywacke. Trettin (1979) has referred these Silurian strata of the Hall Land – Nyeboe Land region (the clastic group of this paper) to the Imina Formation of Canada. In addition, the older rock sequences exposed in the Nyeboe Land fault zone show striking lithological similarity to the platform, slope and trough sequences of the Judge Daly Promontory region described by Kerr (1967b, 1968), Trettin (1971), Christie (1974) and Trettin & Balkwill (1979).

Hurst & Kerr (this volume) have dealt with the correlation of certain Lower Palaeozoic facies across Nares Strait including a suggested on-line correlation of the carbonates at Kap Ammen, Hall Land, with the platform margin (horst block?) that can be traced through Judge Daly Promontory and to the south-west through Ellesmere Island. This platform margin was a stable feature during the Upper Ordovician and Lower Silurian and it separates the platform carbonates to the south-east from a variable sequence of clastic rocks to the north-west (Trettin 1979, Trettin & Balkwill 1979). Downflexing of this platform margin occurred in late Lower Silurian time and trough turbidites in both Ellesmere Island and Greenland covered the carbonates, as illustrated by the Hall Land section (Fig. 3). It is of interest that the present-day southern limit of the Silurian turbidites (Imina Formation), although affected by erosion, is also essentially on-line across Nares Strait from central Hall Land to Judge Daly Promontory.

The recognition of Cambro-Ordovician platform, slope and trough deposits in an uplifted fault block makes it possible to locate the earlier position of this platform margin more accurately in Greenland than hitherto. This margin was located in Cambro-Ordovician time in northern Nyeboe Land at the site of the Nyeboe Land fault zone. Extensive cover of the Silurian turbidites in and south of the fault zone prevent a more precise definition. In Ellesmere Island the Cambro-Ordovician platform margin is represented by a facies boundary between carbonates of the Cornwallis Group on the south-east and slope deposits — mudstones, turbidites, cherts and resedimented conglomerates — of the Hazen Formation to the north-west (Kerr 1968, Trettin 1979, Trettin & Balkwill 1979). On Judge Daly

Promontory this facies boundary is a linear feature, parallel with and slightly north-west of the later Upper Ordovician–Lower Silurian position.This linear feature is associated with faulting (U. Mayr, pers. comm.) and is on-line with the Nyeboe Land fault zone; the position of the same margin in Greenland (Fig. 7).

Wulff Land anticline

The only major fold structures in eastern Ellesmere Island which involve Lower Palaeozoic strata and bring up Proterozoic rocks in their cores are a belt of NE-trending, fault-disturbed anticlines traceable throughout Judge Daly Promontory sub-parallel to the coast of Kennedy Channel. In the south-west large wavelength folds have a constant strike and form a broad anticlinorium (Kerr 1967b, 1973); to the north-east the fold pattern is complicated by faulting, but a major anticline exposing Proterozoic rocks reaches the north-eastern tip of the promontory (Christie 1974). The folds are parallel with the margin of the Franklinian geosyncline and they deform rocks of facies corresponding to those of the Nyeboe Land fault zone; thus correlation with the Wulff Land anticline is strongly suggested. The structural composition of the North Greenland fold belt adjacent to and north of the Wulff Land anticline is unknown as the area is under the Lincoln Sea. Therefore it is uncertain whether structures of the Wulff Land type bringing up Proterozoic strata form a discrete belt in Greenland as they do on Judge Daly Promontory.

Nyeboe Land fault zone

Major Tertiary faults are common on Judge Daly Promontory in north-eastern Ellesmere Island: dislocations that hold an important position in discussions about strike-slip movement along Nares Strait (Wilson 1963, Christie 1964, 1967, 1974, Kerr 1967a, Miall 1981, Mayr & de Vries, this volume). In north-eastern Judge Daly Promontory two major high-angle faults (Fig. 7) delimit a region that is highly dissected by faults that are sub-parallel to the mid-Palaeozoic Ellesmerian fold pattern. The two major faults transect the coast at the northern end of the promontory and they strike into the southern end of the Robeson Channel. Likewise in Greenland, the main dislocation of the Nyeboe Land fault zone is a high-angle fault of assumed Tertiary age, sub-parallel to the Palaeozoic folds and it strikes into the opposite end of Robeson Channel. Thus correlation is suggested between the fault zones of Nyeboe Land and Judge Daly Promontory, and they are proposed here to be parts of a single fracture line — the Judge Daly – Nyeboe Land fracture line. It should be noted that the Judge Daly fault zone was used by Kerr (1967a) specifically to describe the south-eastern fault on Judge Daly Promontory; in this paper it is used for convenience to include both faults (Fig. 7).

Tertiary sedimentary strata of the Eureka Sound Formation are preserved as down-faulted outliers between the two major faults on Judge Daly Promontory and in a faulted coastal outcrop (Christie 1974). Thus it is possible to demonstrate a Tertiary tectonic history involving Paleocene folding, uplift and erosion, Eocene southerly-directed thrusting, followed by Oligocene or Miocene strike-slip movement and extensional down-faulting (Miall 1981, Mayr & de Vries, this volume). In the absence of post-Franklinian deposits no comparable chronological detail can be recognised in the Nyeboe Land fault zone. Thus it is uncertain if the main tectonic event in Greenland causing uplift of the northern coastal block should be correlated with the Paleocene or the Eocene compressive tectonism known from Judge Daly Promontory. Furthermore, no data exist at the present time for strike-slip movements in the Nyeboe Land fault zone although it should be stressed that no detailed structural analysis has been undertaken.

In addition to the Tertiary chronology that can be elucidated, it is most probable that the Judge Daly fault zone also had an older history. Christie (1974) suggests that all major lineaments on north-eastern Judge Daly Promontory may have been active in pre-Tertiary time and that such movements accounted in part for the Tertiary sedimentary basins. Furthermore, although the north-western part of Judge Daly Promontory is not yet mapped in detail, the Cambro-Ordovician platform margin has been mapped as a very sharp facies contact which is also associated with faulting (U. Mayr, H. P. Trettin, pers. comm.); a relationship that suggests some syndepositional fault control.

The Nyeboe Land fault zone has been correlated by some workers (Newman 1977, Newman, this volume) with a major thrust zone in central Ellesmere Island on the west side of Kane Basin. This correlation has been used in support of a pre-drift position of Greenland and Ellesmere Island involving 250 km of strike-slip displacement along Nares Strait. The main dislocation of this thrust zone is the Parrish Glacier thrust that, according to Mayr & de Vries (this volume), is a low-angle, southerly-directed thrust associated with imbricate thrust sheets which has transported Proterozoic and Lower Palaeozoic strata over Paleocene conglomerates. Correlation of this thrust with the Nyeboe Land fault zone violates the continuity of the Cambro-Ordovician platform margin outlined in this paper (as well as the Upper Ordovician – Silurian margin described by Hurst & Kerr in this volume) which, coincidently or not, occurs on both sides of Nares Strait in the same area as high-angled fault zones. Furthermore, a problem of 'timing' arises if widely separated Tertiary (?Eocene) dislocation zones are used to conclude major (late Paleocene – Eocene) displacement of Greenland and Ellesmere Island.

Harder Fjord fault zone

A second major fault zone, the Harder fjord fault zone, occurs in the North Greenland fold belt and although not described in detail in this paper it is within the region shown in Fig. 7. This fault zone has a long and complex tectonic history (Soper et al. 1980) possibly going back to the early Palaeozoic (Hurst & Surlyk 1980, Surlyk et al. 1980). The sense of net displacement is downthrow to the south and the mountainous terrain of northern Peary Land has been likened to a horst (Fränkl 1955, Haller & Kulp 1962). The main fault planes are typically high-angled and the fault zone has been active in Paleocene or later Tertiary time since Paleocene strata are caught up in the zone (Croxton et al. 1980). There is also evidence of pre-Tertiary magmatism along it (Dawes & Soper 1979, Soper et al. 1980, Parsons 1981). Evidence of Recent fault movement occurs in the form of earthquake epicentres and solfatara activity in its eastern part (Dawes & Peel 1981).

The Harder Fjord fault zone is on direct strike with the Lake Hazen fault zone of Ellesmere Island (Higgins et al., this volume) and while the two fault zones are separated by 200 km of open water, both are traceable for much longer distances in Greenland and Ellesmere Island respectively, and their persistence, strike and nature invite correlation. The Lake Hazen and the Harder Fjord fault zones have affected and preserved Late Palaeozoic, Mesozoic and Tertiary strata of the Sverdrup and Wandel Sea Basins, respectively. The Lake Hazen fault zone shows many features in common with its Greenland counterpart, i.e. a long and complex tectonic history, high-angle faults along which the Grant Land Mountains have been uplifted as a thrust block in the north, and a comparable Tertiary, post-Paleocene age for the latest movements (Trettin 1971).

Gravity field of northern Nares Strait

Sobczak & Stephens (1974) have described the Bouguer anomaly gravity field of the northern Nares Strait region on the basis of 70 field stations in northern Ellesmere Island, Greenland and the Lincoln Sea (Fig. 9). The most notable features of this in the context of the geological features described in this paper are: 1) that the primary gravity anomalies trend sub-parallel to the regional structural elements of the Ellesmere–Greenland fold belt and 2) that a prominent north-easterly trending gravity gradient is continuous from Judge Daly Promontory across the Robeson Channel to the northern coastal area of Nyeboe Land. The Hazen Plateau and Lincoln Sea highs (maximum anomaly 50 mgal) and a parallel belt of gravity lows over Judge Daly Promontory and northern Nyeboe Land (minimum anomaly –95 mgal) are separated by a steep horizontal gradient (maximum –3.68 mgal/km) that is interpreted by Sobczak & Stephens (1974: 1) as representing "an abrupt step-like thickening of the crust of 10 km or more from the northwest to the southeast". These authors conclude that the change in crustal thickness may represent the crust–mantle boundary of the stable craton as earlier depicted in the region by Riddihough et al. (1973).

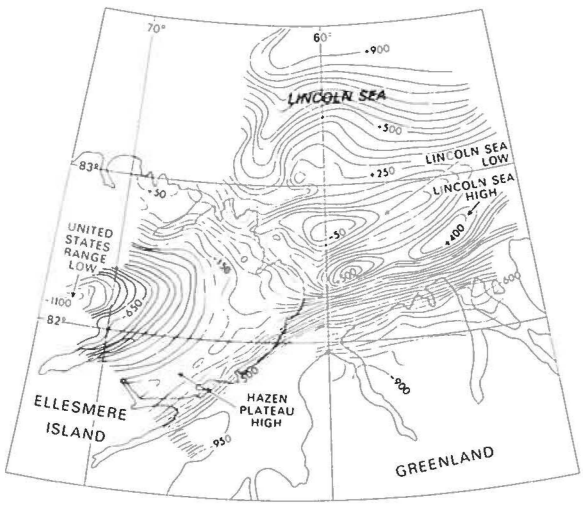


Fig. 9. Bouguer anomaly map of northern Nares Strait region redrawn after Sobczak & Stephens (1974). Gravity contours are of 50 mgal intervals.

The significant point for the present discussion is that this strong gravity gradient of more than 100 mgal coincides regionally with the Franklinian platform margin and with the Judge Daly – Nyeboe Land fracture line as defined in this paper. Also in detail there exists a remarkable correlation with fault lineaments. Thus in Nyeboe Land, the primary gravity anomaly trends are exactly parallel with the Nyeboe Land fault zone and they transect the northern coastline and the regional structural trend of the fold belt at an oblique angle. Thus the "step-like thickening of the crust" mentioned by Sobczak & Stephens (1974) may be coincident with a block fault system at the margin of the Franklinian geosyncline, movements along which caused downflexing and subsidence that controlled the trough sedimentation. Whatever the true origin of the gravity gradient, its spatial correlation with the platform margin and a major fracture system strongly suggests that it represents a structure at least as old as early Palaeozoic. A deep-seated change in crustal character or thickness associated with the passage from normal continental crust to attenuated continental or oceanic crust remains one possibility. Sobczak & Stephens (1974) suggest that it may be related to an ancient plate boundary.

Inferences about Nares Strait

This paper defines four geological features in northern Greenland that can be correlated with corresponding features in Ellesmere Island. This allows inferences to be made on the relative positions of the two landmasses as well as on the proposed structure of the Strait. The four features are:

1) Cambro-Ordovician platform margin
2) Mid-Palaeozoic Wulff Land anticline
3) Tertiary Nyeboe Land fault zone
4) Tertiary Harder Fjord fault zone

Both fault zones, 3) and 4), almost certainly represent the sites of Palaeozoic fractures.

Transcurrent movement

It is not necessary to invoke strike-slip movement between Greenland and Ellesmere Island in order to effect an acceptable correlation of the four geological features outlined above. This suggests a relative position of the two landmasses, in at least Palaeozoic and Tertiary time, that was not far removed from present-day geography. This conclusion refutes conventional reconstructions that involve major movement between Greenland and Ellesmere Island (e.g. Srivastava 1978). This is also supported by other geological features in the northern Nares Strait region discussed elsewhere in this volume by Higgins et al. and Hurst & Kerr, and also by the steep gravity gradient that crosses Robeson Channel without offset (Sobczak & Stephens 1974).

Considered individually, features of Tertiary age are of limited value as indicators of any early Tertiary (Paleocene–Eocene) displacement other than providing a minimum age of any motion. However, in this case the Nyeboe Land fault zone most probably represents the site of a much older (pre-drift) fracture. Particular importance is attached to these four geological features since they record events which represent a span of 500 m.y. from the early Palaeozoic to the Cenozoic and yet each correlates directly across Nares Strait. If Sobczak & Stephen's (1974) interpretation of the steep gravity gradient across the Robeson Channel is correct then the geological record represented by on-line correlatable features is extended well into the Precambrian. However, each of the features transects the Robeson Channel at the same low angle and a *precise* control of the amount of strike-slip net movement which could have occurred is impossible.

The most suitable feature is the Cambro-Ordovician platform margin, since this undoubtedly formed prior to the suggested late Phanerozoic strike-slip motion and prior to the mid-Palaeozoic and Tertiary tectonism of the Franklinian geosyncline. The geological reliability of the feature on both sides of Nares Strait suggests that net sinistral transcurrent movements of up to 50 km do not disrupt the facies pattern unacceptably. Displacements greater than this produce an increasingly problematical mismatch of the platform margin; a net displacement greater than 100 km can be ruled out.

Structure

Kerr (1967a) suggested that Nares Strait is a Tertiary rift valley and he proposed a major fault parallel to the northern coasts of Hall Land and Nyeboe Land. The inference of normal faulting in the Robeson Channel–Lincoln Sea area was, in the absence of onshore data, based solely on the morphology of the coastline. The Hand Bugt fault indicates that a main tectonic event in the coastal region of Hall Land and Nyeboe Land was compressional and not extensional causing coastal down-faulting. Assuming the compression to be Tertiary, the onshore evidence, at least on the Greenland side, does not support an extensional rift valley in the Lincoln Sea area.

The Judge Daly – Nyeboe Land fault line is sub-parallel to the regional fold pattern and southern margin of the Ellesmere–Greenland fold belt. It is thus oblique to Nares Strait. The two major faults in north-eastern Judge Daly Promontory can be traced to the south-west as part of a regional thrust and fault belt which is parallel with and follows the Kennedy Channel coast into the Kane Basin area. Just south of 80°N this belt swings markedly to the west and the dislocations strike inland (Kerr 1967a, Mayr & de Vries, this volume). Therefore, like the mid-Palaeozoic (Ellesmerian) fold structures, this regional thrust-fault belt has also a section that is more or less parallel to Nares Strait.

Christie (1964, 1967) drew attention to the striking linearity of major faults in Judge Daly Promontory and he demonstrated strike-slip displacement of major fold structures. However, he concluded that movements "such as would occur if continental blocks were to shift tens or hundreds of kilometres" have not occurred (Christie 1967: 3, 1974). Mayr & de Vries (this volume) report evidence of sinistral strike-slip motion amounting to about 19 km within the Judge Daly fault zone.

The fault zone correlation suggested in this paper raises the question of how far strike-slip displacements recorded in Judge Daly Promontory or elsewhere along the fault line have influenced the relative positions of Greenland and Ellesmere Island as separate 'plates' and whether such displacements can be used as a direct indication of the amount of strike-slip motion that has occurred along Nares Strait. Since the Judge Daly – Nyeboe Land fault line is oblique to Nares Strait, any strike-slip displacements along it may not necessarily have been translated into overall motion through Nares Strait (Baffin Bay to Lincoln Sea), and such movements may not have contributed to the separation of Greenland and Ellesmere Island. Major tectonic stresses propagated northwards into the Nares Strait region as a

result of the crustal evolution of Baffin Bay and the surrounding oceans, may well have been translated into dislocations oriented oblique to Nares Strait. In this context the deep-seated structure that crosses the Robeson Channel (Sobczak & Stephens 1974) could have hindered in some way the passage of tectonic forces through Nares Strait having the effect of transposing them along fractures parallel to it, i.e. along the Judge Daly – Nyeboe Land fracture line.

The recognition of the Judge Daly – Nyeboe Land and the Harder Fjord – Lake Hazen fracture lines indicates that extensive dislocations of comparable length to Nares Strait, but oblique to it, were of considerable importance in this part of the Arctic in Tertiary time. Moreover, the bedrock geology of the two oblique dislocation lines demonstrates a complex Tertiary history involving compressional and extensional tectonism, including various types of fault and thrust displacements, to say nothing of a pre-Tertiary history that included tectonism and in at least Peary Land extrusive and intrusive magmatism. This contrasts markedly with the Nares Strait lineament across which correlations of geological features of Precambrian to Cenozoic age can be satisfactorily made without invoking strike-slip or vertical displacements and along which no evidence exists of regional shearing or compressive structures which might be connected to a Tertiary subduction zone.

Therefore, without invoking a tantalising and complicated tectonic history for Nares Strait, e.g. a post-Devonian dextral displacement followed by a more or less equal amount of sinistral displacement in response to the generally accepted models for opening of the surrounding oceans, the geology of the northern Nares Strait region suggests that main crustal displacements in the late Phanerozoic took place along fractures oblique to Nares Strait rather than along it.

Acknowledgements

Special thanks are directed to R. L. Christie, Geological Survey of Canada, organiser of Operation Grant Land 1965–66 for support and coordination of the Greenland field work from base-camps in Ellesmere Island. I am grateful to GGU and GSC staff, A. K. Higgins, J. M. Hurst, U. Mayr, J. S. Peel and H. P. Trettin for discussion on the geology of the northern Nares Strait region, and to the above persons and D. A. C. Gardner (Gulf Canada Resources Inc.) for helpful comments to the manuscript. The aerial photographs are published with the permission (A. 495/79) of the Geodætisk Institut, Copenhagen, Denmark. The paper is published with the permission of the Director of the Geological Survey of Greenland.

References

Allaart, J. H. 1965. The Lower Paleozoic sediments of Hall Land, North Greenland. – Unpubl. rep., Grønlands geol. Unders.: 11 pp.

Balkwill, H. R. 1978. Evolution of Sverdrup Basin, Arctic Canada. – Bull. Am. Ass. Petrol. Geol. 62: 1004–1028.

Bendix-Almgreen, S. E. & Peel, J. S. 1974. Early Devonian vertebrates from Hall Land, North Greenland. – Rapp. Grønlands geol. Unders. 65: 13–16.

Berry, W. B. H., Boucot, A. J., Dawes, P. R. & Peel, J. S. 1974. Late Silurian and early Devonian graptolites from North Greenland. – Rapp. Grønlands geol. Unders. 65: 11–13.

Carey, S. W. 1958. The tectonic approach to continental drift. – In: Carey, S. W. (convener), Continental drift. A symposium: 177–355. – Univ. Tasmania, Hobart.

Christie, R. L. 1964. Geological reconnaissance of north-eastern Ellesmere Island, District of Franklin. – Mem. geol. Surv. Can. 331: 79 pp.

Christie, R. L. 1967. Operation Grant Land (1966), northern Ellesmere Island. – Pap. geol. Surv. Can. 67–1: 2–3.

Christie, R. L. 1974. Northeastern Ellesmere Island: Lake Hazen region and Judge Daly Promontory. – Pap. geol. Surv. Can. 74–1A: 297–299.

Christie, R. L., Dawes, P. R., Frisch, T., Higgins, A. K., Hurst, J. M., Kerr, J. W. & Peel, J. S. 1981. Geological evidence against major displacement in the Nares Strait. – Nature, Lond. 291: 478–480.

Croxton, C. A., Dawes, P. R., Soper, N. J. & Thomsen, E. 1980. An occurrence of Tertiary shales from the Harder Fjord Fault, North Greenland fold belt, Peary Land. – Rapp. Grønlands geol. Unders. 101: 61–64.

Dawes, P. R. 1966. Lower Palaeozoic geology of the western part of the North Greenland fold belt. – Rapp. Grønlands geol. Unders. 11: 11–15.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R. & Haller, J. 1979. Historical aspects in the geological investigation of northern Greenland. Part 1: New maps and photographs from the 2nd Thule Expedition 1916–1918 and the Bicentenary Jubilee Expedition 1920–1923. – Meddr Grønland 200(4): 38 pp.

Dawes, P. R. & Peel, J. S. 1981. The northern margin of Greenland from Baffin Bay to the Greenland Sea. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 201–264. – Plenum Press, New York & London.

Dawes, P. R. & Soper, N. J. 1973. Pre-Quaternary history of North Greenland. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 117–134.

Dawes, P. R. & Soper, N. J. 1979. Structural and stratigraphic framework of the North Greenland fold belt in Johannes V. Jensen Land, Peary Land. – Rapp. Grønlands geol. Unders. 93: 40 pp.

Du Toit, A. L. 1937. Our wandering continents: hypothesis of continental drifting. – Oliver and Boyd, Edinburgh: 366 pp.

Fränkl, E. 1955. Rapport über die Durchquerung von Nord Peary Land (Nordgrönland) im Sommer 1953. – Meddr Grønland 103(8): 61 pp.

Haller, J. 1971. Geology of the East Greenland Caledonides. – Interscience Publ., London: 413 pp.

Haller, J. & Kulp, J. L. 1962. Absolute age determinations in East Greenland. – Meddr Grønland 171(1): 77 pp.

Higgins, A. K., Mayr, U. & Soper, N. J. 1982. Fold belts and metamorphic zones of northern Ellesmere Island and North Greenland. – This volume.

Hilgenberg, O. C. 1966. Bestätigung der Kennedy-Channel-Scherung durch die Bruchstruktur von Grönland und Nordost-Kanada. – Geotekt. Forsch. 22: 1–74.

Hurst, J. M. 1979. Uppermost Ordovician and Silurian geology of north-west Peary Land, North Greenland. – Rapp. Grønlands geol. Unders. 88: 41–49.

Hurst, J. M. & Kerr, J. W. 1982. Upper Ordovician to Silurian facies patterns in eastern Ellesmere Island and western North Greenland and their bearing on the Nares Strait lineament. – This volume.

Hurst, J. M. & Surlyk, F. 1980. Notes on the Lower Palaeozoic clastic sediments of Peary Land, North Greenland. – Rapp. Grønlands geol. Unders. 99: 73–78.

Kerr, J. W. 1967a. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Kerr, J. W. 1967b. Stratigraphy of central and eastern Ellesmere Island, Arctic Canada. Part I. Proterozoic and Cambrian. – Pap. geol. Surv. Can. 67–27(I): 63 pp.

Kerr, J. W. 1968. Stratigraphy of central and eastern Ellesmere Island, Arctic Canada. Part II. Ordovician. – Pap. geol. Surv. Can. 67–27(II): 92 pp.

Kerr, J. W. 1973. Geology, Kennedy Channel and Lady Franklin Bay, District of Franklin. 1: 250,000. – Map. geol. Surv. Can. 1359A.

Koch, L. 1920. Stratigraphy of Northwest Greenland. – Meddr dansk geol. Foren. 5(17): 78 pp.

Mayr, U. & de Vries, C. D. S. 1982. Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago. – This volume.

Miall, A. D. 1981. Late Cretaceous and Paleogene sedimentation and tectonics in the Canadian Arctic Islands. – In: Miall, A. D. (ed.), Sedimentation and tectonics in alluvial basins. – Spec. Pap. geol. Ass. Can. 23: 221–272.

Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Newman, P. H. 1982. A geological case for movement between Canada and Greenland along Nares Strait. – This volume.

Parsons, I. 1981. Volcanic centres between Frigg Fjord and Midtkap, eastern North Greenland. – Rapp. Grønlands geol. Unders. 106: 69–75.

Peel, J. S. 1974. Lower Cambrian fossils from Nyeboe Land, North Greenland fold belt. – Rapp. Grønlands geol. Unders. 65: 17 only.

Peel, J. S. 1979. *Serrodiscus* from northern Nyeboe Land, North Greenland. – Rapp. Grønlands geol. Unders. 91: 116 only.

Pettijohn, F. J. 1954. Classification of sandstones. – J. Geol. 62: 360–365.

Poulsen, V. 1969. An Atlantic Middle Cambrian fauna from North Greenland. – Lethaia 2: 1–14.

Riddihough, R. P., Haines, G. V. & Hannaford, W. 1973. Regional magnetic anomalies of the Canadian Arctic. – Can. J. Earth Sci. 10: 157–163.

Sobczak, L. W. & Stephens, L. E. 1974. The gravity field of northeastern Ellesmere Island, part of northern Greenland and Lincoln Sea with map. Lincoln Sea 1: 500 000. – Earth Physics Branch, Gravity Map Series 114: 9 pp.

Soper, N. J., Higgins, A. K. & Friderichsen, J. D. 1980. The North Greenland fold belt in eastern Johannes V. Jensen Land. – Rapp. Grønlands geol. Unders. 99: 89–98.

Soper, N. J., Dawes, P. R., & Higgins, A. K. 1982. Cretaceous–Tertiary magmatic and tectonic events in North Greenland and the history of adjacent ocean basins. – This volume.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–351.

Surlyk, F. 1982. Nares Strait and the down-current termination of the Silurian turbidite basin of North Greenland. – This volume.

Surlyk, F., Hurst, J. M. & Bjerreskov, M. 1980. First age-diagnostic fossils from the central part of the North Greenland foldbelt. – Nature, Lond. 286: 800–803.

Taylor, F. B. 1910. Bearing of the Tertiary mountain belt on the origin of the earth's plan. – Bull. geol. Soc. Am. 21: 179–226.

Trettin, H. P. 1971. Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, Arctic Archipelago. – Bull. geol. Surv. Can. 203: 134 pp.

Trettin, H. P. 1979. Middle Ordovician to Lower Devonian deep-water succession at southeastern margin of Hazen Trough, Cañon Fiord, Ellesmere Island. – Bull. geol. Surv. Can. 272: 84 pp.

Trettin, H. P. & Balkwill, H. R. 1979. Contributions to the tectonic history of the Innuitian Province, Arctic Canada. – Can. J. Earth Sci. 16: 748–769.

Wegener, A. 1924. The origin of continents and oceans. – Methuen & Co., London: 212 pp. (Transl. 3rd edit. by J. G. A. Skerl).

Wilson, J. T. 1963. Hypothesis of Earth's behaviour. – Nature, Lond. 198: 925–929.

Wilson, J. T. 1965. A new class of faults and their bearing on continental drift. – Nature, Lond. 207: 343–347.

Brecciated lineaments in Washington Land, North Greenland, and their relation to Nares Strait

JOHN M. HURST and J. WILLIAM KERR

Hurst, J. M. & Kerr, J. W. 1982. Brecciated lineaments in Washington Land, North Greenland, and their relation to Nares Strait. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 193–198.

Linear crush zones at Kap Jefferson suggest strike-slip movement transverse to Nares Strait. Other parallel lineaments, fjords and valleys occur in North Greenland and may have similar structures in them. These may have been areas of incipient strike-slip movement related to an early compressional phase in the formation of the Nares submarine rift valley. These lineaments are approximately perpendicular to Palaeozoic folds and inferred gneissic trends. They may be incipient transform faults, controlled by the interplay between regional plate tectonic stresses at the time, and the influence of pre-existing structures.

J. M. Hurst, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark and J. W. Kerr, Department of Energy, Mines and Resources, Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303–33rd St. N.W., Calgary, Alberta, Canada, T2L 2A7. J. W. K. present address: J. William Kerr & Associates Ltd., Consulting Geologists, Suite 407, 630–8th Ave. S.W., Calgary, Alberta, Canada, T2P 1G6.

Kerr (1967a, b, 1980a, b, 1981) considered Nares Strait to be a submarine rift valley resulting from faulting during periods of crustal compression and extension when Greenland and Ellesmere Island rotated apart. Little evidence of this structure exists on land north of latitude 80°N on the Greenland side. The purpose of this paper is to focus attention on actual and inferred faults and crush zones in western North Greenland bordering Nares Strait, between Kane Basin and Lincoln Sea (Fig. 1), and to briefly discuss their relevance to Nares Strait.

Structures

Kap Jefferson is situated 40 km north of latitude 80°N on the western coast of Washington Land at the northern perimeter of Kane Basin (Fig. 1). Here there is a complex succession of Silurian basin slope deposits on the southern limb of the Hazen Trough (LBF and CSF; Fig. 2). These basin slope deposits lie conformably upon platform carbonates of Ordovician–Silurian age (AFF), and partly interfinger south-eastward with platform dolomites (ABF). This facies relation was outlined by Hurst (1980a, b) and is similar to the situation on Ellesmere Island (Kerr 1967a, Trettin et al. 1972) as well as on Devon Island (Morrow & Kerr 1977).

Several major lineaments at Kap Jefferson are aligned N10°W, trending at an angle of about 70° to the trend of Nares Strait (Figs 2 and 3). A second minor set of lineaments are aligned at N45°W (Fig. 3), but are not visible on the ground and can only be recognised on aerial photographs. The major lineaments cut through all platform deposits, and individual lineaments form topographic depressions up to 200 m wide. In some cliff exposures (Fig. 4) they have been eroded to a depth of 70 m. They preferentially form linear valleys and are followed by straight stream segments. The rocks in the major lineaments are extremely shattered, brecciated, and invaded by calcite veins, and grade laterally into undisturbed rock (Fig. 5). The crush zones consist almost entirely of fragments of the surrounding formation and there is little apparent offset of bedding across them (Fig. 5). The crush zones undoubtedly are fault breccias. Little vertical movement has occurred, however, because outcrop patterns have not been disrupted, and because beds appear to continue through the breccias. It is suggested that there has been some lateral displacement within these zones. Lateral displacement would be very hard or impossible to detect because the beds are so close to being horizontal. Vertical displacement, even

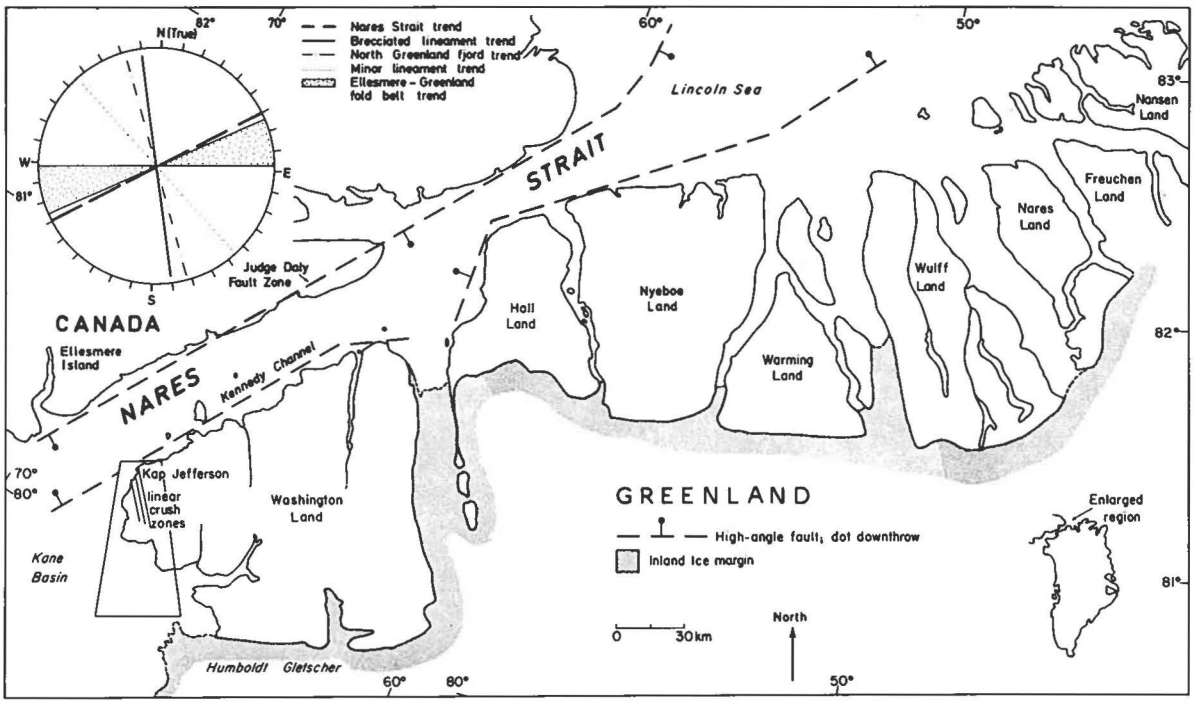


Fig. 1. Simplified map of western North Greenland and eastern Ellesmere Island, Canada, showing the northern part of the Nares submarine rift valley. Faults are after Kerr (1967a). Note the parallel alignment of the fjords in North Greenland with respect to each other and to the crush zones at Kap Jefferson. These are nearly at right angles to the pre-existing regional geological strike and nearly at right angles to the Nares submarine rift valley. Azimuths of the main structural features of the region shown in top left. Inset bottom left indicates region shown in Figs 2 and 3.

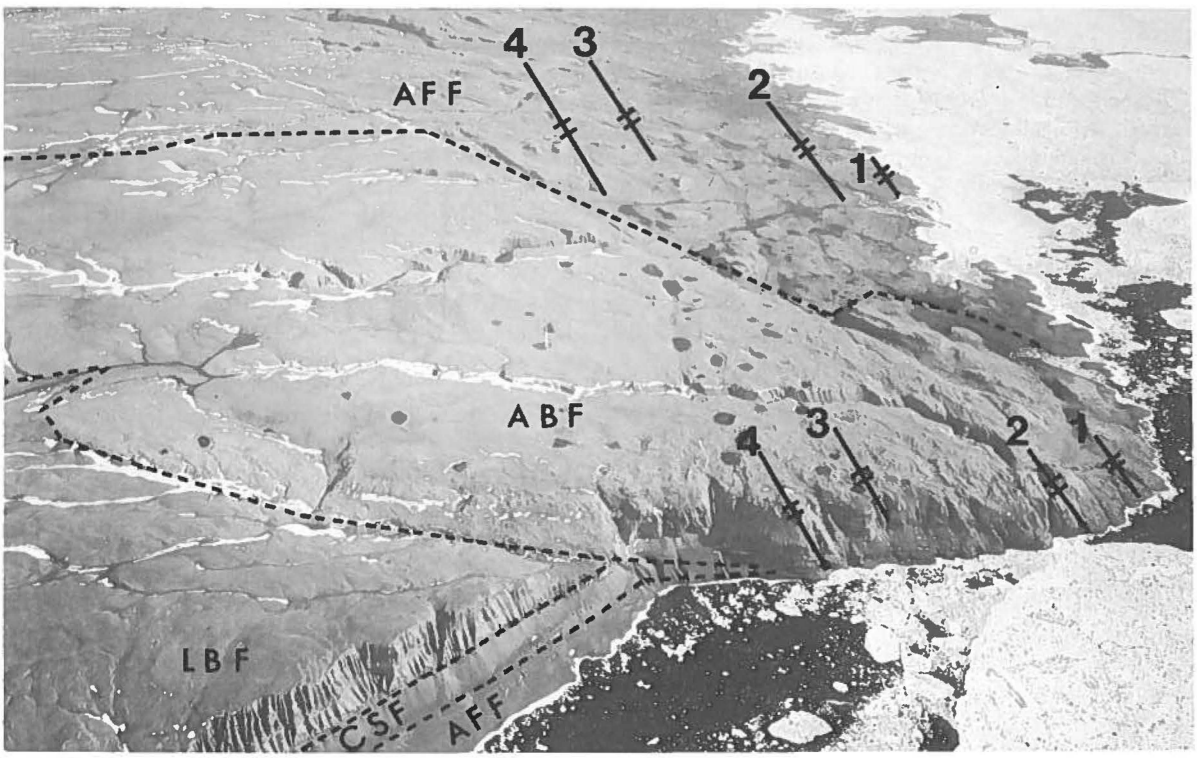


Fig. 2. Geological setting of Kap Jefferson, western Washington Land, viewed southwards (see Fig. 1). Four pronounced lineaments (1 to 4) cut the Aleqatsiaq Fjord Formation (AFF) and the Adams Bjerg Formation (ABF). Cape Schuchert Formation (CSF) and Lafayette Bugt Formation (LBF). Aerial photograph 545K–S, no. 11781; copyright Geodætisk Institut, Denmark.

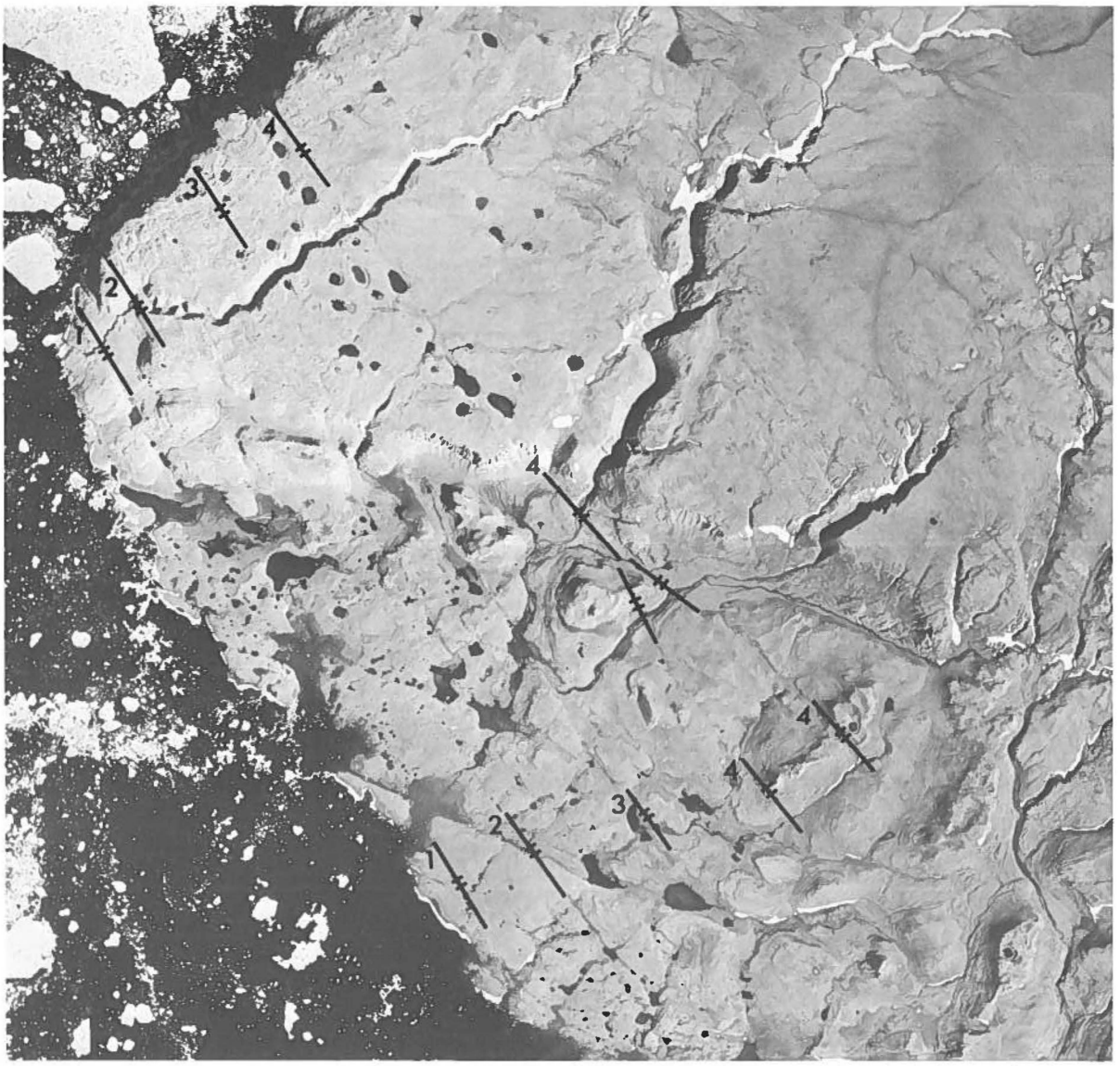


Fig. 3. The four pronounced major lineaments (1 to 4) cut by minor lineaments. Geology as in Fig. 2. Aerial photograph 240V, no. 179, copyright Geodætisk Institut, Denmark.

slight, would be readily observable. Throughout the general region south of Kap Jefferson where the platform dolomites are exposed (Fig. 2) approximately 50 km² are affected by crush zones, giving the regional rock surface a very fractured appearance. The four major lineaments at Kap Jefferson (Figs 2 and 3) are 100 to 200 m wide and contain within them numerous crush zones. Numerous other separate and minor crush zones occur between the four major lineaments.

The lineaments at Kap Jefferson are aligned at about N10°W. The most impressive feature of the North Greenland coast between Washington Land and Freuchen Land is a series of major fjords with approximately similar alignment. Also within the land area many major valleys have a similar alignment, some containing glacial tongues (Fig. 1). The parallelism between the linear crush zones and topographic features suggests that these features all have a similar origin. The main valleys may be glacially carved, but those glaciers probably selectively followed pre-existing pathways and lines of geological and geomorphological weakness. It is suggested here that the NW-trending fjords and valleys of the North Greenland coast may be major lineaments controlled by fault zones or fracture zones, similar to those at Kap Jefferson, that were modified by later glacial erosion.

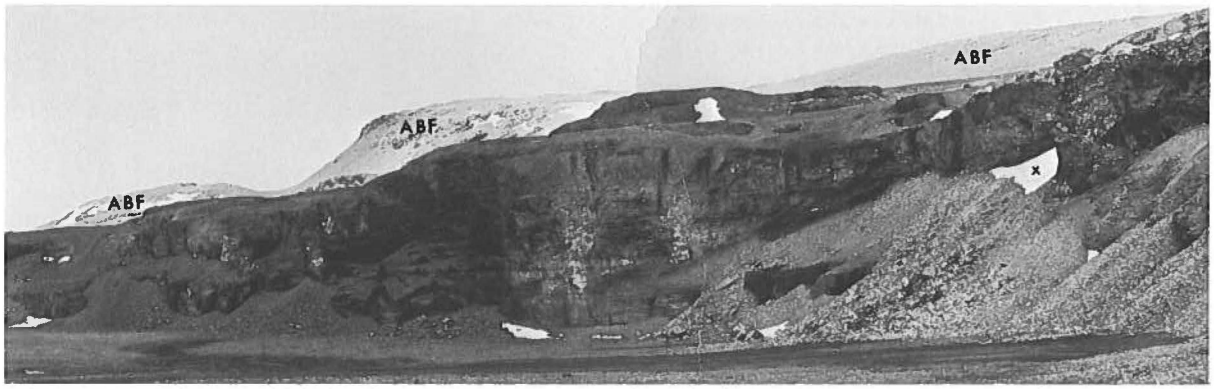


Fig. 4. Valley formed by major lineament 1 of Figs 2 and 3. Note normally-bedded Aleqatsiaq Fjord Formation passing into brecciated zone. Valley width 300 m, valley height 80 m. Adams Bjerg Formation (ABF). X marks the snowpatch in the centre of the area enlarged in Fig. 5.

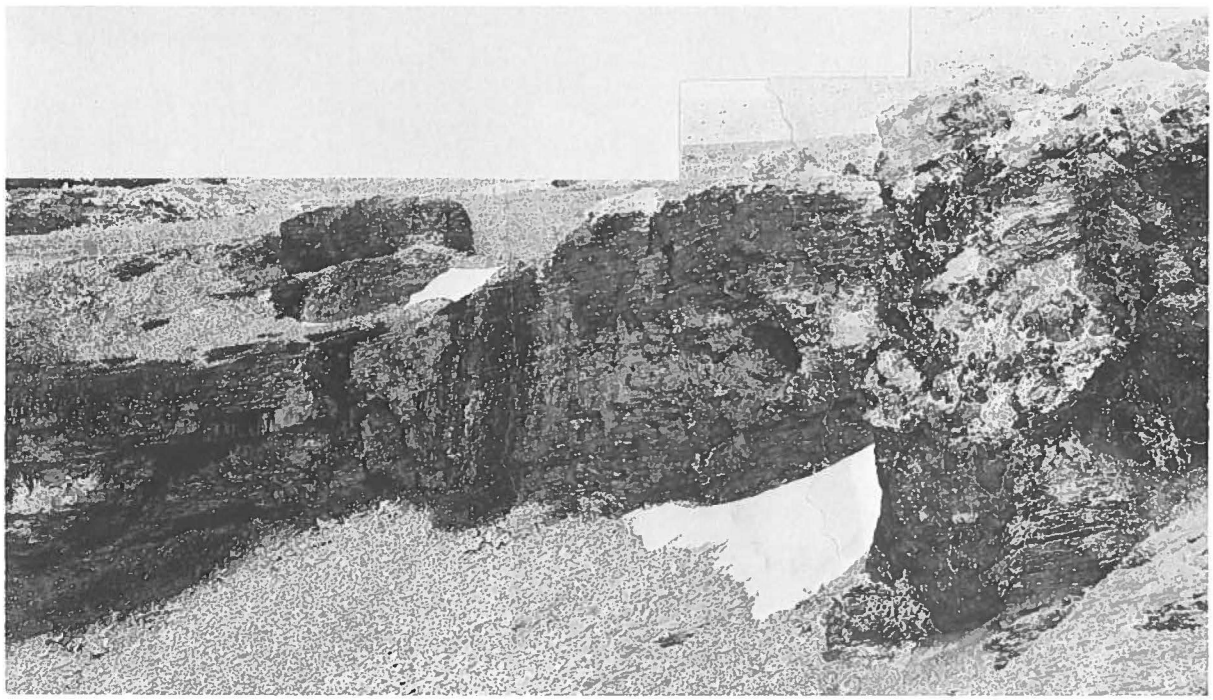


Fig. 5. Close view of crushed rock (Fig. 4) passing into less disturbed and finally undisturbed strata. Cliff height 20 m.

Interpretation

It is generally accepted that Nares Strait is a fault controlled feature of Cretaceous–Tertiary age. Many have considered that it is the locus of a strike-slip fault with hundreds of kilometres of left-lateral displacement (Wegener 1924, Wilson 1963, Srivastava 1978). Kerr (1967a) suggested that Nares Strait formed by a combination of minor strike-slip displacement and rotation apart of Greenland and Ellesmere Island. This involved first a period of extension in the southern Nares Strait region resulting from extension faults from Baffin Bay propagating northward in late Cretaceous to early Tertiary time. At this time extension occurred south of latitude 80°N only, with the region north of that subjected to compression and some left-lateral strike-slip. This is transpression in the sense of Harland (1971, 1973). Subsequently, in mid-Tertiary time the entire

length of Nares Strait became a zone of extension faulting and the Nares submarine rift valley achieved its form as a downdropped structure. In this latter time Baffin Bay and Lincoln Sea were connected continuously by the extension zone, and the rotation apart of Ellesmere Island and Greenland ceased.

The structural trends that apparently existed prior to the Cretaceous–Tertiary rifting event are shown approximately on Fig. 1. The alignment of the lineations at Kap Jefferson and the major fjords and valleys farther north in Greenland is approximately at right angles to the local structures, which presumably also existed before rifting. The lineations, major fjords and valleys may have formed along jointing directions at right angles to the Palaeozoic structural trends while the Nares rift structure was forming. The fjords may be largely erosional and faults are only inferred. However, the crush zones in the Silurian rocks at Kap Jefferson certainly are fault controlled, and have some small displacement at least. These crush zones are post-depositional, and there is no evidence that they correspond to zones of facies changes. Indeed, the major facies change between platform rocks (carbonate) and basin slope rocks (shale) at Kap Jefferson is oblique to the crush zones (Fig. 2).

The age of formation of the lineaments at Kap Jefferson is uncertain. They clearly are post-depositional, unrelated to facies changes. It is inferred that they formed during the Cretaceous–Tertiary plate tectonic deformation outlined by Kerr (1981). Thus, they presumably are related directly to the formation of the Nares submarine rift valley. An analogue to that rifted zone was documented by Surlyk (1977, 1978) in East Greenland, where a dormant north-trending rift system is cut by transverse trending fjords which are inferred to be the site of cross faults. The crush zones at Kap Jefferson may have formed in the following way.

In an early stage of rifting extension took place south of lat. 80°N and transpression took place to the north (Fig. 6A). South of lat. 80°N extension faults in the Nares Strait region were controlled by gneissic and schistose trends of the basement (Kerr 1967a). Details of basement structural trends are unknown north of this latitude, where the region is covered by Palaeozoic

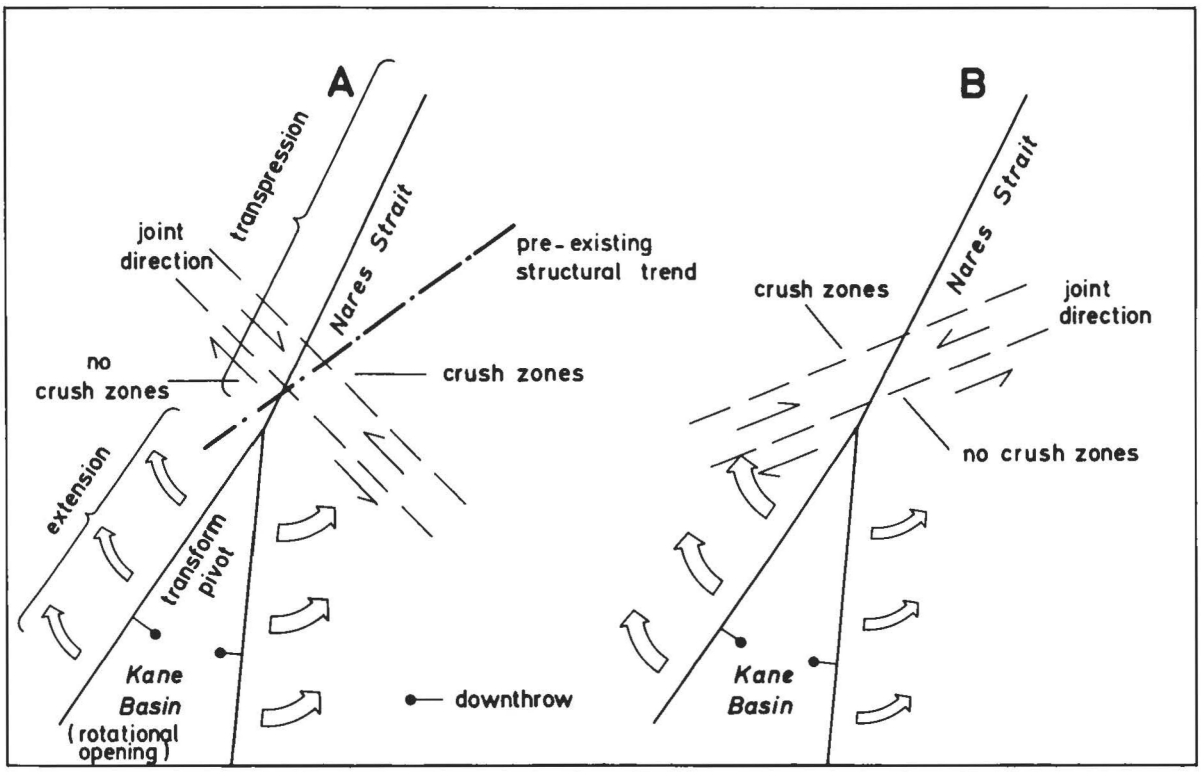


Fig. 6. Structural development of Nares Strait in the Kane Basin – Kennedy Channel area. A: Actual configuration of structures. Rotational opening and foundering in a zone of extension in the south merged across a transform pivot into a zone of transpression in the north. These were controlled by the pre-existing structural trend of folded Lower Palaeozoic sediments (which may reflect basement trends). Mainly strike-slip occurred along Kennedy Channel and mainly crush (incipient strike-slip) along cross trending joints, resulting in crush zones at Kap Jefferson. B: Hypothetical situation in which pre-existing structural trends offered little or no interference to rift advance. Both the pre-existing structural trends (N-trending), and the cross joints could allow extension without transpression.

sediments, but folds in these sediments cross Nares Strait acutely (Fig. 1), and basement trends are inferred to have similar orientation. The lineaments at Kap Jefferson may have been controlled by the northward extension of the structures forming Nares Strait. The lineaments at Kap Jefferson may reflect interference of basement and fold trends with the initiating stresses of the Nares submarine rift valley. These crush zones appear to only occur north of the zone of transformation, lending support to Kerr's (1967a) argument that compression occurred in that region. These crush zones may be incipient transform faults. They may have developed in the early or compressional phase of rifting in the manner shown in Fig. 6A. The compressional phase was then replaced by an extensional phase in which the extensional structures broke northward from southern Nares Strait to completely separate Ellesmere Island and Greenland.

It is interesting to speculate on why similar crush zones have not been reported on the coast of Ellesmere Island. This may have to do with the angular orientation of the extensional structures initiating Nares Strait, and the pre-existing structures and cross joints within them. A direction of jointing perpendicular to the pre-existing structural trend in the Kap Jefferson region was a possible zone of release for the counter-clockwise rotation of the Greenland block. Because the angle between these joints and the tendency of movement of Greenland was less than 90°, crush resulted from incipient transpression and strike-slip occurred along the structural trend. If the orientation of pre-existing structures had been slightly different (Fig. 6B) then crush zones might not have developed, and instead there might have been grabens formed along the cross trending joints.

In summary, the lineaments at Kap Jefferson are crush zones suggesting some small degree of lateral movement. They are approximately perpendicular to pre-existing fold trends and inferred basement trends, with those trends possibly controlling their orientation. They appear to be related to the formation of the Nares submarine rift valley, resulting from transpression as the result of the interplay between pre-existing structures and the orientation of the initiating stresses. Parallel aligned fjords and valleys farther north between Washington Land and Freuchen Land may have similar origin.

Acknowledgements

We thank P. R. Dawes and A. K. Higgins for critically reading the manuscript. J. M. H. publishes with the permission of the Director of the Geological Survey of Greenland. The aerial photographs are published with the permission (A. 495/79) of the Geodætisk Institut, Copenhagen, Denmark.

References

Harland, W. B. 1971. Tectonic transpression in Caledonian Spitsbergen. – Geol. Mag. 108: 27–42.

Harland, W. B. 1973. Tectonic evolution of the Barents Shelf and related plates. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 599–608.

Hurst, J. M. 1980a. Paleogeographic and stratigraphic differentiation of Silurian carbonate buildups and biostromes of North Greenland. – Bull. Am. Ass. Petrol. Geol. 64: 527–548.

Hurst, J. M. 1980b. Silurian stratigraphy and facies distribution in Washington Land and western Hall Land, North Greenland. – Bull. Grønlands geol. Unders. 138: 95 pp.

Kerr, J. W. 1967a. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Kerr, J. W. 1967b. A submerged continental remnant beneath the Labrador Sea. – Earth planet. Sci. Lett. 2: 283–289.

Kerr, J. W. 1980a. Did Greenland drift along Nares Strait? – Bull. Can. Petrol. Geol. 28: 279–283.

Kerr, J. W. 1980b. A plate tectonic contest in Arctic Canada. – In: Strangway, D. W. (ed.), The continental crust and its mineral deposits. – Spec. Pap. geol. Ass. Can. 20: 457–486.

Kerr, J. W. 1981. Evolution of the Canadian Arctic Islands: a transition between the Atlantic and Arctic Oceans. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins, 5, The Arctic Ocean: 105–199. – Plenum Press, New York & London.

Morrow, D. W. & Kerr, J. W. 1977. Stratigraphy and sedimentology of Lower Paleozoic formations near Prince Alfred Bay, Devon Island. – Bull. geol. Surv. Can. 254: 122 pp.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Surlyk, F. 1977. Mesozoic faulting in East Greenland. – In: Frost, R. T. C. & Dickers, A. J. (eds), Fault tectonics in N.W. Europe. – Geol. Mijnb. 56: 311–327.

Surlyk, F. 1978. Jurassic basin evolution of East Greenland. – Nature, Lond. 274: 130–133.

Trettin, H. P., Frisch, T. O., Sobczak, L. W., Weber, J. R., Niblett, E. R., Law, L. K., DeLaurier, J. M. & Whitham, K. 1972. The Innuitian Province. – In: Price, R. A. & Douglas, R. J. W. (eds), Variations in tectonic styles in Canada. – Spec. Pap. geol. Ass. Can. 11: 83–179.

Wegener, A. 1924. The origin of continents and oceans. – Methuen & Co., London: 212 pp. (Transl. 3rd edit. by J. G. A. Skerl).

Wilson, J. T. 1963. Continental drift. – Scient. Am. 280(4): 86–100.

A geological case for movement between Canada and Greenland along Nares Strait

PATRICIA H. NEWMAN

Newman, P. H. 1982. A geological case for movement between Canada and Greenland along Nares Strait. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 199–204.

Nares Strait, part of a 1400 km long lineament between Canada and Greenland, may be one arm (along with Lancaster Sound and Baffin Bay) of a triple junction.

Study of the geological literature on the area from northern Baffin Island to northern Greenland revealed a number of possibly important markers and structural features. The most important of these are: 1) the structural boundary between deformed and undeformed rocks at the margin of the Palaeozoic fold belt of northern Ellesmere Island and Greenland and 2) the large asymmetrical northwest-trending graben systems of northern Baffin Island and Greenland, whose fault blocks tilt to the northeast and southwest respectively. Correlation of these features across Nares Strait suggests that movement along the Strait of a minimum of 200 km has occurred since the mid-Palaeozoic, and a maximum of 300 km since late Precambrian times. It seems probable that the movement occurred in a number of episodes, beginning with rifting in the Proterozoic and ending in the Tertiary Eurekan orogeny.

P. H. Newman, Department of Geology, Dalhousie University, Halifax, Nova Scotia, Canada B3H 3J5. Present address: Department of Geology, University of Auckland, Auckland, New Zealand.

This paper is based on an M.Sc. thesis (Newman 1977, see also Newman, this volume) the geological part of which centred on an extensive literature survey that covered the northern Baffin Bay – Nares Strait region (Fig. 1). This survey was conducted largely during 1976–1977. The conclusions of this work, therefore, do not take into account most of the work carried out and published on the region since 1977. One purpose of the study was to see if, by obtaining a broad overview of the geology of the entire region, any correlations relating to movement along Nares Strait missed by people working on a more detailed scale, might become apparent.

This paper is not intended to be a detailed description of the pro-movement geological criteria. It is simply to direct attention to a few apparently major problems with a direct, no-movement correlation across Nares Strait, and to propose a possible alternative solution for discussion.

Detailed descriptions of the geology of the region have been published elsewhere and in this volume, e.g. Thorsteinsson & Tozer (1970) and Dawes (1976) give good overall coverage of the Canadian and Greenland sides of Nares Strait, respectively. Hence, there follows only a brief simplified description of the regional geological and tectonic setting (and condensed reference list) in order to set a general base for discussion.

Outline of regional tectonics

Reconstructions based primarily on interpretation of magnetic lineations in the Labrador Sea, the North Atlantic and the Arctic Ocean (Kristoffersen & Talwani 1977, Le Pichon et al. 1977, Srivastava 1978) suggest that there was some movement of Greenland with respect to Canada. Burke & Dewey (1973) proposed the existence of a plume-generated triple junction southeast of Devon Island. According to them, one arm spread and became Baffin Bay; one became inactive before spreading (Lancaster Sound); and left-lateral transform motion occurred along the third (Nares Strait).

There is little direct evidence for spreading in Baffin Bay. No definite central ridge has been found, though Keen et al. (1972) and Johnson (1971) mentioned a possible ridge in northern Baffin Bay. Jackson et al. (1979) showed that magnetic lineations could be mapped in Baffin Bay, and suggested that the anomalies fitted numbers 13 to 21. The crust in the central part of the Bay is oceanic (Keen et al. 1971) and spreading or opening of some sort could have occurred in Baffin Bay at approximately the same time as known Tertiary spreading in the Labrador Sea (Kristoffersen & Talwani 1977, Srivastava 1978). Martin (1973) suggested that the oceanic crust, like that of the Red Sea, is a product of rifting rather than spreading.

The inactive arm of the proposed triple junction, Lancaster Sound, is bounded by steep faults and contains a thick (about 10 km) sedimentary sequence similar to that of aulacogens (Jackson et al. 1977).

Some of the geological evidence for lateral motion along the third arm of the possible triple junction, Nares Strait, is examined here.

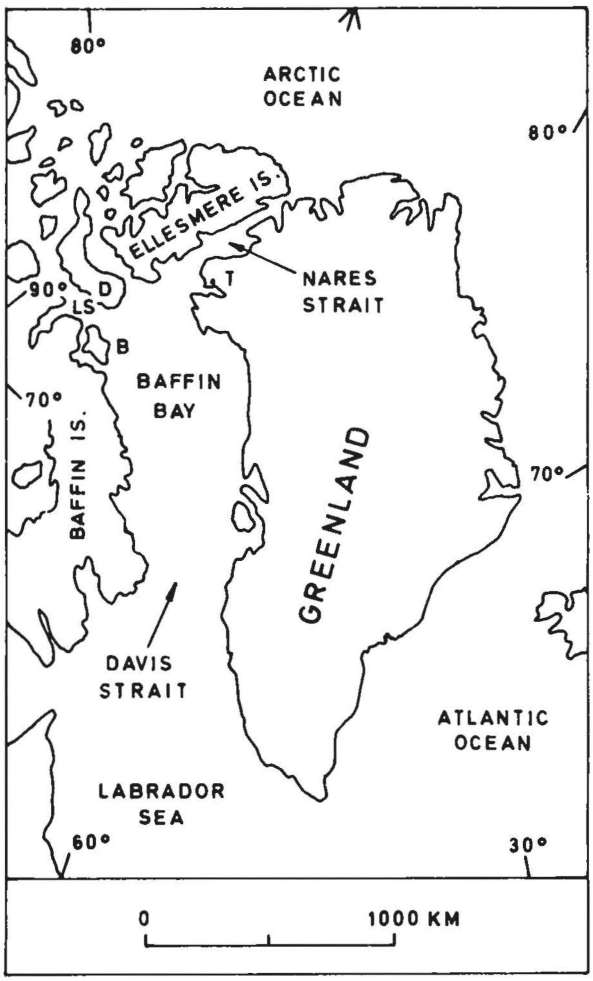


Fig. 1. Index map of the Baffin Bay – Nares Strait region. D = Devon Island, LS = Lancaster Sound, B = Bylot Island, T = Thule.

Outline of regional geology

Precambrian rocks are essentially restricted to the southern half of the map area (Fig. 2). These are predominantly granulite-facies gneisses of the Churchill Province, but thick sequences of relatively undisturbed Proterozoic strata (c. 1200 to c. 600 m.y.), including sandstone units, volcanic rocks, red beds and carbonate rocks are preserved in large northwest–southeast trending graben structures on northern Baffin Island and in the Thule district of Greenland (Jackson & Davidson 1975, Dawes 1976). A few small outcrops of a volcanic–sedimentary sequence similar to the lower portion of these major sequences are preserved on the tips of peninsulas of southeastern Ellesmere Island (Frisch et al. 1978). While these outcrops appear to have a much less clearly-defined structural setting, with only minor faulting, associated basic dykes generally trend north to northeast. Some faults associated with the graben of Baffin Island and the Thule district have been reactivated a number of times since the Proterozoic.

Proterozoic to Palaeozoic rocks surrounding the crystalline shield form a distinctive linear belt running roughly northeast. This essentially continuous succession of Proterozoic to Devonian strata forms the central stable platform and to the north the Franklinian geosyncline. The geosyncline did not undergo severe orogenesis until the late Devonian to mid-Pennsylvanian when it was transformed into the Ellesmere–Greenland fold belt (Thorsteinsson & Tozer 1970, Dawes 1976).

Carboniferous to Tertiary clastic and carbonate rocks deposited on the eroded fold belt form the Sverdrup Basin of west and north-central Ellesmere Island. The Tertiary Eurekan orogeny was the main event of deformation of the Sverdrup Basin, and it extensively overprints the Palaeozoic Ellesmerian structures of the Ellesmere–Greenland fold belt in both Canada and Greenland (Thorsteinsson & Tozer 1970, Dawes 1976).

In the Ellesmere–Greenland fold belt, the major thrusting on Ellesmere Island is directed southward, whereas in Greenland, it is to the north. Due to the extensive overprinting of structures by the later Tertiary orogeny, there is some debate as to how much of this thrusting belonged to the earlier Palaeozoic event. Thorsteinsson & Tozer (1970) believe that thrusting did not begin until the Tertiary on Ellesmere Island. In Greenland at least some of the major thrusts are of Tertiary age (Dawes & Soper 1971).

Correlation with 250 km displacement

The most frequently mentioned magnitude of the left-lateral displacement of Greenland relative to Canada is 200–300 km (Wegener 1924, Carey 1958, Clarke 1968, Johnson 1971, Keen & Barrett 1973, Srivastava 1978). Fig. 3 illustrates reconstruction with 250 km of displacement. Two striking correlations of the simplified geology are immediately apparent.

The large Proterozoic northwest-trending graben and horst systems of northern Baffin and Bylot Islands and the Thule district are brought into conjunction. The sedimentary sections appear to be very similar both in lithology and sequence. The thicknesses of the rocks of North-West Greenland and Bylot Island are comparable, although the Borden Peninsula section is substantially greater (Newman 1977). Both the Canadian and Greenland systems have been reactivated a number of times since the Precambrian (Blackadar 1957, Jackson & Davidson 1975, Dawes 1976). Both graben systems are asymmetrical, but in an opposite sense. The fault blocks of the North Baffin Rift Zone tilt strongly northeast (Jackson & Davidson 1975); those of the Thule district tilt southwest (Davies et al. 1963, Dawes 1976).

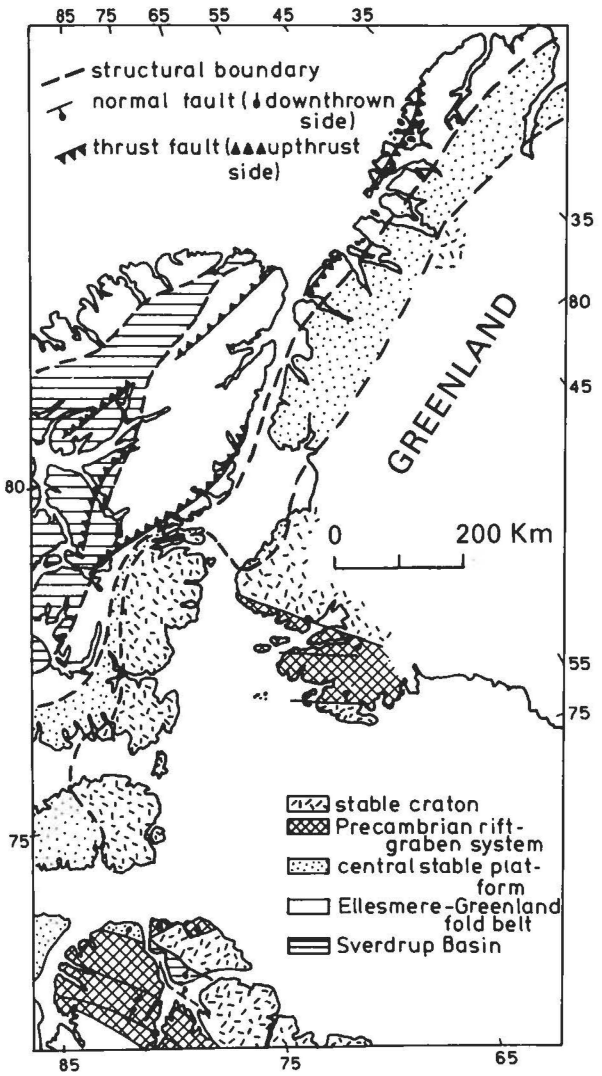


Fig. 2. Simplified geology/structure map of the Nares Strait region.

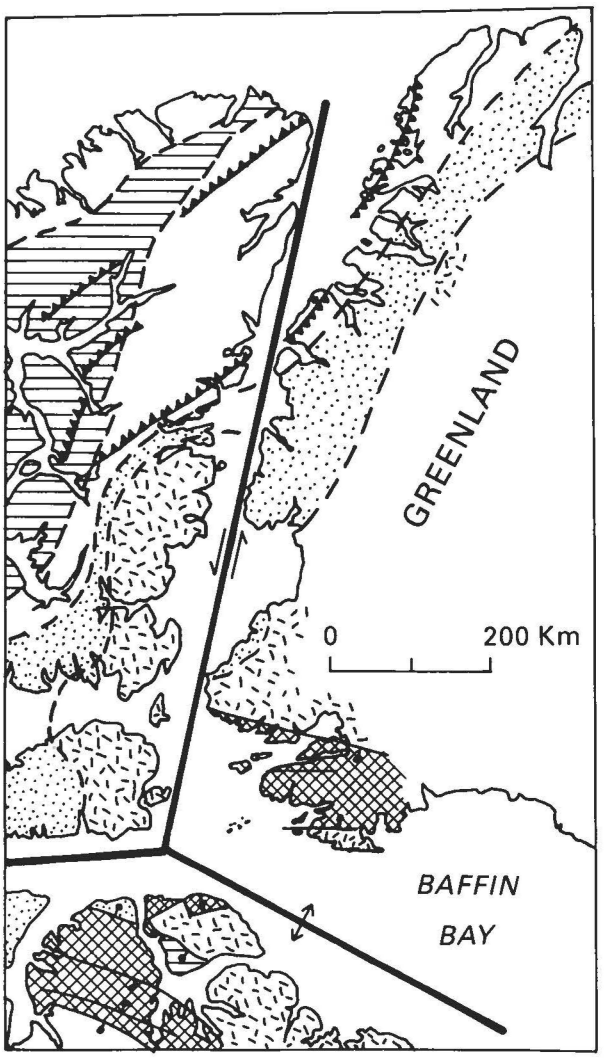


Fig. 3. 250 km reconstruction of simplified geology/structure map, showing possible triple junction. Legend as in Fig. 2.

These might be interpreted as opposite margins of a rift system. The greatest drop would be in the central rift, in this case, Baffin Bay (Fig. 4). On the 250 km reconstruction (Fig. 3) the central rift between the two graben zones of North-West Greenland and northern Baffin Island is oriented at an angle of about 120° to Lancaster Sound and Nares Strait, as is typical of the triple junction configuration.

In this orientation (Fig. 3), the Proterozoic sedimentary and volcanic rocks of southeastern Ellesmere Island have no equivalent directly across Nares Strait in Greenland. While there are very obvious sedimentological similarities between the Ellesmere and Greenland sections, the author believes that in the overall picture their differences are more important than their similarities, i.e. the differences in lateral extent, structural setting, strike of structural features and dykes, and completeness of section. In other words, while these two sequences were presumably once parts of a common sedimentary system, they were not necessarily directly opposite one another as in the present-day configuration of Nares Strait.

Given a triple-rift system, one could expect a common, continuous sedimentological system to all three rifts, with similar resultant rocks. The major differences would be in lateral extent, structural strike, thickness, and possibly volcanic content, determined by the extent of activity along each arm. Presumably, an arm/rift that did little more than open slightly and stop would have little, if anything, in the way of a sedimentary record of this exposed (e.g. Lancaster Sound). An arm along which there was continuing activity, primarily in a lat-

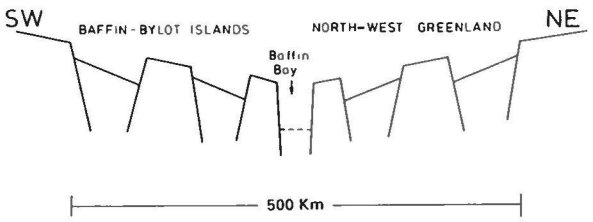


Fig. 4. Cartoon cross-section of reconstructed Baffin Bay with the Baffin–Bylot and Thule graben systems, based on literature description of asymmetrical tilted fault blocks. From Newman (1977).

eral sense, would possibly preserve and expose a series of small, narrow basins (e.g. Nares Strait, and the small Proterozoic outcrops of southeastern Ellesmere Island). The arm along which there was continuous rifting and opening activity would appear to be the arm along which, or parallel to which, one would expect to see sediments formed and preserved to the greatest lateral extent in a series of parallel graben (e.g. Baffin Bay and the Baffin–Bylot and Thule graben systems).

The second striking correlation in the 250 km configuration (Fig. 3) is the pronounced alignment of the highly folded and faulted Proterozoic to Devonian sedimentary sequence of the Ellesmere–Greenland fold belt. In all the literature surveyed during this study, there was only one marker horizon which all workers seemed able to identify and locate clearly, and agree on, and that was the structural boundary between deformed and undeformed rocks north of the stable craton. This boundary is shown on Figs 2 and 3, and it correlates well on the 250 km reconstruction.

It is interesting to note that the thrust faults on either side of the Strait are usually in opposing directions, i.e. Greenland shows predominantly northerly tectonic transport, while in Ellesmere Island it was mostly to the south. Rotational movement of Greenland with respect to Canada may provide a possible explanation (Kerr 1967, Srivastava 1978), but according to Freund (1965), reversal of thrust directions is to be expected across a strike-slip fault (Fig. 5).

In the 250 km reconstruction (Fig. 3), the boundary between the Precambrian crystalline basement and the Palaeozoic platform cover does not correlate well across the Strait. Considering the difference in width of stable platform deposits between Ellesmere Island and Greenland (Fig. 2), however, it is impossible to find a position which fits both boundaries. The southern boundary with the Precambrian basement is a depositional and erosional boundary, and is therefore dependent on the shape of the original depositional surface, and the depth of subsequent erosion. It seems more probable that this marker, rather than the structural boundary to the north, would have had an 'original' sharp bend in it. Scattered outliers of Palaeozoic carbonate rocks on southeastern Ellesmere Island (Christie 1962, Frisch et al. 1978) indicate that the Palaeozoic cover there was once much more extensive than at present. Therefore, the positions on the 250 km reconstruction of the original southern boundaries of the Palaeozoic sediments would possibly be not so far apart.

Using the geological evidence cited above, movement between Canada and Greenland of about 250 km appears to give a reasonable correlation. However, some Precambrian features correlate well to a maximum of about 300 km displacement, and the Palaeozoic markers, such as the fold boundary, will correlate to a minimum of 200 km if Nares Strait is allowed to widen slightly.

Timing

Most workers supporting movement between Canada and Greenland suggest that it occurred during the late Cretaceous to Tertiary. This interpretation is based on the timing of spreading (from magnetic lineations) in the Labrador Sea to the south, and that of various sections of the northern mid-Atlantic Ridge (Le Pichon et al. 1971, 1977, Srivastava 1978). Dawes (1973), however, pointed out that if the main overturning of fold structures in opposing directions on either side of the Strait is truly of Palaeozoic age, then Nares Strait was a prominent and influential feature at least that far back.

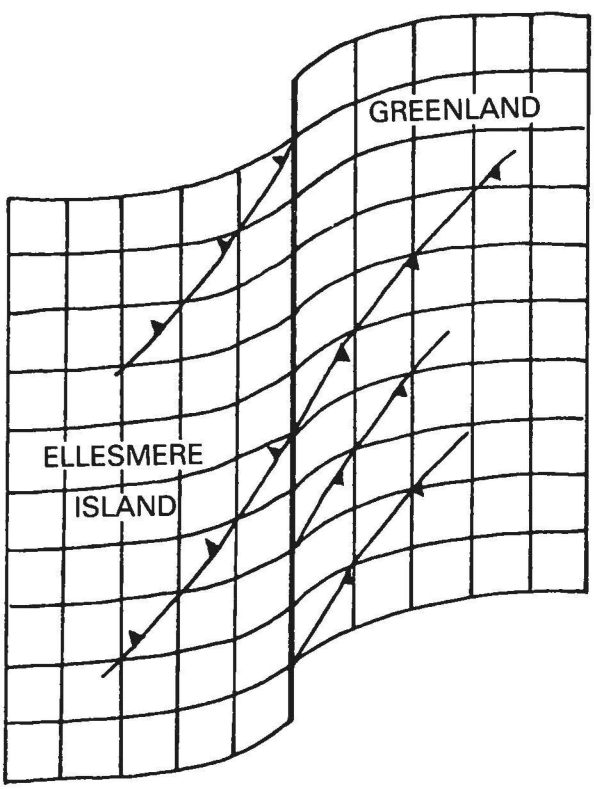


Fig. 5. Predicted orientation of thrust faults across a strike-slip fault (modified from Freund 1965).

Accepting that the boundary of the Palaeozoic fold belt was originally straight, and that the Baffin and Thule Proterozoic graben systems are related and were originally joined, there is no good geological fit with movement of less than 200 km. Le Pichon et al. (1977) believe that a maximum of 100 km of displacement occurred in the last 63 m.y. Preserved Carboniferous and younger sediments occur west of Nares Strait in the Sverdrup Basin, and therefore there are no distinctive younger markers to use in correlating across the Strait. The Tertiary orogeny reactivated many of the older structures, but these need not have been in their original relative positions. The crustal weaknesses would have continued to exist wherever the continents lay relative to one another and would have been susceptible to reactivation. Left-lateral displacement could have occurred along Nares Strait during both main orogenies. Perhaps the Ellesmerian orogeny moved Greenland 100 to 150 km, and the Eurekan orogeny the last 100 km.

Srivastava (1978) believes that 250 km of movement occurred in the last 63 m.y. Perhaps the Palaeozoic Ellesmerian orogeny and any possible earlier ones only formed compressional features and did not involve any lateral movement along Nares Strait. If the Palaeozoic fold boundary does correlate at the 250 km reconstruction as suggested, then the last 250 km of movement is of post-Palaeozoic age. The Tertiary Eurekan orogenic forces, at a low angle to the Palaeozoic compression, reactivated many of the older structures and could have initiated strike-slip displacement along the Strait. Thorsteinsson & Tozer (1970) believe that the Ellesmerian orogeny produced only folds, and that the prevalent southward thrusting did not begin until the Tertiary. This may indicate that substantial left-lateral movement did not occur until the Tertiary. Even if thrusting did occur during the Palaeozoic, it was strongly reinforced during the Eurekan orogeny (Kerr 1967, Dawes 1976) and perhaps was a more dominant feature of the later orogeny.

Conclusions

On the basis of the few large-scale features discussed above, a case can be made for considerable movement along Nares Strait. The left-lateral displacement between Greenland and Canada, to a total of about 250 km, probably occurred in a number of movements between late Precambrian time and the Tertiary. Major rifting activity took place during the Proterozoic and Devonian–Carboniferous, and substantial displacement during the late Cretaceous–Tertiary.

If a theory of no, or little, movement, i.e. direct geological correlation across Nares Strait, is to be maintained, then it must provide a reasonable explanation for the rather major features and problems discussed here.

References

Blackadar, R. G. 1957. The Proterozoic stratigraphy of the Canadian Arctic Archipelago and northwestern Greenland. – In: Gill, J. E. (ed.), The Proterozoic in Canada. – Roy. Soc. Can. Spec. Publ. 2: 93–100.

Burke, K. & Dewey, J. F. 1973. Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks. – J. Geol. 81: 406–433.

Carey, S. W. 1958. The tectonic approach to continental drift. – In: Carey, S. W. (convener), Continental drift. A symposium: 177–355. – Univ. Tasmania, Hobart.

Christie, R. L. 1962. Geology, southeast Ellesmere Island, District of Franklin (map with marginal notes) 1'' to 4 miles. – Map geol. Surv. Can. 12–1962.

Clarke, D. B. 1968. Tertiary basalts of the Baffin Bay area. – Unpubl. Ph. D. thesis, Univ. Edinburgh, Scotland: 122 pp.

Davies, W. E., Krinsley, D. B. & Nicol, A. H. 1963. Geology of the North Star Bugt area, Northwest Greenland. – Meddr Grønland 162(12): 68 pp.

Dawes, P. R. 1973. The North Greenland fold belt: a clue to the history of the Arctic Ocean basin and the Nares Strait lineament. – In: Tarling, D. H. & Runcorn, S. K. (eds), Implications of continental drift to the earth sciences 2: 925–947. – Academic Press, London & New York.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R. & Soper, N. J. 1971. Significance of K/Ar age determinations from northern Peary Land. – Rapp. Grønlands geol. Unders. 35: 60–62.

Freund, R. 1965. A model of the structural development of Israel and adjacent areas since Upper Cretaceous times. – Geol. Mag. 102: 189–205.

Frisch, T., Morgan, W. C. & Dunning, G. R. 1978. Reconnaissance geology of the Precambrian Shield on Ellesmere and Coburg islands, Canadian Arctic Archipelago. – Pap. geol. Surv. Can. 78–1A: 135–138.

Jackson, G. D. & Davidson, A. 1975. Bylot Island map-area, District of Franklin. – Pap. geol. Surv. Can. 74–29: 12 pp.

Jackson, H. R., Keen, C. E. & Barrett, D. L. 1977. Geophysical studies on the eastern continental margin of Baffin Bay and in Lancaster Sound. – Can. J. Earth Sci. 14: 1991–2001.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor spreading in central Baffin Bay. – Can. J. Earth Sci. 16: 2122–2135.

Johnson, J. S. 1971. A contribution to the structure of northern Baffin Bay and Lancaster Sound. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 103 pp.

Keen, C. E. & Barrett, D. L. 1973. Structural characteristics of some sedimentary basins in northern Baffin Bay. – Can. J. Earth Sci. 10: 1267–1278.

Keen, M. J., Loncarevic, B. D. & Ewing, G. N. 1971. Continental margin of eastern Canada: Georges Bank to Kane Basin. – In: Maxwell, A. E. (ed.), The sea 4: 251–291. – Wiley-Interscience, New York.

Keen, M. J., Johnson, J. [S.] & Park, I. 1972. Geophysical and geological studies in eastern and northern Baffin Bay and Lancaster Sound. – Can. J. Earth Sci. 9: 689–708.

Kerr, J. W. 1967. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Kristoffersen, Y. & Talwani, M. 1977. Extinct triple junction south of Greenland and the Tertiary motion of Greenland relative to North America. – Bull. geol. Soc. Am. 88: 1037–1049.

Le Pichon, X., Hyndman, R. D. & Pautot, G. 1971. Geophysical study of the opening of the Labrador Sea. – J. geophys. Res. 76: 4724–4743.

Le Pichon, X., Sibuet, J.-C. & Francheteau, J. 1977. The fit of the continents around the North Atlantic Ocean. – Tectonophysics 38: 169–209.

Martin, R. 1973. Geological history of Baffin Bay – continental drift before sea-floor spreading and the exploration for hydrocarbons. – In: Hood, P. J. (ed.), Earth science symposium on offshore eastern Canada. – Pap. geol. Surv. Can. 71–23: 599–620.

Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Newman, P. H. 1982. Marine geophysical studies of southern Nares Strait. – This volume.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Thorsteinsson, R. & Tozer, E. T. 1970. Geology of the Arctic Archipelago. – In: Douglas, R. J. W. (ed.), Geology and economic minerals of Canada. – Econ. Geol. Rep. geol. Surv. Can. 1: 547–590.

Wegener, A. 1924. The origin of continents and oceans. – Methuen & Co., London: 212 pp. (Transl. 3rd edit. by J. G. A. Skerl).

Cretaceous–Tertiary magmatic and tectonic events in North Greenland and the history of adjacent ocean basins

NORMAN J. SOPER, PETER R. DAWES and ANTHONY K. HIGGINS

Soper, N. J., Dawes, P. R. & Higgins, A. K. 1982. Cretaceous–Tertiary magmatic and tectonic events in North Greenland and the history of adjacent ocean basins. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 205–220.

Greenland emerged as a separate entity during the break-up of the Laurasian plate in Palaeogene time. Research in the last decade has revealed some of the magmatic and tectonic events which were associated with the development of Greenland's northern margin: late Cretaceous basic dyke swarms, an explosive volcanic province (Kap Washington Group) dated as end-Cretaceous, a thrust zone of late Paleocene – Eocene age and a later Tertiary period of fault reactivation related to regional uplift. The north-eastern (Wandel Sea) margin was affected by compressional and extensional deformations.

The evidence for this sequence of events is outlined together with chemical data which characterise the igneous activity as alkalic, of 'within-plate' type. A coherent pattern can be discerned between the onshore geological history and the spreading history of adjacent ocean basins, when the revised polarity time scale of Hailwood et al. (1979) is used as a link. We show how the geological evidence from North Greenland provides a number of important constraints on the timing and magnitude of displacements between Greenland and Canada in the Labrador Sea and Baffin Bay and also along Nares Strait.

N. J. Soper, Department of Geology, University of Sheffield, Mappin Street, Sheffield S1 3JD, England; P. R. Dawes and A. K. Higgins, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark.

In North Greenland a Proterozoic – Lower Palaeozoic platform sequence passes northwards into a clastic trough which was subjected to deformation and metamorphism in mid-Palaeozoic time. This comprises the eastward extension of the Franklinian mobile belt of Arctic Canada. Superimposed on the Franklinian belt are structures of Tertiary age which delimit an important regional tectonic zone extending from the Canadian Arctic Islands (Eurekan orogeny: Thorsteinsson & Tozer 1970, Trettin et al. 1972) through North Greenland to Svalbard (West Spitzbergen orogeny: Harland & Horsfield 1974). The deformation in this zone is only locally intense. The Canadian and Svalbard segments have been extensively studied, but the Tertiary events in North Greenland have been documented only in the last decade. In 1969 a volcanic province (Kap Washington Group) was discovered on the north coast of Peary Land (Dawes & Soper 1970, 1973) and subsequently dated isotopically at about the Cretaceous–Tertiary boundary (Larsen et al. 1978). The volcanic rocks are in thrust contact with Palaeozoic metasediments of the Franklinian trough, and this important phase of compressive deformation perpendicular to the North Greenland continental margin is also of Palaeogene age. Weaker structures parallel to the Wandel Sea margin have been shown to affect the youngest (Paleocene) sediments of the Wandel Sea Basin cover sequence (Troelsen 1950, Håkansson 1979, Håkansson et al. 1981), and other volcanic and tectonic features of Tertiary age have come to light during the systematic mapping programme which is currently being undertaken by the Geological Survey of Greenland.

It is evident that these events affected the North Greenland continental margin during the inception and development of the adjacent ocean basins in late Mesozoic and Cenozoic time (Fig. 1). The spreading geometry of the Labrador Sea, Norwegian Sea and Eurasia Basin has been investigated intensively in recent years by magnetic anomaly matching (Kristoffersen & Talwani 1977, Talwani & Eldholm 1977, Phillips et al. 1978, Srivastava 1978, Feden et al. 1979), but incompatibilities exist between the various models proposed and no unique geometrical solution has emerged for the development of the oceans around Greenland in terms of rigid-plate theory. It is here that geological evidence of within-plate deformations can provide useful constraints on displacement models. The prime example is the very subject of this symposium — Nares Strait. Since the geological evidence from the lands bordering the Strait allows only tens rather than hundreds of kilometres wrench displacement (Christie et al. 1981, Dawes & Kerr, this volume) models which

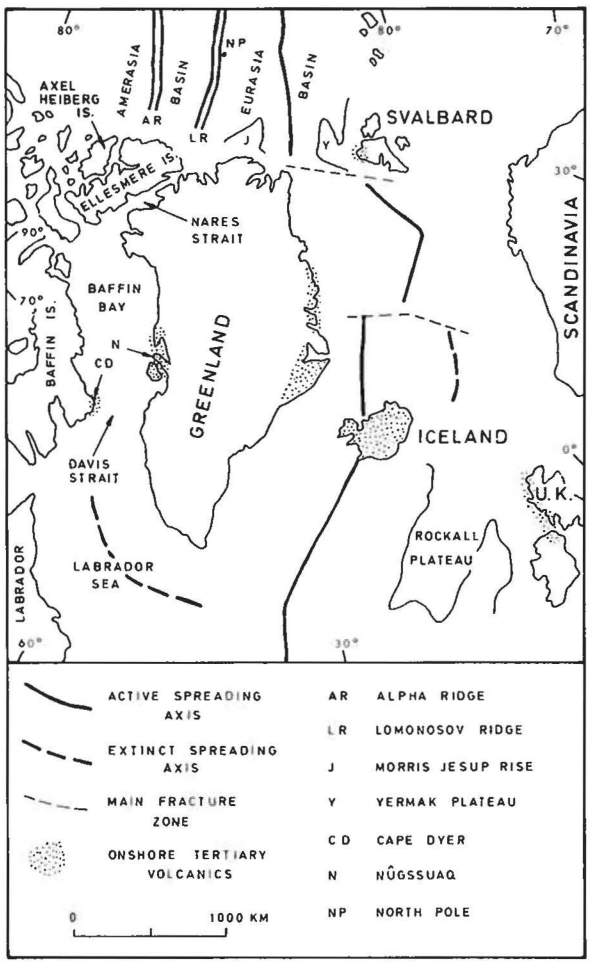


Fig. 1. Index map showing position of Greenland in the North Atlantic – Arctic Ocean system. Outline of submarine plateaux demarcated by 1000 m isobath.

require very large displacements can be eliminated. On the other hand, reconstructions that allow no movement along Nares Strait and treat the North American and Greenland continental masses as entirely rigid result in unacceptable overlap of continental rocks in some part of the North Atlantic region. In addition, credible plate tectonic models must provide appropriate geotectonic settings to account for geological features which would otherwise remain unexplained — for example the existence of an explosive alkalic volcanic province on the north coast of Greenland.

In order to relate the onshore geological record to the spreading history of adjacent oceans, a chronological framework is required and this is provided by the polarity time scale. By the mid-70s it had become apparent that the Heirtzler et al. (1968) time scale assigned ages to magnetic epochs in the Palaeogene which were discrepantly high by about ten per cent. The studies of spreading geometry listed above continued to use the Heirtzler scale and thus relationships with geologically dated events are not always apparent. Among the revised time scales recently proposed we adopt that of Hailwood et al. (1979) which is based in the Palaeogene on a biostratigraphically controlled reversal stratigraphy from North Atlantic sediment cores.

In this paper we show that the Cretaceous–Tertiary geological record in North Greenland and the spreading history of the adjacent ocean basins, in so far as they are presently understood, form a coherent pattern if viewed in the light of the revised polarity time scale, and that the parallel lines of evidence provide useful constraints on the interpretation of each other.

Cretaceous–Tertiary events in North Greenland

Much remains to be learned of the post-Palaeozoic history of North Greenland, but there is sufficient evidence to identify several major magmatic and tectonic features of late Mesozoic–Tertiary age (Fig. 2).

Basic dyke swarms

Dolerite dykes are extensively developed in the North Greenland fold belt where they cut the mid-Palaeozoic structures and they occur less frequently on the platform to the south (Ellitsgaard-Rasmussen 1955, Jepsen 1969, Dawes & Soper 1970). These dykes are restricted to the eastern part of the fold belt, extending as far west as the vicinity of Victoria Fjord at long. 48°W. In the fold belt approximately coast-normal dykes predominate (NNE to NW) and there is a marked concentration in a 100 km wide zone of the north coastal region (Fig. 2). On the islands in the central part of this zone the proportion of northerly-trending dykes reaches 20 per cent; large inclined dolerite sheets and scattered E–W-trending dykes also occur here. The northerly swarm dies out southwards towards the Harder Fjord fault zone, to be replaced in and south of the fault by WNW–ESE and NW–SE swarms of en echelon dykes extending SSE for about 120 km (Higgins et al. 1981). Some dykes in the Frederick E. Hyde Fjord area are aligned in the fault zone.

The dykes are normally straight discordant dolerites with chilled margins, usually ranging up to 25 m in thickness but occasionally attaining 200 m. Particularly towards the north coast there is evidence of shearing and retrogression (Dawes & Soper 1979), and within a few hundred metres of the Kap Cannon thrust the dykes develop schistose fabrics and greenschist-facies assemblages.

Petrographically the undeformed dykes are ophitic textured dolerites and olivine dolerites, occasionally feldsparphyric or vesicular. Samples occur with relatively fresh olivine, but there is a gradation in the degree of alteration to complete replacement of the olivine by

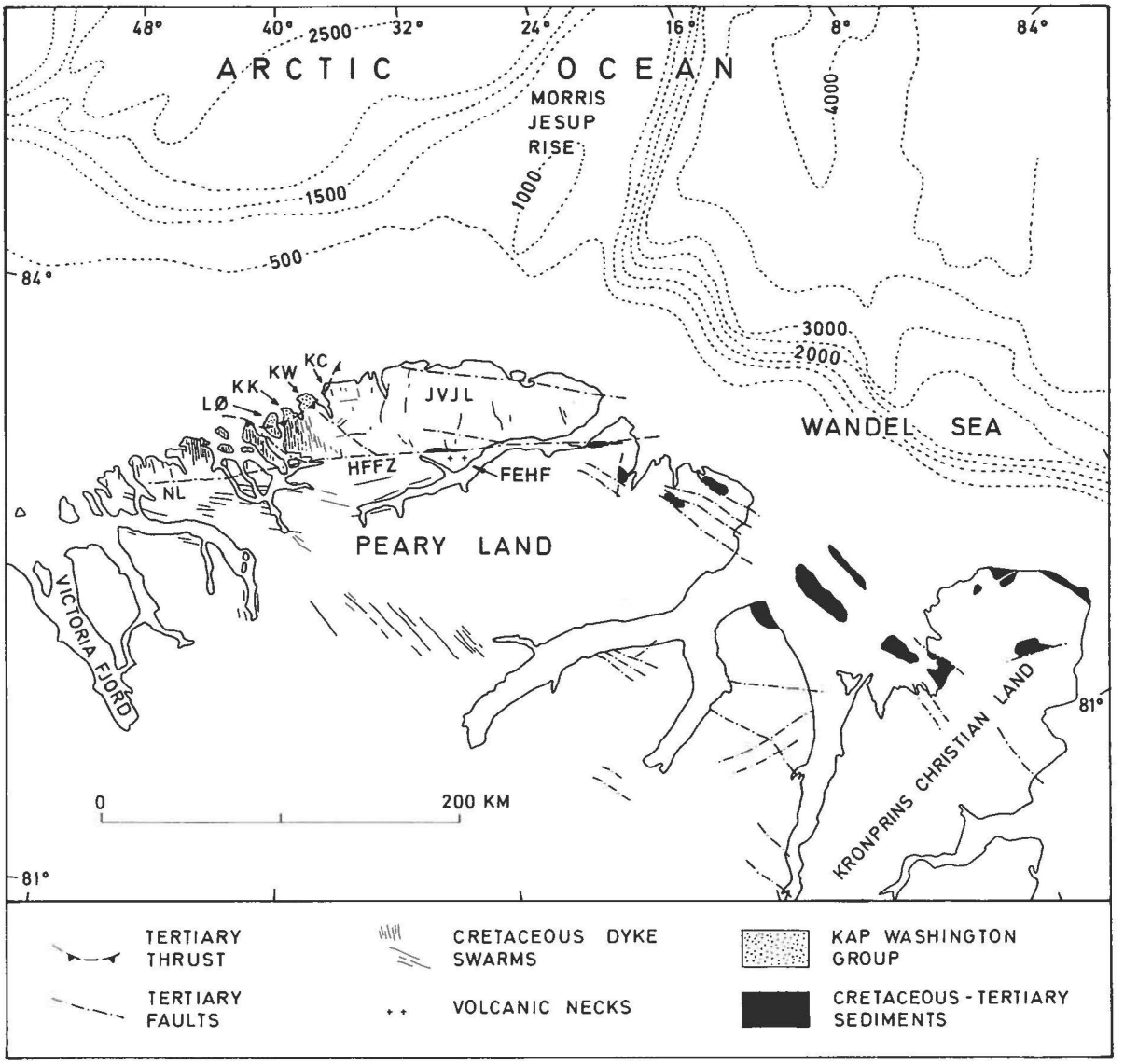


Fig. 2. Cretaceous–Tertiary magmatic and tectonic features of the northern margin of Greenland. Bathymetry in metres from Johnson et al. (1979). NL, Nansen Land; JVJL, Johannes V. Jensen Land; LØ, Lockwood Ø; KK, Kap Kane; KW, Kap Washington; KC, Kap Cannon; HFFZ, Harder Fjord fault zone; FEHF, Frederick E. Hyde Fjord. Data from the authors' field work and from references quoted in the text.

magnetite, iddingsite, etc. A single pyroxene is present, evidently a strongly titaniferous augite, and both the pyroxene and feldspar usually show some degree of alteration. Ilmenite is abundant in many samples.

A selection of major and trace element analyses of dykes is given in Table 1; average compositions of dyke swarms and sills are given in Table 2. Here, comment is confined to the essential chemical characteristics of the suite. Its alkalic character is indicated by exceptionally high TiO_2 (usually > 4 per cent), high $Na_2O + K_2O$ (> 4 per cent) and high P_2O_5 and total iron. All analyses fall in the alkalic field of an alkali–silica plot (Fig. 3) although there is a scatter extending into the subalkaline field as defined by Irvine & Baragar (1971). All the dykes are olivine-normative; some contain nepheline and hypersthene in the norm. This is thought to be a consequence of alkali and silica mobility during alteration. The alkalic affinity of the suite is confirmed by the immobile element distribution. For example, on a TiO_2–Zr/P_2O_5 plot, adopted by Floyd & Winchester (1975) for the discrimination of alkali and tholeiitic basalts, the 40 dykes for which Zr has been determined show a pronounced 'vertical' trend characteristic of alkalic compositions (Fig. 4).

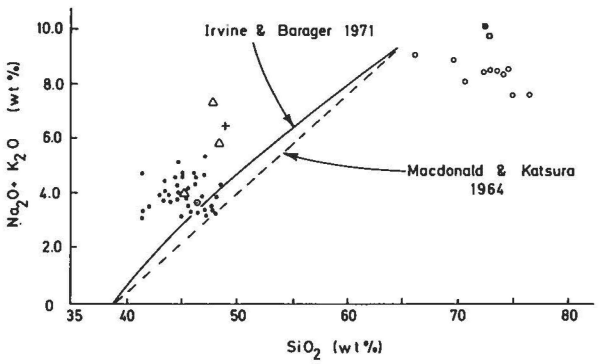


Fig. 3. Alkali–silica diagram for 50 igneous samples from the northern margin of Greenland. Full dots: dolerite dykes and sills; triangles: samples from a single composite dyke; open circles: Kap Washington rhyolitic volcanics; cross: Kap Washington basalt lava. Circle with dot represents average of 28 alkalic basalts from Hawaii (Macdonald & Katsura 1964), open square is average of four peralkaline rhyolites (pantellerites) from Ethiopia (Barberi et al. 1975) and the solid square is average of nine peralkaline rhyolites (comendites) from Kangerdlugssuaq, East Greenland (Brooks & Rucklidge 1976). Full and dashed lines divide the alkaline and tholeiitic basalts of Macdonald & Katsura (1964) and alkaline and subalkaline fields of Irvine & Barager (1971).

The dyke swarms were emplaced in continental crust, and show trace element distributions believed to be characteristic of 'within-plate' magmatism (Fig. 5). We infer a continental rift environment for the dyke swarms and return to the geotectonic implications of this in a later section.

The age of the main swarm is defined on geological evidence. Steeply inclined N–S-trending sheets and dykes cut Carboniferous–Permian and early Cretaceous sediments on western Lockwood Ø and Kap Kane but fail to penetrate the Kap Washington Group volcanic rocks (Brown & Parsons 1981) which are dated at about the end of the Cretaceous (Larsen et al. 1978, Batten et al. 1981). The main dyke swarm is almost certainly of late Cretaceous age and there is nothing to suggest that the compositionally identical WNW–ESE and E–W sets are not of similar age. A K–Ar whole-rock age of 66 Ma has been obtained on an E–W dyke (Dawes & Soper 1971).

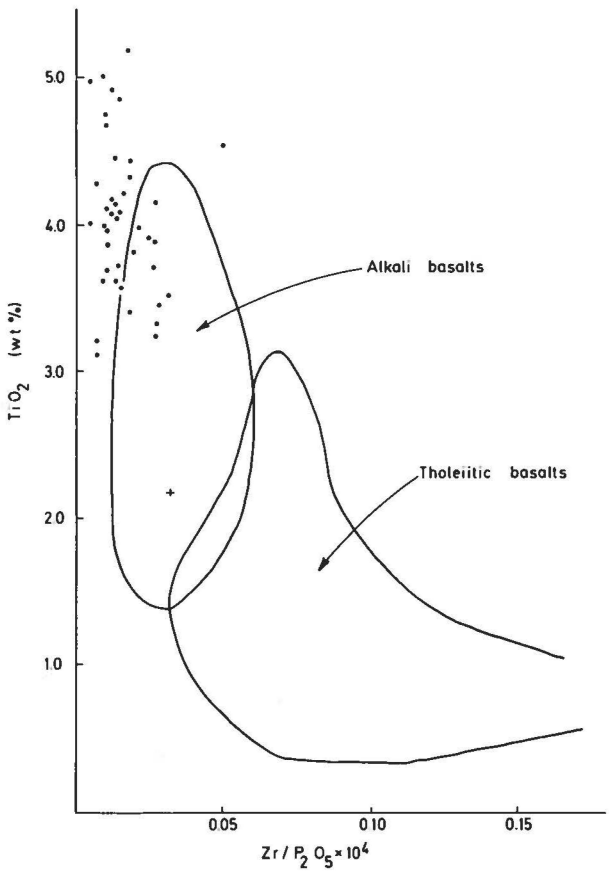


Fig. 4. A TiO_2–Zr/P_2O_5 diagram showing plots of 40 dolerite dykes and a single sample of Kap Washington basalt (cross). Fields of alkali and tholeiitic basalts after Floyd & Winchester (1975).

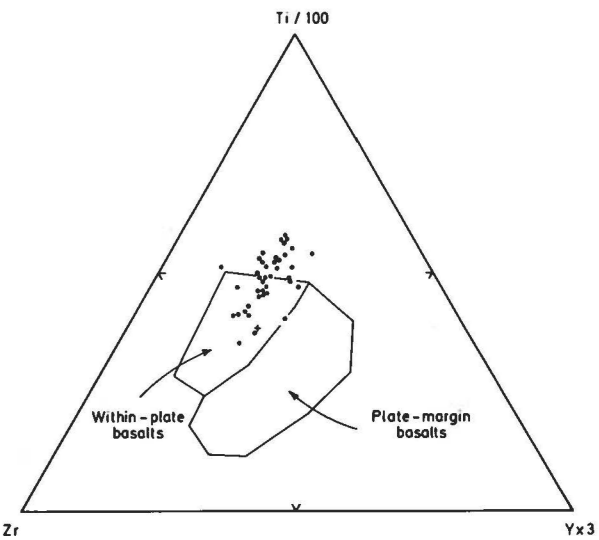


Fig. 5. A discrimination diagram using Ti, Zr and Y showing plots of 40 dolerite dykes and a single sample of Kap Washington basalt (cross). Fields of 'within-plate' and 'plate-margin' basalts after Pearce & Cann (1973).

Table 1. Chemical analyses of dolerite dykes.

	1	2	3	4	5	6	7	8	9	10	11
GGU sample No.	11351	11398	53435	53436	53473	53482	53483	53545	100606	100610	100620
SiO_2	41.98	47.71	44.45	44.49	46.17	43.01	47.18	44.11	46.38	44.50	46.08
TiO_2	4.98	3.72	4.43	5.27	3.81	4.00	3.68	4.92	4.07	3.63	3.89
Al_2O_3	11.64	12.93	13.18	12.44	14.80	13.84	13.43	12.68	14.60	14.66	14.31
Fe_2O_3	5.23	3.34	2.63	3.79	1.78	5.64	4.85	5.45	2.72	3.91	2.30
FeO	10.29	10.47	11.82	11.33	11.21	8.57	8.24	9.56	10.45	9.49	12.18
MnO	0.30	0.22	0.24	0.25	0.23	0.22	0.27	0.24	0.21	0.22	0.21
MgO	5.87	6.25	5.77	5.73	5.51	7.59	4.61	5.94	5.38	6.22	5.46
CaO	10.60	9.71	9.68	9.67	9.67	8.79	8.17	9.90	9.20	9.71	9.36
Na_2O	2.67	2.53	3.57	2.80	3.42	2.56	3.84	3.24	3.35	3.51	3.21
K_2O	0.92	0.88	1.04	1.00	0.96	1.39	1.50	0.69	1.30	0.80	1.17
P_2O_5	2.35	0.56	1.31	1.16	1.22	1.29	1.73	1.12	1.28	0.91	0.81
H_2O+	1.71	1.08	1.66	1.58	0.99	2.82	1.79	1.09	1.54	2.28	0.98
Total	98.54	99.40	99.78	99.51	99.77	99.73	99.29	99.94	99.77	99.84	99.96
CIPW weight norm											
or	5.44	5.20	6.26	5.91	5.73	8.45	8.86	4.08	7.80	4.85	6.97
ab	22.59	21.41	25.12	23.69	29.28	22.26	32.49	26.29	28.47	23.98	26.86
an	17.06	21.33	17.14	18.42	22.37	22.83	14.98	18.02	21.28	22.39	21.40
ne	–	–	3.03	–	–	–	–	0.61	0.16	3.46	0.30
di	16.72	19.15	19.04	18.18	14.94	10.98	11.86	19.65	13.74	17.20	16.73
hy	4.70	18.59	–	6.97	0.10	0.33	5.81	–	–	–	–
ol	12.22	1.55	14.70	9.06	14.70	20.91	9.76	14.14	14.83	15.98	15.54
mt	2.87	2.58	2.77	2.82	2.48	2.69	2.42	2.77	2.51	2.55	2.76
il	9.46	7.07	8.57	10.01	7.31	7.81	6.99	9.34	7.86	7.05	7.46
ap	5.44	1.30	3.08	2.69	2.85	3.08	4.01	2.59	3.01	2.15	1.90
py	–	–	0.80	–	0.80	0.90	–	–	0.80	0.50	0.10
Trace elements (ppm)											
Sr	589	471	758	699	662	574	563	678	741	695	516
Rb	26	20	14	14	16	23	24	11	21	19	25
Zn	n.d.	n.d.	149	n.d.	119	89	n.d.	n.d.	140	152	102
Cu	20	50	20	11	46	31	11	42	24	47	78
Ni	8	51	16	17	37	26	4	41	24	63	56
Ba	626	441	1071	708	1035	1961	1260	669	980	971	639
Nb	n.d.	n.d.	32	n.d.	23	29	n.d.	n.d.	28	26	29
Zr	108	148	230	209	233	113	173	130	156	120	215
Y	47	33	42	42	34	31	24	36	32	31	37

Samples 1 and 2 collected by K. Ellitsgaard-Rasmussen in 1950, 3 to 11 by P. R. Dawes and N. J. Soper in 1969.
Samples 1, 2, 4, 7 and 8: major elements by Ib Sørensen, GGU, 1978; trace elements by J. C. Bailey, University of Copenhagen, 1980. Samples 3, 5, 6, 9, 10 and 11: all major and trace elements (except H_2O^+) by J. G. Holland, University of Durham, U. K. 1972; H_2O^+ by Ib Sørensen, GGU, 1978. n.d. = not determined. Norms are calculated with Fe_2O_3 adjusted to 0.15.

The Kap Washington Volcanic Group

Limited observations made in 1969 (Dawes & Soper 1970, 1973) revealed the existence on the northern coast of Peary Land, within the North Greenland fold belt, of a varied suite of post-Palaeozoic volcanic rocks, for which the name Kap Washington Group was adopted. It was established that the volcanic rocks are tilted southwards, sheared and in places cleaved and in thrust contact with Palaeozoic metasediments of the fold belt (Fig. 6). Observations in 1980 showed that their outcrop is confined to the northern parts of the peninsulas between Kap Cannon and Lockwood Ø (Fig. 2), a more limited area than had been inferred from aerial photograph interpretation (Dawes 1976).

Field identifications of rhyolitic lavas, welded ash-flows, andesites and basic flows, water-lain tuffs, accretionary lapilli tuffs, agglomerates and breccias, with some intrusives, were made in 1969 and an explosive calc-alkaline character was inferred for the suite. Sampling was unfortunately confined largely to the felsic flows. These are typically composed of a microcrystalline matrix, with phenocrysts of perthitic K–Na feldspar and partly resorbed quartz. Flow banding characterises some rhyolitic types, although generally



Fig. 6. The Kap Washington volcanics at Kap Cannon (KC) viewed from the south-west. The volcanics are structurally overlain by metamorphosed Lower Palaeozoic and ?older strata of the North Greenland fold belt. The Kap Cannon thrust is marked by a dashed line. Photograph taken May 1969.

textures show considerable evidence of shearing and alteration. The single basic flow sampled is an olivine basalt with drusy areas which contain alkali feldspar. Detailed mapping of the volcanic rocks has now been undertaken by Brown & Parsons (1981) who report that the group has a total thickness in excess of 5 km and that considerable stratigraphic variation occurs from area to area.

Chemical analyses so far available are all, except one, of felsic flows (Table 3). These are difficult to characterise chemically, but they fall in the subalkaline field of an alkali–silica plot (Fig. 3); the single analysed sample from a basic flow is thoroughly alkaline and this is confirmed by its immobile element distribution (Fig. 4). The relatively high values of Nb, Zr and Y conform to normal values for alkaline rocks, while two samples show normative acmite; a clear indication of peralkaline character. These rocks correspond to comendites. An average composition of ten acid volcanics is given in Table 2.

Table 2. Average major element compositions of dolerite intrusions and Kap Washington Group volcanics.

	E–W dykes	Northerly dykes	Sills	Kap Washington volcanics	
No. of analyses	16	16	4	1	10
SiO_2	45.88	45.08	45.75	49.04	72.64
TiO_2	3.95	4.34	4.01	2.18	0.35
Al_2O_3	13.78	13.42	12.31	14.17	12.99
Fe_2O_3	4.95	4.25	3.94	6.23	3.01
FeO	8.83	9.88	10.99	4.59	1.10
MnO	0.21	0.23	0.23	0.20	0.11
MgO	5.67	5.54	5.60	7.09	0.14
CaO	9.18	9.40	8.44	7.21	0.38
Na_2O	3.14	3.03	2.85	3.58	4.18
K_2O	0.98	1.07	1.41	2.86	4.17
P_2O_5	0.95	1.20	1.12	0.42	0.04
H_2O+	2.18	1.89	3.09	2.43	0.89
Total	99.70	99.33	99.74	100.00	100.00

Details of the 11 analyses of Kap Washington Group volcanics are given in Table 3. E–W dyke average composed of samples GGU 11398, 53482*, 53483, 53487, 53493, 53534*, 100626, 100627*, 218978, 218994, 254647, 254737, 255781, 270831, 270832 and 270833. Northerly dyke average composed of samples GGU 11323, 11345, 11351, 11363, 53425, 53435*, 53436, 53473*, 53555, 100606*, 100610*, 100620*, 100622, 218991, 255697 and 255714. Sill average composed of samples 53454, 53455, 53459 and 255702. Analyses of the samples marked by an asterisk by J. G. Holland, University of Durham, U.K. 1972; the rest by Ib Sørensen, GGU, 1978 and 1980.

Table 3. Chemical analyses of Kap Washington Group volcanics.

	1	2	3	4	5	6	7	8	9	10	11
GGU sample No.	53443	53444	53446	53447	53448	53450	53451	53452	53453	53466	53449
SiO_2	69.80	72.99	66.28	70.82	76.65	74.12	74.58	73.49	75.06	72.59	49.04
TiO_2	0.49	0.19	0.27	0.45	0.59	0.27	0.31	0.37	0.35	0.21	2.18
Al_2O_3	13.14	14.25	18.21	14.39	13.63	10.02	11.73	10.58	10.84	13.08	14.17
Fe_2O_3	4.05	1.51	1.81	3.21	0.36	6.25	3.08	4.55	4.48	0.82	6.23
FeO	0.70	1.49	2.16	1.18	0.59	0.40	1.30	1.50	0.80	0.88	4.59
MnO	0.37	0.05	0.05	0.14	0.01	0.13	0.06	0.16	0.07	0.06	0.20
MgO	0.12	0.19	0.14	0.18	0.08	0.07	0.06	0.11	0.20	0.29	7.09
CaO	1.29	0.02	0.02	0.02	0.02	0.03	0.02	0.42	0.16	1.83	7.21
Na_2O	3.56	5.23	2.50	3.82	7.17	3.51	4.03	3.71	4.00	4.27	3.58
K_2O	5.33	3.24	6.59	4.26	0.36	4.91	4.46	4.77	3.59	4.16	2.86
P_2O_5	0.11	0.02	0.03	0.02	0.03	0.03	0.02	0.03	0.04	0.05	0.42
H_2O+	1.04	0.82	1.94	1.51	0.51	0.26	0.35	0.31	0.41	1.76	2.43
CIPW weight norm											
Q	23.69	28.28	24.34	29.88	33.40	32.77	31.42	30.39	34.55	28.08	–
C	–	2.16	7.10	3.54	1.45	–	0.26	–	0.19	–	–
or	31.85	19.32	39.71	25.59	2.13	29.08	26.48	28.31	21.28	25.00	17.32
ab	30.46	44.60	21.58	32.83	61.01	24.30	34.27	27.91	34.02	36.81	24.32
an	4.16	–	–	–	–	–	–	–	0.53	4.30	14.49
ne	–	–	–	–	–	–	–	–	–	–	3.65
ac	–	–	–	–	–	2.31	–	2.15	–	–	–
ns	–	–	–	–	–	0.67	–	0.26	–	–	–
di	1.39	–	–	–	–	–	–	1.68	–	3.95	15.83
hy	6.05	4.58	5.81	6.18	0.64	9.74	5.90	8.15	7.36	0.96	–
ol	–	–	–	–	–	–	–	–	–	–	16.64
mt	0.84	0.55	0.74	0.79	0.18	–	0.78	–	0.93	0.32	2.00
il	0.95	0.36	0.53	0.87	1.12	0.51	0.59	0.70	0.66	0.40	4.25
ap	0.25	0.12	0.12	0.12	0.12	0.18	0.12	0.07	0.09	0.12	1.00
Trace elements (ppm)											
Sr	17	35	15	47	74	8	7	24	23	202	323
Rb	146	144	422	135	13	288	206	159	141	159	91
Zn	204	251	451	214	46	347	252	286	294	29	121
Cu	5	4	6	6	3	8	6	8	8	4	150
Ni	1	4	3	2	4	2	4	4	4	3	96
Ba	224	239	195	62	55	20	40	47	147	874	446
Nb	82	106	137	78	71	299	151	161	170	46	29
Zr	604	804	1092	530	444	1837	908	1023	1070	242	133
Y	55	69	105	71	37	145	86	98	103	20	28

Samples collected by P. R. Dawes and N. J. Soper in 1969.
All major and trace elements (except H_2O^+) by J. G. Holland, University of Durham, 1972. H_2O^+ by Ib Sørensen, GGU, 1978. Major analyses recalculated to 100%. Norms are calculated with Fe_2O_3 adjusted to 0.15.

Preliminary petrological studies by Brown & Parsons (1981) confirm the thoroughly alkalic nature of the province. They report riebeckite trachytes, riebeckite comendites or pantellerites and alkalic basalts as the chief lava types; the presence of the soda amphibole riebeckite in both siliceous lavas and ash-flow tuffs demonstrates that the Kap Washington Group is of peralkaline character.

Peralkaline rocks are well known world-wide from regions associated with crustal attenuation and rift tectonics (e.g. Africa – Red Sea region, Gass 1970, Barberi et al. 1972) and are known from the North Atlantic (e.g. East Greenland, Brooks & Rucklidge 1976). We infer an extensional rift environment for the volcanicity, linking the origin of the province to that of the slightly earlier swarms of alkalic basic dykes.

The Kap Washington Group volcanic rocks overlie a sequence of shales and sandstones which have yielded angiosperm leaf fragments that are not older than mid-Cretaceous in age (B. E. Koch, pers. comm. 1981). Shales interbedded with the volcanic rocks about 3 km above the base have yielded pollen and spores of end-Cretaceous (Campanian or Maastrichtian) age (D. J. Batten, pers. comm. 1981). Thus at least the basal 3

km of the volcanic pile is of late Cretaceous age and we adopt an age range of 75–65 Ma for the volcanicity — at or just prior to the Cretaceous–Tertiary boundary (Batten et al. 1981). The upper age limit is within the experimental error of the earlier Rb–Sr whole-rock age on five rhyolite samples of 63 Ma (Larsen et al. 1978).

Kap Cannon thrust zone

The outcrop of the Kap Washington volcanic rocks is terminated to the south at a major south-dipping thrust zone along which the highly-deformed metasediments of the North Greenland fold belt are displaced relatively northwards over the volcanic rocks (Dawes & Soper 1970, Fig. 6). Mylonitisation occurs along the thrust plane, which has been traced from Kap Cannon as far west as Lockwood Ø. Widespread retrograde metamorphism affects the metasediments along the north coast, especially near the thrust, and K–Ar mineral ages of 84–42 Ma from Palaeozoic mica schists (Dawes & Soper 1971) support the idea of a weak Tertiary overprinting connected with the thrusting.

Mapping in 1980 revealed at least two further steep thrusts or reversed faults to the north of the main Kap Cannon thrust; the thrust slices contain Carboniferous–Permian sediments and Palaeozoic flysch in addition to the volcanic rocks. The volcanic rocks record only the effects of Tertiary deformation; there is little folding but locally strong cleavage development, brecciation and shearing. Displacement on the thrusts is not known. Inclination of the main thrust plane ranges from 40° at Kap Cannon in the east to near vertical on the west side of Lockwood Ø. The structures may be of upthrust type rather than far-travelled nappes.

The thrusting episode post-dates the Kap Washington Group. K–Ar whole-rock ages of 32 and 35 Ma (early Oligocene) from somewhat cleaved rhyolite samples (Dawes & Soper 1971) are perhaps best interpreted as indicating erosion and cooling during post-thrusting uplift. A late Paleocene – Eocene age for the compression event is adopted.

Wandel Sea Basin structure

South-east-trending faults and open folds affect the Wandel Sea Basin cover sequences of eastern Peary Land and Kronprins Christian Land (Haller 1970, Dawes 1976, Håkansson 1979, Håkansson et al. 1981). These structures are parallel to the Wandel Sea continental margin and Spitzbergen fracture zone. Faults of similar trend are present in adjacent parts of the Palaeozoic platform and several major examples have been mapped recently in the fold belt north of the Harder Fjord fault zone (Fig. 2). Håkansson (1979) has shown that the main block faulting affected eastern Peary Land after deposition of the youngest recorded cover sediments (?Paleocene) and that the sense of displacement across the faults implied by the geometry of drag folds is opposite to that determined from matching the stratigraphy. He suggests that faults of pre-Tertiary (perhaps Jurassic) age were reactivated in post-Paleocene time by tension followed by compression, or vice versa. As discussed below the latter is more consistent with displacement models.

Harder Fjord fault zone

This major E–W-trending fault zone has been traced for over 250 km across the North Greenland fold belt from the Wandel Sea coast to Nansen Land (Fig. 2). Its eastern portion takes the form of a narrow graben in which representatives of the post-Franklinian cover sequence are preserved. These comprise non-marine upper Permian, marine Cretaceous, and non-marine early Tertiary (Paleocene) sediments (Croxton et al. 1980, Wagner et al. in press), as well as greenstones, subaqueous extrusives in part, of unknown age. Doleritic dykes are associated with the central part of the fault zone and, also in this area, a few kilometres to the south of the fault several breccia pipes and necks occur, which contain basement gneiss xenoliths and volcanic lithologies comparable to some of those in the Kap Washington Group (Soper et al. 1980, Parsons 1981).

The origin of the fault zone is uncertain. Surlyk et al. (1980) postulate that the fault had a considerable influence on the early development of the Franklinian trough sedimentation, although the direct field evidence of this has been questioned (Higgins et al. 1981). The fault zone probably influenced the pattern of emplacement of the late Cretaceous basic dyke swarms, the breccia pipes and necks and the greenstones could be directly linked to the existence of the fault, and there were substantial movements subsequent to the deposition of the early Tertiary sediments. Some basic dykes aligned in the fault zone in the Frederick E. Hyde Fjord area show evidence of post-dyke crushing (Dawes & Soper 1979). Recent solfatara activity is recorded from its eastern end and earthquake activity along the fault zone indicates continuing movement (Dawes & Peel 1981).

Dawes (1973) noted the parallelism of this and other large faults in Peary Land to the Spitzbergen fracture zone, although recent work has not revealed clear evidence of wrench displacement. For example, 25 km south of Lockwood Ø the plane of the northern branch of the Harder Fjord fault zone is well exposed, inclined 60° to the south. Downthrow is to the south and equivalent to the whole thickness of the Polkorridoren Group — as much as 2 km (Higgins et al. 1981). It is thought to be associated with a period of stress relaxation after the Kap Cannon thrusting episode and probably marks the main uplift of the north Peary Land mountain chain in mid- to late Tertiary time.

Spreading history of the ocean basins around Greenland

The chronology of sea-floor spreading in the oceans around Greenland is now established in outline, although the origin of Baffin Bay and the whole Amerasia Basin remains unclear. There are problems concerning the timing of the onset of spreading in many areas and the degree to which thinning and extension of the continental crust were involved. We briefly review the spreading history of the North-East Atlantic, Eurasia Basin and Labrador Sea – Baffin Bay.

North-East Atlantic

In Mesozoic time zones of crustal extension were already developing on the sites of Greenland's future rifted margins. A shallow seaway was initiated between Greenland and Scandinavia in late Palaeozoic time and a marine connection was maintained through it between the Boreal Ocean and western Europe for much of the Jurassic and Cretaceous, associated on the Greenland side with coast-parallel extensional faulting (Birkelund & Perch-Nielsen 1976, Surlyk 1977).

Anomaly 24 (52 Ma) is the earliest recognised with certainty in the Norwegian Sea (Vogt et al. 1970) and between South-East Greenland and Rockall (Vogt & Avery 1974) and the general view has been that spreading commenced between Greenland and Europe shortly before anomaly 24 time (early Eocene), preceded by an abortive spreading phase in the Cretaceous which opened Rockall Bight. However, anomaly 24 lies oblique to the shelf edge in some areas and as much as 100 km to seaward of it. Phillips et al. (1978) have interpreted the intervening areas as oceanic crust generated by pre-anomaly 24 spreading. These areas could equally represent thinned and subsided continental crust as found, for example, off South-East Greenland by Featherstone et al. (1977). Two lines of evidence confirm the initiation of spreading at anomaly 24 time. The eruption of very large volumes of tholeiitic basalt onto the East Greenland continental margin can realistically only be associated with major rifting at the onset of spreading. The main basalt pile in this area has been dated palynologically as late Paleocene – early Eocene (55–53 Ma, Soper et al. 1976) and this is consistent with a fission-track age of 54.6 ± 1.7 Ma on the Skaergaard intrusion (Brooks & Gleadow 1977). The reversed magnetic polarity of the basalt sequence confirms its eruption in the 24–25 reversed epoch (Faller 1975). Secondly, a pronounced shift in spreading geometry occurred in the Labrador Sea at anomaly 24–25 time and this must reflect a major change in plate configuration. The evidence thus confirms that the initiation of the main phase of spreading between Greenland and Europe took place at about the Paleocene–Eocene boundary (53 Ma) (Talwani & Eldholm 1977).

Eurasia Basin and Spitzbergen fracture zone

The ocean floor of the Eurasia Basin was generated by spreading on the Nansen axis which connects to the North Atlantic via the Spitzbergen fracture zone and the Knipovich Ridge (De Geer line). The outermost ridge flank anomaly recognised in the Eurasia Basin is 24 (52 Ma), but there is room for anomalies 25 to 27 between it and the continental slope (Vogt et al. 1979). The intervening zone may be floored by oceanic crust as old as 62 Ma or by thinned and subsided continental crust. Limited geological evidence bearing on this early extensional phase is discussed subsequently. Notable features in the present context are the Morris Jesup Rise and Yermak Plateau, northern extensions of the North Greenland and Svalbard continental shelves. These are interpreted by Feden et al. (1979) as aseismic ridges generated by a Yermak hot spot in the period between anomalies 18 and 13 (40–35 Ma) and subsequently rafted apart by post-anomaly 13 spreading. Johnson et al. (in press) interpret that part of the Yermak plateau south of 82°N as continental crust. These authors, and Phillips et al. (1978), Feden et al. (1979) and Kovacs (this volume), all associate the Yermak hot spot with a triple rift junction north of Greenland, active until anomaly 13 time, whose third arm was represented by the Nares Strait lineament. These models require substantially greater displacement on the Nares Strait line than is permitted by the geological constraints (Christie et al. 1981, Dawes & Kerr, this volume).

Geometrical problems are also associated with the De Geer line along which the Barents Shelf has been displaced dextrally with respect to the Wandel Sea margin of North-East Greenland. Early reconstructions which juxtaposed the coastlines or continental margins (Bullard et al. 1965, Bott & Watts 1971, Pitman & Talwani 1972) involved unacceptably large overlaps in this region. Reconstructions by means of magnetic anomaly matching (Talwani & Eldholm 1977, Phillips et al. 1978) produced small overlaps between the Wandel Sea and Barents shelves until about anomaly 13 time, implying oblique compression until the early Oligocene followed by oblique extension.

Labrador Sea and Baffin Bay

The origin of the Labrador Sea – Baffin Bay seaway is still poorly understood. The nature of the crust is uncertain and so therefore is the position of the continent–ocean boundary; many reconstructions show coastline mismatches too large to be readily accommodated by within-plate deformations.

Linear magnetic anomalies in the southern Labrador Sea are contiguous with ocean floor anomalies 20–24 in the North Atlantic with which they define an ancient triple junction (Vogt et al. 1969, Laughton 1971, Vogt & Avery 1974). It is widely accepted that the Labrador Sea was opened by a phase of sea-floor spreading which

terminated after anomaly 13 time — early Oligocene (Kristoffersen & Talwani 1977). However, the extent of this spreading and its time of onset are in doubt. A number of authors, most recently Srivastava (1978), recognise anomalies as old as 32 (75 Ma) and Srivastava suggests a northward propagation of the spreading axis, with anomalies 31 and 28 the oldest present in the central and northern sectors of the Labrador Sea. Conversely van der Linden (1975) showed, mainly on the basis of seismic refraction data, that the north-western margin of the Labrador Sea could be floored by thinned continental crust and Hinz et al. (1979) confirmed that the continent–ocean boundary cannot be defined in this region by means of conventional geophysical characteristics. On the basis of borehole and seismic reflection data, Grant (1980, this volume) established that an early Cretaceous unconformity, presumably overlying continental basement, occurs in the same area as anomaly 28 in Srivastava's model, suggesting that the outer magnetic anomalies in the Labrador Sea have been incorrectly identified. Spreading may well have commenced significantly later than 75 Ma, but preceded in Cretaceous time by extensional thinning and subsidence of the continental crust.

The origin of Baffin Bay is even more problematical. Two schools of thought have emerged. One regards the thin (c. 5 km) crust of the deeper parts of the bay as oceanic, generated by spreading from a northward extension of the Labrador Sea axis (Keen & Barrett 1972, Keen et al. 1974, Kristoffersen & Talwani 1977, Srivastava 1978). This is supported by Jackson et al. (1979), who report linear magnetic anomalies in central Baffin Bay which they correlate with ocean floor anomalies 13–21 and a central linear gravity low which they associate with the extinct spreading axis. The other school interprets the bay as a thinned and subsided continental region (Kerr 1967, 1980, van der Linden 1975), thus avoiding some of the geometrical problems associated with a 'pure spreading' hypothesis.

Two lines of geological evidence must be considered in relation to the problem. One concerns the small sinistral wrench displacement — as little as 25 km — which appears permissible on the Nares Strait line (Christie et al. 1981). This effectively rules out a 'pure spreading' origin for Baffin Bay. It is impossible to accommodate the required displacement of Greenland plus the attached part of Ellesmere Island with respect to the remainder of the Canadian Arctic Archipelago without invoking within-plate deformations greatly in excess of those associated with the Eurekan structures. Much of Baffin Bay must be floored by continental crust. Smaller displacements can, however, be accommodated and are required by the conventional ideas of opening of the Labrador Sea.

The second line of evidence concerns the similarity between the basalt provinces of East and West Greenland, not only as regards the composition and volume of the picritic-tholeiitic extrusives, but also the pre-eruptive sedimentary history (Higgins & Soper 1981). By analogy with the geotectonic setting of the East Greenland province, we support the idea that the West Greenland basalts mark the major rifting episode in the Davis Strait (Clarke & Upton 1971), an idea adopted by Srivastava (1978) in his model. The Nûgssuaq basalts have been precisely dated palynologically as Danian (Jürgensen & Mikkelsen 1974), a date supported by Ar 40–39 ages in the range 62–57 Ma (Parrott & Reynolds 1975). The dominantly reversed polarity of the West Greenland basalts (Athavale & Sharma 1975) thus indicates eruption in the 3 Ma reversed epoch between anomalies 26 and 27 (61–58 Ma). This is the most probable date at which rifting propagated through the Davis Strait into Baffin Bay — somewhat earlier than envisaged by Kristoffersen & Talwani (1977) or Srivastava (1978). It should be noted that a coast-parallel basic dyke swarm of supposed Jurassic age in South-West Greenland has been associated by Watt (1969) with earlier tensional rifting in the region.

A complex history for the Labrador Sea and Baffin Bay must therefore be envisaged, with the attenuation and 'oceanisation' of continental crust as particularly important, plus some phases of spreading. Assuming spreading occurred in Baffin Bay it must have been extremely slow (close to the rotation pole) until anomaly 21 time, with a change in geometry at anomaly 24 (Srivastava 1978) and a final phase of slow, oblique spreading in the period of anomalies 21–13. The displacement vector for this last phase adopted by Srivastava (1978) is parallel to Nares Strait and the wrench motion required there to satisfy the Jackson et al. (1979) model for central Baffin Bay is in excess of 100 km, which can be ruled out. The vector derived by Kristoffersen & Talwani (1977) is about 35° oblique to Nares Strait and can in principle be resolved into sinistral slip parallel to the Strait and compression normal to the northern Ellesmere – North Greenland sector of the Eurekan fold belt. However, the slip component is still geologically unacceptably large and the displacement vector inclined at such an acute angle to the assumed Baffin Bay spreading axis as to render the model geometrically improbable. The geological constraints on the total wrench offset on Nares Strait thus raise serious problems for any hypothesis for the origin of Baffin Bay which involves significant sea-floor spreading and it questions the interpretations of Jackson et al. (1979) who identify sea-floor magnetic anomalies and a central spreading axis in the Bay.

Makarov Basin

Almost every conceivable origin has at some time been proposed for the Alpha Ridge (continental crust, spreading centre, subduction zone, incipient island arc) and for the Makarov Basin which lies between that and the Lomonosov Ridge (see Sweeney 1978). Recent interpretations of aeromagnetic data obtained in 1977–78

(Taylor et al. 1981, Kovacs, this volume) have tentatively identified sea-floor spreading type anomalies in the Makarov Basin and suggest that spreading began at anomaly 34 time (late Cretaceous) and continued to anomaly 19 (42 Ma, late Eocene). The implications of a model which involves simultaneous spreading on the Makarov and Nansen axes in the Eocene, but which also accepts strictly limited displacement on the Nares Strait line have yet to be explored.

Discussion

The extent to which the Cretaceous–Palaeogene history of North Greenland can be related to the changing geotectonic pattern outlined above, is discussed in the following.

Late Cretaceous events (pre-anomaly 29, 65 Ma)

A reconstruction at 65 Ma, (the Cretaceous–Tertiary boundary) shows North Greenland, the Barents Shelf and Lomonosov continental fragment in juxtaposition, prior to the onset of spreading on the Nansen axis but, if the proposals of Taylor et al. (1981) and Kovacs (this volume) are correct, with spreading already taking place in the Makarov Basin (Fig. 7A). The North Greenland dyke swarms were emplaced into this continental region. Their alkalic composition, 'within-plate' chemical character and spatial relationship to the later explosive alkalic volcanicity of the Kap Washington Group all point to their association with a continental rift system. The main concentration of coast-normal dykes aligns rather better with the Makarov than the Nansen axis. If Johnson et al. (in press) are correct in interpreting part of the Yermak Plateau as continental, and if the Nansen spreading axis did not propagate across it until about anomaly 13 time, then the dykes are unlikely to represent the southern termination of the Nansen axis. The only credible alternative, on present evidence, would then be that they represent the termination of the Makarov axis. However, the model of Johnson et al. requires that between the times of anomalies 24 and 13 the spreading axes of the Eurasia and North-East Atlantic basins were linked by a transform fault system through Svalbard. The improbability of this, together with the existence of pre-anomaly 24 crust of unknown origin in the Eurasia Basin leads us to favour a link between the coast-normal dyke swarm and early extensional rifting along the Nansen axis as implied in Fig. 7.

We note the contrast in composition between the North Greenland dykes and the tholeiitic basalts exposed on the Barents margin of the Eurasia Basin in Svalbard and Franz Josef Land (Tyrell & Sandford 1933, Gusev 1979), which are of comparable but poorly defined age (Dibner 1957, Gayer et al. 1966, Burov et al. 1976), and the fact that the alkaline olivine plateau basalt lavas of the Woodfjorden area in northern Spitzbergen, are now dated as late Tertiary (Prestvik 1978). On the grounds of both their age and alkalic composition, we do not relate either the basic dykes or the Kap Washington volcanics directly to the Yermak hot spot, as suggested by Feden et al. (1979). The geographical coincidence of the alkalic rift magmatism and later hot spot activity may well have petrogenetic significance, but exploration of this problem must await the completion of petrological studies of the North Greenland rocks and also sampling of the hot spot products from the submarine plateaux. Furthermore, in view of the late Cretaceous age assignment of the Kap Washington volcanics, it hardly seems likely that the Paleocene tuffs of Svalbard (Dypvik & Nagy 1978) were derived from the Kap Washington volcanism as has been suggested (Feden et al. 1979).

We regard the alkaline–peralkaline volcanicity of the Kap Washington Group as a further manifestation of the crustal tension and rift magmatism represented by the dykes. The timing of this final phase of alkalic magmatic activity has geotectonic implications of regional importance. The dyke swarms and rift volcanics show that the late Cretaceous tectonic regime in the region was extensional — as indeed it was throughout the whole crustal segment from the Sverdrup Basin through North Greenland to the northern Barents Shelf. The volcanic activity is therefore likely to pre-date the onset of spreading in the Labrador Sea, since the anti-clockwise rotation of Greenland with respect to North America while still pinned to the Eurasia plate clearly induced compression in the region, as evidenced by the later episode of Kap Cannon thrusting. The age of the Kap Washington Group may thus be regarded as a maximum for the onset of true spreading in the Labrador Sea, and places this event no earlier than end-Cretaceous.

This represents a significant revision of the time at which major anti-clockwise displacement of Greenland commenced. Supporting evidence comes from the Canadian Arctic Archipelago where a fundamental but hitherto unexplained change in tectonic regime also took place at about the end of the Cretaceous. The Mesozoic history of continuous subsidence of the Sverdrup Basin, with a sediment fill derived from extrabasinal sources, changed in latest Cretaceous time to one of differential vertical movements with uplifted blocks supplying mainly locally derived non-marine sediments of the Eureka Sound Formation (Trettin et al. 1972, Balkwill 1978). Basic intrusive and effusive activity, centred on Axel Heiberg Island, ceased at the same time. These changes are consistent with our proposed timing for the onset of spreading in the Labrador Sea, and also with Srivastava's location of the rotation pole for the early opening phase (prior to anomaly 25, 56 Ma) immediately south of the Sverdrup Basin.

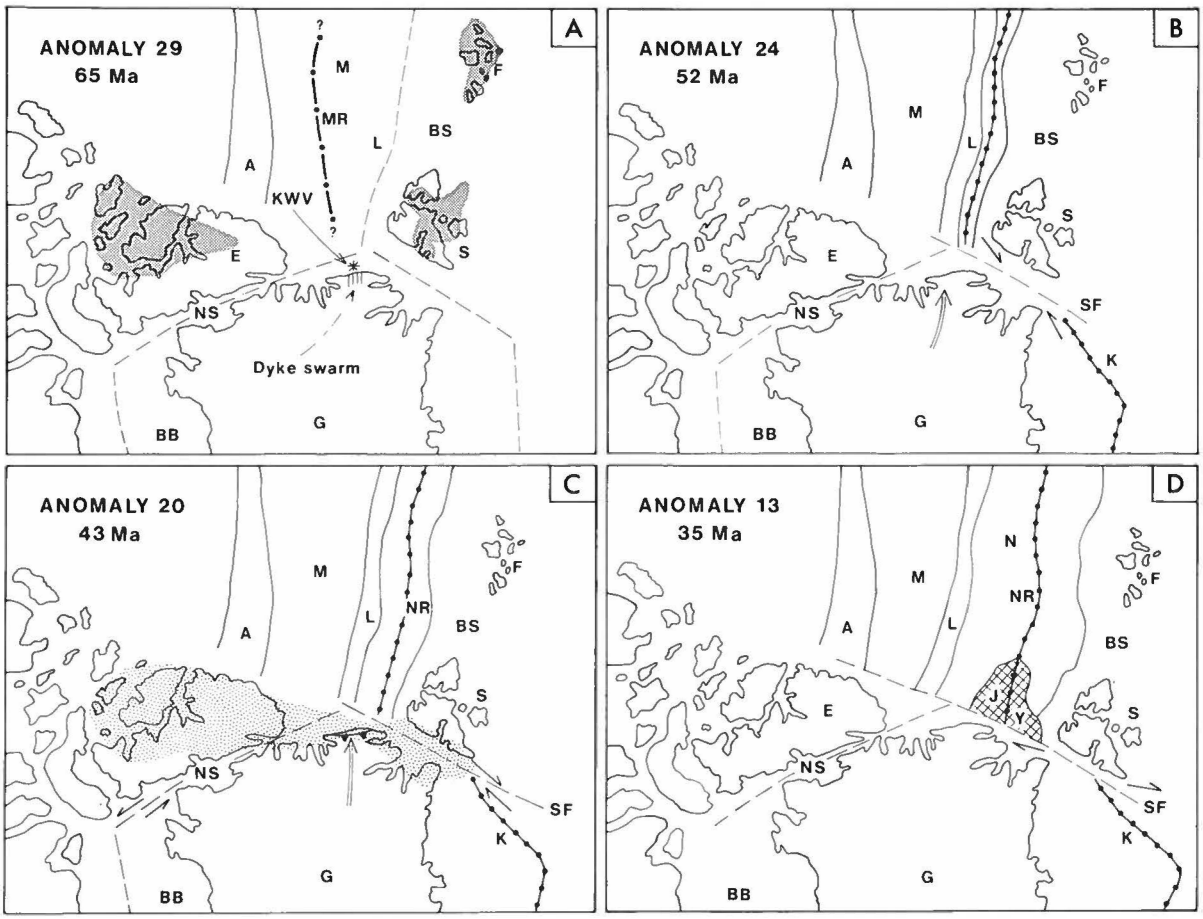


Fig. 7. Cartoon reconstructions of northern Greenland, eastern Canadian Arctic and the Barents Shelf. Fracture zones are shown dashed, active spreading axes are dotted, dot-dash line represents a possible extinct spreading axis, and plate movement direction is shown by single arrows. The fit of northern Greenland and the Barents Shelf approximately follows Phillips et al. (1978); fit of Greenland and Ellesmere Island approximately after Le Pichon et al. (1977).
G – Greenland, E – Ellesmere Island, S – Svalbard, F – Franz Josef Land, BS – Barents Shelf, A – Alpha Ridge, MR – Makarov Ridge, L – Lomonosov Ridge, NR – Nansen (Gakkel) Ridge, K – Knipovich Ridge, BB – Baffin Bay, M – Makarov Basin, N – Nansen Basin, SF – Spitzbergen Fracture Zone, NS – Nares Strait fracture.

A. Reconstruction at anomaly 29 time (Cretaceous–Tertiary boundary) showing main Mesozoic magmatic events. KWV – Kap Washington volcanics (75–65 Ma); shaded areas represent mainly basaltic magmatism (Jurassic–Cretaceous).
B. Reconstruction at anomaly 24 time (early Eocene). The double arrow represents the inferred compression direction in North Greenland.
C. Reconstruction at anomaly 20 time (middle Eocene) showing minor inferred strike-slip displacement along Nares Strait, the Kap Cannon thrust and the areas affected by the Eurekan orogeny in Canada and Greenland and the West Spitzbergen – Wandel Sea orogenic zone (shaded).
D. Reconstruction at anomaly 13 time (early Oligocene). Sea-floor spreading ceases between Greenland and North America. Arrows on Spitzbergen fracture zone indicate the inferred extensional displacement. J – Morris Jesup Rise, Y – Yermak Plateau.

Anomaly 29 to anomaly 13 (65 Ma to 35 Ma)

This represents the period of sea-floor spreading in the Labrador Sea. Spreading commenced on the Nansen axis perhaps as early as anomaly 27 time (61 Ma) and in the North-East Atlantic as late as 52 Ma. Thus for a period in the Eocene and early Oligocene (52–35 Ma) Greenland underwent displacements with respect to all adjacent continental masses. A geometrical consequence of this was the eastward migration of the southern end of the Nansen axis past the North Greenland continental margin, presumably along a major transform fault associated with the Spitzbergen fracture zone (Fig. 7B). A second consequence was the compression of North Greenland and Ellesmere Island against the Lomonosov continental fragment, producing the E–W portion of the Eurekan fold belt.

The main associated structure in North Greenland is the Kap Cannon thrust complex whose Eocene age is rather poorly defined geologically but entirely consist-

ent with the spreading chronology adopted here. Trettin & Balkwill (1979) have emphasised the apparent discrepancy in timing between the opening of the Labrador Sea and Baffin Bay (previously thought to be late Cretaceous – Paleocene) and the supposedly associated compressional phase of the Eurekan orogeny in the eastern Canadian Arctic Islands (middle Eocene to early Miocene). Bearing in mind that the early Miocene date refers to the post-Eurekan cover sequence, and thus represents a minimum age for the closure of deformation, and the younger (early Oligocene) age for the cessation of spreading in the Labrador Sea inherent in the revised polarity time scale, our interpretation largely resolves the discrepancy. If, as we propose, spreading did not commence in the Labrador Sea until about the Cretaceous–Tertiary boundary and Greenland remained 'pinned' against the Barents Shelf until Eocene time, Eurekan compression could have been associated with the anti-clockwise rotation of Greenland and Ellesmere Island with respect to the North American plate and have been restricted to Eocene – early Oligocene time. Any sinistral displacement on the Nares Strait lineament is likely to have taken place during this period (Fig. 7C). The proposal that spreading was taking place simultaneously on the Makarov axis does not materially change this scenario, but does prove difficult to reconcile with limited displacement on the Nares Strait line.

The Yermak and Morris Jesup aseismic ridges were generated in the period 41–35 Ma (Fig. 7D) and a speculative connection can be made between compression and perhaps over-riding of the southern flank of the Yermak hot spot by the North Greenland continental margin, and excess basalt discharge from the adjacent section of the Nansen spreading axis. Termination of this process synchronously with the end of spreading in the Labrador Sea must otherwise be seen as unrelated coincidence.

Compressional folding of the Wandel Sea Basin sediments of eastern Peary Land on ESE axes in post-Paleocene time can be associated with oblique compression of North-East Greenland against the Barents Shelf continental margin prior to anomaly 13, as required by most spreading models. These folds are presumably counterparts of the better dated (Eocene – early Oligocene) main phase structures of the West Spitzbergen orogen (Harland & Horsfield 1974). Deformation was more intense in Spitzbergen, with easterly directed thrusting, and has been similarly interpreted as the effect of dextral transpression across the Spitzbergen fracture zone (Harland 1965, 1969, Lowell 1972). One might predict the intensification of the Peary Land deformation offshore towards the Wandel Sea shelf edge, with SW-directed thrusting.

Post-anomaly 13

After spreading had ceased in the Labrador Sea, North-East Greenland ceased to be pinned against the Barents Shelf and oblique extension commenced along the Spitzbergen fracture zone, as evidenced by the overlaps of a few tens of kilometres shown in this region on recent anomaly 13 reconstructions (Talwani & Eldholm 1977, Phillips et al. 1978). Onshore structures are compatible with limited extension, rather than a hundred or more kilometres demanded by earlier models. The principal extensional structure in Svalbard is the Forlandsundet graben, whose sediment fill Harland (1969) considered to post-date the West Spitzbergen orogeny. We associate block faulting of the Wandel Sea succession with this phase; of the two alternative structural interpretations proposed by Håkansson (1979) reactivation of earlier structures by compression followed by extension is the preferred model. Again, more intense extensional faulting might be expected offshore.

In northernmost Greenland, major uplift took place in post-anomaly 13 time, with significant displacement on E–W fractures such as the Harder Fjord fault zone. Feden et al. (1979) have discussed the elevation of this region as a response to the development of the Yermak hot spot. Isostatic response to crustal thickening due to the Kap Cannon compressive phase is also a possibility. Current resurgence of the hot spot suggested by Feden et al. (1979) is perhaps evidenced by the solfataric activity (see Troelsen 1949) along E–W fault lines in the region.

Conclusions and the Nares Strait problem

The presence on the northern coast of Greenland in Peary Land of late Cretaceous intrusive and extrusive rocks cannot be used specifically (as can other geological features described in this symposium) to infer the magnitude of possible offset along Nares Strait. They do, however, provide some restraint on reconstructed relative positions of northern Greenland and Ellesmere Island at the time of their genesis. In many pre-drift reconstructions involving substantial strike-slip motion along Nares Strait, Peary Land is placed adjacent to north-eastern Ellesmere Island, a region devoid of comparable magmatic rocks (e.g. Feden et al. 1979). While an argument based on the absence of rocks can never be conclusive, the geotectonic interpretations of this paper suggest that the lack of magmatic products in this part of Ellesmere Island is not coincidental.

The Greenland alkaline magmatism was associated with continental rifting that pre-dated the opening of the Eurasia Basin. Of particular interest is the conspicuous coast-normal basic dyke swarm that intensifies northwards towards the outer coast and that is regarded as the southernmost effect of the suturing that separated the Lomonosov Ridge from the Barents Shelf. Dykes occur in Greenland as far west as about long. 48° and

thus in reconstructions of the type mentioned above, it is surprising that no corresponding magmatic products are known from north-eastern Ellesmere Island. We suggest that the late Cretaceous magmatism of the region, manifested onshore in north coastal Peary Land, occurred in a specific geological location far removed from north-eastern Ellesmere Island.

This paper shows that the magmatic and tectonic history of North Greenland in Cretaceous and Tertiary time relates well to the timing and pattern of plate displacements when viewed in the context of the Hailwood et al. (1979) polarity time scale. Broadly, three tectonic periods can be recognised on Greenland's northern margin: a late Cretaceous phase of extensional tectonism associated with continental rift magmatism, a Paleocene–Eocene phase of compression and northward thrusting, and a later Tertiary period of vertical displacements in a mildly extensional tectonic environment. The Palaeogene compressional phase is seen as a consequence of anti-clockwise rotation of Greenland due to the opening of the Labrador Sea. We have shown, by two independent lines of evidence, that this compressive phase commenced at about the Cretaceous–Tertiary boundary, suggesting that major spreading in the Labrador Sea also commenced then, some 10 Ma later than previously supposed. At that time Greenland was still part of the Eurasia plate and the compressional buildup was not manifested in thrust displacement of the northern margin against the Lomonosov continental fragment until the north-eastern (Wandel Sea) margin began to become 'unpinned' from the Barents Shelf in Eocene time.

Recognition of a coherent relationship between events on the continental margins of Greenland and spreading patterns does not in itself go far towards 'solving' the Nares Strait conflict. But in combination with the constraints on the magnitude of displacement along Nares Strait recognised by Christie et al. (1981) it does provide an essential framework within which the problem should be viewed. The geological evidence shows that Greenland did undergo significant anti-clockwise rotation with respect to the North American plate in early Tertiary time, and this precludes an origin for the Labrador Sea and Baffin Bay in which sea-floor spreading plays no part. The problem is to define the extent of spreading, and how the resulting displacements have been accommodated. Clearly, within-plate strains are important, not only in this context but generally: the failure of rigid-plate theory to produce mutually compatible reconstructions of the North-East Atlantic, Labrador Sea and Eurasia Basin illustrates the point.

Geologists would welcome a magnetic anomaly superposition study of these three oceans directed towards producing a series of finite difference displacements which not only satisfy the constraints provided by the geological evidence, but serve specifically to quantify the within-plate deformations at each stage. Some recent studies have made a start in this direction, but better integration of the onshore geological evidence with the interpretation of offshore geophysical data is required.

Acknowledgements

Two of us, N. J. S. and P. R. D., thank John D. C. Peacock and Bruce K. Reid, leaders of the Joint Services Expedition to Peary Land for support and coordination of the geological work into the 1969 expedition programme. P. R. D. thanks C. K. Brooks and J. C. Bailey for discussions on the chemical composition of the Kap Washington Group volcanics. C. K. Brooks, G. L. Johnson, L. M. Larsen and H. P. Trettin critically read the manuscript and offered helpful suggestions for which we are grateful. The paper is published with the permission of the Director of the Geological Survey of Greenland.

References

Athavale, R. N. & Sharma, P. V. 1975. Paleomagnetic results on early Tertiary lava flows from West Greenland and their bearing on the evolution history of the Baffin Bay-Labrador Sea region. – Can. J. Earth Sci. 12: 1–18.

Balkwill, H. R. 1978. Evolution of Sverdrup Basin, Arctic Canada. – Bull. Am. Ass. Petrol. Geol. 62: 1004–1028.

Barberi, F., Tazieff, H. & Varet, J. 1972. Volcanism in the Afar depression: its tectonic and magmatic significance. – Tectonophysics 15: 19–29.

Barberi, F., Ferrara, G., Santacroce, R., Treuil, M. & Varet, J. 1975. A transitional basalt-pantellerite sequence of fractional crystallization, the Boina Centre (Afar Rift, Ethiopia). – J. Petrol. 16: 22–56.

Batten, D. J., Brown, P. E., Dawes P. R., Higgins, A. K., Koch, B. E., Parsons, I. & Soper, N. J. 1981. Peralkaline volcanicity on the Eurasia Basin margin. – Nature, Lond. 294: 150–152.

Birkelund, T. & Perch-Nielsen, K. 1976. Late Palaeozoic–Mesozoic evolution of central East Greenland. – In: Escher, A. & Watt, W. S. (eds.), Geology of Greenland: 304–339. – Geol. Surv. Greenland, Copenhagen.

Bott, M. H. P. & Watts, A. B. 1971. Deep structure of the continental margin adjacent to the British Isles. – Rep. Inst. Geol. Sci. 70/14: 89–109.

Brooks, C. K. & Gleadow, A. J. W. 1977. A fission-track age for the Skaergaard intrusion and the age of the East Greenland basalts. – Geology 5: 539–540.

Brooks, C. K. & Rucklidge, J. C. 1976. Tertiary peralkaline rhyolite dikes from the Skaergaard area, Kangerdlugssuaq, East Greenland. – Meddr. Grønland 197(3): 27 pp.

Brown, P. E. & Parsons, I. 1981. The Kap Washington Group volcanics. – Rapp. Grønlands geol. Unders. 106: 65–68.

Bullard, E., Everett, J. E. & Smith, A. G. 1965. The fit of the continents around the Atlantic. – In: Blackett, P. M. S., Bullard, E. & Runcorn, S. K., A symposium on continental drift. – Phil. Trans. Roy. Soc. Lond. 258A: 41–51.

Burov, J. P., Krasilscikov, A. A., Firsov, L. V. & Klubov, B. A. 1976. The age of Spitsbergen dolerites. – Norsk Polarinst. Årbok 1975: 101–108.

Christie, R. L., Dawes, P. R., Frisch, T. [O.], Higgins, A. K., Hurst, J. M., Kerr, J. W. & Peel, J. S. 1981. Geological evidence against major displacement in the Nares Strait. – Nature, Lond. 291: 478–480.

Clarke, D. B. & Upton, B. G. J. 1971. Tertiary basalts of Baffin Bay: field relations and tectonic setting. – Can. J. Earth Sci. 8: 248–258.

Croxton, C. A., Dawes, P. R., Soper, N. J. & Thomsen, E. 1980. An occurrence of Tertiary shales from the Harder

Fjord Fault, North Greenland fold belt, Peary Land. – Rapp. Grønlands geol. Unders. 101: 61–64.

Dawes, P. R. 1973. The North Greenland fold belt: a clue to the history of the Arctic Ocean basin and Nares Strait lineament. – In: Tarling, D. H. & Runcorn, S. K. (eds), Implications of continental drift to the earth sciences 2: 925–947. – Academic Press, London & New York.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R. & Kerr, J. W. 1982. The case against major displacement along Nares Strait. – This volume.

Dawes, P. R. & Peel, J. S. 1981. The northern margin of Greenland from Baffin Bay to the Greenland Sea. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 201–264. Plenum Press, New York & London.

Dawes, P. R. & Soper, N. J. 1970. Geological investigations in northern Peary Land. – Rapp. Grønlands geol. Unders. 28: 9–15.

Dawes, P. R. & Soper, N. J. 1971. Significance of K/Ar age determinations from northern Peary Land. – Rapp. Grønlands geol. Unders. 35: 60–62.

Dawes, P. R. & Soper, N. J. 1973. Pre-Quaternary history of North Greenland. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 117–134.

Dawes, P. R. & Soper, N. J. 1979. Structural and stratigraphical framework of the North Greenland fold belt in Johannes V. Jensen Land, Peary Land. – Rapp. Grønlands geol. Unders. 93: 40 pp.

Dibner, V. D. 1957. The geological structure of Franz Josef Land. – In: Markov, F. G. & Nalivkin, D. V. (eds), Geologiia Sovetskoi Arktiki 81: 11–20. – Inst. Geologii Arktiki Trudy (in Russian).

Dypvik, H. & Nagy, J. 1978. Early Tertiary bentonites from Svalbard; a preliminary report. – Polarforschung 48: 139–150.

Ellitsgaard-Rasmussen, K. 1955. Features of the geology of the folding range of Peary Land North Greenland. – Meddr Grønland 127(7): 56 pp.

Faller, A. M. 1975. Palaeomagnetism of the oldest Tertiary basalts in the Kangerdlugssuaq area of East Greenland. – Bull. geol. Soc. Denmark 24: 173–178.

Featherstone, P. S., Bott, M. H. P. & Peacock, J. H. 1977. Structure of the continental margin of south-eastern Greenland. – Geophys. J. Roy. astr. Soc. 48: 15–27.

Feden, R. H., Vogt, P. R. & Fleming, H. S. 1979. Magnetic and bathymetric evidence for the "Yermak hot spot" northwest of Svalbard in the Arctic Basin. – Earth planet. Sci. Lett. 44: 18–38.

Floyd, P. A. & Winchester, J. A. 1975. Magma type and tectonic setting discrimination using immobile elements. – Earth planet. Sci. Lett. 27: 211–218.

Gass, I. G. 1970. The evolution of volcanism in the junction area of the Red Sea, Gulf of Aden and Ethiopian rifts. – Phil. Trans. Roy. Soc. Lond. 267A: 369–381.

Gayer, R. A., Gee, D. G., Harland, W. B., Miller, J. A., Spall, H. R., Wallis, R. H. & Winsnes, T. S. 1966. Radiometric age determinations on rocks from Spitsbergen. – Norsk Polarinst. Skr. 137: 39 pp.

Grant, A. C. 1980. Problems with plate tectonics: the Labrador Sea. – Bull. Can. Petrol. Geol. 28: 252–278.

Grant, A. C. 1982. Problems with plate tectonic models for Baffin Bay – Nares Strait: evidence from the Labrador Sea. – This volume.

Gusev, B. V. 1979. Comparative characteristics of the magnetism of plateau basalts in Siberia, Taymir, Franz Josef Land and other regions of the Earth. – U.S. Dept. Air Force, WP–AFB, Ohio, Translation nr. FTD–1D (RS) T–1265–79: 20 pp. (From Russian).

Hailwood, E. A., Bock, W., Costa, L. [I]., Dupeuble, P. A., Müller, C. & Schnitker, D. 1979. Chronology and biostratigraphy of Northeast Atlantic sediments, DSDP Leg 48. – Initial Rep. Deep Sea Drilling Proj. 48: 1119–1141. – U. S. Gov. Print. Off., Washington, D. C.

Haller, J. 1970. Tectonic map of East Greenland (1: 500,000). An account of tectonism, plutonism, and volcanism in East Greenland. – Meddr Grønland 171(5): 286 pp.

Harland, W. B. 1965. The tectonic evolution of the Arctic–North Atlantic region. – In: Blackett, P. M. S., Bullard, E. & Runcorn, S. K., A symposium on continental drift. – Phil. Trans. Roy. Soc. Lond. 258A: 59–75.

Harland, W. B. 1969. Contribution of Spitsbergen to understanding of tectonic evolution of North Atlantic region. – In: Kay, M. (ed.), North Atlantic – geology and continental drift, a symposium. – Mem. Am. Ass. Petrol. Geol. 12: 817–851.

Harland, W. B. & Horsfield, W. T. 1974. West Spitsbergen orogen. – In: Spencer, A. M. (ed.), Mesozoic-Cenozoic orogenic belts: data for orogenic studies. – Spec. Publ. geol. Soc. Lond. 4: 747–755.

Heirtzler, J. R., Dickson, G. O., Herron, E. M., Pitman, W. C. & Le Pichon, X. 1968. Marine magnetic anomalies, geomagnetic field reversals, and motions of the ocean floor and continents. – J. geophys. Res. 73: 2119–2136.

Higgins, A. C. & Soper, N. J. 1981. Cretaceous–Palaeogene sub-basaltic and intrabasaltic sediments of the Kangerdlugssuaq area, central East Greenland. – Geol. Mag. 118: 337–354.

Higgins, A. K., Friderichsen, J. D. & Soper, N. J. 1981. The North Greenland fold belt between central Johannes V. Jensen Land and eastern Nansen Land. – Rapp. Grønlands geol. Unders. 106: 35–45.

Hinz, K., Schlüter, H.-U., Grant, A. C., Srivastava, S. P., Umpleby, D. & Woodside, J. 1979. Geophysical transects of the Labrador Sea: Labrador to Southwest Greenland. – Tectonophysics 59: 151–183.

Håkansson, E. 1979. Carboniferous to Tertiary development of the Wandel Sea Basin, eastern North Greenland. – Rapp. Grønlands geol. Unders. 88: 73–84.

Håkansson, E., Heinberg, C. & Stemmerik, L. 1981. The Wandel Sea Basin from Holm Land to Lockwood Ø, eastern North Greenland. – Rapp. Grønlands geol. Unders. 106: 47–63.

Irvine, T. N. & Baragar, W. R. A. 1971. A guide to the chemical classification of common volcanic rocks. – Can. J. Earth Sci. 8: 523–548.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor spreading in central Baffin Bay. – Can. J. Earth Sci. 16: 2122–2135.

Jepsen, H. F. 1969. Basiske intrusiver i det sydlige Pearyland og deres geologiske milieu. – Unpubl. thesis. Univ. Aarhus, Denmark: 66 pp.

Johnson, G. L., Monahan, D., Grønlie, G. & Sobczak, L. [W.] 1979. General bathymetric chart of the oceans (GEBCO), 1:6 000 000, Sheet 5.17, The Arctic Ocean. – Can. Hydrogr. Serv., Ottawa.

Johnson, G. L., Sundvor, E. & Myhre, A. M. (in press). The Yermak plateau. – Norsk Polarinst. Skr.

Jürgensen, T. & Mikkelsen, N. 1974. Coccoliths from volcanic sediments (Danian) in Nûgssuaq, West Greenland. – Bull. geol. Soc. Denmark 23: 225–230.

Keen, C. E. & Barrett, D. L. 1972. Seismic reflection studies in Baffin Bay: an example of a developing ocean basin. – Geophys. J. Roy. astr. Soc. 30: 253–271.

Keen, C. E., Keen, M. J., Ross, D. I. & Lack, M. 1974. Baffin Bay: small ocean basin formed by sea-floor spreading. – Bull. Am. Ass. Petrol. Geol. 58: 1089–1108.

Kerr, J. W. 1967. A submerged continental remnant beneath the Labrador Sea. – Earth planet. Sci. Lett. 2: 283–289.

Kerr, J. W. 1980. A plate tectonic contest in Arctic Canada. – In: Strangway, D. W. (ed.), The continental crust and its

mineral deposits. – Spec. Pap. geol. Ass. Can. 20: 457–486.

Kovacs, L. C. 1982. Motion along Nares Strait recorded in the Lincoln Sea: aeromagnetic evidence. – This volume.

Kristoffersen, Y. & Talwani, M. 1977. Extinct triple junction south of Greenland and the Tertiary motion of Greenland relative to North America. – Bull. geol. Soc. Am. 88: 1037–1049.

Larsen, O., Dawes, P. R. & Soper, N. J. 1978. Rb/Sr age of the Kap Washington Group, Peary Land, North Greenland, and its geotectonic implication. – Rapp. Grønlands geol. Unders. 90: 115–119.

Laughton, A. S. 1971. South Labrador Sea and the evolution of the North Atlantic. – Nature, Lond. 232: 612–617.

Le Pichon, X., Sibuet, J-C. & Francheteau, J. 1977. The fit of the continents around the North Atlantic Ocean. – Tectonophysics 38: 169–209.

Lowell, J. D. 1972. Spitsbergen Tertiary orogenic belt and the Spitsbergen fracture zone. – Bull. geol. Soc. Am. 83: 3091–3102.

Macdonald, G. A. & Kutsura, T. 1964. Chemical composition of Hawaiian lavas. – J. Petrol. 5: 82–133.

Parrott, R. J. E. & Reynolds, P. H. 1975. Argon-40/Argon-39 geochronology: age determinations of basalts from the Labrador Sea area. – Geol. Soc. Am. Abs. with Prog. 7: 835 only.

Parsons, I. 1981. Volcanic centres between Frigg Fjord and Midtkap, eastern North Greenland. – Rapp. Grønlands geol. Unders. 106: 69–75.

Pearce, J. A. & Cann, J. R. 1973. Tectonic setting of basic volcanic rocks determined using trace element analyses. – Earth planet. Sci. Lett. 19: 290–300.

Phillips, J. D., Feden, R., Fleming, H. S. & Tapscott, C. 1978. Aeromagnetic studies of the Greenland-Norwegian Sea and Arctic Ocean. – Unpubl. manuscript: 64 pp.

Pitman, W. C. & Talwani, M. 1972. Sea-floor spreading in the North Atlantic. – Bull. geol. Soc. Am. 83: 619–646.

Prestvik, T. 1978. Cenozoic plateau lavas of Spitsbergen – a geochemical study. – Norsk Polarinst. Årbok 1977: 129–143.

Soper, N. J., Downie, C., Higgins, A. C. & Costa, L. I. 1976. Biostratigraphic ages of Tertiary basalts on the east Greenland continental margin and their relation to plate separation in the Northeast Atlantic. – Earth planet. Sci. Lett. 32: 149–157.

Soper, N. J., Higgins, A. K. & Friderichsen, J. D. 1980. The North Greenland fold belt in eastern Johannes V. Jensen Land. – Rapp. Grønlands geol. Unders. 99: 89–98.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Surlyk, F. 1977. Stratigraphy, tectonics and palaeogeography of the Jurassic sediments of the areas north of Kong Oscars Fjord, East Greenland. – Bull. Grønlands geol. Unders. 123: 56 pp.

Surlyk, F., Hurst, J. M. & Bjerreskov, M. 1980. First age-diagnostic fossils from the central part of the North Greenland foldbelt. – Nature, Lond. 286: 800–803.

Sweeney, J. F. 1978. Arctic geophysical review. – Publ. Earth Physics Branch 45 (4): 108 pp.

Talwani, M. & Eldholm, O. 1977. Evolution of the Norwegian-Greenland Sea. – Bull. geol. Soc. Am. 88: 969–999.

Taylor, P. T., Kovacs, L. C., Vogt, P. R. & Johnson, G. L. 1981. Detailed aeromagnetic investigation of the Arctic Basin, 2. – J. geophys. Res. 86: 6323–6333.

Thorsteinsson, R. & Tozer, E. T. 1970. The geology of the Arctic Archipelago. – In: Douglas, R. J. W. (ed.), Geology and economic minerals of Canada. – Econ. Geol. Rep. geol. Surv. Can. 1: 547–590.

Trettin, H. P. & Balkwill, H. R. 1979. Contributions to the tectonic history of the Innuitian Province, Arctic Canada. – Can. J. Earth Sci. 16: 748–769.

Trettin, H. P., Frisch, T. O., Sobczak, L. W., Weber, J. R., Niblett, E. R., Law, L. K., DeLaurier, J. M. & Whitlam, K. 1972. The Innuitian Province. – In: Price, R. A. & Douglas, R. J. W. (eds.), Variations in tectonic styles in Canada. – Spec. Pap. geol. Ass. Can. 11: 83–179.

Troelsen, J. C. 1949. Contributions to the geology of the area round Jørgen Brønlunds Fjord, Peary Land, North Greenland. – Meddr. Grønland 149(2): 29 pp.

Troelsen, J. [C]. 1950. Geology. – In: Winther, P. C. et al., A preliminary account of the Danish Pearyland Expedition, 1948–9. – Arctic 3: 6–8.

Tyrell, G. W. & Sandford, K. S. 1933. Geology and petrology of the dolerites of Spitsbergen. – Proc. Roy. Soc. Edin. 53: 284–321.

van der Linden, W. J. M. 1975. Crustal attenuation and sea-floor spreading in the Labrador Sea. – Earth planet. Sci. Lett. 27: 409–423.

Vogt, P. R. & Avery, O. E. 1974. Detailed magnetic surveys in the northeast Atlantic and Labrador Sea. – J. geophys. Res. 79: 363–389.

Vogt, P. R., Avery, O. E., Morgan, W. J., Johnson, G. L., Schneider, E. D. & Higgs, R. H. 1969. Morphology, magnetic anomalies and evolution of the Northeast Atlantic and Labrador Sea, Part III – Evolution. – Trans. Am. geophys. Un. 50: 189 only.

Vogt, P. R., Ostenso, N. A. & Johnson, G. L. 1970. Magnetic and bathymetric data bearing on sea-floor spreading north of Iceland. – J. geophys. Res. 75: 903–920.

Vogt, P. R., Taylor, P. T., Kovacs, L. C. & Johnson, G. L. 1979. Detailed aeromagnetic investigation of the Arctic Basin. – J. geophys. Res. 84: 1071–1089.

Wagner, R. H., Soper, N. J. & Higgins, A. K. in press. A late Permian flora of Pechora affinity in North Greenland. – Rapp. Grønlands geol. Unders. 108.

Watt, W. S. 1969. The coast-parallel dike swarm of southwest Greenland in relation to the opening of the Labrador Sea. – Can. J. Earth Sci. 6: 1320–1321.

Fragmentation of the Canadian Arctic Archipelago, Greenland, and surrounding oceans

LAWRENCE W. SOBCZAK

Sobczak, L. W. 1982. Fragmentation of the Canadian Arctic Archipelago, Greenland, and surrounding oceans. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 221–236.

Fragmentation of the Canadian Arctic Archipelago, Greenland and surrounding ocean basins is indicated mainly by a pattern of northeast-trending fractures and northwest-trending arches and rifts. Based on a comparison of this structural pattern with experimentally produced fracture patterns, a new evolutionary concept for the Cenozoic era is proposed for the region. The Eurasia Basin and North Atlantic sea-floors developed as rift-generated tensional zones separated by a later developed dextral transform shear zone, here named the Nansen Shear Zone. Sub-parallel to and some 2000 km southwest of this shear couple is another less well-developed system comprised of M'Clure Strait and Baffin Bay which were initiated by rifting and later connected by a dextral transform shear zone, here named the Parry Channel Shear Zone. The landmass between these two transform couples (Greenland and the Queen Elizabeth Islands) probably was and is acted upon by a sinistral force couple as a result of spreading between these couples and is probably partially responsible for generating internal deviatoric tensional and compressional stresses noted from earthquake data. Fractures and arch axes within the Queen Elizabeth Islands have similar directions to respective features within the North Atlantic sea-floor; stresses may have similar orientations in both regions.

Nares Strait may be a poorly-developed sinistral transform shear zone that joins rift-generated ocean basins: the Eurasia Basin to the north and Baffin Bay to the south.

L. W. Sobczak, Department of Energy, Mines and Resources, Earth Physics Branch, Gravity and Geodynamics Division, 1 Observatory Crescent, Ottawa, Ontario, Canada, K1A OY3.

Contribution from the Earth Physics Branch No. 943.

A fragmentation model for the region of the Canadian Arctic Archipelago, Greenland, and adjacent ocean areas is proposed on the basis of analogy with experimental results from Brown (1928) and Riedel (1929) summarized by De Sitter (1956); this fragmentation has occurred during and since the time of the Eurekan Deformation in the Cenozoic. In particular the azimuths of arches, fractures and rifts are utilized to interpret palaeostress fields that in turn may provide constraints on the evolution of the region. As well, the variation in thicknesses of the crystalline crust below sedimentary basins such as the Sverdrup Basin is considered to indicate possible stress directions and behaviour of crustal development. Theoretical calculation of the stress field from earthquake data for the Beaufort Sea region and Byam Martin Channel area in the Queen Elizabeth Islands (Hasegawa 1977, Hasegawa et al. 1979) also provides additional support for the model.

The fragmentation concept, here applied to the last 60 Ma or so, advocates that a northeastern piece of the North American plate (about 2000 km wide) split along two boundaries — one in the North Atlantic Ocean – eastern Arctic Ocean and the other in Baffin Bay – Parry Channel – M'Clure Strait (Fig. 1). These boundaries are considered to consist of early developed tensional rift zones at either end of a later developed dextral transform shear zone as described in this paper. The landmass between these boundaries (Queen Elizabeth Islands and Greenland) was fractured primarily by northeast-trending faults, grabens and troughs and was affected by northwest-trending uplifts, ridges, arches and folds.

Locations of fractures, arches and rifts are shown on a polar stereographic projection for a portion of the Arctic region in order to compare general trends and similarities. Azimuths of these features are taken from an arbitrary rectangular coordinate system superimposed on the polar stereographic projection. This paper does not utilize spherical geometry. It is considered unnecessary because the error in azimuths is about one degree using the rectangular coordinate system laid over the polar stereographic projection. Thus this method does not jeopardize the azimuths and arguments used throughout this paper.

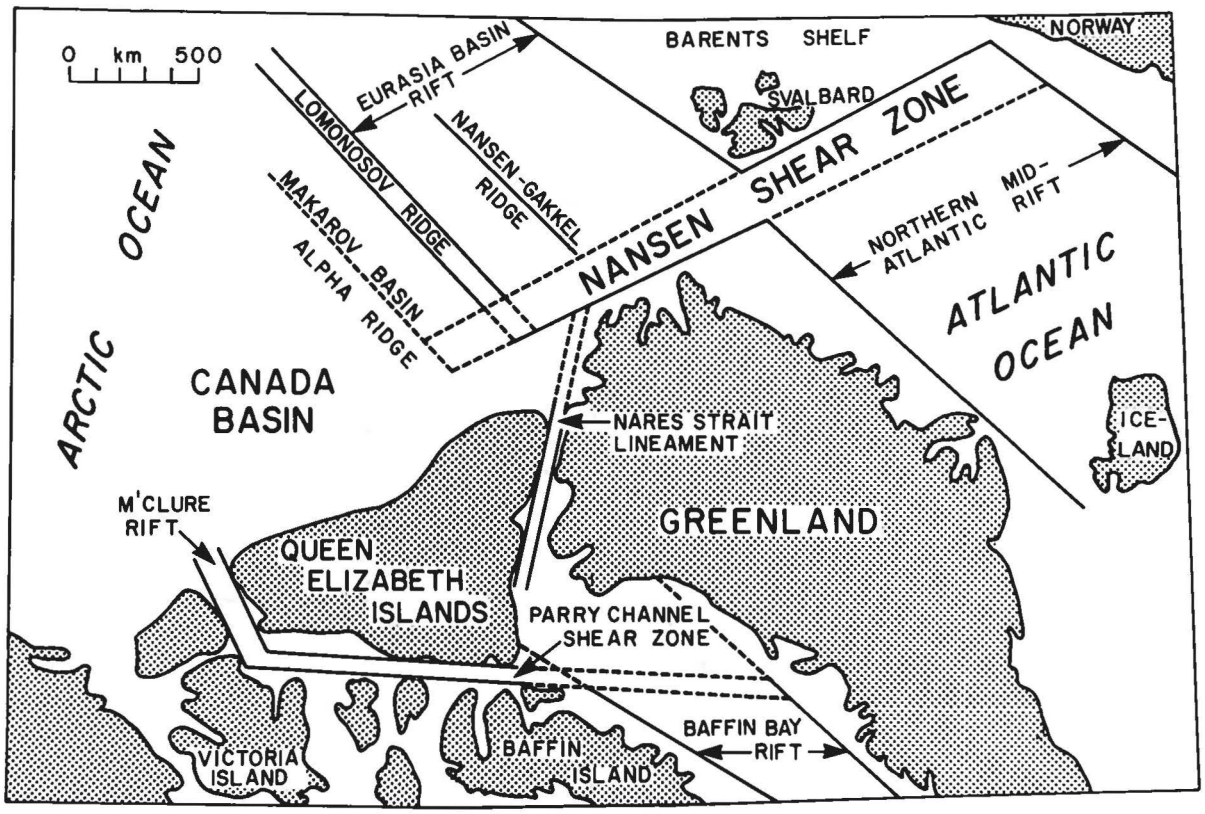


Fig. 1. Location map of major rift and shear zones of the Arctic region. Eurasia Basin Rift – Nansen Shear Zone – Northern mid-Atlantic Rift form one well-developed dextral transform couple, and M'Clure Rift – Parry Channel Shear Zone – Baffin Bay Rift form a poorly-developed dextral transform couple. The Nares Strait Lineament joins these two transform couples in two triple junctions, perhaps as a sinistral transform couple.

Geological and tectonic setting

The study region (Fig. 1) includes the Canadian Arctic Archipelago, Greenland and the neighbouring parts of the Atlantic and Arctic Oceans and Baffin Bay. The archipelago (Figs 2 and 3, Kerr 1981) formed from several major depositional subsiding basins laid in an overlapping fashion more or less one above the other on a Precambrian continental crystalline basement; the basins are progressively younger toward the north. Before and during the major period of deposition of the Franklinian Geosyncline and the Sverdrup Basin two prominent structural highs persisted. The Boothia Uplift has a north–south strike normal to the axis of the future Sverdrup Basin and a northward plunge. The Pearya Geanticline has a northeast strike, sub-parallel to the axis and was a magmatically quiet zone during the formation of the Sverdrup Basin. The Cornwallis Fold Belt (Kerr 1977), which developed prior to and during the Ellesmerian Orogeny in four pulses of uplift that together constitute the Cornwallis Disturbance, lies sub-parallel to and over the northern end of the Boothia Uplift. The Ellesmerian Orogeny also produced structural features normal (Rens Fiord Uplift, Fig. 2) and sub-parallel (Minto Arch, Parry Islands Fold Belt, Ellesmere–Greenland Fold Belt, and the northern Ellesmere Fold Belt, Fig. 2) to the future axis of the Sverdrup Basin.

The Sverdrup Basin developed in various phases after the Ellesmerian Orogeny as a subsiding basin (Sweeney 1977) commencing in Early Mississippian time (Balkwill 1978); basin development ended with the Eurekan Orogeny in the mid-Tertiary. Sweeney (op. cit.) proposed that initial subsidence was partly due to lithospheric cooling but later subsidence episodes were initiated by extensional forces operating in the Triassic and in late Early Cretaceous time. These forces resulted in the downdropping of graben blocks beneath the basin. Balkwill (1978) suggested that crustal foundering accompanied by fracturing and tapping of upper mantle fluids prompted primary subsidence and enlargement of basin space. Basalt flows and mafic dykes and sills are systematically intercalated with and transect all but the Cenozoic sediments of the Sverdrup Basin. Many of these dykes are coextensive with faults or are sub-parallel to local faults, folds, uplifts and arches, de-

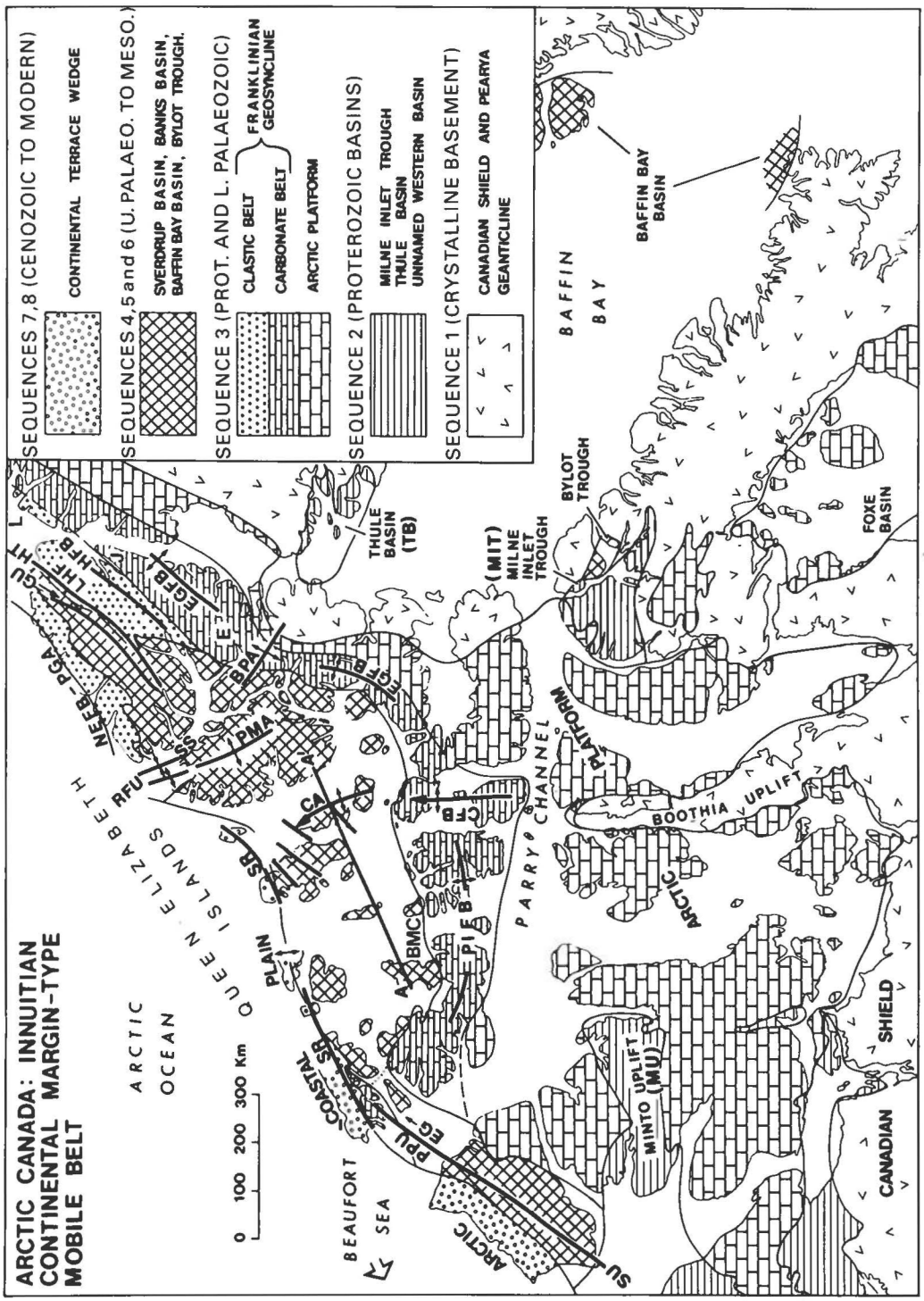
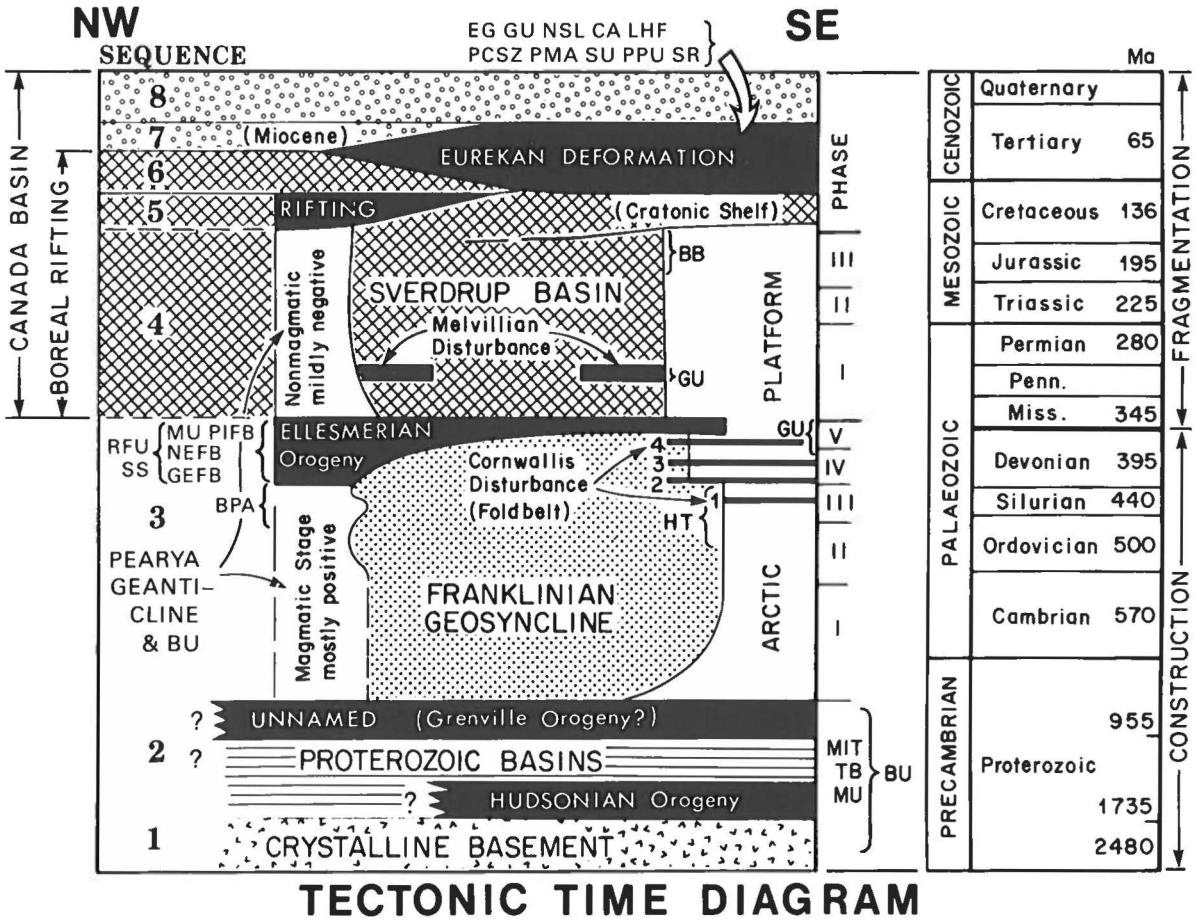


Fig. 2. Structures and structural provinces of the Canadian Arctic Archipelago (after Kerr 1981). A–A' is location of profile shown in Fig. 4. BMC = Byam Martin Channel, BPA = Bache Peninsula Arch, CA = Cornwall Arch, CFB = Cornwallis Fold Belt, E = Ellesmere Island, EG = Eglinton Graben, EGFB = Ellesmere–Greenland Fold Belt, GU = Grantland Uplift, HFB = Hazen Fold Belt, HT = Hazen Trough, J = Judge Daly Promontory, L = Lincoln Sea, LHF = Lake Hazen Fault, NEFB–PGA = Northern Ellesmere Fold Belt–Pearya Geanticline, PIFB = Parry Islands Fold Belt, PMA = Princess Margaret Arch, PPU = Prince Patrick Uplift, RFU = Rens Fiord Uplift, SR = Sverdrup Rim, SS = Svartevaeg Syncline, SU = Storkerson Uplift.

veloped during the Eurekan Deformation (Figs 2 and 3) (Balkwill & Bustin 1980) and in places these overlie similar tectonic features produced by the older Ellesmerian Orogeny.

Noteworthy features produced during the development of the Queen Elizabeth Islands include variations in crustal thickness and an inverse relationship between thicknesses of sediments and crystalline crustal rocks.



TECTONIC TIME DIAGRAM

Fig. 3. Tectonic time diagram for the Canadian Arctic Islands (after Kerr 1980a) which resembles a cross-section. The stippled, cross-hatched and circled ornaments show regions where there was continuous or nearly continuous sedimentation. The solid black pattern indicates tectonic events with widespread unconformities. See Figs 2 and 7 for explanation of abbreviations.

The latter feature is a product of isostatic compensation. Oceanward crustal thinning below passive continental margin basins has also been advocated (Sobczak & Weber 1973, Sobczak 1975a, b) with the mass deficiency of the water and sediments being isostatically compensated by a mantle anti-root according to the Airy hypothesis. Analysis of gravity and seismic data for the Sverdrup Basin area (Sobczak & Overton in prep.) suggests that the sediments and water are compensated regionally by a mantle anti-root, and that the crystalline crust thins below the thickest sedimentary section and thickens towards the thinner sections in the axial region of the basin (Fig. 4). The seismic results also suggest that folds (long wavelength undulations) are prominent at the crust–mantle boundary along the axis of the basin. These features can be explained by a crust which was stretched by tensional forces perpendicular to the continental margin and folded, probably simultaneously, by compressional forces along the axis of the basin. Tensional forces caused necking of the crystalline crust whereby the top sagged (tectonic subsidence) and became a repository for sediments (loading subsidence) and the bottom arched upward to be replaced by mantle material in accordance with isostasy.

During and after the Eurekan Deformation, fragmentation was active in the oceanic areas of the North Atlantic Ocean and Baffin Bay. The evolution of the North Atlantic Ocean has been described by Grønlie et al. (1979) who have summarized the work of several authors including, in particular, that of Talwani & Eldholm (1977). The principal structural elements involved are numbered in Fig. 5 and listed with Fig. 7. Sea-floor spreading along the Mohns Ridge (#16, #13) has been more complex. In the latter area at least two spreading centres, now extinct, are believed to have been active before the establishment, about 10 Ma ago, of the presently spreading Iceland – Jan Mayen Ridge (# 18). One of the extinct axes (#14) located in the Norway Basin was active between 55 and 26 Ma ago after which period a new spreading axis developed to the west be-

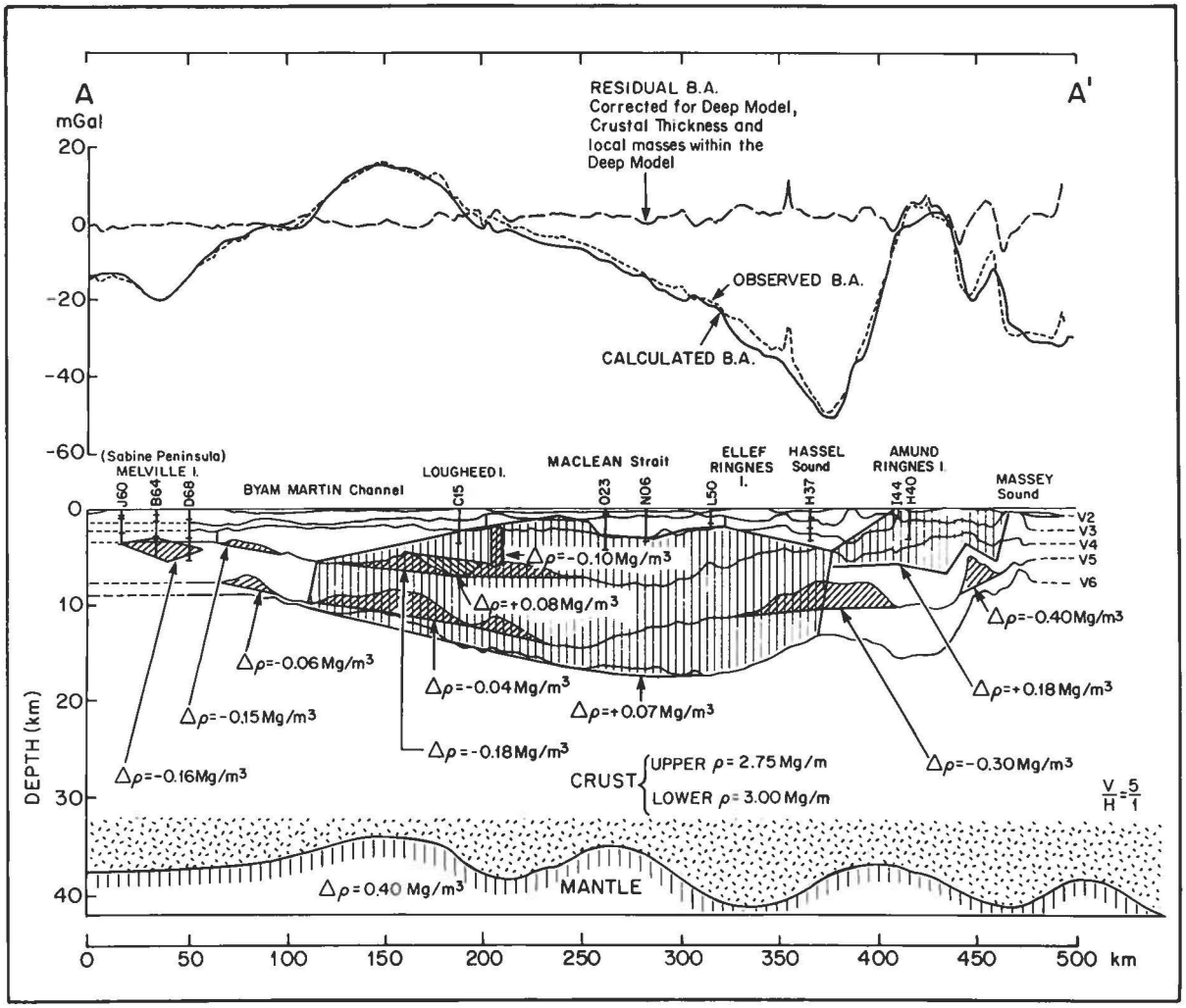


Fig. 4. Profile A–A′ shows observed modified Bouguer anomaly, calculated anomaly and corrected anomaly for all masses combined: Sobczak & Overton's (in prep.) deep model for Sverdrup Basin (layers 1 to 3) and Franklinian Geosyncline (layers 4 and 5), Forsyth et al.'s (1979) crust–mantle model, and local masses indicated by seismic and gravity data. The sedimentary thickness varies from 9 to 17.4 km, the crystalline continental thickness varies from 18 to 33 km and the total crustal thickness varies from 34 to 42 km (Sobczak & Overton in prep.). For location of profile see Fig. 2.

tween 23 and 17 Ma ago, forming the Jan Mayen Ridge (#17) which is considered to include a piece of Greenland (Grønlie et al. 1979). North of the Greenland Fracture Zone sea-floor spreading along the Knipovich Ridge (#9) has been active since about 36 Ma ago. No major opening existed before that time and motion between Greenland and Svalbard was primarily one of shear along the Greenland (#8) and Senja (#11) Fracture Zones.

For the Baffin Bay area, Jackson et al. (1977, 1979) have presented geophysical evidence which supports the existence of oceanic crust formed by sea-floor spreading in the central part of the bay. Preliminary magnetic evidence suggests that sea-floor spreading occurred between approximately 38 to 60 Ma ago along an extinct axis (#22) along the centre of the bay that is probably related to a negative gravity anomaly.

Since the end of the Eurekan Deformation (about 20 Ma ago) the Queen Elizabeth Islands area has been relatively stable (Kerr 1981). Recent crustal activity is indicated by earthquakes recorded during the period 1908 to 1975 (Basham et al. 1977, Wetmiller & Forsyth 1978). The epicentres appear to be concentrated over the northwestern continental margin where thick sedimentary basins are developing (Sobczak 1975a, b) over older structures such as the Sverdrup Rim (#48) and Boothia Uplift (#26) and over basin areas such as Baffin Bay and Sverdrup Basin (Byam Martin Channel area). For the Byam Martin Channel area Forsyth et al. (1979) have described various geophysical features

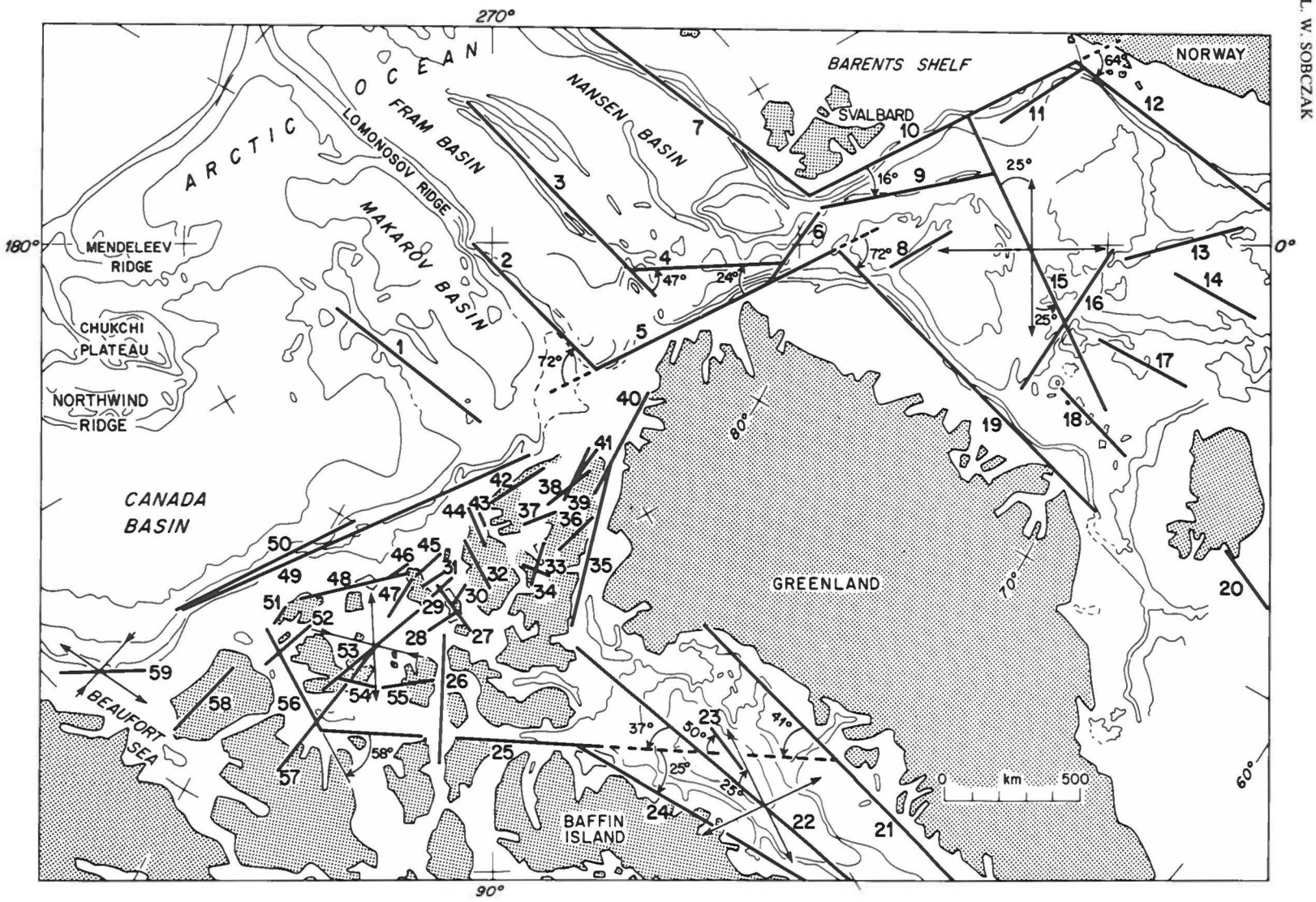


Fig. 5. Location of features on the northeastern part of the North American plate and surrounding oceans. In the text and Fig. 7 the azimuths of numbered structural features are taken relative to a local rectangular co-ordinate system in which 0° is towards the top of the figure, 180° towards the bottom, 90° towards the right side and 270° towards the left side. Bathymetric contours are 500, 1000, 2000 m, etc. See Fig. 7 for key to numbered features.

trending northeast such as the Byam Martin Channel earthquake swarm (#57, Hasegawa 1977) and magnetic linears (#46, 47, 53, Reford 1967) which are coextensive with mafic dykes and faults.

Terminology

Structural features discussed in this paper have been grouped into two broad categories, viz. fractures and arches. *Fractures* include all types of breaks at the surface of the crust such as 1) faults (all types with minor displacements), 2) rifts (zones of breaks) where adjacent crustal blocks have been displaced from an initial fault in a direction normal to the fault(s), usually over considerable distances (>100 km); these include rift-generated ocean basins bounded by continental margins which were the original zones of rifts, 3) shear zones, in which adjacent crustal blocks have been translated relative to one another in a direction parallel to an initial fault or fault zone between them and may be separated by faults with accompanying grabens > 50 km wide. *Arches* is applied here very generally as a term to include 1) mid-oceanic ridges (upwelling of mantle material), 2) uplifted areas (e.g. Boothia Uplift), and 3) antiformal structures.

Large-scale structural features are shown in Fig. 1 and are named as follows: 1) Eurasia Basin Rift, a rift-generated ocean basin, extends in width from the Lomonosov Ridge to the Barents Shelf and includes the Fram Basin, Nansen Basin and Nansen–Gakkel Ridge, 2) Northern mid-Atlantic Rift, another rift-generated ocean basin, extends in width from Greenland to Norway and includes most of the region of the northern Atlantic Ocean, 3) Nansen Shear Zone extends between Svalbard and Greenland, 4) Baffin Bay Rift, a proposed rift-generated bay, is the deeper portion of the Baffin Bay between Greenland and Baffin Island, 5) M'Clure Rift, which occurs along M'Clure Strait, forms the western portion of the Parry Submarine Rift Valley (Kerr 1981) and is here divided into a rift zone (M'Clure Rift) and a shear zone, Parry Channel Shear Zone. The M'Clure Rift is considered an incipient rift zone because probably little normal separation has taken place and no oceanic crust occurs in the middle of the strait, 6) Parry Channel Shear Zone is considered to extend along Parry Channel from M'Clure Strait to Baffin Bay and includes both the eastern portion of Parry Submarine Rift Valley and the Lancaster Aulacogen (Kerr 1981). The geological data used by Kerr (1981) to define the Parry Submarine Rift Valley and the Lancaster Aulacogen may also apply to a shear zone. For example, in the Lancaster Aulacogen faulted sediments are overlain by undeformed sediments and the uppermost 2 km of the sedimentary section are not cut by major faults. Some of the faults die out upwards, so there are folds rather than faults near surface but the uppermost part of the column is not folded. Thus the faults originated at depth and were propagated upwards and could be interpreted in terms of Brown's and Riedel's shear zone model discussed below.

Rifts and shear zone models

Experiments on shear zones have revealed some interesting facts which appear to apply to the regions of the Eurasia Basin, northern Atlantic Ocean, Baffin Bay, Parry Channel and M'Clure Strait. Brown (1928) and Riedel (1929), as discussed in De Sitter (1956), showed in their experiments with wet clay or paraffin-petroleum jelly mixture on two wooden blocks (Fig. 6a) that a shearing motion originating at the bottom of the mixture propagated upward in a wedge-shaped zone widening towards the surface. The first cracks observed on the surface were obviously tension cracks making a 45° to 47° angle with the shear plane (Fig. 6a). These cracks rotated during movement, increasing the angle to 50° or 60°. The tension cracks were parallel to the principal normal stress produced by the applied shearing stress. After the development of the tension cracks a zone of shear developed at the surface in which a second set of cracks developed at angles of 10° to 15° to the direction of shear.

These experimental structures appear to be analogous to observed, well-developed features in the North Atlantic region and to similar less clearly developed features in the Canadian Arctic Archipelago. Thus, in spite of the great differences in the elastic parameters of the crust and those of the experimental media there appears to be justification for drawing an analogy between the patterns and nature of faults produced in models and observed in the crust.

Eurasia Basin – Northern mid-Atlantic Rift system

The Eurasia Basin – Northern mid-Atlantic Rift system is well developed and about 57 Ma old (Grønlie et al. 1979). The edges of these rifts and the transform shear zone that joins them are approximated by straight lines drawn along the shelf breaks (Figs 1 and 5). These rift edges are taken to be analogous to the tension cracks in Brown's and Riedel's experiments. If their history followed this model the Eurasia Basin Rift and Northern mid-Atlantic Rift developed 57 Ma ago, initially at an angle of about 45° to the shear zone (Fig. 6b) and rotated during spreading to their present positions which make angles of 64° to 72° with respect to the Nansen Shear Zone (Fig. 6c) which developed later on the surface around 36 Ma ago (Fig. 5). In addition, two frac-

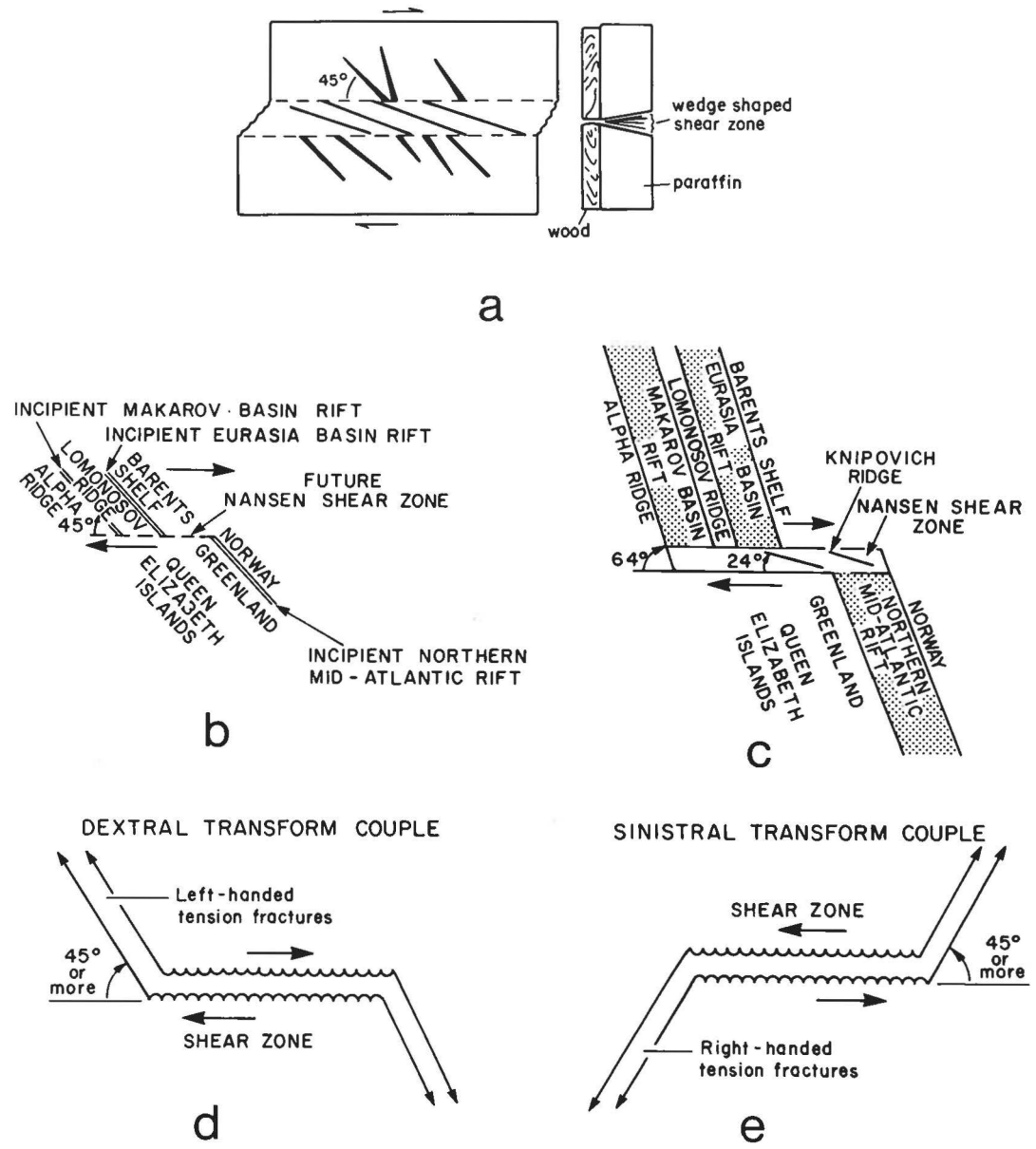


Fig. 6. (a) Fracture diagram explaining Brown's (1928) and Riedel's (1929) experiments. 45° tension cracks (rifts) appear first, then the shear zone develops and finally cracks develop at 10° to 15° in the shear zones, (b) schematic early development of the Eurasia Basin – Northern mid-Atlantic Rifts, (c) schematic early development of the Nansen Shear Zone (dextral) and later development of the Eurasia Basin – Northern mid-Atlantic Rifts, (d) generalized dextral transform couple, (e) generalized sinistral transform couple.

tures (Knipovich Ridge, north and south) identified as two narrow central deeps with flanking ridges developed about 36 Ma ago in the Nansen Shear Zone with angles of 16° to 24° to the shear zone (Fig. 5), again similar to the experimental results (Fig. 6c). Thus the Eurasia Basin – Northern mid-Atlantic Rift system appears to comprise tensional fractures radiating from a probable dextral transform shear zone (Fig. 6d) and is called here a dextral transform couple. The rifts may be described as left-handed as they trend to the left from a position facing the shear zone. The Makarov Basin may be another left-handed tensional rift zone lying subparallel to the Eurasia Basin Rift (Figs 6b, c) if the Nansen Shear Zone extends that far west.

M'Clure – Baffin Bay Rift system

This is a weakly-developed system (Figs 1 and 5) and is discussed in part by Jackson et al. (1979) and Kerr (1981). It lies some 2000 km southwest of, and more or less sub-parallel to, the Eurasia Basin – Northern mid-Atlantic Rift system. The M'Clure and Baffin Bay Rifts developed initially about 60 Ma ago, approximately at the same time as the Eurasia Basin and Northern mid-Atlantic Rift system. Then, about 38 to 25 Ma ago, the proposed Parry Channel Shear Zone developed and jointed the M'Clure – Baffin Bay Rift system as a weakly-developed, probably dextral, transform shear zone. In this paper Parry Channel is considered a shear zone even though geological evidence (Kerr 1981, pers. comm.) indicates extensional features across the channel. Parry Channel exhibits features analogous to the shear zone in Brown's (1928) and Riedel's (1929) experiments. In the experiments the shear zone (Fig. 6a), consisting of inward, steeply-dipping faults, progressively widens from depth towards the surface as the faults propagate upwards. Graben structures could develop within the shear zone if adjacent landmasses drifted apart while being translated parallel to the shear zone. Generally the inward-dipping faults are shown by Kerr (1981: fig. 17) for Lancaster Aulacogen which in this paper is considered to be part of the Parry Channel Shear Zone. The channel is about 100 km wide and if Kerr's geological section is extrapolated downwards assuming the dips of the faults to be constant (45°) then the shear zone could extend to a depth of about 50 km into the upper mantle. The faults in Parry Channel apparently die out within the sediments of sequences 6 and 7 (Fig. 3) and do not reach the surface (Kerr 1981). Sequence 3 is depicted as having different thicknesses on either side of the faults (Kerr 1981, fig. 17) and this might be interpreted to result from horizontal (shear) displacements along Parry Channel. The geological evidence, therefore, is not inconsistent with a shear zone developed as a wedge-shaped system of upward propagating faults, although evidence for shear movement may not be apparent at the surface until much later than the attendant rift zones, and may have very little indication of translational movement on the surface.

Determination of the azimuths of the rift zones is sometimes problematical. For example, the azimuth of the centre line of the Baffin Bay Rift may be taken as approximately 308° from bathymetry or 324° as indicated by the central negative gravity anomaly (Jackson et al. 1979). Using the latter value the M'Clure and Baffin Bay Rifts form angles of 58° and 50° respectively to the later developed proposed Parry Channel Shear Zone. Brown's and Riedel's models predict that rifts initially form at angles of 45°. These larger angles suggest that small angular rotations of 13° and 5° of the M'Clure and Baffin Bay Rifts and adjacent crustal blocks have taken place. If the magnitude of the angular rotation is any measure of the expected separation across the rifts then the width of Baffin Bay would be expected to be less than the 100 km width of the M'Clure Rift. However, it is 500 km in places! This paradox may result from either considerably greater crustal stretching during the initial development of the Baffin Bay Rift or because the Baffin Bay Rift is also a complementary rift of the Nares Strait lineament. Some support of the idea of crustal stretching is provided by Umpleby (1979) who has proposed that the margin of the Labrador Shelf is formed on downdropped, stretched continental crust, and by Le Pichon & Sibuet (1981) who have presented a simple stretching model for the continental margin in the northeast Atlantic. Their model explains the relationship between initial subsidence, thermal subsidence, continental crust thinning and gives the reason for the transition from continental stretching to oceanic accretion.

Nares Strait lineament

The Nares Strait lineament between Greenland and Ellesmere Island has been a focus of discussion for years (Kerr 1980b); some people advocate no movement along it, others favour a considerable amount of sinistral movement (up to 400 km). The present study tends to support the case for little or no movement, more or less in accord with Dawes' & Kerr's summary in this volume, who indicate that the boundaries between the various facies in Proterozoic and Palaeozoic rocks on either side of Nares Strait are in alignment. Riddihough et al. (1973) noted that magnetic anomalies are continuous across Nares Strait with no apparent offset along the Strait. A strong gravity gradient with a change in the gravity field of more than 100 mgal from relatively high values over Lincoln Sea and Ellesmere Island to lower values over Greenland (Sobczak & Stephens 1974), lies sub-parallel to the structural grain of the Palaeozoic strata and the north coast of Greenland. If this gradient is related to structures formed prior to 60 Ma ago then, because it is not offset along Nares Strait (35, azimuth of 14°, Fig. 5), very little movement along the Strait is indicated. On the other hand movement parallel to the gradient (40, azimuth 28°, Fig. 5) might not have disturbed the gradient significantly. For example, Nares Strait, about 50 km wide, could have been created by a sinistral movement of about 200 km along the gradient.

However, if this movement did occur along the gradient in the last 60 Ma then the Greenland side of the Nansen Shear Zone would be offset about 115 km. As there is no such apparent offset, then also very little movement is likely to have taken place along the gradient. Preliminary analysis of this gradient by the author suggests that the crust below Greenland, most of Nares Strait and Judge Daly Promontory of Ellesmere Island is either thicker by 6 to 8 km than the crust below Lincoln Sea and northern Ellesmere Island areas, or

lower in mean crustal density by 0.07 Mg/m³. Regardless of the cause of the gradient and providing the gradient has not occurred recently as a result of glacial loading, a major transition must take place along a line which crosses Nares Strait at an oblique angle and along which there has been no significant transcurrent movement.

Nares Strait is considered in Fig. 1 to be a shear zone as produced in Riedel's (1929) experiment; that is, there may be many inward and steeply-dipping faults that propagate upwards in a wedge-shaped shear zone that has undergone very little surface shear movement. In the latter respect it would be similar to Parry Channel. Initially, development of a shear zone would not require strike-slip movement at the surface and the shear movement would be manifested initially in the development of rift zones at either end. In this case the Baffin Bay Rift would form one rift zone at the south end of the shear making an angle of 50° with the shear zone, and the Eurasia Basin Rift would form the complementary rift making an angle of 61°; this system would be a sinistral transform couple according to the definition (Fig. 6e). As well, a northward extension of the Nares Strait lineament intersects the Nansen–Gakkel Ridge near the northward extension of the northeastern edge of the Nansen Shear Zone. Also the combined width of the Eurasia Basin and Makarov Basin Rifts is more than 100 km wider than the Northern mid-Atlantic Rift along the Nansen Shear Zone; this excess width on the Eurasian side of the shear zone may suggest that additional rifting sub-parallel to the Northern mid-Atlantic Rift may have taken place in Baffin Bay. The pattern that emerges across the entire region is that of three separate transform couples interlinked at two triple junctions (Parry Channel, Baffin Bay, Nares Strait junction and Eurasia Basin, Nares Strait, northern Atlantic Ocean junction).

Orientations of structures and tectonic forces with time

An examination of the orientations of structures formed during various time intervals and of recent principal forces (Figs 5 and 7) may provide a further insight into the fragmentation of the Arctic region. All the structural features are plotted on a polar stereographic projection (Fig. 5), but for convenience in measuring azimuths and making comparisons between the trends of various features an arbitrary rectangular coordinate system has been superimposed on this polar stereographic projection; the top and bottom of the map (Fig. 5) parallel to the 0°–180° azimuth of the polar stereographic projection are taken as 0° and 180° respectively and the left and right sides 270° and 90° in the rectangular system used here. Such a planar representation for the real surface produces only minor distortions as was discussed for a comparable projection, a polar azimuthal equidistant projection by Sweeney & Haines (1978). Distortions of distance between parallels of latitude increase southwards by a factor $(90°-\lambda)\text{rad.}/\sin(90°-\lambda)$ where λ is latitude. For example, a change of 30° of latitude represents a distortion of 47 km for a surface distance of 1000 km. As most of the azimuths were determined between 70° to 85° latitude a distortion of 39.4 km in a distance of 1675 km would represent an error in azimuth of 1.3 degrees which is considered insignificant in this study.

Azimuths of structures and forces have been grouped into broad time categories (Recent, Sverdrup Basin, and Franklinian Geosyncline intervals, Fig. 7). Azimuths of recent tectonic events are shown at the hub of Fig. 7, and include azimuths of deviatoric forces and faults and focal depths calculated in Byam Martin Channel and Beaufort Sea by Hasegawa (1977) and Hasegawa et al. (1979). In Byam Martin Channel four large (magnitude 5.1 to 5.7) earthquakes in 1972 with focal depths between 10 and 30 km yielded average deviatoric tension (T_1) and compression (P_1) azimuths of 340° and 90° respectively and a strike of the preferred, nearly normal (85°), fault plane (F_1) of 35°. These azimuths in the rectangular coordinate system used here are 357°, 106° and 50° respectively (Figs 5 and 7). In the Beaufort Sea a large (magnitude 4.2) earthquake in 1975 with a focal depth of 40 km gave azimuths in the rectangular coordinate system of (T_2) 204°, (P_2) 42°, and (F_2) 90°.

The azimuths of the folds and fractures that formed during development of the Sverdrup Basin interval (mainly Eurekan Deformation) are shown radiating from the hub of Fig. 7. Fractures (including shear zones) trend between 14° and 94° with a mean value close to the present fault direction of 50° in Byam Martin Channel (Hasegawa 1977), that is between the deviatoric tension (T_1) and compression (P_1) forces of the Byam Martin Channel earthquakes, while arches and rifts trend between 297° and 335° between T_1 and P_1 forces. The force diagram for the Beaufort Sea earthquake appears to be unrelated to the fractures and arches of the Sverdrup Basin interval as these fall in different quadrants of the force diagram of the Beaufort Sea (Fig. 7). Some of these azimuths were approximated from the general trend of many individual fractures, while some are more reliable such as those determined for the Cornwall Island faults over the Cornwall Arch by Balkwill (1974); the average azimuth of 54 faults is 35° and the average fold axis azimuth is 323°. Azimuths of fold axes is about half the variation in direction of the fractures. Apparently azimuths of folds and fractures in the Queen Elizabeth Islands are similar to those within the North Atlantic Ocean (Fig. 7). This suggests that both regions probably have been subjected to similar stresses regardless of the fact that two different types of crusts (oceanic and continental) are present. Ages of

most oceanic structural features are indicated in Figs 5 and 7.

During the formation of the Franklinian Geosyncline (i.e. mainly during the Ellesmerian Orogeny, Fig. 3) azimuths of arches (237° to 333°) have a much larger variation (96°) than those produced in the Sverdrup Basin interval. The Cornwallis Fold Belt, which was uplifted as much as 7700 m (Kerr 1977), was probably influenced by the underlying Boothia Uplift, a northward-trending, Precambrian salient which plunges to the north below the Sverdrup Basin. This uplift existed during Precambrian time, persisted during the Phanerozoic and influenced the tectonics within the region. It is parallel to the alignment of the deviatoric tension force T_1 and normal to the deviatoric compression force P_1 (Hasegawa 1977). With the exception of the Boothia Uplift, no significant fractures dating from the Franklinian Geosyncline interval have been found.

Fractures and folds

It is difficult from simple fracture-fold theory to determine the exact nature of external forces from the pattern of fractures and arches (De Sitter 1956, Billings 1958). During the Sverdrup Basin interval, fractures generally occur in the first quadrant (0° to 90°) in both old, continental crust (Queen Elizabeth Islands) and new, oceanic crust (northern Atlantic Ocean). However, obvious fractures did not also occur in the fourth quadrant (270° to 360°), but, instead, arches predominate in this quadrant. Part of the explanation of this distribution may lie in the fact that most of the northwest-trending features are mid-oceanic ridges along which magma has intruded along zones of weakness or fractures. The two continental arch features, the Cornwall Arch and Princess Margaret Arch (Fig. 2), may also be zones of weakness that are largely intruded by igneous rocks. Along the Cornwall Arch evidence from drill holes and gravity data (Sobczak & Overton in prep.) suggest that a significant amount of mafic igneous rock is intruded along and may have largely caused the Cornwall Arch. On the other hand folds can develop at right angles to a principal compressive force (Billings 1958). Perhaps prior to and during the opening of the Eurasia Basin – Northern mid-Atlantic Rifts and the M'Clure – Baffin Bay Rifts compressive forces acted along the axis of the Sverdrup Basin (azimuth 50°) and may have been partially responsible for structures such as the Cornwall Arch and Princess Margaret Arch.

Discussion

Structures, earthquakes analysis and experimental models may indicate the tectonic forces that fragmented the Arctic region. Tectonic forces that affected the North American continent and adjacent oceans during the Sverdrup Basin interval produced arches and rifts whose azimuths trend northwest and fractures whose azimuths trend generally to the northeast (Fig. 7). However, during the Franklinian Geosyncline interval there were, apparently, no fractures, and azimuths of arches have a much greater range varying from the southwest to the northwest, making interpretation of the force field more difficult. The Boothia Uplift, a north-trending crystalline structure, persisted as a positive element throughout the Sverdrup Basin and Franklinian Geosyncline intervals with an azimuth that was consistent and intermediate to those of the arches and fractures (Fig. 7). With the exception of large rifts, structures produced in the Sverdrup Basin interval have similar azimuths, so it appears that they may have been produced by similar tectonic forces. Tertiary and early Mesozoic forces which caused these structures probably had azimuths similar to the present-day deviatoric forces (Fig. 7) as determined from four large earthquakes in Byam Martin Channel by Hasegawa (1977). Fault plane solutions for these earthquakes indicated dextral strike-slip motion in a northeasterly direction which is more or less the mean azimuth of the fractures for the Sverdrup Basin interval. Azimuths of tensional and compressional deviatoric forces for the Beaufort Sea earthquake lie 53° and 64° respectively counterclockwise from those determined in Byam Martin Channel (Fig. 7) (Hasegawa et al. 1979) and may be representative of the present-day force field in the Canada Basin.

Azimuths of fractures and arches produced during the Sverdrup Basin interval vary by as much as 80° and 38° respectively; these variations may relate to a rotation of up to ± 40° of a deviatoric force system having an initial orientation similar to that shown for recent forces in Byam Martin Channel (Fig. 5). Alternatively the magnitude of forces along the tensional and compressional directions may have varied with time. Older underlying structures and the structural grain may have also diverted or influenced the fracture–arch directions.

Arches and rifts produced during the Sverdrup Basin interval have similar azimuths which is paradoxical because arches are usually compressive features whereas rifts are tensional. This suggests that fragmentation within the Arctic region may be the result of two types of interacting deviatoric tectonic forces such as a regional tension-tension force couple acting longitudinally and latitudinally as proposed for the North Atlantic Ocean and Baffin Bay areas (Fig. 5) and a local (Queen Elizabeth Islands) tension-compression force couple as indicated for the Byam Martin Channel (Fig. 5) (Hasegawa 1977). The tension-tension forces may be a result of a mantle convection current or a revolving ellipsoidal earth with centrifugal forces that act radially from the North Pole. Probably one or both of these gradually forced the continents to move southward as

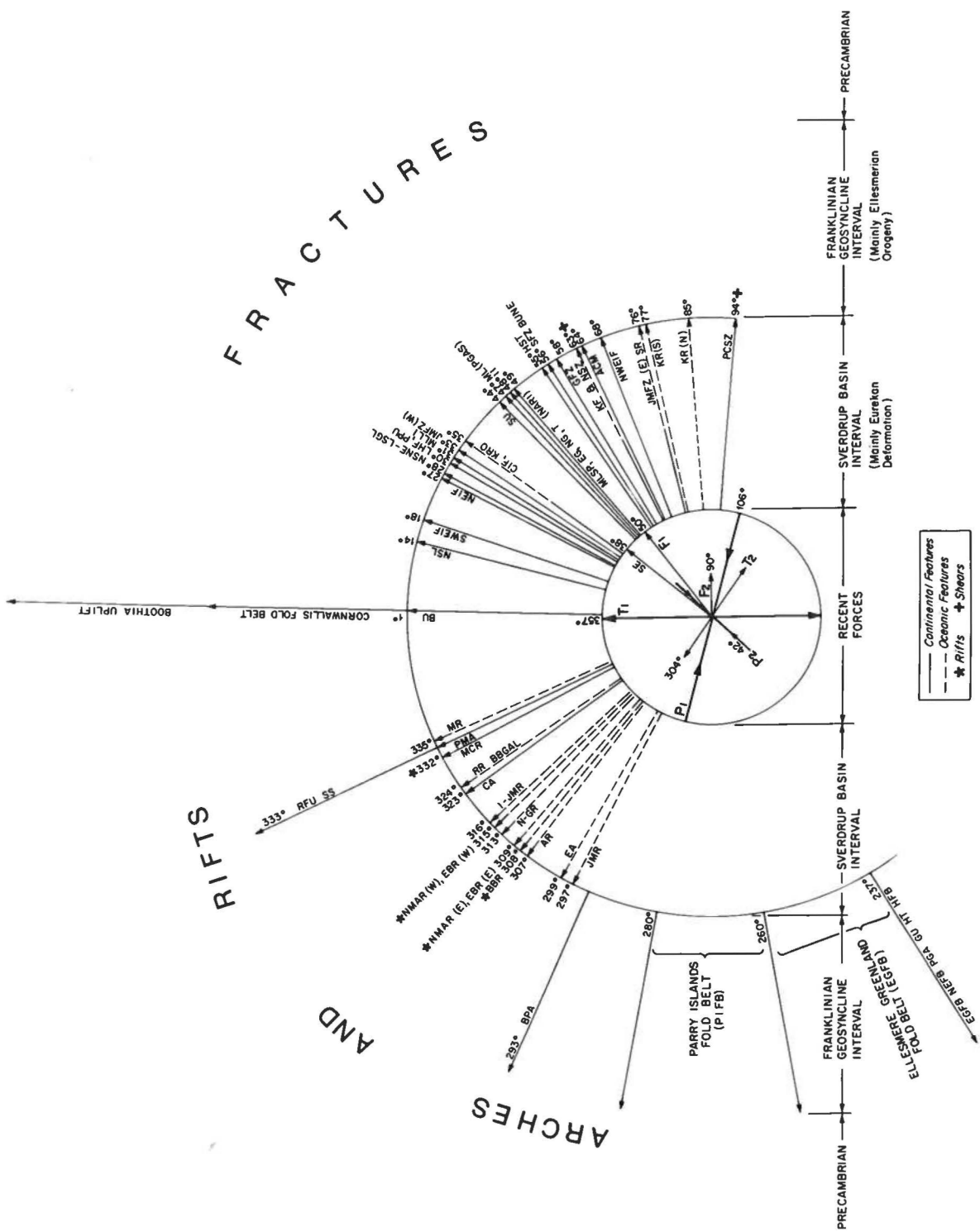


Fig. 7. Azimuth plot of structural features and forces relative to a rectangular coordinate system. For the Byam Martin Channel earthquake with focal depths between 10 and 30 km (Hasegawa 1977), deviatoric tensional and compressional forces shown as T_1 and P_1 and fault direction as F_1. Similarly for the Beaufort Sea with a focal depth of 40 km (Hasegawa et al. 1979) these quantities are shown as T_2, P_2, and F_2.

Explanation of abbreviations from Figs 3 and 7; numbers are from Fig. 5.

Arches

36	EGFB:	Ellesmere–Greenland Fold Belt (237° to 260°, Kerr 1981)
42	NEFB–PGA:	Northern Ellesmere Fold Belt – Pearya Geanticline (237°, Kerr 1981)
38	HFB:	Hazen Fold Belt (237°, Kerr 1981)
38	HT:	Hazen Trough (237°, Kerr 1981)
38	GU:	Grantland Uplift (237°, Kerr 1981)
54, 55	PIFB:	Parry Islands Fold Belt (260°–280°, Tozer & Thorsteinsson 1964)
33	BPA:	Bache Peninsula Arch (293°, Kerr 1981)
17	JMR:	Jan Mayen Ridge (297°, 17–23 Ma, Grønlie et al. 1979)
14	EA:	Extinct axis (299°, 26–55 Ma, Grønlie et al. 1979)
1	AR:	Alpha Ridge (307°, Fig. 5)
2	LR:	Lomonosov Ridge (315°, Fig. 5)
3	N–GR:	Nansen–Gakkel Ridge (313°, Fig. 6)
18	I–JMR:	Iceland – Jan Mayen Ridge (316°, 10 Ma, Grønlie et al. 1979)
27	CA:	Cornwall Arch (323°, Balkwill 1974)
20	RR:	Reykjanes Ridge (324°, 10 Ma, Grønlie et al. 1979)
23	BBGAL:	Baffin Bay gravity anomaly low (324°, Jackson et al. 1979)
44	RFU:	Rens Fiord Uplift (333°, Trettin 1969)
43	SS:	Svartevaeg Syncline (333°, Trettin 1969)
32	PMA:	Princess Margaret Arch (333°, Thorsteinsson 1974)
15	MR:	Mohns Ridge (335°, Sobczak & Sweeney 1978, 57 Ma, Grønlie et al. 1979)
26	CFB:	Cornwallis Fold Belt (1°, Fortier et al. 1963, Kerr 1977)
26	BU:	Boothia Uplift (1°, Fortier et al. 1963, Kerr 1977)

Rifts

21, 22, 24	BBR:	Baffin Bay Rift (308°, 38–60 Ma, Jackson et al. 1977, 1979, Fig. 5)
12	NMAR(E):	Northern mid-Atlantic Rift (East), (309°, 57 Ma, Fig. 5)
7	EBR(E):	Eurasia Basin Rift (East), (309°, Fig. 5)
19	NMAR(W):	Northern mid-Atlantic Rift (West), (315°, 57 Ma, Fig. 5)
2	EBR(W):	Eurasia Basin Rift (West), (315°, Fig. 5)
56	MCR:	M'Clure Rift (332°, Fig. 5)

Fractures and other linear features

35	NSL:	Nares Strait Lineament (14°, Kerr 1981)
34	SWEIF:	Southern Ellesmere Island faults (18°, Thorsteinsson 1974)
39	NEIF:	Northern Ellesmere Island faults (27°, Thorsteinsson 1974)
40	NSNE–LSGL:	Nares Strait north end – Lincoln Sea gravity linear (28°, Sobczak & Stephens 1974)
41	LHF:	Lake Hazen Fault (30°, Christie 1964)
51	PPU:	Prince Patrick Uplift (30°, Thorsteinsson & Tozer 1960)
47	MLLI:	Magnetic linears Lougheed Island area (31°, Forsyth et al. 1979)
16	JMFZ(W):	Jan Mayen Fracture Zone (West) (33°, 26–36 Ma, Grønlie et al. 1979)
30	CIF:	Cornwall Island Faults (35°, Balkwill 1974)
6	KRO:	Knipovich Ridge offset (35°, Sobczak & Haines 1978)
57	SE:	Seismic epicentres (38°, Byam Martin Channel, Hasegawa 1977)
58	SU:	Storkerson Uplift (44°, Kerr 1981)
46	ML(PGAS):	Magnetic linears (Prince Gustaf Adolf Sea) (47°, Forsyth et al. 1979)
45	II:	Igneous intrusions (strike of gravity anomaly) (48°, Sobczak & Overton in prep.)
53	MLSP:	Magnetic linears Sabine Peninsula (49°, Forsyth et al. 1979)
53	NG:	Narrow grabens (reflection seismic), (49°, Forsyth et al. 1979)
29	T(NARI):	Trough (northwest of Amund Ringnes Island), (49°, Sobczak & Overton in prep.)
52	EG:	Eglinton Graben (49°, Tozer & Thorsteinsson 1964)
53	BMCSF:	Byam Martin Channel Seismic Fault (50°, Hasegawa 1977)
31	HST:	Hassel Sound Trough (55°, Sobczak & Overton in prep.)
28	BUNE:	Boothia Uplift north end (56°, Sobczak & Overton in prep.)
11	SFZ:	Senja Fracture Zone (56°, 36 Ma, Grønlie et al. 1979)
8	GFZ:	Greenland Fracture Zone (58°, 36 Ma, Grønlie et al. 1979)
50	KF:	Kaltag Fault (63°, Kerr 1981)
5, 10	NSZ:	Nansen Shear Zone (63°, Fig. 5)
49	ACM:	Arctic Continental Margin (64°, Fig. 5)
37	NWEIF:	Northwestern Ellesmere Island faults (68°, Thorsteinsson 1974)
48	SR:	Sverdrup Rim (76°, Meneley et al. 1975)
13	JMFZ(E):	Jan Mayen Fracture Zone (East), (76°, 26–36 Ma, Grønlie et al. 1979)
9	KR(S):	Knipovich Ridge (South), (77°, 36 Ma, Fig. 5)
4	KR(N):	Knipovich Ridge (North), (85°, Fig. 5)
59	BSSF:	Beaufort Sea Seismic Fault (90°; Hasegawa et al. 1979, Fig. 5)
25	PCSZ:	Parry Channel Shear Zone (94°, 25–38 Ma, Jackson et al. 1979, Fig. 5).

indicated by the opening of the Eurasia Basin along meridians of longitude from the North Pole. The continents were forced to either fracture, stretch, or both in order to occupy a larger southward circumference than when the continents were closer to the pole. Thus tensional forces would be experienced by the continents both radially and tangentially as indicated for the North Atlantic Ocean and Baffin Bay (Fig. 5). Analogous to Brown & Riedel's experiments (Fig. 6a) the crust first rifted in the North Atlantic – Eurasia Basin at a 45° angle to the direction of tensional forces which were probably of similar strength and then sheared dextrally along the Nansen Shear Zone. At the same time a sub-parallel poorly-developed, rift-transform shear zone was probably formed some 2000 km to the southwest (M'Clure Strait, Parry Channel and Baffin Bay). The plate between these dextral transform couples (Fig. 6d) continued to be affected by more or less longitudinal tensional forces but the latitudinal forces became compressional due to a sinistral force couple resulting from the formation of two transform couples. Under this stress regime, in the rectangular frame of reference, fractures developed in the northeast quadrant at some acute angle to the principal force and arch features developed in the northwest quadrant. However, most of these arch features are mid-oceanic ridges produced by upwelling mantle material probably along zones of weaknesses which may be new fractures. In the Queen Elizabeth Islands the Cornwall Arch and Princess Margaret Arch may result from intruded mafic material along zones of weaknesses and the effects of an underlying crystalline crust that compressed along and stretched perpendicular to the axis of the Sverdrup Basin (Sobczak & Overton in prep.).

A crystalline crust that can stretch, and over a period of time behave more as a plastic than a rigid body, is another important tectonic feature to consider in the fragmentation and evolution of the region. It can best explain the undulating Moho and the thickness variation of the crystalline crust below the Sverdrup Basin (6 to 10 km thinner below the thickest sedimentary column, Fig. 4, Sobczak & Overton in prep.), the observed subsidence (Sweeney 1977), isostatic equilibrium and crustal thinning in the continental margin areas (Sobczak 1975a, b). As well it can offer an additional mechanism for the somewhat erratic fragmentation of the region. If the lower crystalline crust can be considered to behave like a balloon with strong and weak spots in it, then the weaker, more flexible spots could experience more stretching, fragmentation and subsidence. Perhaps this plastic variation can partly explain differences in widths between sub-parallel rift zones such as the Baffin Bay and M'Clure Rifts. Further, a plastic crust can explain arch features such as the Cornwall Arch and Princess Margaret Arch. If, in the past, the compressional force was rotated from present-day directions counter-clockwise more along the axis of the Sverdrup Basin, with the tensional force perpendicular to the Arctic continental margin, then as the crystalline crust stretched below in line with the tensional force the sediments and crystalline crust would subside and fold perpendicular to the compressive force. This picture is consistent with folding noted at the crust–mantle boundary more or less along the axis of the Sverdrup Basin (Fig. 4).

Perhaps similar circumstances may also apply to the Alpha Ridge. The Alpha Ridge, an accordian-like feature whose origin is unknown, is sub-parallel to the ridges of the North Atlantic Ocean and Queen Elizabeth Islands and nearly normal to the compressive force determined by Hasegawa et al. (1979) in the Beaufort Sea (Fig. 5). It lies mid-way in the Arctic Ocean between the coast of Alaska and Barents Shelf. The crests of the Alpha Ridge and Lomonosov Ridge and the upper part of the Barents Shelf, fall on a straight line which has a decreasing gradient from the Barents Shelf of 2000 m per 15° of latitude (Sobczak & Sweeney 1978). The crests of the ridges may thus be considered to have subsided at the rate of about 1.2 metres per kilometre of distance from the Barents Shelf, a subsidence which may be a function of time of separation. The Lomonosov Ridge and the Barents Shelf probably separated about 55 Ma ago (Vogt et al. 1979). If subsidence in the Sverdrup Basin can be considered typical for subsiding areas in the Arctic region, Sweeney (1977) showed a mean subsidence rate for the central region of the Sverdrup Basin of about 2.4 km/55 Ma and of about 1 km/55 Ma for the northwest margin area. The submerged ridge heights of 2.0 km depth for the Alpha Ridge and 1.5 km for the Lomonosov Ridge may correspond to a similar rate of subsidence with time as these features rifted away from the Barents Shelf. If ridge heights can be used as an indicator of separation, then the Alpha Ridge may have separated before the separation of the Lomonosov Ridge and Barents Shelf because of its greater depth. Also, high amplitude magnetic linears, which probably relate to mafic intrusions, lie parallel to the ridge axis and trend southeastward over the continental margin to the coast of Ellesmere Island (Riddihough et al. 1973). DeLaurier (1978) concluded that the Alpha Ridge did not fit the spreading hypothesis because cooling models would predict very little present relief. As well, spreading ridges are more or less parallel to rift edges (Fig. 7) but no obvious sub-parallel rift edge occurs on the North American side of the Alpha Ridge. Instead the central axis of the Canada Basin lies more or less sub-parallel to the direction of major fracturing.

A speculative concept that appears to emerge is that rift zones developed on the Eurasian side of the Alpha Ridge and more complicated stretched zones (the Canada Basin and Sverdrup Basin) on the North American side. Thus the Alpha Ridge is probably a former continental margin which rifted away from the Lomonosov Ridge and Barents Shelf. Initially, the Alpha Ridge and Canada Basin may have stretched

more or less perpendicular to the ridge axis (radially from the pole) while rifting was taking place on the Eurasian side of the ridge, but as rifting terminated and landmasses on either side started to drift apart the Alpha Ridge was then probably compressed more or less perpendicular to the ridge and stretched latitudinally. Effects of stretching in and around the Canada Basin may take the form of down-faulted, inward-dipping blocks around the continental margins and Alpha Ridge similar to structures proposed by Umpleby (1979) for the Labrador Shelf, and by Le Pichon & Sibuet (1981) for passive margins; oceanic crust or heavily oceanized continental crust may be present in the centre of the basin. The greatest amount of separation is postulated to have taken place about the north-trending axis of the basin which may coincide with a buried, low-relief ridge.

Another feature that appears to emerge from this study is the variation of forces for the general region between 70° and 80° latitude (Fig. 5). From east to west, areas are alternatively under the influence of either latitudinal tension (North Atlantic Ocean and Beaufort Sea) or latitudinal compression (Queen Elizabeth Islands). If this pattern continues to the west to the region of Mendeleev Ridge, Chukchi Plateau and Northwind Ridge on the other side of the Canada Basin then it may also be under compression. This compression may have produced these more or less longitudinal trending features. In the Queen Elizabeth Islands the Boothia Uplift is more or less normal to the existing compressive force in Byam Martin Channel (Hasegawa 1977). A large number of seismic epicentres along the uplift (Basham et al. 1977, Wetmiller & Forsyth 1978) may indicate a response to this force.

Concluding remarks

The model for Cenozoic fragmentation of the Canadian Arctic Archipelago and surrounding region proposed in this paper is based on a pattern of northeast-trending fractures and faults and northwest-trending folds and rifts. In the context of this symposium Nares Strait is considered to represent a prominent example of one of the northeast-trending faults of this fracture system. It may represent a poorly-developed sinistral transform shear zone composed of many inward steeply-dipping faults, although very little surface shear movement is envisaged. This conclusion is supported by the strong gravity gradient that can be traced without apparent offset from northeastern Ellesmere Island across northern Nares Strait to Greenland. The fragmentation model in which Nares Strait links the rift-generated oceans of Baffin Bay to the south with the Eurasia Basin to the north supports the case against major strike-slip motion between Ellesmere Island and Greenland along Nares Strait.

Acknowledgements

I thank J. Wm. Kerr, H. S. Hasegawa, R. L. Coles, A. K. Goodacre, J. F. Sweeney and R. P. Riddihough for reading and commenting on earlier versions of the manuscript. Special thanks are extended to M. D. Thomas for suggestions and editorial assistance. Thanks are also extended to S. Cumyn and B. Draper for drafting and C. Lacasse and M. Whissell for typing.

References

Balkwill, H. R. 1974. Structure and tectonics of Cornwall Arch, Amund Ringnes and Cornwall Islands, Arctic Archipelago. – In: Aitken, J. D. & Glass, D. J. (eds), Canadian Arctic Geology: 39–62. – Geol. Ass. Can./Can. Soc. Petrol. Geol, Toronto.

Balkwill, H. R. 1978. Evolution of Sverdrup Basin, Arctic Canada. – Bull. Am. Ass. Petrol. Geol. 62: 1004–1028.

Balkwill, H. R. & Bustin, R. M. 1980. Late Phanerozoic structures, Canadian Arctic Archipelago. – Palaeogeogr., Palaeoclim., Palaeoecol. 30: 219–227.

Basham, P. W., Forsyth, D. A. & Wetmiller, R. J. 1977. The seismicity of northern Canada. – Can. J. Earth Sci. 14: 1646–1667.

Billings, M. P. 1958. Structural geology. – Prentice-Hall Inc., New York: 514 pp. (2nd edit.).

Brown, R. W. 1928. Experiments relating to the results of horizontal faulting. – Bull. Am. Ass. Petrol. Geol. 12: 715–720.

Christie, R. L. 1964. Geological reconnaissance of northeastern Ellesmere Island, District of Franklin. – Mem. geol. Surv. Can. 331: 79 pp.

Dawes, P. R. & Kerr, J. W. 1982. The case against major displacement along Nares Strait. – This volume.

De Sitter, L. U. 1956. Structural geology. – McGraw-Hill Book Company, New York: 551 pp. (2nd edit.).

DeLaurier, J. M. 1978. The Alpha Ridge is not a spreading centre. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 87–90.

Forsyth, D. A., Mair, J. A. & Fraser, I. 1979. Crustal structure of the central Sverdrup Basin. – Can. J. Earth Sci. 16: 1581–1598.

Fortier, Y. O., Blackadar, R. G., Glenister, B. F., Greiner, H. R., McLaren, D. J., McMillan, N. J., Norris, A. W., Roots, E. F., Souther, J. G., Thorsteinsson, R. & Tozer, E. T. 1963. Geology of the north-central part of the Arctic Archipelago, Northwest Territories (Operation Franklin). – Mem. geol. Surv. Can. 320: 671 pp.

Grønlie, G., Talwani, M. & Chapman, M. 1979. Geophysical studies in the Norwegian-Greenland Sea. – Norsk Polarinst. Skr. 170: 61 pp.

Hasegawa, H. S. 1977. Focal parameters of four Sverdrup Basin, Arctic Canada, earthquakes in November and December of 1972. – Can. J. Earth Sci. 14: 2481–2494.

Hasegawa, H. S., Chou, C. W. & Basham, P. W. 1979. Seismotectonics of the Beaufort Sea. – Can. J. Earth Sci. 16: 816–830.

Jackson, H. R., Keen, C. E. & Barrett, D. L. 1977. Geophysical studies on the eastern continental margin of Baffin Bay and in Lancaster Sound. – Can. J. Earth Sci. 14: 1991–2001.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor spreading in central Baffin Bay. – Can. J. Earth Sci. 16: 2122–2135.

Kerr, J. W. 1977. Cornwallis Fold Belt and the mechanism of basement uplift. – Can. J. Earth Sci. 14: 1374–1401.

Kerr, J. W. 1980a. A plate tectonic contest in Arctic Canada. – In: Strangway, D. W. (ed.), The continental crust and its mineral deposits. – Spec. Pap. geol. Ass. Can. 20: 457–486.

Kerr, J. W. 1980b. Did Greenland drift along Nares Strait? – Bull. Can. Petrol. Geol. 28: 279–289.

Kerr, J. W. 1981. Evolution of the Canadian Arctic Islands: a transition between the Atlantic and Arctic Oceans. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 105–199. – Plenum Press, New York & London.

Le Pichon, X. & Sibuet, J.-C. 1981. Passive margins: a model of formation. – J. geophys. Res. 86: 3708–3720.

Meneley, R. A., Henao, D. & Merritt, R. K. 1975. The northwest margin of the Sverdrup Basin. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 531–544.

Reford, M. S. 1967. Aeromagnetic interpretation, Sverdrup Basin. – Depart. Indian Affairs & Northern Development, Oil & Gas Technical Rep. 1975: 678–7–10–1 to 678–7–10–5.

Riddihough, R. P., Haines, G. V. & Hannaford, W. 1973. Regional magnetic anomalies of the Canadian Arctic. – Can. J. Earth Sci. 10: 157–163.

Riedel, W. 1929. Zur Mechanik geologischer Brucherscheinungen. – Centr. bl. Min. Abt. B: 354–368.

Sobczak, L. W. 1975a. Gravity and deep structure of the continental margin of Banks Island and Mackenzie Delta. – Can. J. Earth Sci. 12: 378–394.

Sobczak, L. W. 1975b. Gravity anomalies and passive continental margins, Canada and Norway. In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 743–761.

Sobczak, L. W. & Haines, G. V. 1978. Arctic bathymetry chart. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): plate 1.

Sobczak, L. W. & Overton, A. in prep. Shallow and deep crustal structure of the western Sverdrup Basin. – Data presented at the Geol. Ass. Can. & Min. Ass. Can. joint meeting 1980. Prog. with Abs. 5: 81 only.

Sobczak, L. W. & Stephens, L. E. 1974. The gravity field of northeastern Ellesmere Island, part of northern Greenland and Lincoln Sea with map. Lincoln Sea 1:500 000. – Earth Physics Branch, Gravity Map Series 114: 9 pp.

Sobczak, L. W. & Sweeney, J. F. 1978. Bathymetry of the Arctic Ocean. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 7–14.

Sobczak, L. W. & Weber, J. R. 1973. Crustal structure of the Queen Elizabeth Islands and polar continental margin, Canada. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 517–525.

Sweeney, J. F. 1977. Subsidence of the Sverdrup basin, Canadian Arctic Islands. – Bull. geol. Soc. Am. 88: 41–48.

Sweeney, J. F. & Haines, G. V. 1978. Arctic geophysical review – an introduction. – In: Sweeney, J. F. (ed.), Arctic geological review. – Publ. Earth Physics Branch 45(4): 1–6.

Talwani, M. & Eldholm, O. 1977. Evolution of the Norwegian-Greenland Sea. – Bull. geol. Soc. Am. 88: 969–999.

Thorsteinsson, R. 1974. Carboniferous and Permian stratigraphy of Axel Heiberg Island and western Ellesmere Island, Canadian Arctic Archipelago. – Bull. geol. Surv. Can. 224: 115 pp.

Thorsteinsson, R. & Tozer, E. T. 1960. Summary account of the structural history of the Canadian Arctic Archipelago since Precambrian time. – Pap. geol. Surv. Can. 60–7: 23 pp.

Tozer, E. T. & Thorsteinsson, R. 1964. Western Queen Elizabeth Islands, Arctic Archipelago. – Mem. geol. Surv. Can. 332: 242 pp.

Trettin, H. P. 1969. Pre-Mississippian geology of northern Axel Heiberg and northwestern Ellesmere Islands, Arctic Archipelago. – Bull. geol. Surv. Can. 171: 82 pp.

Umpleby, D. C. 1979. Geology of the Labrador Shelf. – Pap. geol. Surv. Can. 79–13: 34 pp.

Vogt, P. R., Taylor, P. T., Kovacs, L. C. & Johnson, G. L. 1979. Detailed aeromagnetic investigation of the Arctic Basin. – J. geophys. Res. 84: 1071–1089.

Wetmiller, R. J. & Forsyth, D. A. 1978. Seismicity of the Arctic, 1908–1975. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 15–24.

The evolution of the Nares Strait lineament and its relation to the Eurekan orogeny

JOHN W. PEIRCE

Peirce, J. W. 1982. The evolution of the Nares Strait lineament and its relation to the Eurekan orogeny. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 237–252.

If one accepts that the Labrador Sea was closed in Early Cretaceous time, then this assumption creates extra space northwest of Greenland. Where Ellesmere Island lay within it relative to Greenland and also to North America becomes a major problem in geological interpretation. Whatever solution is chosen creates conflicts with the current interpretations of the geology of the eastern Sverdrup Basin and/or the geology of Nares Strait.

The main elements of the tectonic model presented herein are: 1) Early to Middle Cretaceous: incipient rifting in the Labrador Sea terminated in an RRR triple junction north of Bylot Island and established the Nares Strait and Lancaster Sound lineaments; 2) Late Cretaceous to Early Paleocene: Greenland and Ellesmere Island rotated together counter-clockwise relative to North America, causing crustal shortening in the Sverdrup Basin (the first phase of the Eurekan orogeny); 3) Late Paleocene to Early Oligocene: responding to a new pole of rotation, Greenland moved left-laterally relative to Ellesmere Island along the Wegener Transform Fault; 4) Middle Oligocene: Greenland and Ellesmere Island moved northwestwards some 40–50 km (the main phase of the Eurekan orogeny); 5) Middle Oligocene to Present: compressive stresses relaxed followed by rapid subsidence of all basins.

This model requires that the Early Cretaceous Sverdrup Basin was much wider than today. An analogy is drawn with the thin-skinned tectonic model for the southern Appalachians as a possible mechanism to reconcile this requirement with the known geology. As major lateral Tertiary motion on the Wegener Fault is at odds with the geological interpretations across Nares Strait, this model suggests that either an alternative geological interpretation of Nares Strait (perhaps in the context of a fault zone) be found, alternative locations to accommodate the motion be found, or else the currently accepted tectonic history of the North Atlantic is seriously in error.

J. W. Peirce, Petro-Canada Exploration Inc., P. O. Box 2844, Calgary, Alberta, Canada, T2P 3E3.

Nares Strait is the long linear waterway which separates Ellesmere Island and Greenland (Fig. 1). Parts of it are commonly referred to as Robeson Channel, Hall Basin, Kennedy Channel, Kane Basin and Smith Sound (see Fig. 3).

The debate (see Kerr 1980a for a review) about the amount of left-lateral displacement along this lineament began with early papers by Taylor (1910) and Wegener (1924). It has remained unresolved since then, largely because correlations of geological and structural features on either side of the Strait require considerable interpretation. Those favouring little or no significant motion along Nares Strait include most of the geologists who have worked in the area. Kerr (1967: 483) claimed that correlation of the stratigraphy across the Strait "conclusively disproves the suggestion that Greenland has drifted hundreds of kilometers to the northeast by strike-slip movement". Dawes (1973) felt that the geological arguments against major motion were inconclusive, but after several seasons of field work in the area, he (pers. comm. 1980) strongly supports the consensus that the geological correlations between Ellesmere Island and Greenland are consistent with little or no net lateral offset.

Those favouring major left-lateral offset (200 km or more) along the Wegener Fault in Nares Strait approach the problem from a regional perspective using plate tectonic theory but lack first-hand knowledge of the geology on either side of Nares Strait. Kristoffersen & Talwani (1977) and Srivastava (1978) demonstrated that an extinct triple junction south of Greenland and the magnetic anomalies in the Labrador Sea require major left-lateral motion along Nares Strait if all the plates have behaved rigidly to a first order approximation. Sclater et al. (1977) demonstrated that a comprehensive reconstruction of the North Atlantic plates predicts a similar motion along Nares Strait.

Recent work by the Greenland Geological Survey reported in this volume (Dawes & Kerr, Dawes et al., Frisch & Dawes, Higgins et al., Hurst & Kerr, Peel & Christie, Peel et al.) has strengthened the geological arguments supporting no major offset. Newman (1977, this volume) has suggested alternative interpretations, but clearly the most convincing geological interpreta-

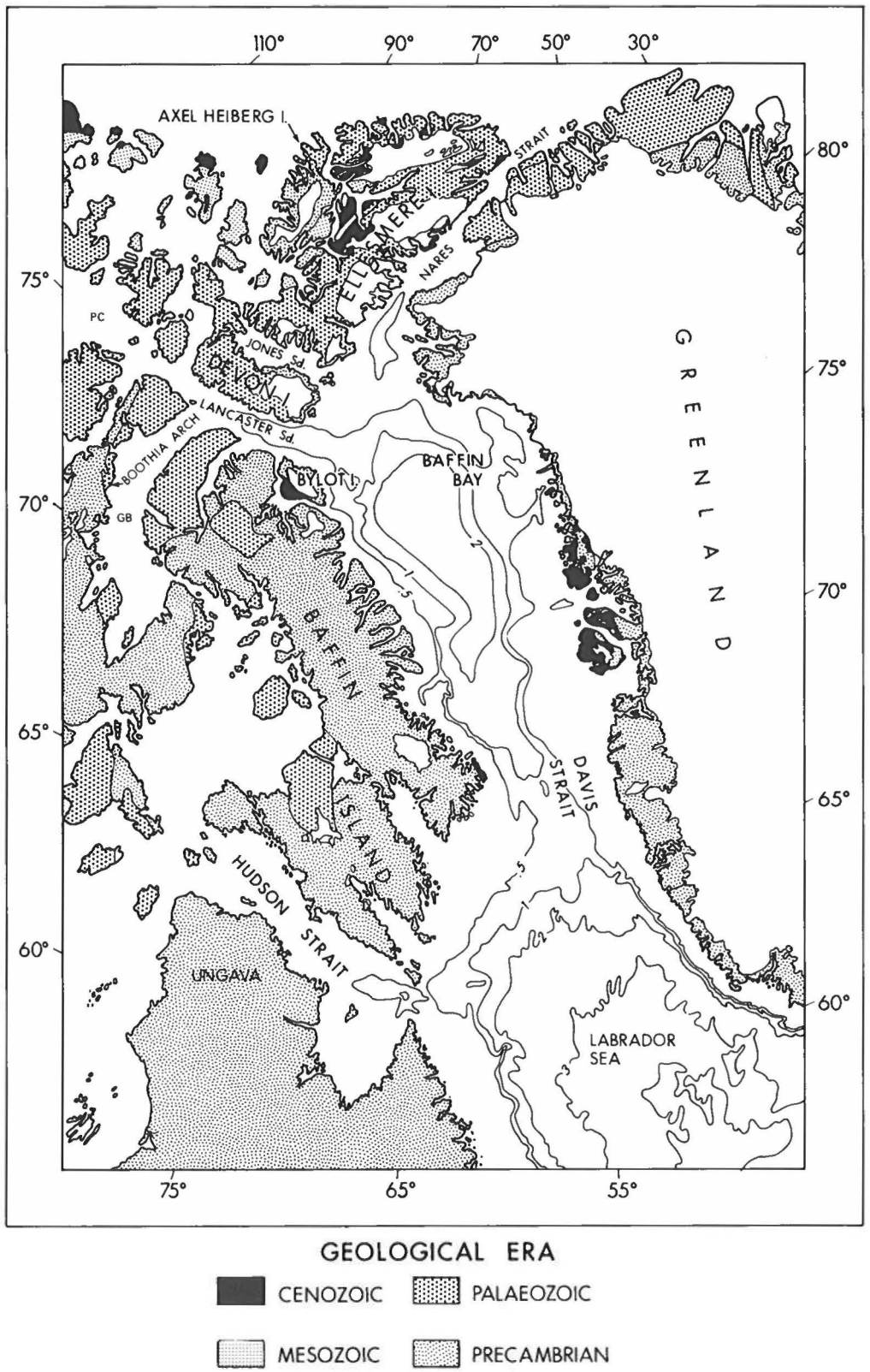


Fig. 1. Index map showing general geology and place names. Judge Daly Promontory is the long peninsula on Ellesmere Island which runs parallel to Nares Strait at 65°W (see Fig. 3). The Lake Hazen intermontane basin is the small elongated basin with some Tertiary sediments which lies just north of Judge Daly Promontory. GB = Gulf of Boothia, PC = Parry Channel.

tions are at odds with any regional model requiring major offset on the Wegener Fault.

There is little geophysical evidence which bears directly on the correlations across Nares Strait. There is a magnetic anomaly near the edge of the Palaeozoic folding on Ellesmere Island which appears to intersect the coast at a higher angle than the thrust faults. Riddihough et al. (1973) cited this as evidence against extensive left-lateral displacement along the Wegener Fault. However, a careful look at their published map reveals that this anomaly ends at the coast of Greenland and does not continue into the craton. Furthermore, 250 km to the northeast there is an anomaly which does continue into the craton at the same angle proposed by their correlation. In my opinion, there is a better correlation with displacement of 250 km than without, but neither correlation is particularly convincing.

The contours of the regional gravity field in the Lincoln Sea just northwest of Greenland run roughly parallel to the strike of the Palaeozoic folds. Sobczak (this volume) has argued that the lack of a major break in the gravity gradient is evidence against significant offset. However, the lack of seismicity in Nares Strait (Wetmiller & Forsyth, this volume) and the isostatic rebound pattern in the area (England, this volume) strongly suggest that Nares Strait is not tectonically active today. Therefore, it is likely that the gravity field reflects more recent events than the Wegener Fault, primarily the lithospheric flexure associated with the Ellesmerian and Greenland ice caps.

Keen & Peirce (this volume) have tested three models for the formation of Baffin Bay which are alternatives to sea-floor spreading there. As none of these models could explain the subsidence and crustal thickness of Baffin Bay and simultaneously remove the requirement for significant Tertiary motion on the Wegener Fault, they suggest that the geology in Nares Strait be reinterpreted in the context of a fault zone rather than a discrete offset.

The Nares Strait debate is not about the observations; it is a debate about what those observations mean. The geophysicists espouse a regional explanation which is an extension of their view of North Atlantic plate motions. The geologists argue for a different regional explanation which is an extension of their correlations made on a more local scale. My initial assumptions in this paper are on a North Atlantic scale, and therefore my model is inherently prejudiced toward the 'major offset' perspective. Many aspects of the regional geology have been considered in this overall tectonic model which have not been discussed previously in this particular debate. However, some of the regional geology — notably the amount of observed compression in the eastern Sverdrup Basin and the geology around Nares Strait — are undeniably at odds with this model as they are presently understood. My model offers a different perspective for these important problems along with testable implications.

With this preamble, there are several pieces to the Nares Strait puzzle which I consider key to understanding the larger tectonic picture. These are:

1) It is probable that the cratonic cores of Ellesmere Island and Greenland were once joined to that of Baffin Island as part of the North American craton.
2) The stratigraphy of the Eclipse Trough on Bylot Island indicates that an initial graben had opened by the Middle Cretaceous. Sedimentation was interrupted by the Bylot unconformity in the Early Paleocene (McWhae 1981, Miall et al. 1980).
3) The Alexis volcanics (Umpleby 1979) of offshore Labrador are Early Cretaceous in age (McWhae & Michel 1975, McWhae et al. 1981).
4) The first phase of the Eurekan orogeny (Balkwill 1978) is synchronous with the initial opening of the Labrador Sea (Srivastava 1978). Compressive motion north of the pole of rotation can explain the orientation and timing of the early Eurekan structures such as the Princess Margaret Arch (see Fig. 3).
5) The Nares Strait lineament does not form a small circle about the last pole of opening of the Labrador Sea (Srivastava 1978), apparently contradicting the idea that the Wegener Fault was a transform fault.
6) The separation of Greenland from Norway (Kristoffersen & Talwani 1977, Sclater et al. 1977, Talwani & Eldholm 1977) was synchronous with the late phase of opening of the Labrador Sea. Therefore active plate margins must have existed northeast and northwest of Greenland during this time (Srivastava & Falconer, this volume).
7) The second phase of the Eurekan orogeny produced about 40–50 km of crustal shortening in a NW–SE orientation (Balkwill 1978, McWhae 1981). This compression post-dated the last phase of opening of the Labrador Sea.
8) The major marine unconformities are the Early Cretaceous Labrador unconformity, the Middle Cretaceous Avalon unconformity (only well developed south of Baffin Island), the Late Cretaceous to Paleocene Bylot unconformity, the Oligocene Baffin Bay unconformity (best developed in northern Baffin Bay), and the Miocene Beaufort unconformity. These names were informally adopted by McWhae (1981) to aid discussion in the context of this area. Although they occur on a world-wide basis (Vail et al. 1977), all but the Avalon and Beaufort unconformities appear to have a major tectonic component in their formation in eastern Canada.

These points will be discussed in chronological order below.

Initial plate positions

One premise for determining the initial positioning of plates is to close up the Mesozoic and Cenozoic ocean basins. This puts Greenland up against Baffin Island near the position originally proposed by Bullard et al. (1965) (Fig. 2). The problem which this premise creates is that an extra wedge of space appears northwest of Greenland which no longer exists today. What was the spatial relationship between Greenland and the Arctic Islands? If one leaves Ellesmere Island in its present position relative to North America, then how can one explain the geological correlations across Nares Strait or remove the intervening wedge of material? Likewise if Ellesmere Island is put up against Greenland to avoid these arguments, how can one account for some 200 km of compression at the northern rim of the Sverdrup Basin? Thus one critical assumption for any tectonic model of the eastern Canadian Arctic is how close Greenland was to Labrador in Early Cretaceous time.

Kerr (1981) took a middle course between the conflicting choices presented above by minimizing the closure of the Labrador Sea. This interpretation was indirectly supported by Grant (1980) who argued that the oceanic crust in the Labrador Sea is much narrower than Srivastava (1978) proposed. Grant's interpretation makes assumptions about the formation of unconformities with which I cannot agree. However, one of his main points was that the existence of the Labrador unconformity implied continental crust underneath. That conclusion is no longer necessarily valid if an Early Cretaceous rifting phase occurred.

For this paper, I have assumed that Srivastava's (1978) model for the Labrador Sea is essentially correct, although I disagree with some critical details, particularly the time of initial rifting. Kristoffersen & Talwani (1977) and Sclater et al. (1977) arrived at nearly the same answer as Srivastava using different data sets. Also Tapscott (1979) has done a detailed statistical analysis of the Greenland/Rockall/North America plate

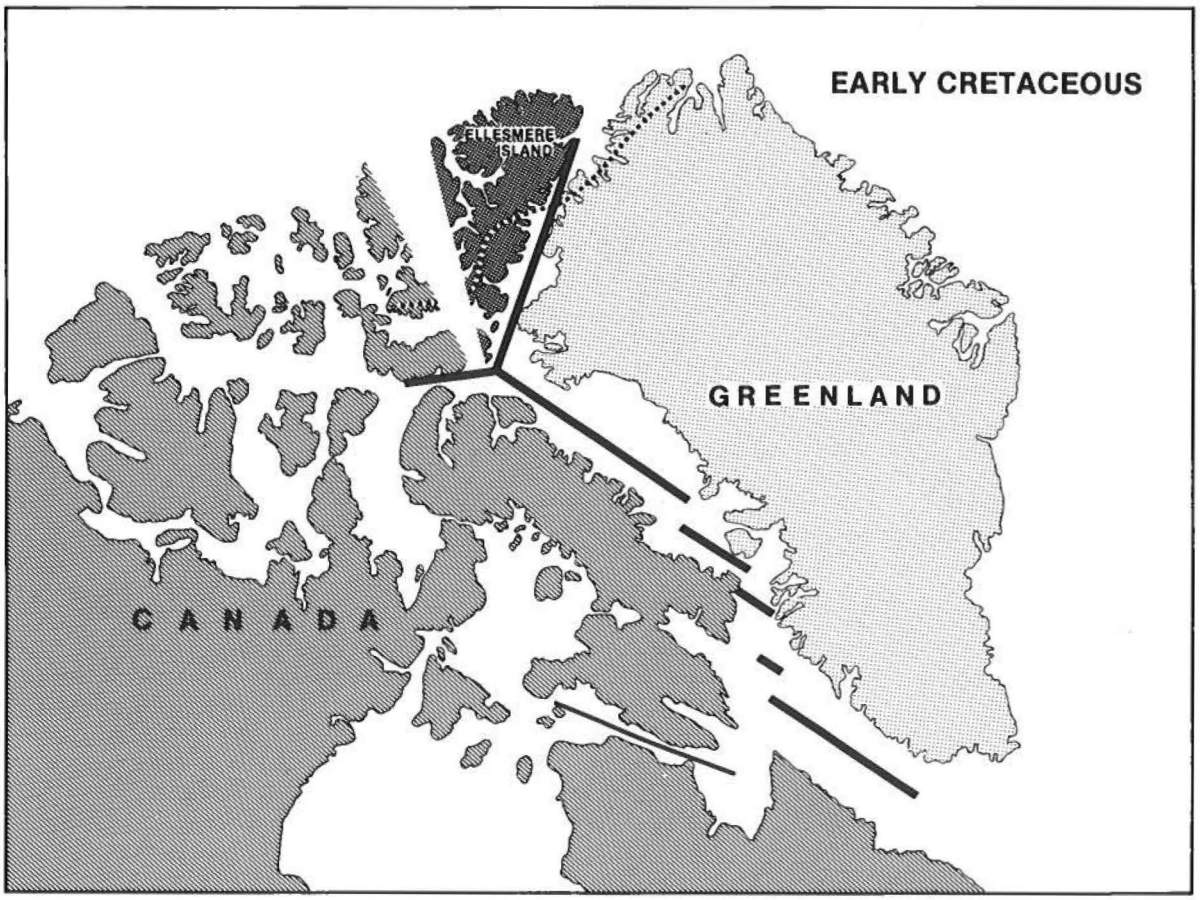


Fig. 2. Initial positioning of the plates around the Labrador Sea. Greenland's position is that given by Srivastava (1978). Ellesmere Island is assumed to be part of Greenland. Baffin Island has been closed up against Ungava. The heavy lines indicate the incipient rifting pattern which developed in the Early Cretaceous; thin line indicates extension in Hudson Strait; dotted line indicates the southern edge of the Palaeozoic fold belt. The open wedge west of Ellesmere Island shows diagrammatically the amount of extra space in the Sverdrup Basin predicted for this time.

system which supports some of Srivastava's poles. To my mind, these independently similar solutions give considerable credence to the validity of the sea-floor spreading model in this area.

Having assumed an initial position for Greenland closely juxtaposed to Labrador, the next critical assumption is where Ellesmere Island lay initially. In the absence of unambiguously definitive criteria, I have chosen to reassemble the Precambrian cratonic pieces. This puts Greenland against Baffin Island with Ellesmere Island rotated clockwise up against northern Greenland. This position eliminates the need to destroy material between Ellesmere and Greenland in some subsequent event, but it does imply that the Sverdrup Basin was much wider in the early Mesozoic. This is contrary to present interpretations of the geology of the eastern Sverdrup Basin, and will be discussed further below.

The position chosen (Fig. 2) for Ellesmere Island requires about 200 km of left-lateral offset along the Wegener Fault since the end of the Jurassic. An initial position with only 25 km offset could have been chosen if the Bullard et al. (1965) initial fit position for Greenland had been used, but Keen & Peirce (this volume) have shown that such a choice cannot be carried forward in time without creating serious overlaps of Devon and Ellesmere Islands and creating apparently unacceptable compression in Baffin Bay. Thus this reconstruction (Fig. 2) appears to be at odds with the geological correlations across Nares Strait summarized by Dawes & Kerr (this volume). The implication is that either an acceptable alternative model be found for the Labrador Sea or an alternative explanation be found for the geology of Nares Strait. The concept of a fault zone rather than a discrete offset in Nares Strait (Keen & Peirce, this volume) is a promising perspective. Another alternative is that the motion has been accommodated elsewhere, but no satisfactory interpretation has yet been offered.

A further, more minor assumption has been made in the initial position shown in Fig. 2. The overlap between Baffin Island and Greenland can be largely eliminated by closing Hudson Strait about a pole in the Gulf of Boothia. Although no direct evidence supports this ad hoc rotation, there exists a narrow channel striking ENE–WSW between Baffin Island and the islands off the northern end of the Ungava Peninsula which limits any possible motion of Baffin Island and makes this hypothesis quite plausible. If allowance is made for stretching and thinning of the continental margins as well, then the overlap problem in Davis Strait can be completely eliminated even if no oceanic crust exists there as Menzies (this volume) has suggested.

Incipient rifting phase

The principal clues for the timing of the incipient rifting phase are: 1) the stratigraphy of the Eclipse Trough, Bylot Island; 2) the age of the Alexis volcanics off the coast of Labrador; and 3) the Labrador unconformity. All of these indicate that incipient rifting began in the Early Cretaceous, although significant extension did not occur until much later.

In the Eclipse Trough, the base of the visible section rests on the Labrador unconformity. It is not known if a significant sedimentary section lies below the Labrador unconformity within the trough. These sediments are tentatively correlated with the Bjarni Formation (Umpleby 1979) further south (which lies on the Alexis volcanics) by McWhae (1981) and Miall et al. (1980). This evidence suggests that sufficient extension must have occurred by Middle Cretaceous time to allow graben formation.

The Alexis volcanics have been drilled at several locations off Labrador. Their petrological and magnetic characteristics are intermediate between continental and oceanic affinities and are consistent with an early rift phase tectonic setting (Srivastava et al. 1977, McWhae et al. 1981). The most reliable K/Ar ages for these volcanics are between 118 and 122 m.y. (McWhae et al. 1981).

Although the causal relationship between marine unconformities and tectonic activity is far from clear, the presence of the Labrador unconformity in a non-marine section as far north as Baffin Bay strongly suggests a tectonic component in its formation (Miall et al. 1980). The same unconformity is closely associated with the Alexis volcanics in Labrador where the overlying Bjarni Formation becomes at times more marine in character. McWhae (1981) estimated the time interval represented by the Labrador unconformity as 133–120 m.y.

Burke & Dewey (1973) speculated that a triple junction once existed in northern Baffin Bay. I suggest that this three ridge (RRR) triple junction was part of the incipient rifting phase. Lineaments were established through Lancaster Sound (perhaps as far as the Parry Channel) and through Nares Strait. Graben structures developed along these lineaments. The Paleocene volcanic sands at Judge Daly Promontory (Miall 1981) suggest that some Cretaceous volcanism may have occurred in Nares Strait and, by analogy, in Lancaster Sound as well. However volcanism in both these areas must have been limited in volume or there would be more evidence of it present today. This suggests that very limited extension occurred, consistent with the concept of an unsuccessful rifting phase. It is interesting to note that the present orientations of the three limbs of the triple junction are not symmetrical. However, rotation of Greenland and Ellesmere Island back against North America puts the three limbs of the triple junction nearly 120 degrees apart (compare Figs 2 and 4).

The concept of a long period of incipient rifting has parallels in the early opening history of the North Atlantic (e.g. the Triassic volcanics of New England and the Maritimes) and in the present-day tectonic style of the East African rift valley. It has been recognized in this area by both McWhae (1981) and Kerr (1981). Furthermore Sclater et al. (1977) treat Greenland as a separate plate as early as 95 m.y. ago in their analysis of North Atlantic plate motions, although its early motion was minor.

The incipient rifting phase persisted until the end of the Cretaceous. The presence of the Avalon unconformity (Jansa & Wade 1975) as a minor marker in the northern Labrador Sea during this time seems to be related to events farther south and east such as the opening of the Bay of Biscay and the development of Rockall Trough and does not appear to have a local tectonic component. At the end of the Cretaceous the incipient rifting phase was superseded by a more successful pattern of rifting, dubbed the Eurekan rifting episode by Kerr (1967, 1981).

Early phase of the Eurekan orogeny

The Eurekan rifting episode began in the Maastrichtian about 70 m.y. ago (anomaly 32 time, Srivastava 1978,

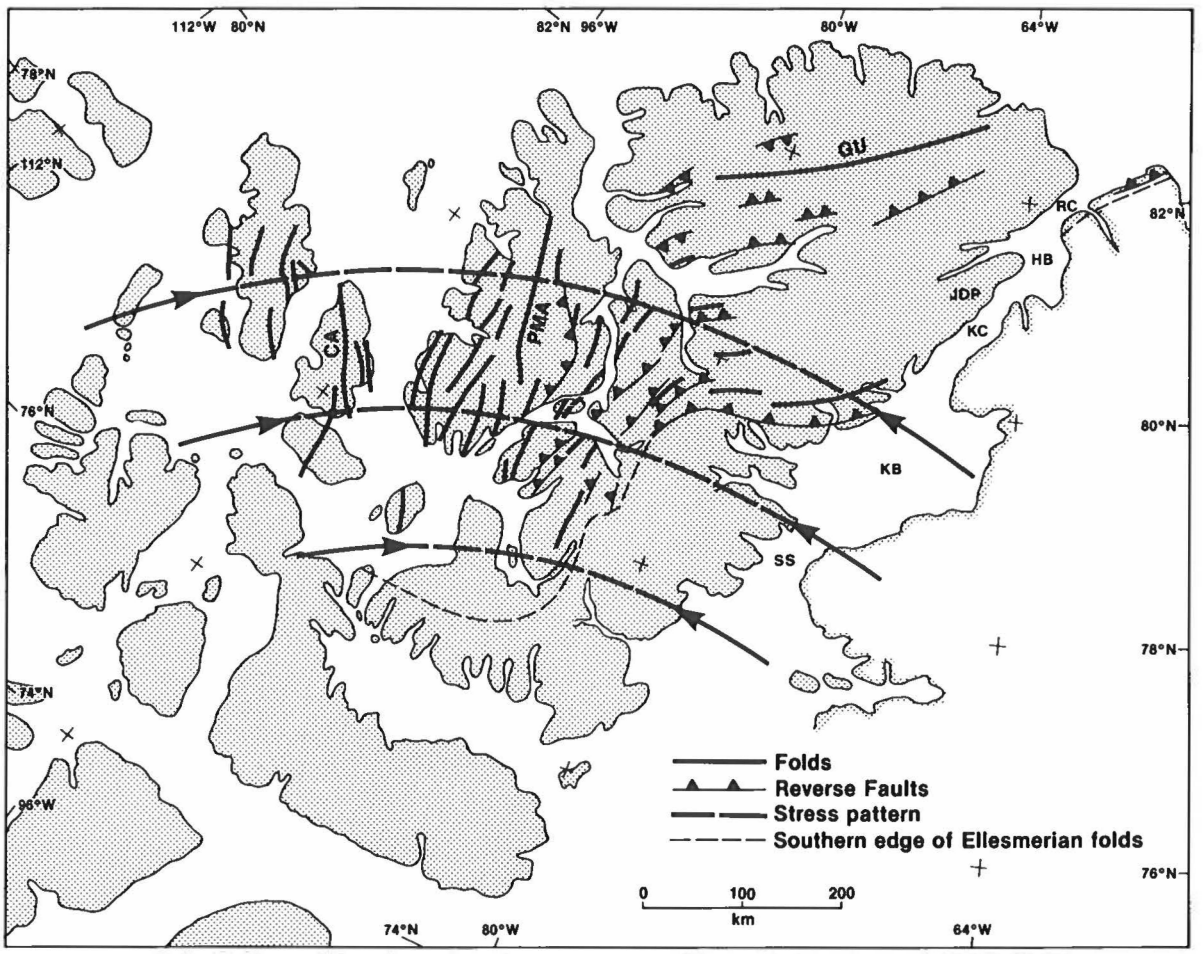


Fig. 3. Predicted stress pattern in the eastern Sverdrup Basin during the initial opening of the Labrador Sea. Eurekan folds and thrust faults are shown. No attempt has been made to sort out early Eurekan structures from those formed during the main phase of the orogeny as most structures probably moved at both times. The NE–SW structures on northern Ellesmere Island appear to be Palaeozoic lineaments reactivated during the main phase of the Eurekan orogeny. The structural data are from King (1969), Balkwill (1978) and McWhae (1981); base map is from King (1969). CA = Cornwall Arch, PMA = Princess Margaret Arch, GU = Grantland Uplift, JDP = Judge Daly Promontory, RC = Robeson Channel, HB = Hall Basin, KC = Kennedy Channel, KB = Kane Basin, SS = Smith Sound.

modified using the time scale of LaBrecque et al. 1977). Greenland and Ellesmere (and the Nares Strait lineament) rotated in a counter-clockwise sense away from the North American coast. As the pole of rotation was near northern Baffin Island (Sclater et al. 1977, Srivastava 1978), compression and uplift occurred north of the pole of rotation. The compressive stress pattern followed small circles about the pole of rotation. As folds were produced normal to this stress pattern, the fold pattern produced was a fan converging towards the pole of rotation. Fig. 3 shows the predicted stress pattern and the orientation of the Eurekan folds in the Sverdrup Basin. The folds were restricted to the less competent rocks of the Sverdrup Basin and did not extend into the Precambrian terrain. Some strike-slip faulting parallel to the predicted stress pattern is shown in McWhae's (1981) map near the southern margin of the basin.

Using the rotation poles given by Srivastava (1978), the amount of compressive strain predicted for the centre of the Sverdrup Basin (approximately 900 km north of the pole of rotation) is 200 km. This amounts to crustal shortening of nearly 25 per cent. During the early phase of the Eurekan orogeny, major arches such as the Cornwall Arch and Princess Margaret Arch did develop. Balkwill (1978) argued that these early Eurekan structures were caused by uplift with little compression. In contrast, this interpretation suggests that compression was the primary cause of the early phase of the Eurekan orogeny. Also a strong unconformity below the Eureka Sound Formation on the northern rim of the basin indicates a major pulse of tectonic activity in the Sverdrup Basin at this time (see fig. 8 in Meneley et al. 1975). There is also some evidence from deep seismic refraction work that a wave-like pattern with a 10 km amplitude and a N–S orientation has developed on the Mohorovičíc discontinuity underneath the Sverdrup Basin (Forsyth et al. 1979).

Taken together, the known features which may represent crustal compression can only account for 5–10 per cent shortening. Accounting for the remainder is a major unsolved problem. There is an interesting analogy to be drawn between this problem and a similar one regarding compression in the southern Appalachians of the United States. Were the basement rocks underlying the sediments involved in the deformation (thick-skinned deformation) or were the sediments decoupled from basement rocks as they deformed (thin-skinned deformation) (Rodgers 1970)? Geological interpretations of the southern Appalachians have indirectly suggested thrusting of less than 100 km (e.g. map of Williams 1978), but recent geophysical evidence from the COCORP program suggests thin-skinned deformation with at least 260 km of thrusting (Cook et al. 1979). If similar processes operated in the Sverdrup Basin, attaining compression of up to 25 per cent in the sedimentary section would not be a problem. Such a hypothesis provides a possible answer to arguments against major compression which are based on surface geology, but how and if the deeper crust was shortened remains unanswered.

During this time the Baffin Bay area was very close to the pole of rotation. The stress pattern was compressive in the north and mildly tensional in the south. In the north, the Bylot unconformity can be explained by uplift and erosion. In the south, and particularly further south in Labrador, there is no tectonic explanation for it. As a similiar unconformity has been noted in the North Sea (Kent 1981) and observed on a world-wide basis (Vail et al. 1977), it appears that the primary cause of the Bylot unconformity was a eustatic lowering of sea level and its effect was amplified by tectonic events in northern Baffin Bay and the Sverdrup Basin.

Opening of Baffin Bay

Between the time of anomalies 25 and 21 (59–50 m.y.), the pole of rotation for Greenland with respect to North America moved from Baffin Island to equatorial Africa (13°N, 2°E, Srivastava 1978). Although the magnetic anomalies in Baffin Bay are difficult to correlate at best (Jackson et al. 1979), the history of these changes is clearly recorded in the Labrador Sea (Srivastava 1978). At the same time spreading in the Atlantic migrated northwards between Greenland and Europe (Kristoffersen & Talwani 1977, Talwani & Eldholm 1977) and into the Arctic Ocean (Vogt et al. 1979). By anomaly 24 time the Wegener Transform Fault and its equivalent on the northeast side of Greenland, the Nansen Transform Fault, were established (Peirce 1980), and Greenland was separated from the surrounding plates.

The volcanics near Davis Strait are clearly related in age (dated at 58–60 m.y. by Parrott & Reynolds 1975) to the early opening of Baffin Bay. There is debate about whether or not they are related to a hot spot or mantle plume (Morgan 1972). Hall (1981) includes them in the Thulean volcanic line and relates them to the Icelandic hot spot. However, either such a hot spot must have been 1000 km in diameter, more than five times the commonly accepted scale of hot spots, or there must have been considerable lateral transport similar to that suggested by Morgan (1978) for some other hot spots. Moreover, Clarke (1977) argues convincingly that the Davis Strait volcanics are geochemically distinct from the others in the Thulean line. If they are not part of the Thulean volcanics, then the period of eruption was too short to qualify as a hot spot. I agree with Clarke (1977) that there was no Davis Strait hot spot and that these volcanics were erupted in response to the changing stress pattern as spreading migrated north into Baffin Bay.

At anomaly 25 time (Fig. 4), Ellesmere Island became part of the North American plate and was separated from Greenland by the Wegener Transform Fault.

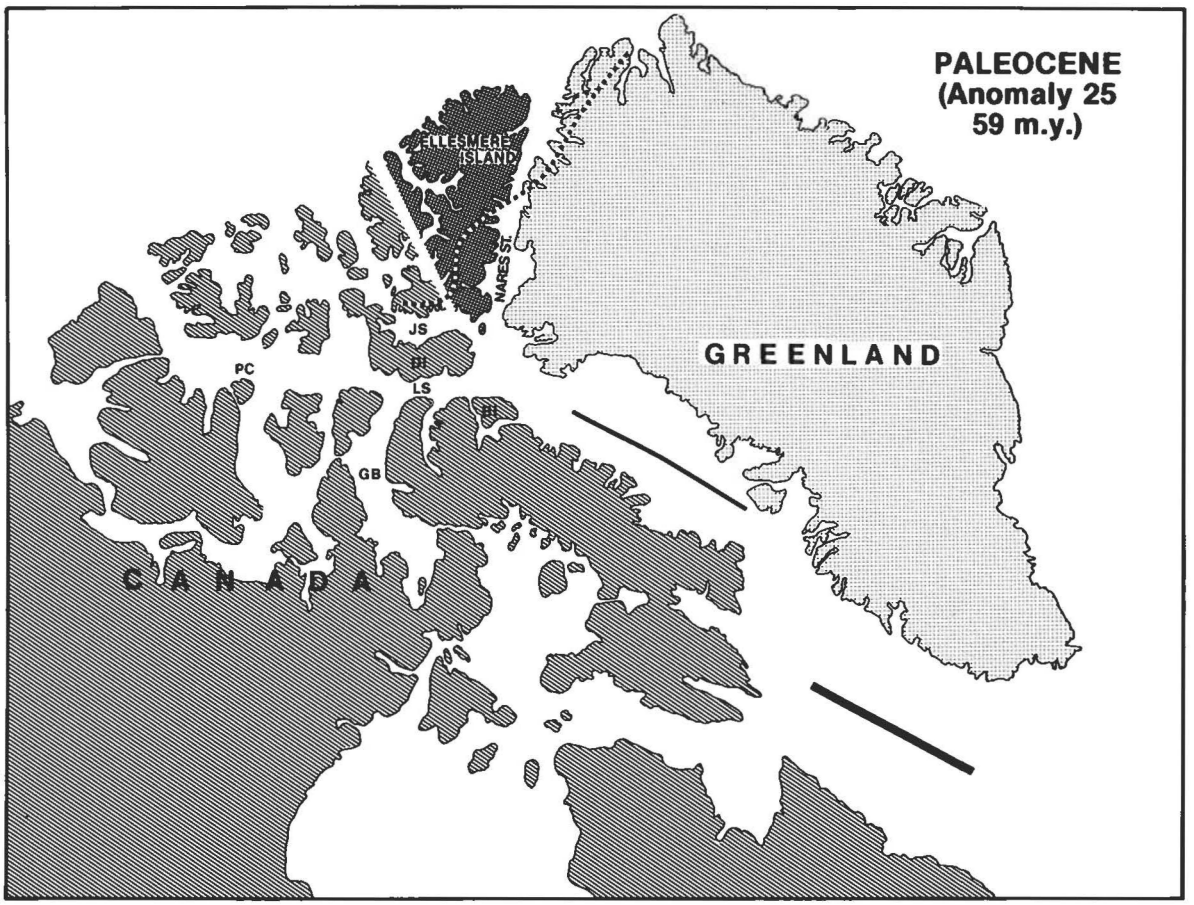


Fig. 4. Positioning of the Labrador Sea plates at anomaly 25 time (Paleocene). Note closure in the Sverdrup Basin since the Early Cretaceous. The heavy line indicates the Labrador Sea spreading centre; thin line indicates extension in Baffin Bay at this time. PC = Parry Channel, DI = Devon Island, LS = Lancaster Sound, JS = Jones Sound, BI = Bylot Island, GB = Gulf of Boothia.

According to the poles derived by Srivastava (1978), about 200 km of left-lateral offset occurred during the Late Paleocene and Early Eocene (anomalies 25–21) and 120 km of offset occurred during the Middle Eocene to Early Oligocene (anomalies 21–13, Fig. 5). The total offset is 320 km, not 250 km as Srivastava says in the text. About a third of this distance was accommodated in Baffin Bay by stretching and attenuation of the continental margins (Keen & Peirce, this volume). A similar amount of stretching may have occurred between North America and Ellesmere Island and contributed to the widening of grabens such as Lancaster and Jones Sounds.

Lancaster Sound was by now a failed arm rift. The other two limbs of the original RRR triple junction had become active plate boundaries while Lancaster Sound subsided and received a huge influx of Tertiary sediments.

Main phase of Eurekan compression

McWhae (1981) has suggested that a triple junction existed at the mouth of Jones Sound during the Eocene. This triple junction is required to explain his hypothesis that Greenland moved northwestwards during the Eocene causing the main phase of the Eurekan orogeny.

The Eocene triple junction is rejected on two counts. First, the timing is wrong. Balkwill (1978) showed that the Eurekan orogeny extended into post mid-Eocene time because rocks of that age were deformed by it. Secondly, the geometry is impossible. Had such a triple junction existed, it must have been a ridge-fault-fault (RFF) junction with the Baffin Bay spreading centre, the Wegener Transform Fault, and a Jones Sound Transform Fault as its limbs. The vector and stability diagrams (Fig. 6) demonstrate that such a triple junction could not have maintained itself unless the spreading was either very asymmetric or very oblique to the strike of the spreading centre.

Kerr (1980b, 1981) has suggested that a quadruple

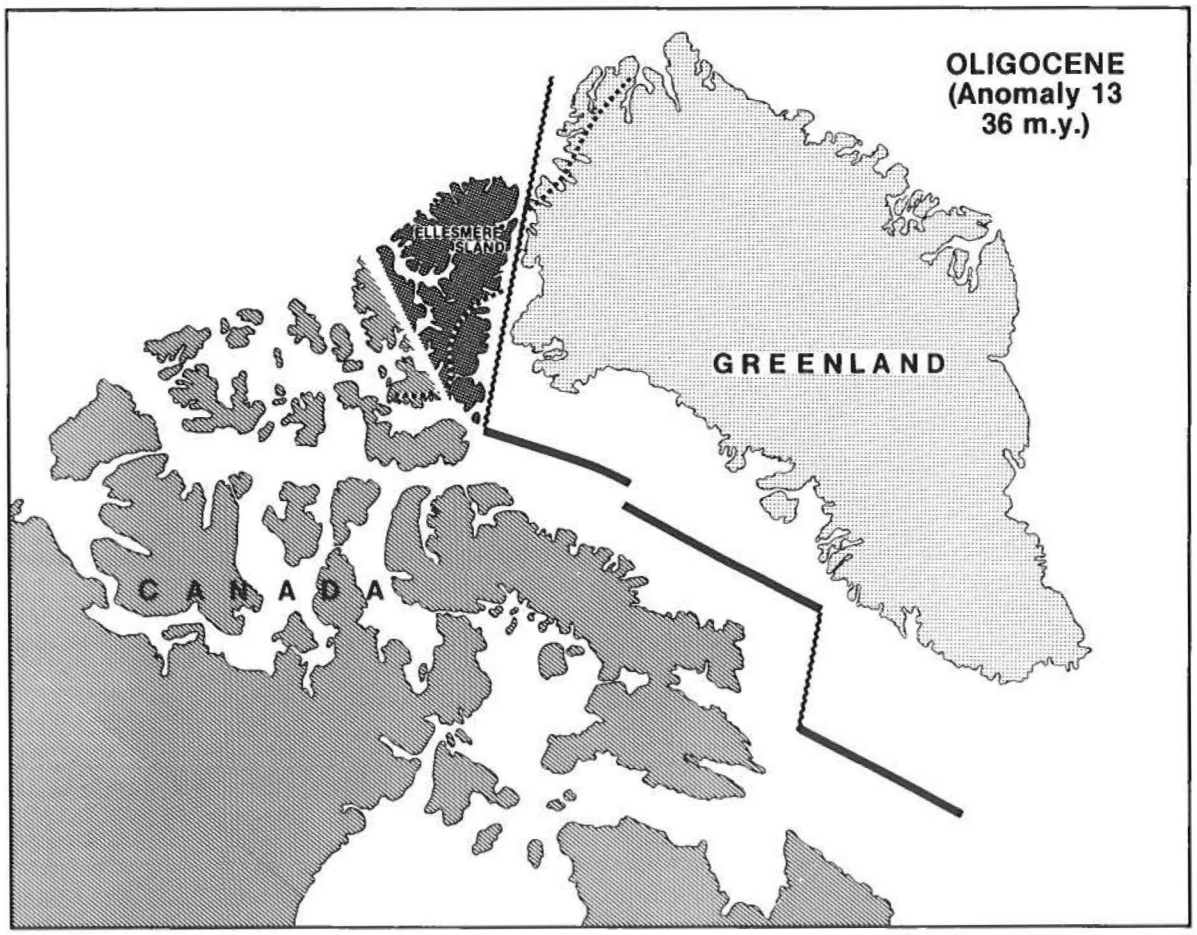


Fig. 5. Positioning of the Labrador Sea plates at anomaly 13 time (Oligocene). The heavy lines indicate the spreading centres which ceased to be active at this time. There probably was no discrete spreading centre which extended through Davis Strait. The wavy lines indicate transform faults. Note how the southern edge of the Palaeozoic fold belt has been apparently offset.

junction existed at this location during the opening of Baffin Bay. As a quadruple junction is inherently unstable (it will degenerate into two triple junctions, McKenzie & Morgan 1969), this hypothesis is also rejected.

However, McWhae's (1981) hypothesis can be modified to fit the constraints if the northwesterly motion of Greenland is delayed until Early Oligocene time (anomaly 13, 36 m.y.) when spreading had ceased in the Labrador Sea (Kristoffersen & Talwani 1977, Sclater et al. 1977, Srivastava 1978). I hypothesize that as the Wegener Transform Fault locked, Greenland moved northwestwards, relative to North America, about 50 km. There is some evidence for motion in this sense in Davis Strait in the form of possible fault offsets and folds with keystone structures associated with faulting (fig. 13 in McWhae 1981). Ellesmere Island was pushed ahead of Greenland producing the main phase of the Eurekan orogeny. Presumably the folding did not occur at Nares Strait because the crust on either side was relatively competent. Instead the stress was transmitted across eastern Ellesmere Island to the less competent rocks of the Sverdrup Basin where the strain could be accommodated. This stress reactivated early Eurekan structures in the Sverdrup Basin and Palaeozoic structures in northern Ellesmere Island. Similar variations in the style of deformation have been observed in the Himalayan collision and have been interpreted to be related to crustal age and strength (Molnar & Tapponnier 1981).

Srivastava (1978) breaks the motion of Greenland relative to North America between anomaly 24 and the Present into two stages of motion (see arrows on his fig. 21d) which ended at anomaly 13 time. Although the generally accepted strike of the Wegener Transform Fault is parallel to Nares Strait, the last pole of rotation postulated by Srivastava (1978) predicts a strike 25° different from this (see his fig. 21d). A careful examination of the data (particularly his fig. 2) shows that there is good control on the amount of rotation during this time, but there is poor control on the azimuth to the

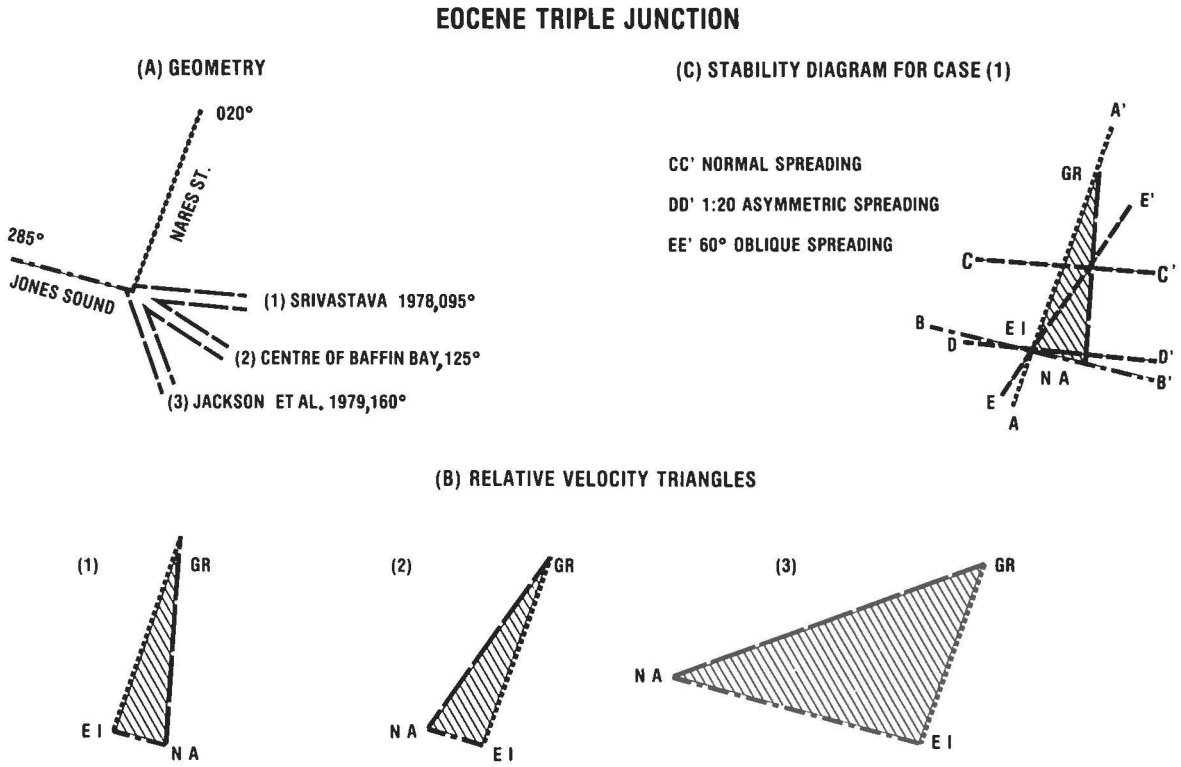


Fig. 6. Vector and stability diagram (McKenzie & Morgan 1969) for the Eocene triple junction in Baffin Bay hypothesized by McWhae (1981). NA = North America, EI = Ellesmere Island, GR = Greenland. Orientations of EI/NA and GR/EI vectors based on strikes of Jones Sound and Nares Strait. GR/NA spreading assumed perpendicular to Baffin Bay spreading axis. Relative velocities unknown. The lines of no motion in the stability diagram are AA', BB' and CC'. For the triple junction geometry to be stable these must intersect in a point. This junction has an unstable geometry unless the spreading is 60° oblique to the spreading axis which would be oriented along line EE', or spreading was asymmetric in a 1:20 ratio, being faster on the eastern side (DD').

pole of rotation because the fracture zones are poorly defined. The first well-defined pole is the total pole of rotation for anomaly 25 to the present. This pole does predict the strike of Nares Strait (Fig. 7). There remains, however, a discrepancy between that pole and that estimated from the data in the North Atlantic by Sclater et al. (1977) for the same motion. If we accept Srivastava's total pole as the best estimate of motion between Greenland and North America between anomaly 25 and anomaly 13, when spreading ceased in the Labrador Sea, then the combination of that pole with a rotation of +1.5° about a pole at 15°N, 130°W produces the overall pole computed by Sclater et al. (1977). This new combination of poles predicts the strike of Nares Strait, moves Greenland to the northwest relative to North America, and explains the 50 km of compression in Ellesmere and Axel Heiberg Islands during the main phase of the Eurekan orogeny.

Kristoffersen & Talwani (1977) did suggest northwesterly motion of Greenland, but they were basing their conclusions only on magnetic anomalies near the triple junction and they had no Labrador Sea anomalies to work with. Sclater et al. (1977) considered the fit of all the North Atlantic pieces of the puzzle without the benefit of Srivastava's (1978) correlations in the Labrador Sea. Their poles of opening are in excellent agreement with those of Srivastava (1978), except that their pole for the opening of the Labrador Sea between anomalies 13 and 21 is significantly further east than Srivastava's pole for the same period. The addition of a small amount of northwesterly motion of Greenland as suggested above reconciles the three results very well.

This model predicts some left-lateral motion in the vicinity of Jones Sound, as indicated on McWhae's (1981) maps, but the timing is later. The amount of motion is based primarily on Balkwill's (1978) and McWhae's estimates of crustal shortening during the main phase of the Eurekan orogeny. There is no direct evidence in the magnetic anomaly correlations for the 50 km move of Greenland. However, as the motion would have been nearly parallel to the Labrador Sea anomalies, and as there are no sharply-defined fracture zones to be offset, such motion could have been easily overlooked. On the other hand, it is unlikely that more than about 50 km of translation would go undetected.

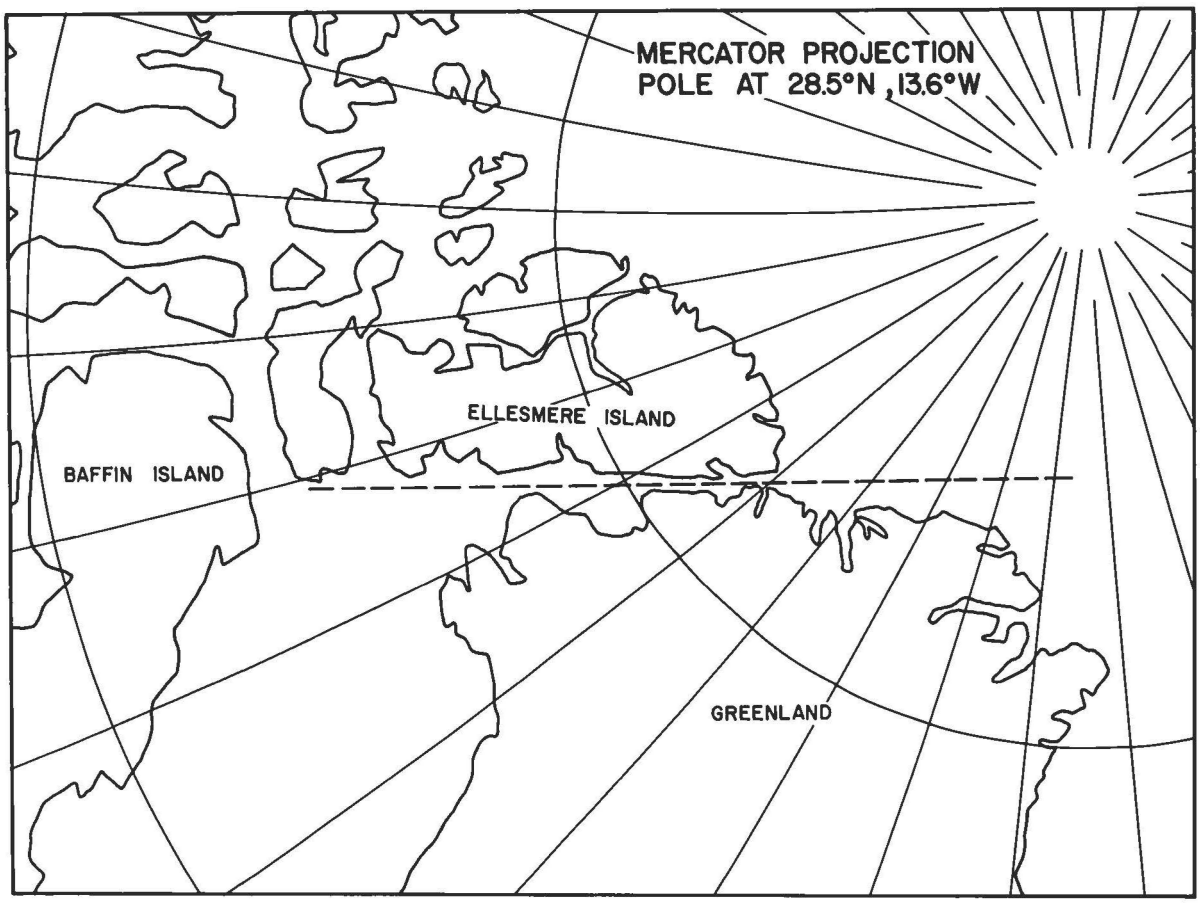


Fig. 7. Mercator projection of Nares Strait area about pole of rotation at 28.5°N, 13.6°W, which is Srivastava's (1978) total pole of rotation for anomaly 25 to the present. Note that Nares Strait appears as a parallel of 'latitude' relative to this pole and is therefore a small circle about this pole. Major transform faults form small circles about their poles of rotation. The strike predicted by Srivastava's (1978) pole for his last direction of spreading (anomaly 21 to 13) is about 25° more northeasterly than the strike of Nares Strait (see case (1), Fig. 6).

Sedimentary section in Lancaster Sound and Nares Strait

The tectonic time diagram in Fig. 8 outlines the changing tectonic styles of each area and schematically shows the extent of the major unconformities.

The incipient rifting probably migrated northwards up the Labrador coast. Therefore, I expect the Labrador unconformity to be diachronous, being younger to the north. When the triple junction was established in northern Baffin Bay in the Early–Middle Cretaceous, the structural lineaments in Nares Strait and Lancaster Sound were established. These features resembled the East African Rift Valley of today — extensional grabens without any oceanic crust developed in the centre and surrounded by uplifted highlands. The oldest sedimentary rocks in both areas are likely to be Palaeozoic blocks preserved within the grabens (Daae & Rutgers 1975, Kerr 1980b). Above these there should be continental equivalents of the Bjarni Formation similiar to those found in the Eclipse Trough. If the highlands were dipping away from the central grabens, the sediment supply may have been limited. On the other hand, the grabens may have acted as the major line of drainage for the area concentrating sedimentation within it. The palaeocurrent evidence cited by Miall et al. (1980) suggests that the Eclipse Trough was a tributary feeding north into the more major lineament of Lancaster Sound. There is also the possibility of volcanics associated with the rifting occurring anywhere in the Early–Middle Cretaceous section. The volcanic sands at Judge Daly Promontory (Miall 1981, see Fig. 1) might be derived from such Cretaceous volcanics.

The incipient rifting phase probably persisted until Late Cretaceous, although it is unlikely that any large amount of subsidence occurred to allow accumulation of a thick section. With the onset of sea-floor spreading in the Labrador Sea, both Lancaster Sound and Nares

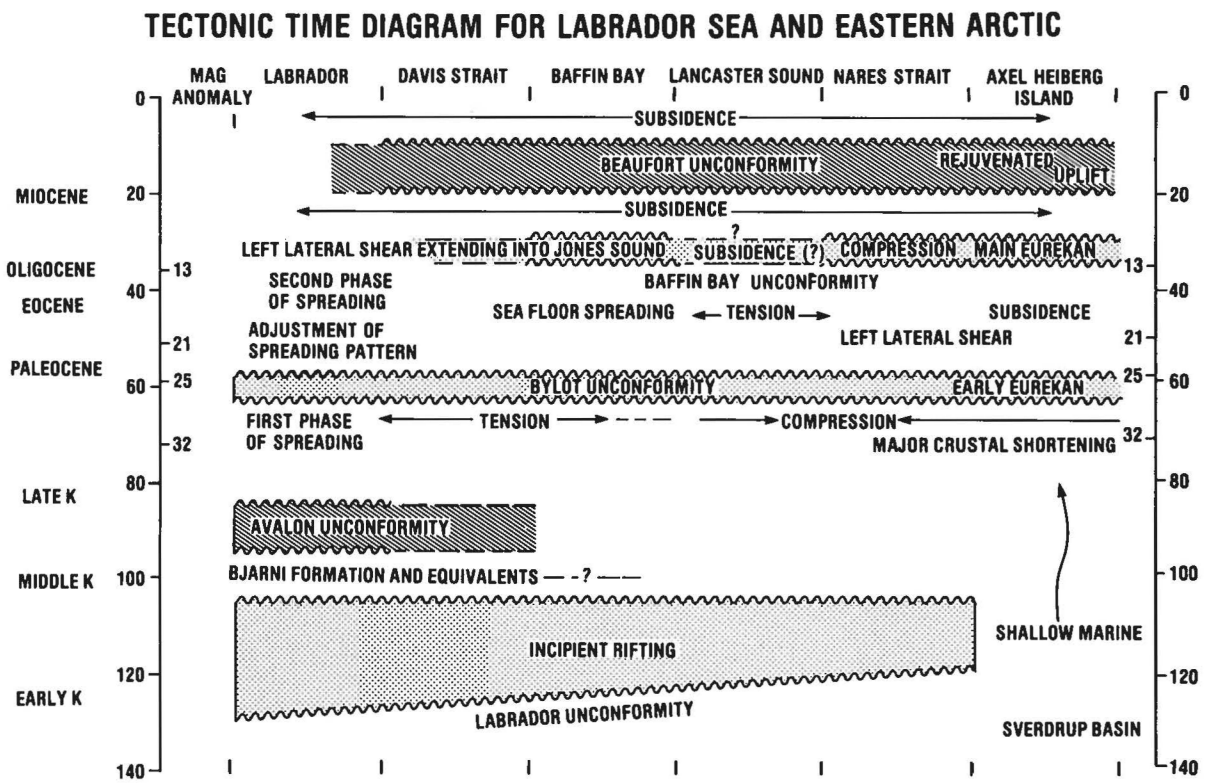


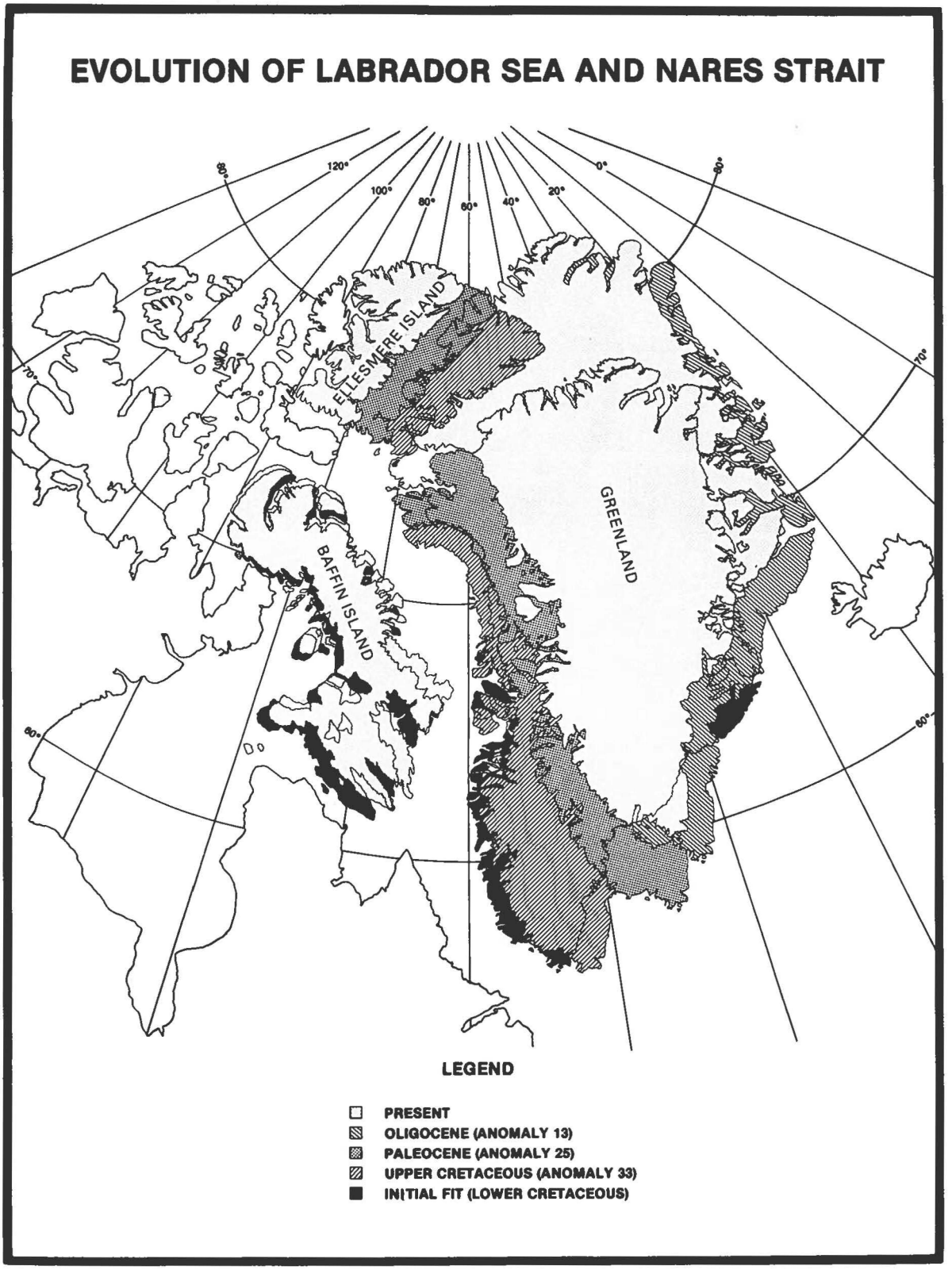
Fig. 8. Tectonic time diagram for the eastern Canadian Arctic and Labrador Sea. The major unconformities are informally named after McWhae (1981). The three unconformities where local tectonics may have enhanced their development are shaded with dots. The two unconformities without a local tectonic component are shaded with diagonal lines. All five unconformities occur in many parts of the world, indicating that eustatic changes of sea level may have contributed to their formation (Vail et al. 1977). The validity of these eustatic sea level changes is a matter of lively debate. The ages of the magnetic anomalies are based on the time scale of LaBrecque et al. (1977). K indicates Cretaceous.

Strait were placed in a compressive stress regime where the amount of compression was proportional to the distance from the pole of rotation. Therefore both these lineaments should have experienced more compression, and probably more erosion, during the time represented by the Bylot unconformity than did the Eclipse Trough. Lancaster Sound was close enough to the pole of rotation that the amount of compression was probably minimal — perhaps an absence of subsidence would be a more accurate description.

When the spreading pattern changed in the Labrador Sea and Baffin Bay began to open, Ellesmere Island became detached from the Greenland plate. Compression in Lancaster Sound relaxed and there was tensional stress to widen the structure and accelerate subsidence. As the other two limbs of the original triple junction were still active plate margins, Lancaster Sound became a failed rift. Subsidence and sedimentation have been rapid since then, except for a brief hiatus during the Baffin Bay unconformity and a possible hiatus during the Miocene Beaufort unconformity.

Lancaster Sound has at least six kilometres of sedimentary section in it (Daae & Rutgers 1975, Kerr 1980b). McWhae (1981) has argued that the bulk of the section in Lancaster Sound is Cretaceous in age based on an attempt to correlate the Bylot unconformity into Lancaster Sound on seismic sections and by analogy with the Eclipse Trough. However, the structural his-

Fig. 9. Summary diagram showing the major stages in the evolution of the Labrador Sea and Nares Strait. The complete outlines of Greenland and Baffin Island are shown in their initial and present-day positions. Intermediate positions shown where they are not covered by a later position. The initial fit position from Bullard et al. (1965) is shown for reference, but Keen & Peirce (this volume) have suggested that Srivastava's Upper Cretaceous position is a more likely initial fit geologically. The motion of Greenland is generally to the northeast and then to the northwest. The Oligocene position of Greenland is explained in the text. The Upper Cretaceous and Paleocene positions are from Srivastava (1978). The initial position of Ellesmere Island is not known but the model predicts that it was near the Upper Cretaceous position shown. The Upper Cretaceous position assumes that Ellesmere Island and Greenland moved as one until the Paleocene. The Oligocene and Paleocene positions of Ellesmere Island are the same and are explained in the text. The motion of Ellesmere Island is generally to the northwest.


EVOLUTION OF LABRADOR SEA AND NARES STRAIT
ELLESMERE ISLAND
GREENLAND
BAFFIN ISLAND
LEGEND
PRESENT
OLIGOCENE (ANOMALY 13)
PALEOCENE (ANOMALY 25)
UPPER CRETACEOUS (ANOMALY 33)
INITIAL FIT (LOWER CRETACEOUS)

tory outlined above suggests that it is unlikely that rapid subsidence and sedimentation could have begun before the Paleocene. Therefore, an alternative interpretation is that the majority of the Lancaster Sound section is Tertiary in age. In short, I suggest that the Bylot unconformity may be considerably deeper than McWhae (1981) has suggested. Kerr (1980b) arrives at the same conclusion, although his interpretation of the tectonic history is different.

Nares Strait has had a more complicated history than Lancaster Sound. Until the Paleocene their histories were quite similar and continental equivalents of the Bjarni Formation should be expected at the bottom of the Nares section as well. However, during opening of Baffin Bay, while Lancaster Sound was beginning to subside rapidly, the extensive left-lateral motion on the Wegener Transform Fault must have sheared and deformed these early sediments extensively. Furthermore, it is unlikely that there was an overall subsidence of Nares Strait while it was an active transform fault, although local deeps may have been developed in the fault zone.

During the main phase of the Eurekan orogeny, Nares Strait was under compressive stress while Lancaster Sound was relatively unaffected. It is likely that the Baffin Bay unconformity cut quite deeply into whatever section had accumulated in Nares Strait. As there were continued pulses of uplift in the eastern Sverdrup Basin into the Miocene (Balkwill 1978), it is unlikely that the compressive stresses there relaxed before then. Miall (1979) cites the absence of Paleocene sediments in the Lake Hazen intermontane basin on Ellesmere Island as evidence that subsidence in the area did not begin until the Eocene or Oligocene. Nares Strait probably did not begin to subside rapidly until the Neogene, well after Lancaster Sound began to subside. However, as the rate of subsidence decreases exponentially (under similar conditions of sediment loading), during the Miocene and Pliocene Nares Strait may have been subsiding more rapidly then than Lancaster Sound.

Conclusion

Both Nares Strait and Lancaster Sound are features which were localized along older lineaments established during an Early to Middle Cretaceous incipient rifting phase. Because northern Baffin Bay was in a compressive stress regime through the Late Cretaceous and especially in the Paleocene (early phase of Eurekan orogeny) it is unlikely that major subsidence could have occurred until the Late Paleocene. At this time Lancaster Sound became a true failed rift while Nares Strait became a major transform fault. Given the post mid-Eocene age of the main phase of the Eurekan orogeny, I have hypothesized that Greenland moved northwestwards about fifty kilometres after spreading ceased in the Labrador Sea. This additional motion of Greenland appears to make some of the key geological and geophysical pieces of the Nares Strait puzzle fit together, although the arguments about the correlation of geological trends across Nares Strait will undoubtedly continue. Fig. 9 shows these motions superimposed on one diagram summarizing the evolution of the Labrador Sea and Nares Strait.

Acknowledgements

As with most papers, the initial ideas for this paper were established by discussions with others. In this case J. R. H. McWhae and J. W. Kerr deserve the credit for getting me interested in the problem. I would like to acknowledge the Atlantic Geoscience Centre for providing me with their version of SUPERMAP, which was used for the plate reconstructions, and U. Kerrison and H. Desireau for solving the problems of getting it running on our facilities. My thanks also to R. A. Meneley, P. R. Dawes and an anonymous reviewer for valuable criticisms, to J. Pendergast for typing several drafts of the manuscript, and to Petro-Canada for allowing me to publish these ideas.

References

Balkwill, H. R. 1978. Evolution of Sverdrup Basin, Arctic Canada. – Bull. Am. Ass. Petrol. Geol. 62: 1004–1028.

Bullard, E., Everett, J. E. & Smith, A. G. 1965. The fit of the continents around the Atlantic. – In: Blackett, P. M. S., Bullard, E. & Runcorn, S. K., A symposium on continental drift. – Phil. Trans. Roy. Soc. Lond. 258A: 41–51.

Burke, K. & Dewey, J. F. 1973. Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks. – J. Geol. 81: 406–433.

Clarke, D. B. 1977. The Tertiary volcanic province of Baffin Bay. – In: Baragar, W. R. A., Coleman, L. C. & Hall, J. M. (eds), Volcanic regimes in Canada. – Spec. Pap. geol. Ass. Can. 16: 445–460.

Cook, F. A., Albaugh, D. S., Brown, L. D., Kaufman, S., Oliver, J. E. & Hatcher, R. D. 1979. Thin-skinned tectonics in the crystalline southern Appalachians; COCORP seismic-reflection profiling of the Blue Ridge and Piedmont. – Geology 7: 563–567.

Daae, H. D. & Rutgers, A. T. C. 1975. Geological history of the Northwest Passage. – Bull. Can. Petrol. Geol. 23: 84–108.

Dawes, P. R. 1973. The North Greenland fold belt: a clue to the history of the Arctic Ocean basin and the Nares Strait lineament. – In: Tarling, D. H. & Runcorn, S. K. (eds), Implications of continental drift to the earth sciences 2: 925–947. – Academic Press, London & New York.

Dawes, P. R. & Kerr, J. W. 1982. The case against major displacement along Nares Strait. – This volume.

Dawes, P. R., Frisch, T. & Christie, R. L. 1982. The Proterozoic Thule Basin of Greenland and Ellesmere Island: importance to the Nares Strait debate. – This volume.

England, J. 1982. Postglacial emergence along northern Nares Strait. – This volume.

Forsyth, D. A., Mair, J. A. & Fraser, I. 1979. Crustal structure of the central Sverdrup Basin. – Can. J. Earth Sci. 16: 1581–1598.

Frisch, T. & Dawes, P. R. 1982. The Precambrian Shield of northernmost Baffin Bay: correlation across Nares Strait. – This volume.

Grant, A. C. 1980. Problems with plate tectonics: the Labrador Sea. – Bull. Can. Petrol. Geol. 28: 252–278.

Hall, J. M. 1981. The Thulean volcanic line. – In: Kerr, J. W. & Fergusson, A. J. (eds), Geology of the North Atlantic borderlands. – Mem. Can. Soc. Petrol. Geol. 7: 231–244.

Higgins, A. K., Mayr, U. & Soper, N. J. 1982. Fold belts and metamorphic zones of northern Ellesmere Island and North Greenland. – This volume.

Hurst, J. M. & Kerr, J. W. 1982. Upper Ordovician and Silurian facies patterns in eastern Ellesmere Island and western North Greenland and their bearing on the Nares Strait lineament. – This volume.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor spreading in central Baffin Bay. – Can. J. Earth Sci. 16: 2122–2135.

Jansa, L. F. & Wade, J. A. 1975. Geology of the continental margin off Nova Scotia and Newfoundland. – In: van der Linden, W. J. M. & Wade, J. A. (eds), Offshore geology of eastern Canada. – Pap. geol. Surv. Can. 74–30: 51–105.

Keen, C. E. & Peirce, J. W. 1982. The geophysical implications of minimal Tertiary motion along Nares Strait. – This volume.

Kent, P. E. 1981. The history of the northeast Atlantic margin in a world setting. – In: Kerr, J. W. & Fergusson, A. J. (eds), Geology of the North Atlantic borderlands. – Mem. Can. Soc. Petrol. Geol. 7: 1–10.

Kerr, J. W. 1967. Nares submarine rift valley and the relative motion of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Kerr, J. W. 1980a. Did Greenland drift along Nares Strait? – Bull. Can. Petrol. Geol. 28: 279–289.

Kerr, J. W. 1980b. Structural framework of Lancaster Aulacogen, Arctic Canada. – Bull. geol. Surv. Can. 319: 24 pp.

Kerr, J. W. 1981. Evolution of the Canadian Arctic Islands: a transition between Atlantic and Arctic Oceans. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 105–199. – Plenum Press, New York & London.

King, P. B. 1969. Tectonic map of North America. 1:5 000 000. – U.S. Geol. Surv.

Kristoffersen, Y. & Talwani, M. 1977. Extinct triple junction south of Greenland and the Tertiary motion of Greenland relative to North America. – Bull. geol. Soc. Am. 88: 1037–1049.

LaBrecque, J. L., Kent, D. V. & Cande, S. C. 1977. Revised magnetic polarity time scale for Late Cretaceous and Cenozoic time. – Geology 5: 330–335.

McKenzie, D. & Morgan, W. J. 1969. Evolution of triple junctions. – Nature, Lond. 224: 125–133.

McWhae, J. R. H. 1981. Structure and spreading history of the northwestern Atlantic region from the Scotian Shelf to Baffin Bay. – In: Kerr, J. W. & Fergusson, A. J. (eds), Geology of the North Atlantic borderlands. – Mem. Can. Soc. Petrol. Geol. 7: 299–332.

McWhae, J. R. H. & Michel, W. F. E. 1975. Stratigraphy of Bjarni H–81 and Leif M–48, Labrador Shelf. – Bull. Can. Petrol. Geol. 23: 361–382.

McWhae, J. R. H., Elie, R., Laughton, K. C. & Gunther, P. R. 1981. Stratigraphy and petroleum prospects of the Labrador Shelf. – Bull. Can. Petrol. Geol. 28: 460–488.

Meneley, R. A., Henao, D. & Merritt, R. K. 1975. The northwest margin of the Sverdrup Basin. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 557–587.

Menzies, A. W. 1982. Crustal history and basin development of Baffin Bay. – This volume.

Miall, A. D. 1979. Tertiary fluvial sediments in the Lake Hazen intermontane basin, Ellesmere Island, Arctic Canada. – Pap. geol. Surv. Can. 79–9: 25 pp.

Miall, A. D. 1981. Late Cretaceous and Paleogene sedimentation and tectonics in the Canadian Arctic Islands. – In: Miall, A. D. (ed.), Sedimentation and tectonics in alluvial basins. – Spec. Pap. geol. Ass. Can. 23: 221–272.

Miall, A. D., Balkwill, H. R. & Hopkins, W. S. 1980. Cretaceous and Tertiary sediments of the Eclipse Trough, Bylot Island area, Arctic Canada, and their regional setting. – Pap. geol. Surv. Can. 79–23: 20 pp.

Molnar, P. & Tapponnier, P. 1981. A possible dependence of tectonic strength on the age of the crust in Asia. – Earth planet. Sci. Lett. 52: 107–114.

Morgan, W. J. 1972. Deep mantle convection plumes and plate motions. – Bull. Am. Ass. Petrol. Geol. 56: 203–213.

Morgan, W. J. 1978. Rodriguez, Darwin, Amsterdam,.., a second type of hotspot island. – J. geophys. Res. 83: 5355–5360.

Newman, P. H. 1977. The offshore and onshore geology and geophysics of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Newman, P. H. 1982. A geological case for movement between Canada and Greenland along Nares Strait. – This volume.

Parrott, P. J. E. & Reynolds, P. H. 1975. Argon-40/Argon-39 geochronology: age determinations of basalts from the Labrador Sea area. – Geol. Soc. Am. Abs. with Prog. 7: 835 only.

Peel, J. S. & Christie, R. L. 1982. Cambrian–Ordovician platform stratigraphy: correlations around Kane Basin. – This volume.

Peel, J. S., Dawes, P. R., Collinson, J. D. & Christie, R. L. 1982. Proterozoic – basal Cambrian stratigraphy across Nares Strait: correlation between Inglefield Land and Bache Peninsula. – This volume.

Peirce, J. W. 1980. The Nares-Nansen triple junction. – Trans. Am. geophys. Un. 61: 357 only.

Riddihough, R. P., Haines, G. V. & Hannaford, W. 1973. Regional magnetic anomalies of the Canadian Arctic. – Can. J. Earth Sci. 10: 157–163.

Rodgers, J. 1970. The tectonics of the Appalachians. – Wiley-Interscience, New York: 271 pp.

Sclater, J. G., Hellinger, S. & Tapscott, C. [T.] 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. – J. Geol. 85: 509–552.

Sobczak, L. W. 1982. Fragmentation of the Canadian Arctic Archipelago, Greenland, and surrounding oceans. – This volume.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Srivastava, S. P. & Falconer, R. K. H. 1982. Nares Strait: a conflict between plate tectonic predictions and geological interpretation. – This volume.

Srivastava, S. P., Falconer, R. K. H., Haworth, R. T., Peirce, J. W., Clark, M. J. & Clarke, D. B. 1977. Correlation between magnetic anomalies and the bedrocks in the offshore regions of eastern Canada. – Trans. Am. geophys. Un. 58: 747 only.

Talwani, M. & Eldholm, O. 1977. Evolution of the Norwegian-Greenland Sea. – Bull. geol. Soc. Am. 88: 969–999.

Tapscott, C. T. 1979. The evolution of the Indian Ocean triple junction and the finite rotation problem. – Unpubl. Ph. D. thesis, Mass. Inst. Tech./Woods Hole Ocean. Inst.: 210 pp.

Taylor, F. B. 1910. Bearing of the Tertiary mountain belt on the origin of the earth's plan. – Bull. geol. Soc. Am. 21: 179–226.

Umpleby, D. C. 1979. Geology of the Labrador Shelf. – Pap. geol. Surv. Can. 79–13: 34 pp.

Vail, P. R., Mitchum, R. M., Todd, R. G., Widmier, J. M., Thompson, S., Sangree, J. B., Bubb, J. N. & Hatlelid, W. G. 1977. Seismic stratigraphy and global changes of sea level. – In: Payton, C. E. (ed.), Seismic stratigraphy – applications to hydrocarbon exploration. – Mem. Am. Ass. Petrol. Geol. 26: 49–212.

Vogt, P. R., Taylor, P. T., Kovacs, L. C. & Johnson, G. L. 1979. Detailed aeromagnetic investigation of the Arctic Basin. – J. geophys. Res. 84: 1071–1089.

Wegener, A. 1924. The origin of continents and oceans. – Methuen & Co., London: 212 pp. (Transl. 3rd edit. by J. G. A. Skerl).

Wetmiller, R. J. & Forsyth, D. A. 1982. Review of seismicity and other geophysical data near Nares Strait. – This volume.

Williams, H. 1978. Tectonic-lithofacies map of the Appalachian orogen. 1:1 000 000. – Memorial Univ. of Newfoundland, Canada.

Geophysics

Marine geophysical study of southern Nares Strait

PATRICIA H. NEWMAN

Newman, P. H. 1982. Marine geophysical study of southern Nares Strait. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 255–260.

Bathymetry and gravity contour maps of southern Nares Strait (herein termed North Water Bay) based mainly on previously unpublished data, together with seismic reflection and magnetic data, indicate the preservation of two major sedimentary basins. These basins are elongate approximately parallel to major onshore faults. Both are filled with disturbed strata.

Kane Basin to the north is probably filled with five to ten km of gently northeast-dipping sediments, disturbed by minor northeast-trending faults.

Gravity modelling indicates a possible, fault-bounded, basalt-filled 'fracture' off southeastern Ellesmere Island. This structure may continue northeastward into Nares Strait as a sediment-filled trough. This indicates only that Nares Strait is a fault-bounded feature, and not necessarily that there has been any horizontal motion, major or otherwise, along it.

P. H. Newman, Department of Geology, Dalhousie University, Halifax, Nova Scotia, Canada B3H 3J5. Present address: Department of Geology, University of Auckland, Auckland, New Zealand.

Marine geophysical data provided by the Bedford Institute of Oceanography, Dartmouth, Nova Scotia were analysed in the area from 75°N to 80°N and 64°W to 84°W (Fig. 1), as part of the research for an M.Sc. thesis (Newman 1977). These data provide clues relating to the question of the existence of a transform fault through Nares Strait, the linear seaway separating Greenland and Ellesmere Island.

The main study area is defined to the south by a bathymetric sill separating it from Baffin Bay (Fig. 2), and to the north by the narrows of Smith Sound. For the purpose of this paper it is termed North Water Bay.

The main data base for the research analysis comprised cruises Baffin 70–021, Hudson 71–032, Hudson 74–026 and Hudson 76–023, though data from earlier cruises were reviewed and used for qualitative control.

Seismic data

Seismic coverage is shown in Fig. 3. Work in western North Water Bay and Kane Basin was restricted due to heavy ice. In general, the 1971 records are good (Keen & Barrett 1973), but those of 1974 and 1976 are poorer. A major problem with all the records is the water depth multiple. In the deep ocean waters, with depths of thousands of metres, this is not a serious problem, but in areas of relatively shallow depths any sub-bottom structures below a few hundred metres are obscured.

From magnetic data, Keen & Barrett (1973) and Hood & Bower (1975) proposed the existence of a single, deep sedimentary basin trending northwestward from Thule Air Base. However, seismic data indicate two major sedimentary basins, separated by a basement high (Ross & Falconer 1975), and these basins are here named the Steensby Basin and the North Water Basin (Fig. 7).

The thickness and somewhat disturbed nature of the sediments in the arcuate North Water Basin can be seen well in line 74–12 (Fig. 4). A syncline can be seen between 'a' and 'b', and 'c' and 'd' (Figs 3 and 4), as well as on several other lines. Farther south this feature becomes shallower and much more disturbed. Keen & Barrett (1973) infer shallow magnetic basement here, and thus the southern termination of the North Water Basin. To the north, the trace of the syncline is lost because of poor records.

The 1971 sonobuoy in the west-central portion of North Water Basin (Fig. 3) found about 1.5 km of velocity 3.02 km/sec overlying a velocity of 4.64 km/sec of undetermined thickness. The velocity of 4.64 km/sec (density in the order of 2.5 g/cc; Ludwig et al. 1970) is relatively high for sedimentary rocks, possibly indicating old, dense sedimentary rocks (Keen & Barrett 1973). This would affect the timing of any tectonic activity along Nares Strait.

North of North Water Basin through Smith Sound in similar water depth to that found farther south, the seismic records emphasize the extreme ruggedness and lack of any unconsolidated sediments seen on the bathymetric records.

The eastern boundary of the North Water Basin,

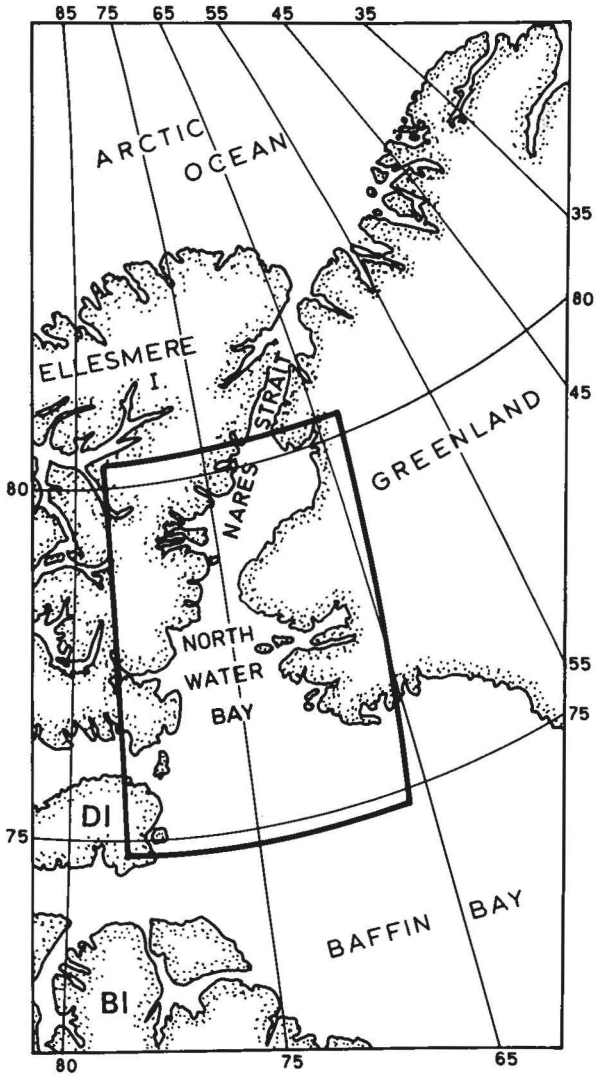


Fig. 1. Index map showing area of marine geophysical study. DI = Devon Island, BI = Baffin Island.

curving around the broad shelf of Greenland (Fig. 2), is defined by a sharp shelf-break and rugged topography. Bedrock is locally exposed. Line 74–12 (Fig. 4) shows the rugged, bare bedrock coming off the Greenland shelf. A bedrock sample of Precambrian granitic gneiss was obtained by electric drill in this area in 1974 by the Bedford Institute (R. K. H. Falconer, pers. comm.). It was found to have a density of 2.62 g/cc and a P-wave velocity of 6.2 km/sec.

A reversed refraction line south of North Water Basin (Fig. 3) defined three sediment layers, with a total thickness of 7 km, and maximum velocity of 4.16 km/sec, over a refractor of velocity 5.5 km/sec. This 'basement' refractor was interpreted as probably carbonate, but possibly Precambrian crystalline basement (Keen & Barrett 1972).

Free air gravity

A free air gravity map (Newman 1977) has been prepared from the marine data from the four Bedford Institute cruises in North Water Bay between 1970 and 1976, and from Dominion Observatory Bouguer land data from the coasts of Devon and Ellesmere Islands south of latitude 78°N. No land data were available north of 78°N on Ellesmere Island, or for any part of northern Greenland. A simplified version of this map is shown in Fig. 5.

Ross (1973) published free air and Bouguer maps of Baffin Bay as far north as approximately 77°30'N from 1970 and 1971 data. The present paper includes more recent and detailed data, and extends his work farther north.

In the absence of any detailed seismic data, a number of 2–D gravity models were attempted for selected profiles in this area to get some feel for the basement topography and density, and the density and thicknesses of any overlying sediments (Newman 1977). This modelling exercise illustrated well the fact that gravity mod-

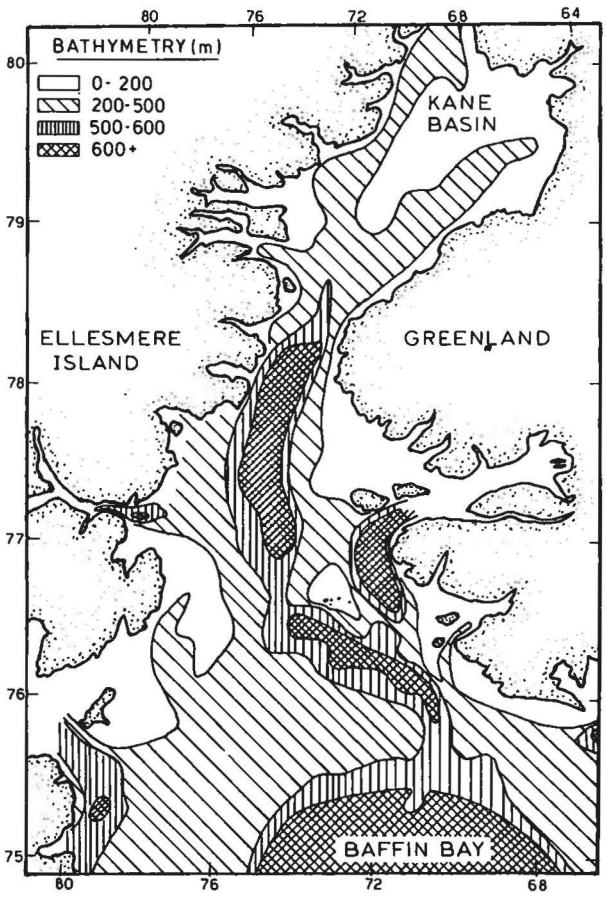


Fig. 2. Simplified bathymetric contour map of southern Nares Strait region.

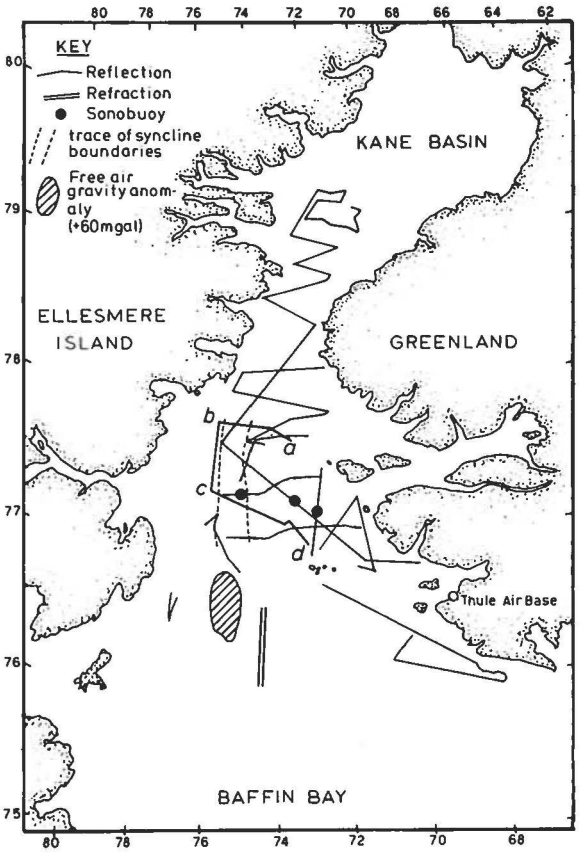


Fig. 3. Seismic coverage map of the North Water Bay area, showing trace of syncline on reflection profiles and the position of the free air gravity high.

elling cannot provide unique solutions, even when partly controlled by seismic reflection profiling. It was still felt, however, that these models could contribute a useful qualitative picture. Fig. 6 shows one of the 'best-fit' models for each of the three profiles located on Fig. 5.

For the models, densities of 2.7, 2.1–2.5, and 1.03 g/cc were assumed for crystalline basement, sediments, and water, respectively. From the few refraction data of the area and the more extensive studies of Lancaster Sound and the Baffin Island Shelf (Jackson et al. 1977) these sediment densities seem reasonable.

A detailed description of the line running northwest from Thule Air Base (Fig. 6, model 2) has been published by Ross & Falconer (1975). As there is good seismic control of basement topography, modelling was used to get an indication of density. It was found that a rough fit could be obtained if all the basement was assumed to have a density of 2.7 g/cc, and the sediment 2.4 g/cc. A better fit was obtained by assuming a density of 2.4 g/cc in the Steensby Basin, and a 2.1 g/cc layer overlying a layer of 2.4 g/cc in the North Water Basin. To fit the depression in the observed gravity near the western end of the line, it was necessary to assume a narrow 4 to 5 km deep body of 2.4 to 2.5 g/cc density below the seismically determined basement. There is no seismic evidence to support either the division of the North Water Basin into two layers or the existence of the deeper, narrow sedimentary basin, but due to the water depth multiple and the poorer quality of the records at depth, the seismic data do not refute these features either.

Profile 1 (Fig. 6), across southern Smith Sound can be explained by inferring about 2 km of material of either 2.5 or 2.2 g/cc density, filling a possibly fault-bounded depression in the basement. A deep (up to 5 km of 2.5 g/cc density), narrow basin of this sediment density is required to fit the small negative anomaly within the larger anomaly. This is possibly a fracture zone or narrow, fault-bounded depression filled with sediments, similar in position and form to that at the western end of profile 2 (Fig. 6). It appears that the syncline noted on the seismic reflection records (Fig. 4) is associated with this fracture in the models. This suggests that the faults were either active during sedimentation, or have been since.

The most intense gravity low of the mapped area (–90 mgal) occurs over the very shallow Kane Basin (Fig. 5). This may indicate a deep sedimentary section, and modelling was consistent with this (Newman 1977). Even with very low density sediment (1.9 g/cc), which seismic reflection profiles do not substantiate, at least 2.5 km of sediment are required. With more dense (2.5 g/cc) sediments, up to 10 km are necessary.

A relatively sharp positive gravity anomaly (max. greater than +80 mgal) located east of southeastern Ellesmere Island (Figs 3 and 5) is crossed by profile 3. Ross (1973) also mentioned this and stated that modelling indicated a shallow basement with a density difference of 0.5 g/cc. He postulated the presence of a basalt-filled fracture zone. Profile 3 (Fig. 6) was modelled for the study of this feature. It is interpreted as an essentially block-faulted basement with a deep depression beneath the free air gravity high, all overlain by high density material. Using basement density of 2.7 g/cc, a density of 3.1 g/cc for the overlying material gave the best fit, requiring a depth of about 10 km beneath the most intense gravity high. The same fit could be achieved with a basement density of 2.6 g/cc, overlain by material of density 3.0 g/cc, which is equally reasonable. Using a lower density, e.g. 2.9 g/cc, for the overlying material, gave unreasonable depths (25 km) of material required, and densities higher than 3.1 g/cc approach those densities expected of the upper mantle.

Magnetics

Magnetic data have been collected on all Bedford Institute's cruises in the area, but due to the large iono-

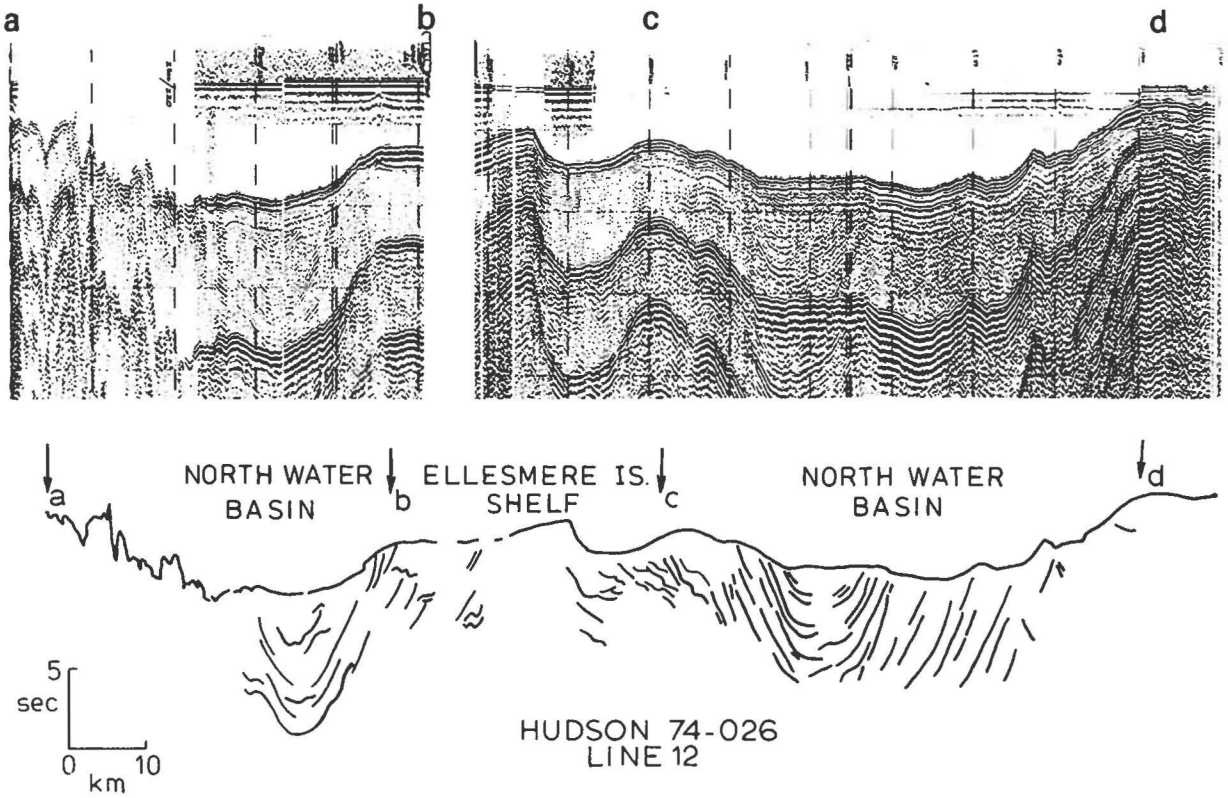


Fig. 4. Above: CSS Hudson line 74–12 seismic reflection profile. Sea surface is the heavy line seen across the top of parts of the record. Below: line drawing of line 74–12. Scale is the same for both profile and line drawing. The locations of points a–d are shown on Fig. 3.

spheric disturbances in these northern latitudes, interpretation and contouring of the data are problematic. However, in conjunction with aeromagnetic data, the marine data can be used qualitatively to characterize particular features.

All lines, marine and aeromagnetic, show a smooth negative anomaly trending northwest through the centre of North Water Bay (Fig. 7). Hood & Bower (1975) interpreted this as a half-graben containing over 20 km of sediments in the east, and shallowing to about 10 km in the west. This is the same feature that has been interpreted to be two sedimentary basins from seismic and gravity data. The basement high, which is distinct on seismic and gravity maps, can be detected only as a small increase in the magnetic signature, offset a little to the west.

As the tracks cross onto the Greenland continental shelf, the anomalies change suddenly to short wavelength positive anomalies. They indicate shallow, probably rugged basement. Farther north these anomalies disappear. Possibly this indicates smoothing or burial of the shelf beneath sediment cover.

To the south of the sedimentary basin anomalies, the magnetic anomalies increase again to form a high which can be followed from east to west almost continuously to both mainlands.

All magnetic lines crossing the region of the positive gravity anomaly off southeastern Ellesmere Island, even with no diurnal correction, are characterized by high amplitude, short wavelength anomalies. This is possibly an indication of the presence of basaltic material, as suggested by Ross (1973) and the gravity modelling.

Discussion

North Water Bay shows a number of interesting features, primarily associated with the North Water Basin (Fig. 7). This is a large arcuate basin, visible on the bathymetry (Fig. 2), free air gravity (Fig. 5) and seismic reflection data (Fig. 4). It is filled with disturbed sediments and trends roughly parallel to the major Precambrian fault and dyke systems on either side of Nares Strait (i.e. trending approximately northeasterly in the west and southeasterly in the east) (Dawes 1976, Frisch et al. 1978). Seismic refraction velocities, and densities from gravity modelling suggest that the sediments in

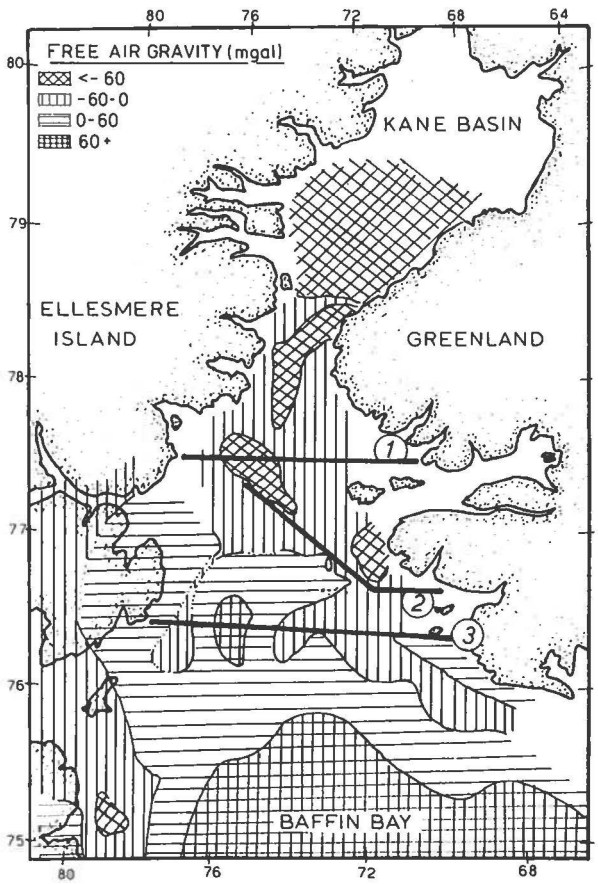


Fig. 5. Simplified free air gravity contour map of the North Water Bay area with location of 2–D model profiles shown in Fig. 6.

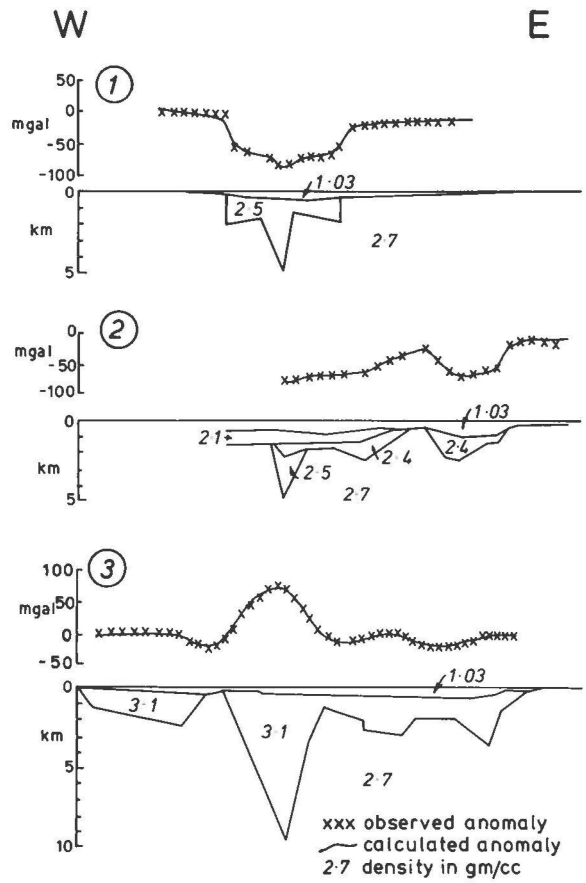


Fig. 6. Free air gravity 2–D models; locations are given in Fig. 5.

these basins are fairly consolidated and therefore probably of Tertiary age or older. The present topography of the basins is probably due to glacial erosion.

Seismic reflection records show relatively disturbed sediments throughout the basin, but in particular show a very clear, sharply-bounded syncline in the west, trending approximately north–south into the southern end of Nares Strait (Figs 3 and 4). Gravity modelling of several lines across this basin (Fig. 6) indicates two major layers of sedimentary density (2.1 and 2.4 g/cc) and this is borne out by seismic refraction results (Keen & Barrett 1973). However, the 'best-fit' models also require a deep, relatively narrow trough filled with fairly high density sediment (2.5 g/cc) directly beneath the syncline seen on the reflection records. This could indicate a fault origin for the syncline and either syntectonic deposition of the sediments in the syncline or considerable (and continuing) compaction of the basal sediments, or both.

Gravity modelling and the short wavelength, high amplitude magnetic anomalies over the anomalous free air gravity high to the south of North Water Basin (Figs

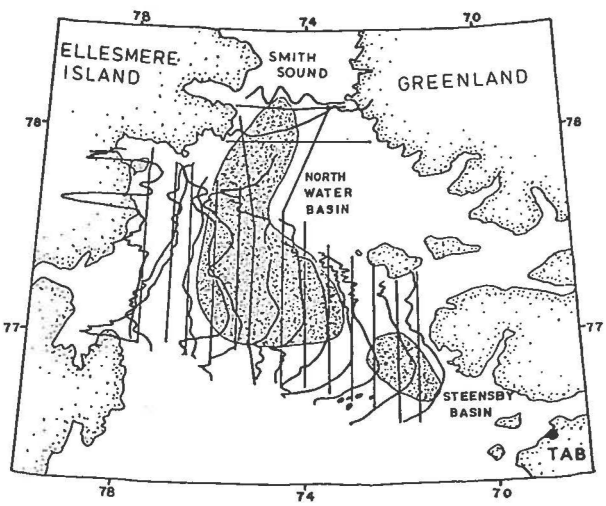


Fig. 7. Positions of North Water and Steensby Basins, as defined by bathymetry, gravity, and seismic reflection, shown in relation to aeromagnetic profiles (modified from Ross & Falconer 1975). Magnetic anomalies are shown positive to the west or north of the flight lines. TAB = Thule Air Base.

5 and 6) tend to confirm Ross's (1973) interpretation of this feature as a basalt-filled fracture. This feature is directly in line with the sediment-filled depression and syncline of western North Water Basin to the north (Fig. 3). It is possible that this basalt(?) and sediment filled 'fracture' trends into Nares Strait, though there is as yet no direct evidence of this. The synclinal sediments of North Water Basin cannot be traced to the north into Smith Sound, which is bathymetrically very rough and shows little seismic penetration. Although the irregular magnetic signature is similar to the typical Precambrian crystalline crustal signature (Hood & Bower 1975), gravity modelling indicates that there is probably a thick sequence (more than 2 km) of high density sediment located in Smith Sound (Newman 1977).

Kane Basin, with a depth of less than 200 m, forms a shallow sill between Baffin Bay and the Arctic Ocean. From seismic work and gravity modelling, it appears to have been the site of substantial sediment deposition. The minimum thickness of sediment possible to explain the less than – 90 mgal gravity anomaly is about 2.5 km of very low density material (too low as indicated by seismic reflection character). Five to ten km of higher density material is more likely. The sediments appear to dip gently northwards, with minor northerly-trending faults. Since the dip is visible only on the lines striking NNE, and the faults only on those striking ESE, this northerly designation may well approach northeast.

Conclusions

Faulting appears to have made a significant contribution to the structure of the offshore, as well as onshore, southern Nares Strait. In particular, a deep (10 km+ ?), relatively narrow, fault-bounded trough, filled with sediments to the north and basalt(?) to the south appears to exist and trend into Nares Strait, though it cannot definitely be traced as far as Smith Sound with the available data.

The minor, shallow northeast(?)-trending faults of Kane Basin, and the fact that the syncline of western North Water Basin appears to continue to the ocean bottom surface, may indicate some minor continuing activity along basement faults, or they may simply be the result of compaction of sediment under the influence of the faulted basement topography. The present aseismic nature of Nares Strait (Basham et al. 1977) may indicate the latter.

From these data, it appears that Nares Strait is a reasonably old, fault-bounded feature, but there is as yet no control on the horizontal movement along these faults.

References

Basham, P. W., Forsyth, D. A. & Wetmiller, R. J. 1977. The seismicity of northern Canada. – Can. J. Earth Sci. 14: 1646–1667.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Frisch, T., Morgan, W. C. & Dunning, G. R. 1978. Reconnaissance geology of the Precambrian Shield on Ellesmere and Coburg islands, Canadian Arctic Archipelago. – Pap. geol. Surv. Can. 78–1A: 135–138.

Hood, P. J. & Bower, M. E. 1975. Northern Baffin Bay: low level aeromagnetic profiles obtained in 1974. – Pap. geol. Surv. Can. 75–1A: 89–93.

Jackson, H. R., Keen, C. E. & Barrett, D. L. 1977. Geophysical studies on the eastern continental margin of Baffin Bay and in Lancaster Sound. – Can. J. Earth Sci. 14: 1991–2001.

Keen, C. E. & Barrett, D. L. 1972. Seismic refraction studies in Baffin Bay: an example of a developing ocean basin. – Geophys. J. Roy. astr. Soc. 30: 253–271.

Keen, C. E. & Barrett, D. L. 1973. Structural characteristics of some sedimentary basins in northern Baffin Bay. – Can. J. Earth Sci. 10: 1267–1278.

Ludwig, W. J., Nafe, J. E. & Drake, C. L. 1970. Seismic refraction. – In: Maxwell, A. E. (ed.), The sea 4(1): 53–84. – Wiley-Interscience, New York.

Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.

Ross, D. I. 1973. Free air and simple Bouguer gravity maps of Baffin Bay and adjacent continental margins. 1:2 000 000. – Pap. geol. Surv. Can. 73–37: 11 pp.

Ross, D. I. & Falconer, R. K. H. 1975. Geological studies of Baffin Bay, Davis Strait, and adjacent continental margins. – Pap. geol. Surv. Can. 75–1A: 181–183.

Review of seismicity and other geophysical data near Nares Strait

ROBERT J. WETMILLER and DAVID A. FORSYTH

Wetmiller, R. J. & Forsyth, D. A. 1982. Review of seismicity and other geophysical data near Nares Strait. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 261–274.

Nares Strait is aseismic even to small magnitude levels, in contrast to nearby active areas in Baffin Bay, the Lincoln Sea and the Sverdrup Basin. Activity immediately to the south of the Strait is infrequent and minor. The nature of Lg propagation suggests a discontinuity in the continental crust at the southern terminus of Nares Strait while seismicity in the Lincoln Sea seems most logically related to an extension of the activity in the Sverdrup Basin.

Intense seismic activity in Baffin Bay appears to be confined south of latitude 75°N. Focal mechanisms for recent earthquakes in Baffin Bay and north of Greenland do not support the hypothesis of large left-lateral displacements along Nares Strait.

Free air gravity anomalies indicate that some of the seismicity of northern Baffin Bay and Baffin Island may be related to glacial rebound or uncompensated sedimentary loads acting on zones of weakness. The aseismic nature of Nares Strait under the action of stresses due to local glacial unloading or regional north–south compression argues against the presence of a relatively young shear zone of major crustal or lithospheric dimension.

R. J. Wetmiller and D. A. Forsyth, Department of Energy, Mines and Resources, Earth Physics Branch, Division of Seismology and Geothermal Studies, 1 Observatory Crescent, Ottawa, Ontario, Canada, K1A OY3.

Contribution from the Earth Physics Branch No. 919.

Two basic premises of plate reconstructions of Greenland and North America are that a major structural zone exists along Nares Strait and that this zone has resulted from or enabled relative movement between Greenland and Baffin Island. The timing of this motion is considered to be mainly post-Late Cretaceous. A corollary to these assumptions is that since the motion involves intermediate size continental plates, the resulting structures must be significant discontinuities in the crust and lithosphere.

Small magnitude seismicity in the Nares Strait region has been monitored since the mid-1960s by the Canadian Seismograph Network. The relationships between seismicity, geology, gravity and glacial rebound information are considered here to see if they support the hypothesis of such a significant structural zone.

Seismicity data

The seismicity data for the Nares Strait region shown in Fig. 1 are taken from the Earth Physics Branch's Earthquake Data files which contain information on all known earthquakes in or near the Canadian Arctic. The events plotted occurred in the interval from 1933 to December 1979. The map is divided into two time periods at 1960, because significant improvements in the earthquake coverage took place in the 1960s with the expansion of the Canadian Seismograph Network in northern Canada. Prior to the 1960s little information is available on earthquakes with magnitude less than 5.0 and epicentres must be considered generally uncertain by 100–300 km (Leblanc & Wetmiller 1974). However, by the late 1960s earthquakes with magnitudes as low as 3.7 are almost completely detected and the epicentre uncertainties are reduced to less than 50 km in general (Basham et al. 1977). This amount of uncertainty still precludes a correlation of earthquakes with specific geological features but it does allow a correlation of seismic activity with geological provinces and thereby sheds some light on current tectonic processes. It should be noted, however, that almost no information is available on the focal depth of the earthquake activity which is generally thought to be occurring at shallow depths in the crust (Hashizume 1973, Qamar 1974, Basham et al. 1977) but it may extend to depths of 65 km (Stein et al. 1979).

The earthquake history of the region prior to 1960 is characterized by a few large poorly located events. None are in Nares Strait. The largest and first is a M 7 1/4 event in November, 1933. This is the largest earthquake in northern Canada, one of the largest anywhere in the Arctic and obvious evidence of significant tec-

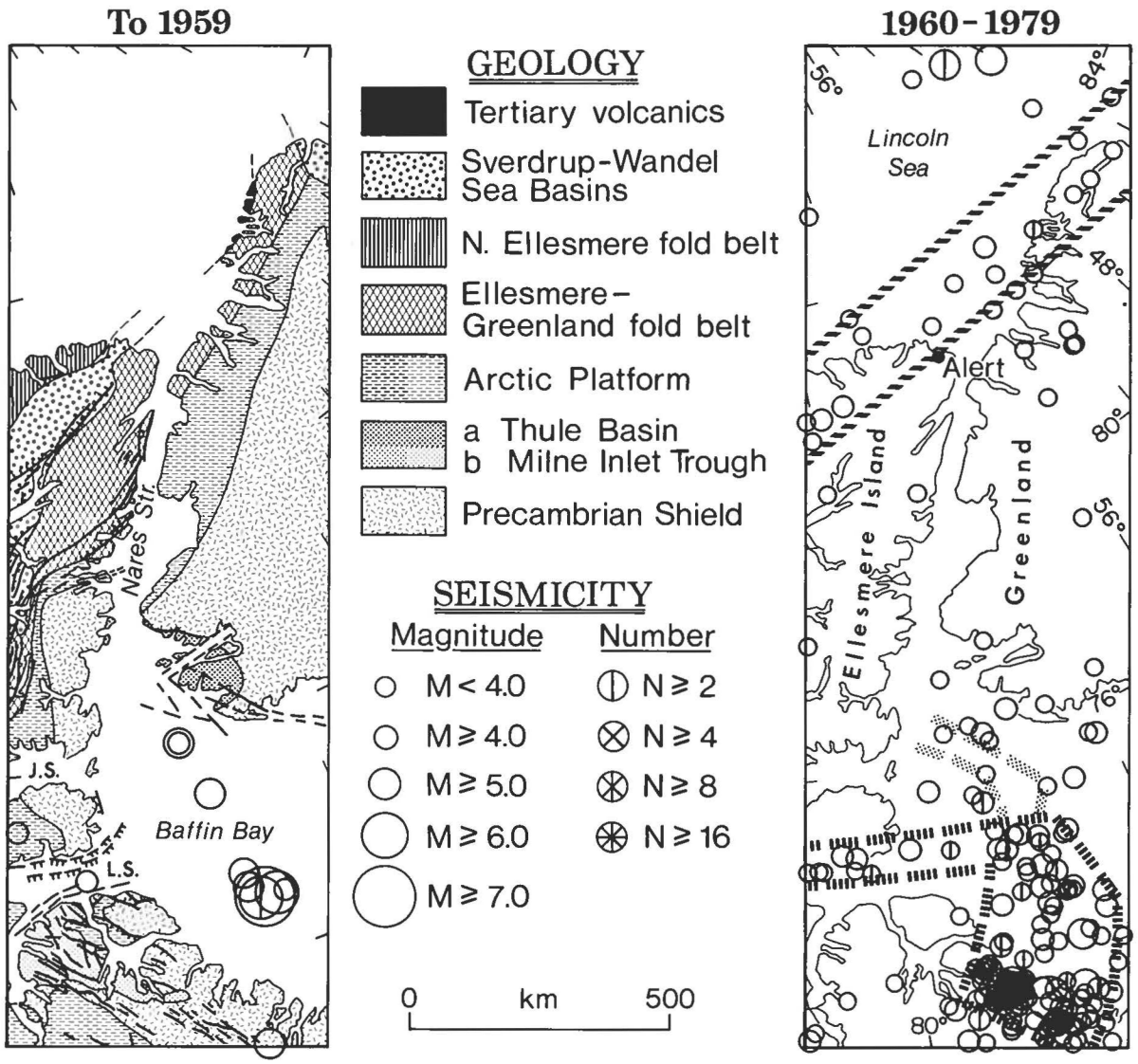


Fig. 1. Seismicity of the Nares Strait region prior to 1960 and since 1960. Geology after Kerr (1967) and Dawes & Soper (1973). Hashed lines on right figure indicate anomalous zones distinguished by epicentre trends or interrupted Lg propagation. L. S. – Lancaster Sound; J. S. – Jones Sound.

tonic activity in northern Baffin Bay within a few hundred kilometres of Nares Strait. Stein et al. (1979) consider this event to be the result of glacial rebound in a coastal area acting on pre-existing faults at a depth of 65 km. The pre-existing faults are thought to be those of the rift system which allowed the separation of Baffin Island and Greenland via the opening of Baffin Bay. The linear trend of epicentres down central Baffin Bay for the period prior to 1960 (only the northern half of that trend of epicentres is visible in Fig. 1) has been taken as evidence by some researchers (Qamar 1974, Barrett et al. 1971) of the central ridge of the rift zone. However, the large uncertainties in location that must be assigned to the events in this period do not allow the central linear trend to be defined with any real confidence. A strong central trend is not apparent in the more recent and better located data.

The aseismic nature of the area since 1960 indicates an apparent absence of significant dynamic processes along Nares Strait. However, the possibility exists that we could be witnessing a quiescent period in a non-stationary record of seismicity. While the record is too short to distinguish this possibility, it should be noted that such quiescence indicates the area is void of events with magnitudes greater than about 3.7. Seismic activity is concentrated however in northern Baffin Bay, near the epicentre of the M 7 1/4 event, and on adjacent Baffin Island. Fig. 1 shows that the earthquake trend in

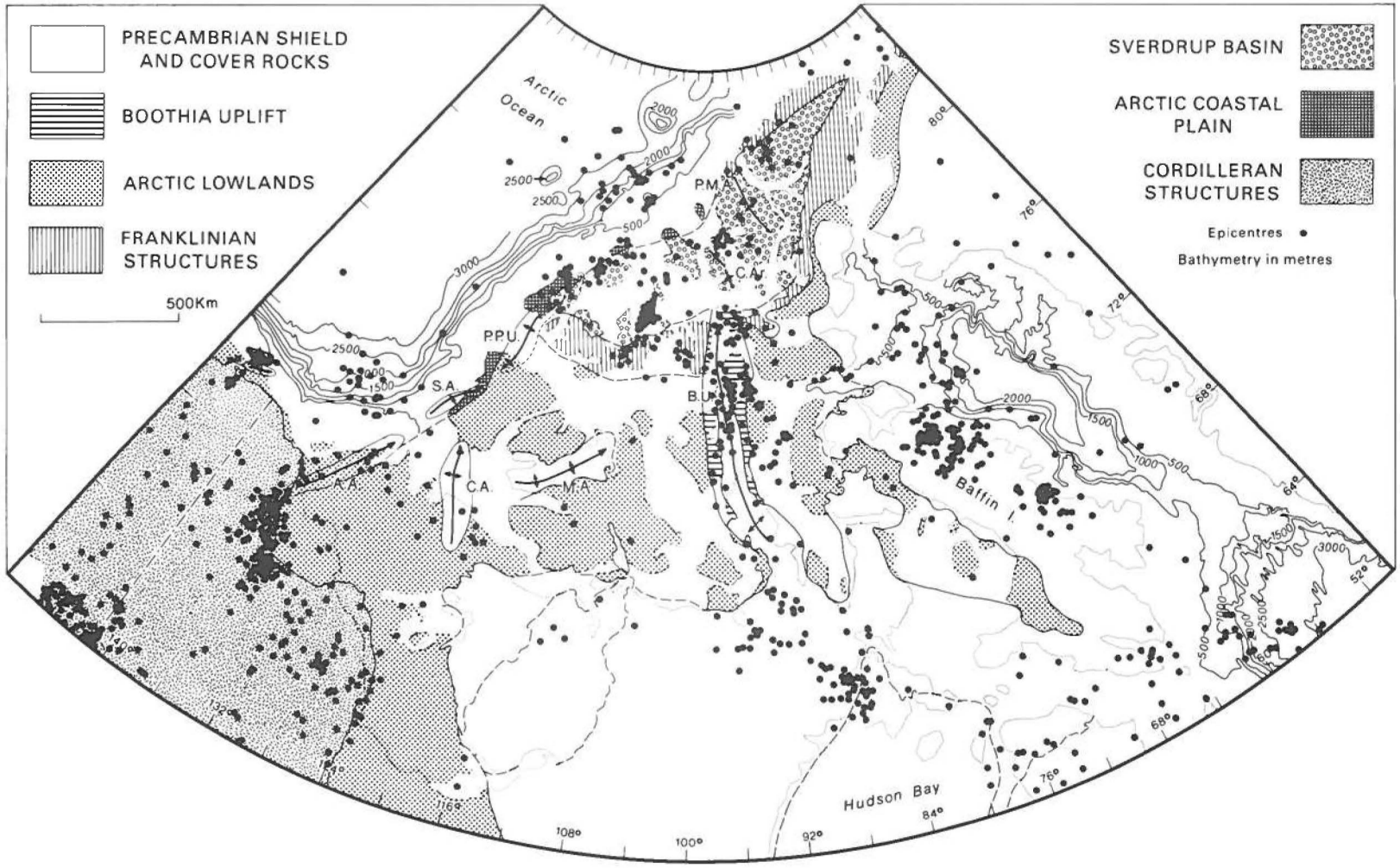


Fig. 2. Recent seismicity and geological structures of the Canadian Arctic and adjacent northern Greenland (after Basham et al. 1977). B.U. – Boothia Uplift; P.M.A. – Princess Margaret Arch; C. Ar. – Cornwall Arch; P. P. U. – Prince Patrick Uplift; S. A. – Storkerson Arch; A. A. – Aklavik Arch; C. A. – Coppermine Arch; M. A. – Minto Arch.

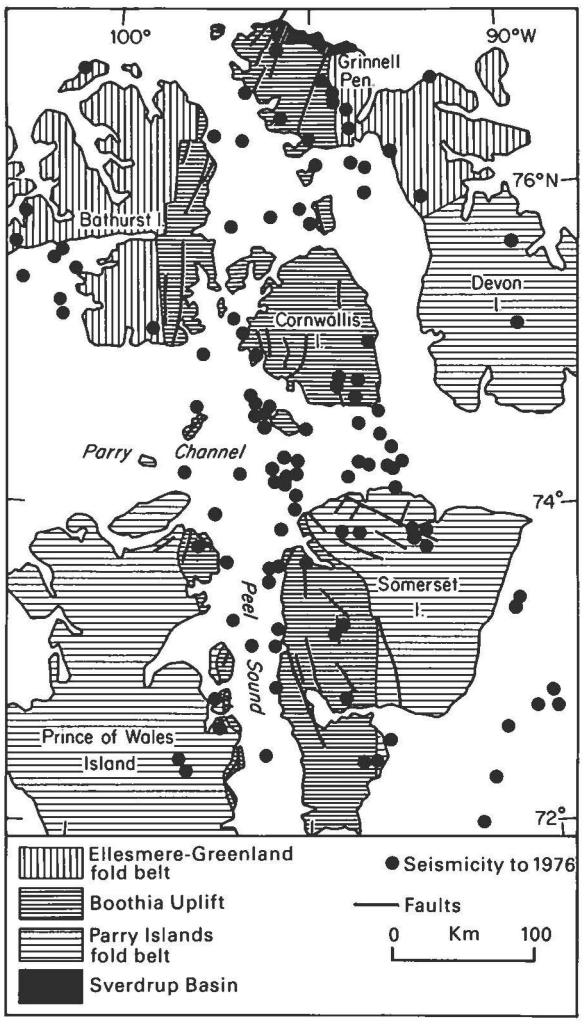


Fig. 3. Seismicity and structural trends of the Cornwallis Fold Belt – Boothia Uplift. Geology after Kerr (1977).

northern Baffin Bay may be cut off by another trend along Lancaster Sound. This trend is only poorly defined by the recent data showing a weak alignment of epicentres parallel to the Sound and a subtle change in the character of the seismic activity in Baffin Bay in line with the Sound. North of Lancaster Sound recent seismic activity is generally less frequent and of smaller magnitude.

Recent seismic activity is scattered elsewhere in the map area without any definite trends except for the possible linear zone of epicentres along the Arctic coast of Greenland and Ellesmere Island in Fig. 1.

The lack of seismic activity along Nares Strait is more remarkable when the extent of seismic activity and relationships to geological structures elsewhere in northern Canada are considered. Fig. 2 shows the recent seismicity in northern Canada superimposed on the geological provinces. Many features such as the Boothia Uplift (Fig. 3) and the structures of the central Sverdrup Basin (Fig. 4), that have geologically demonstrated tectonic activity from the Palaeozoic to the Cretaceous, have contemporary seismic activity associated with them (Forsyth et al. 1979). Thus the lack of seismic activity along Nares Strait, a feature that plate reconstruction theories require to have been active contemporary with and more recently than these other Arctic features, indicates that it is certainly not the present site of any significant dynamic tectonism and casts considerable doubt on the presence of a large crustal or lithospheric rift zone.

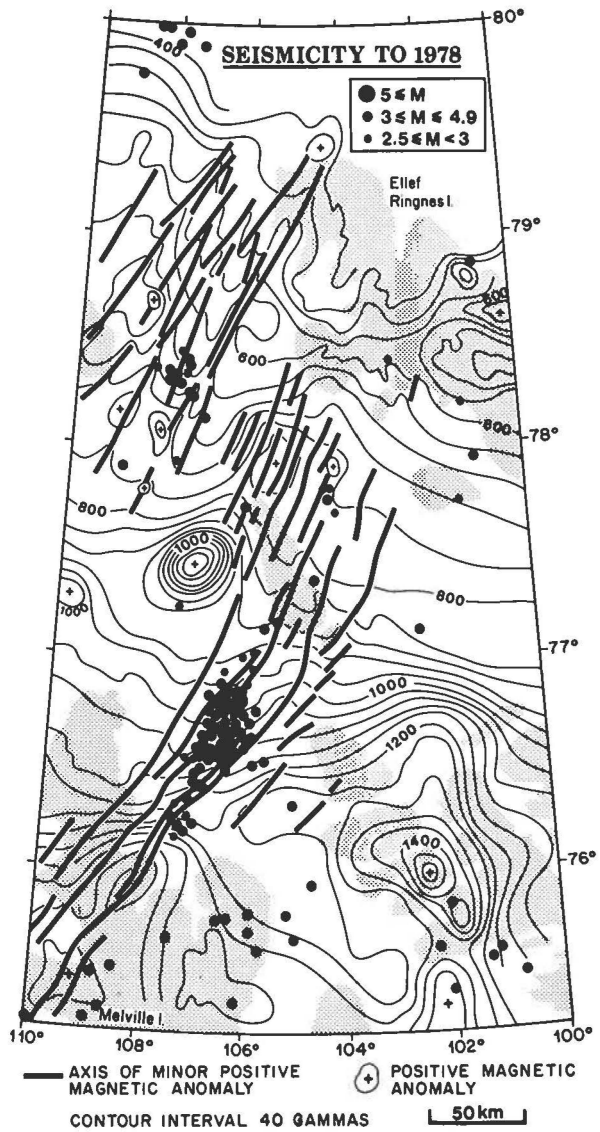


Fig. 4. Seismicity and structural trends of the central Sverdrup Basin.

Lg propagation data

Although Nares Strait has been shown to be aseismic in recent years, at least down to the magnitude 4.0 level, earthquakes in the surrounding areas shed some light on the nature of the crust in the Strait and the possible existence there of any major shear zones. The local earthquake phase, Lg, is sensitive to the gross properties of the Earth's crust (Knopoff et al. 1974) and will not propagate through oceanic crust. This fact has been used to identify areas of oceanic crust in the Arctic (Oliver et al. 1955, Wetmiller 1974) on a regional scale. If there has been large-scale relative motion of crustal plates along Nares Strait, then there should be a discontinuity in the crustal structure along the Strait that could cause a change in any Lg phase that crosses it.

The better data are from recent earthquakes in northern Baffin Bay (Table 1) for which the Lg phase propagates north along the length of Nares Strait to the Canadian seismograph station at Alert on the northern coast of Ellesmere Island (Fig. 1). Fig. 5 shows the distribution of the available data. The Alert seismogram for the May 28, 1972 earthquake is a typical example of the propagation of Lg along Nares Strait. The Lg phase is well developed and indicates that the propagation path from northern Baffin Bay to Alert across Nares Strait is composed of normal continental crust with no major discontinuities. The wavelength of the Lg phase in this case is not more than a few km so that any major break in the gross crustal parameters of more than a few wavelengths (i.e. 10 km) anywhere along the propagation path should result in changes in the Lg phase on the Alert seismogram.

A major break of the crust is apparent in northern Baffin Bay. Fig. 5 shows the evidence for this feature. The Lg phase of any earthquake that crosses the break zone in Fig. 5 is strongly attenuated. The seismograms shown in Fig. 5 are uncorrected for individual station gain, earthquake magnitude or epicentral distance. The best indication of the relative attenuation of the Lg phase in this case is the ratio of the maximum amplitudes on each record of the Lg phase to that of the Sn phase. For propagation in continental areas, the amplitudes of the Lg phase are typically much greater than those of the Sn phase since the energy in the Sn phase travels beneath the crust and is relatively independent of changes in the crustal character. The Resolute record of the May 28, 1972 event is a typical example of the propagation of the Lg phase across the break zone in northern Baffin Bay. In contrast to the Alert record of this event discussed previously, the Resolute record shows that the Lg phase is clearly attenuated; the maximum amplitude of Lg is much smaller than that of Sn. Evidence for this break in the continental crust in northern Baffin Bay was first presented by Wetmiller (1974).

Table 1. Earthquakes used for Lg propagation study of Nares Strait.

Date	Time GMT	Magnitude	Epicentre °N	°W
07 Feb. 1966	00:^{h}14^m:01^s	4.2	75.42	77.00
10 Aug. 1966	23:17:50	4.3	75.00	73.67
07 Sep. 1966	18:02:44	4.7	75.50	65.00
02 June 1971	20:47:49	4.6	74.92	67.56
14 May 1972	14:19:48	4.1	75.07	74.15
28 May 1972	23:08:03	4.2	76.16	72.30
20 Dec. 1978	16:23:58	4.0	76.44	70.16

The existence of the break zone suggests the presence of a significant anomaly in the crustal structure of northern Baffin Bay at the southern end of Nares Strait. The northwest trend of the zone does not support a feature simply continuous with Nares Strait. The break zone is not a seismically active feature; Fig. 5 shows that the earthquake activity in northern Baffin Bay does not seem to be localized on the zone. To the south (in line with Lancaster Sound) the zone merges with the area of high seismicity identified previously. Wetmiller (1974) suggests that the break zone does not continue into the active seismic zone south of Lancaster Sound and does not join the area of oceanic crust in the deep central part of Baffin Bay.

Here again, Lancaster Sound appears to play a fundamental role in the distribution of current seismotectonic features in northern Baffin Bay; it is the southern limit of the break zone and the northern limit of the active seismic zone. The relationship of the break zone and the active seismic zone is intriguing. Perhaps at one time there was an active continuous tectonic zone over both features, possibly a zone of underthrusting running from Baffin Island to Ellesmere Island. Subsequent tectonic processes have then cut the zone at Lancaster Sound, leaving a crustal break and a lower level of seismic activity to the north and a higher level of seismic activity to the south. None of our data suggest that Nares Strait was involved in any of the tectonic activity of northern Baffin Bay.

P nodal solutions

Seven independent P nodal solutions have been published for four earthquakes located in the northern Baffin Bay region south of Nares Strait. A summary of these solutions is given in Table 2. No P nodal solutions are available for Nares Strait itself as the Strait is aseismic but some significant earthquakes have occurred in the Lincoln Sea to the north and east of the Strait. A P nodal solution has been calculated for an event on January 4, 1978 in this area. This earthquake is the closest to the northern end of Nares Strait for which a P nodal solution is possible. It is too far away to be directly related to the present character of the Strait, but by comparison with the P nodal solutions to the

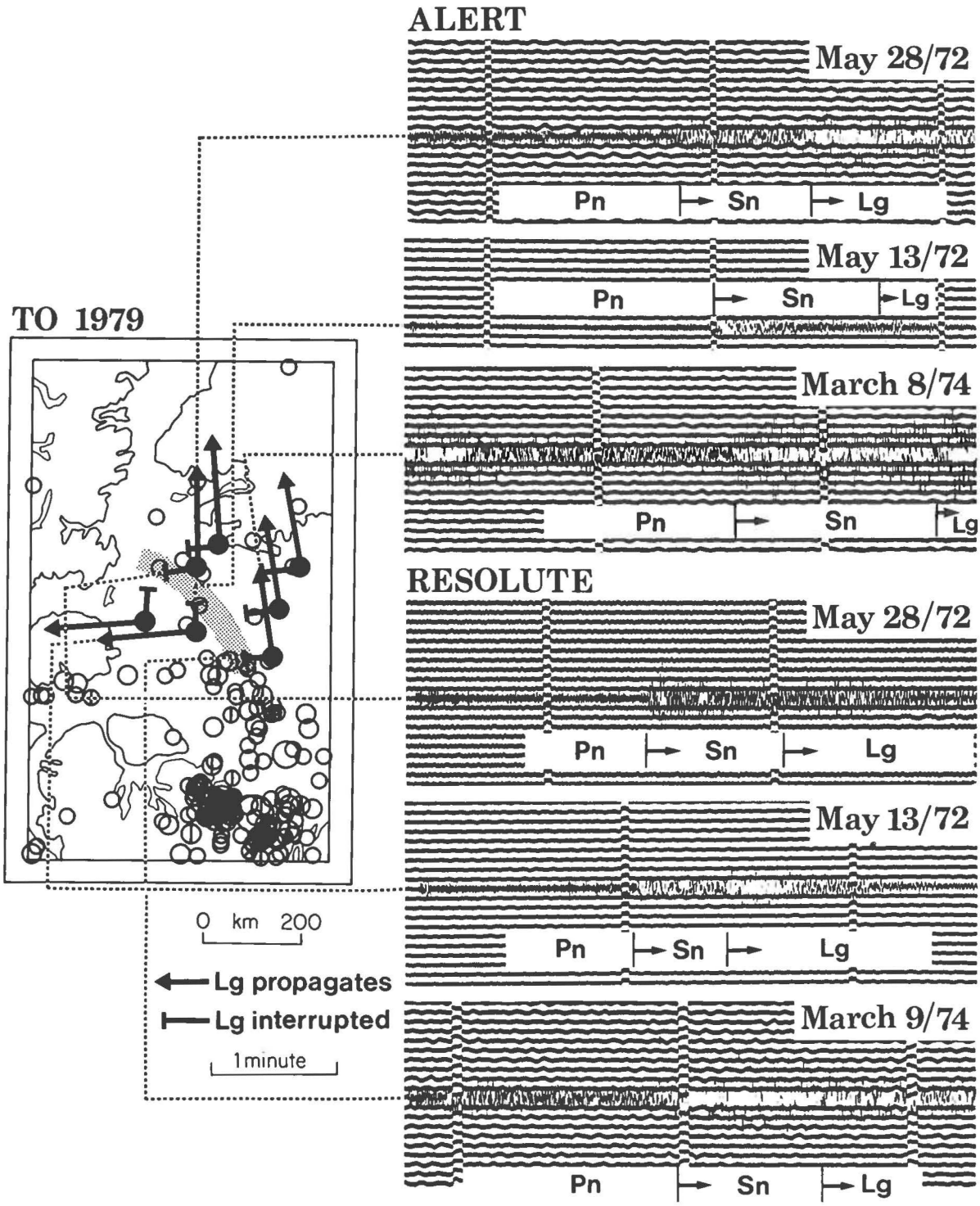


Fig. 5. The nature of Lg propagation from northern Baffin Bay. The minute scale refers to seismogram time scale.

Table 2. Summary of P nodal solutions published for earthquakes near Nares Strait.

Date	Location	Magnitude	Reference
Nov. 1933	Baffin Bay	7.3	Stein et al. (1979)
Sep. 1963	Baffin Island	6.3	Stein et al. (1979) Qamar (1974) Sykes (1970)
Jan. 1972	Baffin Island	4.5	Hashizume (1973)
Nov. 1976	Baffin Bay	5.4	Stein et al. (1979) Wetmiller & Horner (1978)
Jan. 1978	Lincoln Sea	5.0	This study

south, it provides some information on the nature of the stress field to the north and south of the Strait.

The equal area projection of the P nodal solution for the Lincoln Sea event is shown in Fig. 6 and the parameters are summarized in Table 3. First motions at seismograph stations in Canada were read from original records; for stations outside of Canada, data were taken from the Earthquake Data Report (EDR) 78–1 of the National Earthquake Information Service (NEIS). The solution is based entirely on short period data which are sometimes unreliable for such work but the earthquake was too small and the long period seismograph records were too noisy at the time of the event to derive any more reliable data. Readings within 16° of the epicentre are plotted at an angle of emergence of 49°, otherwise angles of emergence are calculated from the Herrin 1968 Tables assuming a focal depth of 15 km. (The EDR lists the event as having a focal depth of 33 km but this is only a default depth used by NEIS when the focal depth is shallow and no data exist to constrain it). The solution indicates almost pure strike-slip motion on either of two steeply dipping planes. There is no independent evidence in this study with which to distinguish either of the nodal planes as the actual fault plane. The principal axis of compression (P in Fig. 6) is nearly horizontal and oriented in the north–south direction. On the whole, the solution is well confined by the data distribution, the relatively large number of reversals (17) are typical of short period data.

Fig. 7 shows the distribution of all eight P nodal solutions that are available near Nares Strait. A common feature of all the solutions is that the principal axis of compression (P) is oriented in the north–south direction. This suggests that the regional tectonic stress is also oriented in the same direction, i.e. sub-parallel to

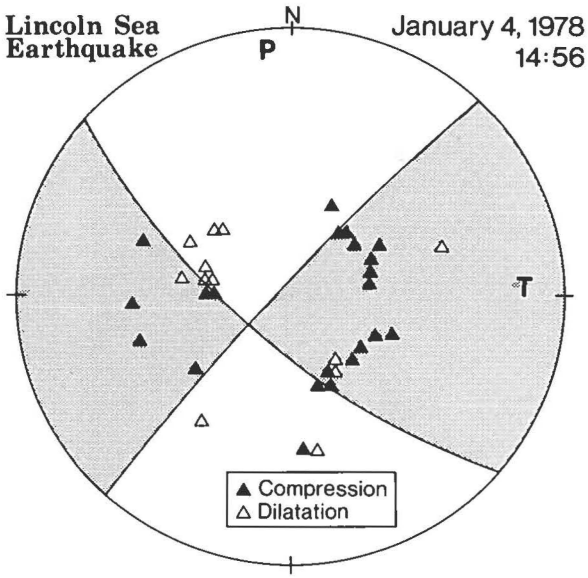


Fig. 6. P nodal solution (lower hemisphere) for event of January 4, 1978. One symbol may represent more than one reading.

the trend of Nares Strait itself. However, the lack of seismicity along Nares Strait shows that any proposed geological structures along the Strait are not responding to this stress while other features in the region are active.

Seven of the P nodal solutions are for earthquakes in the active seismic zone identified previously in northern Baffin Bay and on Baffin Island. An understanding of the present tectonic processes in this zone may indirectly shed some light on the origin of Nares Strait. Three independent P nodal solutions have been published for a 1963 M 6 1/4 earthquake on Baffin Island. The solution by Qamar (1974) in the middle of the 1963 solutions (Fig. 7) is clearly in error and should be disregarded. It is strongly weighted to the first motion at FBC (Frobisher Bay Canada) that, unknown to that author, had a reversed polarity at the time of the earthquake. Without the incorrect reading at FBC, Qamar's (1974) solution for this earthquake would be quite similar to the independent solutions published by Sykes (1970) and Stein et al. (1979). The P nodal solutions for the two earthquakes on Baffin Island are consistent with normal faulting on east–west to northwest-trending planes. The solutions are also consistent in depicting

Table 3. Parameters of the P nodal solution of the January 4, 1978 earthquake in the Lincoln Sea.

Readings	Reversals	A Plane		C Plane		P-Axis		T-Axis	
		az.[1]	dip	az.[1]	dip	az.[1]	dip	az.[1]	dip
80	17	42°	86°NW	130°	76°SW	355°	7°	87°	13°

1) azimuth measured clockwise from north.

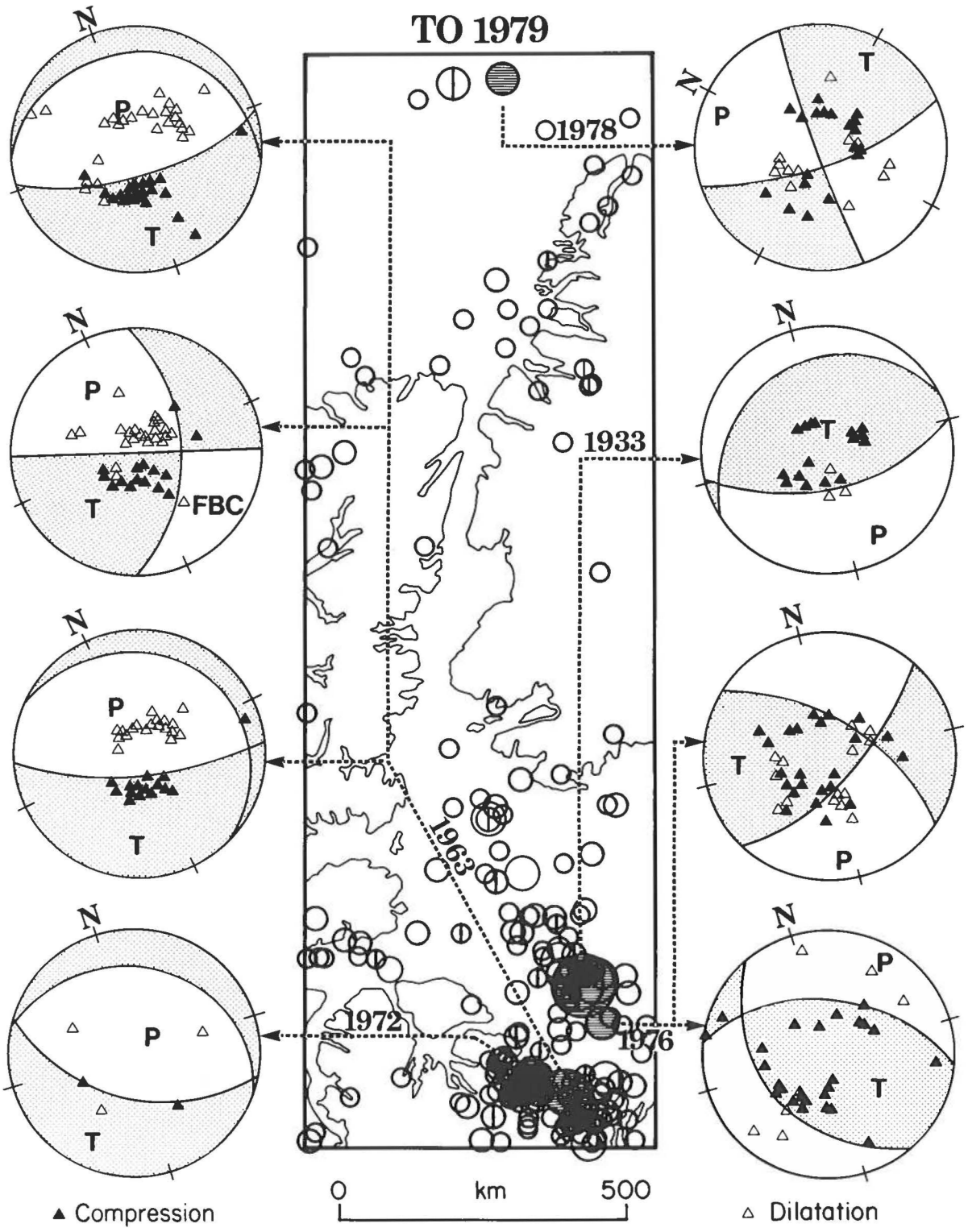


Fig. 7. P nodal solutions in the area of Nares Strait.

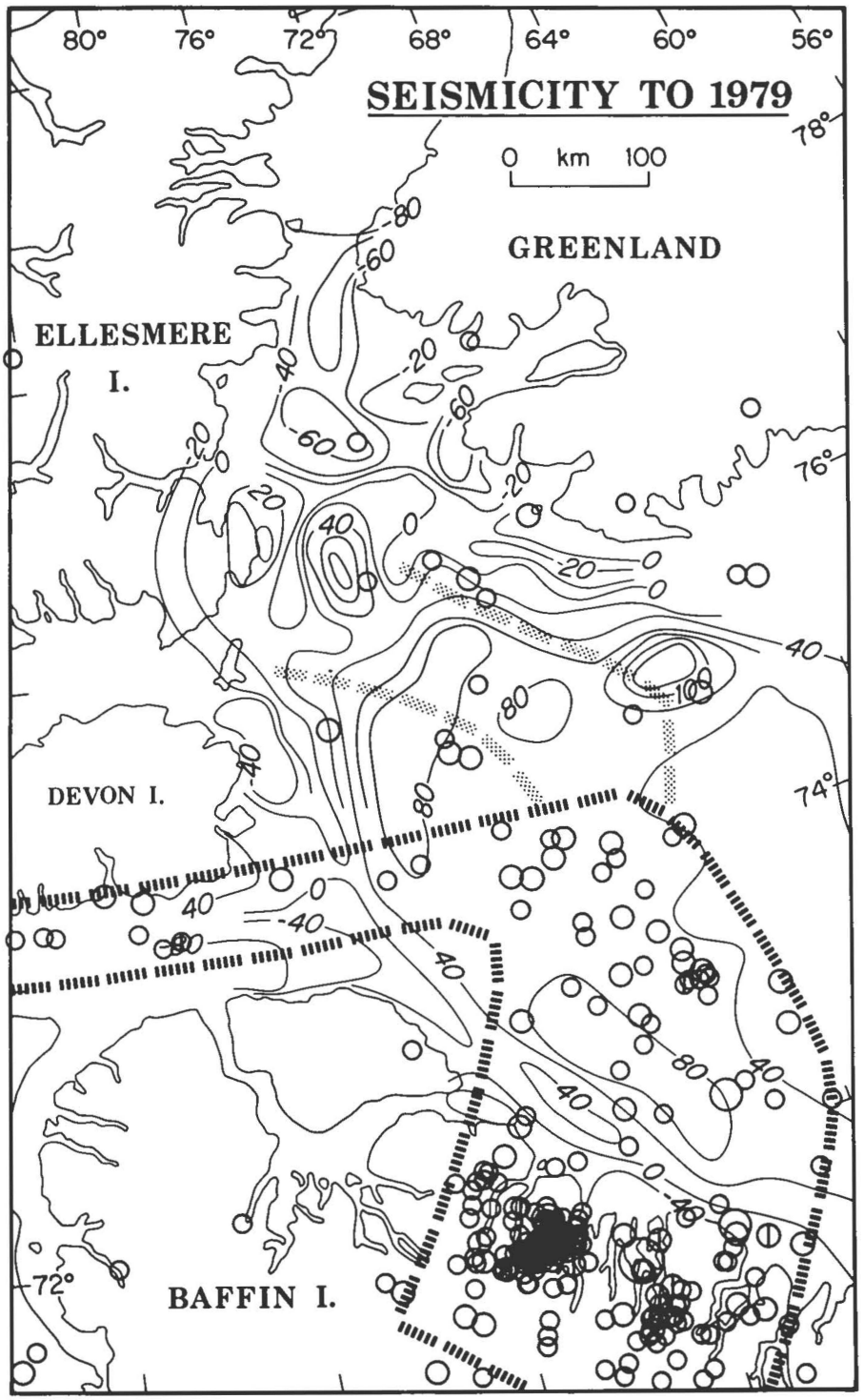


Fig. 8. Free air gravity and seismicity in the northern Baffin Bay – southern Nares Strait region.

one plane dipping very steeply to the south. These planes are similar in orientation and in line with the major fault structures outlining the northeast edge of the Milne Inlet Trough (Jackson et al. 1975). Miall et al. (1980) indicate the Trough is a basin formed by a Cretaceous to mid-Eocene Eurekan period of rifting of a northwest–southeast structural grain that has been periodically rejuvenated since the Proterozoic.

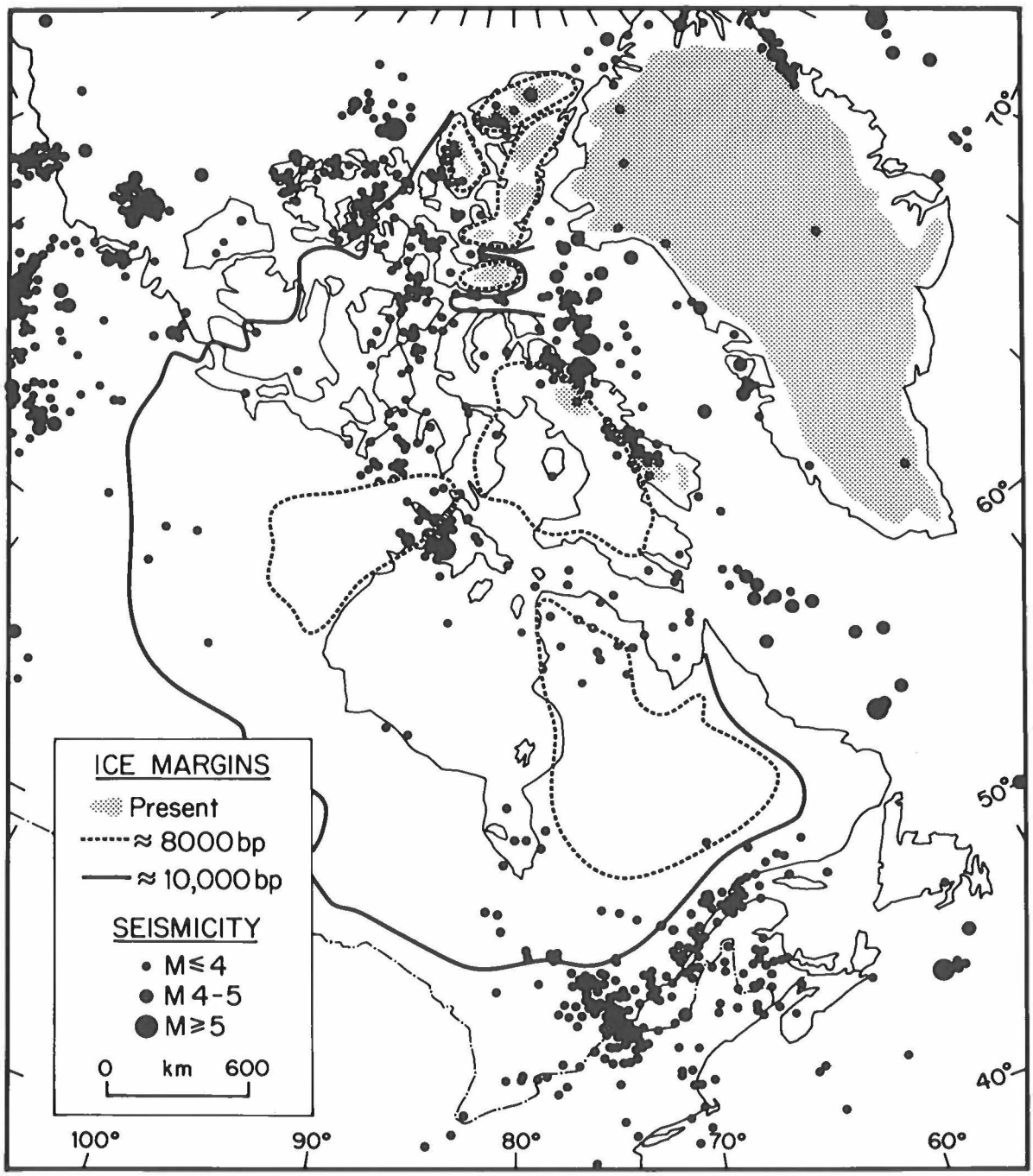


Fig. 9. Distribution of epicentres in northern Canada and Greenland compared to the centres of ice loading during the later stages of the last glaciation.

Two independent and different P nodal solutions have been calculated for the 1976 earthquake in Baffin Bay in the active seismic zone. The difference between the two solutions is mainly due to different focal depths assumed for this earthquake. The lower 1976 solution in Fig. 7 is by Stein et al. (1979) and assumes a focal depth of 33 km. This value is the NEIS default depth and is not an independent calculation. The upper 1976 solution is by Wetmiller & Horner (1978) and assumes a focal depth of 18 km. Again, this is a default depth as there exist no data with which to calculate reliably the focal depth. The difference in the two solutions indi-

cates the sensitivity of the P nodal solutions to focal depth and the care which must be taken in cases where focal depth is poorly known. The two 1976 solutions considered here have in common thrust faulting (Greenland over Baffin Island) on a plane trending roughly northwest–southeast and dipping to the north.

Finally, Stein et al. (1979) have published a P nodal solution for the major 1933 M 7 1/4 earthquake in northern Baffin Bay. Their solution requires thrust faulting on east–west striking planes. They also calculate the focal depth to be 65 km, a surprisingly large value, that is based on a recording of the earthquake at one seismograph station in California.

In all, the four earthquakes in the active seismic zone for which there are P nodal solutions show a simple pattern. Those events located offshore have a solution consistent with thrust faulting, those located onshore have a solution consistent with normal faulting. One interpretation of these observations would be that the thrust activity offshore represents a region of interaction between Greenland and Baffin Island as they respond to a mainly north–south compressive stress. The lack of seismicity along Nares Strait would imply the adjustment is being taken up further west in the Canadian Archipelago and argues against the presence of a relatively young shear zone along Nares Strait.

Sediment loads

Another mechanism which may explain some of the seismic activity in northern Baffin Bay is the effect of recent sediment loading. Fig. 8 shows that there is a large region of positive free air gravity in the northwest-

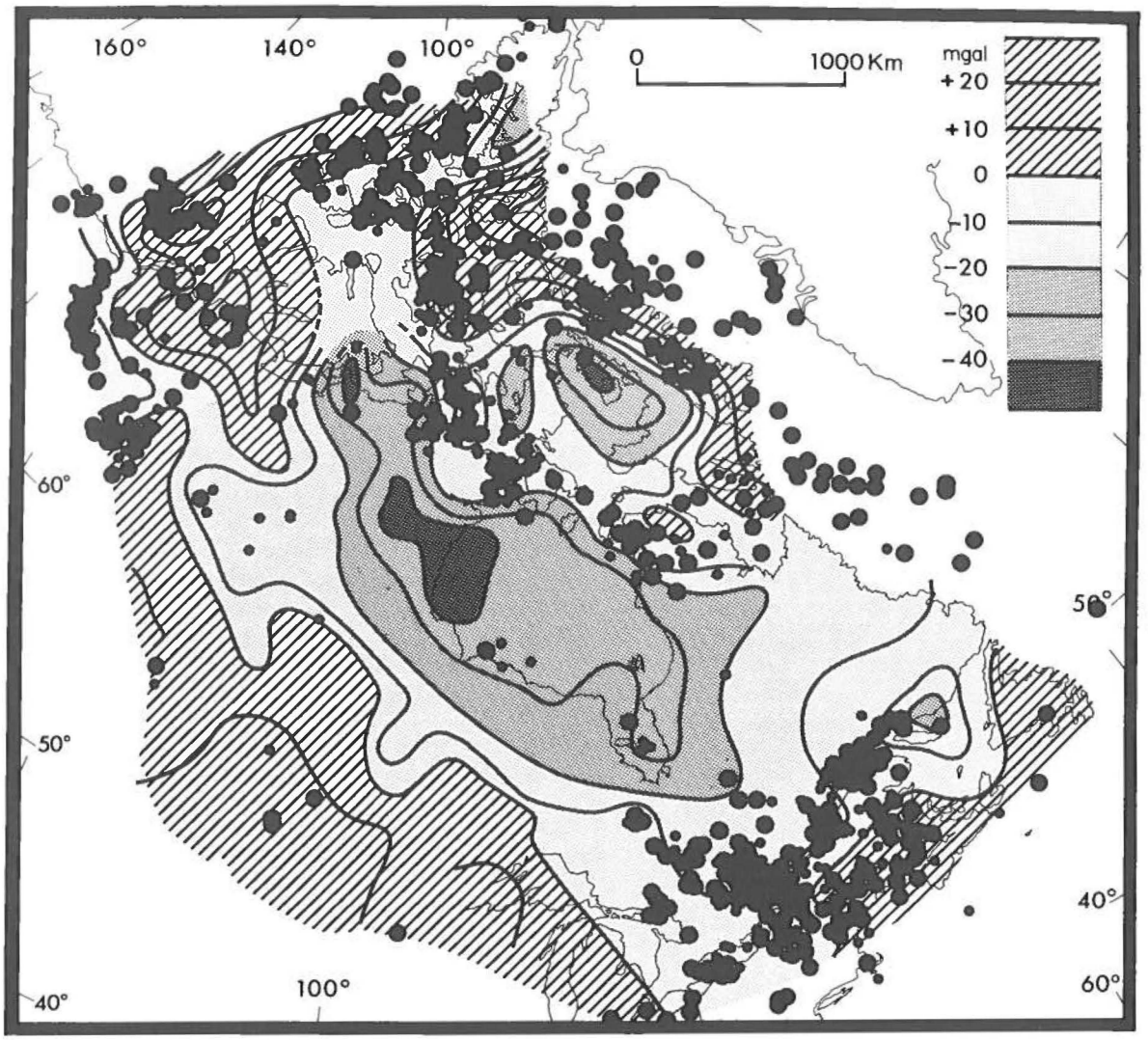


Fig. 10. Distribution of epicentres in northern Canada compared to the smoothed free air anomaly map by Walcott (1972).

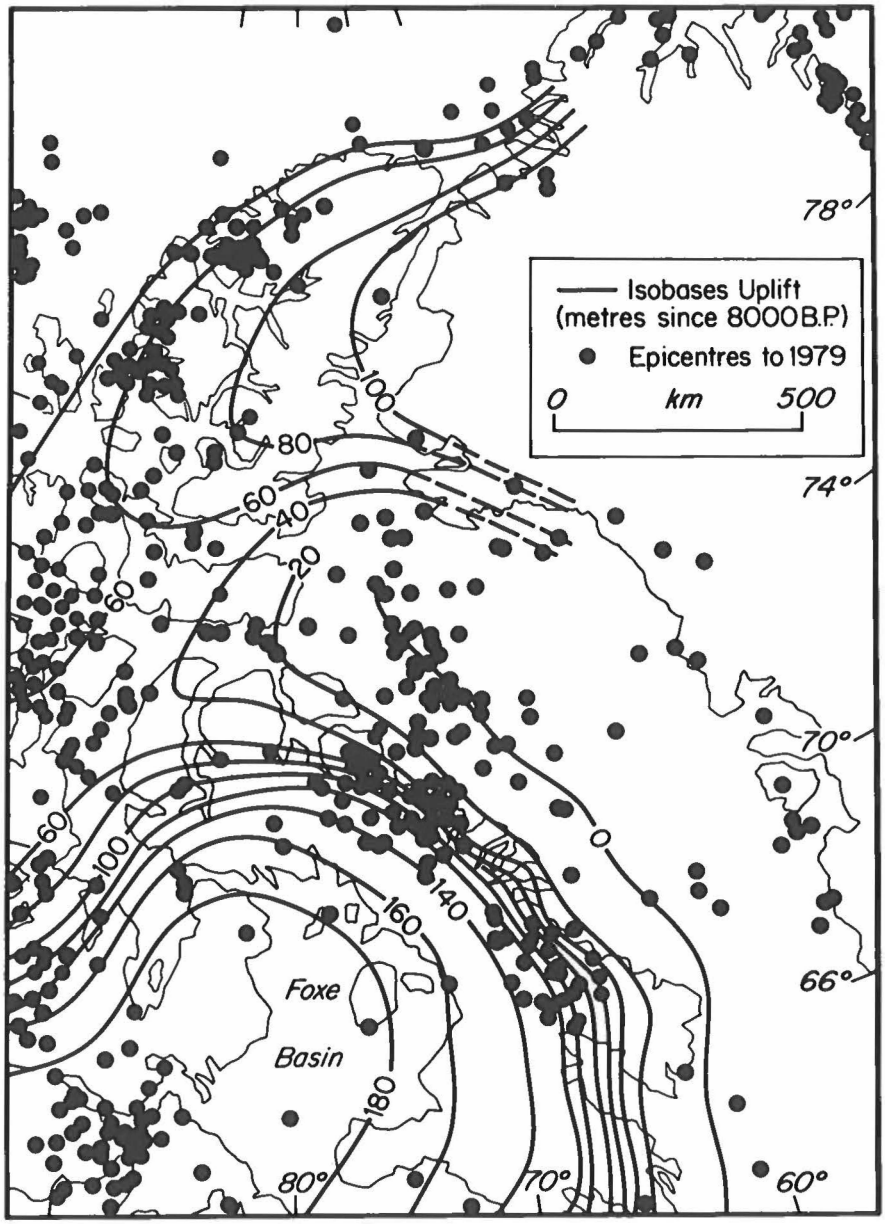


Fig. 11. Glacial isobases and seismicity for the Nares Strait region.

ern part of the Bay. Sobczak (1975) suggests that similar positive free air anomalies along passive continental margins may indicate areas of uncompensated loads of recent sediments. On the Canadian Arctic continental slope there are three such anomalies that correlate with pockets of recent seismic activity (Basham et al. 1977). Whereas the anomalies on the north slope are distinct, the free air anomaly in northern Baffin Bay consists of elliptically-shaped maxima on a broad positive zone circumscribing the northwest area of the Bay.

Fig. 8 shows that most of the recent earthquakes have occurred within the area of the broad positive anomaly. The break zone extending north of Lancaster Sound is aligned with a series of positive anomalies which separate the negative fields over Lancaster Sound and southern Nares Strait.

Considering the thrust nature of the P nodal solutions in the offshore active area, the mass excess indicated by the broad free air gravity high may be evidence of regional crustal thickening caused by the interaction of the Greenland plate with the plate containing Baffin Island. The three smaller elliptical maxima in Fig. 8 may reflect more local sediment loads. Nares Strait appears to be free of any effects due to sediment loading.

Glacial unloading

Stein et al. (1979) consider the pattern of P nodal solutions to be a consequence of glacial rebound at the continental margin reactivating the basement faults remaining from the rifting of Baffin Bay. Basham et al. (1977) have also noted the correlation of seismic activity on Baffin Island and elsewhere in northern Canada with areas of high differential rates of glacial rebound. Fig. 9 shows the epicentral distribution relative to the suggested ice centres of the later stages of the last glaciation (Prest 1970). Note the central areas of the major ice sheets are nearly aseismic with most of the activity centred around the edges. Fig. 10 shows the effect as reflected in the smoothed free air gravity map by Walcott (1972). The broad free air lows over Hudson Bay and the Foxe Basin represent areas that have yet to attain isostatic equilibrium and are aseismic while the fringe or interval areas representing areas of higher rebound gradients are characterized by seismic activity.

Gravity data in Nares Strait are sparse, but uplift isobases from the work of Miller & Dyke (1974) and England (1976) may be compared with the regional seismicity. Fig. 11 shows that for the Foxe Basin – Baffin Island area the relationship between the isobase gradient and seismicity is very similar to that between the free air gravity gradient and seismicity. Note that the uplift over Nares Strait in area Fig. 11 appears to reflect the response of the Greenland sheet and extends uninterruptedly across Nares Strait. A major crustal structural discontinuity along Nares Strait might be expected to influence the glacial response of this area.

Conclusions

Nares Strait is presently aseismic. Focal mechanism solutions from areas immediately south and to the north of the Strait indicate the region is under the influence of a north–south compressive stress, sub-parallel to the Strait itself. Glacial isobases suggest that Nares Strait does not play a significant role in the isostatic rebound of the area. Other major structural features of the Canadian Arctic that have demonstrated tectonic activity contemporary with the activity proposed for Nares Strait, play a significant role in the rebound of the Canadian Arctic and are represented in the present seismicity. The aseismic nature of Nares Strait indicates that it is not presently a zone of active tectonism and contests the hypothesis of a relatively young rift zone of major crustal or lithospheric dimension along the Strait.

Our geophysical evidence indicates that there are currently active tectonic features in northern Baffin Bay. These are probably faults generated by the opening of Baffin Bay that are presently being reactivated by postglacial rebound. However, we can see no evidence to extend these features north into Nares Strait. Instead, our data suggest that the tectonic features in northern Baffin Bay continue west into Lancaster Sound or possibly northwest into Jones Sound. On the basis of this we suggest that movements proposed for Nares Strait in plate reconstructions of Greenland and North America involve deformations west of Ellesmere Island in the Canadian Archipelago.

References

Barrett, D. L., Keen, C. E., Manchester, K. S. & Ross, D. I. 1971. Central part of Baffin Bay is oceanic. – Nature, Lond. 229: 551–552.

Basham, P. W., Forsyth, D. A. & Wetmiller, R. J. 1977. The seismicity of northern Canada. – Can. J. Earth Sci. 14: 1646–1667.

Dawes, P. R. & Soper, N. J. 1973. Pre-Quaternary history of North Greenland. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 117–134.

England, J. 1976. Postglacial isobases and uplift curves from the Canadian and Greenland High Arctic. – Arct. Alp. Res. 8: 61–78.

Forsyth, D. A., Mair, J. A. & Fraser, I. 1979. Crustal structure of the central Sverdrup Basin. – Can. J. Earth Sci. 16: 1581–1598.

Hashizume, M. 1973. Two earthquakes on Baffin Island and their tectonic implications. – J. geophys. Res. 78: 6069–6081.

Jackson, G. D., Davidson, A. & Morgan, W. C. 1975. Geology of the Pond Inlet map-area, Baffin Island, District of Franklin. – Pap. geol. Surv. Can. 74–25: 33 pp.

Kerr, J. W. 1967. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Kerr, J. W. 1977. Cornwallis fold belt and the mechanism of basement uplift. – Can. J. Earth Sci. 14: 1374–1401.

Knopoff, L., Schwab, F., Nakanishi, K. & Chang, F. 1974. Evaluation of Lg as a discriminant among different continental crustal structures. – Geophys. J. Roy. astr. Soc. 39: 41–70.

Leblanc, G. & Wetmiller, R. J. 1974. An evaluation of seismological data available for the Yukon Territory and the Mackenzie Valley. – Can. J. Earth Sci. 11: 1435–1454.

Miall, A. D., Balkwill, H. R. & Hopkins, W. S. 1980. Cretaceous and Tertiary sediments of Eclipse Trough, Bylot Island area, Arctic Canada, and their regional setting. – Pap. geol. Surv. Can. 79–23: 20 pp.

Miller, G. H. & Dyke, A. S. 1974. Proposed extent of late Wisconsin Laurentide ice on eastern Baffin Island. – Geology 2: 125–130.

Oliver, J., Ewing, M. & Press, F. 1955. Crustal structure of the Arctic regions from the Lg phase. – Bull. geol. Soc. Am. 66: 1063–1074.

Prest, V. K. 1970. Quaternary geology of Canada. – In: Douglas, R. J. W. (ed.), Geology and economic minerals of Canada. – Econ. Geol. Rep. geol. Surv. Can. 1: 676–764.

Qamar, A. 1974. Seismicity of the Baffin Bay region. – Bull. seism. Soc. Am. 64: 87–98.

Sobczak, L. W. 1975. Gravity anomalies and passive continental margins, Canada and Norway. In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 743–761.

Stein, S., Sleep, N. H., Geller, R. J., Wang, S.-C. & Kroeger, G. C. 1979. Earthquakes along the passive margin of eastern Canada. – Geophys. Res. Lett. 6: 537–540.

Sykes, L. R. 1970. Focal mechanism solution for earthquakes along the world rift system. – Bull. seism. Soc. Am. 60: 1749–1752.

Walcott, R. I. 1972. Late Quaternary vertical movements in eastern North America: quantitative evidence of glacio-isostatic rebound. – Rev. Geophys. Space Phys. 10: 849–884.

Wetmiller, R. J. 1974. Crustal structure of Baffin Bay from the earthquake-generated Lg phase. – Can. J. Earth Sci. 11: 123–130.

Wetmiller, R. J. & Horner, R. B. 1978. Canadian earthquakes – 1976. – Seism. Ser. Earth Physics Branch 75: 79 pp.

Motion along Nares Strait recorded in the Lincoln Sea: aeromagnetic evidence

LOUIS C. KOVACS

Kovacs, L. C. 1982. Motion along Nares Strait recorded in the Lincoln Sea: aeromagnetic evidence. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 275–290.

Detailed low-level aeromagnetic data taken over the Lincoln Sea and surrounding regions are discussed in an attempt to explain the evolution of the area and its relation to the possible movement of North America and Greenland along Nares Strait. A depth-to-source analysis of the data delineates a major northeastward extension of the Nares Strait extending to the western edge of the Morris Jesup Rise, and identified here as the Arctic Ocean portion of the Wegener Fault or Fracture Zone. Two regions of moderate amplitude, short wavelength anomalies characterized by shallow (< 2 km) magnetic basement are delineated, one to the north of Ellesmere Island and the other to the northwest of the Morris Jesup Rise, as well as a steep escarpment paralleling the Wegener Fault and forming the northwest boundary of a deep, sediment-filled basin. A plate tectonic analysis of these data suggests that the hot spot responsible for the formation of the Morris Jesup Rise–Yermak Plateau in the late Eocene and early Oligocene and the Kap Washington Group volcanics at about the Cretaceous–Tertiary boundary was also active in the late Cretaceous. During this period, prior to any strike-slip movement of Greenland along Nares Strait, a single, Iceland-like volcanic massif was formed north of Greenland and Ellesmere Island contemporaneously with postulated sea-floor spreading in the Makarov Basin. Sinistral movement of North America from Greenland along Nares Strait, and along a second "Ellesmere Fracture Zone" extending northeast through the massif from the edge of the Ellesmere Island shelf then led to the formation in early Tertiary time of a deep sediment-filled basin due to thinning and foundering of the crust northeast of Nares Strait, the emplacement of the western half of the massif between Lomonosov Ridge and Ellesmere Island, and of the eastern half of the massif just west of the Morris Jesup Rise. A shift in the hot spot to the east then led, with the initiation of spreading in the Eurasia Basin, to the later formation of the Morris Jesup Rise and Yermak Plateau. A total left-lateral displacement of about 250 km between these features is noted, similar to the amount suggested along Nares Strait by other geophysical work in Baffin Bay and the Labrador Sea.

L. C. Kovacs, Department of the Navy, Naval Research Laboratory, Washington, D.C. 20375, U.S.A.

Nares Strait, a linear channel separating northwestern Greenland from eastern Ellesmere Island, was considered by Wegener (1922) to be a great strike-slip fault formed by the separation of Greenland from Canada to form Baffin Bay. Since then numerous investigators have proposed various sequences of plate tectonic motion along Nares Strait (see for example, Pitman & Talwani 1972, Churkin 1973, Johnson & Vogt 1973, Ostenso & Wold 1973, Herron et al. 1974, Kristoffersen & Talwani 1977, Sclater et al. 1977, Srivastava 1978, Sweeney et al. 1978, Feden et al. 1979, Jackson et al. 1979). Other investigators (Riddihough et al. 1973, Kerr 1980) however have found little or no geologic evidence for significant offset across Nares Strait. A good review of the work by both mobilists and fixists concerning Nares Strait, as well as a compromise hypothesis, may be found in Kerr (1980) and Srivastava & Falconer (this volume).

Any strike-slip motion of Greenland along Nares Strait would be expected to have left traces in the area of the Lincoln Sea, the marginal Arctic sea north of Greenland and Ellesmere Island (Fig. 1). In fact, if the Nares Strait lineament is the surface expression of a transform fault sometimes called the Wegener Fault (Johnson & Vogt 1973, Ostenso & Wold 1973), plate tectonic considerations would require that the Wegener Fault extends through the Lincoln Sea to the anomaly 13 isochron on the west flank of the Nansen Ridge, the time at which spreading in the Labrador Sea (Srivastava 1978) and Baffin Bay (Jackson et al. 1979) ceased. Evidence of possible plate tectonic motion in the Lincoln Sea would also have to be considered in the light of geologic evidence of nearby igneous activity of latest Cretaceous or earliest Tertiary age (Rb/Sr age of 63 m.y.) from the Kap Washington Group volcanics and their associated metasediments (Dawes 1973, Larsen et al. 1978). This paper examines previously unpublished aeromagnetic data from a detailed 1975 investigation of the Lincoln Sea to see if evidence relating these tectonic episodes exists. I will also consider evidence relating

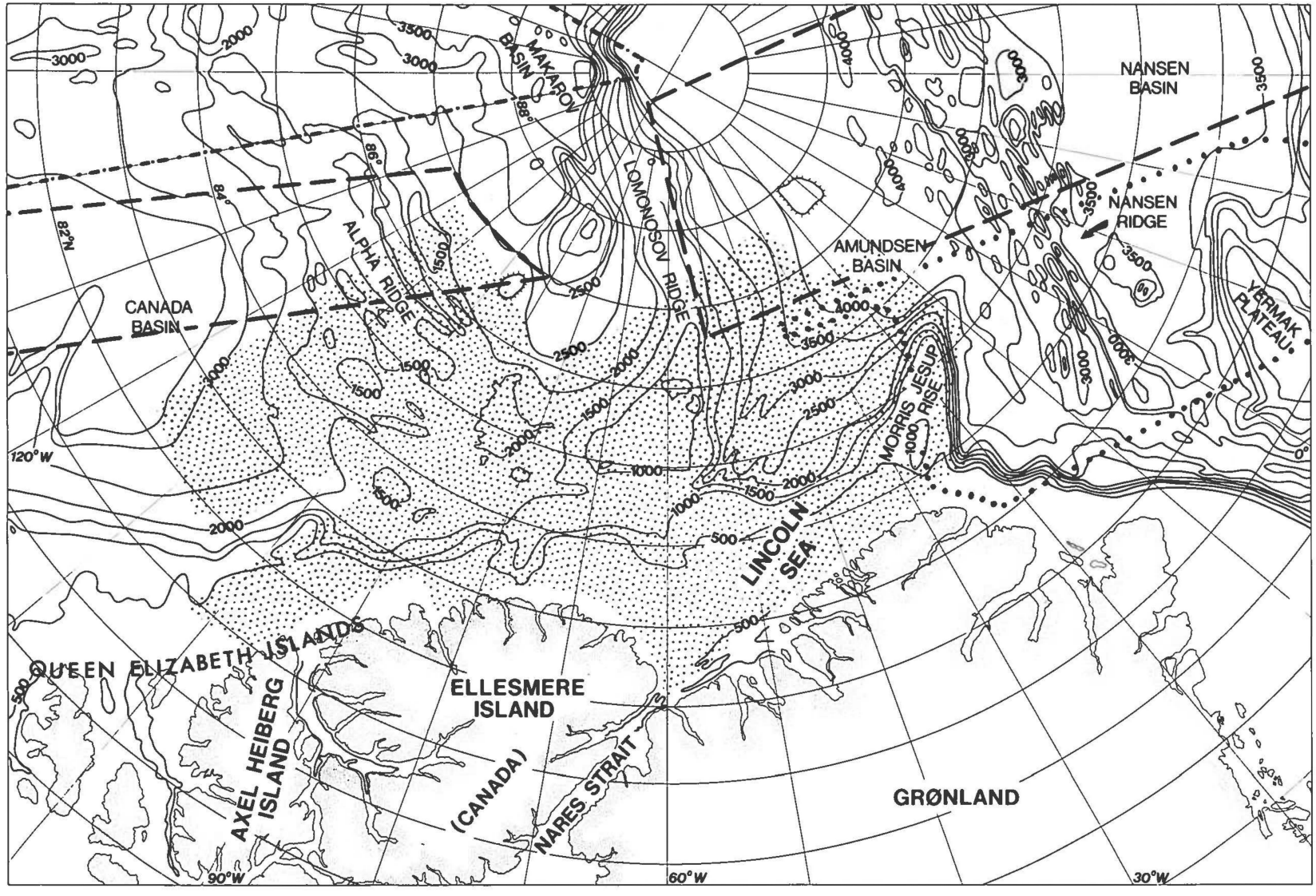


Fig. 1. Lincoln Sea and surrounding regions, showing areas of detailed aeromagnetic coverage. Stippled region denotes area discussed in this paper. Dotted outline refers to data discussed in Feden et al. (1979); dashed outlines refer to data discussed by Vogt et al. (1979a); dot-dash line outlines the region discussed in Taylor et al. (1981). Bathymetry is in metres, from Johnson et al. (1979).

Nares Strait to a RFF (ridge-transform fault-transform fault) triple junction involving the contemporaneous spreading of the Eurasia Basin, Norwegian–Greenland Sea and Baffin Bay/Labrador Sea from late Paleocene (anomaly 24) to early Oligocene (anomaly 13) and the formation of the Morris Jesup Rise–Yermak Plateau by the postulated "Yermak hot spot" (Herron et al. 1974, Vogt et al. 1978, 1979a, b, Feden et al. 1979), as well as recent evidence of sea-floor spreading in the Makarov Basin–Fletcher Abyssal Plain from the Campanian to the late Paleocene (Taylor 1978, Taylor et al. 1981).

Data collection and display

The over 30 000 line kilometres of aeromagnetic data discussed were collected by the U.S. Naval Research Laboratory in April and May of 1975. The area investigated (Fig. 1) includes the Alpha Ridge east of the 120°W meridian, overlapping the Alpha Ridge data reported on by Vogt et al. (1979a), the continental shelves of Axel Heiberg and Ellesmere Islands and the Lincoln Sea shelf north of Greenland to Kap Morris Jesup, the Morris Jesup Rise, (overlapping the Eurasia Basin data reported on by Feden et al. 1979 and Vogt et al. 1979a), the Greenland–Lomonosov Ridge slopes of the Amundsen Basin to the 4000 metre isobath, the Lomonosov Ridge south of 87°30'N and the slope of the Makarov Basin above the 2500 metre isobath. The data tracks were flown perpendicular to the major trend of the Alpha Ridge, an orientation that put the tracks approximately 30° clockwise to the strike of Nares Strait (Fig. 2). Data-track spacing was approximately 12 km over the Lincoln Sea and 22 km over the Alpha Ridge.

The data were collected aboard a modified Lockheed P3–A Orion aircraft equipped with a proton precession magnetometer. Flight altitude and speed were nominally 300 m and 450 km/hr, respectively. Navigational control was maintained using a Litton 51 dual inertial navigation system. This system generally provides positional accuracy with a drift error of .8 km/flight hour or less, so that a typical flight of 10 hours would experience a round trip error of 8 km at the end of the flight. On very long flights (12 hours) the accuracy is occasionally less due to the additional effect of the exponentially increasing amplitude of the Schuler cycle of the gyros. Fortunately, for very long flights the last few hours were usually spent in transiting back to base at high altitude so that no magnetic data were collected. Round-trip errors did not exceed 10 km for a 10 hour flight. Use of the same aircraft and navigation system in the Norwegian–Greenland Sea in 1972, 1973 and 1974, where post-flight correction of the aircraft tracks from Loran–C and satellite fixes was possible, showed overall errors of 3 to 6 km. The magnetic field total intensity, aircraft position and altitude were collected every three seconds, the IGRF–65 anomaly calculated in real-time using an onboard mini-computer, and the time, position, altitude, and residual and total field intensities stored on magnetic tape and in disc file by track.

The aeromagnetic data displayed in Fig. 3 were recalculated using the IGRF–75 reference field (Anonymous, 1976). Because slight DC offsets in the data relative to the IGRF were important in preventing an adequate display of the low-amplitude anomalies over the Lincoln Sea, the data were high pass filtered to remove any long wavelength bias due to diurnal variations and inaccuracies in the IGRF. The filter cut off period was 90 minutes, corresponding to a wavelength of 600 km at the aircraft speed flown. An unfortunate by-product of this operation would be the removal of any two-dimensional anomalies which strike parallel to the tracks. This was minimized by the orientation of the flight lines perpendicular to the dominant palaeomagnetic fabric.

Magnetic field structure

When compared to the latest available bathymetry in the area (Fig. 1, Johnson et al. 1979), interpretation of the aeromagnetic data becomes a study in contrasts (Fig. 3). Anomalies over the portion of the Alpha Ridge investigated here are typical of those reported for other areas of the ridge (Riddihough et al. 1973, Coles et al. 1978, Vogt et al. 1979a, Taylor et al. 1981), having short wavelengths with peak-to-trough amplitudes attaining 1500 nT or more and being sub-linear in character. Anomalies over the Lincoln Sea shelf and on the slope of the Amundsen Basin on the other hand are very subdued, having typical amplitudes of 100 nT or less and wavelengths of 60 km or more. Major isolated high-amplitude linear anomalies are found in three regions along the continental shelf and slope of the Queen Elizabeth Islands and Greenland. The furthest west and the deepest, labelled A–A' in Fig. 4, parallels the Ellesmere Island continental shelf at 83°30'N, 90°W in the region of the 1500+ m channel between Ellesmere Island and the Alpha Ridge (Sobzcak & Sweeney 1978, Johnson et al. 1979). The second (Fig. 4, B–B') trends slightly north of west, approximately parallel to the 1000 m isobath north of Ellesmere Island at 84°N and between 65° and 80°W. The third lies in the shallowest water (Fig. 4 C–C') and trends northwest from 84°N, 30°W, and perpendicular to the Morris Jesup Rise magnetic anomaly (Fig. 4, MJA) reported by Feden et al. (1979) for approximately 100 km before turning north and continuing to 85°20'N, 35°W (Fig. 4, C–C''), where it terminates.

Riddihough et al. (1973) suggested that the Alpha Ridge magnetic trends continue onto the shelf and over the northwest coast of Ellesmere Island on the basis of

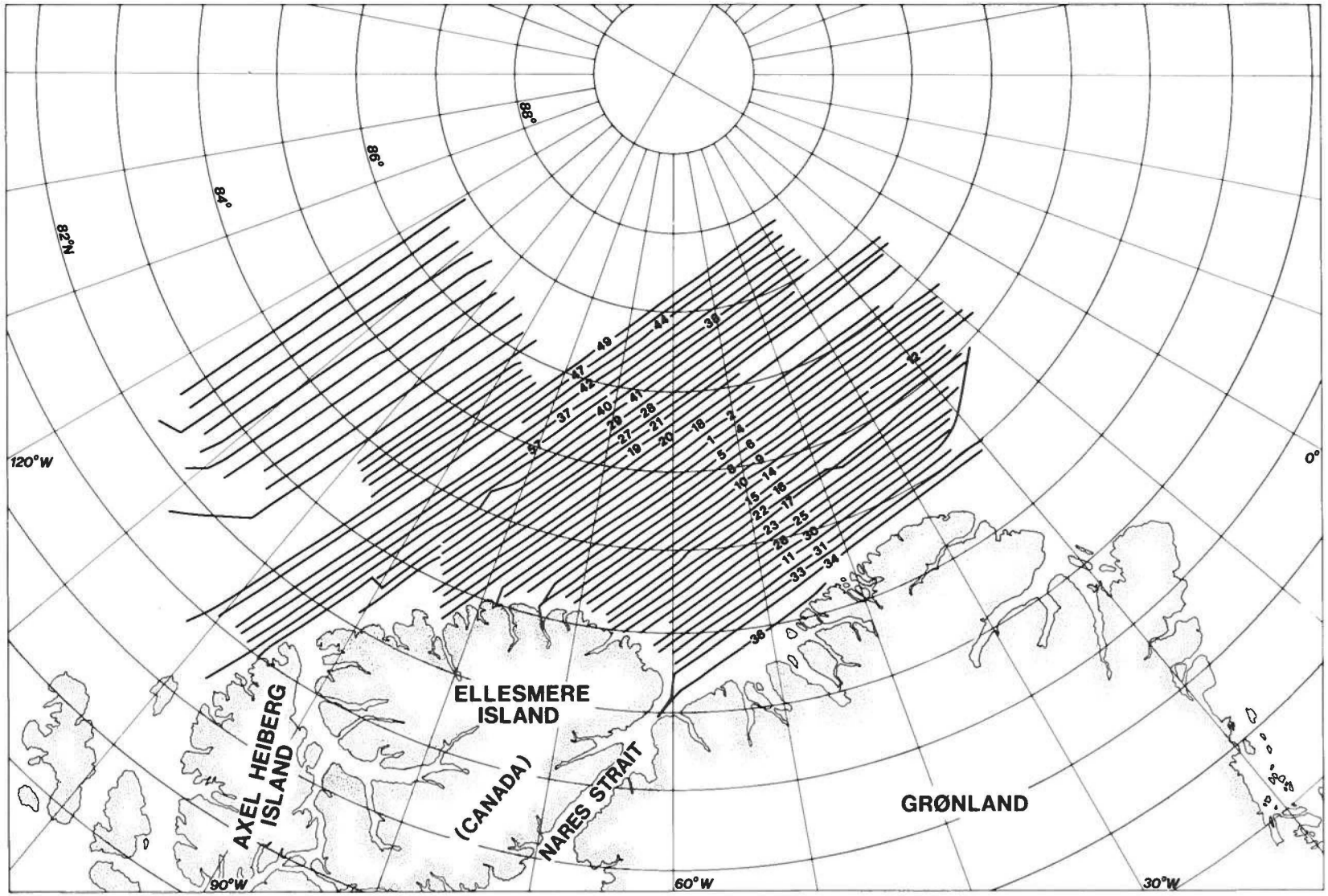


Fig. 2. Aeromagnetic data tracks flown by the Naval Research Laboratory over the Lincoln Sea in 1975. Numbered tracks correspond to numbered data referred to in Figs 7 and 8 and in the text.

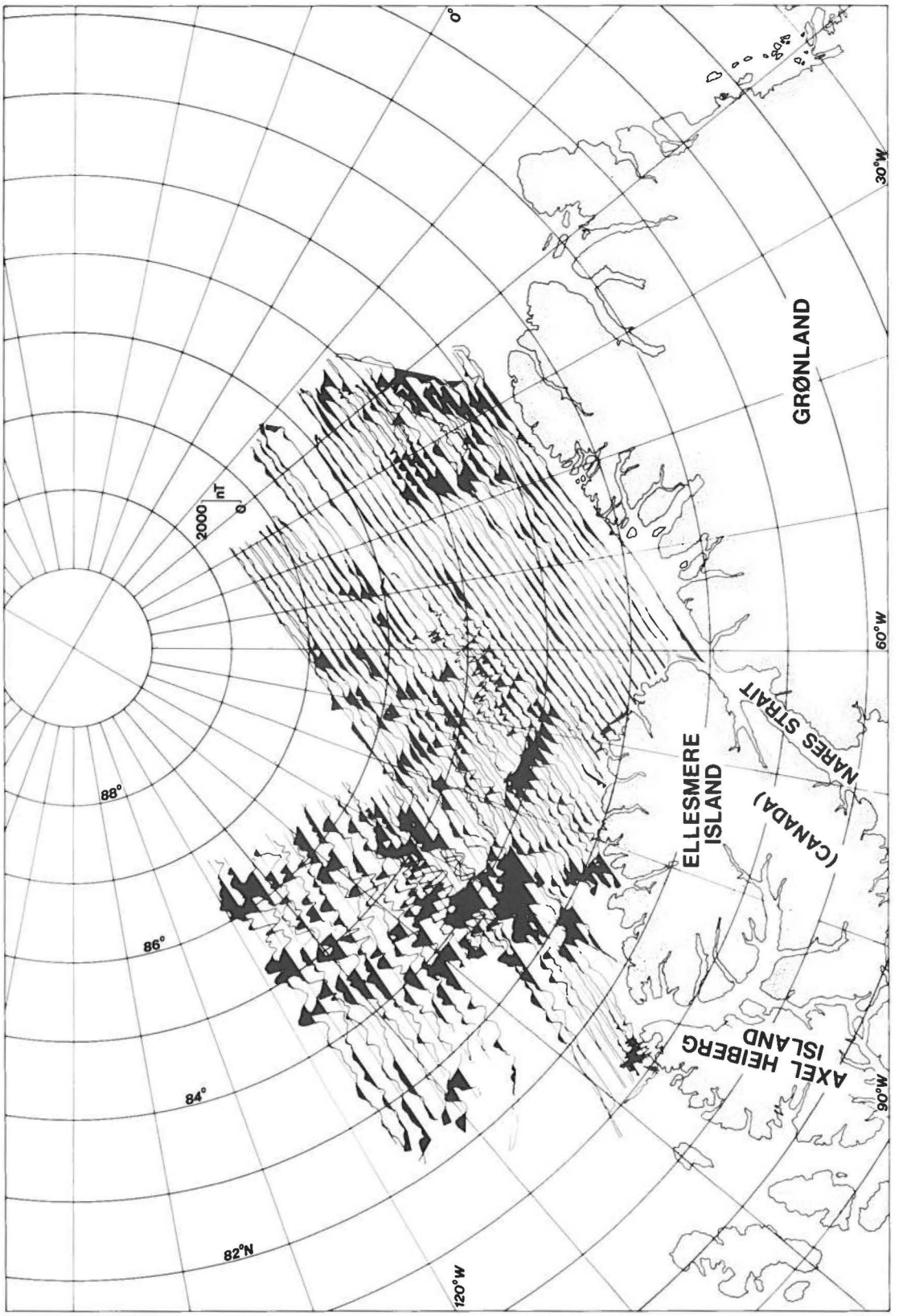


Fig. 3. IGRF–75 residual anomalies over the Lincoln Sea with diurnal variations removed. Black areas are positive anomalies, vertical exaggeration is 4000 nT per degree of latitude.

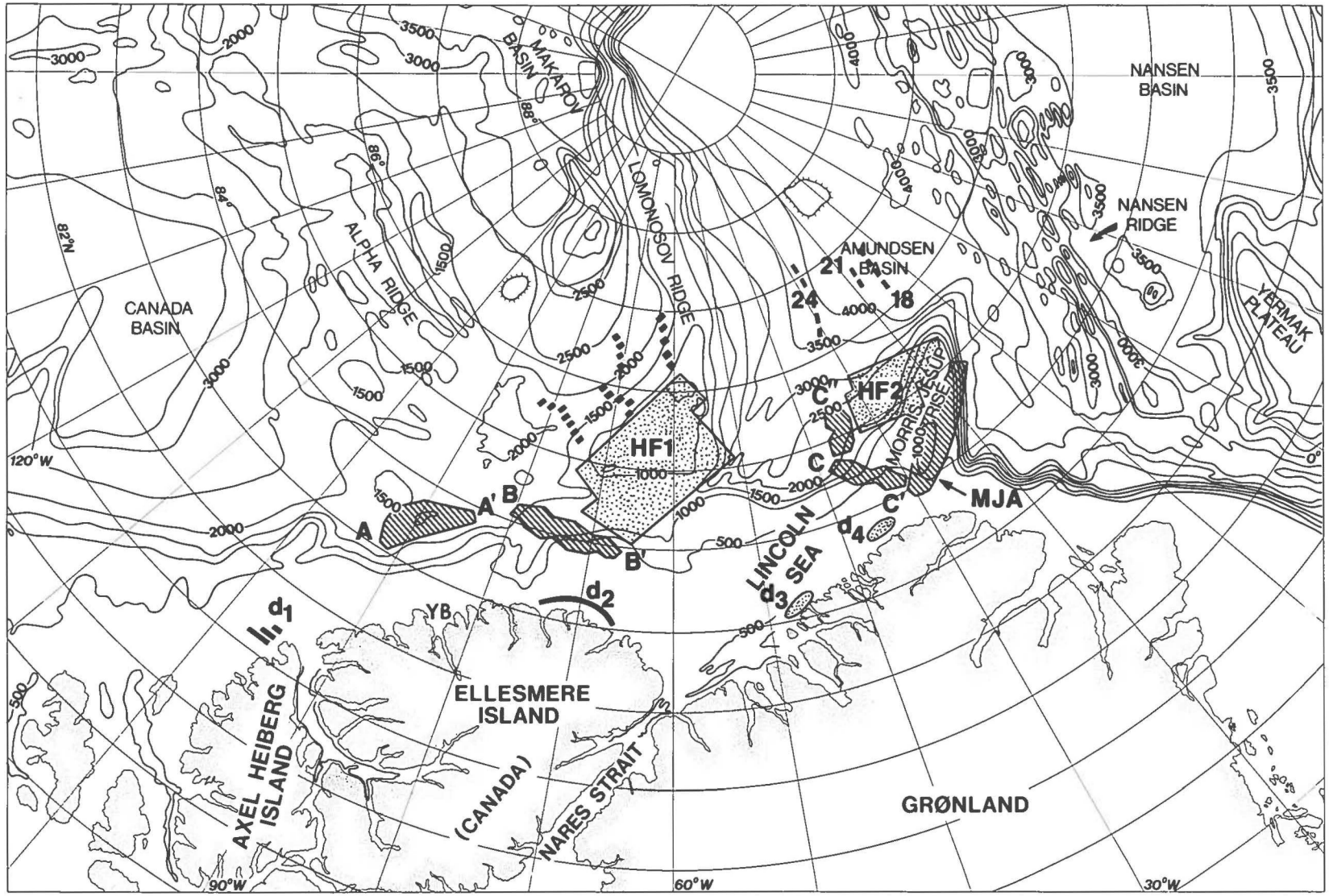


Fig. 4. Major anomaly groups discussed in the text. Stippled areas 'HF1' and 'HF2' are regions of short wavelength moderate amplitude anomalies. Shaded bar 'MJA' is the Morris Jesup Rise anomaly. Short heavy lines labelled 'd_1' and 'd_2' are due to shallow intrusive features (such as dykes). Stippled regions 'd_3' and 'd_4' lie just seaward of the Kap Washington volcanics and are probably associated with them. Unlabelled dashed lines northwest of 'HF1' are tentatively identified in the text as being due to sea-floor spreading. Numbered dashed lines are magnetic lineations identified by Feden et al. (1979) and Vogt et al. (1979a). YB = Yelverton Bay.

vertical magnetic field data collected at high altitude (3.5 km) and wide track spacing (approximately 70 km). The only such magnetic trend appearing in these data is one extending NW from Yelverton Bay at 83°N, 83°W and terminating at the 500 m isobath of the Ellesmere Island shelf. I suggest therefore that the Alpha Ridge group of anomalies is instead bordered by the broad complex high at A–A′ (Fig. 4) which lies at the southeastern end of the Alpha Ridge bathymetric structures.

Sea-floor spreading anomalies 21 to 24 (Feden et al. 1979, Vogt et al. 1979a) of the Nansen Ridge spreading centre are seen in the Amundsen Basin in the northwestern part of the investigation area. They represent the only agreed on spreading anomalies in the area. However, if the Makarov Basin was formed by sea-floor spreading processes as Taylor (1978) and Taylor et al. (1981) suggest, then similar anomalies would be expected to continue eastward toward the slope of the Makarov Basin about 87°N, 90°W and from there southeast into the area discussed in this paper. Unfortunately, this part of the Makarov Basin is the only part of the eastern Amerasia Basin still unsurveyed at low level by U.S. Navy research organizations. Unpublished data from a 1979 Canadian survey of the Lomonosov Ridge (P. J. Hood, pers. comm. 1979) suggest some of these anomalies may continue as discussed, though the situation is confused by a large bathymetric and magnetic high at 87°45′N, 88°W and these unpublished data do not extend across much of the Makarov Basin. Nevertheless, short linear anomalies do extend up the slope of the Makarov Basin parallel to the Lomonosov and Alpha Ridges and gradually blend into a region of moderate amplitude, very short wavelength (< 10 km) anomalies. These high-frequency anomalies (HF1, Fig. 4) strike parallel to B–B′ and extend NE from B–B′ in a band into the broad bathymetric high connecting the Lomonosov Ridge with the Ellesmere Island shelf. A second, similar group of high-frequency, moderate amplitude anomalies (HF2, Fig. 4) appears just east of and parallel to C–C′′ and just west of the Morris Jesup Rise but these are not associated with any obvious bathymetric features in the GEBCO map Arctic bathymetry (Johnson et al. 1979).

Finally, groups of very short wavelength (< 5 km), moderate to high amplitude magnetic anomalies appear off the coasts of Axel Heiberg Island, Ellesmere Island and Greenland. The first group (d_1, Fig. 4), near the northwestern end of Axel Heiberg Island at 81°N, is near coastal Devonian dioritic intrusions (Trettin 1973) and probably associated with them. The second (d_2, Fig. 4) extends parallel to the coast of Ellesmere Island between 65°W and 75°W and is probably due to dykes of indeterminate age. The third and fourth (d_3, d_4, Fig. 4) are of lesser amplitude, lie just west and east, respectively, of the Kap Washington Group volcanics and are probably associated with them (Feden et al. 1979).

Depth-to-source analysis

The tracks labelled in Fig. 2 were analysed for depth to magnetic source using the maximum entropy autocorrelation analysis algorithm of Phillips (1975, 1978). This method assumes that the sources are dykes of semi-infinite extent lying approximately perpendicular to the tracks and having been created by a process such that the magnetization function along the track can be represented as an uncorrelated white noise sequence. Using Burg's maximum entropy spectral analysis techniques (Burg 1975), autocorrelation coefficients for a series of short data sets abstracted from the data are calculated; then, using the well-known relationship between the source depth and the autocorrelation function of a magnetic profile (see, for example, Serson & Hannaford 1957), the source depth is calculated. A complete description of the method can be found in Phillips (1975).

The analysis method is sensitive to two variables: the data point spacing and the number of data points used to calculate the autocorrelation coefficents. The minimum source depth to which the method is sensitive is proportional to the data point spacing; the point spacing should be no more than half the expected minimum source depth. The running window which determines the number of data points used to calculate the autocorrelation coefficients then determines the deepest sources that the method will accurately sense and these should be around 2Z(max)/DELX where DELX is the data point spacing and Z(max) is the maximum depth of interest. The method then calculates multiple depth estimates centred on multiples of this window; inaccurate estimates are rejected on the basis of their lack of close convergence. In the analysis presented here, two separate depth-to-source analyses were made due to the large range in expected source depths: the first at a point spacing and window of 1 and 30 kilometres, respectively (sensitive to source depths ranging from 2 to 15 km), and the second at a point spacing and window of .25 and 7.5 km, respectively (sensitive to depths from .5 to 3.75 km). Though the method would be expected to be sensitive to the strike of magnetic features in relation to the data tracks, in practice most real anomaly data contain three dimensional source information which tends to counteract this. Depth-to-source data calculated from aeromagnetic investigations over the Mohns and Knipovich Ridges (Kovacs & Vogt 1979) have shown excellent agreement with bathymetric and seismic data when the data track was as much as 30 to 45 degrees non-perpendicular to the magnetic fabric. Uncertainties in the calculated source depths are typically one third the data point spacing or less, or 80 m and 300 m respectively for the two analyses made. The results from these analyses were then combined and contoured and are presented in raw (Fig. 5) and generalized forms (Fig. 6).

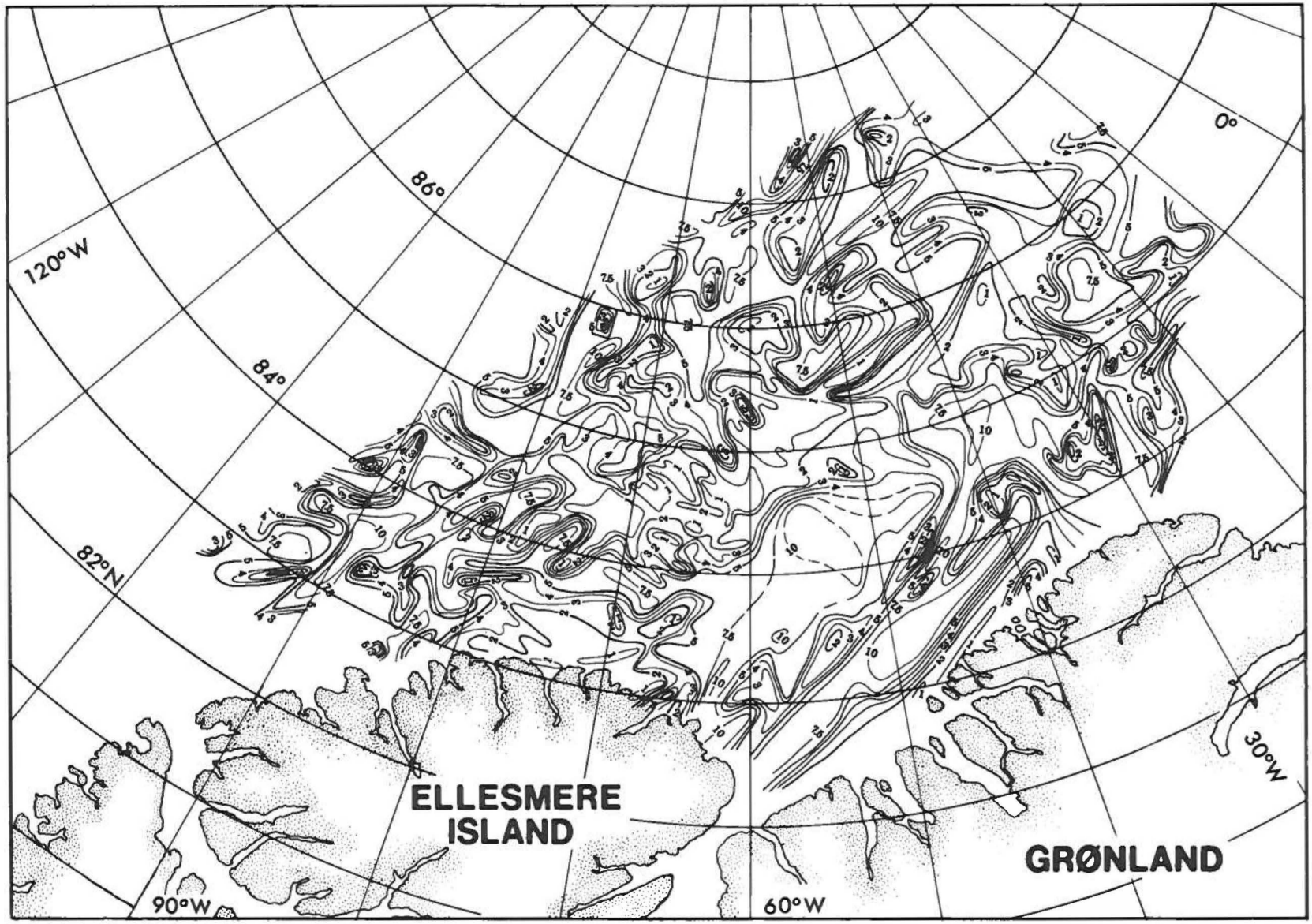


Fig. 5. Contoured depth-to-source magnetic data calculated from the numbered tracks in Fig. 1. Depths used are a combination of results calculated from data sampled every 0.25 km (depths <4 km) and data sampled every 1.0 km (depths >2 km). Contour interval is 1 km from 0 to 5 km, 2.5 km from 5 to 10 km.

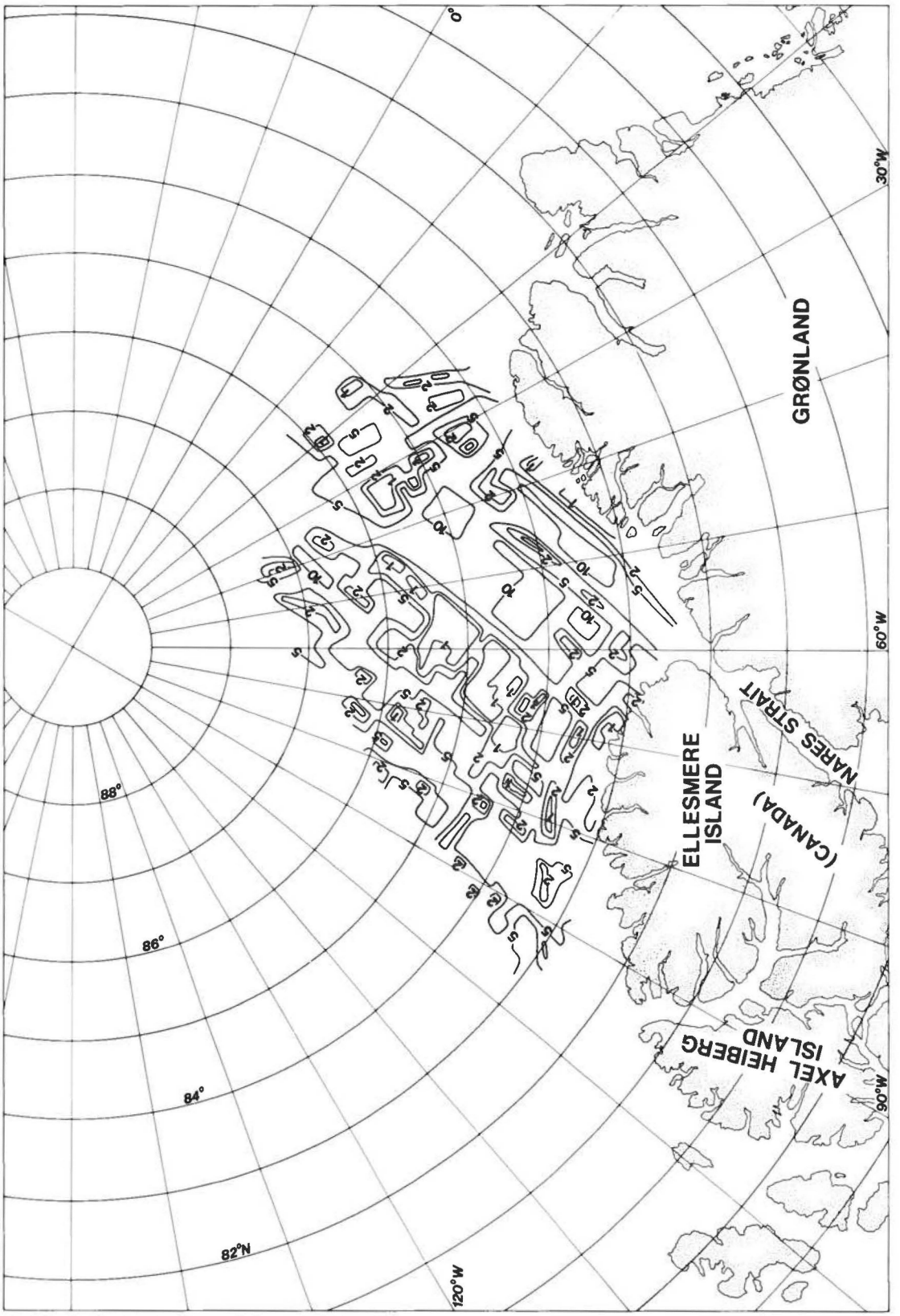


Fig. 6. Generalized depth-to-source data from Fig. 5. 1 km, 2 km, 5 km and 10 km contours.

Discussion and interpretation

In Figs 5 and 6, two major source-depth structures are immediately apparent: a linear region of deep (> 10 km) sources extending some 300 km from Nares Strait toward the Morris Jesup Rise, and a large source-depth high (< 2 km deep) under the short wavelength linear anomalies (HF1) found to the northeast of B–B′ (Fig. 4), and sub-parallel to the previously mentioned source depth low. Source depth highs also exist under the Morris Jesup Rise and along the coasts of Greenland and Ellesmere Island. Additionally, a group of highs are found northwest of the Morris Jesup Rise and northeast of and parallel to C–C′′, under HF2 (Fig. 4). Linear regions of deep sources are found northwest of this region and northwest of and parallel to the large magnetic source high northeast of B–B′. A basin of deep sources may be found centred at 84°N, 55°W between the deep source extension of Nares Strait and the source high northeast of B–B′. Finally, isolated source highs such as the dykes mentioned previously (Fig. 4) are found along the continental shelf of Ellesmere Island and in scattered highs along the Alpha Ridge.

On the basis of their close association with the slope or break in the shelves of Ellesmere Island and Greenland the linear anomaly B–B′ (Figs 4 and 7) is postulated to mark the continent–ocean boundary, similar to the East Coast Magnetic Anomaly discussed by Taylor et al. (1968). C–C′ (Figs 4 and 7) is similarly taken to be associated with the edge of Greenland's continental shelf, rather than with the Morris Jesup Rise anomaly because of its difference in strike and close association with the short wavelength magnetic features similar to HF1 that lie west of the Morris Jesup Rise. If, as postulated by Srivastava (1978) and Jackson et al. (1979), 250 km of left-lateral motion occurred along Nares Strait during the opening of Baffin Bay, then rotating Greenland back to close Baffin Bay brings the isolated magnetic highs B–B′ and C′′–C′ into alignment (Fig. 7B). This similarly brings the two regions (HF1 and HF2, Fig. 4) of moderate amplitude, short wavelength anomalies that lie northeast of B–B′ and C–C′ into alignment. The linear magnetic source deep, found in the depth-to-source analysis (Figs 5 and 6) to extend from and in line with Nares Strait, is therefore taken to be additional evidence of a major transform fault along Nares Strait that extends into the Lincoln Sea to the Morris Jesup Rise. This is labelled the Wegener Fracture Zone in Fig. 9, following Johnson & Vogt (1973) and Ostenso & Wold (1973).

Similarly, a second fracture zone is taken to lie along the boundary between the magnetic basement high under HF1 (Fig. 4) and the magnetic basement low southeast of HF1 and southwest of C–C′′, and coincident with the channel-like bights in the 2000, 2500, 3000 and 3500 m GEBCO map sheet 5.17 isobaths of the Lomonosov Ridge – Amundsen Basin slope (Johnson et al. 1979). This fracture zone, which I call here the Ellesmere Fracture Zone (Fig. 9) extends from its southwestern terminus in the magnetic basement at the 500 m isobath to its northeastern end just seaward of the 3500 m isobath of the Amundsen Basin near the anomaly 24 isochron (Fig. 4). The region of subdued, long wavelength anomalies bordered by the Ellesmere Island shelf, the Ellesmere and Wegener Fracture Zones and the magnetic basement high north and east of C′′–C–C′ that is characterized in Figs 5 and 6 by moderately deep magnetic basement could then be a region of thinned and foundered continental crust, rather than oceanic crust, formed by the left-lateral motion of Arctic Canada in relation to Greenland along the Wegener and Ellesmere Fracture Zones.

A close examination of the magnetic anomalies in regions HF1 and HF2 (Fig. 8) shows some features characteristic of oceanic sea-floor spreading anomalies, including bilaterally symmetric anomalies equidistant from an occasionally visible central magnetic high. This central high also corresponds closely to a 70 km long by 20 km wide low in the middle of the depth-to-source data for region HF1 (Figs 5 and 6). An unabashed plate tectonics interpretation of the magnetic evidence in these areas then suggests that these high-frequency anomalies are sea-floor spreading anomalies with a fossil spreading centre lying under the starred region shown in Fig. 8, and that the following plate tectonic sequence took place. Spreading postulated to have begun in the Santonian or early Campanian (Taylor 1978, Taylor et al. 1981) in the Makarov Basin extended up the slope of the northeastern Ellesmere Island – northern Greenland shelf and terminated in a transition region similar to the present-day Gulf of Tadjurah (Courtillot et al. 1980) at the Kap Washington Group volcanics, beginning the formation of a shallow, possibly emergent, plateau of oceanic crust (Fig. 9A). This was followed shortly by the initiation of spreading in the Labrador Sea in the Maastrichtian at anomaly 32 (Srivastava 1978). By anomaly 25 in the mid-Paleocene (Fig. 9B) spreading had begun in Baffin Bay (Srivastava 1978, Jackson et al. 1979) and by late Paleocene prior to anomaly 24, spreading had begun in the Norwegian–Greenland Sea and Eurasia Basin (Pitman & Talwani 1972, Johnson & Vogt 1973, Herron et al. 1974, Kristoffersen & Talwani 1977, Srivastava 1978, Feden et al. 1979, Jackson et al. 1979, Vogt et al. 1979a, b). By this time, the shallow plateau over what I call here the Kap Washington hot spot was fully formed and was beginning to be split into two pieces (HF1, HF2, Fig. 4) by the formation of the Wegener–Ellesmere fault complex that formed the southwest arm of the RFF triple junction north of Greenland (Fig. 9C). This strike-slip motion between HF1 and HF2 was accompanied by thinning and foundering of the continental crust between the Ellesmere and Wegener Fracture Zones and between HF2 and Ellesmere Island, forming a deep sedimentary basin. Shortly after spreading ceased in the Makarov Basin, the Lomonosov Ridge became fixed to

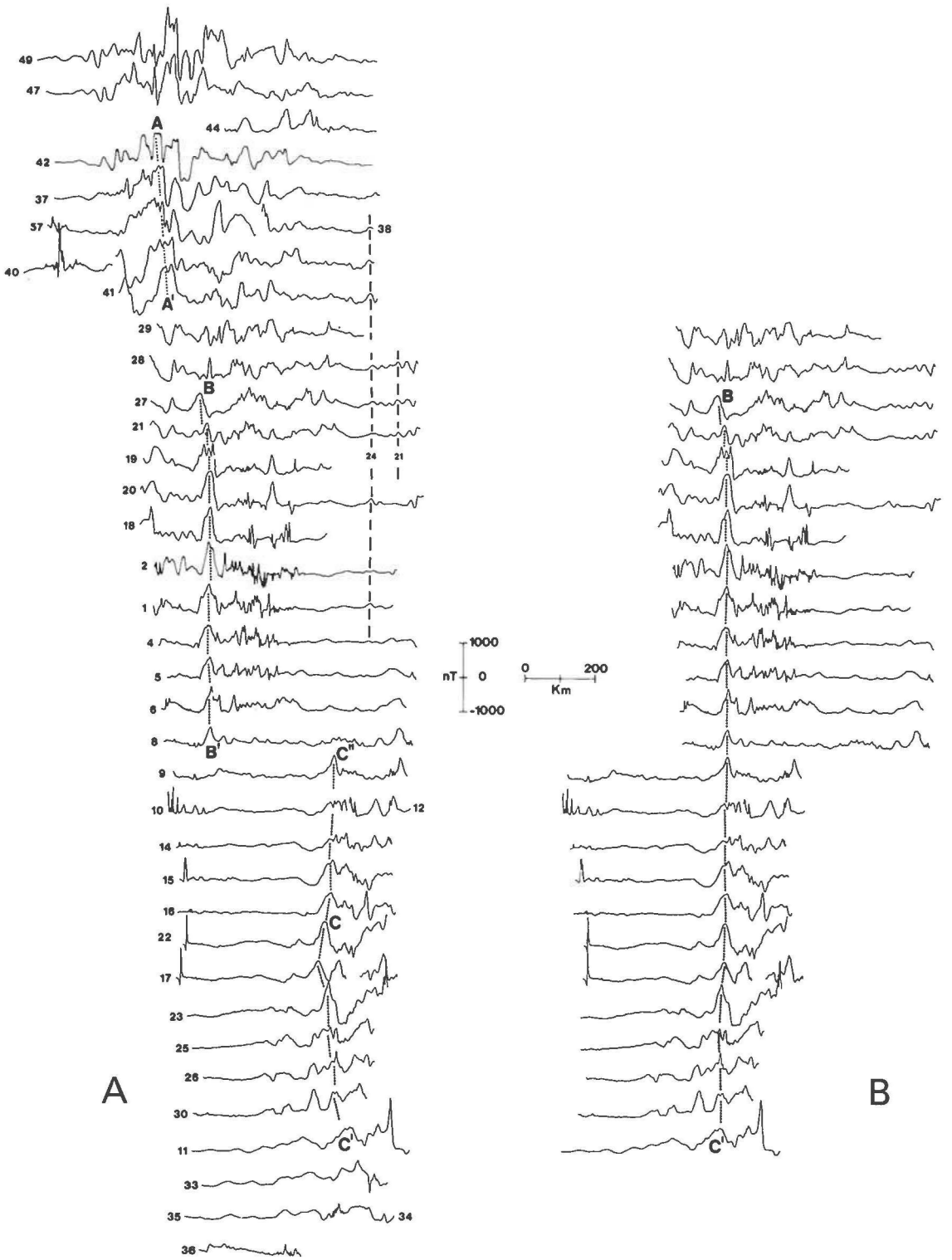


Fig. 7. A: Anomaly versus distance profiles for the tracks numbered in Fig. 2. Southern ends of all tracks are on the left. Features are aligned using a line on Fig. 2 from the intersection of track 49 and 70°W long. to the intersection of track 36 and 45°W long. Dotted lines connect continental edge anomalies A–A′, B–B′ and C′′–C–C′ shown in Fig. 4. Dashed lines are anomalies 21 and 24 as identified by Feden et al. (1979) and Vogt et al. (1979a). B: Anomaly versus distance profiles for track numbers 29 through 36, with tracks 9 through 36 shifted left (south) to align C–C′′ with B–B′.

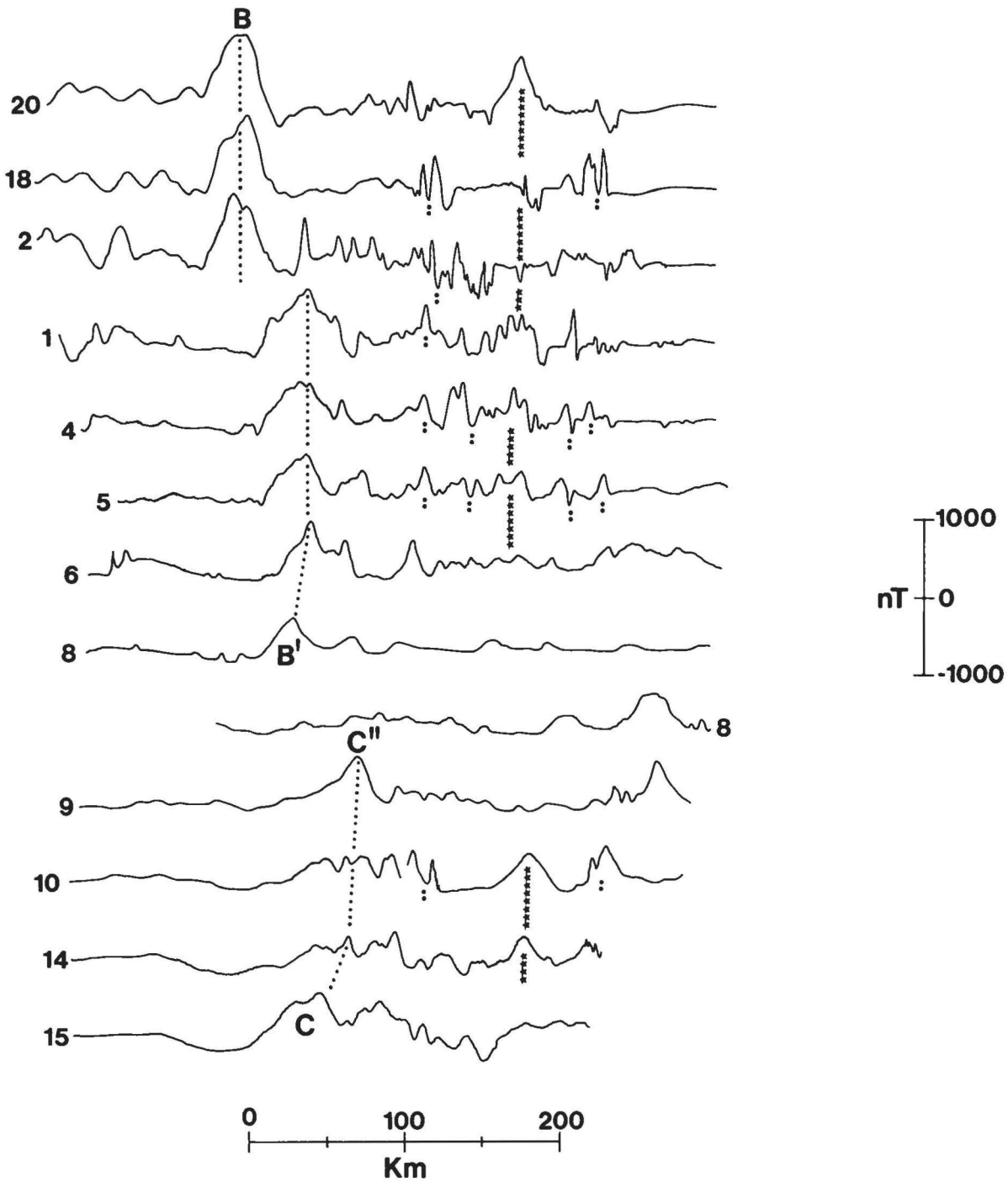


Fig. 8. Anomaly versus distance profiles for tracks 20 through 15 (location in Fig. 2), aligned as in Fig. 7B, but with track 8 cut in the middle and magnetic highs aligned. Postulated relic central anomaly is starred. representative anomalies having bilateral symmetry are dotted. Note the change in relative scale compared to Fig. 7.



Fig. 9. Suggested plate tectonic evolution of the Lincoln Sea region. AR – Alpha Ridge; LR – Lomonosov Ridge; EI – Ellesmere Island; G – Greenland; S – Svalbard; single dashed lines denote active fracture zones; double dashed lines denote landward extensions of active spreading centres. A: Anomaly 34 time, 80 m.y. B: Anomaly 25 time, 59 m.y. C: Anomaly 24 time, 56 m.y. D: Anomaly 13 time, 35 m.y. Age of magnetic anomalies based on time scale of LaBrecque et al. (1977).

Ellesmere Island and Canada across the plateau forming HF1, and the position of the Kap Washington hot spot relative to Greenland shifted slightly eastward of HF2 beginning the formation of the Morris Jesup Rise–Yermak Plateau (Feden et al. 1979, Vogt et al. 1979b). By anomaly 13 time in the early Oligocene (Fig. 9D) spreading in Baffin Bay and the Labrador Sea (Srivastava 1978, Jackson et al. 1979) and strike-slip motion along the Wegener–Ellesmere Faults ceased, the anomalous production of basalt forming the Morris Jesup Rise and Yermak Plateau abruptly ended (Feden et al. 1979) and the basement morphology of the Makarov Basin, Lomonosov Ridge, Lincoln Sea and Morris Jesup Rise became essentially fixed. Continued spreading along the North Atlantic and Arctic spreading ridges (Karasik et al. 1972, Pitman & Talwani 1972, Ostenso & Wold 1973, Herron et al. 1974, Pitman & Herron 1974, Talwani & Eldholm 1977, Srivastava 1978, Sweeney et al. 1978, Feden et al. 1979, Vogt et al. 1979a, b) then led to the formation of the present-day structures.

As in any preliminary hypothesis derived primarily from aeromagnetics and unconstrained by seismic or drill hole data, several problems exist in this interpretation. Srivastava (1978) postulated a northwest component to the motion of Greenland, towards Ellesmere Island in the late Cretaceous and early Tertiary to form the Innuitian fold belt during the Eurekan orogeny; this movement of Greenland is further discussed by Srivastava & Falconer (this volume). If such a compressive motion took place there is no record of it in this author's interpretation of the Lincoln Sea aeromagnetics. In fact, depending on the particular sequence of motions used to rotate Greenland back against Arctic Canada, and the choice of the strike of the anomalies in the area HF2 (Fig. 4), a gap of as much as 50 to 80 km between the supposedly once contiguous blocks HF1 and HF2 appears! This would tend to imply a separation of Greenland from Ellesmere Island by this amount, rather than compressive motion.

Another problem is whether the basement under HF1 and HF2 is totally oceanic, and formed by plate tectonic and hot spot/plume processes. The anomaly record, though appearing very similar to the partially lineated high-frequency anomalies just north of Iceland (Vogt et al. 1980), is too spotty for any reasonable anomaly identifications to be made. In fact, the above interpretation may be overly enthusiastic, a possible alternative being the generation of numerous dykes in the regions HF1 and HF2 which coincidentally formed structures having occasional bilateral symmetry. There is also no bathymetric feature clearly associated with HF2, though the detailed magnetics and the close correspondence of magnetic features suggest this is more likely due to the lack of detailed bathymetric data in the area. A reinterpretation of the bathymetry incorporating the magnetics and depth-to-source data is given in Fig. 10.

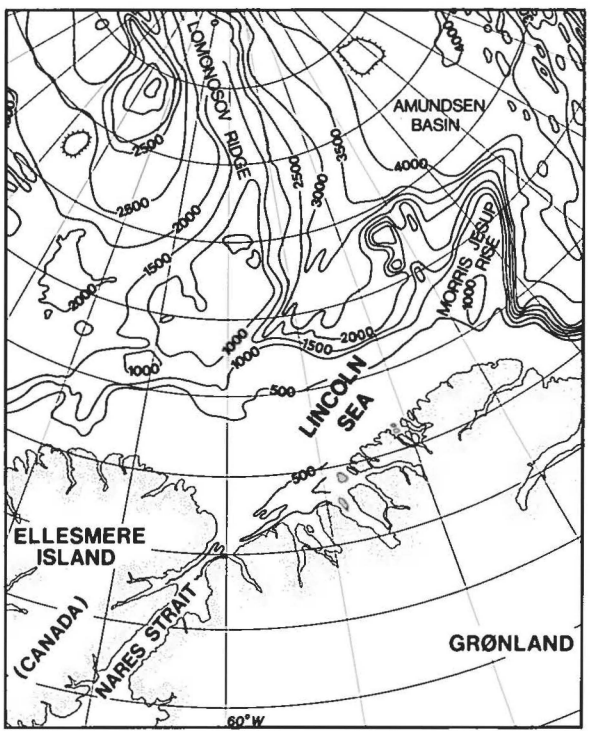


Fig. 10. Revised Lincoln Sea bathymetry (cf. Fig. 1), using discussed depth-to-source and magnetic data, and GEBCO sheet 5.17 (Johnson et al. 1979). Bathymetry in metres.

Another problem with a spreading history in the region northeast of B–B′ (Fig. 4) is its lack of subsidence and the fact that it forms the transition between the Ellesmere Island continental shelf and the generally acknowledged continental structures of the Lomonosov Ridge. Oceanic crust between the Lomonosov Ridge and the Greenland/Ellesmere Island shelf certainly cannot be ruled out, given the complex history of the area, but if it exists at the source depths calculated a large free air gravity anomaly would be expected. Such an anomaly does not appear in the available regional data (Sobczak 1978). This, however may be due to the low pass filtering implicit in its method of presentation (Sobczak 1978), since a free air high of about 60 mgal is shown on one line of a detailed gravity survey of the Lincoln Sea (Sobczak & Stephens 1974). All the processes postulated would be characterized by very low spreading rates and fluctuating rates of basalt production as discussed by Feden et al. (1979) and Jackson et al. (1979), which could be responsible for the relatively shallow and equivocal character of the anomalies. Certainly, other interpretations are possible and theories on the origin of the area will continue to be revised as more geophysical data become available. Hopefully, the interpretation offered here can at least serve as a guide to future work in the Lincoln Sea and its geotectonic connections to the surrounding Arctic regions.

Conclusions

On the basis of a depth-to-source analysis of previously unpublished aeromagnetic data in the Lincoln Sea, a major deep magnetic basement extension of the Nares Strait is delineated and identified as the Wegener Fracture Zone. Shelf edge anomalies and two anomalously shallow blocks of possible oceanic anomalies lying perpendicular to the fracture zone are identified; an approximately 250 km displacement along the fracture zone is postulated to explain the offset in these magnetic features. Another fracture zone is delineated, along with a major sediment filled basin whose deep magnetic basement structure is due to thinning and foundering of the underlying continental crust.

The anomalously shallow oceanic anomalies found northeast of the shelf edge anomalies are postulated to be due to an Iceland-like hot spot and to the very slow spreading rates and complex history of the area, though the magnetic record is not unequivocal enough to rule out a swarm of dykes in an otherwise continental structure.

Acknowledgements

The author thanks R. H. Feden and the officers and crew of the Naval Research Laboratory Flight Detachment for their collection of the data presented here. I also thank R. H. Feden, H. S. Fleming, G. L. Johnson, R. K. Perry and P. R. Vogt for their numerous helpful discussions. P. R. Vogt, G. L. Johnson and R. H. Feden reviewed the manuscript; I thank them and an anonymous reviewer for their many constructive comments. M. F. Phillips and B. Herron did much of the computer processing of the data; C. Rockelli did the typing of the manuscript; I. Jewett drafted the figures. This work was supported by the Office of Naval Research, Arctic Program under contract RR032–05–01.

References

Anonymous 1976. International geomagnetic reference field, 1975. – Trans. Am. geophys. Un. 57: 120–121.

Burg, J. P. 1975. Maximum entropy spectral analysis. – Unpubl. Ph. D. thesis, Stanford Univ., U.S.A.: 123 pp. (available University Microfilms International, Ann Arbor, Michigan).

Churkin, M. 1973. Geologic concepts of Arctic Ocean Basin. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 485–499.

Coles, R. L., Hannaford, W. & Haines, G. V. 1978. Magnetic anomalies and the evolution of the Arctic. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 51–66.

Courtillot, V., Galdeano, A. & Le Mouel, J. L. 1980. Propagation of an accreting plate boundary: a discussion of new aeromagnetic data in the Gulf of Tadjurah and southern Afar. – Earth planet. Sci. Lett. 47: 144–160.

Dawes, P. R. 1973. The North Greenland fold belt: a clue to the history of the Arctic Ocean basin and the Nares Strait lineament. – In: Tarling, D. H. & Runcorn, S. K. (eds), Implications of continental drift to the earth sciences 2: 925–947. – Academic Press, London & New York.

Feden, R. H., Vogt, P. R. & Fleming, H. S. 1979. Magnetic and bathymetric evidence for the "Yermak hot spot" northwest of Svalbard in the Arctic Basin. – Earth planet. Sci. Lett. 44: 18–38.

Herron, E. M., Dewey, J. F. & Pitman, W. C. 1974. Plate tectonics model for the evolution of the Arctic. – Geology 2: 377–380.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor spreading in central Baffin Bay. – Can. J. Earth Sci. 16: 2122–2135.

Johnson, G. L. & Vogt, P. R. 1973. Marine geology of Atlantic Ocean north of the Arctic circle. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 161–170.

Johnson, G. L., Monahan, D., Grønlie, G. & Sobczak, L. [W.] 1979. General bathymetric chart of the oceans (GEBCO), 1:6 000 000, Sheet 5.17, The Arctic Ocean. – Can. Hydrogr. Serv., Ottawa.

Karasik, A. M., Demenitskaya, R. M. & Shelovanov, V. G. 1972. Specific character of magnetic field and development of undersea ridges of the Arctic Basin. – Geologiya i Perspektiuy Neftegazononosti Sovetskoy Arktiki. – U.S. Dept. Air Force, WP–AFB, Ohio, Translation FTD–ID (RS) T–0797–80: 51–55. (From Russian).

Kerr, J. W. 1980. Did Greenland drift along Nares Strait? – Bull. Can. Petrol. Geol. 28: 279–289.

Kovacs, L. C. & Vogt, P. R. 1979. Depth-to-magnetic source analysis of the Greenland/Norwegian Sea and Arctic Ocean. – Trans. Am. geophys. Un. 60: 372 only.

Kristoffersen, Y. & Talwani, M. 1977. Extinct triple junction south of Greenland and the Tertiary motion of Greenland relative to North America. – Bull. geol. Soc. Am. 88: 1037–1049.

LaBrecque, J. L., Kent, D. V. & Cande, S. C. 1977. Revised magnetic polarity time scale for Late Cretaceous and Cenozoic time. – Geology 5: 330–335.

Larsen, O., Dawes, P. R. & Soper, N. J. 1978. Rb/Sr age of the Kap Washington Group, Peary Land, North Greenland, and its geotectonic implication. – Rapp. Grønlands geol. Unders. 90: 115–119.

Ostenso, N. A. & Wold, R. J. 1973. Aeromagnetic evidence for origin of Arctic Ocean Basin. – In: Pitcher, M. G. (ed.), Arctic geology. – Am. Ass. Petrol. Geol. 19: 506–516.

Phillips, J. D. 1975. Statistical analysis of magnetic profiles and geomagnetic reversal sequences. – Unpubl. Ph. D. thesis, Stanford Univ., U.S.A.: 134 pp. (available University Microfilms International, Ann Arbor, Michigan).

Phillips, J. D. 1978. ADEPT: a program to estimate depth to magnetic basement from sampled magnetic profiles. – U.S. geol. Surv. Computer Contribution, Reston, Virginia: 35 pp.

Pitman, W. C. & Herron, E. M. 1974. Continental drift in the Atlantic and Arctic. – In: Kristjansson, L. (ed.), Geodynamics of Iceland and the North Atlantic area: 1–15. – D. Reidel Publ. Co., Dordrecht, Holland.

Pitman, W. C. & Talwani, M. 1972. Sea-floor spreading in the North Atlantic. – Bull. geol. Soc. Am. 83: 619–646.

Riddihough, R. P., Haines, G. V. & Hannaford, W. 1973. Regional magnetic anomalies of the Canadian Arctic. – Can. J. Earth Sci. 10: 157–163.

Sclater, J. G., Hellinger, S. & Tapscott, C. 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. – J. Geol. 85: 509–552.

Serson, P. H. & Hannaford, W. L. W. 1957. A statistical analysis of magnetic profiles. – J. geophys. Res. 62: 1–18.

Sobzcak, L. W. 1978. Gravity from 60°N to the Pole. – In: Sweeney, J. R. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 67–74.

Sobczak, L. W. & Stephens, L. E. 1974. The gravity field of northeastern Ellesmere Island, part of northern Greenland and Lincoln Sea with map. Lincoln Sea 1:500 000. – Earth Physics Branch, Gravity Map Series 114: 9 pp.

Sobzcak, L. W. & Sweeney, J. F. 1978. Bathymetry of the Arctic Ocean. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 7–14.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Srivastava, S. P. & Falconer, R. K. H. 1982. Nares Strait: a conflict between plate tectonic predictions and geological interpretation. – This volume.

Sweeney, J. F., Irving, E. & Geuer, J. W. 1978. Evolution of the Arctic Basin. –In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 91–100.

Talwani, M. & Eldholm, O. 1977. Evolution of the Norwegian-Greenland Sea. – Bull. geol. Soc. Am. 88: 969–999.

Taylor, P. T. 1978. Low-level aeromagnetic data across the western Arctic Basin. – Trans. Am. geophys. Un. 59: 268–269.

Taylor, P. T., Zietz, I. & Dennis, L. S. 1968. Geologic implications of aeromagnetic data for the eastern continental margin of the United States. – Geophysics 33: 755–780.

Taylor, P. T., Kovacs, L. C., Vogt, P. R. & Johnson, G. L. 1981. Detailed aeromagnetic investigation of the Arctic Basin, 2. – J. geophys. Res. 86: 6323–6333.

Trettin, H. P. 1973. Early Paleozoic evolution of northern parts of Canadian Arctic Archipelago. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 57–75.

Vogt, P. R., Feden, R. H., Eldholm, O. & Sundvor, E. 1978. The ocean crust west and north of the Svalbard Archipelago. – Polarforschung 48: 1–19.

Vogt, P. R., Taylor, P. T., Kovacs, L. C. & Johnson, G. L. 1979a. Detailed aeromagnetic investigation of the Arctic Basin. – J. geophys. Res. 84: 1071–1089.

Vogt, P. R., Kovacs, L. C., Johnson, G. L. & Feden, R. H. 1979b. The Eurasia Basin. – Proc. Norwegian Sea symposium, 1979, NSS–3: 1–29. – Norsk Petroleum Forening, Oslo.

Vogt, P. R., Johnson, G. L. & Kristjansson, L. 1980. Morphology and magnetic anomalies north of Iceland. – J. Geophys. 47: 67–80.

Wegener, A. 1922. Die Entstehung der Kontinente und Ozeane. – Friedr. Vieweg & Sohn, Braunschweig: 144 pp. (3rd revised edit.).

Satellite magnetic field over the Nares Strait region

ROBERT A. LANGEL and LEIF THORNING

Langel, R. A. & Thorning, L. 1982. Satellite magnetic field over the Nares Strait region. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 291–293.

The general characteristics of the total field magnetic anomalies from satellite data over the region adjacent to Nares Strait are briefly described. The interpretation of satellite data is still at an early stage, but it is favoured that the regional anomalies in this region reflect variations in composition and/or structure in the lower crust related to the Innuitian Ellesmere–Greenland mobile belt. An elongated negative anomaly occurs over Ellesmere Island parallel to the Innuitian mobile belt and to Nares Strait. This negative anomaly separates major magnetic highs over northern Greenland and the Alpha Ridge of the Canada Basin. The magnetic contrast is interpreted to represent either a change in the thickness of the Precambrian crust or a petrologic change brought about during the formation of the Innuitian mobile belt.

R. A. Langel, Geophysics Branch, National Aeronautics and Space Administration, Goddard Space Flight Center, Greenbelt, Maryland 20771, U.S.A. and L. Thorning, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK-1350 Copenhagen K, Denmark.

During the last few years increasing attention has been directed towards the processing and interpretation of satellite magnetic anomaly fields, and a number of promising results have been presented (see Langel & Thorning in press, for full references). The information used has been OGO data (Orbiting Geophysical Observatories), i.e. scalar total magnetic field measured at altitudes from 400 to 600 km by the OGO 2, 4 and 6 satellites (Cain & Langel 1971, Langel 1974, 1980). This note draws attention to some interesting aspects of the satellite magnetic field over the Nares Strait region (Fig. 1). The nearby areas of Canada and Greenland are treated in more detail by Langel et al. (1980) and Langel & Thorning (in press), respectively.



Fig. 1. Index map of the Nares Strait region. The approximate southern limit of the Innuitian mobile belt is indicated (from Trettin et al. 1972). NP = North Pole.

The magnetic data

The magnetic field measured by the OGO satellites is comprised of several components: the Earth's core field, magnetic field components originating in the crust, external fields from the magnetosphere and ionosphere, and instrument noise. It is possible to isolate those components of the magnetic total field that are thought to have sources in the Earth's crust (Mayhew 1979). The main field has been subtracted using a model of degree and order thirteen derived from the satellite data itself (Langel et al. 1980). All individual passes have been evaluated, and passes showing variations exceeding approximately 20 nT probably caused by magnetospheric and ionospheric sources have been entirely omitted from further analysis. Remaining passes normally still exhibit fields of a few thousand kilometres wavelength of uncertain origin. These were removed by the fitting and subtraction of a linear function to each of the passes. As a consequence the absolute zero level of the data is floating. The resultant magnetic field represents anomalies of crustal origin.

Fig. 2 shows a contoured map of the reduced data over the Nares Strait region redrawn from a larger map

of the entire polar region that is presented in Langel & Thorning (in press).

Interpretation of anomalies

Detailed discussions of the interpretation of satellite magnetic data are given in several of the papers already quoted, and will not be included here. It is sufficient to say that it is a field of geophysics only recently opened up, and much more work is necessary before any final answers can be expected. The forthcoming data from MAGSAT satellites, which are much more detailed than the OGO data hitherto used, will make the work even more rewarding.

Although the map (Fig. 2) is an average anomaly map and thus represents data from varying altitude, the general trends in the magnetic data are quite clear: two magnetic highs divided by an elongated low. The two highs are the North Greenland magnetic anomaly and the Alpha Ridge magnetic anomaly, and both of these are major ones seen on a global scale. The trend of the negative anomaly correlates well with the position of the Innuitian Ellesmere – North Greenland mobile belt.

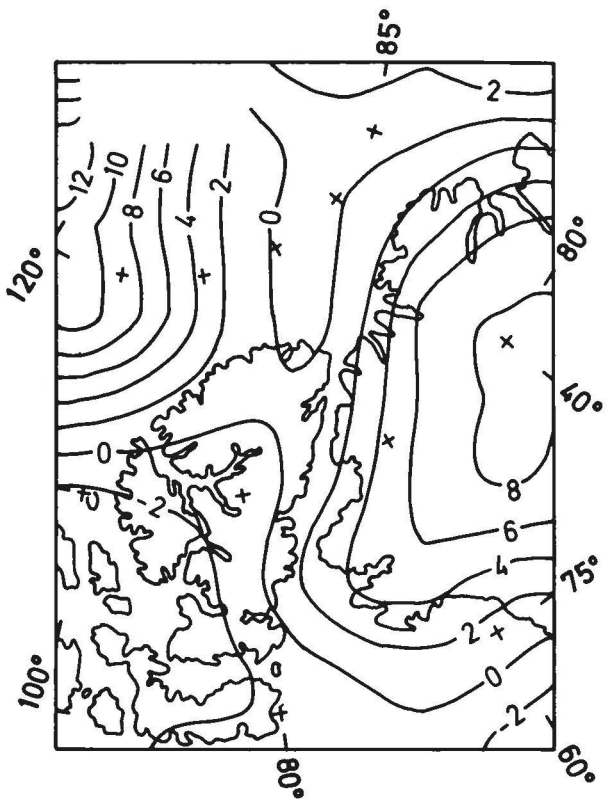


Fig. 2. Satellite magnetic field over the Nares Strait region as derived from OGO satellite measurements. The contours represent average magnetic anomalies in nT, and the data have not been processed into an equivalent source representation. Redrawn from Langel & Thorning (in press).

The sources of such long wavelength anomalies lie in the crust. The mantle just below the Moho is dominantly of non-magnetic mineralogy, no matter what the geothermal gradient (Wasilewski et al. 1979). Where the depth to the Curie isotherm is less than the depth to the Moho, the Curie isotherm defines the lower boundary of the magnetic crust. Thus, the magnetic anomaly fields are a result of variations in crustal thickness, average crustal magnetisation and depth to the Curie isotherm. To date only a few of the regional-scale anomalies delineated in anomaly maps from satellite data have been modelled. Of the models derived, two (Regan & Marsh 1982, Mayhew et al. 1982) indicate distinct regions of intrusions of highly magnetic material into the crust from the mantle. This, however, is not true for all anomalies and Mayhew (in press) has indicated that the anomalies in the western U.S. are anti-correlated with heat flow and probably largely reflect variations in the depth to the Curie isotherm. It would thus seem that many, if not most, anomalies of this scale reflect the structure and petrology of the *lower* crust or variations in the isotherm.

Relation to Nares Strait

Nares Strait, joining Baffin Bay to the Arctic Ocean, is considered by most workers to be the site of a dislocation zone of some importance. Taylor (1910) first suggested the idea of substantial strike-slip motion along the Strait. A late Phanerozoic age for the movement has been favoured by advocates of large displacement between Greenland and Ellesmere Island (e.g. Srivastava 1978 and several papers in this volume). However, it has also been argued (e.g. Burek 1973, Dawes 1973, Fahrig et al. 1973) — notably from different lines of evidence — that Nares Strait is the site of a much older, fundamental fracture of the crust, possibly dating back to the Precambrian.

The magnetic low generally follows the Innuitian mobile belt across the Queen Elizabeth Islands with the eastern boundary of the trough parallel to Nares Strait. As far as the crust adjacent to the magnetic low is concerned, the magnetic high over northern Greenland corresponds to continental crust of up to 42 km thickness (Pawlowicz 1969). To the west of the magnetic low is the high over the Alpha Ridge where the nature of the crust is more uncertain (Langel et al. 1980). Although few in number, the existing heat flow data for Ellesmere Island are not high (A. Judge, pers. comm.) as might be expected from an elevated Curie isotherm and corresponding thin crust. This suggests that the low depicts a structural or petrologic variation in the crust rather than an undulation of the isotherm. The apparent structural boundary along Nares Strait would tend to confirm this interpretation.

The contrast in the magnetic anomaly map between

the northern Greenland high and the Ellesmere Island low is a large one. This would seem to indicate a fairly fundamental difference in crustal structure or petrology between the two locations. As already pointed out, the trend lines in the magnetic anomaly pattern closely follow the trend of the Innuitian mobile belt with the Greenland magnetic high over the thick Precambrian Shield. The magnetic low could then be interpreted as a fundamental change in the Precambrian basement, i.e. thinning, or even disappearance, of the Precambrian rock beneath the Palaeozoic sediments of the fold belt, as depicted in fig. 4 of Dawes (1973). Alternatively, it could reflect a petrologic change brought about during the Palaeozoic deposition and/or during the subsequent orogenesis of the Innuitian mobile belt. If the arguments of Burek (1973), Dawes (1973) and Fahrig et al. (1973) are correct and Nares Strait is the site of an extremely old, possibly Precambrian, fundamental fracture of the crust, then a change in the Precambrian basement seems the most likely alternative. In the absence of seismic refraction data, a definitive interpretation is difficult.

Along strictly speculative lines H. Frey and Langel have noted (in prep.) that in a reconstruction of Pangaea, many of the long wavelength anomalies detected in the satellite data are continuous, or nearly so, across the reconstructed boundaries. If this result can be upheld it implies that the anomaly sources were in existence prior to the break-up of Pangaea. This is true for the Greenland high and a high over the adjacent region of western Scandinavia in the Pangaea reconstruction. If the Alpha Ridge is continental type crust (Eardley 1961, King et al. 1966, Taylor 1978) it is conceivable that it was once located directly adjacent to the North Greenland magnetic high and thus formed part of the postulated "Pearya" landmass from which the sedimentary fill of the North Greenland geosyncline is thought to have originated (Dawes 1973).

Acknowledgements

R. A. L. thanks R. Horner and F. Gertler for their assistance in carrying out the necessary computation and M. A. Mayhew, H. Frey and P. T. Taylor for valuable discussion. L. T. publishes with the permission of the Director of the Geological Survey of Greenland.

References

Burek, P. J. 1973. Structural deduction of the initial age of the Atlantic rift systems. – In: Tarling, D. H. & Runcorn, S. K. (eds), Implications of continental drift to the earth sciences 2: 815–830. – Academic Press, London & New York.

Cain, J. C. & Langel, R. A. 1971. The geomagnetic survey by the polar orbiting geophysical observatories OGO 2 and OGO 4, 1965–1967. – In: Zmuda, A. J. (ed.), World magnetic survey 1957–1969. – Bull. Int. Ass. Geomagnetism Aeronomy 28: 65–75. – Paris.

Dawes, P. R. 1973. The North Greenland fold belt: a clue to the history of the Arctic Ocean basin and the Nares Strait lineament. – In: Tarling, D. H. & Runcorn, S. K. (eds), Implications of continental drift to the earth sciences 2: 925–947. – Academic Press, London & New York.

Eardley, A. J. 1961. History of geologic thought on the origin of the Arctic Basin. – In: Raasch, G. O. (ed.), Geology of the Arctic 1: 223–249. – Univ. Toronto Press.

Fahrig, W. F., Irving, E. & Jackson, G. D. 1973. Test of nature and extent of continental drift as provided by study of Proterozoic dike swarms of Canadian shield. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 583–586.

King, E. R., Zietz, I. & Alldredge, L. R. 1966. Magnetic data on the structure of the central Arctic region. – Bull. geol. Soc. Am. 77: 619–646.

Langel, R. A. 1974. Near-earth magnetic disturbance in total field at high latitudes, 1. Summary of data from OGO 2, 4 and 6. – J. geophys. Res. 79: 2363–2371.

Langel, R. A. 1980: Near-earth satellite magnetic field measurements: a prelude to Magsat. – Trans. Am. geophys. Un. 60: 667–668.

Langel, R. A. & Thorning, L. in press. A satellite magnetic anomaly map of Greenland. – Submitted to Geophys. J. Roy. astr. Soc.

Langel, R. A., Coles, R. L. & Mayhew, M. A. 1980. Comparisons of magnetic anomalies of lithospheric origin measured by satellite and airborne magnetometers over western Canada. – Can. J. Earth Sci. 17: 876–887.

Mayhew, M. A. 1979. Inversion of satellite magnetic anomaly data. – J. Geophys. 45: 119–128.

Mayhew, M. A. in press. Application of satellite magnetic anomaly data to Curie isotherm mapping. – Submitted to J. geophys. Res.

Mayhew, M. A., Thomas, H. H. & Wasilewski, P. J. 1982. Satellite and subsurface expression of anomalous crustal structure in Kentucky and Tennessee. – Earth planet Sci. Lett. 58: 395–405.

Pawlowicz, E. F. 1969. An isostatic study of northern and central Greenland based on gravity values and airborne radar ice-thickness measurements. – Unpubl. Ph. D. thesis, Ohio State Univ, U.S.A.: 98 pp.

Regan, R. D. & Marsh, B. D. 1982. The Bangui anomaly: its geological origin. – J. geophys. Res. 87: 1107–1120.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Taylor, F. B. 1910. Bearing of the Tertiary mountain belt on the origin of the earth's plan. – Bull. geol. Soc. Am. 21: 179–226.

Taylor, P. T. 1978. Low-level aeromagnetic data across the western Arctic Basin. – Trans. Am. geophys. Un. 59: 268–269.

Trettin, H. P., Frisch, T. O., Sobczak, L. W., Weber, J. R., Niblett, E. R., Law, L. K., DeLaurier, J. M. & Whitham, K. 1972. The Innuitian Province. – In: Price, R. A. & Douglas, R. J. W. (eds), Variations in tectonic styles in Canada. – Spec. Pap. geol. Ass. Can. 11: 83–179.

Wasilewski, P. J., Thomas, H. H. & Mayhew, M. A. 1979. The Moho as a magnetic boundary. – Geophys. Res. Lett. 6: 541–544.

Crustal history and basin development of Baffin Bay

ARTHUR W. MENZIES

Menzies, A. W. 1982. Crustal history and basin development of Baffin Bay. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 295–312.

Lack of definitive magnetic anomaly stripe data in Baffin Bay and Davis Strait imposes severe limitations on the extension of the Labrador Sea sea-floor spreading model into these areas. Using refraction and earthquake seismicity data to examine crustal and mantle structure, a case is presented for the existence of oceanic crust in Baffin Bay, and a bridge of continental crust connecting Greenland and Baffin Island across Davis Strait. Reconstruction or closure of Greenland with Labrador and Baffin Island is made by first defining their respective continental–oceanic crust boundaries, and then by juxtaposing these boundaries. The two sets of motion implied by Srivastava's (1978) interpretation of the magnetic anomaly pattern in the Labrador Sea are used to develop an opening motion for Labrador Sea and Baffin Bay. The Baffin Bay opening began in Paleocene time (anomaly 24) and ceased about Eocene time (anomaly 20). This motion is along a strike-slip fault crossing through Davis Strait which resulted in Baffin Bay opening as a rhombochasm with a bridge of continental crust being maintained in Davis Strait.

The suggested origin of Baffin Bay requires appreciable movement to have occurred either along Nares Strait or in the Canadian Arctic Archipelago.

A. W. Menzies, Esso Resources Canada Limited, Exploration Department, Esso Plaza, 237–4th Ave. S. W., Calgary, Alberta, Canada, T2P OH6. Present address: Kuwait Foreign Petroleum Exploration Co., 80 New Bond Street, London, W1Y 9DA, England.

Studies of the formation of the Atlantic Ocean have provided an excellent model for the understanding of the origin of oceanic basins generally. The generalized model of sea-floor spreading as applied to the Atlantic Ocean has remained the primary operating hypothesis for the initial studies of all oceanic basins within the Atlantic area. Thus the widely held view pertaining to the area of this study is that the Labrador Sea was created by the separation of a Greenland 'plate' from the North Atlantic plate. The action was a northerly extension of the spreading evolution of the North Atlantic Ocean that began in early Jurassic time and by late Cretaceous (Senonian) time had extended into the region of the Labrador Sea. A two-stage opening has been postulated (Srivastava 1978) with, in the first stage, Greenland rotating in an anti-clockwise direction throughout Senonian time and then, in the second stage, continuing opening through the Paleocene to at least Eocene time and this included the opening of Baffin Bay with strike-slip movement along Nares Strait. The nature of the transformation of the spreading centre through Davis Strait has been the subject of considerable conjecture and speculation.

The timing of these events is derived from the interpretation of magnetic 'stripes' in the Labrador Sea. Although there is not complete agreement about the interpretation of the data amongst all investigators (see e.g. van der Linden 1975, Hinz et al. 1979), it appears to be generally agreed that anomaly 33 (Santonian) and anomaly 20 (Eocene) are the oldest and youngest probable in the Labrador Sea. Sometime subsequent to anomaly 20 the spreading centre aborted the Baffin Bay–Labrador Sea area and spreading at the present-day site to the east of Greenland took over the dominant role. Greenland has since moved with the North American plate as the North Atlantic opening progressed through the Norwegian–Greenland Sea.

This scenario provides a good first working hypothesis, but leaves some problems. Broadly speaking the major problems are:

1) Davis Strait is a morphological high in an apparently purely extensional regime. Is this a post-extensional, or syn-extensional feature such as Iceland, or is it more fundamental, a geomorphological high of some antiquity?
2) There are no basin-wide, good quality, magnetic stripes in Baffin Bay with which to interpret the evolution of this area, such as are present in the Labrador Sea (Jackson et al. 1979). Magnetic profiles are indeed available, but it appears that the ability to provide dependable correlations of anomalies is lacking.
3) In Baffin Bay there is an apparent lack of good morphological fit of either the coastlines or of any bathymetric contours with which to conjecture or

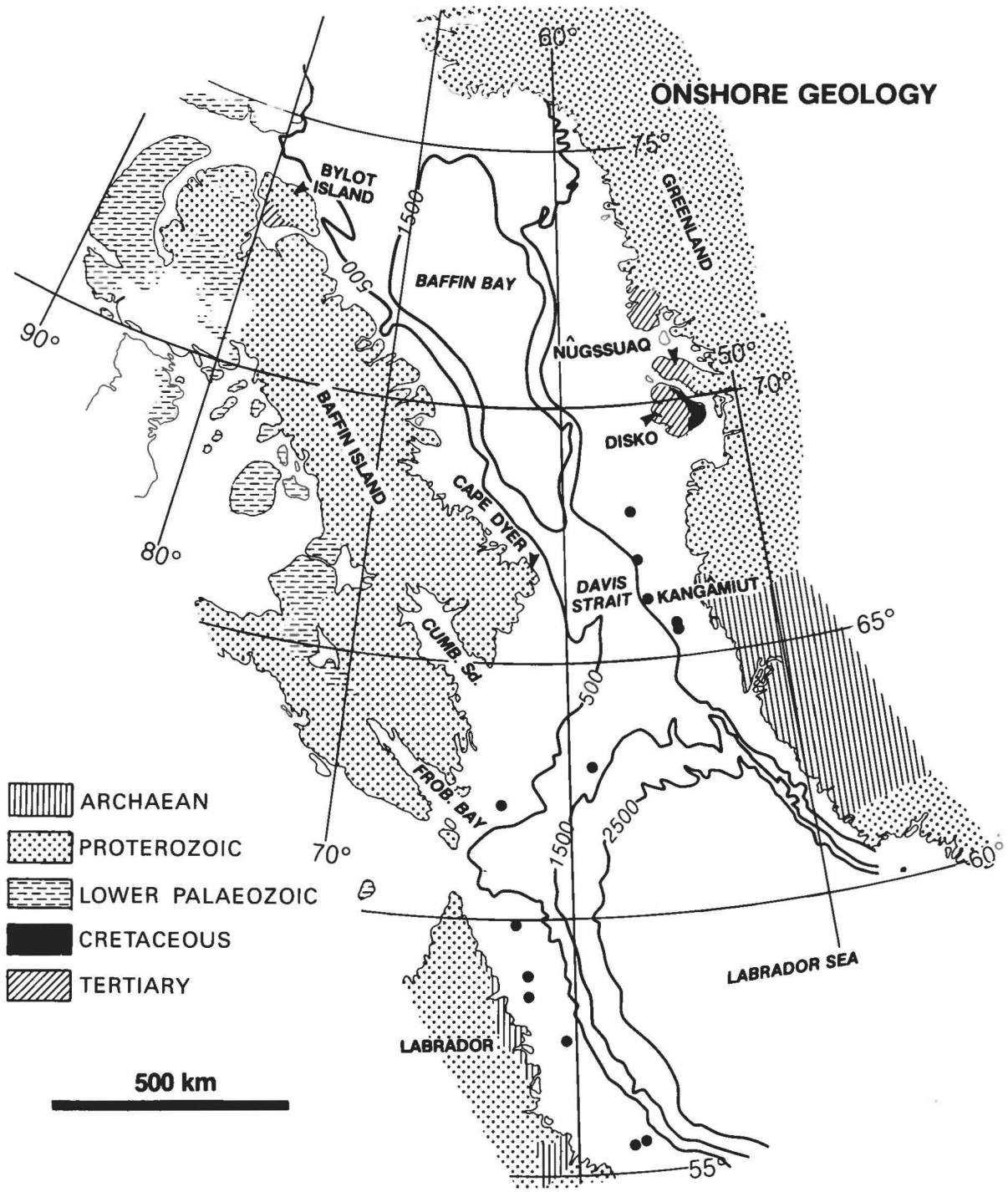


Fig. 1. General geological map of the study area compiled from published sources of the Geological Surveys of Canada (GSC) and Greenland (GGU). Black dots indicate sites of industry boreholes with one, Kangâmiut, named; bathymetry is in metres.

postulate a pre-spreading closure of the Bay. This is quite different from the Labrador Sea where there is a good morphological fit of the coastlines.

4) Whereas the Labrador Sea shows a general symmetry about a central bathymetric ridge and graben system, Baffin Bay shows an asymmetric bathymetric basin and no morphologic or crustal evidence of a ridge or graben.

It seems that the resolution of the problems rests on two parameters, viz. defining the nature of the crust in Baffin Bay and Davis Strait and the accurate definition of the continental–oceanic crust boundary.

Surface geology considerations

The oldest rocks in the study area are of Archaean age and these form a stable block flanked on the north by successively younger rocks (Fig. 1). The oldest of these are metamorphosed early Proterozoic rocks, mainly Aphebian that in places are intimately associated with reworked Archaean material (Douglas 1967, Escher et al. 1970, Jackson & Taylor 1972, Bridgwater et al. 1973, Escher & Watt 1976, Douglas et al. 1979). The late Proterozoic is represented by relatively unmetamorphosed sediments and volcanics that occur on northern Baffin Island and in North-West Greenland. Overlying the Proterozoic rocks on the north end of Baffin Island and occurring as outliers preserved in gra-

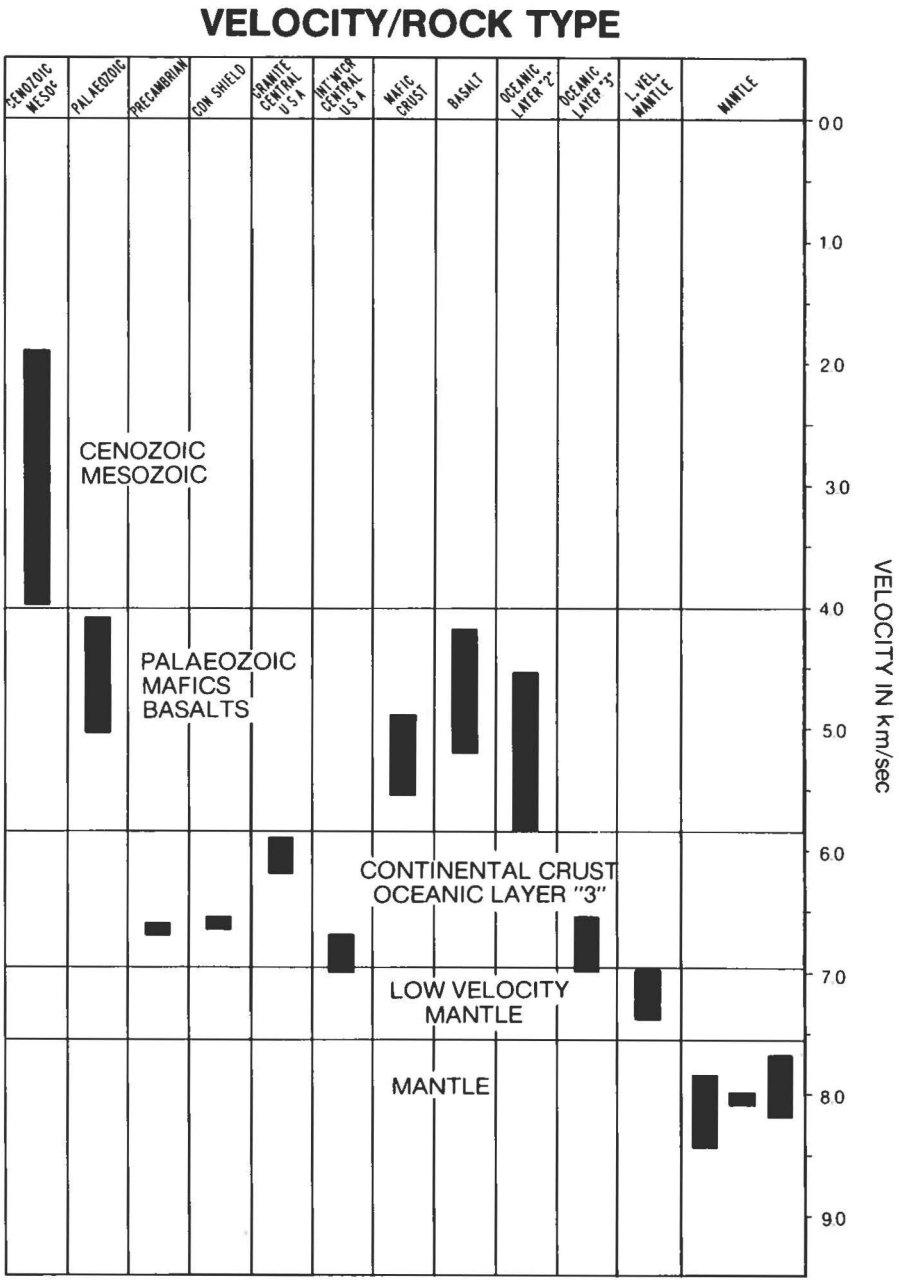


Fig. 2. Compilation diagram of seismic velocities versus rock type. Sources: Healy & Warren 1969, Keen & Barrett 1972, Berry & Fuchs 1973, van der Linden 1975 and Esso Resources Canada Ltd.

bens throughout the map area in Canada are early Palaeozoic carbonates of Cambrian through Silurian age (Beh 1975, Grant 1975). Isolated remnants of this once extensive Palaeozoic cover are also preserved in West Greenland (Escher & Watt 1976). Jurassic basic dykes (Watt 1969) and a small outcrop of possible Jurassic age (King & McMillan 1975) occur in the west coast of Greenland and Labrador, respectively. The next sequence of rocks within the study area are Lower and Upper Cretaceous strata that occur in the Disko region of West Greenland; the Upper Cretaceous is also represented on Bylot Island. Tertiary rocks, both sediments and volcanics, are known from West Greenland and from Cape Dyer and Bylot Island (Clarke & Upton 1971, Andrews et al. 1972, Henderson et al. 1976).

Detailed lithologic and stratigraphic descriptions and interpretations of the region have been published and bedrock summaries of the Canadian and Greenland sides of the study area are found in Douglas (1970) and Escher & Watt (1976). Only some brief comments on the historical significance of some surface outcrops are given here.

The major hiatus that exists between the Lower Palaeozoic carbonates and the late Mesozoic strata indicates that the study area was essentially a high standing craton for the majority of its history. Any sedimentation that took place during this period was most probably of thin continental rocks that were easily removed. This area was most probably a main provenance of the Late Palaeozoic, Mesozoic and Cenozoic strata of the Sverdrup Basin, where a thick and near-complete geologic sequence is present (Balkwill 1978).

The Cretaceous deltaic deposits of the Disko region, West Greenland, prograde from south-east toward north-west (Henderson et al. 1976), although it is postulated that the opening of Baffin Bay is supposed to have progressed from south to north. Additionally the uppermost Cretaceous of this delta is a marine sequence, thickening to the north. Thus the history of this delta indicates the development of a Baffin basin coincident, at least in part (Barremian–Turonian), with the opening of the Labrador Sea, but with the distinction that the marine invasions were from the north. It thus appears that the Baffin Bay sea and the Labrador Sea were converging upon each other, possibly joining at or in the vicinity of Davis Strait (B. MacLean, pers. comm.). This suggests some connection between the geological history of the Arctic Basin and that of the North Atlantic at least as early as early Cretaceous (Barremian) time. In other words, it may be that the

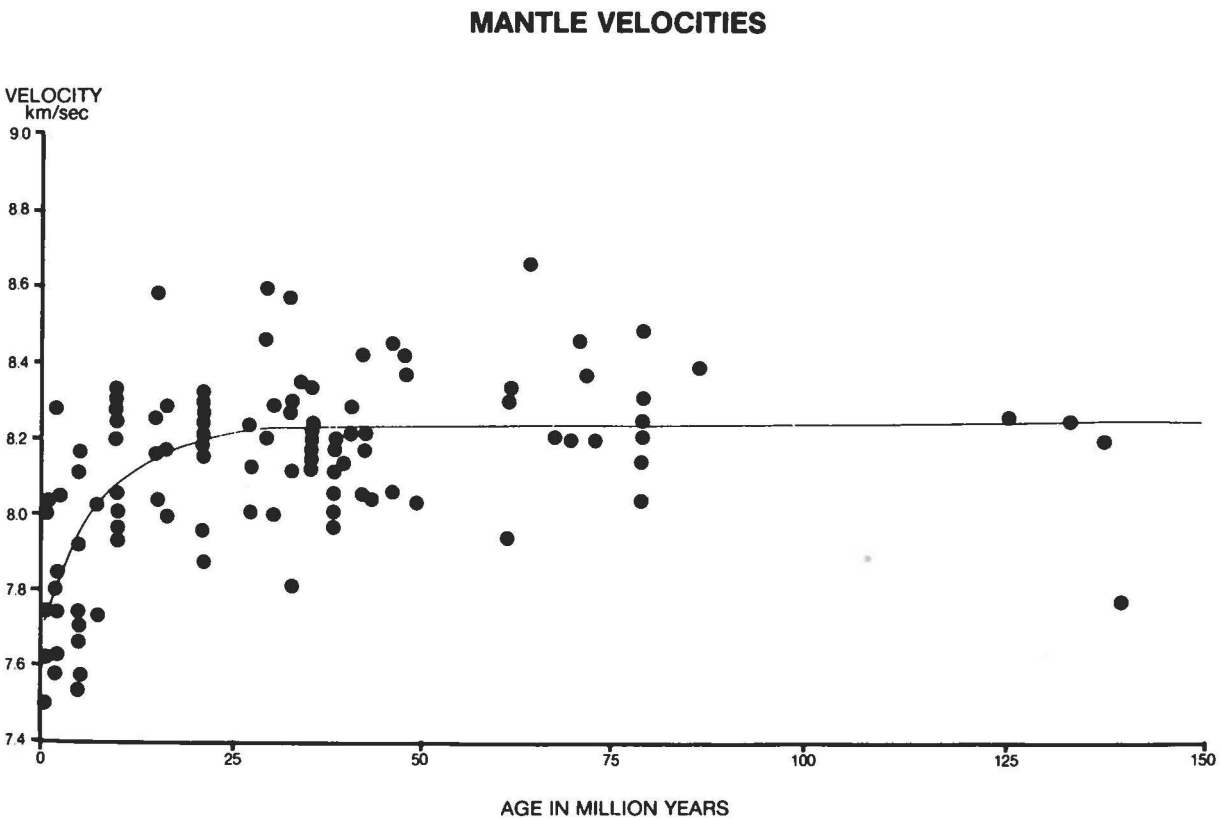


Fig. 3. Diagram showing relationship of mantle velocities and crustal age. From Lewis (1978).

history of Baffin Bay is connected with the Arctic Basin, at least as much, if not more, as with the Labrador Sea – North Atlantic.

The Tertiary volcanics of central West Greenland and Cape Dyer on Baffin Island as well as offshore outcrops indicate a major episode of volcanism beginning in Paleocene time (Clarke & Upton 1971, Clarke 1975, Clarke & Pedersen 1976, MacLean et al. 1978). Cross-bedding and overlap of flows indicate an easterly to north-easterly source for the Cape Dyer sequence and a south-westerly source for the Greenland volcanics. This points to the Davis Strait as a source area for the Tertiary volcanics. From the interpretation of the magnetics of the area it would appear that the volcanic province extends 1000 km from the Nûgssuaq–Disko region to at least Frobisher Bay, and is at least as wide as Davis Strait, i.e. about 200 km. Is Davis Strait an extinct Iceland and/or Faeroe Ridge situation?

Crustal definition

Until recently the interpretation of refraction data was concentrated on trying to relate velocities with rock type and in this way interpret the geology of the crust. It is now accepted that there is nothing sufficiently characteristic about oceanic or continental velocities to distinguish them from each other (Fig. 2). However, mantle velocities are distinguishable from each other. The mantle invariably seems to come in at 7.5 km/sec. or greater (Fig. 3). Parsons & Sclater (1977) indicate that for a crust 46 m.y. or younger the crustal velocity values would range from 7.85 to 8.35 km/sec. Fig. 4 indicates that oceanic crustal velocities range from 6.5 to 7.1 km/sec. Therefore the ability to distinguish mantle velocities makes it possible to attack the problem of oceanic versus continental crust on the basis of depth to mantle and mantle profiles.

Cross-sections A–A′, B–B′ and C–C′ (Figs 6, 7 and 8) show the profile of the mantle and interpreted crustal and sediment thicknesses across Baffin Bay, the Labrador Sea and from Baffin Bay to Labrador Sea (see Fig. 5 for locations). These cross-sections have been compiled from a variety of data both published and unpublished, the main published sources being Hyndman et al. (1971), Keen & Barrett (1972), Keen et al. (1972), Hyndman (1973), McMillan (1973), Pelletier et al. (1974), van der Linden (1975), Jackson et al. (1977, 1979), MacLean (1978), Srivastava (1978), Hinz et al. (1979), Keen & Hyndman (1979) and MacLean & Falconer (1979).

Essential points shown by the cross-sections are given below.

A–A′ (Fig. 6)

1. The velocities in central Baffin Bay ranging from 7.68 to 8.53 km/sec. are clearly mantle.

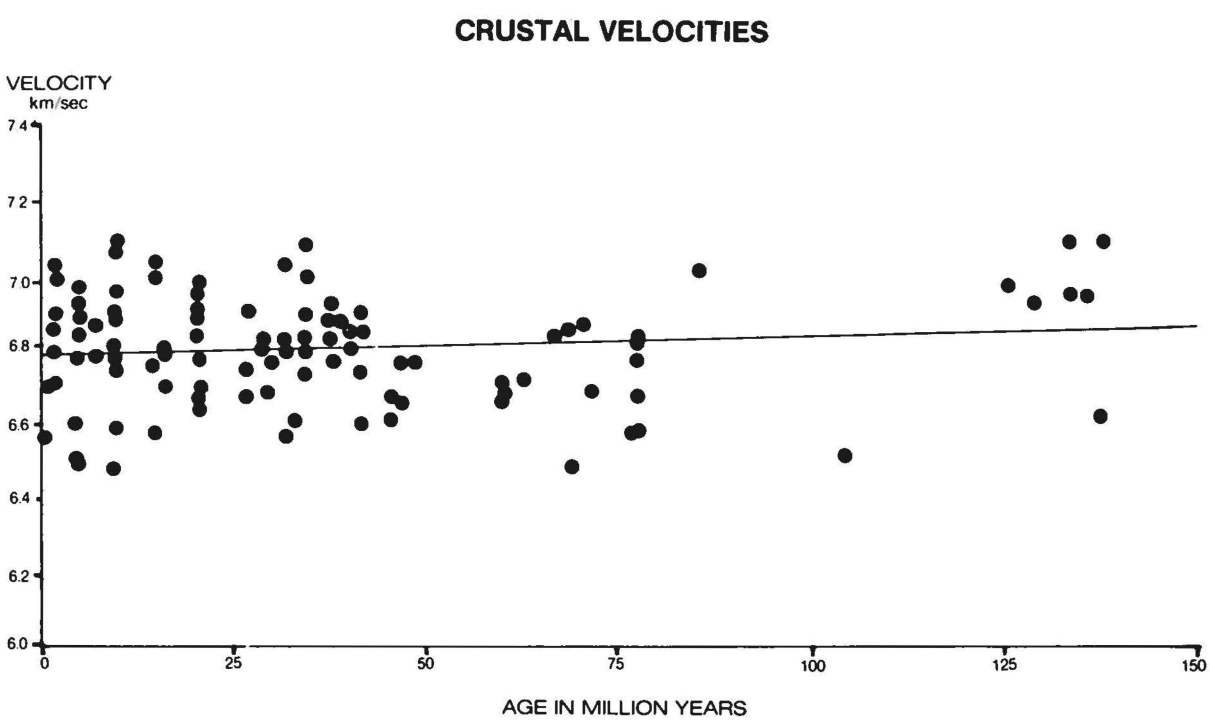


Fig. 4. Diagram showing relationship of crustal velocities and age of oceanic crust. From Lewis (1978).

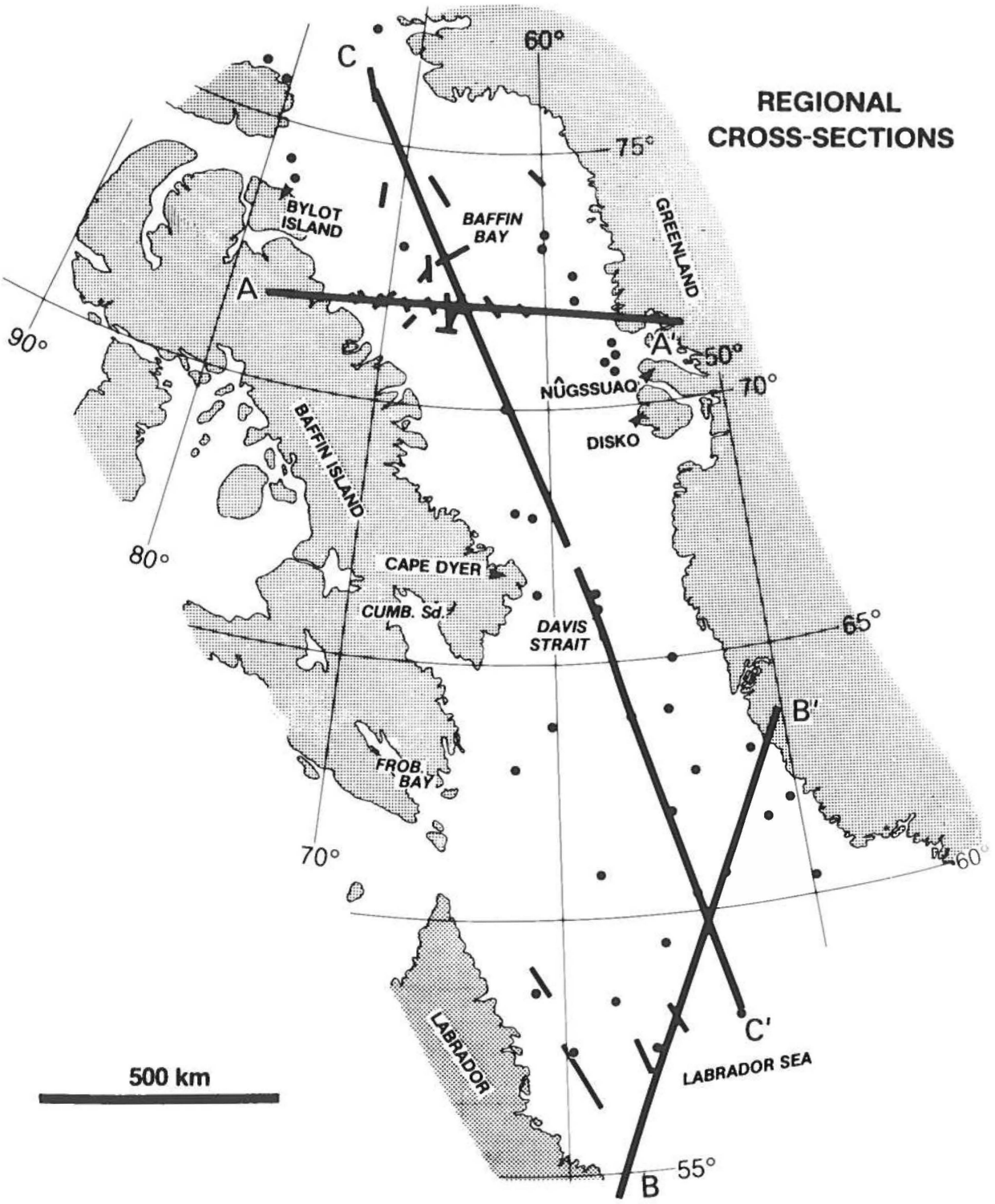


Fig. 5. Map showing location of refraction sonobuoy stations (dots), and seismic refraction profiles (bars). Heavy lines AA′, BB′ and CC′ are the locations of the regional cross-sections shown in Figs 6, 7 and 8.

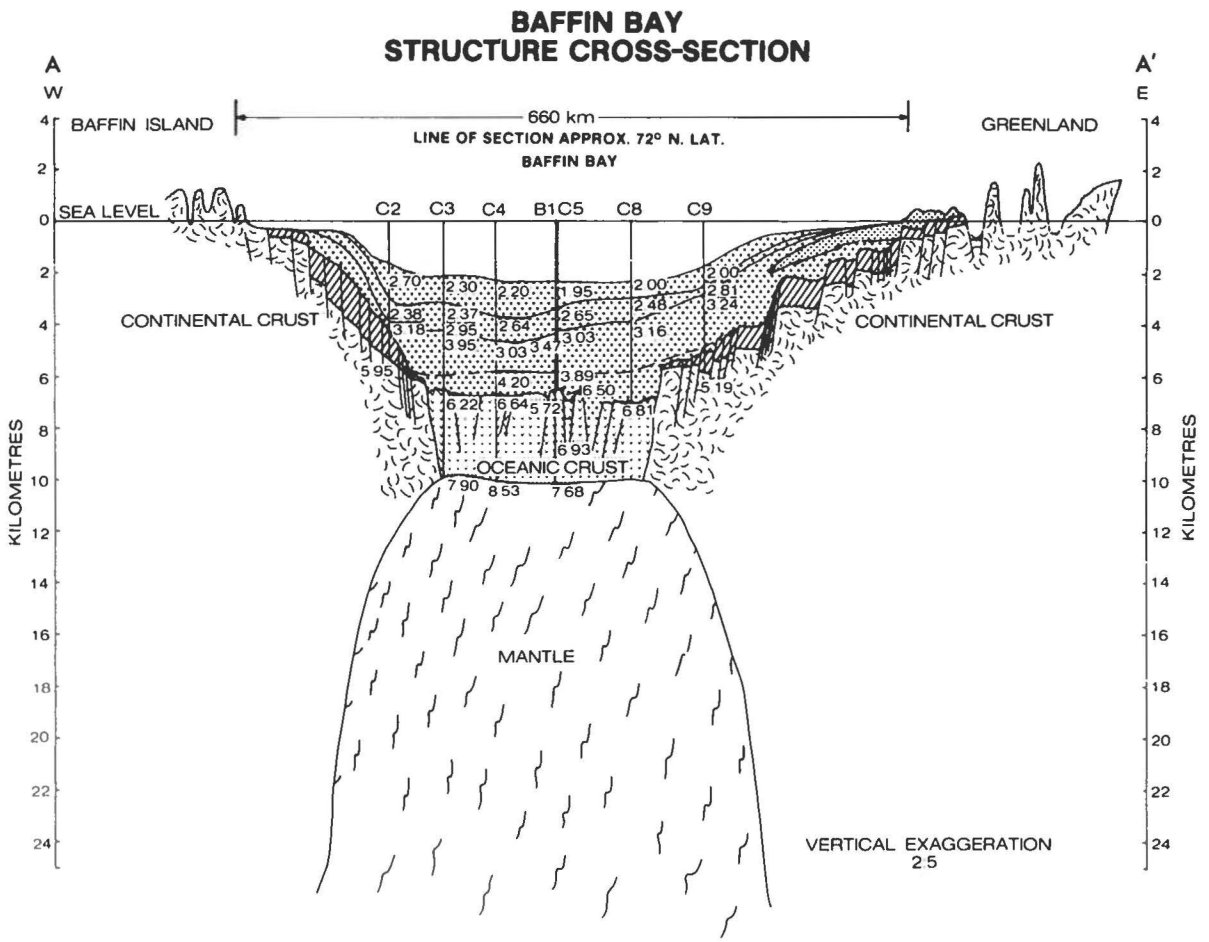


Fig. 6. Regional E–W cross-section of Baffin Bay as located in Fig. 5. Line shading = Mesozoic sediment; heavy dot symbol = Tertiary strata; light dot symbol = Tertiary volcanics. C2, B1, etc. refer to seismic refraction profiles taken from Keen & Barrett (1972). Numbers 2.70, 3.18, etc. are refraction velocities in km/sec.

2. The mantle profile is typical of the change from continental to oceanic crust. Although no depth to mantle is available below Baffin Island and Greenland, these are clearly continental blocks and the depth to mantle must be in excess of 30 km.
3. Depth to mantle of ±10 km is about what one would expect beneath oceanic crust.
4. Thickness of the crust in central Baffin Bay is compatible with the statistical range of values for oceanic crust, i.e. 3–5 km.
5. The submergence versus age relationship demonstrated by Parsons & Sclater (1977) indicates a mid-late Eocene age for this crust, which is compatible with the age postulated for this crust assuming Baffin Bay opened coincidentally with the Labrador Sea, i.e. spreading ceased at about anomaly 20 time.
6. Using crustal velocities and the mantle profile to place the continental–oceanic boundary as shown in Fig. 6, the resulting width of oceanic crust closely approximates the distance between anomalies 24 in the Labrador Sea.

B–B′ (Fig. 7)

1. The mantle profile is typical of the change from continental to oceanic crust.
2. The depth to mantle in central Labrador Sea is again typical of depths beneath oceanic crust.
3. Flanking continental crust velocities are acceptable, but at 6.65 to 7.28 km/sec. they are high for simple continental crust. Because these velocities are in the range of both oceanic crust and low velocity mantle it is proposed that they reflect the alteration of velocities that would occur along a margin where there was significant dyke injection of mantle material.

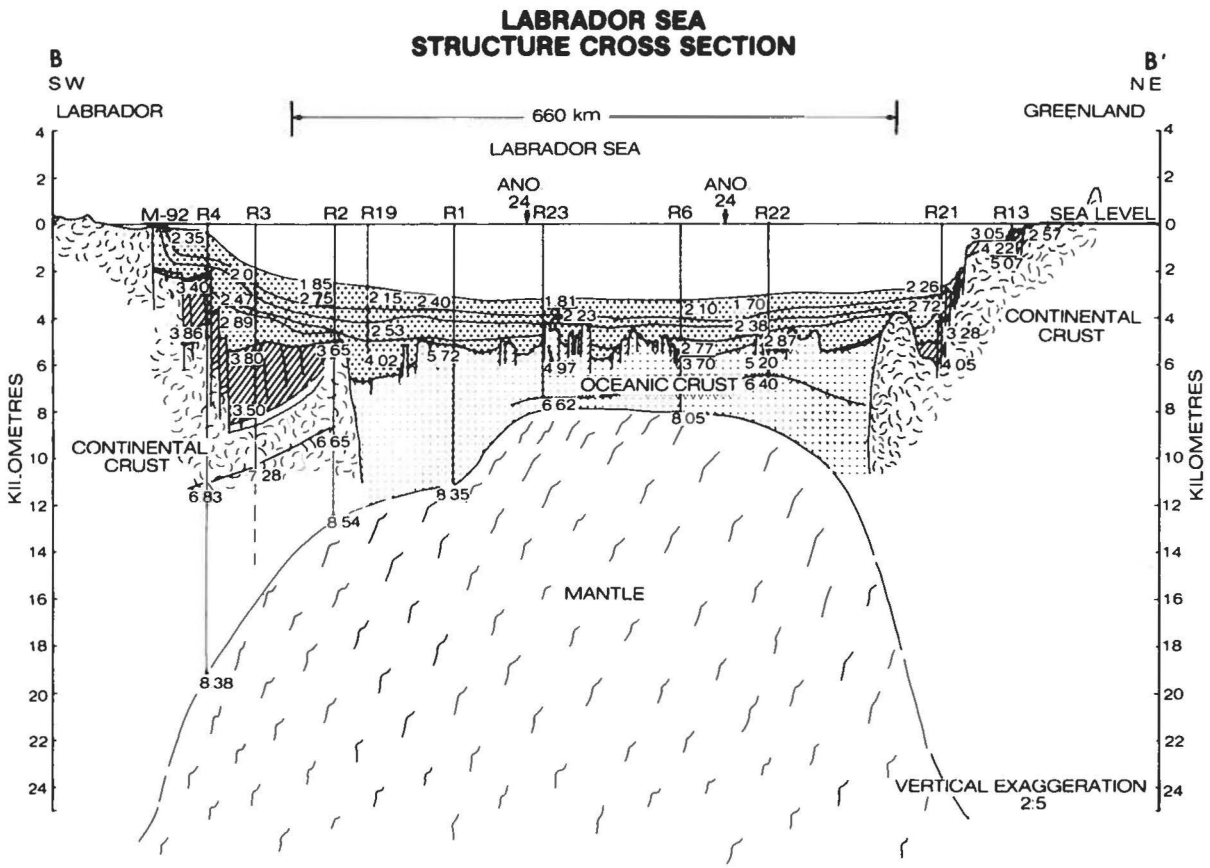


Fig. 7. Regional NE–SW cross-section of Labrador Sea as located in Fig. 5. Line shading = Mesozoic strata, dot symbol = Tertiary strata. The seismic refraction data are M–92, Herjolf eastcan well, Nov. 22, 1978; R1, R2, R3 and R4 from van der Linden (1975), and R6, R13, R19, R21, R22 and R23 from Hinz et al. (1979). Numbers 1.85, 2.35, etc. are refraction velocities in km/sec. Ano. 24 refers to the position of magnetic anomaly 24 of Srivastava (1978).

C–C′ (Fig. 8)

This section shows the tie between A–A′ and B–B′ and highlights the problem of Davis Strait. The single refraction profile D1, a reversed profile, brings forth the following points:

1. The thickness of the post-mantle layers is far too thick for a 'normal' oceanic crust.
2. The depth to mantle is too great for normal oceanic crust.
3. The relatively thin layer of 4 km/sec. material resting on the thick crustal layers is interpreted as Tertiary volcanics.
4. Possible analogous features to this Davis Strait phenomenon are:
   a) The Icelandic volcanic cone
   b) The Faeroe volcanic ridge
   c) Erratic or remnant continental blocks such as the Faeroe Islands, Flemish Cap and Orphan Knoll.

Comparison of Baffin Bay and Labrador Sea geology

The following points are considered significant.

1. Tertiary sediments are twice as thick in central Baffin Bay as in central Labrador Sea.
2. Tertiary sedimentation in the Labrador Sea is typically of coastal prism profile type as seen along open coastal margin sag basins, whereas Baffin Bay sedimentation is basically uniform across the basin. There is no evidence of progradation of sequences from the flanks into the centre of Baffin Bay and it appears to have been filled uniformly across the Bay. This difference in sedimentary style, as interpreted by seismic facies and sequence wedge geometry indicates a possible fundamental difference in the structural genesis of each basin during the Tertiary, even though both were responding to extensional forces

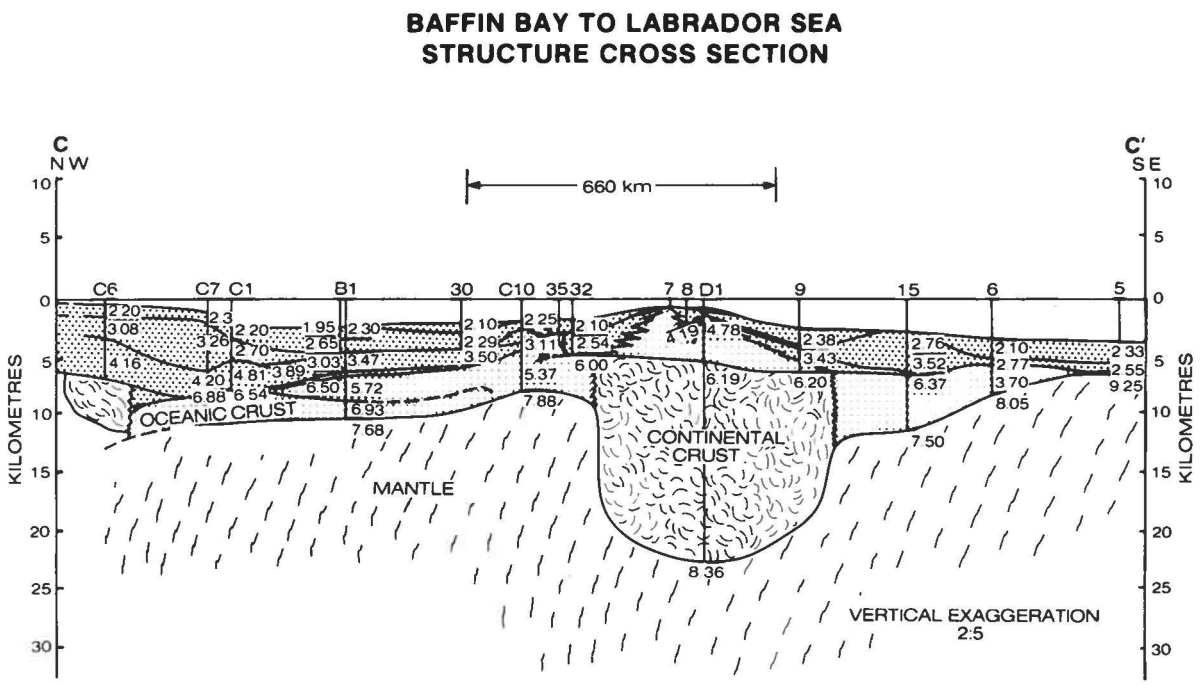


Fig. 8. Regional NW–SE cross-section of Baffin Bay and Labrador Sea illustrating the continental crust under Davis Strait. Dot symbol represents Tertiary strata. The seismic refraction data 7, 8, 30, 32, 35, B1, C1, C6, C7, C10 and D1 are taken from Keen & Barrett (1972); 5, 6, 9 and 15 are from Hinz et al. (1979).

related to plate separation and new crustal formation.

3. There is a good rationale available to reasonably conclude that Baffin Bay, like the Labrador Sea, is floored by oceanic crust.
4. Crustal definition within Davis Strait remains enigmatic. Although the mantle velocity is quite clearly defined, the crustal layer velocity remains ambiguous. Two interpretations are equally valid at this point. Davis Strait is a remnant, or thinned continental bridge between Greenland and Baffin Island; or it is a ridge of volcanics such as the Faeroe Ridge.

Seismicity

In Baffin Bay and Davis Strait interpretations of earthquake short and long period waves by Qamar (1974), Wetmiller (1974), Basham et al. (1977) and others have been made that throw some light on the problem.

Two short period and two long period waves crossing Davis Strait have given apparently conflicting results (Fig. 9). The short period waves as recorded at the Frobisher Observatory are without a long phase, thus indicating probable oceanic crust. The two long period waves indicate transmission through a continental crust. The solution to this conflict must be in the structural genesis of Baffin Bay and the translation of the apparent spreading centre from Labrador Sea through Davis Strait.

Structural trend analysis

For this analysis we are concerned only with major regional features, such as regional foliation in crystalline rocks, and most particularly with anomalous changes in trends and patterns. Major structural trends have been compiled in Fig. 10.

Onshore in coastal Greenland there is a change in the orientation of major lineations at about the contact of the Proterozoic and Archaean blocks. The Proterozoic lineations are generally parallel to the contact, in a sense wrapping around the Archaean block. This trend seems to continue into the more recent structure pattern described by seismic work off the southern end of Baffin Island, suggesting that the recent structures may follow the trend of the fabric of the deeper crustal rocks. A second change of trend occurs off southern Baffin Island where the NW–SE trends of onshore southern Baffin Island do not persist offshore. A third anomaly onshore is the Foxe fold belt which crosses central Baffin Island. This belt runs in an E–W arcuate pattern across the general NW–SE fabric of the island. In the east the belt curves to the south-east, so that when ex-

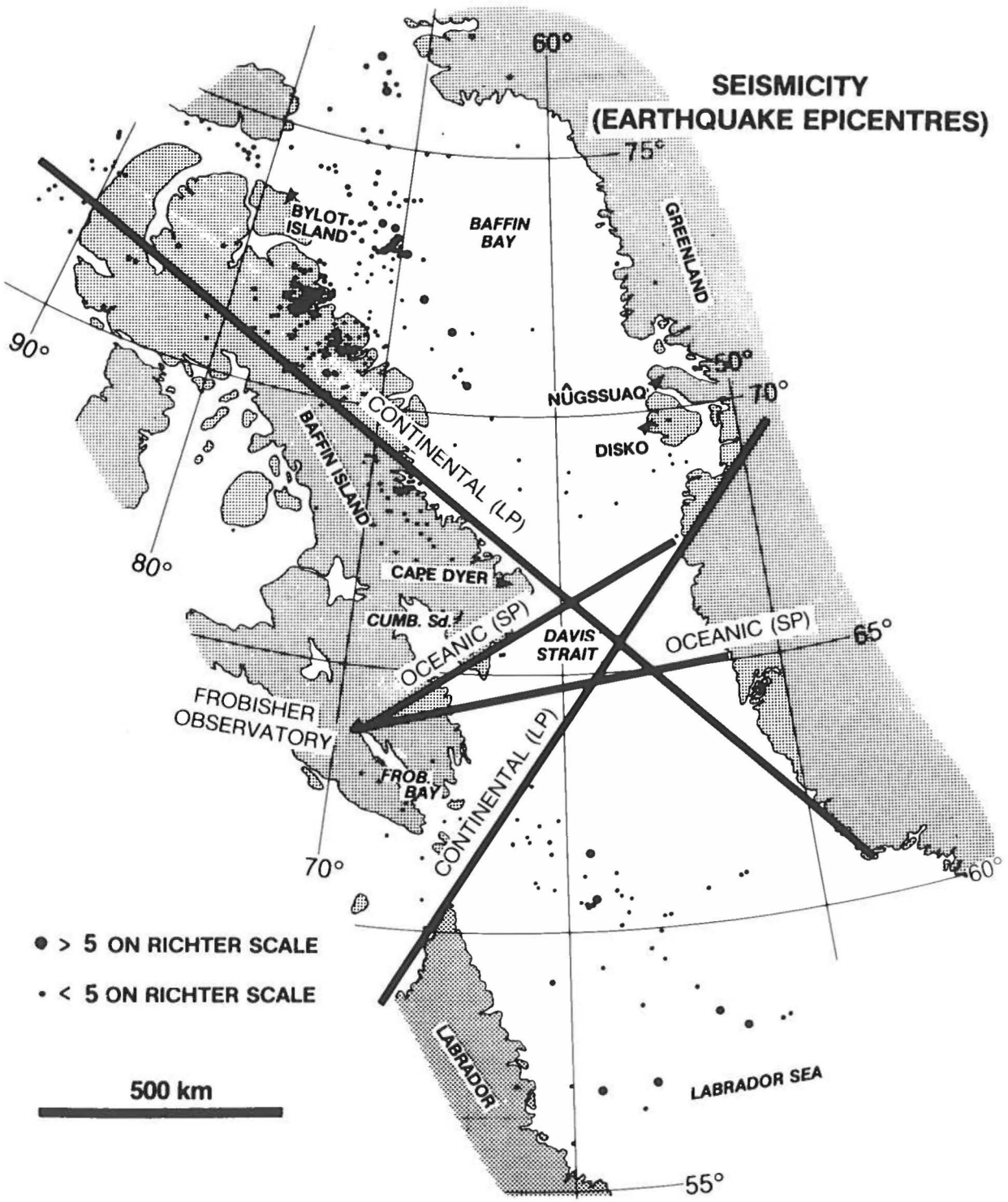


Fig. 9. Seismicity map of the study area. Black dots indicate earthquake epicentres; heavy black lines represent the location of earthquake travel paths across Davis Strait, whose Lg wave presence or absence has an important bearing on the interpretation of continental versus oceanic crust. Data supplied by Earth Physics Branch, Department of Energy, Mines and Resources, Ottawa.

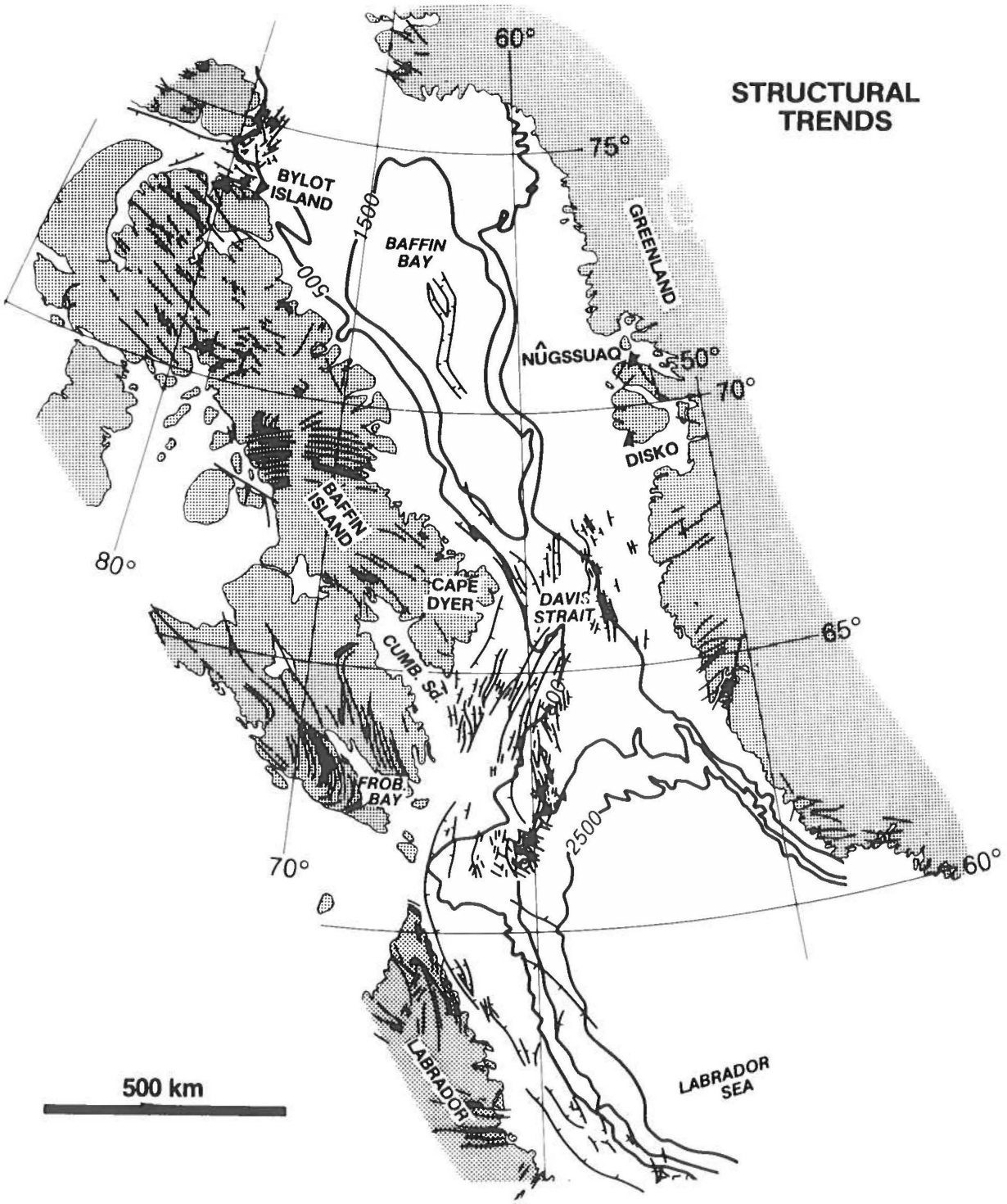


Fig. 10. Map compilation of faults, lineations and foliations to illustrate major onshore and offshore structural trends. Barbs on faults on downthrow side. Onshore interpretation from GSC and GGU maps; offshore interpretation from unpublished data (Esso Resources Canada Ltd.). Bathymetry is in metres.

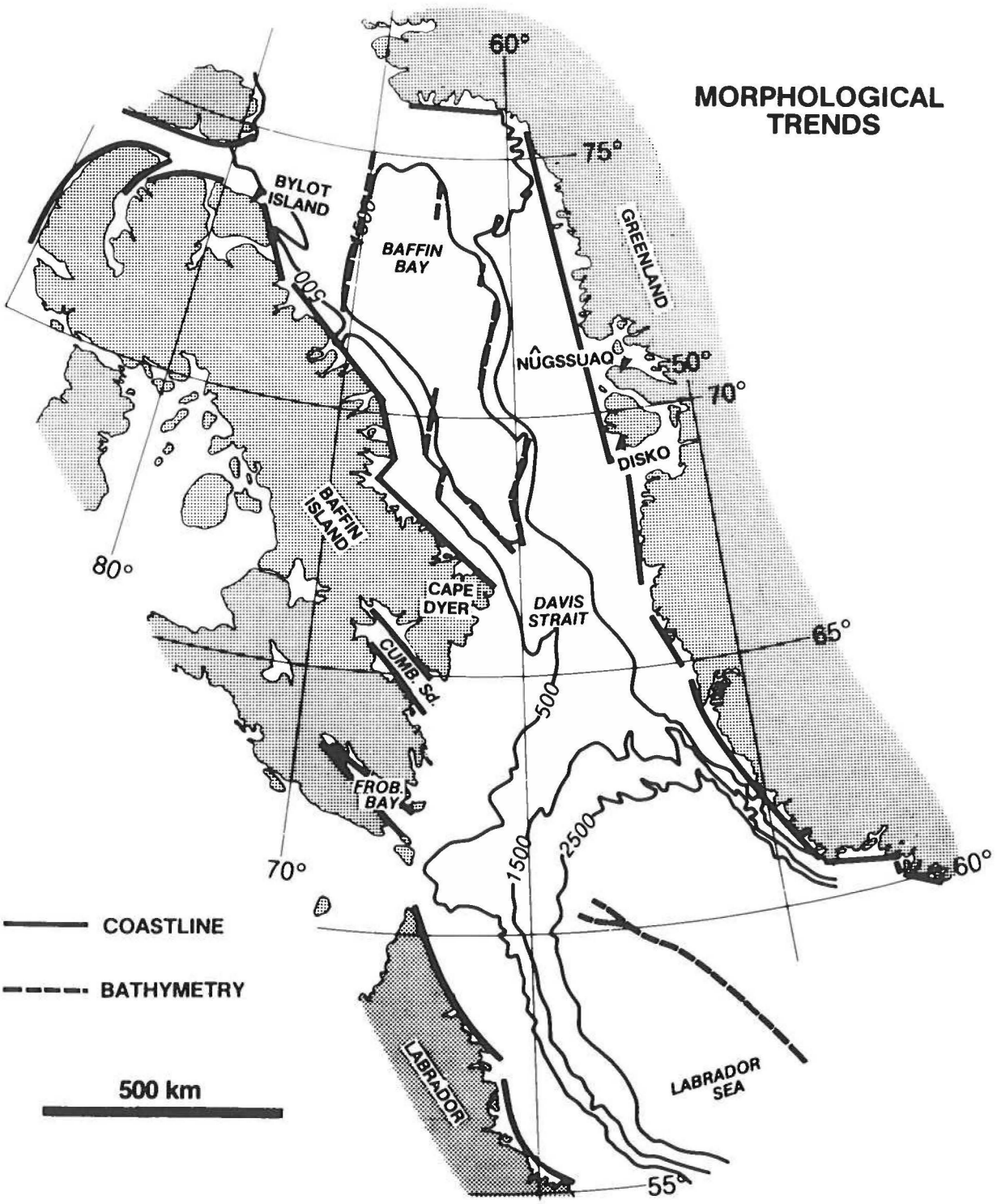


Fig. 11. Morphological trend map based on coastline and bathymetry (in metres).

tended offshore it trends parallel to the east coast of the southern half of the island. In addition, the general coastal orientation of Baffin Island changes south of the fold belt from the dominant NW–SE trend to a more N–S trend, as if skewed relative to the northern half of the island. It would be interesting to know if there is any evidence of Cretaceous tectonism within this ancient fold belt that has affected the regional structure of southern Baffin Island.

A morphological analysis of the bathymetry from various sources (e.g. Loken & Hodgson 1971) shows several anomalous trends and changes in direction of trends (Fig. 11). The most obvious is the Davis Strait high, which provides the morphological distinction between the Labrador Sea and Baffin Bay. In Baffin Bay there are some really striking changes in direction of the bathymetric gradients. It is possible that these have a structural genesis because there is no indication that they are caused by sedimentary constructional patterns (e.g. delta lakes) or destructional erosional patterns (e.g. palaeoslopes). The pattern exhibited is a series of coast-parallel lineations that are offset by N to NNE trends. The most obvious of these trends is the one at the north end of the Bay along the 70°W longitude, followed by the trend in the middle of the Bay which is coincident with the marked offset of the Baffin Island coast south of the Foxe fold belt.

Interpretation of a continental–oceanic crust boundary

From the crustal definition analysis presented it is concluded that Baffin Bay is floored by oceanic crust as is the Labrador Sea. Davis Strait remains ambiguous because of the conflict of the seismicity data. However, because the Kangâmiut well off West Greenland (Fig. 1) bottoms in granitic rocks, thus indicating continental crust at this point, it was decided as a first premise to assume continental crust as most likely for Davis Strait and see what constraint, if any, this would place on our reconstruction.

The continental–oceanic crust boundary shown in Fig. 14 was drawn on the basis of the following:

1) Refraction interpretations as given in cross-sections A–A′, B–B′, C–C′
2) Multichannel, reflection, seismic criteria
3) Structural trend analysis
4) Bathymetric morphological analysis
5) Earthquake seismicity analysis of long waves
6) Magnetic trends
7) Kangâmiut well results

Reconstruction of Baffin Bay and the Labrador Sea

Three reconstruction of the Baffin Bay – Labrador Sea region are presented in Figs 12, 13 and 14. The constraints we placed upon these reconstructions are given below.

1) Closure of Greenland with Labrador must be compatible with major plate motions in the Euro-American plate from 120 m.y. ago to present.
2) Rotation must closely approximate the motion expressed by the magnetic stripes in Labrador Sea.
3) Strike-slip motion along Nares Strait should be kept to a minimum to meet onshore geological data that suggest geological continuity between Ellesmere Island and Greenland (Kerr 1967).
4) Seismicity results in Davis Strait need an explanation.
5) Motions should be compatible with the observed structural and sedimentary styles.

Fig. 12 shows the position at full closure which is achieved by moving back to the approximate position of anomaly 33 (C. Tapscott, pers. comm.). Fig. 13 shows the position at anomaly 24 time (i.e. 57 m.y. on time scale of LaBrecque et al. 1977). The motion requires rotation about a relative pole of rotation located in northern Baffin Island. During this opening the following events are considered to have occurred:

1. As the Labrador rift valley gradually opened there was a marine invasion from the south such that by the end of the Cretaceous we have Atlantic marine conditions in the Frobisher Bay – Cumberland Sound area. Coincidentally there was a marine invasion from the high arctic entering the intracratonic Baffin Bay basin. This northern marine invasion progressed at a rate that permitted the linking up of the Arctic and Atlantic Oceans in late Cretaceous time across the Davis Strait structural high.
2. The Frobisher Bay region, and the Baffin Bay intracratonic basin were essentially extensional regimes. With these two areas under extension the Davis Strait area was placed under mild compression and remained relatively positive while the Cretaceous Labrador oceanic basin and the Cretaceous Baffin intracratonic basin were developing.

In the Tertiary the motion changed and the relative pole of rotation shifted to approximately the position of present-day Algeria. The translation of the opening through Davis Strait was along a NE–SW-trending

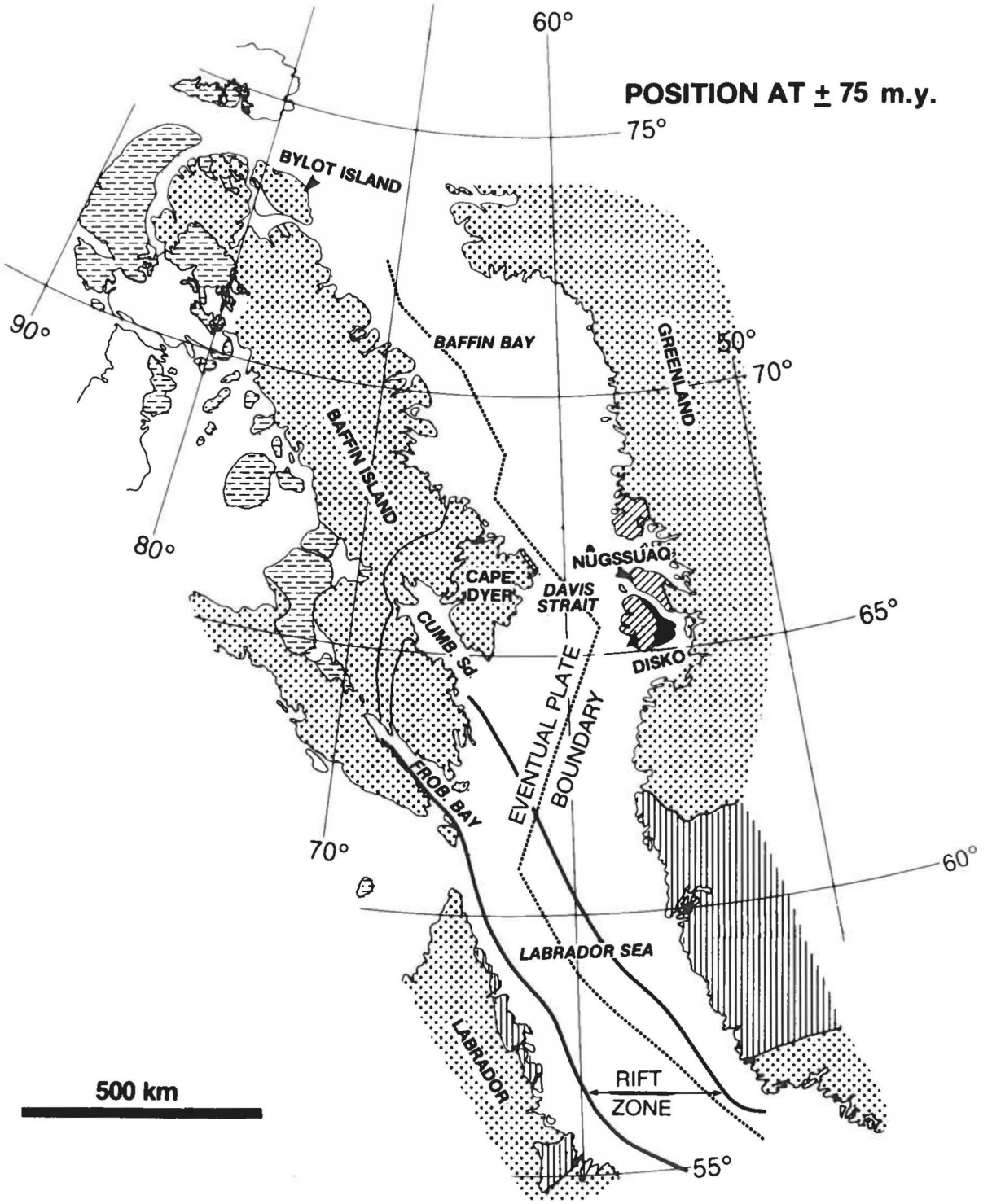


Fig. 12. Interpreted pre-drift position of Greenland and North America at about 75 m.y. (Late Cretaceous). Geology as in Fig. 1.

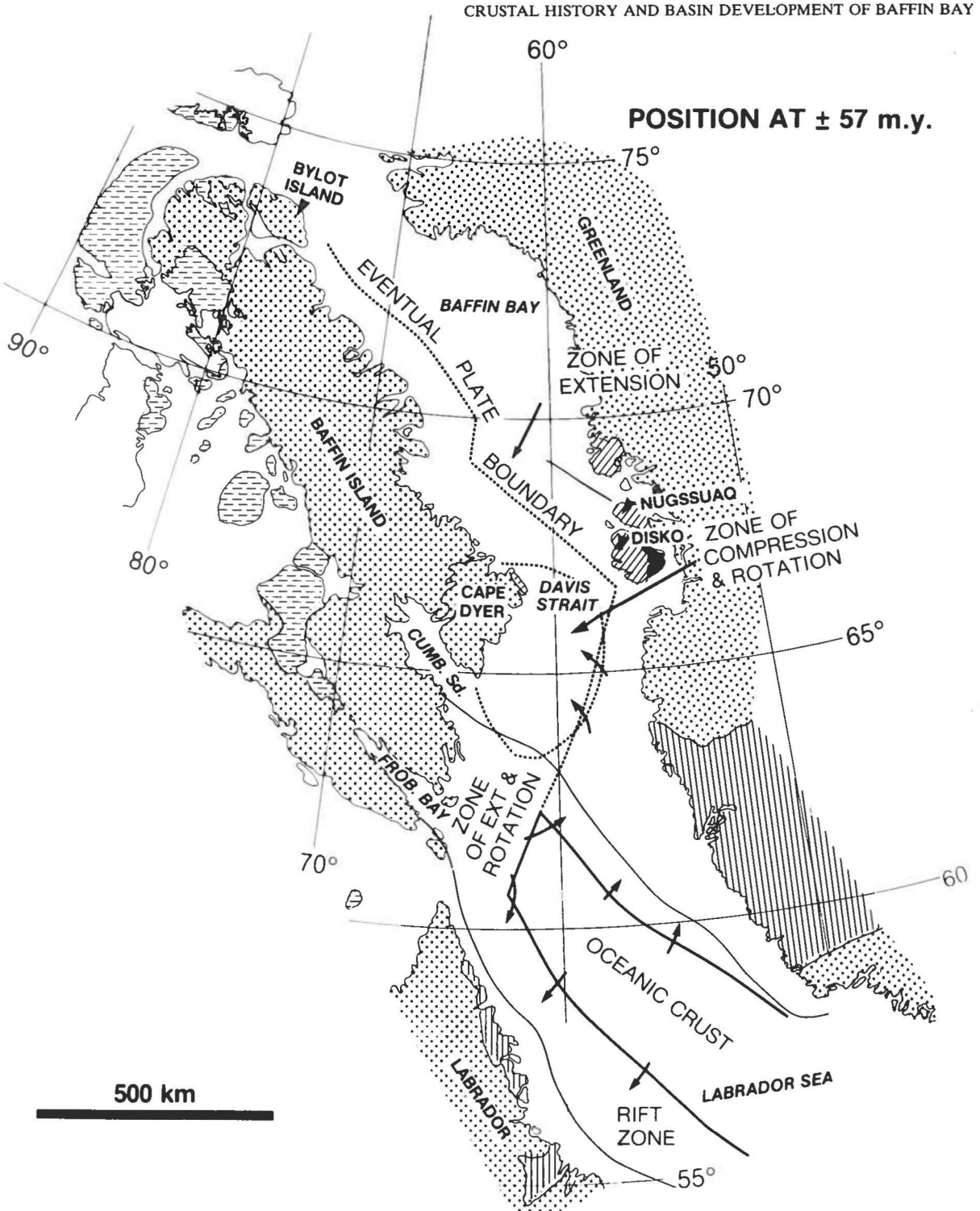


Fig. 13. Interpreted position of Greenland and North America at about 57 m.y. (Paleocene). Geology as in Fig. 1.

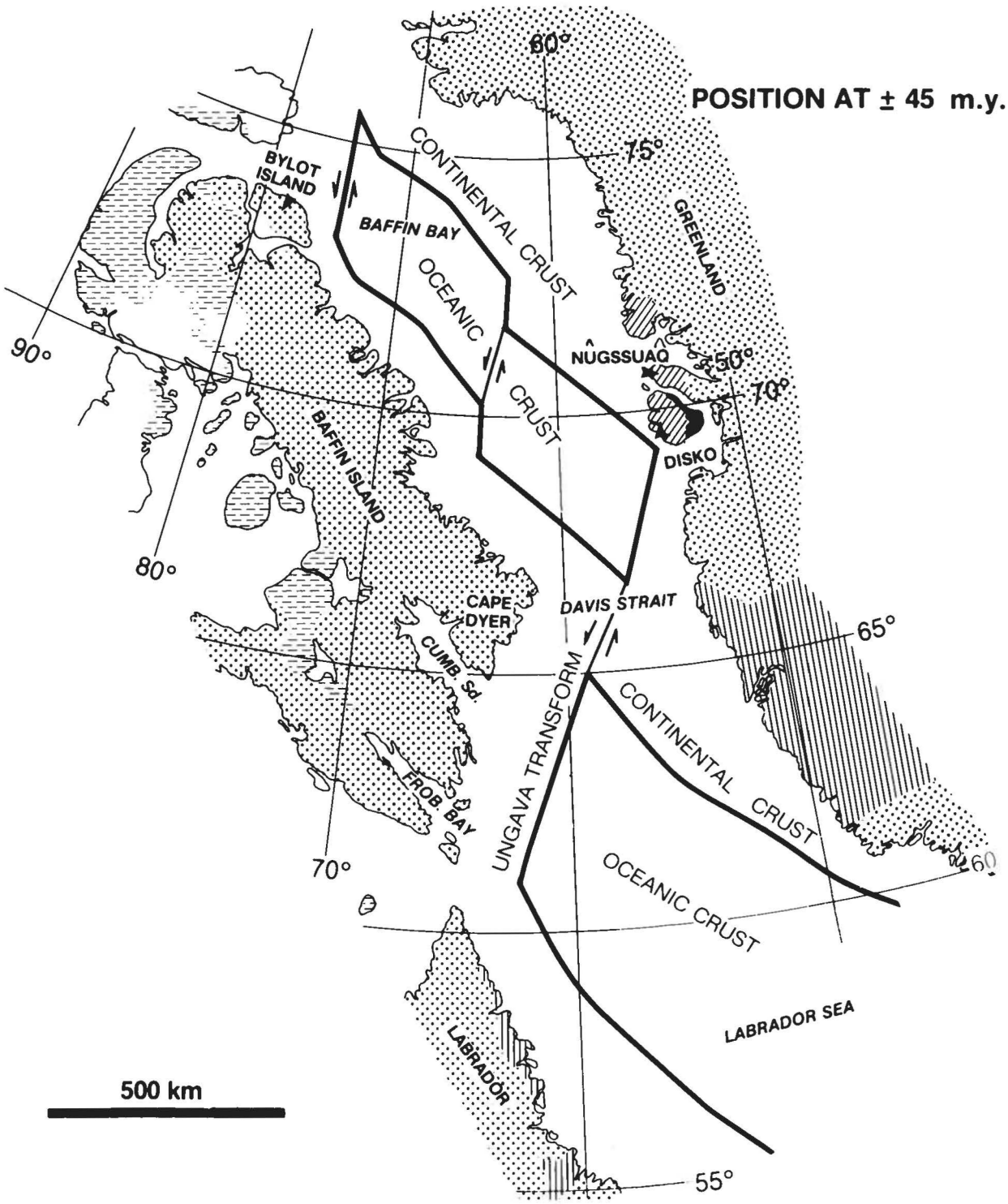


Fig. 14. Interpreted position of Greenland and North America at about 45 m.y. (Eocene) showing interpreted continental–oceanic crust boundary and the position of the Ungava transform. Geology as in Fig. 1.

transform fault. Consequently the Tertiary opening of Baffin Bay and the Labrador Sea was coincident and not progressive from south to north. Baffin Bay opened as a rhombochasm. It is probable that the extension occurred along lineations developed at a much earlier stage. For example the translation through Davis Strait may have followed the zone of weakness at the Proterozoic–Archaean contact, while the opening in southern Baffin Bay occurred along the zone of weakness created by the offshore extension of the Foxe fold belt.

As a result of this motion the Davis Strait continental block was subjected to wrenching and strike-slip motion. However, the motion was not great enough to cause the oceanic crusts of Baffin Bay and the Labrador Sea to link up and a bridge of continental crust was maintained between Greenland and Baffin Island (Fig. 14). During this wrench motion along the Ungava transform (Fig. 14), copeous quantities of basalts were erupted over the area. One might expect that in addition to the location of vents along the main faults there were also volcanic cones along the trend of this transform fault.

Conclusion

The reconstruction described here satisfies all the constraints imposed except for the small motion along Nares Strait. There appears at this stage no acceptable reconciliation between the onshore evidence along Nares Strait that suggests geological continuity between Ellesmere Island and Greenland and the interpretation of crustal history. However, throughout this study several lines of evidence indicate that from Cretaceous through to the present day, the histories of the Arctic Ocean basin and the North Atlantic basin (including Baffin Bay and Labrador Sea) are intimately related. Furthermore, the Nares Strait problem only exists if the Arctic Islands are held to be a separate plate from Greenland. Consequently it is suggested that further resolution of the problem rests with the pursuit of the reconstruction of the Arctic Ocean.

Acknowledgements

I am grateful to Esso Resources Canada Ltd. for permission to use company data and for permission to publish this paper. I thank Richard R. Boudreau and Peter N. Byers of Esso Resources Canada Ltd. for their help and guidance in the analysis of Baffin Bay. I am particularly grateful to Peter R. Dawes for his editorial help and for his untiring efforts in preparing this contribution for publication.

References

Andrews, J. T., Guennel, G. K., Wray, J. L. & Ives, J. D. 1972. An early Tertiary outcrop in north-central Baffin Island, Northwest Territories, Canada: environment and significance. – Can. J. Earth Sci. 9: 233–238.

Balkwill, H. R. 1978. Evolution of Sverdrup Basin, Arctic Canada. – Bull. Am. Ass. Petrol. Geol. 62: 1004–1028.

Basham, P. W., Forsyth, D. A. & Wetmiller, R. J. 1977. The seismicity of northern Canada. – Can. J. Earth Sci. 14: 1646–1667.

Beh, R. L. 1975. Evolution and geology of western Baffin Bay and Davis Strait, Canada. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 453–476.

Berry, M. J. & Fuchs, K. 1973. Crustal structure of the Superior and Grenville Provinces of the northeastern Canadian Shield. – Bull. seism. Soc. Am. 63: 1393–1432.

Bridgwater, D., Watson, J. & Windley, B. F. 1973. The Archaean craton of the North Atlantic region. – Phil. Trans. Roy. Soc. Lond. 273A: 493–512.

Clarke, D. B. 1975. Tertiary basalts dredged from Baffin Bay. – Can. J. Earth Sci. 12: 1396–1405.

Clarke, D. B. & Pedersen, A. K. 1976. Tertiary volcanic province of West Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 364–385. – Geol. Surv. Greenland, Copenhagen.

Clarke, D. B. & Upton, B. G. J. 1971. Tertiary basalts of Baffin Island: field relations and tectonic setting. – Can. J. Earth Sci. 8: 248–258.

Douglas, R. J. W. 1967. Geological map of Canada. 1:5 000 000. – Map geol. Surv. Can. 1250A.

Douglas, R. J. W. 1970. Geology and economic minerals of Canada. – Econ. Geol. Rep. geol. Surv. Can. 1: 838 pp.

Douglas, R. J. W., Poole, W. H., Wade, J. A. & Barss, M. S. 1979. Geology of eastern Canada and adjacent areas. – Map geol. Surv. Can. 1401A.

Escher, A., Henriksen, N., Dawes, P. R. & Weidick, A. 1970. Tectonic/geological map of Greenland. 1: 2 500 000. – Geol. Surv. Greenland, Copenhagen.

Escher, A. & Watt, W. S. 1976. Geology of Greenland. – Geol. Surv. Greenland, Copenhagen: 603 pp.

Grant, A. C. 1975. Geophysical results from the continental margin off southern Baffin Island. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 411–431.

Healy, J. H. & Warren, D. H. 1969. Explosion seismic studies in North America. – In: Hart, P. J. (ed.), The Earth's crust and upper mantle. – Mem. Am. geophys. Un. 13: 208–219.

Henderson, G., Rosenkrantz, A. & Schiener, E. J. 1976. Cretaceous–Tertiary sedimentary rocks of West Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 340–362. – Geol. Surv. Greenland, Copenhagen.

Hinz, K., Schlüter, H.-U., Grant, A. C., Srivastava, S. P., Umpleby, D. & Woodside, J. 1979. Geophysical transects of the Labrador Sea: Labrador to southwest Greenland. – Tectonophysics 59: 151–183.

Hyndman, R. D. 1973. Evolution of the Labrador Sea. – Can. J. Earth Sci. 10: 637–644.

Hyndman, R. D., Clarke, D. B., Hume, H., Johnson, J., Keen, M. J., Park, I. & Pye, G. 1971. Geophysical and geological studies in Baffin Bay and the Labrador Sea. – In: Hood, P. J. (ed.), Earth science symposium on offshore eastern Canada. – Pap. geol. Surv. Can. 71–23: 621–631.

Jackson, G. D. & Taylor, F. C. 1972. Correlation of major Aphebian rock units in the northeastern Canadian Shield. – Can. J. Earth Sci. 9: 1650–1669.

Jackson, H. R., Keen, C. E. & Barrett, D. L. 1977. Geophysical studies on the eastern continental margin of Baffin Bay and in Lancaster Sound. – Can. J. Earth Sci. 14: 1991–2001.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor spreading in central Baffin Bay. – Can. J. Earth Sci. 16: 2122–2135.

Keen, C. E. & Barrett, D. L. 1972. Seismic refraction studies in Baffin Bay: an example of a developing ocean basin. – Geophys. J. Roy. astr. Soc. 30: 253–271.

Keen, C. E. & Hyndman, R. D. 1979. Geophysical review of the continental margins of eastern and western Canada. – Can. J. Earth Sci. 16: 712–747.

Keen, C. E., Barrett, D. L., Manchester, K. S. & Ross, D. I. 1972. Geophysical studies in Baffin Bay and some tectonic implications. – Can. J. Earth Sci. 9: 239–256.

Kerr, J. W. 1967. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

King, A. F. & McMillan, N. J. 1975. A mid-Mesozoic breccia from the coast of Labrador. – Can. J. Earth Sci. 12: 44–51.

LaBrecque, J. L., Kent, D. V. & Cande, S. C. 1977. Revised magnetic polarity time scale for Late Cretaceous and Cenozoic time. – Geology 5: 330–335.

Lewis, B. T. R. 1978. Evolution of ocean crust seismic velocities. – Ann. Rev. Earth planet. Sci. 6: 377–404.

Loken, O. H. & Hodgson, D. A. 1971. On the submarine geomorphology along the east coast of Baffin Island. – Can. J. Earth Sci. 8: 185–195.

MacLean, B. 1978. Marine geological-geophysical investigations in 1977 of the Scott Inlet and Cape Dyer-Frobisher Bay areas of the Baffin Island continental shelf. – Pap. geol. Surv. Can. 78–1B: 13–20.

MacLean, B. & Falconer, R. K. H. 1979. Geological/geophysical studies in Baffin Bay and Scot Inlet-Buchan Gulf and Cape Dyer-Cumberland Sound areas of the Baffin Island Shelf. – Pap. geol. Surv. Can. 79–1B: 231–244.

MacLean, B., Falconer, R. K. H. & Clarke, D. B. 1978. Tertiary basalts of western Davis Strait: bedrock core samples and geophysical data. – Can. J. Earth Sci. 15: 773–780.

McMillan, N. J. 1973. Shelves of Labrador Sea and Baffin Bay, Canada. – In: McCrossan, R. G. (ed.), Future petroleum provinces of Canada, their geology and potential. – Mem. Can. Soc. Petrol. Geol. 1: 473–517.

Parsons, B. & Sclater, J. G. 1977. An analysis of the variation of ocean floor bathymetry and heat flow with age. – J. geophys. Res. 82: 802–827.

Pelletier, B. R., Ross, D. I., Keen, C. E. & Keen, M. J. 1974. Geology and geophysics of Baffin Bay. – Pap. geol. Surv. Can. 74–30(2): 247–258.

Qamar, A. 1974. Seismicity of the Baffin Bay region. – Bull. seism. Soc. Am. 64: 87–98.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

van der Linden, W. J. M. 1975. Crustal attenuation and sea-floor spreading in the Labrador Sea. – Earth planet Sci. Lett. 27: 409–423.

Watt, W. S. 1969. The coast-parallel dike swarm of southwest Greenland in relation to the opening of the Labrador Sea. – Can. J. Earth Sci. 6: 1320–1321.

Wetmiller, R. J. 1974. Crustal structure of Baffin Bay from the earthquake-generated Lg phase. – Can. J. Earth Sci. 11: 123–130.

Problems with plate tectonic models for Baffin Bay – Nares Strait: evidence from the Labrador Sea

ALAN C. GRANT

Grant, A. C. 1982. Problems with plate tectonic models for Baffin Bay – Nares Strait: evidence from the Labrador Sea. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 313–326.

Plate tectonic models for the geologic development of Baffin Bay require major strike-slip displacement through Nares Strait. However, these models are based mainly upon observations made in other areas, such as the Labrador Sea, the North Atlantic, or the Arctic Ocean. This paper reviews a synthesis of available geological and geophysical data from the Labrador Sea that indicate severe problems with plate tectonic models for that area, which must also affect any such model for Baffin Bay. The Labrador Sea data can be interpreted alternatively in terms of vertical crustal movements. Applied to Baffin Bay, such a mechanism involving vertical rather than lateral crustal displacements accommodates the evidence from field studies in the Nares Strait area that there is geological continuity from Ellesmere Island to Greenland.

A. C. Grant, Department of Energy, Mines and Resources, Geological Survey of Canada, Bedford Institute of Oceanography, Dartmouth, Nova Scotia, Canada, B2Y 4A2.

The thesis of this paper is that there are clearly demonstrable problems with plate tectonic models for the Labrador Sea, and that these problems must affect any such model for the geological development of Baffin Bay. Regarding the question of whether Greenland has drifted along Nares Strait, it is only in terms of rigid-plate models that lateral movements are required in Nares Strait. This paper therefore disputes the occurrence of such movements.

This paper is mainly a summary of a recent interpretation of geological and geophysical data from the Labrador Sea that defines specific problems with conventional plate tectonic models for that area (Grant 1980). The conclusions of that study agreed with the then available evidence for geologic continuity across Nares Strait (e.g. Kerr 1967, Dawes 1976, Christie et al. 1978). It is relevant to review that study in the light of the extensive additional evidence for geologic continuity from Ellesmere Island to Greenland that has been presented in this symposium. In addition, exploratory wells off West Greenland, and regional magnetic and gravity maps of the Canadian Arctic Islands and northern Greenland, provide further evidence that is pertinent to the question of displacement along Nares Strait.

Problems with plate tectonic models – regional considerations

The most direct method of demonstrating the reality of problems with plate tectonic-type models for the Baffin Bay – Labrador Sea region (Fig. 1) is to compare some of the models that have been proposed. Grant (1980) assembled a number of pre-drift fits of Greenland relative to Canada (Fig. 2), and divided them into several categories: (1) 'historical' (a, b); (2) fits based on studies of the North Atlantic (c–f); (3) fits based on Arctic Ocean studies (g–i); (4) fits based on studies of Baffin Bay and Greenland (j–m); (5) fits based on studies of the Labrador Sea (n–p). The principal point illustrated by this collection is the great variety in the reconstructions, not only between the categories defined but within them as well. It is clear that the resultant fit depends upon the particular data used, or ignored. For example, each study in categories 4 and 5 has juxtaposed the Tertiary volcanics at Cape Dyer, Baffin Island, with those in the Disko region of West Greenland (Fig. 1). Greenland can be pivoted about this point to close either Baffin Bay, or Labrador Sea, as preferred. Studies centred on the North Atlantic or the Arctic Ocean (categories 2 & 3) arrive at quite different reconstructions (Fig. 2).

In the category labelled 'historical' (Fig. 2a, b), fit b is the well known reconstruction by Bullard et al. (1965). Fit a is by Choubert (1935); it appears to be about as satisfactory as any of the fits in this collection (Fig. 2).

The compilation of reconstructions in Fig. 2 is arbitrary in many respects, and it should not be considered as a criticism of any individual piece of work. The discrepancies among these reconstructions, however, should be recognized as a demonstration of the fact that there are indeed problems with plate tectonic-type models for the Labrador Sea – Baffin Bay region, and that no consensus derives from these studies regarding the geologic history of Nares Strait.

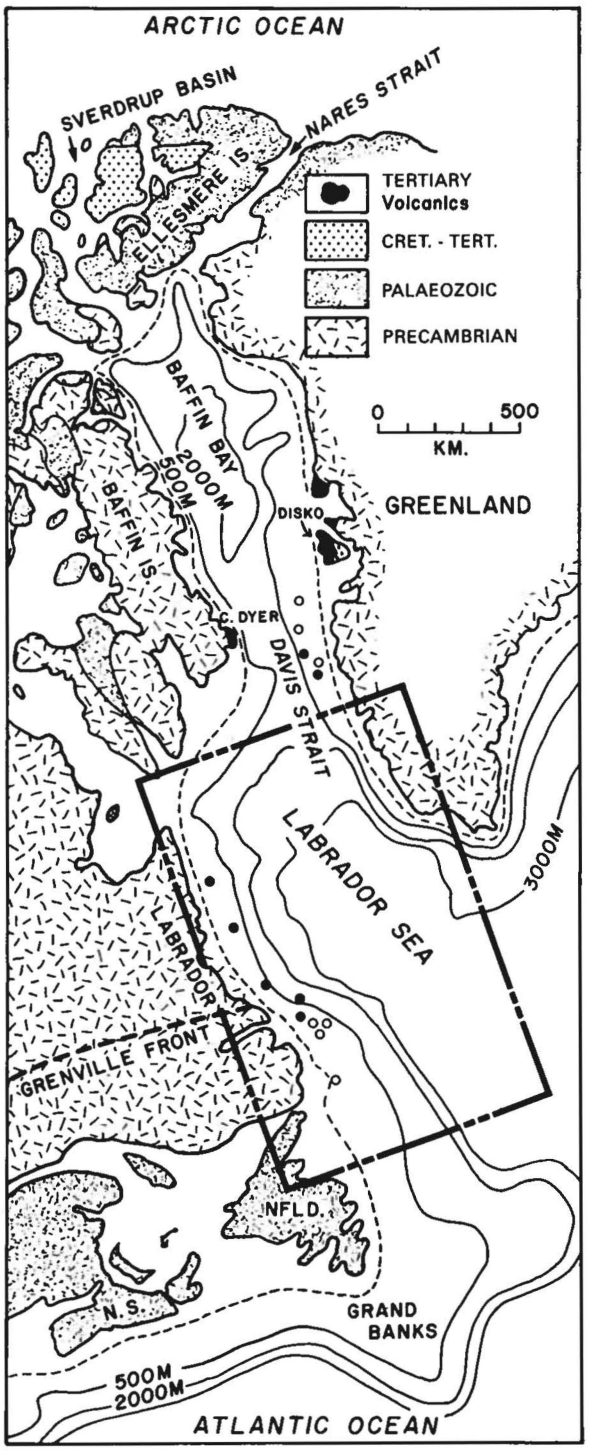


Fig. 1. General geology and bathymetry of the Labrador Sea – Baffin Bay region. The dashed line offshore traces the approximate landward edge of Cretaceous–Tertiary deposits. Circles on shelf mark locations of exploratory wells; filled circles are wells that bottomed in Precambrian rocks. Dashed rectangle shows area of Fig. 3. After Grant (1980).

Specific problems, Labrador Sea

Multichannel seismic data and stratigraphic control from exploratory wells drilled on the Labrador Shelf (Fig. 3) provide evidence of specific problems with plate tectonic models for the Labrador Sea (Grant 1980). The definition of these problems is based upon interpretation of seismic reflectors in two zones of the Labrador margin, as illustrated in Fig. 4. Firstly, a buried erosional surface on the shelf in the depth interval 1 to 2 seconds (2-way time) has been tentatively dated as Late(?) Miocene in age (arrow 1, Fig. 4D), using available seismic and well data. An example of the character of the reflectors expressing this surface is shown in Fig. 5. Mapping this surface shows that it deepens from a minimum depth of about 0.5 km off southern Labrador to more than 2 km off northern Labrador (Fig. 6). This post-Late(?) Miocene differential in subsidence of the shelf from south to north is also evident from drilling results, and in addition, Gradstein & Srivastava (1980) have shown that the rate of subsidence of the Labrador Shelf has been increasing in late Cenozoic time (Fig. 7). This increase in the rate of subsidence, and the differential in subsidence from south to north shown by the Late(?) Miocene erosional surface (Fig. 6), are contrary to the predictions of conventional plate tectonic models for this region. Such models contend that sea-floor spreading ended in the Labrador Sea in Early Oligocene time (e.g. pre-anomaly 13, Srivastava 1978), and therefore the rate of subsidence of the adjacent continental margin should have decreased since then, at a rate proportional to the square root of time (e.g. Sclater et al. 1971). Thus the data compiled in Fig. 7 indicate that the Labrador Shelf has experienced tectonism in late Cenozoic time that is not anticipated or explained by conventional plate tectonic models.

The second point of interpretation concerns the strong, smooth reflector buried deeply beneath the continental slope and rise off northern Labrador (arrow 2, Fig. 4D). An example of the character of this reflector is presented in Fig. 8. Grant (1980) proposed that this reflector represents an erosional unconformity, correlative with the strongly reflective erosional surface that is developed regionally beneath the shelf off southern Labrador (Fig. 4A) and Newfoundland, on rocks ranging in age from early Palaeozoic to Early Cretaceous. This surface on the Grand Banks has been referred to as the Avalon Unconformity (Jansa & Wade 1975), of Early Cretaceous age (Amoco Canada Petroleum Company Ltd & Imperial Oil Ltd 1973). This unconformity is provisionally designated as 'Early Cretaceous' in age off Labrador; a corresponding hiatus occurs in all wells on the Labrador Shelf (Umpleby 1979). This unconformity is downwarped to the north across the zone where the Grenville Front projects seaward beneath the shelf (Fig. 9). South of this zone the unconformity slopes gently seaward (Fig. 4A), but off northern Lab-

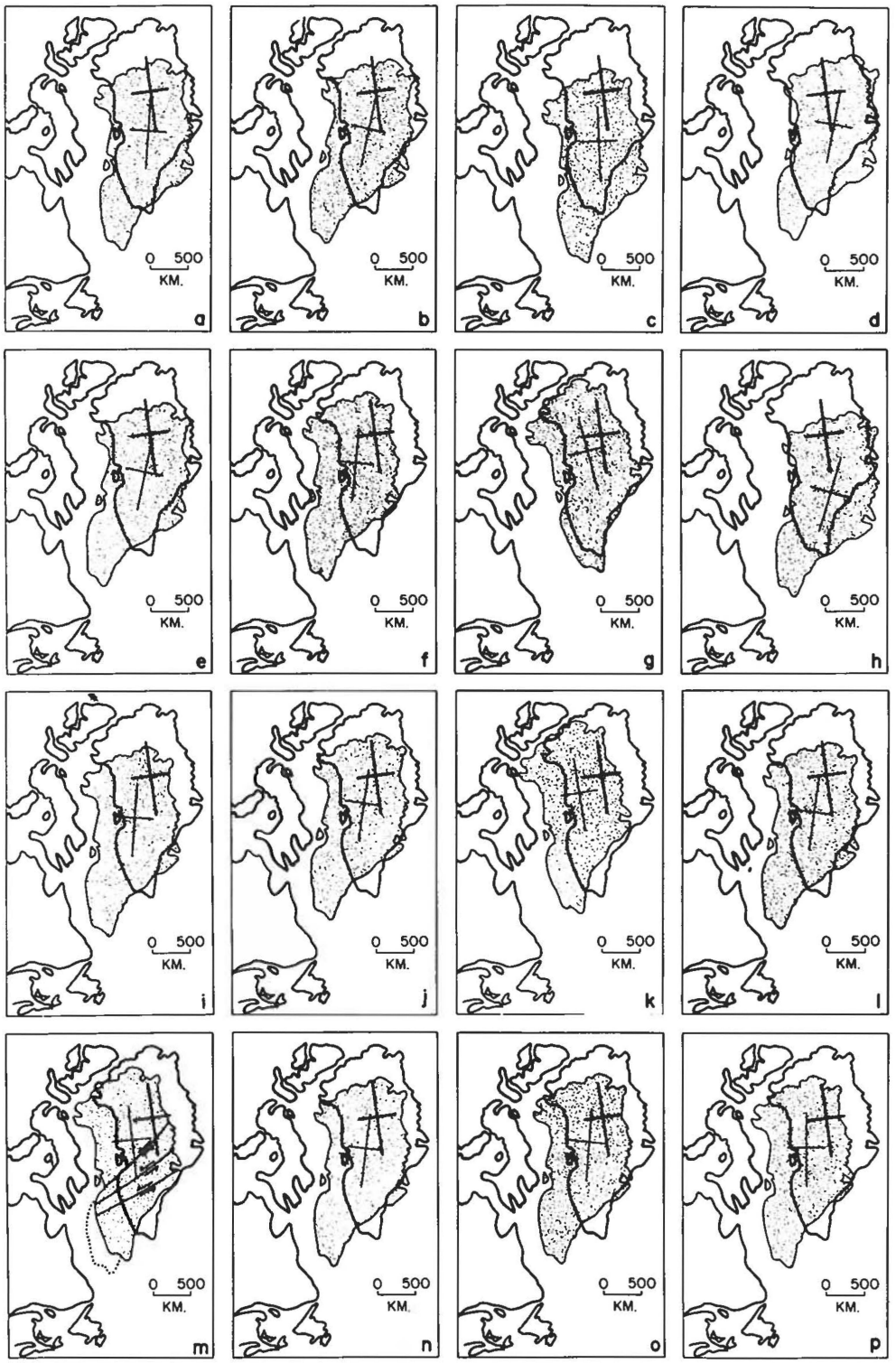


Fig. 2. Pre-drift fits for the Labrador Sea – Baffin Bay region by various authors. a, Choubert (1935) as reviewed by van Houten (1975); b, Bullard et al. (1965); c, Pitman & Talwani (1972); d, Laughton (1972); e, Le Pichon et al. (1977); f, Kristoffersen (1978); g, h, Churkin (1973); i, Harland (1973); j, Keen et al. (1972); k, Martin (1973); l, Bridgwater et al. (1973); m, Beh (1975); n, Le Pichon et al. (1971); o, Hyndman (1973); p, Srivastava (1978). The shaded areas indicate inferred past positions of Greenland; crosses show relative translation and rotation. After Grant (1980).

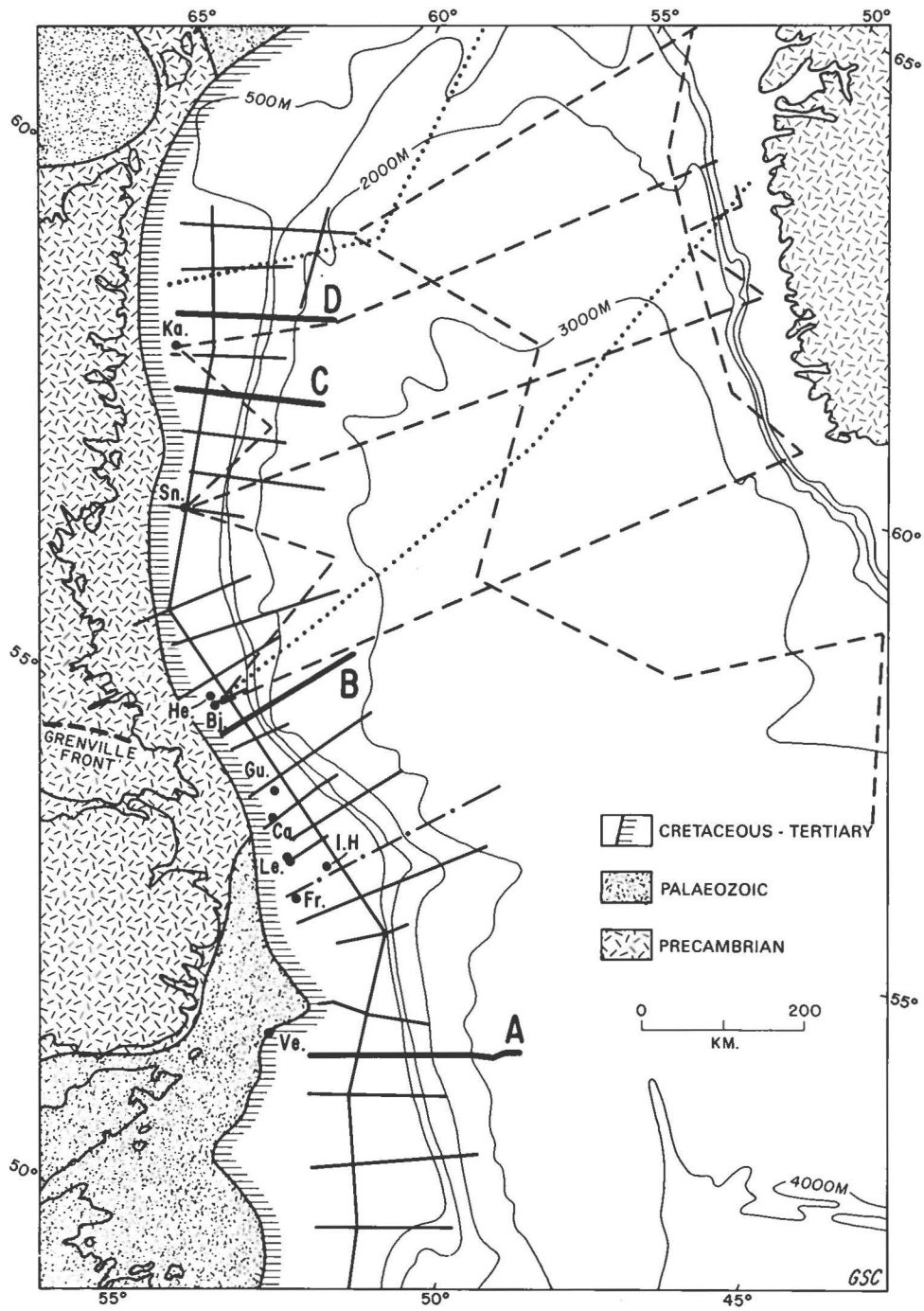


Fig. 3. Geology and bathymetry of the Labrador Sea region, with locations of multichannel seismic lines (solid, Seiscan-Delta; dash-dot, Catalina; dashed, B.G.R.; dotted, C.G.G.) and exploratory wells (Ka, Karlsefni H–13; Sn, Snorri J–90; He, Herjolf M–92; Bj, Bjarni H–81; Gu, Gudrid H–55; Ca, Cartier D–70; Le, Leif M–48 and E–38; I. H., Indian Harbour M–52; Fr, Freydis B–87; Ve, Verrazano L–77). Letters A to D on heavy lines indicate locations of diagrams of seismic profiles in Fig. 4. After Grant (1980).

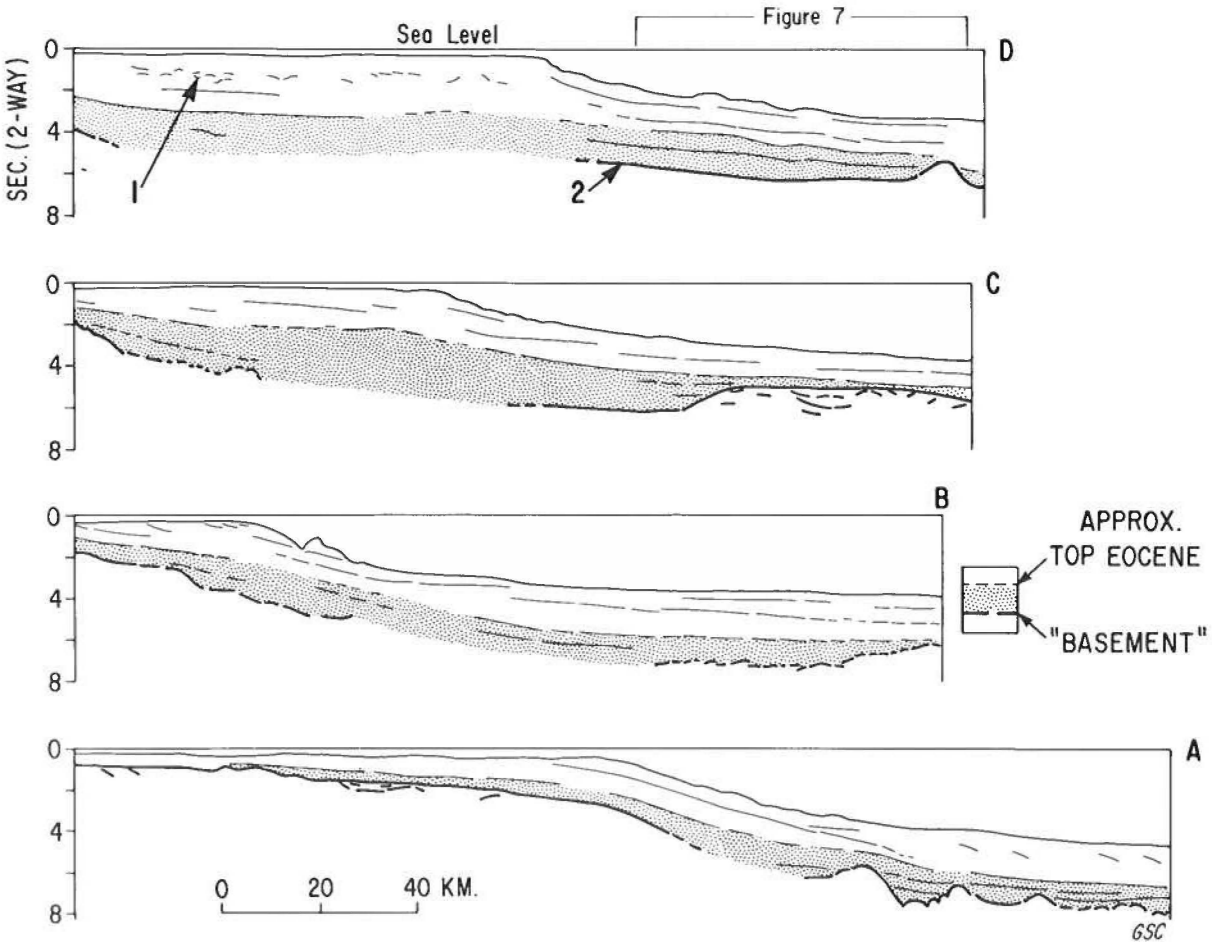


Fig. 4. Diagrams of multichannel seismic profiles of the Labrador margin (locations shown in Fig. 3). The 'basement' reflector is emphasized by a heavy line. Arrows 1 and 2 (profile D) denote points of interpretation discussed in text. After Grant (1980).

rador it is broken by faults and deepens rapidly beneath the inner shelf (Fig. 4B–D). Cretaceous and Early Tertiary sediments on the Labrador Shelf occupy a linear, coast-parallel graben (Cutt & Laving 1977, Umpleby 1979, Miall et al. 1980), dubbed the Erik Graben by N. J. McMillan (pers. comm.). As mapped and described by Cutt & Laving (1977), to the south of the Grenville Front zone this graben is developed in continental basement rocks within the area of the well-preserved Early Cretaceous unconformity (Fig. 9). It can be seen in Fig. 9 that the area mapped as continental crust beneath the slope and rise off northern Labrador is reasonably explained as the downwarped, northward continuation of the continental basement rocks that lie on the seaward side of this graben.

If the reflector mapped beneath the slope and rise off northern Labrador (Fig. 9) indeed represents the 'Early Cretaceous unconformity', then it must be underlain by rocks that are at least 100 m.y. old, and presumably continental in origin. Conflicts of this interpretation with plate tectonic models are indicated in Fig. 9, where magnetic lineations are plotted that have been interpreted (Srivastava 1978) to represent sea-floor spreading-type magnetic anomalies. Anomaly 28, which should represent oceanic crust about 65 m.y. old, lies within the area interpreted to be underlain by continental crust more than 100 m.y. old. This conflict poses questions not only as to whether the magnetic anomalies have been correctly dated, but also whether the magnetic anomalies can be 'true' sea-floor spreading-type anomalies. If the interpretation based on seismic evidence is correct, both these questions must be answered negatively, which prompts examination of linear magnetic anomalies in the Labrador Sea on a more regional scale. This larger question has been touched on by Hinz et al. (1979), who mapped six different types of oceanic basement in the Labrador Sea on the basis of seismic reflection character. The boundaries of these basement zones appear to be independent of the trends and occurrence of linear magnetic anomalies. This indicates that the source of the magnetic anomalies lies deeper than the surface mapped as oceanic base-

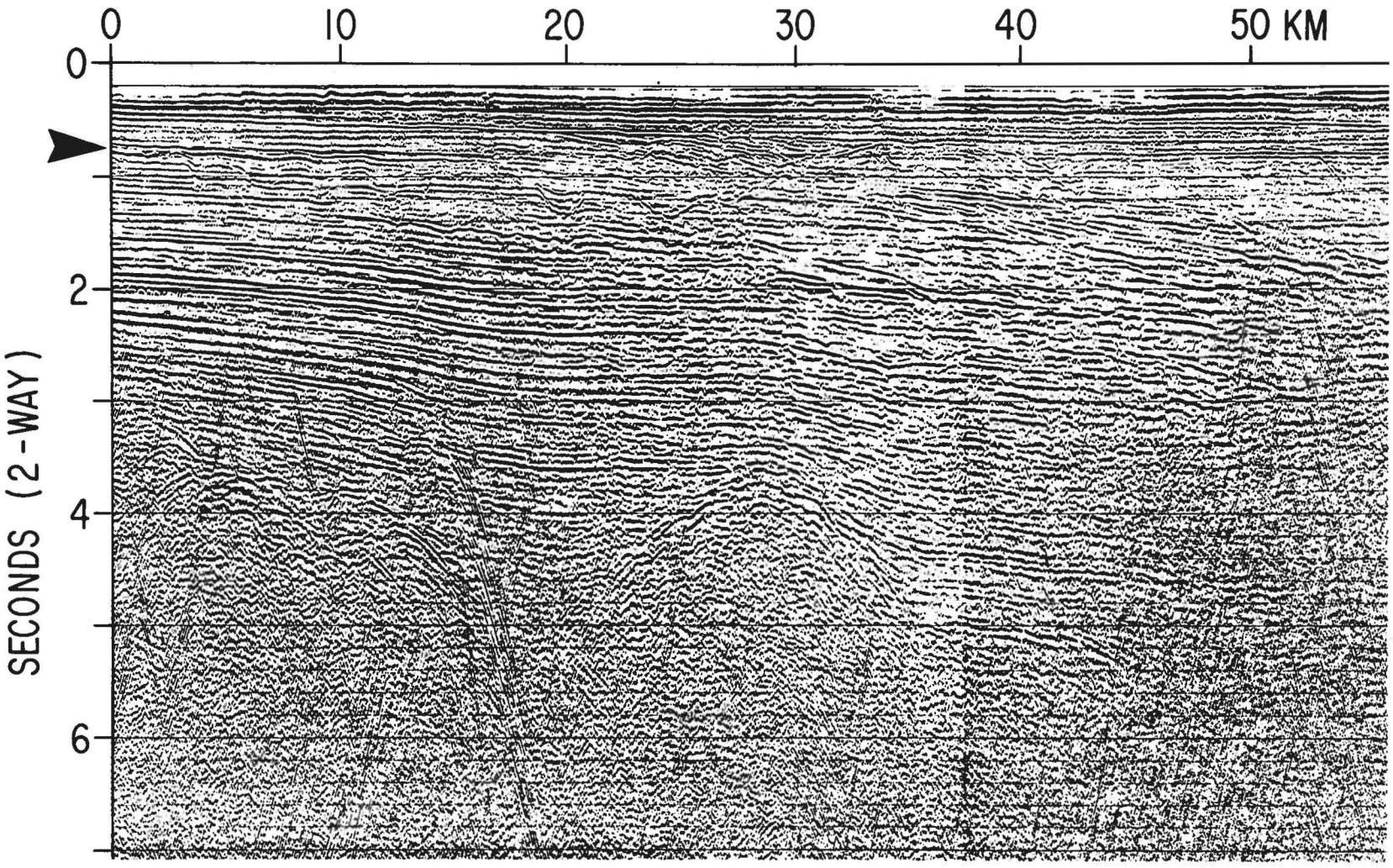


Fig. 5. Photograph of a representative seismic section from the continental shelf off northern Labrador (location shown in Fig. 6), reproduced with permission of the Federal Institute for Geosciences and Natural Resources, Hannover. Arrow indicates level of reflectors interpreted as representing a Late(?) Miocene erosional surface. After Grant (1980).

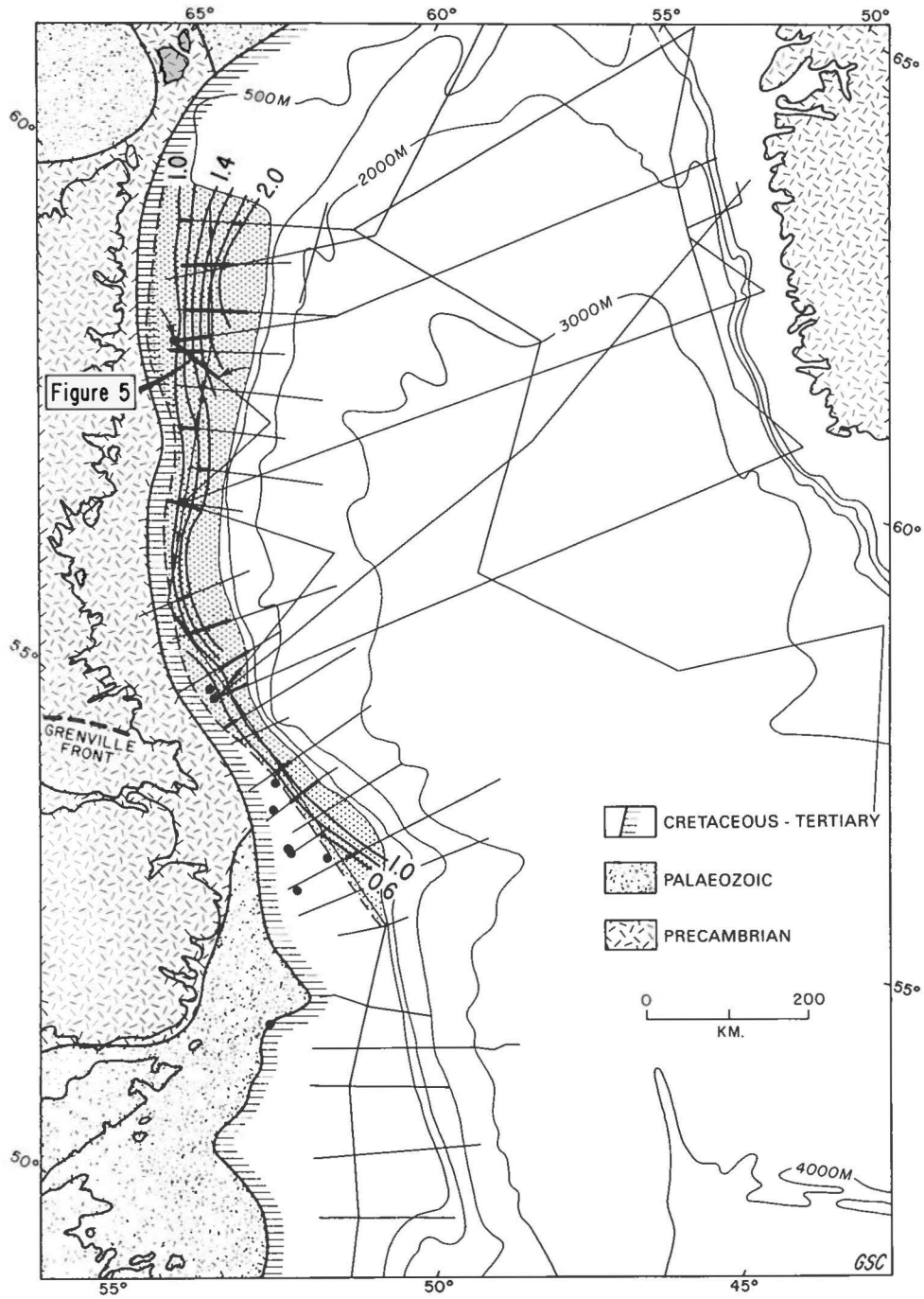


Fig. 6. Generalized bathymetry and geology of the Labrador margin, and lines of multichannel seismic coverage. Stippled area represents the mapped extent of the inferred Late(?) Miocene erosional surface. Heavy parts of the seismic lines show where this surface is represented by an irregular seismic reflector. The contours on this surface (heavy lines) are in seconds of reflection time. Curved arrows mark extent of record section in Fig. 5. After Grant (1980).

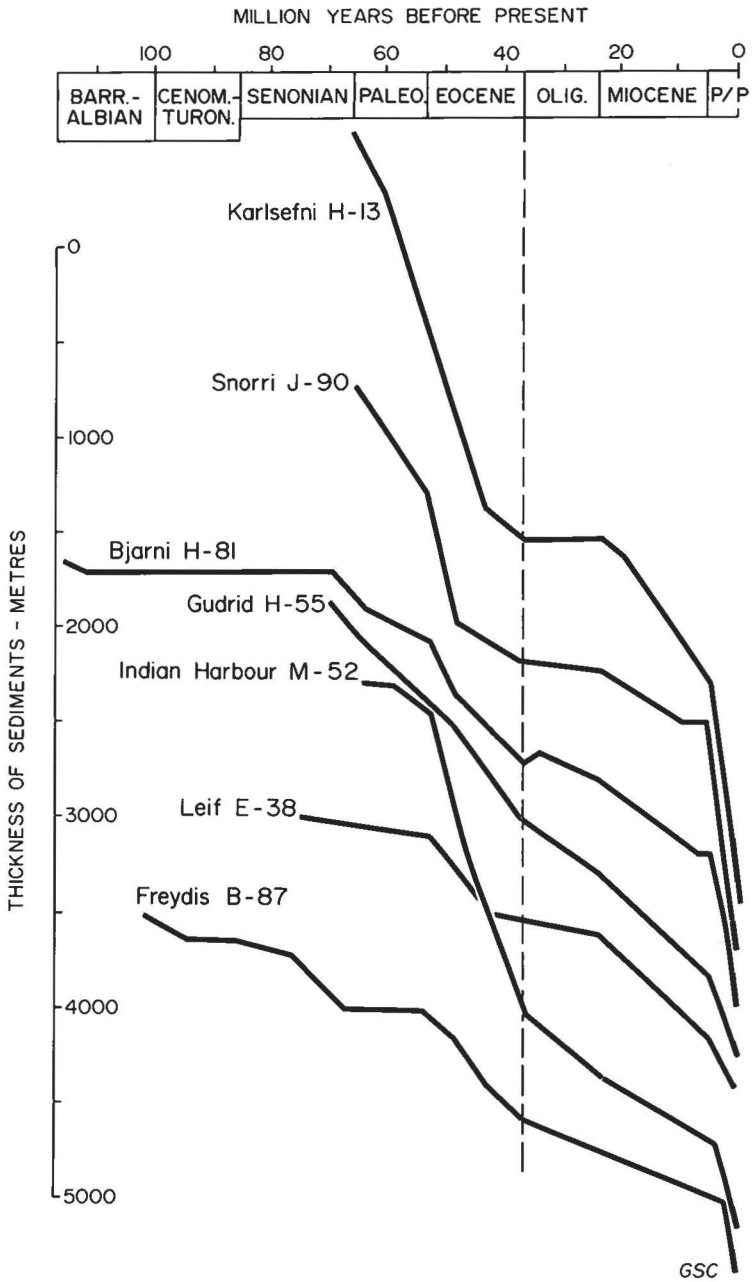


Fig. 7. Curves showing sediment accumulation for wells on the Labrador Shelf (locations in Fig. 3), from data published by Gradstein & Srivastava (1980). After Grant (1980).

ment, and that the formation of oceanic basement and the generation of linear magnetic anomalies are separate processes. It is difficult to reconcile this relationship with a Vine & Matthews (1963) 'tape recorder' mechanism for the formation of oceanic crust. Further, the remarkable parallelism of linear magnetic anomalies and linear gravity anomalies in the Labrador Sea (see Srivastava 1978: fig. 7) suggests that the magnetic anomalies reflect a much more substantial origin than might be expected simply from changes in the polarity of magnetization of the basement rocks. A buried system of linear fractures, which underwent differential vertical displacements and also served as conduits for volcanic extrusion, would be among the simplest sets of conditions sufficient to explain these magnetic and gravity anomalies. Linear fractures, with differential vertical displacements, are well developed in the continental crust beneath the Labrador Shelf; possibly the

basalts that have been drilled on the Labrador Shelf (McMillan 1980) were extruded through such fractures.

The work of Hinz et al. (1979) also indicates problems with plate tectonic-type interpretations of other geophysical data from the Labrador Sea. They concluded that the magnetic and gravity criteria of Talwani & Eldholm (1973) cannot be used to define the continent–ocean crustal boundary in the Labrador Sea (see dotted line in Fig. 9). Moreover, they reinterpreted the refraction seismic data of van der Linden (1975), and deduced a different location for the continent–ocean crustal boundary. This fundamental difference between interpretations of refraction measurements is an indication of the equivocal nature of refraction seismic depths and velocities for defining continental versus oceanic crust.

Implications of these problems

The points of interpretation outlined above indicate that plate tectonic models fail to predict or explain the post-Late(?) Miocene subsidence of the Labrador Shelf, or satisfactorily define the location of the continent–ocean crustal boundary in the Labrador Sea. Specific observations in the Labrador Sea indicate that the dating of sea-floor spreading-type magnetic anomalies is in error by 30–40 million years, or that the anomalies were not produced by sea-floor spreading processes as currently visualized. Examination of magnetic and gravity anomalies on a regional scale raises further questions regarding sea-floor spreading processes, to the extent that alternative mechanisms must be considered in examining the origin of the Labrador Sea. An obvious alternative is that the Labrador Sea formed largely as a result of differential vertical crustal displacements; there is ample evidence of such displacements amounting to several kilometres, recorded by the accumulation of Cretaceous and Tertiary sediments on the Labrador Shelf.

The basic problem with plate tectonic models for the Labrador Sea thus concerns the inferred mechanisms of such models, which derive from assumed relationships of observed physical properties of the crust to crustal origin. In the Labrador Sea there is a proven inability of currently 'standard' geophysical measurements to distinguish continental from oceanic crust; as outlined above, specified velocities and depths to mantle, and the occurrence of linear magnetic anomalies apparently are not sufficient criteria to make this distinction.

If the geologic history of Baffin Bay is linked to that of Labrador Sea, as implied by rigid-plate models, the efficacy of sea-floor spreading-type processes in Baffin Bay must also be questioned. Formation of Baffin Bay by vertical rather than lateral crustal displacements agrees with the evidence for stratigraphic and structural continuity across Nares Strait.

Evidence from the Baffin Bay – Nares Strait region

The northernmost exploratory wells in the Labrador Sea – Baffin Bay seaway are located on the continental shelf off West Greenland, in Davis Strait (Fig. 1). Two of these exploratory wells bottomed in Precambrian basement (Geological Survey of Greenland 1979), which proved that continental crust in these places extends to the shelf edge, and clearly precludes some of the close fits in Davis Strait indicated in Fig. 2. This geologic evidence must be recognized as an additional constraint in attempting plate tectonic-type reconstructions for this region, and in the context of this symposium it can be viewed as supporting the case for geologic continuity across Nares Strait.

Much of this symposium is concerned with new geological observations from the vicinity of Nares Strait; relatively little has been added to the broad geophysical picture of the Baffin Bay region that has existed for some years. Marine geophysical data generally have been interpreted as evidence that Baffin Bay is a small ocean basin formed by sea-floor spreading processes (e.g. Keen et al. 1974). The time of opening of Baffin Bay usually is inferred by extrapolation from the Labrador Sea, as it is difficult to define and identify magnetic lineations in Baffin Bay. Jackson et al. (1979) mapped linear magnetic anomalies in an area of detailed surveying in central Baffin Bay, and concluded that these anomalies were formed by sea-floor spreading processes. They also report a prominent, linear gravity low extending through the survey area parallel to the magnetic lineations. This is the same relationship between gravity and magnetic anomalies that has been noted in the Labrador Sea. Thus the available data suggest that the same problems outlined for the Labrador Sea regarding the validity of traditional geophysical criteria in defining continental versus oceanic crust may also pertain to Baffin Bay, although much additional geophysical and stratigraphic control is required in Baffin Bay to test for these problems more closely.

Magnetic and gravity data from the Canadian Arctic Islands and northern Greenland may relate more directly to the Nares Strait 'question' than available geophysical observations in Baffin Bay. The regional magnetic and gravity data presented in Figs 10 and 11 are generalized after Coles et al. (1976) and Sobczak (1978) respectively. Despite the drastic generalization, these two maps show trends, and lack of trends, that may bear upon the question of whether strike-slip displacement has occurred along Nares Strait.

The major trends in the magnetic field (Fig. 10) are traced by the boundaries of Magnetic Regions drawn by Coles et al. (1976). The southern boundary of their Magnetic Region IV cuts obliquely across the Nares Strait lineament. South of this boundary several linear positive anomalies trend across Nares Strait. Thus the

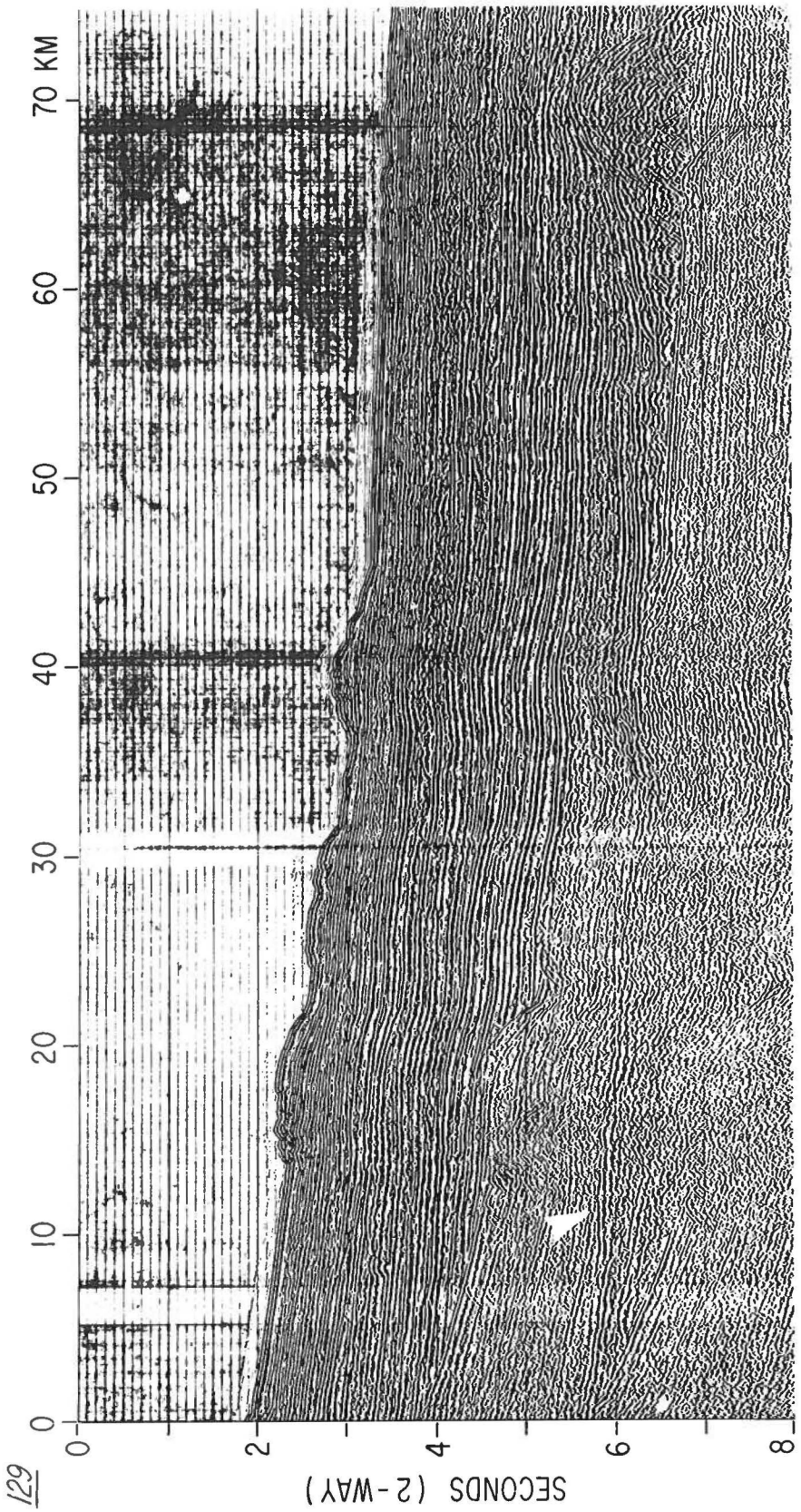


Fig. 8. Photograph of a representative seismic record section from the continental slope off northern Labrador (location shown in Fig. 9), reproduced with permission of Seiscan Delta Ltd., Calgary. Arrow indicates the reflector interpreted to represent the 'Early Cretaceous unconformity'. After Grant (1980).

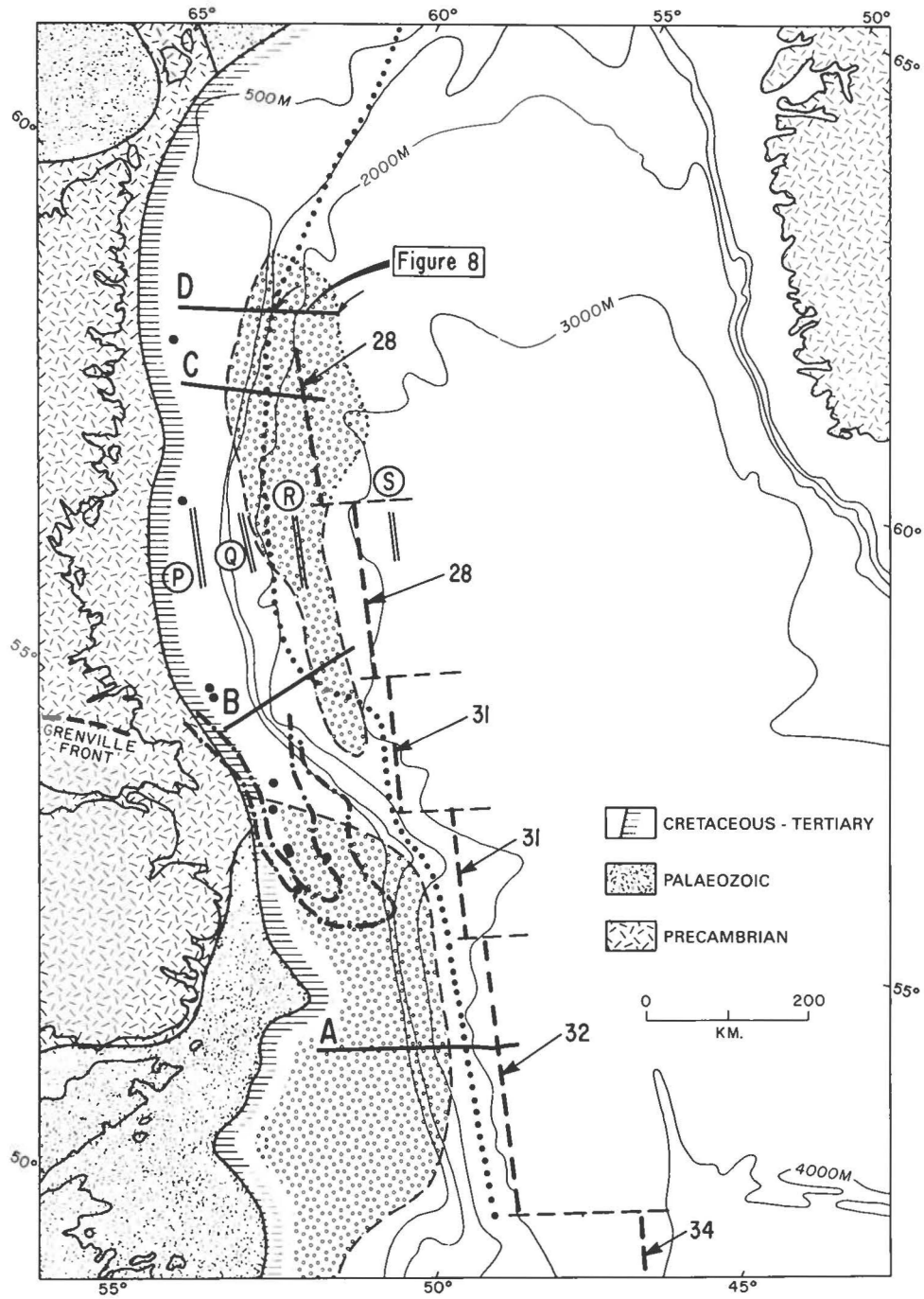


Fig. 9. Generalized bathymetry and geology of the Labrador margin. Shading (circle pattern) indicates inferred minimum extent of smooth basement reflector interpreted to represent an Early Cretaceous unconformity. Double lines labelled P to S (circled) locate refraction seismic profiles (van der Linden 1975). Letters A to D denote locations of seismic profiles in Fig. 4. Dash-dot lines on the shelf represent the extent of Early Cretaceous (inner line) and Paleocene (outer line) sediments (after Cutt & Laving 1977). Numbered dashed lines (28 to 34) represent magnetic anomalies, and the dotted line is the continent–ocean boundary, according to Srivastava (1978). Curved arrows mark extent of record section in Fig. 8. After Grant (1980).

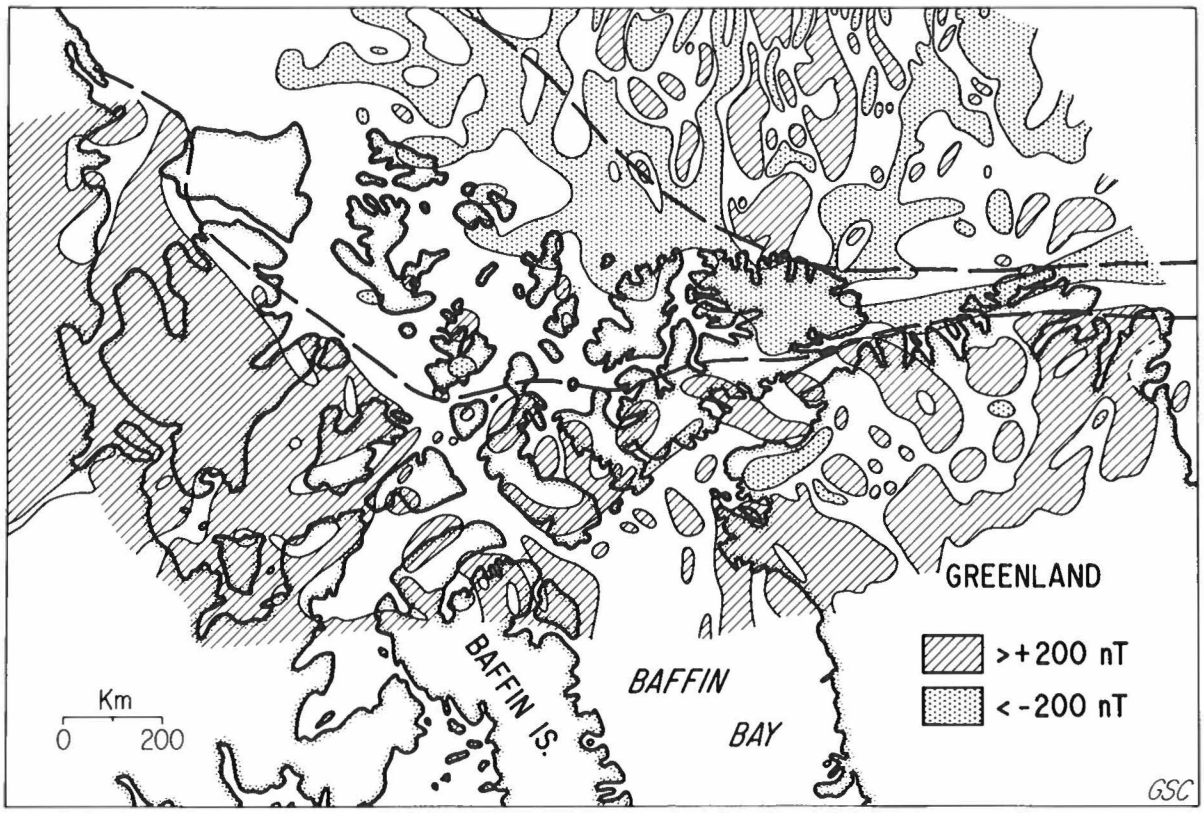


Fig. 10. Vertical component magnetic field residuals, relative to the IGRF. After Coles et al. (1976). Heavy dashed lines are boundaries of magnetic regions defined by these authors.

trends indicated by the magnetic data in Fig. 10 are strong geophysical evidence against displacement in Nares Strait.

The gravity map (Fig. 11) does not show any pronounced trends in the Nares Strait area. This lack of gravity expression can be viewed as an indication that no major crustal disturbance has occurred in the Nares Strait zone, in particular of the subduction type that is required by some of the reconstructions presented in Fig. 2 (e.g. fits a–e, h, i, n).

Conclusions

Discrepancies among plate tectonic-type models for the Labrador Sea – Baffin Bay region indicate that such models do not yield a unique solution to the Nares Strait 'problem'. In the Labrador Sea specific difficulties can be defined with both the predictions and the mechanisms of plate tectonic models, that raise substantial questions concerning the extent to which the Labrador Sea is primarily the product of sea-floor spreading processes. In any 'rigid-plate' scheme these same problems must extend to Baffin Bay, where they may be viewed as a corollary to the geologic arguments for continuity across Nares Strait.

The basic problem with plate tectonic models for this region lies in the ambiguous nature of geophysical data for defining continental versus oceanic crust; available data in the Labrador Sea document the possibility that 'geophysical' oceanic crust may have formed there by processes other than sea-floor spreading. While this may be a condition unique to the Labrador Sea, probably it is worth testing whether geophysical measurements are necessarily more definitive elsewhere. The outstanding requirement at this time is therefore to devise new or refined geophysical measurements of crustal properties that may be more definitive of crustal origin.

Acknowledgements

I thank A. E. Jackson for assistance in compiling data, G. M. Grant and G. L. Cook for preparing illustrations, C. F. M. Lewis, B. D. Loncarevic and R. F. Macnab for reviewing the manuscript. N. J. McMillan has provided much valued comment, but he is not to be blamed for the conclusions of this paper. The Bulletin of Canadian Petroleum Geology granted permission to use Figs 1–9.

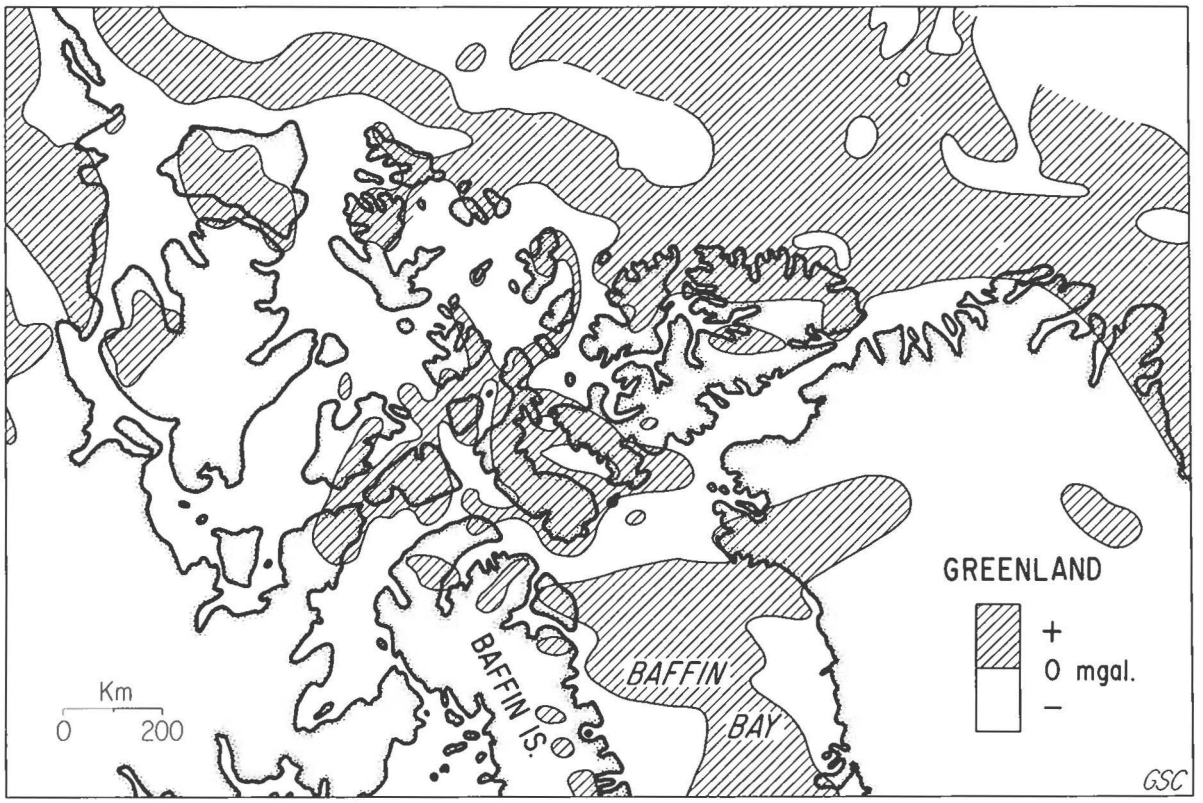


Fig. 11. Free air gravity (Observed/Predicted Field), after Sobczak (1978).

References

Amoco Canada Petroleum Company Ltd & Imperial Oil Ltd 1973. Regional geology of the Grand Banks. – Bull. Can. Petrol. Geol. 21: 479–503.

Beh, R. L. 1975. Evolution and geology of western Baffin Bay and Davis Strait, Canada. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 453–476.

Bridgwater, D., Escher, A., Jackson, G. D., Taylor, F. C. & Windley, B. F. 1973. Development of the Precambrian Shield in West Greenland, Labrador, and Baffin Island. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 99–116.

Bullard, E., Everett, J. E. & Smith, A. G. 1965. The fit of the continents around the Atlantic. – In: Blackett, P. M. S., Bullard, E. & Runcorn, S. K., A symposium on continental drift. – Phil. Trans. Roy. Soc. Lond. 258A: 41–51.

Choubert, B. 1935. Recherches sur la genèse de chaînes Paléozoiques et Antécambriennes. – Rev. Géog. Phy. Géol. Dyn. 8: 5–50.

Christie, R. L., Dawes, P. R. & Frisch, T. 1978. Basal strata of the Proterozoic Thule Group on Ellesmere Island and Greenland. – Geol. Soc. Am. Abs. with Prog. 10: 380 only.

Churkin, M. 1973. Geologic concepts of Arctic Ocean basin. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 485–499.

Coles, R. L., Haines, G. V. & Hannaford, W. 1976. Large scale magnetic anomalies over western Canada and the Arctic: a discussion. – Can. J. Earth Sci. 13: 790–802.

Cutt, B. J. & Laving, G. J. 1977. Tectonic elements and geologic history of the south Labrador and Newfoundland continental shelf, Eastern Canada. – Bull. Can. Petrol. Geol. 25: 1037–1058.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Geological Survey of Greenland 1979. Well data summary sheets, wells drilled in 1976–1977. – Geol. Surv. Greenland, Copenhagen.

Gradstein, F. M. & Srivastava, S. P. 1980. Aspects of Cenozoic stratigraphy and paleoceanography of the Labrador Sea and Baffin Bay. – Palaeogeogr., Palaeoclim., Palaeoecol. 30: 261–295.

Grant, A. C. 1980. Problems with plate tectonics: the Labrador Sea. – Bull. Can. Petrol. Geol. 28: 252–278.

Harland, W. B. 1973. Tectonic evolution of the Barents Shelf and related plates. – In: Pitcher, M. G. (ed.), Arctic geology. – Mem. Am. Ass. Petrol. Geol. 19: 599–608.

Hinz, K., Schlüter, H.-U., Grant, A. C., Srivastava, S. P., Umpleby, D. [C.] & Woodside, J. 1979. Geophysical transects of the Labrador Sea: Labrador to Southwest Greenland. – Tectonophysics 59: 151–183.

Hyndman, R. D. 1973. Evolution of the Labrador Sea. – Can. J. Earth Sci. 10: 637–644.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor spreading in central Baffin Bay. – Can. J. Earth Sci. 16: 2122–2135.

Jansa, L. F. & Wade, J. A. 1975. Geology of the continental margin off Nova Scotia and Newfoundland. – In: van der Linden, W. J. M. & Wade, J. A. (eds), Offshore geology of eastern Canada. – Pap. geol. Surv. Can. 74–30: 51–105.

Keen, C. E., Barrett, D. L., Manchester, K. S. & Ross, D. I. 1972. Geophysical studies in Baffin Bay and some tectonic implications. – Can. J. Earth Sci. 9: 239–256.

Keen, C. E., Keen, M. J., Ross, D. I. & Lack, M. 1974. Baffin Bay: small ocean basin formed by sea-floor spreading. – Bull. Am. Ass. Petrol. Geol. 58: 1089–1108.

Kerr, J. W. 1967. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Kristoffersen, Y. 1978. Sea-floor spreading and the early opening of the North Atlantic. – Earth planet. Sci. Lett. 38: 273–290.

Laughton, A. S. 1972. The southern Labrador Sea – a key to the Mesozoic and Early Tertiary evolution of the North Atlantic. – Initial Rep. Deep Sea Drilling Proj. 12: 1155–1179. – U.S. Gov. Print. Off., Washington, D.C.

Le Pichon, X., Hyndman, R. D. & Pautot, G. 1971. Geophysical study of the opening of the Labrador Sea. – J. geophys. Res. 76: 4724–4743.

Le Pichon, X., Sibuet, J.-C. & Francheteau, J. 1977. The fit of the continents around the North Atlantic Ocean. – Tectonophysics 38: 169–209.

Martin, R. 1973. Geological history of Baffin Bay – continental drift before sea-floor spreading and the exploration for hydrocarbons. – In: Hood, P. J. (ed.), Earth science symposium on offshore eastern Canada. – Pap. geol. Surv. Can. 71–23: 599–620.

McMillan, N. J. 1980. Geology of the Labrador Sea and its petroleum potential. – In: Exploration in new areas and its significance. – Proc. 10th World Petroleum Congress 2: 165–175.

Miall, A. D., Balkwill, H. R. & Hopkins, W. S. 1980. Cretaceous and Tertiary sediments of Eclipse Trough, Bylot Island area, Arctic Canada, and their regional setting. – Pap. geol. Surv. Can. 79–23: 20 pp.

Pitman, W. C. & Talwani, M. 1972. Sea-floor spreading in the North Atlantic. – Bull. geol. Soc. Am. 83: 619–646.

Sclater, J. G., Anderson, R. N. & Bell, M. L. 1971. Elevation of ridges and evolution of the central eastern Pacific. – J. geophys. Res. 76: 7888–7915.

Sobczak, L. W. 1978. Gravity from 60°N to the North Pole. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 67–73.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Talwani, M. & Eldholm, O. 1973. Boundary between continental and oceanic crust at the margin of rifted continents. – Nature, Lond. 241: 325–330.

Umpleby, D. C. 1979. Geology of the Labrador Shelf. – Pap. geol. Surv. Can. 79–13: 34 pp.

van der Linden, W. J. M. 1975. Crustal attenuation and sea-floor spreading in the Labrador Sea. – Earth planet. Sci. Lett. 27: 409–423.

van Houten, F. B. 1975. A precocious Atlantic reconstruction: an historical note. – Geology 3: 194–195.

Vine, F. J. & Matthews, D. H. 1963. Magnetic anomalies over oceanic ridges. – Nature, Lond. 199: 947–949.

The geophysical implications of minimal Tertiary motion along Nares Strait

CHARLOTTE E. KEEN and JOHN W. PEIRCE

Keen, C. E. & Peirce, J. W. 1982. The geophysical implications of minimal Tertiary motion along Nares Strait. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 327–337.

In the light of the geological evidence indicating less than 25 km of left-lateral offset along the Wegener Fault in Nares Strait, we examined three models for the formation of Baffin Bay which do not involve sea-floor spreading in that region. Our purpose was to investigate whether these models present realistic alternatives to sea-floor spreading and simultaneously remove the requirement for extensive lateral separation between Greenland and North America. These models are crustal extension, mantle upwelling and magmatic intrusion. Each model is constrained to match the observed subsidence and present width of the basin and, where possible, the observed crustal thickness. The crustal extension and magmatic intrusion models require extensive left-lateral motion between Greenland and North America, and thus they appear to be at odds with the geological observations. Furthermore, neither the mantle upwelling model nor the magmatic intrusion model can produce the required crustal thinning (85% of continental crust must be eliminated), and there are no analogous examples of such an extreme case of 'oceanization' elsewhere in the world. Thus none of the models can account for both the thin crust in central Baffin Bay and the apparent lack of offset along the Wegener Fault.

We suggest two possible solutions to the conflict between the available observations: 1) Tertiary left-lateral motion has reversed earlier right-lateral motion; or 2) left-lateral motion between Greenland and North America was accommodated elsewhere than along the Nares Strait waterway. The first is shown to be in conflict with geological evidence from the margins of Baffin Bay. There is at present no evidence to support the second, but we suggest that this possibility be critically evaluated in future studies of the geological history of the region. In this context, we suggest that minimal offset across the Nares Strait waterway need not be incompatible with major Tertiary left-lateral motion on the Wegener Fault if one accepts the concept of a fault zone rather than a single discrete offset. Future studies of the extent of horizontal deformation immediately north of Baffin Bay and within the Sverdrup Basin are also required.

A tectonic model, which includes elements of mantle upwelling and crustal extension at the margins and sea-floor spreading in central Baffin Bay, accounts for the observed subsidence and crustal thickness in the centre of the basin, and explains the uplift of eastern Baffin Island and western Greenland. It also satisfies the geometrical requirements of plate tectonic reconstructions for the North Atlantic and Arctic Oceans. However, it requires that an acceptable means of accommodating the opening of Baffin Bay be found before it can be considered valid.

C. E. Keen, Department of Energy, Mines and Resources, Geological Survey of Canada, Bedford Institute of Oceanography, Dartmouth, Nova Scotia, Canada, B2Y 4A2 and J. W. Peirce, Petro-Canada Exploration Inc., P. O. Box 2844, Calgary, Alberta, Canada, T2P 3E3.

The concept of extensive left-lateral motion along the Wegener Fault in Nares Strait during the Tertiary has been revitalized in the last decade to satisfy the need for lateral separation between Greenland and North America in regional plate reconstructions (see Kerr 1980, for a review). Recent geological evidence presented at the symposium (Dawes & Kerr, Dawes et al., Frisch & Dawes, Higgins et al., Hurst & Kerr, Peel & Christie, Peel et al., this volume) strongly favours an interpretation with no more than 25 km of net left-lateral offset along the fault. This apparent impasse prompted us to re-examine the geophysical evidence in search of a new interpretation which is compatible with minimal offset on the fault.

Models for the formation of Baffin Bay

The formation of Baffin Bay and that of Nares Strait (Fig. 1) are intimately related; any tectonic model which explains one must at least not be in conflict with the evidence from the other. Although all recent papers agree that a rifting event is involved in the formation of

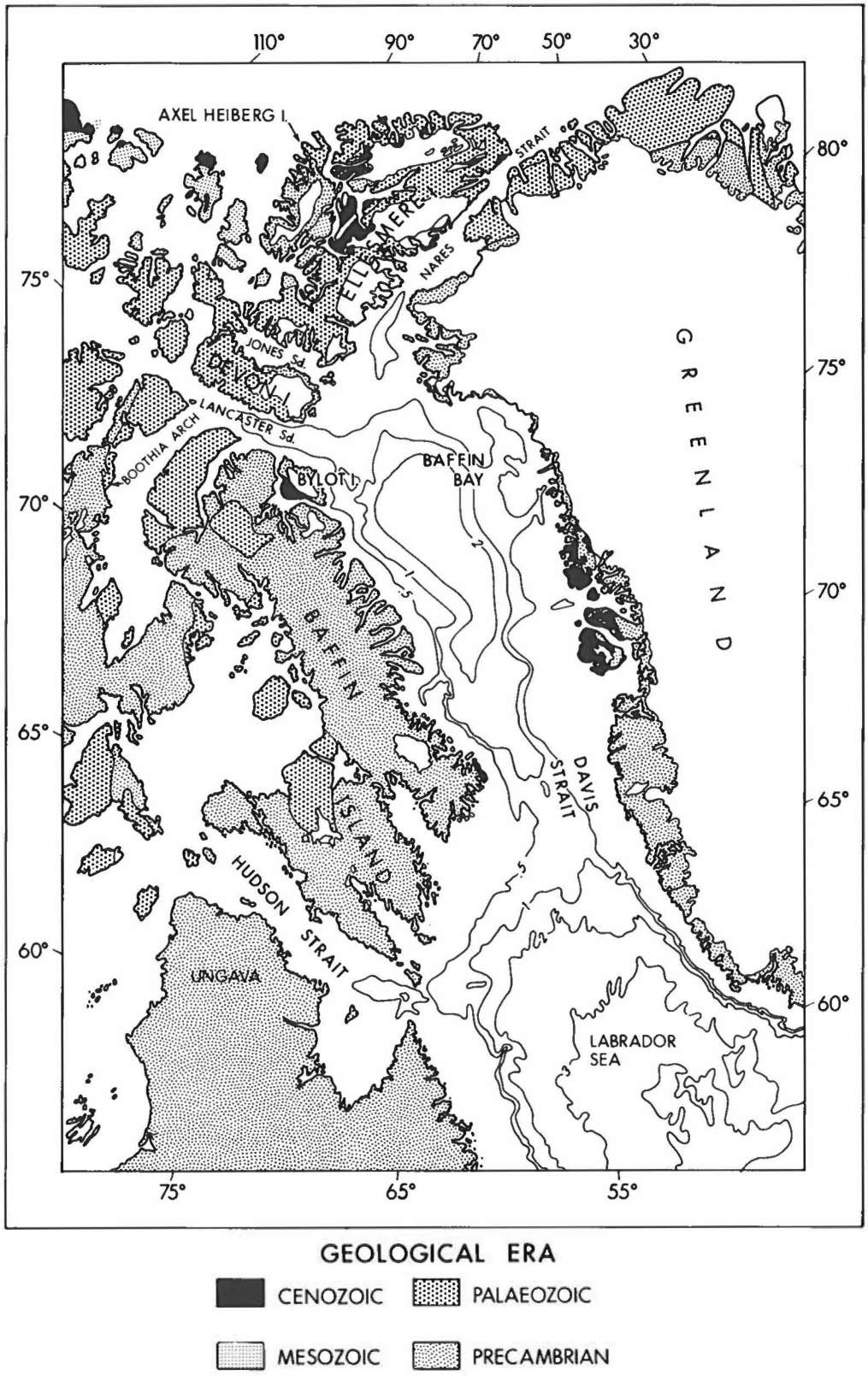


Fig. 1. Index map showing the generalized geology of the area around Baffin Bay and Nares Strait. Water depths are in kilometres.

Baffin Bay, they disagree on its importance. The conventional plate tectonic model is that central Baffin Bay is underlain by oceanic crust produced as Greenland moved away from North America (Keen et al. 1974), although a simple sea-floor spreading mechanism may not be adequate to explain the complex pattern of magnetic and gravity anomalies (Srivastava 1978, Jackson et al. 1979). On the other hand, Kerr (1981: 160) favours a model which combines a "small to moderate amount of lateral movement" with "subsidence or oceanization of a very large intervening segment of continental crust".

The crust in central Baffin Bay consists of two basic units: sediments about 4 km thick and the underlying crystalline crustal rocks not more than 5 km thick. The latter thickness is less than that of most oceanic crust except near spreading centres and in some back-arc basins (e.g. 4 km in parts of the West Philippine Basin (age 40 Ma), Louden 1980). Crustal and upper mantle velocities are 6.5–7.0 and about 8.0 km/sec respectively. Thus, the seismic characteristics of the crust in this region closely resemble those of oceanic crust (Keen & Barrett 1972).

By comparison, intra-cratonic and intra-arc sedimentary basins which are underlain by continental rocks exhibit a relatively thick crust. Some examples of crustal thickness from the best studied of these basins are given in Table 1. In all cases the crystalline crustal thickness is greater than 15 km, with the exception of the Black Sea. This latter region has been used by many workers as evidence of 'oceanization' of continental crust (e.g. Kosminskaya & Pavlenkova 1979). However, recent plate tectonic reconstructions in the Mediterranean–Alpine region suggest that the Black Sea may be the remnant of an ocean basin formed by sea-floor spreading in the Mesozoic, which has not yet been destroyed by the complex plate interactions occurring in that area (Biju-Duval et al. 1977). With this one possible exception, continental sedimentary basins appear to have crystalline crustal thicknesses which are much greater than that observed in Baffin Bay. The same is true of continental rift systems such as the Rhine Graben, the East African Rift System and the Baikal Rift (Table 1). Continental margin sedimentary basins of similar dimensions to the Baffin Bay basin also appear to have a much thicker crust (e.g. see Keen & Barrett 1981).

The total subsidence of the central region of the Bay is estimated to be the present total depth to crystalline basement, about 6 km. The total subsidence is corrected to remove the effect of sediment loading, assuming Airy isostatic compensation (Watts & Ryan 1976). The resulting subsidence of the basin is 3.7 km, or about 800 m less than the depth to 40 Ma old oceanic crust using the standard depth versus age curve for ocean basins (Sclater et al. 1971, Parsons & Sclater 1977). An age of 40 Ma has often been associated with the creation of the central part of Baffin Bay by sea-floor spreading (e.g. Keen et al. 1974, Srivastava 1978). The estimated subsidence will be in error if the crystalline crust was not initially near sea level. However, the arguments presented below will remain valid, unless the initial elevation differed by more than several hundred metres from present sea level.

We consider three numerical models, corresponding to three geological processes which can account for the observed subsidence in the basin. The three processes are: 1) uniform horizontal extension of the lithosphere (McKenzie 1978); 2) mantle upwelling or doming (non-uniform extension) (Sleep 1971, Royden & Keen 1980, Sclater et al. 1980) and 3) magmatic (dyke) intrusion of the lithosphere (Royden et al. 1980). The physical basis for the models and numerical analysis of basin subsidence are described in the above references. Each process produces heating and density changes in the

Table 1. Crystalline crustal thicknesses of several sedimentary basins and some rift systems.

Region	Age of formation	Crystalline crustal thickness-km	Reference
Pannonian	Mid-Tertiary	20	Sclater et al. (1980)
North Sea	Mesozoic	17	Sclater & Christie (1980)
Aegean	Young	20	Makris (1978)
Gulf of St. Lawrence	Palaeozoic	40	Ewing et al. (1966)
Hudson Bay	Palaeozoic	30	Ruffman & Keen (1967)
Black Sea	? Mesozoic	5–8	Neprochnov et al. (1970)
Sverdrup Basin	Palaeosoic–Mesozoic	20–30	Forsyth et al. (1979)
Afar region of Ethiopia	Young	20	Berckhemer et al. (1975)
Rhine Graben	Young	25	Edel et al. (1975)
E. African Rift (Gregory Rift)	Young	18	Long et al. (1973)
Baikal Rift	Young	30	Kosminskaya & Pavlenkova (1979)

Note: Ages are approximate. Many of these features have experienced several episodes of renewed subsidence, extension or volcanism. Crustal thicknesses obtained by removing best estimates of sediment thickness.

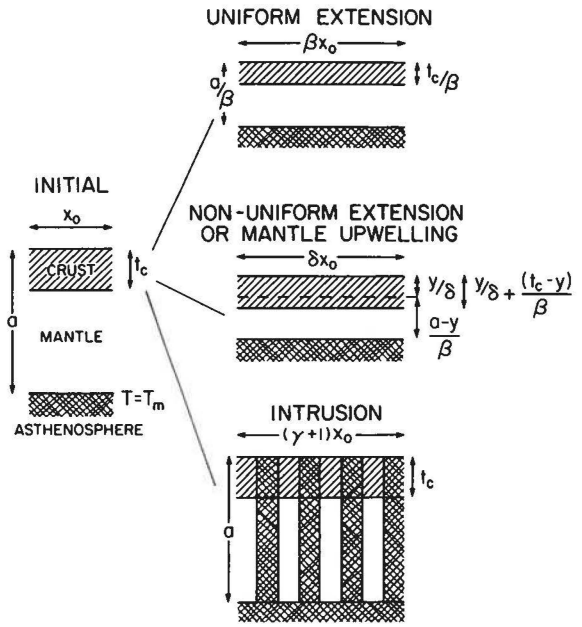


Fig. 2. Schematic representation of the three models considered to describe the formation of Baffin Bay. The initial configuration of the lithosphere is shown on the left. Initial crustal width is x_o, initial crustal thickness is t_c, initial thickness of the thermal lithosphere is as defined by the position of the melting temperature isotherm ($T = T_m$). For our models x_o is unknown, t_c = 35 km, a = 125 km, T_m = 1350°C.

In the uniform extension model (top), the entire lithosphere is stretched by a factor of β and thinned by a factor of $1/\beta$. As the lithosphere is thinned, hot asthenosphere upwells passively from below. After stretching, the crustal width is βx_o and crustal thickness is t_c/β.

In the mantle upwelling model (middle), the shallow lithosphere above a depth y is stretched by a factor δ and thinned by $1/\delta$. The deeper lithosphere below y is thinned by a factor of $1/\beta$. After stretching the horizontal extent of the region is δx_o. If $\delta = \beta$, this model is equivalent to uniform extension.

The intrusion model (bottom) involves the intrusion of dykes of asthenospheric material. The intrusion parameter, γ, is the fraction of dykes replacing original lithosphere. Assuming that the lithosphere extends to accommodate the additional material, no thinning occurs and the horizontal extent of the region is $(1 + \gamma)x_o$.

lithosphere which require uplift or subsidence to maintain isostatic equilibrium. When the tectonic processes become dormant, cooling and subsidence occur. Each numerical model is constrained to match the observed subsidence (3.7 km), the present basin width (220 km), and, where possible, the observed crustal thickness (5 km).

The models are shown schematically in Fig. 2. The β parameter describes the amount of extension in the uniform extension model. Its inverse is the crustal thinning factor. In the mantle upwelling model, δ describes the amount of extension and thinning down to depth y, and β defines the amount of thinning below depth y. The γ parameter in the intrusion model describes the volume fraction of dykes replacing crust (and therefore the amount of extension), but this model does not predict thinned crust. The limiting case in all three models is the total destruction of continental lithosphere, which is equivalent to sea-floor spreading. This extreme is represented numerically by β = infinity in the uniform extension model, $\delta = \beta$ = infinity in the upwelling model, and $\gamma = 1$ in the intrusion model.

The model parameters, observed and computed subsidence, and approximate amount of horizontal separation required by each model are listed in Table 2. It has been assumed that tectonic processes ceased and cooling began about 35 to 40 Ma ago in central Baffin Bay. All values of subsidence are given for a water-filled basin; the observed subsidence has been corrected for sediment loading assuming Airy, pointwise isostatic compensation (Watts & Ryan 1976). The original thickness of continental crust is assumed to be 35 km. The physical constants used in the calculations are the same as those used by Royden & Keen (1980).

The evolution of many rifted continental margins can be described by a combination of these three processes (Royden & Keen 1980, Royden et al. 1980). Extension and thinning of the lithosphere appear to be the primary processes during the rift phase of the development of continental margins (Montadert et al. 1979, Keen & Barrett 1981). The observational data can be satisfied by extension and mantle upwelling which increases towards the ocean–continent boundary, at which point total destruction of the continental lithosphere occurs and sea-floor spreading begins. Magmatic intrusion undoubtedly plays a role in the formation of rifted continental margins, but the importance of this is not well understood.

Uniform extension model

This model produces the Baffin Bay basin by extension and thinning of the lithosphere and the crust. It can satisfy the crustal thickness constraint and predict the observed subsidence of the basin if extension by a factor of 7 (crustal thinning to 5 km) occurred prior to 35 Ma (Table 2).

The model has been used very successfully in describing the geological history of intra-cratonic, intra-arc and continental margin sedimentary basins (Royden & Keen 1980, Sclater & Christie 1980, Sclater et al. 1980). However, these basins require much less extension and thinning ($\beta \leq 3$) than Baffin Bay. The horizontal separation predicted for Baffin Bay by this model is about 200 km, almost as much as that required in conventional plate tectonic models. Also it may not be physically or chemically reasonable to postulate the large extension factor required for Baffin Bay without consideration of either significant volcanism (perhaps Davis Strait is an example) or the triggering of true sea-floor spreading.

Table 2. Results of model calculations.

Model	Parameters				Predicted present uplift[1] or subsidence (km)	Observed present uplift[1] or subsidence (km)	Predicted lateral motion (km)
						Central Basin (cooling starts 35 Ma ago)	
	δ	β	γ	y(km)			
Extension	7	7	0	–	3.51	3.74	214
Mantle upwelling	1	100	0	5.0	3.47	3.74	small
Dyke intrusion	1	1	0.8	–	3.45	3.74	111
						Continental Shelves (cooling starts 60 Ma ago)	
Extension and upwelling	2	7	0	35.0	1.99	1.74	112
						Baffin Island (cooling starts 60 Ma ago)	
Mantle upwelling	1	4	0	35.0	–0.49	~–0.8[2]	

1. Negative sign indicates uplift.
2. Observed uplift of eastern Baffin Island relative to 200 m elevation of western Baffin Island. Figures for West Greenland are not as representative due to the large ice sheet over most of the island.

Mantle upwelling or non-uniform extension model

In this model, upwelling occurs when light, fluid asthenosphere rises diapirically, replacing the denser lithosphere. The resulting thermal and subsidence histories will be the same as those for the non-uniform extension model of Royden & Keen (1980). In order to generate the thin crust, hot material must penetrate the upper crust, assimilating or replacing the original crustal material. Mantle upwelling can produce the correct basin subsidence and it may result in a relatively small amount of lateral separation. However, neither this model nor the intrusion model described below, account for the mechanism by which a large amount of the continental crust is destroyed, leaving the observed thin crustal layer. Possible 'oceanization' models are discussed later.

Magmatic intrusion model

This model requires that γ volume of dykes be intruded into the lithosphere in some randomly distributed manner which is not concentrated at a definable spreading axis. These dykes are assumed to be composed of basic and ultrabasic material from the asthenosphere. In its simplest form, this process does not include phase changes or the segregation of basaltic melt from the asthenosphere (but see petrological implications of such complications discussed below).

While intrusion can explain the observed subsidence of the basin, provided that a very large percentage (80%) of the original lithosphere is replaced by intrusive material, it does not predict the thin crust. Either extension of the lithosphere must occur to accommodate the volume of dykes intruded, giving 110 km of horizontal separation, or the lithosphere must thicken. In neither case is thin crust expected. Furthermore, such processes should lead to a rather heterogeneous crustal structure, perhaps without a well-defined crust–mantle boundary.

Crustal thinning by 'oceanization'

The above results suggest that upwelling might account for the formation of Baffin Bay and not require much lateral separation. However, a plausible means for destroying 85% of the continental crust must be found. There are various ways in which 'oceanization' or crustal thinning might occur apart from extension, many of which have been reviewed by van Bemmelen (1972). Unfortunately, there is little direct evidence supporting any of these hypotheses collectively known as 'oceanization' models. Modes of 'oceanization' can be broadly divided into two groups: those involving sub-crustal erosion and those requiring phase transformations. Both require the segregation of basaltic magma from hot mantle material and its intrusion into the crust. Sub-crustal erosion presumes that the intrusion causes melting of the crustal material followed by its eventual removal and assimilation into the mantle by downward convection. Alternatively, the gabbro-eclogite phase change has been postulated to transform the basalt saturated crustal rocks into eclogite upon cooling.

There are several difficulties with these 'oceanization' models. Both assume that large volumes of basalt are available to effect the necessary chemical and physical changes. As Jarvis & McKenzie (1980) have noted, it is, in general, unlikely that more basaltic magma will be available than that generated at mid-ocean ridges where about 5–6 km of basaltic crust is created (see also Ahern & Turcotte 1979). Modern examples of regions where sufficiently large quantities of basaltic magma are being generated, such as Iceland and the Afar region of Ethiopia, do not exhibit thin crust (Table 1), but appear to consist of about 20 km of basaltic crust overlying the mantle. The phase change hypothesis further suffers from the difficulties of generating a well-defined

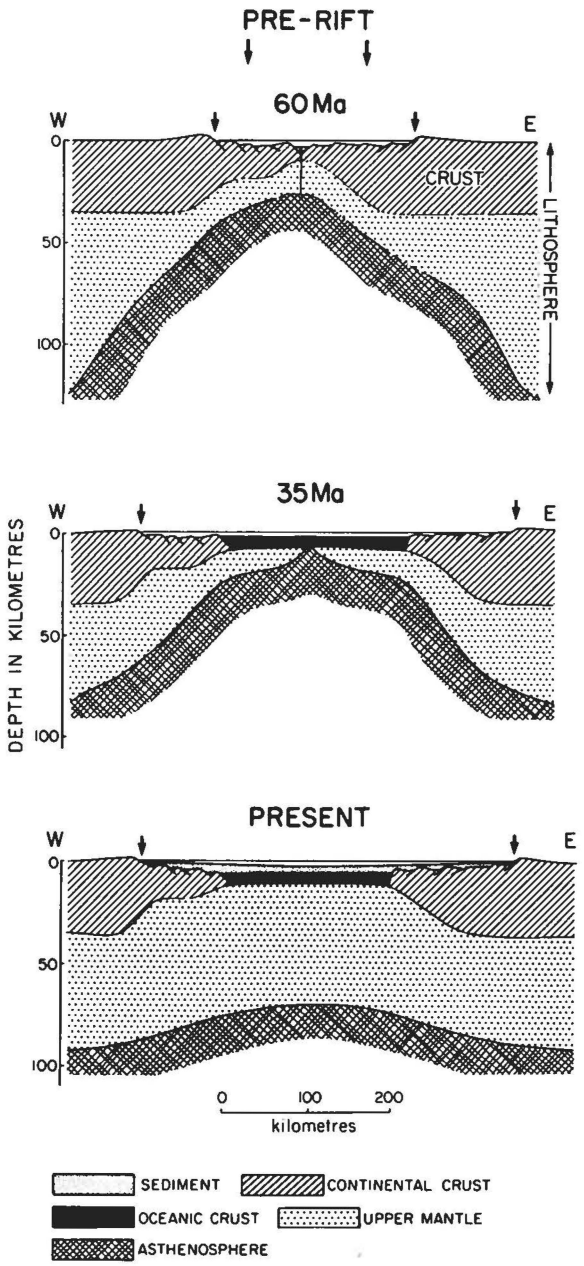


Fig. 3. Schematic representation of three stages in the evolution of Baffin Bay according to our preferred model. Prior to 60 Ma continental rifting between Greenland and Baffin Island occurs. This process may have begun in the Early Cretaceous. The thicknesses of the crust and lithosphere are based on the stretching and thinning parameters given for the continental shelves in Table 2. In particular, note the marginal uplift and initial subsidence of the continental shelves. Sea-floor spreading began about 60 Ma and ceased about 35 Ma ago when Greenland became part of the North American plate. The arrows indicate the amount of horizontal separation attributed to sea-floor spreading and rifting. Since 35 Ma the entire region has been cooling and subsiding, and the lithosphere has been thickening. The total extension of the region is obtained by comparing the separation of the pre-rift arrows with the arrows for 35 Ma.

crust–mantle boundary which is required by the Baffin Bay seismic data, and from the lack of experimental evidence for its occurrence at the temperatures and pressures (Wyllie 1971) which are likely to occur at the shallow mantle depths in Baffin Bay.

Discussion

While the controversy concerning the feasibility of 'oceanization' will undoubtedly continue, it appears to be an unsatisfactory solution to the mode of crustal thinning in central Baffin Bay, where the crust is thin compared to normal oceanic crust. While the three models discussed above can satisfy the basin subsidence, all but the mantle upwelling model require significant lateral motion between Greenland and North America, and therefore do not remove the dilemma created in attempting to reconcile the formation of Baffin Bay with the geological evidence for minimal lateral displacement in the Nares Strait region. The mantle upwelling model suffers from the absence of a realistic mechanism for thinning the crust by 85% from below. While the numerical models we have used are admittedly oversimplified, they do provide reasonable estimates of the subsidence history of the basin and they describe some of the possible alternatives to sea-floor spreading in Baffin Bay. In the absence of a satisfactory oceanization model, we feel compelled to reject these alternative means of forming Baffin Bay and to accept the fact that sea-floor spreading did occur beneath the deep, central basin.

Our preferred model for the formation of Baffin Bay is outlined in Fig. 3. We have assumed that plate tectonic motions and sea-floor spreading occurred between 60 and 35 Ma ago (e.g. Srivastava 1978) and that this was preceded by continental rifting. The rift phase may have begun in the Early Cretaceous as discussed by McWhae (1981) and Peirce (this volume), and probably ended in the Early Paleocene. During this time the present margins of the Bay underwent extension and mantle upwelling. The amount of extension and thinning was most intense near the present ocean–continent boundary and decreased landward. It produced the observed crustal thinning ($\delta = 2$) beneath the shelf of Baffin Island (Jackson et al. 1977, Keen & Hyndman 1979), but little or no crustal change occurred beneath the flanking mainland region. The latter, however, experienced uplift due to the greater lateral extent of mantle upwelling. This uplift is necessary to explain the present uplift of eastern Baffin Island and West Greenland. Sea-floor spreading began about 60 Ma ago, coincident with a major change in spreading direction in the Labrador Sea (Srivastava 1978). The thermal anomaly beneath the adjacent shelf and mainland regions probably began to decay at that time. Given the complicated pattern of the magnetic anomalies in Baffin Bay

(Srivastava 1978, Jackson et al. 1979), the sea-floor spreading mechanism may be somewhat diffuse and is perhaps more similar to that envisioned in back-arc basins (Karig 1971, Isezaki & Uyeda 1973, Isezaki 1975) than to that at most mid-ocean ridge spreading centres. Sea-floor spreading, diffuse or otherwise, formed the deep central part of Baffin Bay and ended about 35 Ma ago. From that time to the present, the entire region has been subsiding.

Geological evidence relating to the timing of the proposed uplift of the mainland regions is sparse and cannot be interpreted unambiguously. In the Davis Strait region, Early Paleocene subaqueous volcanics are now elevated 600 m above sea level, suggesting that uplift occurred after emplacement of these rocks about 57 Ma ago (Parrott & Reynolds 1975). In the Eclipse Trough on Bylot Island, Tertiary (Paleocene?) sediments were deposited in a near-shore marine environment and are now 600 m above sea level (Miall et al. 1980). However, it has proved difficult to establish a reliable age for these sediments. These observations perhaps suggest that the entire coastline of eastern Baffin Island experienced uplift during or since Early Paleocene time. MacLean & Falconer (1979) suggest that the uplift occurred since the Eocene; this would be difficult to reconcile with the timing of tectonic events as proposed here. However, uplift in the Early Paleocene is also consistent with the geological evidence and with the model presented above.

The present amounts of uplift and subsidence predicted for the mainland and shelf regions are given in Table 2. The mainland exhibits a residual amount of its initial uplift. The predicted (500 m) and observed (600 to 800 m) present uplift of eastern Baffin Island agree reasonably well, and this model provides an explanation of the observed topography in that region. Similar uplift is predicted and observed on the Greenland side. Isostatic effects due to glacial rebound are unlikely to exceed 100 m (G. Quinlan, pers. comm. 1980) and we have ignored this relatively short term effect on the subsidence history of the region. Unlike the mainland, the shelf has subsided, and the observed and predicted subsidence are in good agreement.

This composite model provides an excellent explanation of the observed features of the Baffin Bay region. However, the model requires that significant lateral motion (350 km) is produced by crustal extension of the shelves and sea-floor spreading. This must be accommodated to the north of Baffin Bay, primarily during the Tertiary. While some of this extension (perhaps 100 km) can be attributed to graben structures such as Lancaster and Jones Sounds and to the relative motion of Baffin Island and North America, about 250 km of lateral motion is suggested between the Greenland and North American plates.

Difficulties of present pre-drift reconstructions and suggestions for future studies

In a North Atlantic context a probable pre-drift position for Greenland relative to North America and Europe can be derived by closing up the Mesozoic and Cenozoic ocean basins. The geophysical evidence (Bullard et al. 1965, Kristoffersen & Talwani 1977, Sclater et al. 1977, Srivastava 1978, Srivastava & Falconer, this volume) clearly indicates that Greenland moved as a separate plate during the Late Cretaceous and Early Tertiary. We have assumed this to be true in the reconstructions shown in Fig. 4. There is, however, no direct constraint on the position of Ellesmere Island relative to Greenland from these pre-drift reconstructions because the amount of horizontal deformation within the Sverdrup Basin is not well understood. Among the many pre-drift positions for Ellesmere Island which have been proposed (reviewed by Kerr 1980), we have chosen to depict Ellesmere Island against Greenland. In contrast, Srivastava (1978) and others showed a significant gap between them and the implication is that the material in this gap must be destroyed during the opening of the Labrador Sea. As there is no geological evidence to support a subduction zone but there is some evidence for compression (but not enough) in the Sverdrup Basin at the correct time, Peirce (this volume) has argued that the Sverdrup Basin was much wider during the early Mesozoic. If Ellesmere Island and Greenland are juxtaposed, and furthermore if one assumes that significant left-lateral motion occurred on the Wegener Fault in the Tertiary, the pre-drift reconstructions shown in Figs 4a and 4b result. Fig. 4a was obtained using the Bullard et al. (1965) position for Greenland relative to North America, while 4b was derived using Srivastava's (1978) Labrador Sea data to define these positions. If one chooses to allow only minor left-lateral displacement on the Wegener Fault in the Tertiary, then one must choose the Bullard et al. (1965) position for Greenland to avoid overlap of Ellesmere and Devon Islands, as shown in Fig. 4c.

These reconstructions illustrate the difficulties in satisfying both the requirement for minimal offset across Nares Strait and the evidence for some form of sea-floor spreading in Baffin Bay. Fig. 4c demonstrates that it is possible to imagine an initial position of Ellesmere Island relative to Greenland which satisfies the requirement of minimal offset along Nares Strait without overlapping continental blocks. However, if we require that the deep central region of Baffin Bay be closed before sea-floor spreading began there about 60 Ma ago, pre-Tertiary right-lateral motion along Nares Strait is required. This motion could have been reversed by subsequent left-lateral Tertiary motion, resulting in little net offset along the Strait. We do not favour this argument. The closure of Baffin Bay should have

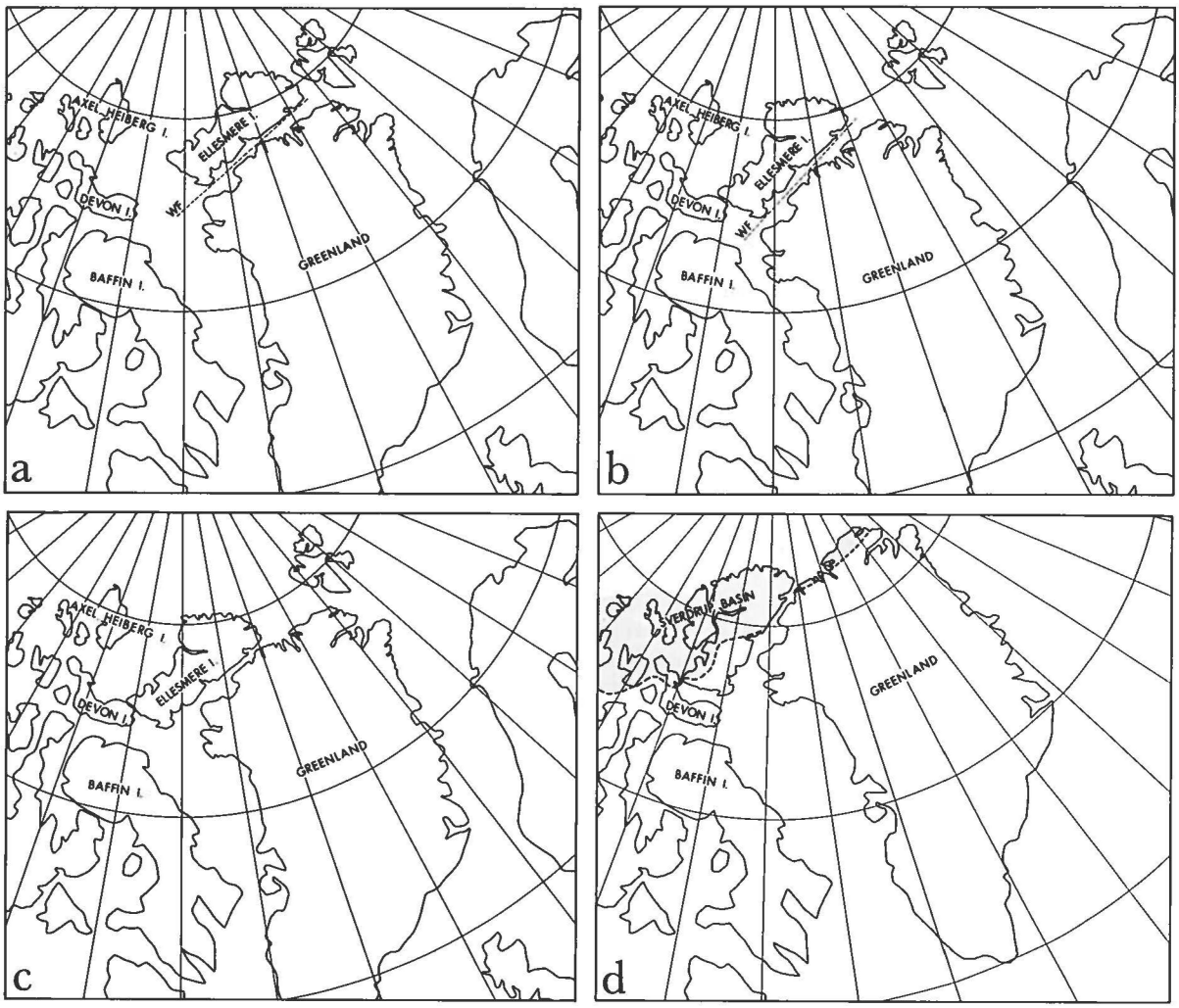


Fig. 4. a. Initial position of Ellesmere Island relative to Greenland assuming the fit of Bullard et al. (1965) for the North Atlantic with Ellesmere Island closely juxtaposed to Greenland. Greenland has been rotated first to the anomaly 25 position of Srivastava (1978), and then Ellesmere Island and Greenland were rotated together to the Bullard et al. (1965) position for Greenland relative to North America. Baffin Island has been rotated 3.8° towards North America about a pole at 70°N, 90°W (Peirce, this volume).

In Fig. 4a,b and c Axel Heiberg Island has been left in its original position and Ellesmere Island has been rotated in its present form to indicate the degree of compression in the Sverdrup Basin during the Eurekan orogeny. The maps are Lambert equal area projections centred at 80°N, 70°W. WF = Wegener Fault.

b. Initial position of Ellesmere Island relative to Greenland using the same rotations as in Fig. 4a except Greenland has been positioned relative to North America according to the anomaly 33 position of Srivastava (1978). Note the difference in the width of Baffin Bay as compared to 4a.

c. Initial position of Ellesmere Island relative to Greenland based on the geological correlations and allowing 25 km of left-lateral motion on the Wegener Fault. Ellesmere Island has been moved 25 km to the northeast and then positioned with Greenland according to the Bullard et al. (1965) fit. This initial fit appears to reconcile the geological evidence with the geophysical evidence, but carrying this picture forward in time requires major right-lateral motion on the Wegener Fault and compression in Baffin Bay. There is no evidence to support the former and some evidence in Baffin Bay indicates extension since the Early Cretaceous.

d. Present geography shown for reference. The present outline of the Palaeozoic and Tertiary fold belts, including the Sverdrup Basin of Canada, is indicated.

caused compression along its margins, and there is no evidence to support this. Indeed, the evidence for graben formation on Bylot Island in the Early Cretaceous (Miall et al. 1980, McWhae 1981) suggests that extension, not compression began at least that early in the region. Similar difficulties apply to the reconstruction of Fig. 4a; Fig. 4b shows Baffin Bay closed, but the 250 km offset along Nares Strait is at odds with most interpretations of the geological data in that region.

We are thus still left with an impasse. We have shown

in this paper that alternative geological models which explain the geological and geophysical data in Baffin Bay without sea-floor spreading are either untenable or require almost as much lateral motion between Greenland and North America as the conventional plate tectonic models. Others have shown that the geological data and their correlation across the Nares Strait waterway require less than 25 km of offset across the Strait. Too often the 'solution' to this impasse has been to disregard either one or the other of these apparently conflicting, but well documented, studies. This is no longer acceptable and one must search elsewhere for solutions to this impasse in future studies. The main question which must be addressed concerns the nature and extent of deformation of the continental lithosphere north of Baffin Bay during rifting and spreading to the south.

One possibility is that the Wegener Fault is a fault zone which behaved in a complex manner, rather than a discrete feature as exemplified by the Nares Strait waterway. Several sub-parallel faults may have accommodated the motion at various times, and within this fault zone there may be several fault blocks, some of which may have been rotated. Such a multiplicity of faults occurs along parts of the San Andreas fault system and has been recently observed on the Queen Charlotte Fault (Riddihough 1980). At least one minor strike-slip fault has been observed in eastern Ellesmere Island (Mayr & de Vries, this volume) which could be interpreted as being part of a wide fault zone. Additionally there may be a zone of plastic deformation on either side of the major faults. This type of deformation has been documented near the Alpine Fault in New Zealand (Walcott 1979).

Furthermore, perhaps the amount and nature of the deformation of the continental lithosphere immediately north of Baffin Bay and within the Sverdrup Basin have been underestimated. Good measurements of crustal and sedimentary thicknesses on the shelf and in the sounds north of Baffin Bay would be helpful in estimating the amount of extension in these regions. Quantitative measures are also needed of the net amount of lateral motion in the Sverdrup Basin since the early Mesozoic. One must also ask if the deformation expressed in the surface rocks is a realistic estimate of the deformation at depth, involving the whole lithosphere.

While it is premature to speculate further on alternative interpretations until more evidence is available, we suggest that these and other alternative perspectives need further consideration.

Conclusions

We have considered three possible models to explain the observed subsidence, width, and crustal thickness in Baffin Bay without appealing to a sea-floor spreading origin. The uniform extension model requires nearly as much lateral motion between Greenland and North America as does a sea-floor spreading model, and there are no analogous basins elsewhere in the world which exhibit the high degree of extension required in Baffin Bay. Neither the magmatic intrusion model (which also requires lateral motion) nor the mantle upwelling model can adequately explain the thin crust observed in Baffin Bay. Therefore none of these three geological models offers a realistic alternative to sea-floor spreading as a means of forming the Baffin Bay basin.

Our preferred explanation of the geophysical evidence from Baffin Bay requires a sea-floor spreading origin for the central basin and moderate extension of each continental margin. Total extension of roughly 350 km must be accommodated by motion north of Baffin Bay. It is simplest to propose that most of this motion took place in Nares Strait but this is at odds with the geological data in that region. Geophysically it is possible that this phase of major left-lateral motion has reversed an earlier phase of right-lateral motion, but this seems unreasonable geologically. We suggest that future work might consider the possibility that the Wegener Fault is a fault zone with complex local tectonics rather than a localized, single offset. Also, estimates of the amount of lateral motion which could be accommodated by extension of the continental lithosphere immediately north of the Baffin Bay basin and by compression within the Sverdrup Basin are critical to solving the geological history of this region.

Thus we favour a tectonic history similar to that discussed by Peirce (this volume) for the eastern Arctic. Such a history requires a sea-floor spreading origin for Baffin Bay and the accommodation of significant lateral motion to the north of the basin. We cannot agree that Baffin Bay was formed by extension, subsidence and modification of continental crust. The Nares Strait dilemma cannot be solved in this manner because large, probably unrealistic amounts of crustal thinning are required. We recognize that an interpretation of the geology around Nares Strait which favours minimal Tertiary motion is compelling. We are not able to offer an alternative interpretation; we only suggest different perspectives which may be promising. The challenge remains to develop fully a unified model which is consistent both with the geophysical evidence in Baffin Bay and the North Atlantic *and* with the geological evidence in Nares Strait.

Acknowledgements

We thank A. C. Grant, B. MacLean, J. R. H. McWhae and S. P. Srivastava for critically reading the manuscript.

References

Ahern, J. L. & Turcotte, D. L. 1979. Magma migration beneath an ocean ridge. – Earth planet. Sci. Lett. 45: 115–122.

Berckhemer, H., Baier, B., Bartelsen, H., Behle, A., Burkhardt, H., Gebrande, H., Makris, J., Menzel, H., Miller, H. & Vees, R. 1975. Deep seismic soundings in the Afar region and on the highland of Ethiopia. – In: Pilger, A. & Rösler, A. (eds), Afar Depression of Ethiopia: 89–107. – Schweizerbart, Stuttgart.

Biju-Duval, B., Dercourt, J. & Le Pichon, X. 1977. From the Tethys ocean to the Mediterranean seas: a plate tectonic model of the evolution of the western Alpine system. – In: Biju-Duval, B. & Montadert, L. (eds), Structural history of the Mediterranean basins: 143–164. – Éditions Technip, Paris.

Bullard, E., Everett, J. E. & Smith, A. G. 1965. The fit of the continents around the Atlantic. – In: Blackett, P. M. S., Bullard, E. & Runcorn, S. K., A symposium on continental drift. – Phil. Trans. Roy. Soc. Lond. 258A: 41–51.

Dawes, P. R. & Kerr, J. W. 1982. The case against major displacement along Nares Strait. – This volume.

Dawes, P. R., Frisch, T. & Christie, R. L. 1982. The Proterozoic Thule Basin of Greenland and Ellesmere Island: importance to the Nares Strait debate. – This volume.

Edel, J. B., Fuchs, K., Gelbke, C. & Prodehl, C. 1975. Deep structure of the southern Rhinegraben area from seismic refraction investigation. – J. Geophys. 41: 333–356.

Ewing, G. N., Dainty, A. M., Blanchard, J. E. & Keen, M. J. 1966. Seismic studies on the eastern seaboard of Canada: the Appalachian system. 1. – Can. J. Earth Sci. 3: 89–109.

Forsyth, D. A., Mair, J. A. & Fraser, I. 1979. Crustal structure of the central Sverdrup Basin. – Can. J. Earth Sci. 16: 1581–1598.

Frisch, T. & Dawes, P. R. 1982. The Precambrian Shield of northernmost Baffin Bay: correlation across Nares Strait. – This volume.

Higgins, A. K., Mayr, U. & Soper, N. J. 1982. Fold belts and metamorphic zones of northern Ellesmere Island and North Greenland. – This volume.

Hurst, J. M. & Kerr, J. W. 1982. Upper Ordovician to Silurian facies patterns in eastern Ellesmere Island and western North Greenland and their bearing on the Nares Strait lineament. – This volume.

Isezaki, N. 1975. Possible spreading centers in the Japan Sea. – J. Mar. geophys. Res. 2: 265–277.

Isezaki, N. & Uyeda, S. 1973. Geomagnetic anomaly of the Japan Sea. – J. Mar. geophys. Res. 2: 51–59.

Jackson, H. R., Keen, C. E. & Barrett, D. L. 1977. Geophysical studies on the eastern continental margin of Baffin Bay and in Lancaster Sound. – Can. J. Earth Sci. 14: 1991–2001.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor spreading in central Baffin Bay. – Can. J. Earth Sci. 16: 2122–2135.

Jarvis, G. T. & McKenzie, D. P. 1980. Sedimentary basin formation with finite extension rates. – Earth planet. Sci. Lett. 48: 42–52.

Karig, D. E. 1971. Origin and development of marginal basins in the western Pacific. – J. geophys. Res. 76: 2542–2560.

Keen, C. E. & Barrett, D. L. 1972. Seismic refraction studies in Baffin Bay: an example of a developing ocean basin. – Geophys. J. Roy. astr. Soc. 30: 253–271.

Keen, C. E. & Barrett, D. L. 1981. Thinned and subsided continental crust on the rifted margin of eastern Canada: crustal structure, thermal evolution, and subsidence history. – Geophys. J. Roy. astr. Soc.: 65: 443–465.

Keen, C. E. & Hyndman, R. D. 1979. Geophysical review of the continental margins of eastern and western Canada. – Can. J. Earth Sci. 16: 712–747.

Keen, C. E., Keen, M. J., Ross, D. I. & Lack, M. 1974. Baffin Bay: small ocean basin formed by sea-floor spreading. – Bull. Am. Ass. Petrol. Geol. 58: 1089–1108.

Kerr, J. W. 1980. Did Greenland drift along Nares Strait? – Bull. Can. Petrol. Geol. 28: 279–283.

Kerr, J. W. 1981. Evolution of the Canadian Arctic Islands: a transition between the Atlantic and Arctic Oceans. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 105–199. – Plenum Press, New York & London.

Kosminskaya, I. P. & Pavlenkova, N. I. 1979. Seismic models of inner parts of the Euro-Asian continent and its margins. – Tectonophysics 59: 307–320.

Kristoffersen, Y. & Talwani, M. 1977. Extinct triple junction south of Greenland and the Tertiary motion of Greenland relative to North America. – Bull. geol. Soc. Am. 88: 1037–1049.

Long, R. E., Sundaralingam, K. & Maguire, P. K. H. 1973. Crustal structure of the East African rift zone. – Tectonophysics 20: 269–281.

Louden, K. E. 1980. The crustal and lithosphere thicknesses of the Philippine Sea as compared to the Pacific. – Earth planet. Sci. Lett. 50: 275–288.

MacLean, B. & Falconer, R. K. H. 1979. Geological/geophysical studies in Baffin Bay and Scott Inlet-Buchan Gulf and Cape Dyer-Cumberland Sound areas of the Baffin Island Shelf. – Pap. geol. Surv. Can. 79–1B: 231–244.

Makris, J. 1978. The crust and upper mantle of the Aegean region from deep seismic soundings. – Tectonophysics 46: 269–284.

Mayr, U. & de Vries, C. D. S. 1982. Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago. – This volume.

McKenzie, D. P. 1978. Some remarks on the development of sedimentary basins. – Earth planet. Sci. Lett. 40: 25–32.

McWhae, J. R. H. 1981. Structure and spreading history of the northwestern Atlantic region from the Scotian Shelf to Baffin Bay. – In: Kerr, J. W. & Fergusson, A. J. (eds), Geology of the North Atlantic borderlands. – Mem. Can. Soc. Petrol. Geol. 7: 299–332.

Miall, A. D., Balkwill, H. R. & Hopkins, W. S. 1980. Cretaceous and Tertiary sediments of Eclipse Trough, Bylot Island area, Arctic Canada, and their regional setting. – Pap. geol. Surv. Can. 79–23: 20 pp.

Montadert, L., de Charpal, O., Roberts, D., Guennoc, P. & Sibuet, J.-C. 1979. Northeast Atlantic passive continental margins: rifting and subsidence processes. – In: Talwani, M., Hay, W. & Ryan, W. B. F. (eds), Deep drilling results in the Atlantic Ocean: continental margins and paleoenvironment: 154–186. – Am. geophys. Un., Washington, D. C.

Neprochnov, Yu. P., Kosminskaya, I. P. & Malovitsky, Ya. P. 1970. Structure of the crust and upper mantle of the Black and Caspian Seas. – Tectonophysics 10: 517–538.

Parrott, R. J. E. & Reynolds, P. H. 1975. Argon-40/Argon-39 geochronology: age determinations of basalts from the Labrador Sea area. – Geol. Soc. Am. Abs. with Prog. 7: 835 only.

Parsons, B. & Sclater, J. G. 1977. An analysis of the variation of ocean floor bathymetry and heat flow with age. – J. geophys. Res. 82: 802–827.

Peel, J. S. & Christie, R. L. 1982. Cambrian–Ordovician platform stratigraphy: correlations around Kane Basin. – This volume.

Peel, J. S., Dawes, P. R., Collinson, J. D. & Christie, R. L. 1982. Proterozoic – basal Cambrian stratigraphy across Nares Strait: correlation between Inglefield Land and Bache Peninsula. – This volume.

Peirce, J. W. 1982. The evolution of the Nares Strait lineament and its relation to the Eurekan orogeny. – This volume.

Riddihough, R. P. 1980. Recent plate tectonics along Canada's west coast. – Geol. Ass. Can., Min. Ass. Can. Prog. with Abs. 5: 78 only.

Royden, L. & Keen, C. E. 1980. Rifting process and thermal evolution of the continental margin of eastern Canada determined from subsidence curves. – Earth planet. Sci. Lett. 51: 343–361.

Royden, L., Sclater, J. G. & von Herzen, R. P. 1980. Continental margin subsidence and heat flow: important parameters in formation of petroleum hydrocarbons. – Bull. Am. Ass. Petrol. Geol. 64: 173–187.

Ruffman, A. [S.] & Keen, M. J. 1967. A time-term analysis of the first arrival data from the seismic experiment in Hudson Bay, 1965. – Can. J. Earth Sci. 4: 901–928.

Sclater, J. G. & Christie, P. A. F. 1980. Continental stretching: an explanation of the post-mid-Cretaceous subsidence of the central North Sea Basin. – J. geophys. Res. 85: 3711–3739.

Sclater, J. G., Anderson, R. N. & Bell, M. L. 1971. The elevation of ridges and the evolution of the central eastern Pacific. – J. geophys. Res. 76: 7888–7915.

Sclater, J. G., Hellinger, S. & Tapscott, C. 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. – J. Geol. 85: 509–552.

Sclater, J. G., Royden, L., Horvath, F., Burchfiel, B. C., Semken, S. & Stegena, L. 1980. Subsidence and thermal evolution of the intra-Carpathian Basins. – Earth planet. Sci. Lett. 51: 139–162.

Sleep, N. H. 1971. Thermal effects of the formation of Atlantic continental margins by continental break-up. – Geophys. J. Roy. astr. Soc. 24: 325–350.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Srivastava, S. P. & Falconer, R. K. H. 1981. Nares Strait: a conflict between plate tectonic predictions and geological interpretation. – This volume.

van Bemmelen, R. W. 1972. Geodynamic models. An evaluation and synthesis. – Developments in geotectonics 2. – Elsevier Publ. Co., Amsterdam: 267 pp.

Walcott, R. I. 1979. Plate motion and shear strain rates in the vicinity of the southern Alps. – In: Walcott, R. I. & Cressell, M. M. (eds), The origin of the southern Alps. – Bull. Roy. Soc. N. Z. 18: 5–12.

Watts, A. B. & Ryan, W. B. F. 1976. Flexure of the lithosphere and continental margin basins. – Tectonophysics 36: 25–44.

Wyllie, P. J. 1971. The dynamic Earth. – J. Wiley & Sons, New York: 416 pp.

Nares Strait: a conflict between plate tectonic predictions and geological interpretation

SHIRI P. SRIVASTAVA and ROBIN K. H. FALCONER

Srivastava, S. P. & Falconer, R. K. H. 1982. Nares Strait: a conflict between plate tectonic predictions and geological interpretation. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 339–352.

The majority of published plate tectonic models concerning relative movement of North America and Greenland predict sinistral movement of 100 to 250 km along Nares Strait since the mid-Upper Cretaceous. Some models also predict an initial gap of up to 150 km across it in late Cretaceous time. If geological and structural data across Nares Strait convincingly preclude such movements, the plate tectonic models for the area must be revised. Different plate configurations and the problems they pose are presented when a) the Queen Elizabeth Islands are considered part of the Greenland plate and Parry Channel is considered the fossil boundary between the North American and Greenland plates, b) the North American and Greenland plates are considered non-rigid to accommodate the relative motions between them, thereby involving virtually no movement along Nares Strait and c) there is a combination of the previous two situations. The pre-drift reconstructions of Greenland relative to North America show that, if no strike-slip motion is allowed along Nares Strait, serious overlaps occur between the continental crust of the two plates. It is concluded from these reconstructions that either strike-slip motion did take place along Nares Strait as the plate tectonic models have predicted or far more deformation of the plates, not supported from their presently known geology, must have taken place. More detailed geological and geophysical information is needed in the regions surrounding Nares Strait before a definite answer to this problem can be obtained.

S. P. Srivastava and R. K. H. Falconer, Department of Energy, Mines and Resources, Geological Survey of Canada, Bedford Institute of Oceanography, Dartmouth, Nova Scotia, Canada, B2Y 4A2. R. K. H. F. present address: Georesearch Associates, Offshore Research and Development Consultants, P. O. Box 2351, Wellington, New Zealand.

Nares Strait is the most prominent physiographic feature between Greenland and Ellesmere Island (Fig. 1). It has attracted the attention of a large number of geologists, geomorphologists and geophysicists since the suggestion made by Taylor (1910) that it was the locus of a strike-slip fault along which North America had moved southwest away from Greenland. Taylor's idea of a strike-slip fault since then has been adopted by a majority of the geophysicists concerned with the evolution of the North Atlantic, the Labrador Sea, Baffin Bay and the Norwegian–Greenland Sea. It has now been well documented from the results of the Deep Sea Drilling Project and from systematic geophysical surveys of the oceanic regions in the world (for details e.g. see Cox 1973) that large parts of the present oceanic regions were formed due to the separation of lithospheric plates (which include the present landmasses) from each other. Thus, by identifying boundaries of the different lithospheric plates involved and by delineating their direction of motion one can reconstruct their palaeopositions. One of the basic concepts is that the plates involved remained rigid throughout the development of the oceanic regions in question, unless sufficient geological evidence exists to suggest otherwise. Based on this assumption several palaeogeographic reconstructions of Greenland relative to North America and Eurasia have been proposed (e.g. Bullard et al. 1965, Pitman & Talwani 1972, Kristoffersen & Talwani 1977, Le Pichon et al. 1977, Sclater et al. 1977, Kristoffersen 1978, Srivastava 1978).

In many of these reconstructions, Nares Strait has been considered the boundary between the North American and Greenland plates. Few geophysical measurements exist in this region to support such an assumption. Nonetheless, the physiography of this region favours it be a structural boundary, like a transform fault as was pointed out by Wilson (1965). Geologists, on the contrary, have interpreted that structural and lithological units continue across the Strait with little or no offset.

It is beyond the scope of the present paper to describe in detail the land geology bordering Nares Strait; however, a brief description relevant to this paper has been included from the work of Dawes (1973, 1976) and Kerr (1967). The rocks surrounding Nares Strait which form part of Ellesmere Island and northern Greenland fall into four groups; the crystalline basement rocks forming part of the Greenland–Canadian Shield, the un-

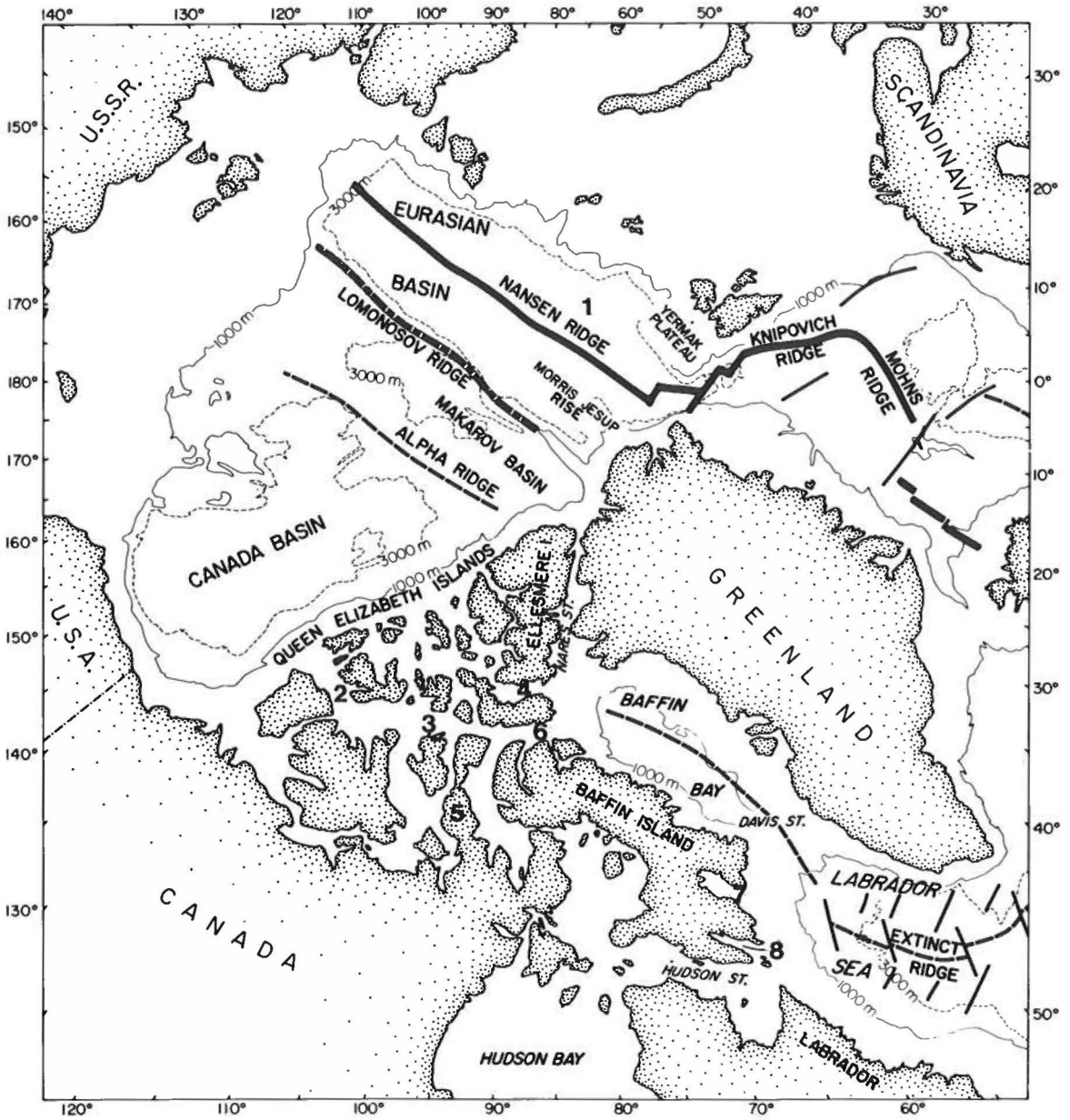


Fig. 1. Map of the regions surrounding Nares Strait. 1 = Nansen Basin, 2 = Viscount Melville Sound, 3 = Parry Channel, 4 = Jones Sound, 5 = Boothia Peninsula, 6 = Lancaster Sound, 7 = Cumberland Sound, 8 = Frobisher Bay.

metamorphosed rocks of Proterozoic and Lower Palaeozoic age forming the stable platform overlying the Shield, and the folded mainly Lower Palaeozoic and Devonian sediments of the Franklinian geosyncline that have been deformed into the Ellesmere–Greenland fold belt. The fourth group is represented by less severely deformed Upper Palaeozoic, Mesozoic and Tertiary strata (Sverdrup Basin) that mainly occupy the central region of Ellesmere Island, but which are not present in Greenland on the shores of Nares Strait. However, a similar Carboniferous to Tertiary sedimentary succession (Wandel Sea Basin) occurs in eastern North Greenland while a volcanic province consisting of mainly rhyolitic lavas and tuffs occurs along the northernmost coast on the shore of the Eurasian Basin. The age of the volcanic rocks has been isotopically dated at about the Cretaceous–Tertiary boundary (Larsen et al. 1978).

Even though different rock units of the Franklinian geosyncline on Ellesmere Island can be correlated with similar rock units on Greenland, the folded rocks show major structural differences. The main difference be-

tween them is that in Greenland axial planes of folds dip toward the craton (Dawes 1973, 1976) whereas in Ellesmere Island they more commonly dip towards the Arctic Ocean (Kerr 1967), but there are some exceptions. Such a marked difference in structural style must in some way be controlled by the structure of the Strait. Nares Strait must have been a significant structural site when the difference in fold pattern emerged, probably in the Palaeozoic (Dawes 1973).

According to Dawes et al. (1980), rock units of the Proterozoic Thule Group are present on opposite sides of northern Baffin Bay and Smith Sound on NW Greenland and SE Ellesmere Island. They maintain that certain rock units are so similar in lithology and thickness that unit to unit correlation of both sedimentary and volcanic rocks has been established. This correlation, they suggest, strongly supports the concept of a single intracratonic basin that shows little or no lateral displacement of its units. On the contrary, the projection of the boundary between the folded and unfolded rocks of the Franklinian geosyncline onto a median line in the Strait suggests sinistral displacement of at least 200 km. Whether one should use this boundary to show strike-slip movement along Nares Strait is debatable because of the structural differences which exist between the fold belt on the Canadian and Greenland sides. In general, geologists maintain the opinion that less than 50 km or so of transcurrent motion has taken place along Nares Strait. This has created a dilemma in the minds of many. A detailed account of this controversy has recently been given by Kerr (1980b).

The question therefore arises whether one can explain the evolution of the oceanic regions adjacent to Nares Strait using sea-floor spreading theory and at the same time keep the northern part of Greenland more or less in its present position relative to Ellesmere Island. We have tried to answer this question first by examining the position of Greenland relative to North America and in particular of Ellesmere Island during the successive stages of evolution of the Labrador Sea, Baffin Bay and the Norwegian–Greenland Sea; secondly by examining the consequences if the Queen Elizabeth Islands in the Canadian Arctic are regarded as attached to Greenland throughout their geological history; and thirdly by examining the possibility of non-rigid behaviour of the plates involved.

The regions surrounding Nares Strait

Nares Strait lies near three oceanic regions whose tectonic development has a direct bearing on the dilemma of Nares Strait. These are the Arctic Ocean Basin to the north (this includes Canada Basin, Makarov Basin, Eurasian Basin and the Alpha, Nansen and Lomonosov Ridges), the Norwegian–Greenland Sea to the east and Baffin Bay – Labrador Sea to the south (Fig. 1). Of the three regions the Arctic Ocean Basin is the least explored except for its eastern part, the Eurasian Basin. Recent geophysical measurements carried out in the Eurasian Basin and Norwegian–Greenland Sea (Vogt et al. 1978, 1979, 1981, Feden et al. 1979) have shown that it started to form at the same time as the Greenland Sea when the Lomonosov Ridge (presumably a continental fragment) separated from Scandinavia and northeastern USSR. Geophysical measurements in the Norwegian–Greenland Sea (Talwani & Eldholm 1977) and the Labrador Sea (Srivastava 1978) show that these regions were formed by a sea-floor spreading process due to the separation of Greenland from Eurasia and North America, respectively. These plates have therefore played an important role in the development of oceanic basins not only to the east and west of Greenland but also to the north of it. Nares Strait seems to be the obvious physiographic feature which connects the three basins. It is for this reason that it has been regarded as a boundary between the North American and Greenland plates along which the two plates have moved.

We will first review the geophysical evidence which has been used to substantiate sea-floor spreading in these regions. Fig. 2 is a composite diagram of the magnetic lineations and fracture zones which have been identified in the Labrador Sea, the northeast Atlantic, the Norwegian–Greenland Sea and in the Eurasian Basin. These results have been compiled mainly from published work by Vogt & Avery (1974), Talwani & Eldholm (1977), Srivastava (1978), Grønlie & Talwani (1978), Voppel et al. (1979), Voppel & Rudloff (1980) and Vogt et al. (1981). We have used the more recent time scale of Berggren et al. (1978) rather than the older one of Heirtzler et al. (1968) in assigning ages to each of the magnetic anomalies. Features worth noticing from this diagram are: a) anomalies older than 24 (53 m.y.) lie mainly in the Labrador Sea and in the North Atlantic south of Greenland, b) anomaly 13 (36 m.y.) does not lie in the Labrador Sea but continues north of Charlie–Gibbs Fracture Zone in the northeast Atlantic towards Iceland and c) anomaly 21 (48 m.y.) and possibly 24 are the oldest identifiable anomalies which lie in the Eurasian Basin.

The occurrence of anomaly 24 in the Eurasian Basin, Norwegian–Greenland Sea, Labrador Sea and in the North Atlantic both north and south of the Charlie–Gibbs Fracture Zone clearly shows that spreading in these regions was simultaneous. This implies formation of a triple junction not only south of Greenland but also north of it. The evolution of the Arctic Basin as a whole is not yet well understood. However, from the pattern of the magnetic lineations in the Eurasian Basin (Vogt et al. 1979, Feden et al. 1979) it is most likely that a FFR (Fault, Fault, Ridge) triple junction (e.g. see Pierce, this volume) existed north of Greenland. The triple junction south of Greenland was RRR (Ridge, Ridge, Ridge) type (Laughton 1971, Kristoffersen &

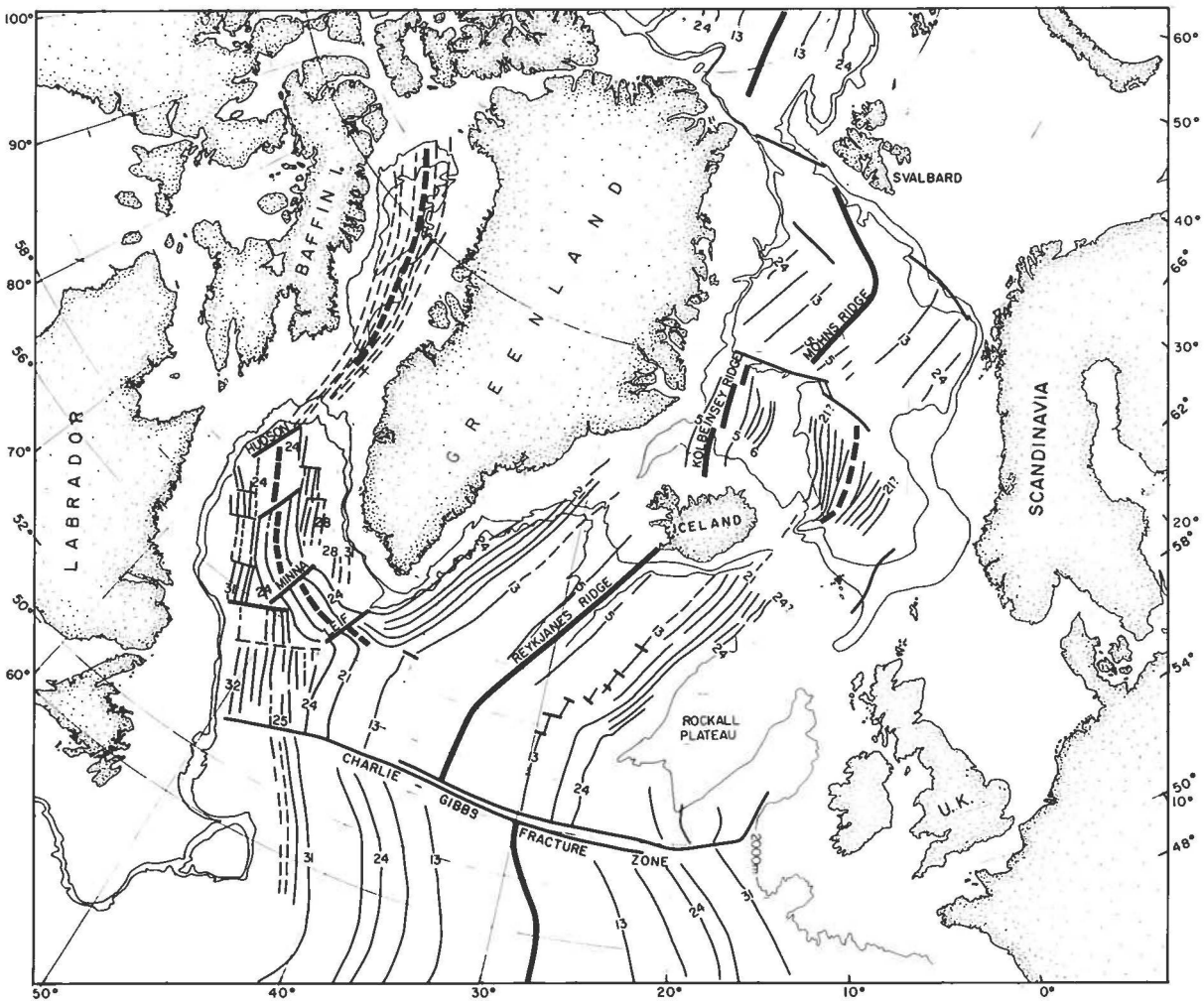


Fig. 2. Magnetic lineations and fracture zones in the North Atlantic Ocean, Labrador Sea and Norwegian–Greenland Sea compiled from sources listed in the text.

Talwani 1977). The absence of anomaly 13 in the Labrador Sea and its continuation towards Iceland has been interpreted as establishing an indication of the time when sea-floor spreading stopped in the Labrador Sea, thus ending the active presence of two triple junctions.

The existence of a triple junction (FFR) north of Greenland is based on the assumption that the Lomonosov Ridge was coupled to the North American plate rather than the Greenland plate (Srivastava & Falconer 1979). Le Pichon et al. (1977), on the other hand, considered that the Lomonosov Ridge was rigidly attached to Greenland through the Canadian Arctic Islands. Though their fit did not involve any lateral motion along Nares Strait, it did across the Boothia Arch with some suggestion of overlap between the Arctic Islands that form part of the Canadian Shield.

We will briefly examine some of the problems arising from different reconstructions.

Nares Strait: a plate boundary

Palaeogeographic positions of Greenland and Eurasia relative to North America at different geological times are shown in Fig. 3. These were obtained by using the pole positions for different regions as given by Srivastava (1978). Reconstructions given by others (Bullard et al. 1965, Pitman & Talwani 1972, Le Pichon et al. 1977, Sclater et al. 1977) based on criteria other than magnetic lineations and fracture zones exist, and differ only in details from the ones shown in Fig. 3.

The sequence of events considered in deriving the palaeogeographic reconstructions is summarized in Table 1. The palaeogeographic reconstructions in Fig. 3 show that Nares Strait has not only been a site of transcurrent motion between northern Greenland and Ellesmere Island, but also a region of compression or sub-

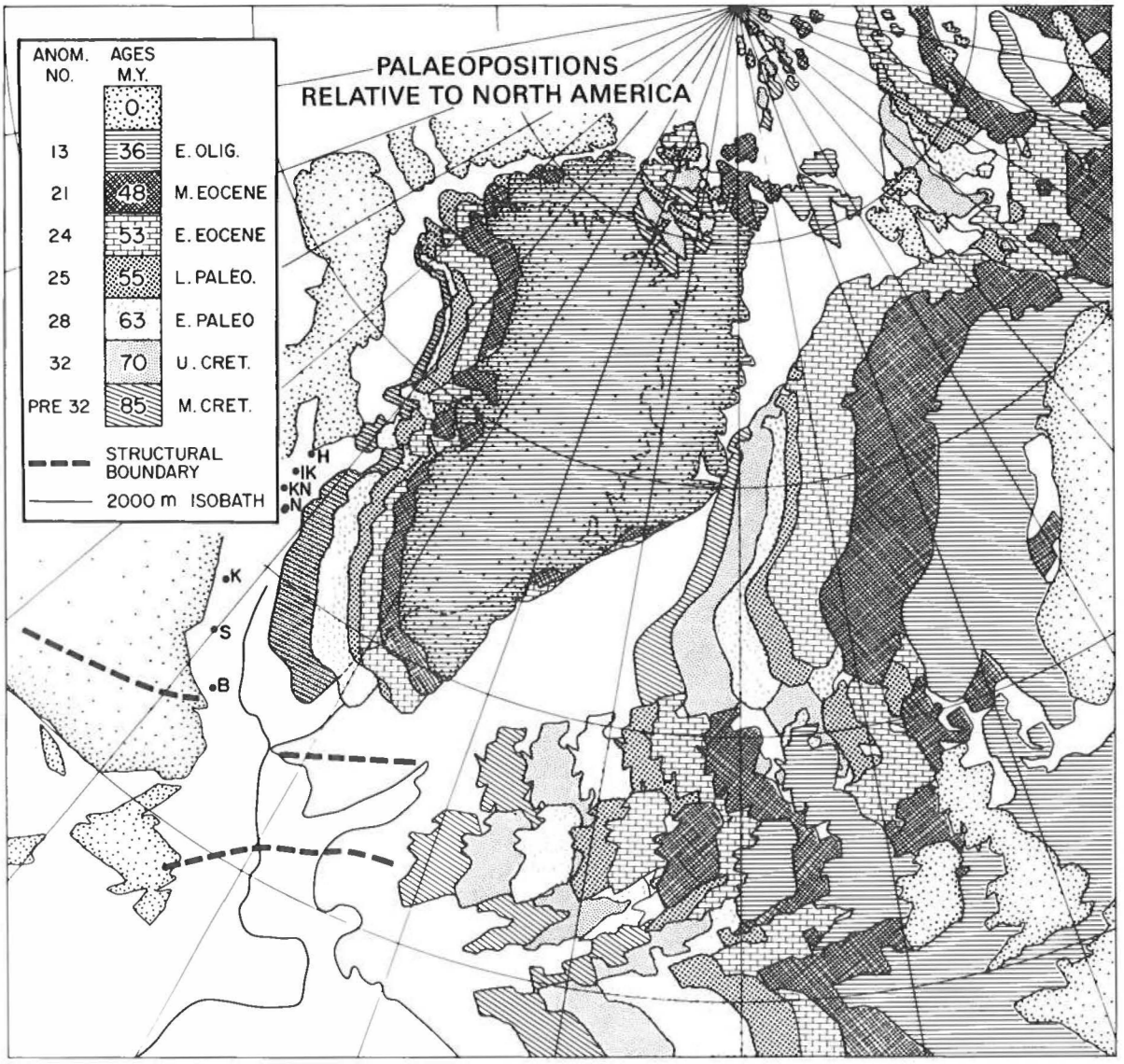


Fig. 3. The palaeogeographic positions of Greenland and Eurasia relative to North America during the successive stages of evolution of the North Atlantic, the Labrador Sea and the Norwegian–Greenland Sea based on the poles of rotation as given by Srivastava (1978). Also shown are the positions of five wells (H–Hellefisk, IK–Ikermiut, KN–Kangâmiut, and N–Nukik 1 and Nukik 2) on the Greenland Shelf and of three wells (K–Karlsefni, S–Snorri, and B–Bjarni) on the Labrador Shelf. Wells on the West Greenland Shelf are rotated to conform with pre-drift plate positions illustrated. The Queen Elizabeth Islands in the Canadian Arctic have been regarded as part of the North American craton in this reconstruction.

duction. We will examine briefly each of these situations.

Fig. 4 shows positions of Greenland relative to North America prior to active sea-floor spreading in the Labrador Sea (Fig. 4a) and at the time of anomaly 25 (55 m.y., Fig. 4b) during which major volcanism took place in the North Atlantic (formation of Iceland–Faeroe Ridge), the northern Labrador Sea, (formation of Davis Strait Sill) and southern Eurasian Basin (formation of Morris Jesup Rise and Yermak Plateau). For the sake of simplicity we have not tried to show the plate boundaries in these reconstructions. Similar reconstructions with plate boundaries have been published by Srivastava (1978). The regions surrounding Nares Strait are our main concern here. At the time of initial opening of the Labrador Sea (Fig. 4a) the position of Greenland shows a gap of 220 km between it and the northern tip of Ellesmere Island, a net displacement from its present position of 330 km to the southwest near the mouth of Smith Sound and of 600 km to the southwest of its southernmost tip. This, however, does not necessarily imply formation of new crust of these magnitudes in

Table 1. Summary of the tectonic events in the North Atlantic, Labrador Sea, Eurasian Basin and Norwegian–Greenland Sea.

Age m.y.	Mag. Ano. No.	North Atlantic	Labrador Sea & Baffin Bay	Eurasian Basin, Norwegian–Greenland Sea
110?		Iberia started to separate from North America thus creating the Newfoundland Basin between Grand Banks and Iberia (1), (2), (3)	Volcanism on the southern Labrador Shelf and on the coast associated with the initial stages of sea-floor spreading in the Labrador Sea (11)	–
95		Bay of Biscay started to open due to rotation of Iberia from Europe (1), (2), (3)	–	–
85	34	Rockall Trough opened due to separation of Rockall and Greenland away from Europe (1) to (4)	–	Commencement of spreading in Makarov Basin (8), (9)
70	32	Opening in Rockall Trough stopped and shifted to the west between southern Labrador coast and Rockall Bank as well to the south between Grand Banks and British Isles (4)	Active sea-floor spreading commenced in the southern Labrador Sea. This gave rise to compressive motion between Queen Elizabeth Islands and northern Greenland (4)	Shear motions between eastern Greenland and Europe and some compression between northern Greenland and Svalbard (4)
63	28	Opening continuing (4)	Active sea-floor spreading commenced in the northern Labrador Sea. More compression between Queen Elizabeth Islands and northern Greenland or along the northern margin of Makarov and Canada Basins (4)	– do –
55	25	– do – Volcanism forming Thulean Rise	Opening in Baffin Bay starting. Volcanism in Davis Strait, on Baffin Island as well as on West Greenland coast. Compression and translational motion between Greenland and Queen Elizabeth Islands (4)	Volcanism in eastern Greenland near Scoresby Sund and on Voring Plateau, Faeroe Islands, on Morris Jesup Rise and Yermak Plateau (4), (10)
53	24	Sea-floor spreading continuing	Active sea-floor spreading commences in Baffin Bay. Translation motion between Greenland and Queen Elizabeth Islands. Drastic change in direction of motion between Greenland and North American plates (4), (5)	Spreading in Makarov Basin stopped and Lomonosov Ridge got firmly coupled to its eastern flank. Active sea-floor spreading started in Nansen Basin, Norwegian–Greenland Sea. Strike-slip motion between Svalbard and northern Greenland (4), (6), (8), (9)
48	21	– do –	Change in direction of motion between Greenland and North America. Strike-slip motion continued between Greenland and Queen Elizabeth Islands (4), (6)	Jump in the ridge axis to the west and formation of Iceland–Faeroe Ridge (7)
36	13	– do –	Extinction of sea-floor spreading in Baffin Bay and Labrador Sea (4), (6)	Extinction of sea-floor spreading in the Norwegian Sea. Break of the land bridge between northern Greenland and Svalbard. Commencement of sea-floor spreading west of the Norwegian Basin (6)

(1) Kristoffersen (1978), (2) Sclater et al. (1977), (3) Le Pichon et al. (1977), (4) Srivastava (1978), (5) Kristoffersen & Talwani (1977), (6) Talwani & Eldholm (1977), (7) Voppel et al. (1979), (8) Srivastava & Falconer (1979), (9) Taylor et al. (1980), (10) Feden et al. (1979), (11) McWhae & Michel (1975).

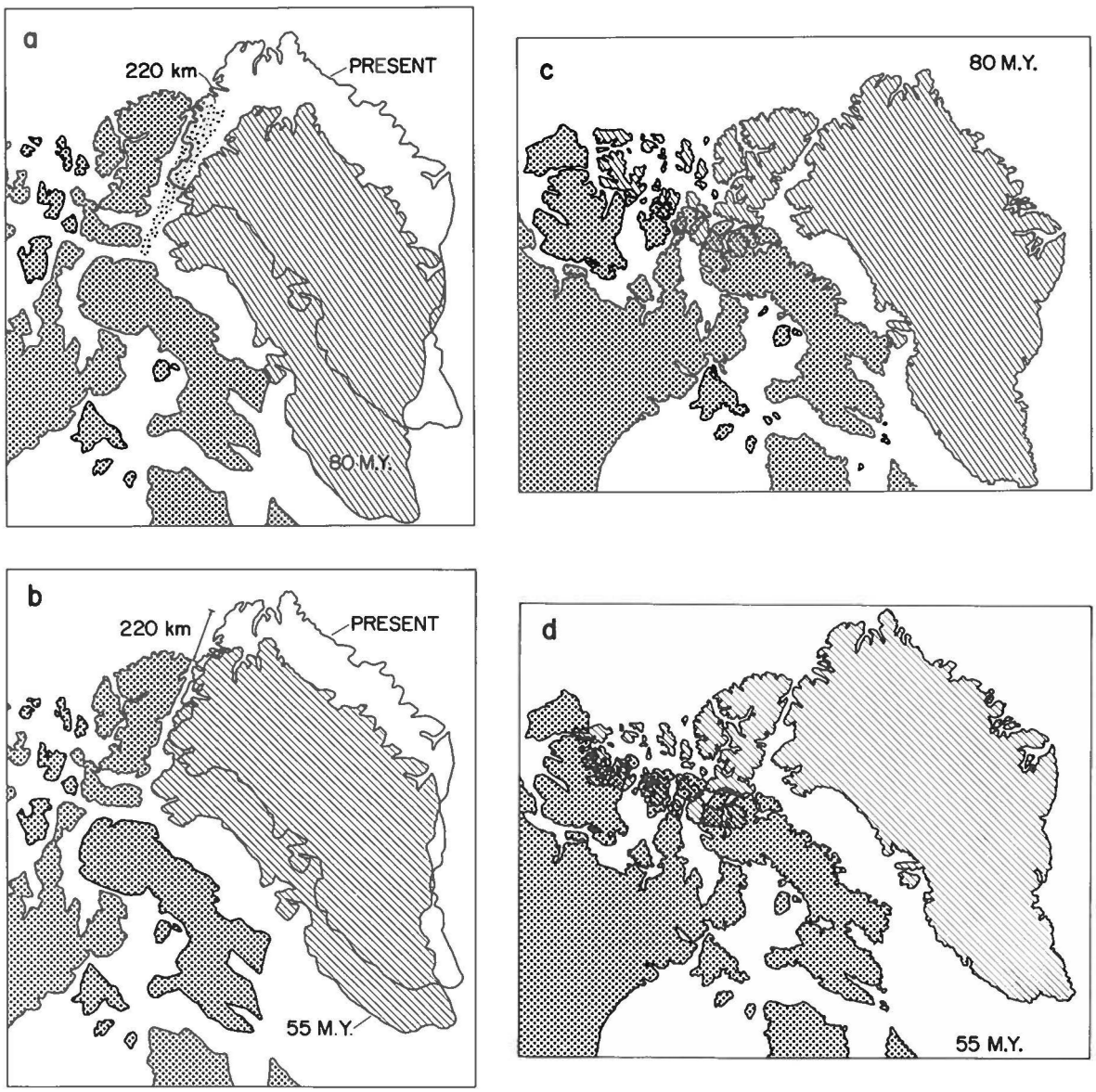


Fig. 4. Palaeogeographic position of Greenland relative to North America prior to active sea-floor spreading in the Labrador Sea (80 m.y.) and at anomaly 25 time (55 m.y.) for the cases when the plate boundary between North America and Greenland lies along Nares Strait (a, b) and when it lies along Parry Channel (c, d).

these regions since the inception of sea-floor spreading in the Labrador Sea. The motion of Greenland in the north was due to translation as well as compression while in the south it was mainly due to extension. Reconstruction at anomaly 25 time (Fig. 4b) shows a left-lateral displacement of 220 km between Greenland and Ellesmere Island along Nares Strait. From the time of initial spreading (Fig. 4a) to anomaly 25 time (Fig. 4b) Greenland moved north with a pivot at the mouth of Lancaster Sound. This perhaps gave rise to compression between Ellesmere Island and northern Greenland. The question arises as to what happened to the material which lay between Ellesmere Island and Greenland. There are three possibilities: a) either it was subducted below Greenland or b) the gap never existed and the required amount of motion was taken up within the Canadian Arctic Islands — probably in the Sverdrup Basin or c) the plate kinematics are poorly controlled and the compressional movement is an artifact. According to Balkwill (1978), the main compression in the Sverdrup Basin took place between mid-Eocene and early Miocene with a maximum lateral shortening of 40

km. This is contrary to what the reconstruction in Fig. 4b suggests. The likelihood of each of these possibilities is further examined later.

Lancaster Sound, Viscount Melville Sound: a plate boundary

If it is assumed that no lateral displacement took place along Nares Strait and that the Queen Elizabeth Islands were always firmly coupled in their present configuration to Greenland rather than to North America, then somewhat different situations are found in the reconstructions (Fig. 4c and 4d). The large overlap between the islands located on either side of Lancaster Sound and Viscount Melville Sound, which results from such reconstructions, is — from a geological point of view — as undesirable, if not more, as the strike-slip motion along Nares Strait.

Viscount Melville and Lancaster Sounds, which form a long arcuate feature, have been regarded as the boundary between the North American and Greenland–Queen Elizabeth Islands plates in these reconstructions mainly because next to Nares Strait they comprise the most prominent physiographic feature. It is equally likely that such a boundary could lie somewhere else among the Arctic Islands, or that the boundary has migrated with time during various episodes of the evolution of the Labrador Sea and Arctic Basin.

The North American plate: a non-rigid plate

Kerr (1981) proposed a configuration of the North American plate including the Canadian Arctic Archipelago relative to Greenland where he has closed part of the Labrador Sea and Baffin Bay without invoking any lateral displacement between Greenland and Ellesmere Island. This is shown in Fig. 5a. The dark solid lines show the regions where compression and crustal extension took place in the North American plate.

The physiography of a number of channels in the Canadian Arctic suggests that they may have been the sites of crustal extension during the early stages of sea-floor spreading (Beh 1975, van der Linden 1977), thus forming grabens and half grabens. Crustal structure under most of these straits still remains unknown, though Kerr (1980a) has interpreted the structure under Lancaster Sound from some neighbouring seismic reflection data.

In his reconstruction, Kerr (1981) has assumed Baffin Bay to be underlain largely by continental crust thereby implying that little or no sea-floor spreading took place in this region. The small movement of Greenland relative to Baffin Island needed in his reconstruction is accommodated by crustal stretching in the formation of Hudson Strait, Frobisher Bay and Cumberland Sound. No refraction data exist to support that the crust under these regions is attenuated. Similarly, the relative movements needed between Ellesmere Island and northern Greenland to close a portion of the Labrador Sea are accommodated by opening a portion of the Sverdrup Basin.

To see the amount of compression that must have occurred within the Sverdrup Basin, we have modified Kerr's reconstruction by considering North America as a rigid plate, with the results shown in Fig. 5b. It shows that about 130 km and 50 km of compression is required at the north and south end of Nares Strait respectively. On the other hand, Balkwill (1978) suggested a maximum compression of 40 km only.

A large portion of the Labrador Sea which to us appears to be underlain by oceanic crust remains open in Kerr's reconstruction. If one regards it as underlain by continental crust then a serious situation develops between Greenland and Eurasia. It has been well accepted by a majority of the workers that the oceanic regions between eastern Greenland and Eurasia and between Eurasia and North America south of Greenland were developed by sea-floor spreading (e.g. see Talwani & Eldholm 1977). Thus, if we close the Norwegian–Greenland Sea with respect to North America to the time of initial opening (Srivastava 1978) and use the position of Greenland as given by Kerr (1981) we find that most of Svalbard overlies northern Greenland (Fig. 5c). Alternatively if we close the Norwegian–Greenland Sea using the poles of rotation given by Talwani & Eldholm (1977) in Fig. 5c a large portion of the North Atlantic south of Greenland, which is definitely oceanic in origin, will remain open.

Fig. 5d shows a reconstruction where Fig. 4a and 5a have been combined. It shows that accommodation of some movement through a deformation of the North American plate (between 50 and 130 km, Fig. 5b) results in a smaller gap (100 km against 220 km) at the northern end of Nares Strait, and less overlap across Davis Strait but still requires 220 km of translational motion along Nares Strait.

The Greenland plate: a non-rigid plate

The reconstructions in Fig. 5c and 5d show that even if some of the relative motion between the Canadian Arctic Islands and northern Greenland is accommodated by deforming the North American plate, considerable lateral motion between northern Greenland and Ellesmere Island is still required. One way to overcome this problem would be to treat Greenland also as a non-rigid plate.

Beh (1975) considered such a situation and post-

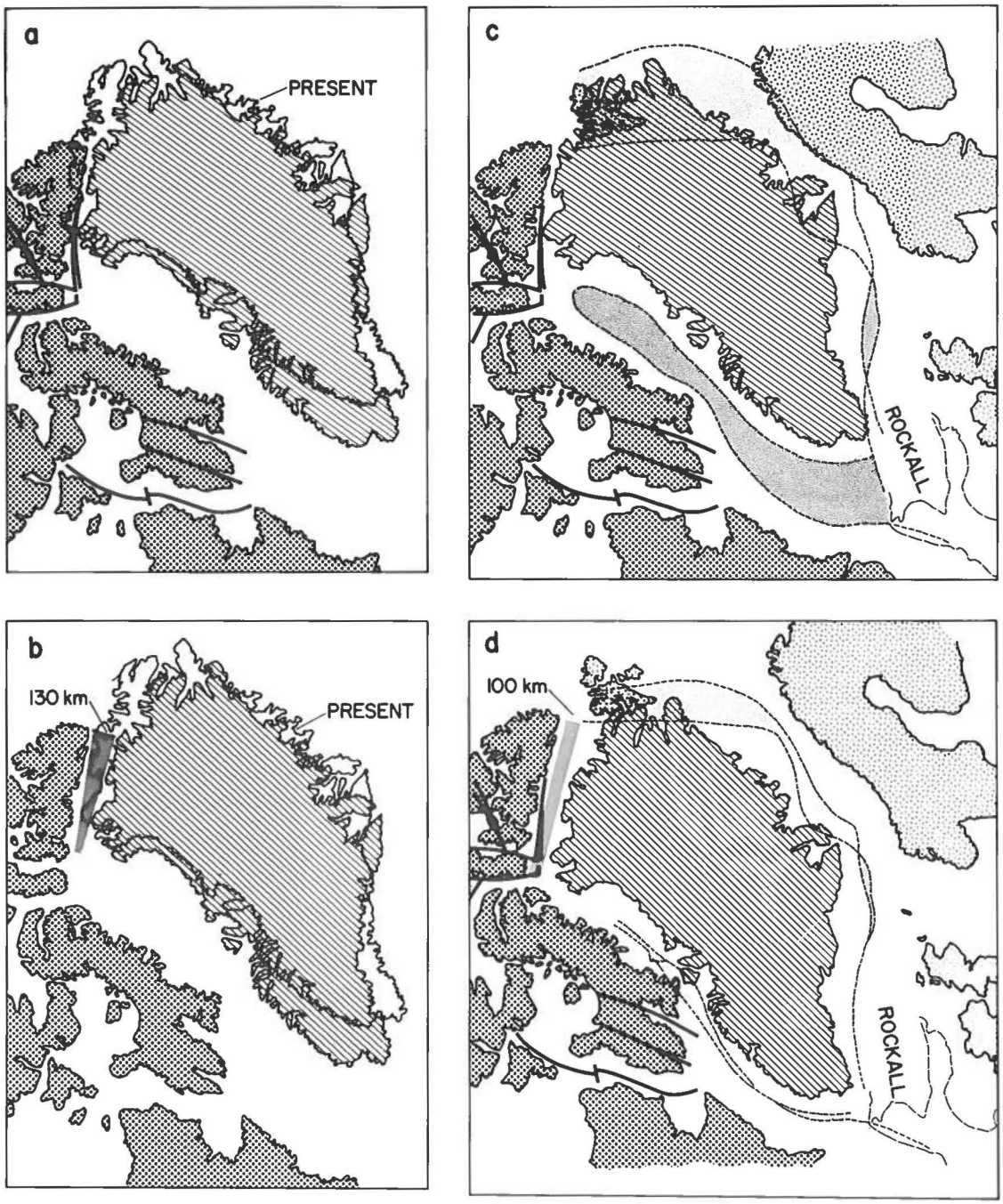


Fig. 5. Palaeogeographic positions of Greenland (striped pattern) relative to North America (dotted pattern): a) as given by Kerr (1981) where part of the relative movement has been accommodated by deforming the continental part of the present North American plate along thick lines, b) Kerr's (1981) reconstruction where the North American plate is considered a rigid plate, c) as given by Kerr (1981) together with the position of the Eurasian plate (open circle pattern) relative to North America as given by Srivastava (1978), and d) at the time of initial opening (Fig. 4a) where North America is considered a non-rigid plate as proposed by Kerr (1981).

ulated a number of shear zones running across the central part of Greenland. He based his interpretation on the published geological information of the Canadian Arctic Islands, on the physiography of the channels running among the islands as well as to the south of them, and the fact that a major part of the Labrador Sea

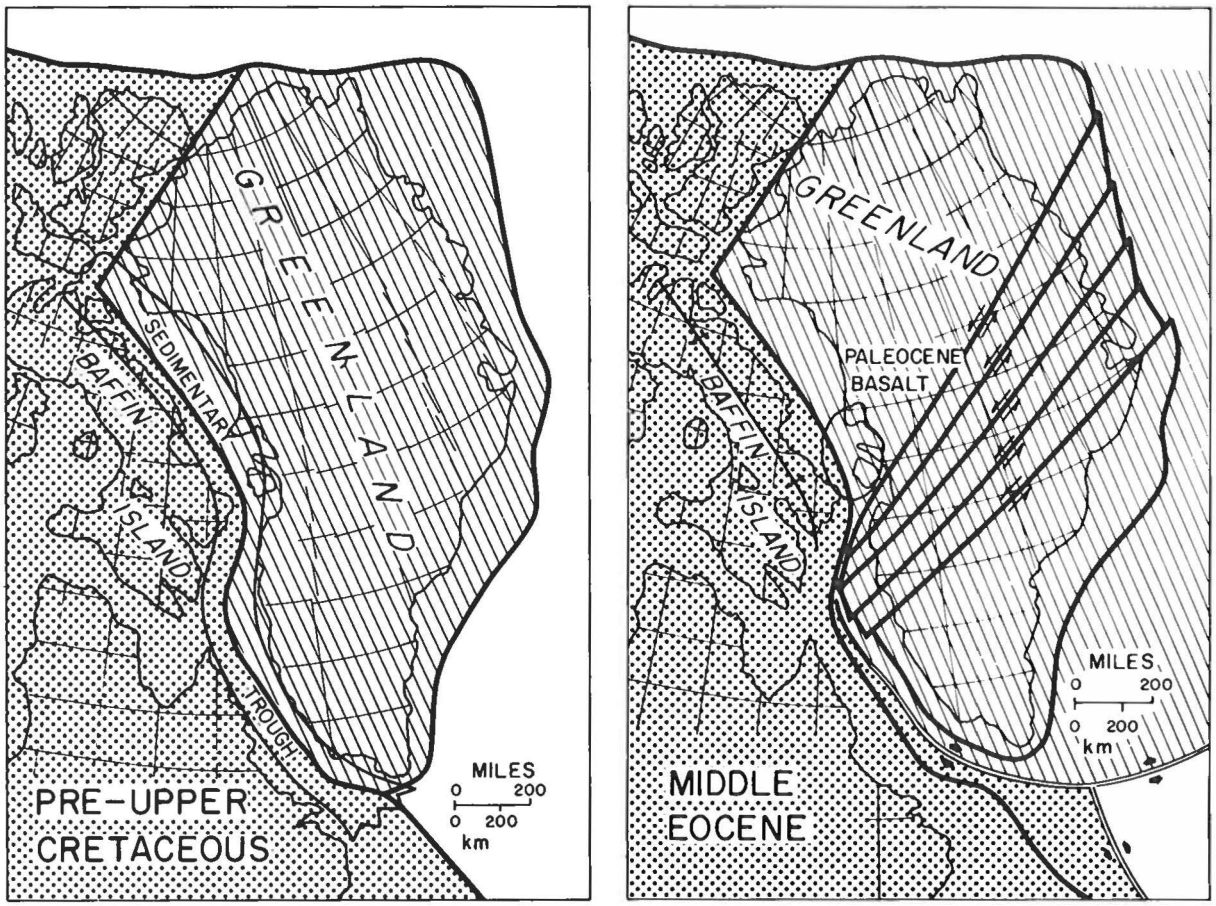


Fig. 6. Palaeogeographic position of Greenland relative to North America in pre-Upper Cretaceous (left) and Middle Eocene times, as proposed by Beh (1975).

and some part of Baffin Bay were formed by sea-floor spreading. Figures 6a and 6b show relative positions of the North American and Greenland plates at pre-Upper Cretaceous and Middle Eocene times. These two reconstructions seem to have overcome the two major objections which geologists have against other reconstructions, namely a) there is hardly any overlap between the landmasses across Davis Strait, and b) the northeastern part of Ellesmere Island remains in its present position relative to northern Greenland. These are achieved by treating the North American and Greenland plates as non-rigid.

According to Beh (1975) sea-floor spreading in the Labrador Sea and Baffin Bay has been accommodated by strike-slip motion along Nares Strait since Middle Eocene (anomaly 21 time) and along some shear zones running across central Greenland for earlier times. Bak et al. (1975) show the presence of some shear zones on the west coast of Greenland but their continuation right across Greenland is not certain. Also, these shear zones are Precambrian features associated with the Proterozoic orogenesis and whether they were rejuvenated during Tertiary is not certain. Beh implies lateral displacement of Greenland relative to Ellesmere Island but in fact his net displacement of the northern part of Ellesmere Island relative to Greenland between mid-Cretaceous (pre sea-floor spreading) and Recent times (post sea-floor spreading in the Labrador Sea) remains small. This is achieved by keeping a large portion of the Sverdrup Basin and the region underlain by Jones, Lancaster, and Cumberland Sounds, Frobisher Bay and Hudson Strait closed by mid-Eocene and then opening them during a later stage of opening in the Labrador Sea. This is contrary to what Kerr (1981) has considered in his reconstructions. Kerr regards the Sverdrup Basin as having been far bigger, prior to sea-floor spreading in the Labrador Sea, than it is today. However he does not give a time frame for the different stages of evolution of the Labrador Sea and Baffin Bay for comparison with Beh's (1975) suggestion.

The pre-drift position of Greenland relative to Ellesmere Island in our reconstruction (Fig. 4a) shows a gap between them of 220 km in the north and of 50 km in the south (near Jones Sound). By late Paleocene (Fig.

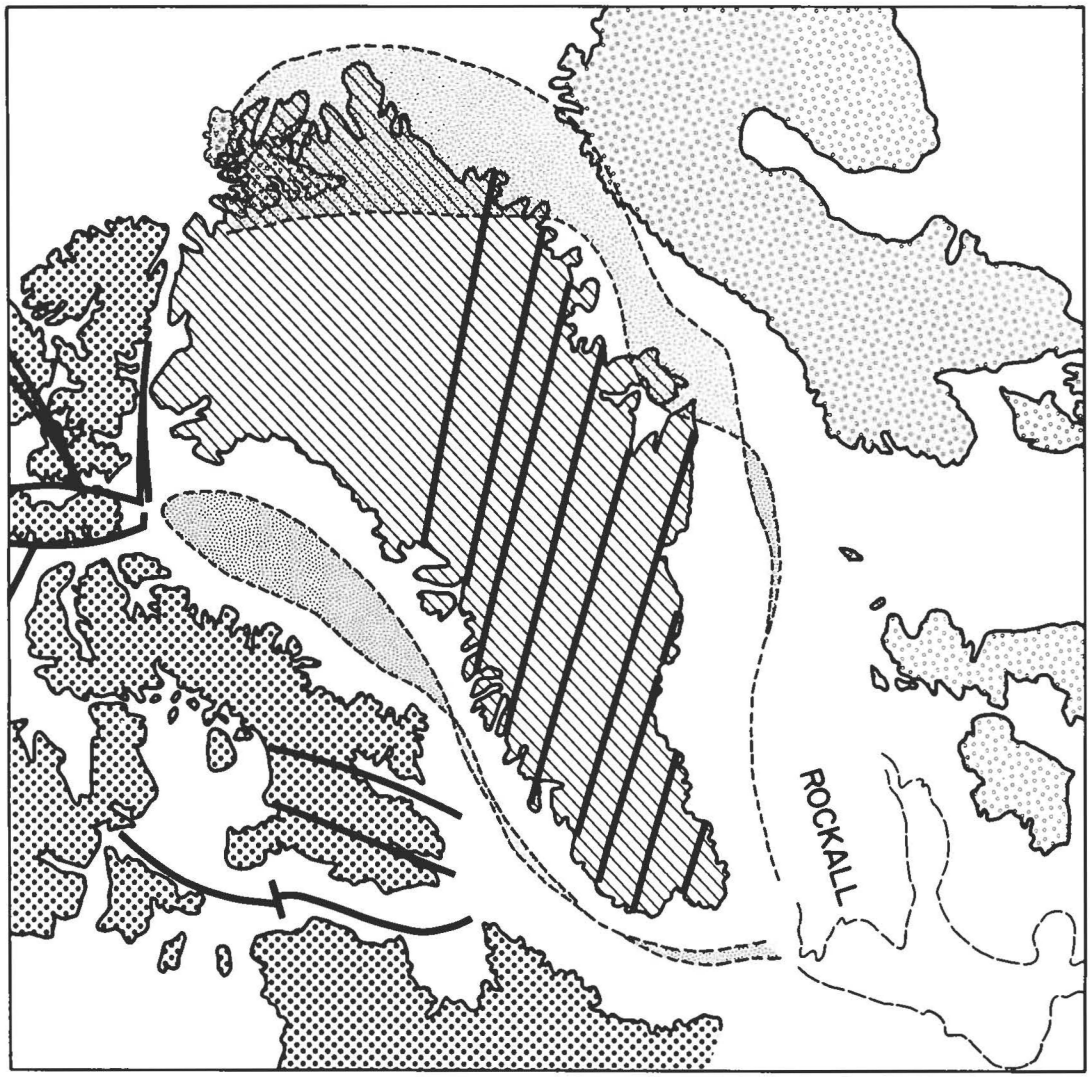


Fig. 7. Palaeogeographic position of Greenland relative to North America where both plates are considered non-rigid and have been deformed along thick lines. The reconstruction combines those given in Figs 4a, 5a and 5d.

4b) the gap is virtually closed and Greenland lies close to Ellesmere Island. Except for the Cretaceous–Tertiary volcanics in northern Greenland (Larsen et al. 1978) there is no other geological evidence either on Ellesmere Island or northern Greenland to support the subduction of material under them. It is thus very likely that the region of the gap was occupied by the southward extension of the far bigger Sverdrup Basin and the subsequent northward movement of Greenland was taken up by folding and faulting within this basin. During the subsequent stages of development of the Labrador Sea, Greenland moved along Nares Strait until early Oligocene when sea-floor spreading ceased in the Labrador Sea and it attained its present position.

If little or no lateral motion took place along Nares Strait as Kerr (1981) maintains, then the only choice would be to move the southern part of Greenland along the shear zones as suggested by Beh (1975). This is shown in Fig. 7 where we have closed the Labrador Sea by introducing shear motion across Greenland along 'hypothetical' shear zones and keeping Baffin Bay and Davis Strait virtually open. We have considered Kerr's (1981) configuration for the North American plate in this construction. A large, undesirable overlap exists between northern Greenland and Svalbard by doing so. This clearly shows that even if we treat the two plates as non-rigid plates strike-slip motion along Nares Strait is essential to avoid such overlaps.

Discussion

We have presented different models of plate configurations in the vicinity of Nares Strait. None of them explains satisfactorily the evolution of the Labrador Sea, Baffin Bay and the North Atlantic Ocean while showing little or no lateral displacement of Greenland relative to Ellesmere Island along Nares Strait. This is because the boundaries of the plates are not well defined throughout. According to plate tectonic theory, provided the plates remain rigid their movements on the surface of the Earth can be described by the poles of rotation over a given interval of time. Plates do break and change their direction of motion which results in a shift of their pole of rotation. In either case, if these changes can be mapped precisely, the movements of the plates and their palaeopositions can be calculated. On the other hand, if the plates acted non-rigidly, their movements cannot be described by use of the same technique and the problem becomes much more complicated in plate tectonic theory. The question of Nares Strait seems to fall somewhere between the two extremes as we have tried to show from the reconstructions. Geophysical measurements in the Labrador Sea show that its evolution can be described satisfactorily using plate tectonic theory by treating Greenland and North America as rigid plates (Srivastava 1978). However, problems arise when we extrapolate the movements of these plates to regions north of the Labrador Sea where the boundaries of these plates are ill defined and the geology of the Canadian Arctic Islands shows that some distortion of the plate and/or shift in the plate boundary was taking place during the same period. Thus models of plate configurations with different boundaries have been considered, which can be grouped into two classes: a) those which explain the evolution of the oceanic regions satisfactorily and show large-scale movements along Nares Strait and b) those which show no displacement along Nares Strait and do not explain the development of the oceanic regions consistent with the geophysical observations.

As we have shown in the preceding sections advocates of models of class a) have largely considered plates rigid while those of models of class b) have considered plates non-rigid and have based their reconstruction largely on matching the present land geology across Nares Strait. The only exception to this is the reconstruction by Beh (1975) who has tried to compromise the situation by introducing large lateral displacement within the Greenland plate along some 'hypothetical' shear zones. Some evidence does indeed exist on the west coast of Greenland for the presence of some shear zones which are related to Proterozoic orogenesis (Bak et al. 1975) but not to Tertiary movement as implied in Beh's (1975) reconstructions.

So what is the solution? The answer to this question rests mainly on the geological and geophysical evidence which has been used in constructing the models. If the geological mapping of land areas surrounding Nares Strait and their interpretation is accurate (for detailed discussion on this see Newman and other papers in this volume), then it is very unlikely that large strike-slip motion along Nares Strait took place as the plate tectonic models have implied. On the other hand, it is equally possible that some refinement in the plate tectonic models may result with the acquisition of additional geophysical data in regions where such is now lacking, but not to the extent to eliminate the strike-slip motion along Nares Strait altogether. In that case there are two possibilities: a) the interpretation of geophysical data in the Labrador Sea, Baffin Bay and Norwegian–Greenland Sea is incorrect, or b) the required motions between the plates were accommodated within the plate in the form of folding, faulting and graben formation. Each possibility will be discussed in turn.

a) Published geophysical information indicates strongly that the marine regions in question were largely formed by sea-floor spreading although in some areas, like northern Baffin Bay, the possibility exists that they may have been formed by stretching of continental crust. Northern Baffin Bay lies close to the pole of rotation for the Labrador Sea from pre-opening time to late Paleocene (Srivastava 1978). During this period, when active sea-floor spreading was taking place in the Labrador Sea, the region of northern Baffin Bay was merely stretched due to large tensional forces. Thus, true oceanic crust was formed in one region while in another region it was the continental crust that was stretched. In either case the amount of movement to be accounted for in plate reconstructions still remains the same. The problem then reduces to one of simple plate kinematics, yielding solutions which show not only large-scale strike-slip motion along Nares Strait but also some compression between Greenland and Ellesmere Island. These solutions regard Nares Strait as a boundary between North America and Greenland plates. We have shown in the preceding sections that the problem even gets worse if we consider the boundary to be somewhere else.

b) Geological information in the Canadian Arctic Archipelago shows that the Sverdrup Basin was a site of crustal subsidence from early Carboniferous to late Cretaceous time and then was subjected to severe folding and faulting during the Eurekan orogeny which ceased prior to the Miocene (Balkwill 1978, Trettin & Balkwill 1979). Crustal subsidence, folding and faulting are often all manifestations of interaction between plates, and their occurrence in the development of the Sverdrup Basin as it exists today must be related to the interaction of the North American plate with the Greenland and Eurasian plates. In general, the tectonic events based on the land geology of the Canadian Arctic Archipelago agree well with the type of interaction between the North American and Greenland plates as predicted from plate tectonic models but they differ in detail. For example, the pre-drift position of Greenland

relative to North America supports a larger extent of the Sverdrup Basin during this time. The Basin underwent folding and faulting subsequently, but the total amount of deformation, which includes compression as well as translation, predicted by the plate tectonic model does not agree with what can be interpreted from the presently known land geology. The situation is complicated by the fact that while the sizes of the plates were increasing in the south due to active sea-floor spreading in the Labrador Sea, they were decreasing in the north. Thus, in the south the plates were acting as rigid plates, while at the same time they were acting as non-rigid plates in the north.

Another source of difficulty in relating the predicted motions based on plate tectonic models with the land geology lies in the fact that a great portion of Greenland is ice-covered and all of its geology cannot be extrapolated from the west to the east coasts. To examine if the translational motion predicted from plate tectonic models along Nares Strait could be accommodated by shear motion within Greenland, several hypothetical shear zones were used in one of our reconstructions (Fig. 7). In spite of the introduction of these shear zones, we find that a serious overlap between Svalbard and northern Greenland still remains if no motion is allowed along Nares Strait.

It is thus concluded from the reconstructions presented here that if no motion is allowed along Nares Strait then either the kinematics of the plate tectonic evolution of the North Atlantic, based on magnetic isochrons and other geophysical data, is in error or much more deformation of the North American plate must have taken place than can be inferred from the presently known geology of the Arctic Archipelago. Sobczak (1980, this volume) using the trends of the folds and faults shows possibilities of large fragmentation of the Canadian Arctic Archipelago. Apparently, detailed geophysical information on Baffin Bay, Davis Strait and the oceanic regions surrounding and including Nares Strait, as well as detailed geology of the Arctic Archipelago and ice-covered regions of Greenland, is needed before an acceptable answer to the Nares Strait dilemma can be found.

Acknowledgements

We thank Miss H. R. Jackson, Dr. R. T. Haworth and Mr. R. F. Macnab for critically reading the manuscript.

References

Bak, J., Korstgård, J. & Sørensen, K. 1975. A major shear zone within the Nagssugtoqidian of West Greenland. – Tectonophysics 27: 191–209.

Balkwill, H. R. 1978. Evolution of Sverdrup Basin, Arctic Canada. – Bull. Am. Ass. Petrol. Geol. 62: 1004–1028.

Beh, R. L. 1975. Evolution and geology of western Baffin Bay and Davis Strait, Canada. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margins and offshore petroleum exploration. – Mem. Can. Soc. Petrol. Geol. 4: 453–476.

Berggren, W. A., McKenna, M. C., Hardenbol, J. & Obradovich, J. D. 1978. Revised Paleogene polarity time scale. – J. Geol. 86: 67–81.

Bullard, E., Everett, J. E. & Smith, A. G. 1965. The fit of continents around the Atlantic. – In: Blackett, P. M. S., Bullard, E. & Runcorn, S. K., A symposium on continental drift. – Phil. Trans. Roy. Soc. Lond. 258A: 41–51.

Cox, A. 1973. Plate tectonics and geomagnetic reversals. – W. H. Freeman & Co., San Francisco: 702 pp.

Dawes, P. R. 1973. The North Greenland fold belt: a clue to the history of the Arctic Ocean basin and the Nares Strait lineament. – In: Tarling, D. H. & Runcorn, S. K. (eds), Implications of continental drift to the earth sciences 2: 925–947. Academic Press, London & New York.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R., Frisch, T. & Christie, R. L. 1980. Archean-Proterozoic history and correlation of lands bordering northernmost Baffin Bay. – Geol. Ass. Can., Min. Ass. Can. Prog. with Abs. 5: 48 only.

Feden, R. H., Vogt, P. R. & Fleming, H. S. 1979. Magnetic and bathymetric evidence for the "Yermak hot spot" northwest of Svalbard in the Arctic Basin. – Earth planet. Sci. Lett. 44: 18–38.

Grønlie, G. & Talwani, M. 1978. Geophysical atlas of the Norwegian-Greenland Sea. – Verma Res. Ser. IV, Lamont-Doherty Geol. Observ., Columbia Univ., New York.

Heirtzler, J. R., Dickson, G. O., Herron, E. M., Pitman, W. C. & Le Pichon, X. 1968. Marine magnetic anomalies, geomagnetic field reversals, and motions of the ocean floor and continents. – J. geophys. Res. 73: 2119–2136.

Kerr, J. W. 1967. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.

Kerr, J. W. 1980a. Structural framework of Lancaster aulacogen, Arctic Canada. – Bull. geol. Surv. Can. 319: 24 pp.

Kerr, J. W. 1980b. Did Greenland drift along Nares Strait? – Bull. Can. Petrol. Geol. 28: 279–289.

Kerr, J. W. 1981. Evolution of the Canadian Arctic Islands: a transition between the Atlantic and Arctic Oceans. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 105–199. – Plenum Press, New York & London.

Kristoffersen, Y. 1978. Sea-floor spreading and the early opening of the North Atlantic. – Earth planet. Sci. Lett. 38: 273–290.

Kristoffersen, Y. & Talwani, M. 1977. Extinct triple junction south of Greenland and the Tertiary motion of Greenland relative to North America. – Bull. geol. Soc. Am. 88: 1037–1049.

Larsen, O., Dawes, P. R. & Soper, N. J. 1978. Rb/Sr age of the Kap Washington Group, Peary Land, North Greenland, and its geotectonic implication. – Rapp. Grønlands geol. Unders. 90: 115–119.

Laughton, A. S. 1971. South Labrador Sea and the evolution of the North Atlantic. – Nature, Lond. 232: 612–617.

Le Pichon, X., Sibuet, J.-C. & Francheteau, J. 1977. The fit of the continents around the North Atlantic Ocean. – Tectonophysics 38: 169–209.

McWhae, J. R. H. & Michel, W. F. E. 1975. Stratigraphy of Bjarni H–81 and Leif M–48, Labrador Shelf. – Bull. Can. Soc. Petrol. Geol. 23: 361–382.

Newman, P. H. 1982. A geological case for movement between Canada and Greenland along Nares Strait. – This volume.

Peirce, J. W. 1982. The evolution of the Nares Strait lineament and its relation to the Eurekan orogeny. – This volume.

Pitman, W. C. & Talwani, M. 1972. Sea-floor spreading in the North Atlantic. – Bull. geol. Soc. Am. 83: 619–646.

Sclater, J. G., Hellinger, S. & Tapscott, C. 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. – J. Geol. 85: 509–552.

Sobczak, L. W. 1980. Fragmentation of the Canadian Arctic Archipelago and Greenland. – Geol. Ass. Can., Min. Ass. Can. Prog. with Abs. 5: 81 only.

Sobczak, L. W. 1982. Fragmentation of the Canadian Arctic Archipelago, Greenland, and surrounding oceans. – This volume.

Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.

Srivastava, S. P. & Falconer, R. K. H. 1979. Review of plate tectonic models of the evolution of the Arctic Basin. – Trans. Am. geophys. Un. 60: 373 only.

Talwani, M. & Eldholm, O. 1977. Evolution of the Norwegian-Greenland Sea. – Bull. geol. Soc. Am. 88: 969–999.

Taylor, F. B. 1910. Bearing of the Tertiary mountain belt on the origin of the earth's plan. – Bull. geol. Soc. Am. 21: 170–226.

Taylor, P. T., Vogt, P. R., Kovacs, L. C. & Thomson, G. L. 1980. Tectonic implications from the west-Arctic Ocean Basin aeromagnetic surveys. – Trans. Am. geophys. Un. 61: 277 only.

Trettin, H. P. & Balkwill, H. R. 1979. Contributions to the tectonic history of the Innuitian Province, Arctic Canada. – Can. J. Earth Sci. 16: 748–769.

van der Linden, W. J. M. 1977. How much continent under the ocean? – J. Mar. geophys. Res. 3: 209–224.

Vogt, P. R. & Avery, O. E. 1974. Detailed magnetic surveys in the northeast Atlantic and Labrador Sea. – J. geophys. Res. 79: 363–389.

Vogt, P. R., Feden, R. H., Eldholm, O. & Sundvor, E. 1978. The ocean crust west and north of the Svalbard Archipelago: synthesis and review of new results. – Polarforschung 48: 1–19.

Vogt, P. R., Taylor, P. T., Kovacs, L. C. & Johnson, G. L. 1979. Detailed aeromagnetic investigation of the Arctic Basin. – J. geophys. Res. 84: 1071–1089.

Vogt, P. R., Perry, R. K., Feden, R. H., Fleming, H. S. & Cherkis, N. Z. 1981. The Greenland-Norwegian Sea and Iceland environment: geology and geophysics. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 493–598. – Plenum Press, New York & London.

Voppel, D. & Rudloff, R. 1980. On the evolution of the Reykjanes Ridge south of 60°N between 40 and 12 million years before present. – J. Geophys. 47: 61–66.

Voppel, D., Srivastava, S. P. & Fleischer, U. 1979. Detailed magnetic measurements south of the Iceland-Faeroe Ridge. – Dtsch. Hydrogr. Z. 32: 154–172.

Wilson, J. T. 1965. A new class of faults and their bearing on continental drift. – Nature, Lond. 207: 343–347.

Economic Geology

Nares Strait and the petroleum explorer

NEIL J. McMILLAN

McMillan, N. J. 1982. Nares Strait and the petroleum explorer. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 355–361.

It is important to be able to make intelligent guesses concerning the effect of lateral movement along Nares Strait on the resulting sedimentary section and the heat regime. A prognosis of the theoretical well — Lincoln No. 1 — would need to contain forecasts concerning the conditions of hydrocarbon generation in the sediments.

Case No. 1 for the postulated sedimentary section assumes 250 km of left-lateral movement along the Nares Strait lineament. Rocks are Lower Cretaceous to Pleistocene. Attendant heat is assumed to be high. Case No. 2 assumes very little movement along the lineament and, therefore, a relatively quiet tectonic history. Rocks are Cambrian to Pleistocene. Attendant heat is assumed to be average.

A simulated analysis of the geochemistry of the situation in Case No. 1 reveals that maximum expulsion of hydrocarbons occurs at 4000 metres' depth. In Case No. 2 the maximum expulsion would occur at 5000 metres. Therefore, it is tentatively concluded that Case No. 1 would provide the best conditions for hydrocarbon generation and entrapment at approximately 1000 metres shallower than Case No. 2.

Large-scale movement along the Nares Strait lineament and the attendant higher heat regime would be a beneficial circumstance for the petroleum explorer.

N. J. McMillan, Aquitaine Company of Canada Ltd., 1700 Selkirk House, 555–4th Ave. S.W., Calgary, Alberta, Canada, T2P 3J6. Present address: Canterra Energy Ltd., P. O. Box 1051, Calgary, Alberta, Canada, T2P 2K7.

It is an honour for me to be considered and requested to make a contribution to this up-to-date, thorough, and thoughtful volume. The Nares Strait lineament has attracted the attention of almost all scholars concerned with the shifting of the Earth's surface and I will not become entangled in the debate as to whether there has been substantial lateral movement along it or practically none. Rather, I will try to illustrate how a controversy such as this can make modelling for oil prospecting more hazardous than usual. There is some comfort in writing these pages which deal with possible future petroleum discoveries. It will be decades before exploration will confirm or refute my opinions.

In the mid-1960s — before the explosion of interest in sea-floor spreading — the Canadian Society of Petroleum Geologists invited Dr. Robert S. Dietz to Calgary to expand on his famous 1961 article in Nature on continental drift. He presented a thoroughly revolutionary dissertation to an audience of about 300 which was followed by a lively and controversial question period. Toward the end of the meeting the chairman asked for a show of hands to determine the number who did or did not 'believe' Dietz. About 275 hands went up to oppose him accompanied by loud cheers. The few who agreed with him raised their hands to the polite ridicule of the other scientists. Here was a case of science by vote. Since then Dietz's theory has become almost universally accepted and a vote today would certainly conclude that the majority then was wrong. I notice that a kind of running score is also being kept in this volume (Johnson & Srivastava).

A central theme in the Nares Strait symposium and one exemplified by several papers in this volume is the thesis that the tectonic evolution of Nares Strait is closely interrelated with that of Baffin Bay and Labrador Sea (Fig. 1). Hence the Nares Strait debate has very important implications for the interpretation of the history and age of the thick sedimentary sections known to occur in the offshore areas between Greenland and Canada and in the seaways around southern Nares Strait. The Strait itself is also a possible site for thick sedimentary accumulations, and it links the sedimentary basins of the Baffin Bay area to those of the Arctic Ocean. It is one such basin at the northern extremity of Nares Strait — in the Lincoln Sea — that I shall focus on in this assessment, reflecting that any forecasts on the hydrocarbon generation of the sediments in this basin also probably are valid for the basins in the southern extremity of Nares Strait in northern Baffin Bay.

Nares Strait and Lincoln Basin

The Lincoln Basin is a large volume of mainly non-magnetic rocks situated in the Lincoln Sea (Fig. 2). Some of the basin's characteristics can be deduced from the data supplied by Sobczak & Stephens (1974) and

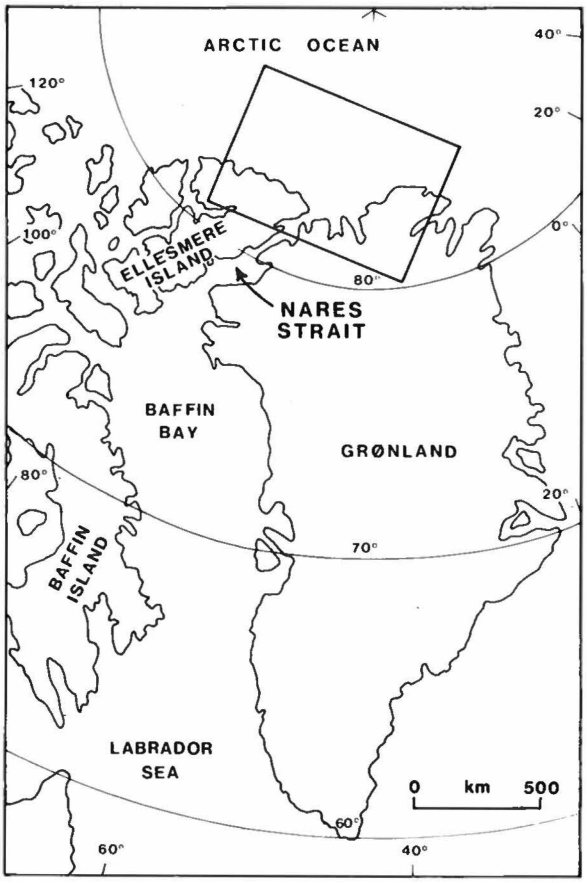


Fig. 1. Index map of the Nares Strait – Labrador Sea region; the frame indicates the area shown in Fig. 2.

Kovacs (this volume). It is an attractive basin for petroleum accumulation because of its great areal extent and depth. It is unattractive for exploration drilling with present technology because of the permanent shifting ice cover. Magnetometer and gravity surveys in the basin combined with field mapping on Ellesmere Island and Greenland allow projections and extrapolations to be made on the nature of the sediments in the basin. These projections and extrapolations are affected profoundly by the choice made in the amount of left-lateral displacement along the Nares Strait lineament. If there was practically no movement (0–25 km) the Lincoln Basin may have been a fairly quiet shelf tectonically and underlain by a stratigraphic section representing all parts of Phanerozoic time. If there has been 250 km of strike-slip movement since Campanian time the shelf may be underlain by mostly Cretaceous and Cenozoic clastics, resting on a basement of Palaeozoics injected by Mesozoic and possibly Cenozoic intrusives. Accordingly a different palaeoheat regime for each of the two possibilities needs to be envisaged when considering the petroleum potential.

To clarify and summarise some aspects of the problems in deciding how attractive the Lincoln Basin is for the occurrence, migration, and entrapment of petroleum a site for the theoretical Lincoln No. 1 well has been selected. This location appears to be on an arch; however, it could just as likely be on a horst within the basin. Case No. 1 (Fig. 3) portrays the situation where a Cretaceous–Cenozoic section rests on Palaeozoic and older continental crust which has been thinned by attenuation during large movement along Nares Strait. Case No. 2 (Fig. 4) portrays the situation that may exist if there were little or no displacement. In this case a Cenozoic–Mesozoic column may rest on a thick non-magnetic Palaeozoic section that in turn rests on magnetic 'granitic' crust. The continental crust in this case has its original thickness and there has been very little disruption or attenuation in the area caused by substantial movement along Nares Strait.

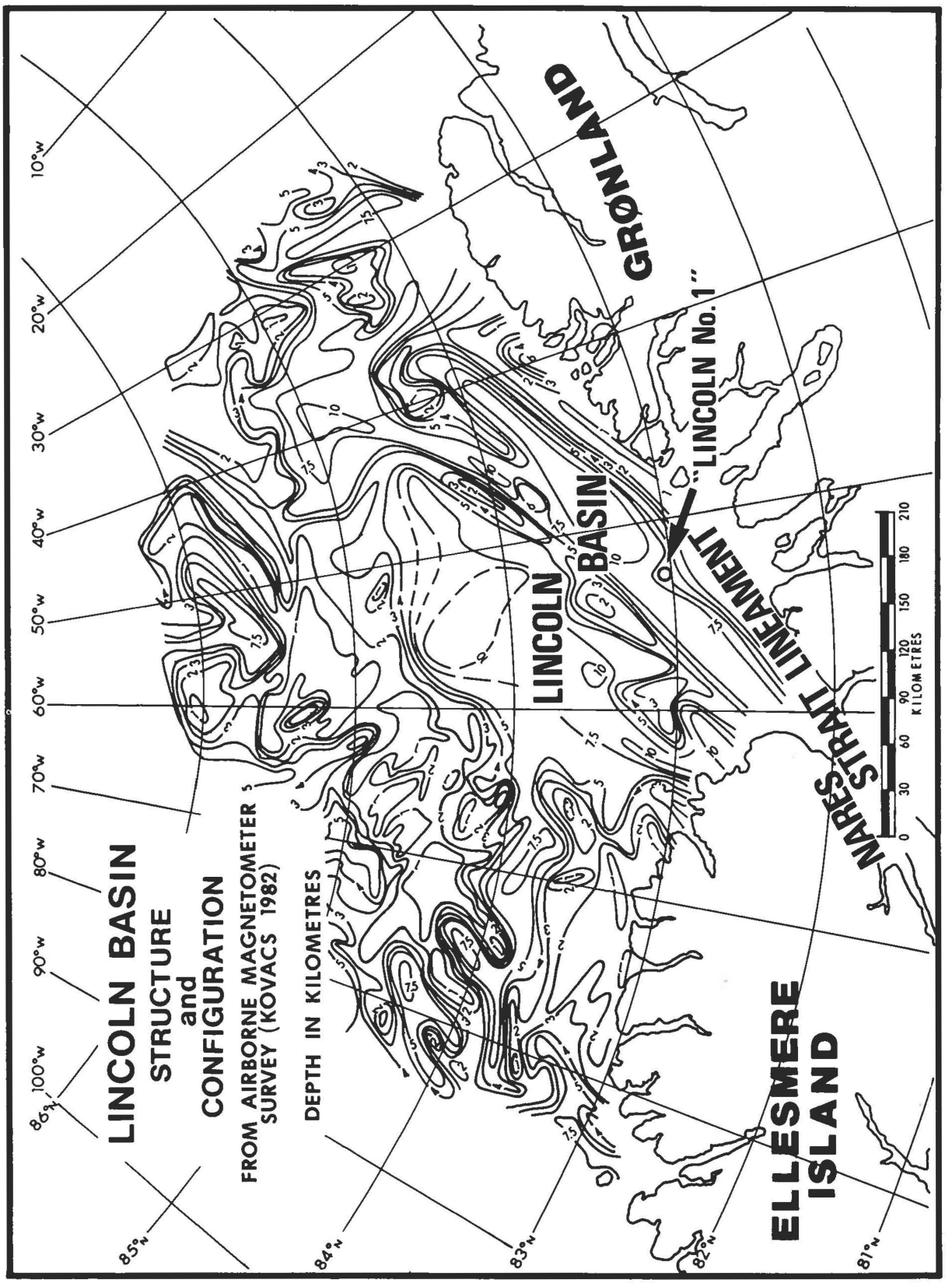
Case No. 1

The overview of the theoretical section of Fig. 3 assumes that great displacement occurred along Nares Strait accompanied by attenuation. It embraces many of the concepts, but not necessarily the timing, offered by Kovacs (this volume). A deep Cretaceous graben-like basin formed between Greenland and Ellesmere Island in a wide Nares Strait. The graben and the sediments partly filling it were moved northeastward until the end of Eocene time. Some basic volcanic rocks would be incorporated in the mainly sand-shale sequence. The continental crust may have been thinned and in places may be practically absent with oceanic crustal materials in its place as judged from the high gravity anomalies (Sobczak & Stephens 1974). Such a situation would be accompanied by high heat flow ranging between 1.60 and 1.80 units (1 unit = 1μcal/cm² per sec.) until at least the end of the Paleocene after which it diminished to 1.40 units by the end of the Eocene. The present-day thermal gradient is estimated to be 30°C/km. The sea-bottom temperatures used on Fig. 3 are meant to conform with the approximations of palaeoclimate given by Clark (1981).

Case No. 2

The hypothetical overview portrayed on Fig. 4 is adopted to contrast with Case No. 1. The assumption is that the continental crust is thick and has not been thinned tectonically by left-lateral movement along Nares Strait. Further, the Lincoln No. 1 site is north of the main Palaeozoic folding as known on northern Green-

Fig. 2. Structure and configuration of the Lincoln Basin based on the aeromagnetic data reported on in this volume by Kovacs.


LINCOLN BASIN
STRUCTURE
and
CONFIGURATION
FROM AIRBORNE MAGNETOMETER SURVEY (KOVACS 1982)
DEPTH IN KILOMETRES
LINCOLN BASIN
GRØNLAND
"LINCOLN No. 1"
NARES STRAIT LINEAMENT
ELLESMERE ISLAND
KILOMETRES
0
30
60
90
120
150
180
210
10°W
20°W
30°W
40°W
50°W
60°W
70°W
80°W
90°W
100°W
85°N
84°N
83°N
82°N
81°N

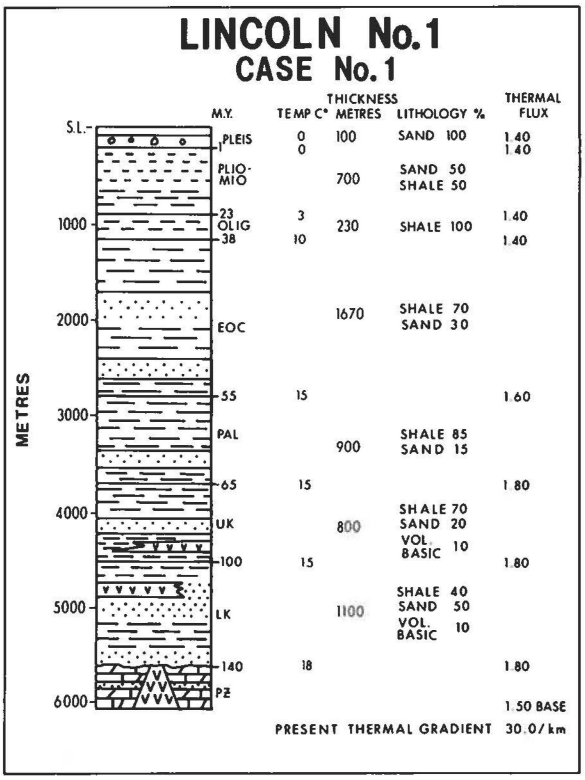


Fig. 3. Postulated sedimentary section, composition, thickness, temperature of the sediments at the time of sedimentation and the thermal flux in μcal/cm ²/sec. The assumption made is that Lower Cretaceous rocks were deposited in a graben in thinned continental crust during large-scale movement along the Nares Strait lineament.

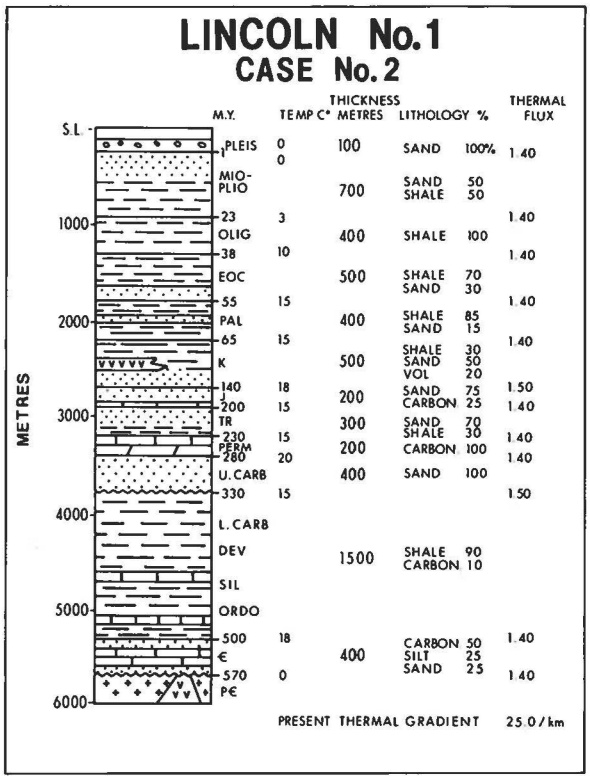


Fig. 4. Postulated sedimentary section, composition, thickness, temperature of the sediments at the time of sedimentation and the thermal flux in μcal/cm ²/sec. The assumption made is that except for a brief erosional interval in the Carboniferous, sedimentation was fairly continuous. The continental crust was thick and there was very little disruption because of little or no movement along the Nares Strait lineament.

land. The Lincoln Basin is a downwarp on the craton which preserved most of the sediments deposited since the Precambrian. One unconformity of short duration (5 m.y.) is recorded in the Carboniferous. Volcanics may be preserved in the Cretaceous sediments. The source of the high gravity anomalies mapped by Sobczak & Stephens (1974) is unknown. No substantial movement is considered to have taken place along Nares Strait. Such a situation perhaps was accompanied by average heat flow (1.40 units) for most of Phanerozoic time. Exceptions may be during part of the Carboniferous and possibly in the early Mesozoic when there were tectonic readjustments or regional reheating. The present-day thermal gradient for Case No. 2 is estimated to be 25°C/km. As with Case No. 1 the sea-bottom temperatures chosen are meant to describe the palaeoclimate described by Clark (1981). The lithology of the Palaeozoic column is considered to be similar to that recorded by Dawes & Peel (1981) and summarised by Churkin & Trexler (1981: fig. 3, column 33).

Diagenetic evolution of the organic matter in the postulated sedimentary sections

The procedure in this hypothetical situation in an attempt to evaluate the ability of the sediments to yield hydrocarbons is one devised by du Rouchet (1980). The method is found to be useful in Canada as well as elsewhere. It can be applied in well-established petroleum producing areas as well as frontier areas.

The method employs a computer programme based on simulations of the depositional history, geothermal history, and the kinetic behaviour of the organic matter decompositions contained in the sediments. The method helps to define the times at which oil and gas were formed throughout the history of any particular basin and it describes the present situation. The reader is referred to du Rouchet's (1980) description for details.

In these specific cases it has been assumed that the

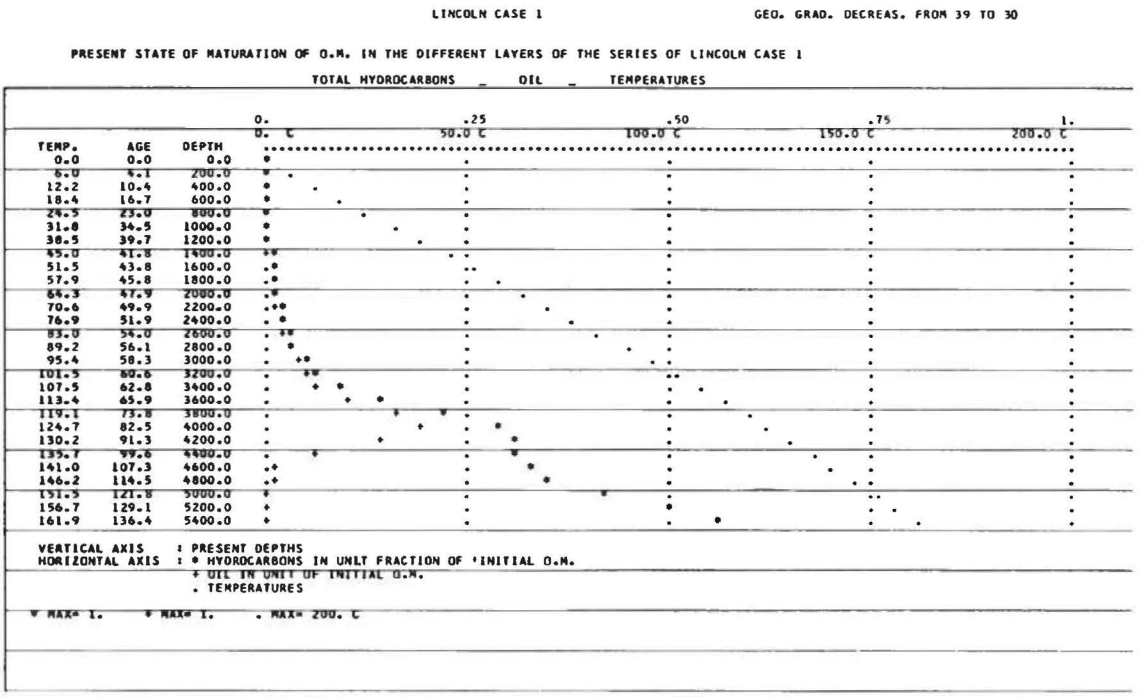


Fig. 5. Present state of maturation of Type II organic matter contained in rocks of Fig. 3. Estimates of amounts of oil and total hydrocarbons plotted and related to present depth, age and temperature.

organic matter is Type II, i.e. a mixture of terrigenous and algal-marine derived organic matter. This assumption seems compatible with the fact that from the Cambrian to the Present Type II organic matter is the most widely known oil source-rock in the geological series all over the world.

Present state for Case No. 1

The present state of the organic matter is recorded on Fig. 5 which is a reproduction of part of the computer print-out. This print-out supposes the presence of one per cent of Type II organic matter in each level of the series. It records oil and gas production from this organic matter for each burial depth. The present temperature and age of deposition are also given (refer also to Fig. 3).

Both oil and gas are being released now from sediments of the Paleocene at the 3000 metre level at 95°C with an amount of organic matter being converted to hydrocarbon of approximately 5 per cent. Maximum formation of oil is at the 4000 metre level at 125°C (in the Upper Cretaceous) where the total amount of the organic matter being converted approaches 30 per cent (2/3 of that amount being oil (20 per cent)). At 150°C (5000 metres) gas is being produced both directly from decomposition of organic matter and from cracking of oil — almost 40 per cent of the organic matter being converted.

Neglecting substantial upward migration of the oil and gas through fracture systems it is reasonable to suggest that any suitable reservoirs in the Paleocene and Upper Cretaceous may contain oil or gas. The Lower Cretaceous would probably have a tendency to contain gas (refer also to Fig. 3).

Present state for Case No. 2

The present state for Case No. 2 is shown on the reproduction of part of the computer print-out contained in Fig. 6. These results should be referred to Fig. 4.

The 95°C temperature level occurs at 3800 metres in this case and the 125°C temperature level is at 5000 metres. At 5100 metres (the deepest simulated in this programme) the amount of oil being released from the organic matter begins decreasing and the amount of gas is increasing. Hydrocarbons probably are being released from all of the lower section to 5700 metres (see Fig. 4.) but the temperature at the Precambrian unconformity likely will be less than 150°C. The fractions of organic matter being converted to hydrocarbons are the same as they are in Case No. 1.

Generated hydrocarbons will be restricted to any reservoirs that are present in the Lower Palaeozoic section.

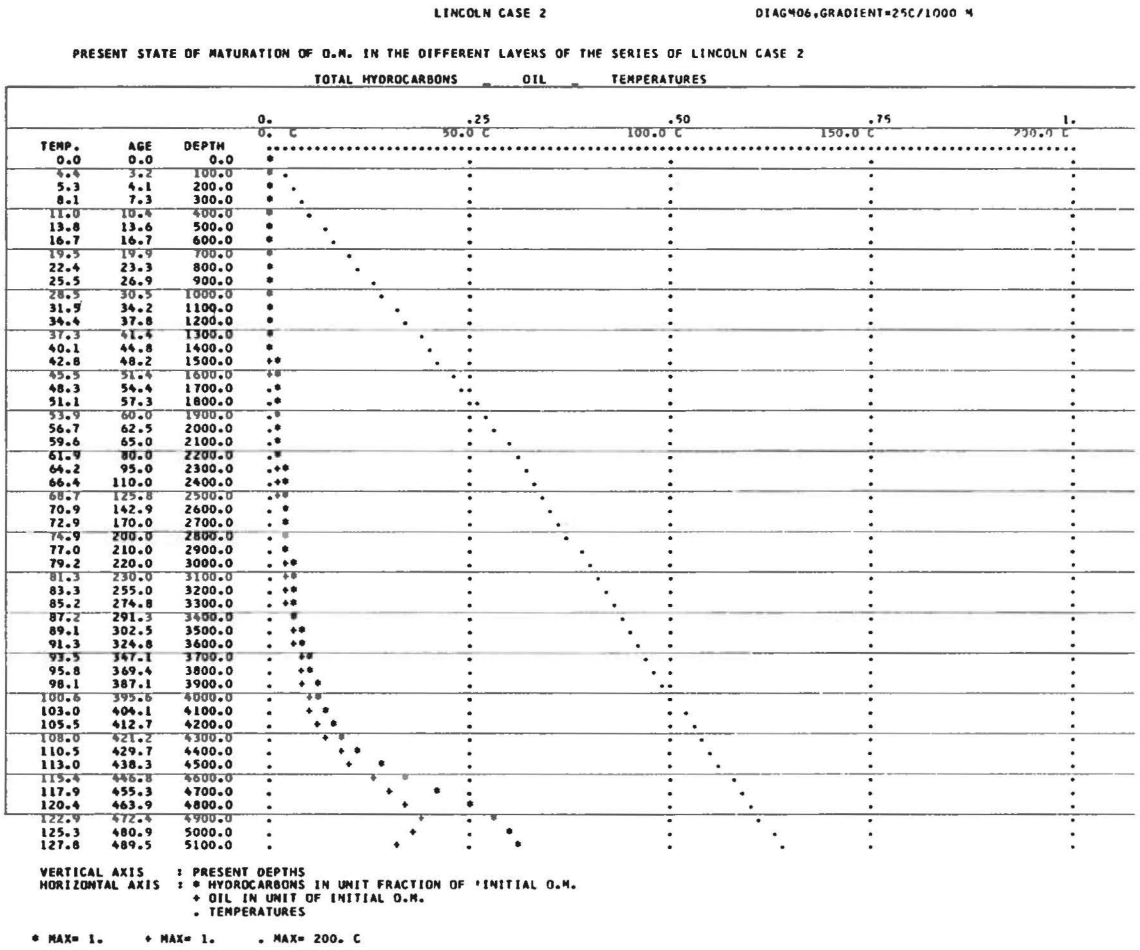


Fig. 6. Present state of maturation of Type II organic matter contained in rocks of Fig. 4. Estimates of amounts of oil and total hydrocarbons plotted and related to present depth, age and temperature.

Summary of the present state and history

The organic matter in the entire section of Case No. 1 from approximately 3000 metres to 5700 metres has been exposed to generating conditions from the past to the present time.

In Case No. 2 favourable conditions have only barely been accomplished and presently extend from 3800 metres to 5700 metres.

Conclusions

There are great hazards in drawing even tentative conclusions about the petroleum potential of the area under the two alternative sets of circumstances suggested. One can make only a guess as to the type of organic matter. The total amount of organic matter surely would not be equal in each of the cases. Great pitfalls loom when estimates are made for the palaeotemperature of deposition of sediments. The same forebodings can be feared when the kinds of sediments that are postulated to have been deposited are chosen. The estimate of the palaeoheat regime as well as the present thermal gradient are open to debate.

Nevertheless petroleum prospectors must make these kinds of choices, and be bold about it when comparing and contrasting widely separated basins in the world. It is clear that Case No. 1 provides the best situation for the generation of hydrocarbons. Furthermore, in Case No. 1 prospecting can begin almost 1000 metres shallower than in Case No. 2.

Therefore, a tectonic regime involving a left-lateral movement of 250 km along the Nares Strait lineament as envisaged by Kovacs (this volume) will provide favourable conditions for petroleum generation in the Lincoln Basin. The explanation is that formation of grabens and thinning of the continental crust provide a higher heat regime. If there has been little or no displacement along the Nares Strait lineament, there may

not have been graben formation and crustal thinning to cause the favourable heat regime for oil generation in Lincoln Basin.

The conclusion that has been made here concerning the Lincoln Basin may apply equally to the situation in the northern part of Baffin Bay.

Acknowledgements

The comparisons could not have been made without the help and patience of M. Jean du Rouchet of Société Nationale Elf Aquitaine, Pau, France. The procedure and programme which provided the conclusions belong to him. I thank him for his generosity. I am also grateful for fruitful discussions with Peter R. Dawes and J. W. Kerr, but take responsibility myself for all conclusions. I also am grateful for the assistance given by drafting personnel, technicians and secretaries of Canterra Energy Ltd., Calgary.

References

Churkin, M. & Trexler, J. H. 1981. Continental plates and accreted oceanic terranes in the Arctic. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 1–20. – Plenum Press, New York & London.

Clark, D. L. 1981. Geology and geophysics of the Amerasian basin. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 599–634. – Plenum Press, New York & London.

Dawes, P. R. & Peel, J. S. 1981. The northern margin of Greenland from Baffin Bay to the Greenland Sea. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 201–264. – Plenum Press, New York & London.

Dietz, R. S. 1961. Continent and ocean basin evolution by spreading of the sea floor. – Nature, Lond. 190: 854–857.

Du Rouchet, J. 1980. Le programme diagen, deux procedures pour apprecier l'evolution chimique de la matiere organique. – Bull. Cent. Rech. Explor. Prod. Elf Aquitaine 4: 813–831.

Johnson, G. L. & Srivastava, S. P. 1982. The case for major displacement along Nares Strait. – This volume.

Kovacs, L. C. 1982. Motion along Nares Strait recorded in the Lincoln Sea: aeromagnetic evidence. – This volume.

Sobczak, L. W. & Stephens, L. E. 1974. The gravity field of northeastern Ellesmere Island, part of northern Greenland and Lincoln Sea with map. Lincoln Sea. 1: 500 000. – Earth Physics Branch, Gravity Map Series 114: 9 pp.

Summary

The case for major displacement along Nares Strait

G. LEONARD JOHNSON and SHIRI P. SRIVASTAVA

Johnson, G. L. & Srivastava, S. P. 1982. The case for major displacement along Nares Strait. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 365–368.

Nares Strait is a prominent physiographic lineament between Greenland and Ellesmere Island, Canada. That it has attracted the attention of geologists, geophysicists and geomorphologists is attested to by this volume. This summary is based solely on those articles presented in the volume that suggest that lateral motion has occurred along Nares Strait; the amount of movement indicated for the Strait ranges from 200 to 320 km.

The Nares Strait dilemma is not yet resolved. Additional dedicated geophysical and geological investigations in the region surrounding Nares Strait will have to be undertaken before a satisfactory solution to the problem can be found. Hopefully, this volume will instigate such activities.

G. L. Johnson, Department of the Navy, Office of Naval Research, Arlington, Virginia 22217, U.S.A. and S. P. Srivastava, Department of Energy, Mines and Resources, Geological Survey of Canada, Bedford Institute of Oceanography, Dartmouth, Nova Scotia, Canada, B2Y 4A2.

Nares Strait, the linear seaway between Greenland and Ellesmere Island, has played an important role in continental drift studies of the Arctic ever since Taylor (1910) suggested it was the site of major transcurrent movement. However, land geologists and marine geophysicists have had a different outlook on the origin and significance of Nares Strait for a long time. This is further attested to by the data presented in this volume and by the two summary papers concluding the volume dealing with the cases for and against major movement along Nares Strait (cf. Dawes & Kerr, this volume). The difference in conclusions of the two summary papers is not because the two papers are in any way biased, but they tend to emphasize the differences in the approach to the problem by the two groups. Land geologists who have mapped the rock units on one or both sides of the Strait have maintained the view that Nares Strait is mainly a graben feature and little or no lateral displacement has taken place along it. Marine geophysicists, on the other hand, have regarded the Strait as a convenient lineament along which to slip the Greenland continental block to explain the evolution of the seas surrounding Greenland (e.g. Arctic Ocean, North Atlantic, Labrador Sea – Baffin Bay, Fig. 1). We agree that the geology of the land surrounding the Strait offers certain constraints for invoking lateral motion along it, but because the structural and lithological boundaries across it lack definite continuities different models involving movement up to 320 km of lateral displacement along it have been proposed by various workers.

The appropriate papers are summarized below in tabular form. Data outside those directly applicable to the problem of whether motion and how much motion occurred along Nares Strait are omitted from this summary. The selection is from a purely objective evaluation and apologies are due to any author who may feel slighted.

Authors	*Amount of displacement*	*Evidence*
Monahan, D. & Johnson, G. L.	240 km	1. Presence of a deep linear trough must be explained. 2. Land geologists can accept 50–100 km offset of marker beds, an additional 60–70 km can be accommodated by geometric uncertainties. 3. The remaining kilometres may be accommodated by post-Palaeozoic expansion of the Northwest Territories platelet by crustal thinning in Lancaster Sound and Jones Sound and in the Sverdrup Basin.

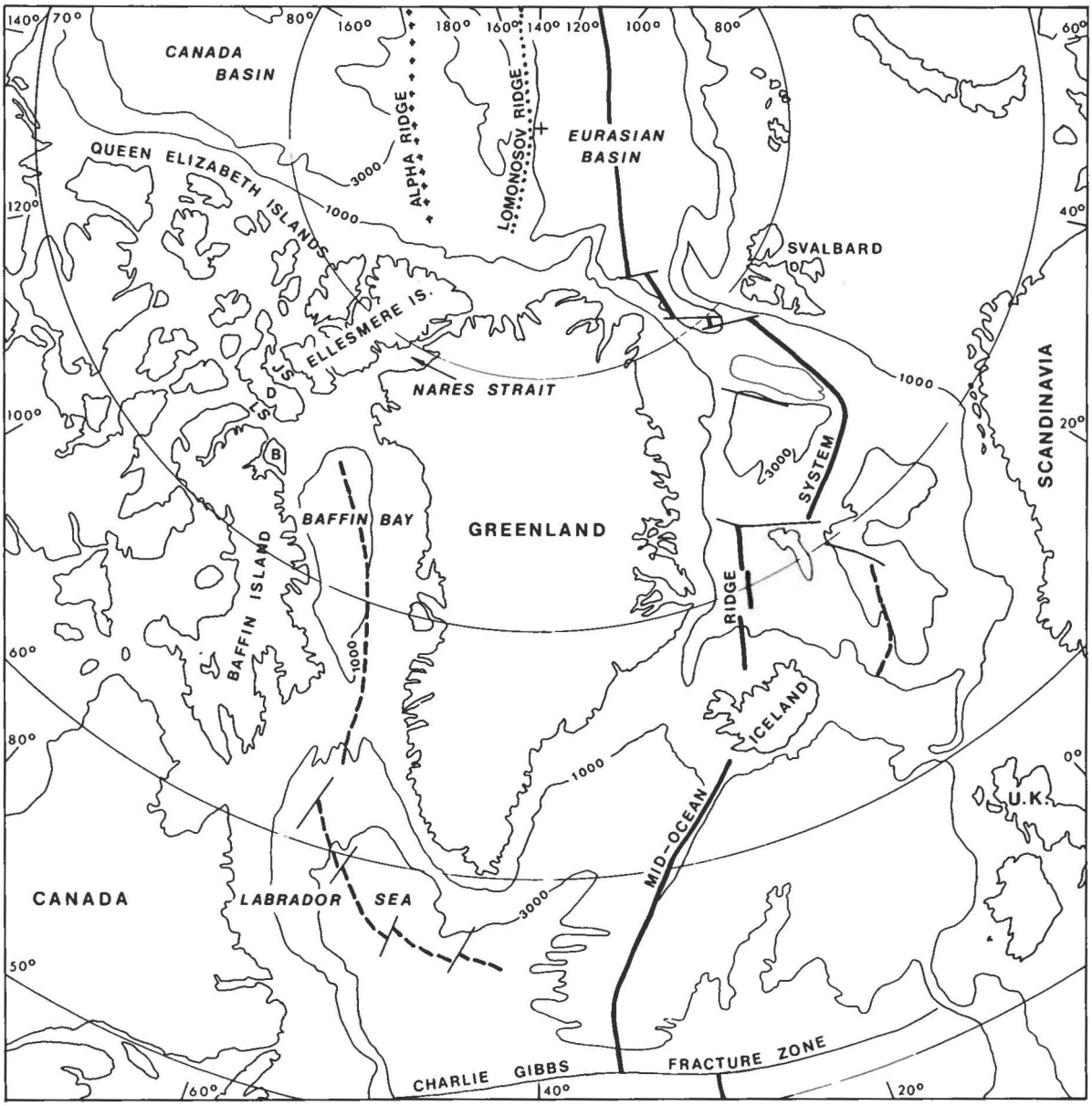


Fig. 1. Index map showing the setting of Nares Strait in the North Atlantic – Arctic Ocean system. J = Judge Daly Promontory, JS = Jones Sound, D = Devon Island, LS = Lancaster Sound, B = Bylot Island. Heavy full lines = active spreading centres, dashed lines = extinct spreading centres, dotted line = ridge of continental origin, line of crosses = ridge of uncertain origin, thin full lines = fracture zones. Bathymetry is in metres.

Authors	*Amount of displacement*	*Evidence*
Mayr, U. & de Vries, C. D. S.	not determined	Late Eocene to early Miocene sinistral strike-slip movement of 19 km parallel to Nares Strait is suggested from the Judge Daly fault zone on the western landward boundary (Ellesmere Island).
Peirce, J. W.	320 km	1. Stratigraphic evidence on Bylot Island of a mid-Cretaceous and younger graben. First phase of the

Authors	*Amount of displacement*	*Evidence*
		Eurekan orogeny created by counter-clockwise rotation of Greenland and Ellesmere Island resulting in crustal shortening of the Sverdrup Basin. 2. Sea-floor spreading in Baffin Bay caused Greenland to move left-laterally some 320 km; during the mid-Oligocene Greenland and Ellesmere Island moved northwest 40–50 km (main phase Eurekan orogeny).
Newman, P. H. (a & b)	200–300 km	1. Geophysical measurements near the southern end of Nares Strait indicate it to be a fault bounded feature, filled with sediments to the north and basalt to the south. 2. Correlation of the structural boundary between deformed and undeformed rocks at the margin of the mid-Palaeozoic fold belt of northern Ellesmere Island and Greenland. 3. Correlation of Proterozoic successions and asymmetrical graben systems of Bylot Island – northern Baffin Island and North-West Greenland. 4. Nares Strait may be one arm of a triple junction (with Baffin Bay and Lancaster Sound) along which transform motion occurred.
Menzies, A. W.	about 240 km	Geophysical observations in the Labrador Sea and Baffin Bay show that these regions were formed by sea-floor spreading, thus requiring large strike-slip movement along Nares Strait, unless the Canadian Arctic Islands are assumed to be a platelet.
Kovacs, L. C.	250 km	1. Depth to magnetic source indicates that Nares Strait extends ~325 km as a deep structural lineament parallel to the northern coast of Greenland. 2. Offset of basement highs with characteristic magnetic signatures on either side and perpendicular to Nares Strait (Wegener Fault) indicates a 250 km movement during the Tertiary opening of Baffin Bay.
Keen, C. E. & Peirce, J. W.	about 250 km	1. Crustal extension, mantle upwelling and magmatic intrusion, considered as mechanisms for the formation of Baffin Bay, cannot satisfactorily match the available data from the Bay. Thus Baffin Bay formed by conventional sea-floor spreading requiring large strike-slip movement along Nares Strait. 2. If no more than 25 km of sinistral strike-slip movement has occurred along Nares Strait, possibilities which need serious consideration are: a) Wegener Fault is a zone, not a single feature and therefore motion may be taken up over a larger area. b) Lateral motion might be accommodated by extension of continental lithosphere in northern Baffin Bay and/or by compression within the Sverdrup Basin.

Author	*Amount of displacement*	*Evidence*
Srivastava, S. P. & Falconer, R. K. H.	about 250 km	1. Substantial movement required by kinematics of North Atlantic as a whole. 2. Alternative is that motion was accommodated by folding and faulting within the Northwest Territories and/or shear within Greenland. However, evidence of events of this magnitude is lacking for the present.

Conclusion

Nares Strait is a prominent physiographic lineament between Greenland and Ellesmere Island which has attracted the attention of many geologists, geomorphologists and geophysicists. To the reader of this volume it will be quickly apparent that there is a considerable diversity of opinion as to the amount of movement, if any, on Nares Strait, from none at all to as large as 320 km. This is because of the criteria used in deciphering motion or no motion along this lineament. When one looks at a small region of the Strait, either geologically or geophysically, the answer invariably comes out that little or no lateral motion has taken place in this region. On the other hand, significant lateral motions are suggested along the Strait when one considers large-scale tectonic developments of the surrounding regions (e.g. papers by Monahan & Johnson, Peirce, Menzies, Srivastava & Falconer).

In general, the mobilists adhere to the concept of sea-floor spreading in which large parts of the present oceanic regions were formed due to the separation of the plates (Fig. 1). The second tenet, which is not so universally adhered to, is that the plates behave in a rigid manner. Non-rigidity within the Greenland or northern part of the North American plates would help to alleviate the problem of large lateral motion along Nares Strait, but it may not be acceptable to all land geologists. We hope the number of articles presented in this volume will give readers a chance to assess what is known of the Nares Strait region and the problems one faces in trying to explain the evolution of the oceanic regions surrounding Greenland and at the same time maintaining the continuity of the geological boundaries across Nares Strait. Ideally, this volume will lead to a more rigorous examination of geology and geophysics surrounding Nares Strait by all earth scientists to decide what additional work is needed to solve the Nares Strait enigma.

References

Dawes, P. R. & Kerr, J. W. 1982. The case against major displacement along Nares Strait. – This volume.

Keen, C. E. & Peirce, J. W. 1982. The geophysical implications of minimal Tertiary motion along Nares Strait. – This volume.

Kovacs, L. C. 1982. Motion along Nares Strait recorded in the Lincoln Sea: aeromagnetic evidence. – This volume.

Mayr, U. & de Vries, C. D. S. 1982. Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago. – This volume.

Menzies, A. W. 1982. Crustal history and basin development of Baffin Bay. – This volume.

Monahan, D. & Johnson, G. L. 1982. Physiography of Nares Strait: importance to the origin of the Wegener Fault. – This volume.

Newman, P. H. 1982a. A geological case for movement between Canada and Greenland along Nares Strait. – This volume.

Newman, P. H. 1982b. Marine geophysical study of southern Nares Strait. – This volume.

Peirce, J. W. 1982. The evolution of the Nares Strait lineament and its relation to the Eurekan orogeny. – This volume.

Srivastava, S. P. & Falconer, R. K. H. 1982. Nares Strait: a conflict between plate tectonic predictions and geological interpretation. – This volume.

Taylor, F. B. 1910. Bearing of the Tertiary mountain belt on the origin of the earth's plan. – Bull. geol. Soc. Am. 21: 179–226.

The case against major displacement along Nares Strait

PETER R. DAWES and J. WILLIAM KERR

Dawes, P. R. & Kerr, J. W. 1982. The case against major displacement along Nares Strait. – In: Dawes, P. R. & Kerr, J. W. (eds), Nares Strait and the drift of Greenland: a conflict in plate tectonics. – Meddr Grønland, Geosci. 8: 369–386.

Nares Strait is a conspicuous lineament that suggests it is the site of a fault. The conflict about magnitude of displacement has developed because the interpretation of geological data from the immediate Nares Strait region is at variance with the common interpretations of the geophysical data from the surrounding oceans, which predict Nares Strait as the site of great displacement.

We summarise here the geological and geophysical data presented in this volume that indicate little or no strike-slip, vertical or oblique displacement along Nares Strait, and the conclusion is based on five categories of evidence:

1) there is present-day continuity of geological features from Ellesmere Island to Greenland,
2) the Strait is aseismic, suggesting that it is not a major crustal or lithospheric fracture,
3) a pre-drift reconstruction showing large dextral strike-slip displacement (e.g. 250 km) introduces unacceptable juxtaposing of the Precambrian and Palaeozoic rock provinces,
4) a pre-drift reconstruction showing Greenland and Ellesmere Island separated obliquely by a wide gap is refuted by the geology of the Strait,
5) there are no unequivocal geological or geophysical data from the immediate Strait area, which suggest that Greenland and Ellesmere Island had an initial relative position radically different from present-day geography.

On the basis of 9 regional stratigraphic and structural features and by the definition of 20 geological–geophysical markers that can be traced without apparent offset from Ellesmere Island to Greenland, we conclude that any net left-lateral, strike-slip motion along Nares Strait has been minor — in the range of 0 to 25 km. This argues strongly against the hitherto widely accepted *'pure'* sea-floor spreading origin of Baffin Bay and the Labrador Sea and suggests that other mechanisms must have played an important part in the formation of these oceanic basins.

P. R. Dawes, Grønlands Geologiske Undersøgelse, Øster Voldgade 10, DK–1350 Copenhagen K, Denmark and J. W. Kerr, Department of Energy, Mines and Resources, Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303–33rd St. N.W., Calgary, Alberta, Canada, T2L 2A7. J. W. K. present address: J. William Kerr & Associates Ltd., Consulting Geologists, Suite 407, 630–8th Ave. S.W., Calgary, Alberta, Canada, T2P 1G6.

The Nares Strait seaway separating Greenland and Ellesmere Island, Canada, shows remarkable linearity over several hundred kilometres (Fig. 1), suggesting it to be the site of a fault (Taylor 1910). Broadly speaking, two very different interpretations have been made of the tectonic history of Nares Strait; one mainly from the geological evidence from lands bordering the Strait and the other from the geophysical evidence collected elsewhere from the surrounding seas and even globally. The geological evidence (supported by geophysical data from Ellesmere Island and Greenland) indicates that the net result of any strike-slip movement along the Strait is minor. Alternatively, the geophysical data from the surrounding oceans (Arctic Ocean, Labrador Sea and Baffin Bay), which rely heavily on the theory of sea-floor spreading, demand much greater displacement, consistently in a left-lateral (sinistral) sense, but varying from about 200 to 400 km. This conflict, which has been debated for many years and summarised by Kerr (1980), is continued by the papers in this volume. In general, the positions of workers have not changed with geological evidence in the Strait being used to support minor displacement and offshore geophysics being used to support great displacement. There are of course exceptions to the general nature of the conflict as outlined above; hence this volume also contains geologically based papers that conclude a case for major displacement (e.g. Newman a) in addition to a mainly geophysically oriented paper from outside the immediate region that concludes little or no movement (e.g. Grant).

This summary is concerned with assessing the geological and geophysical data from the immediate Nares Strait region in terms of the spatial relationship of

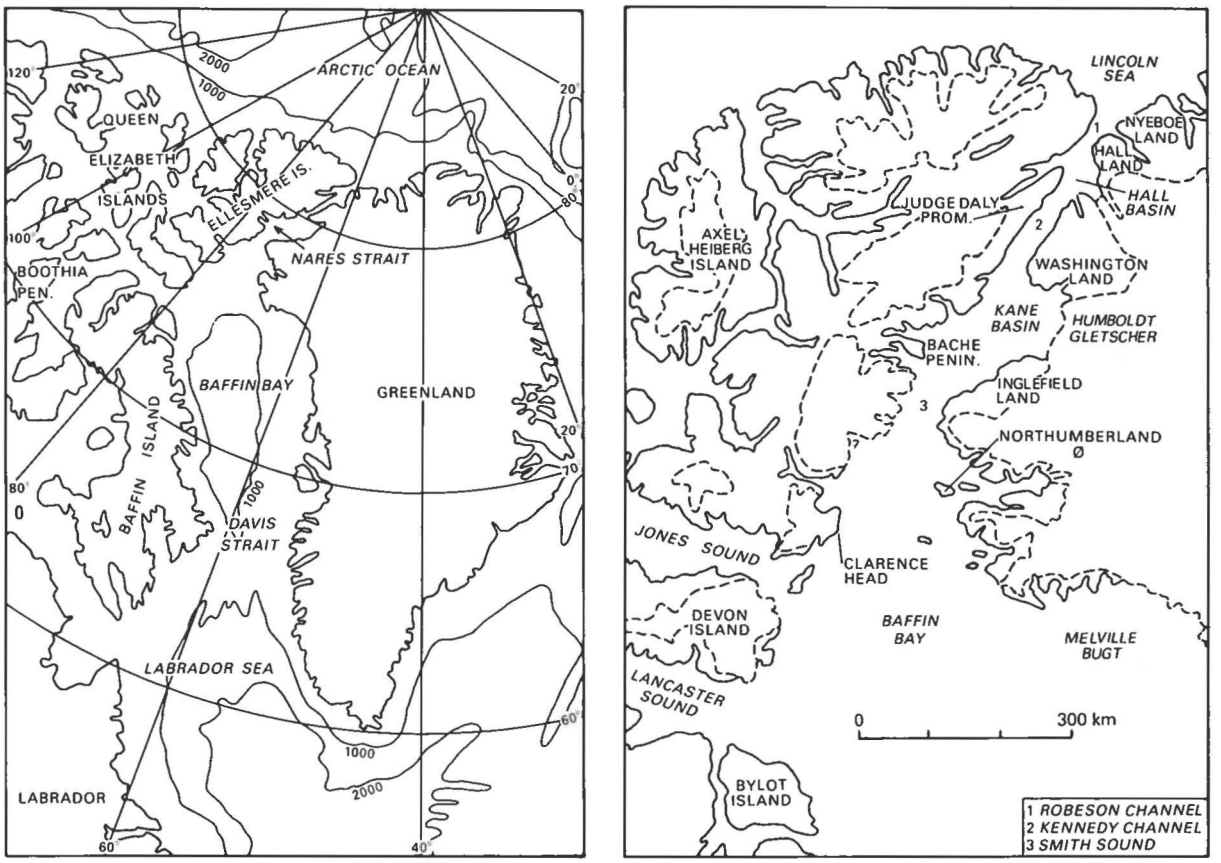


Fig. 1. Index maps of the Nares Strait region. Bathymetry is in metres.

Greenland and Ellesmere Island. We interpret these data as entirely consistent with there having been little or no strike-slip or other displacements along the Strait. We also show that reconstructions with Baffin Bay closed create problems with the Precambrian and Palaeozoic geology which are incompatible with regional stratigraphy and stratigraphic and structural principles.

The main stratigraphic–structural units of the Nares Strait region are illustrated in Fig. 2.

Geological–geophysical markers

The logical way of investigating possible displacement along Nares Strait is to define on Ellesmere Island and Greenland geological boundaries or structures (by both geological and geophysical methods) that are correlatable. Twenty of these geological–geophysical markers can be recognised and these are summarised in Table 1 and illustrated in Fig. 3.

This symposium has made it clear that there is a demand on geologists to use the available geological data to define the amount of sinistral displacement that could have occurred along the Strait. Hence, we have attempted in Table 1 to estimate the maximum net displacement that each marker permits. The markers represent stratigraphical and structural features which have varying reliability as tools for assessing strike-slip motion. Some are precise, well-defined markers, others more poorly defined and imprecise. The accuracy of the estimate of maximum motion that can be determined from any marker depends on a number of factors, the most obvious of which are summarised below.

Definition: the value of a marker depends on how precisely it has been defined by geological and/or geophysical mapping.

Exposure: an exposed marker is of more value than the same marker when in the sub-surface.

Type: steeply-inclined vertical, narrowly-defined, linear bodies or geological contacts give more reliable estimates than shallowly-inclined, broadly or gradationally defined structures or contacts.

Form: continuous and straight markers are more valuable than non-continuous and irregular markers.

Age: accurately dated markers are of more value than markers of uncertain age.

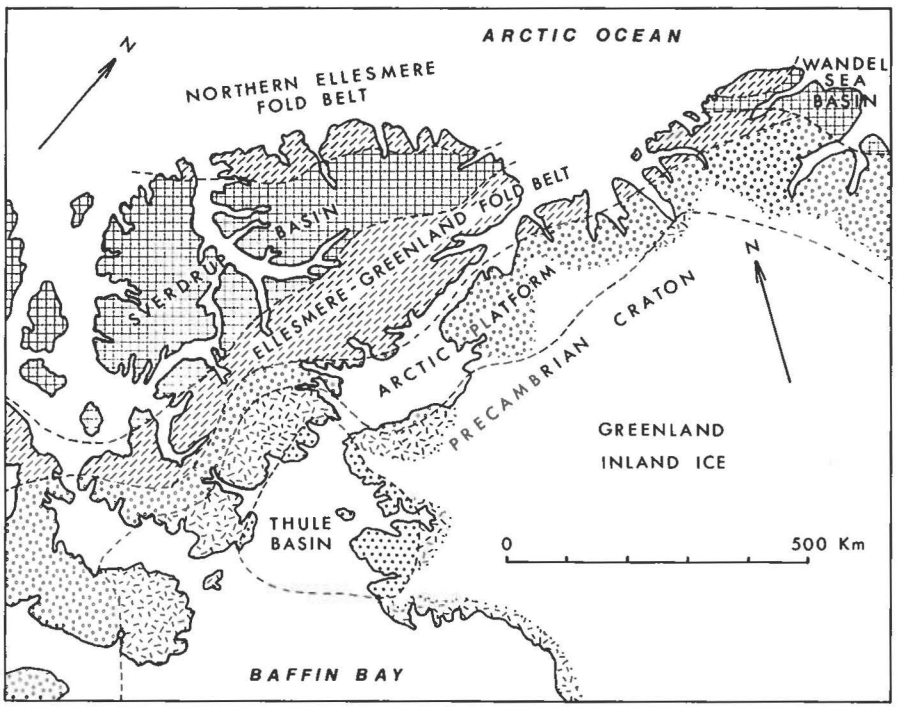


Fig. 2. Main stratigraphic–structural units of the Nares Strait region. The Ellesmere–Greenland and the northern Ellesmere fold belts broadly correspond with the Proterozoic–Devonian Franklinian geosyncline (basin) (cf. Fig. 5B). The northern Ellesmere fold belt includes magmatic rocks — the Pearya geanticline.

Time of formation: markers that existed prior to the supposed displacement are of more value than those which formed during or after displacement.

Angle of intersection: markers transecting the Strait at a high angle of intersection are more valuable than those intersecting at a low angle.

Width of Nares Strait: markers intersecting the Strait at its narrowest parts are of more value than those cutting elsewhere.

The markers recognised in the Nares Strait region vary in reliability. On one extreme are accurately dated, well-exposed, vertical geological facies boundaries which cross the Strait at one of its narrowest parts (e.g. marker 10, Table 1). At the other extreme are broad, magnetic or seismic zones reflecting sub-surface structures of uncertain age and origin that cross the Strait in its widest part (e.g. markers 17 and 20). The restraint on motion imposed by each marker is proportional to the reliability of that marker.

1. Geological continuity between Ellesmere Island and Greenland

The data indicating that geological features are continuous without offset from Ellesmere Island to Greenland are presented in terms of 9 regional geological features (Table 2) and the 20 geological–geophysical markers (Table 1). On the basis of these 29 lines of evidence, there is no reason to envisage that the present-day relative position of Ellesmere Island and Greenland is significantly different from what it was in the past.

In Table 1, restraints in kilometres are given for the most precise markers and these are estimates of the maximum amount of net dextral motion (i.e. restored to an original position) possible from present-day geography. The figures quoted do not suggest that such displacements have occurred. Rather, each figure indicates the maximum displacement that the marker can allow at the present state of definition of that marker. All these markers are dealt with by papers in this volume and the reader is referred in Table 1 to the relevant papers.

Several of the markers, particularly the geophysically defined ones (17 to 20), have not been discussed in the volume in terms of a quantitative estimate of displacement, and no estimate is included for these in Table 1. However, with the possible exception of the linear zone of earthquake epicentres (marker 20 — which reflects a fracture line, the history of which could entirely postdate the supposed displacement), we conclude that all the markers refute major strike-slip movement along Nares Strait of the amount used in many conventional reconstructions (e.g. 200 km or more). All markers, and the regional features given in Table 2, provide a reliable approximate correlation of the position of Greenland and Ellesmere Island. A more exact measure of relative position is provided by the more precise and narrowly defined markers.

Table 1. The geological–geophysical markers that cross Nares Strait without offset

Reference number	Geological–geophysical marker	Estimate of maximum strike-slip displacement possible	Area of intersection with Nares Strait	Reference this volume or otherwise
1	Precambrian (Archaean) marble and marble-rich meta-sedimentary belt intruded by diagnostic meta-igneous rocks.	100 km	Smith Sound	Frisch & Dawes
2	Northern margin of the Thule Basin indicated by rapid thickness variation in the Proterozoic Thule Group.	75 km	Smith Sound	Dawes et al.
3	Bache Peninsula arch; a structural basement high separating the Thule Basin on the south from the Arctic platform on the north.	100 km	Smith Sound	Peel et al.
4	Northern zero-edge (0 isopach) of the Proterozoic Rensselaer Bay Formation (Thule Group).	100 km	Kane Basin	Peel et al.
5	Cambro-Ordovician platform margin; a linear facies boundary separating southern carbonate and northern trough clastics, that is the site of faults.	100 km	Robeson Channel	Dawes
6	Lower Ordovician evaporite belt represented by two distinct stratigraphic levels; Baumann Fiord Formation (Canada) and Poulsen Cliff and Nygaard Bay Formations (Greenland).	50 km	Kane Basin	Peel & Christie
7	Middle Ordovician evaporite belt; Bay Fiord Formation (Canada) and Cape Webster Formation (Greenland).	50 km	Kennedy Channel	Peel & Christie
8	Cambro-Ordovician hinge line, over which there is a marked thickening of deposits northwards.	150 km	Kennedy Channel	Peel & Christie
9	Upper Ordovician – Silurian platform margin associated with a linear horst; a steep facies front between southern carbonates and northern clastics.	100 km	Robeson Channel	Hurst & Kerr a
10	Lower Silurian (middle Llandovery) near vertical abrupt facies contact separating platform carbonates with platform-edge reefs from mudstone.	25 km	Kennedy Channel	Hurst & Kerr a

11	Silurian carbonate buildup ('reef') belt; part of a regional belt traceable through the Queen Elizabeth Islands and across North Greenland.	100 km	Hall Basin and Kennedy Channel	Hurst & Kerr a, Kerr (1967a, this volume)
12	Southern boundary of significant folding (Ellesmerian) of the mid-Palaeozoic Ellesmere–Greenland fold belt.	50 km	Kennedy Channel to Robeson Channel	Higgins et al.
13	Southern limit of amphibolite-facies rocks in the Ellesmere–Greenland fold belt.	100 km	Robeson Channel and Lincoln Sea	Higgins et al.
14	Lake Hazen – Harder Fjord fracture line; a high-angle fault system of probable early Palaeozoic age reactivated in the Tertiary.	150 km	Lincoln Sea	Dawes, Higgins et al.
15	Judge Daly – Nyeboe Land fracture line; a high-angle fault system of probable early Palaeozoic age, reactivated in the Tertiary.	100 km	Robeson Channel and Lincoln Sea	Dawes
16	Judge Daly Promontory – Wulff Land anticlinal zone exposing Proterozoic strata in the margin of the Ellesmere–Greenland fold belt.	100 km	Robeson Channel	Dawes
17	Region of distinctive magnetic character based on vertical field residuals; the southern boundary of Region IV strikes from Judge Daly Promontory to Nyeboe Land.	?	Robeson Channel	Grant, Coles et al. (1976)
18	Kane Basin magnetic anomaly; a distinctive magnetic high at about 79°30'N between Bache Peninsula and Humboldt Gletscher.	?	Kane Basin	Grant, Riddihough et al. (1973) Coles et al. (1976)
19	Prominent steep gravity gradient indicating fundamental crustal change from Lincoln Sea to Greenland; trend lines strike from Judge Daly Promontory to Nyeboe Land.	?	Robeson Channel and Lincoln Sea	Sobczak, Dawes,
20	An anomalous, linear zone of earthquake epicentres along the northern coast region of Ellesmere Island and Greenland.	?	Lincoln Sea	Wetmiller & Forsyth

The markers are arranged generally from the oldest to the youngest and they are located on Fig. 3. The estimates of strike-slip displacement quoted are rounded off values that indicate the maximum amount of net dextral motion from present-day geography which can be reconciled with present knowledge and definition of the markers. All markers are consistent with zero net strike-slip displacement along Nares Strait, although marker 20 depicts a younger fracture line, the history of which could entirely post-date the supposed displacement discussed at the symposium.

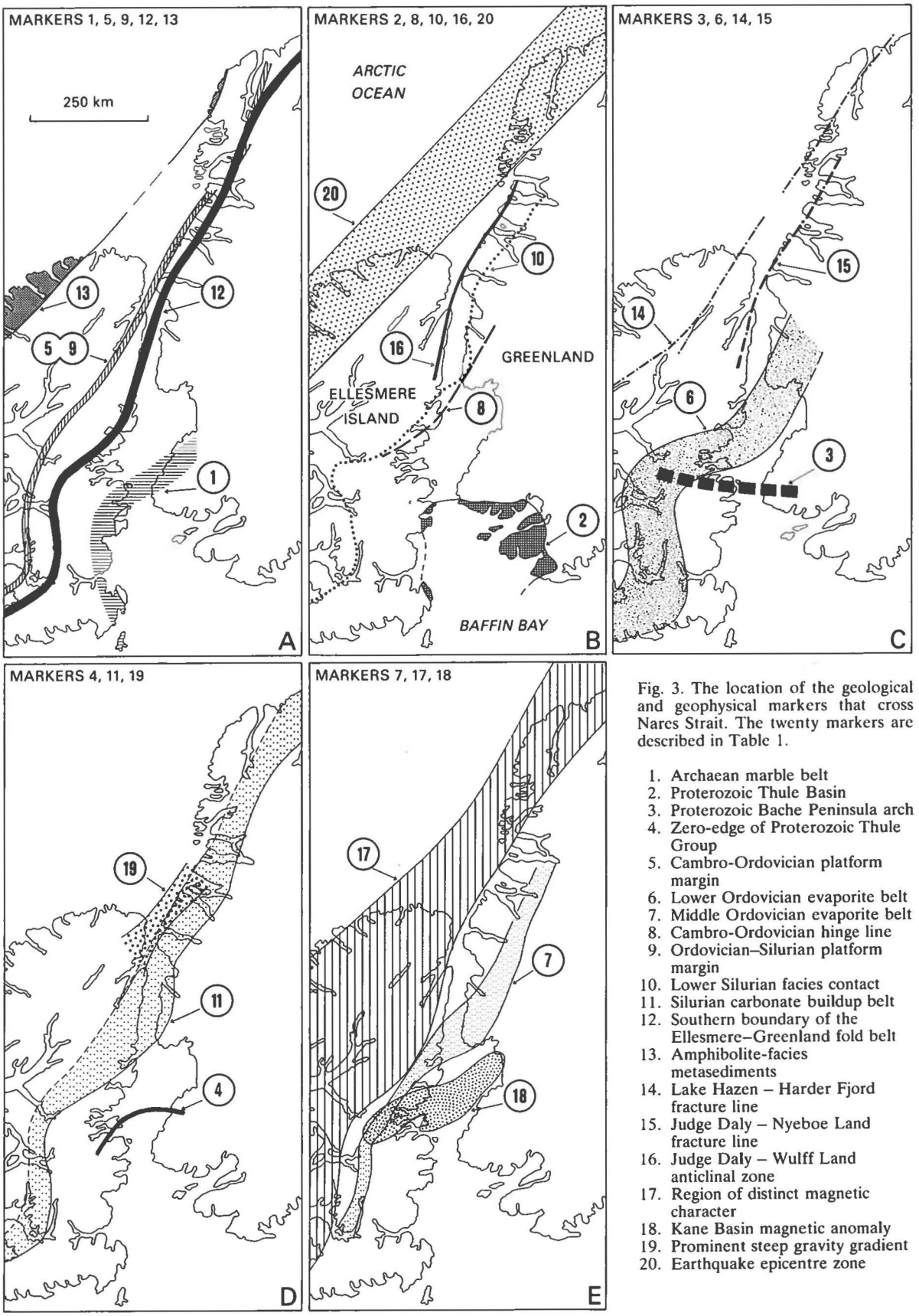


Fig. 3. The location of the geological and geophysical markers that cross Nares Strait. The twenty markers are described in Table 1.

1. Archaean marble belt
2. Proterozoic Thule Basin
3. Proterozoic Bache Peninsula arch
4. Zero-edge of Proterozoic Thule Group
5. Cambro-Ordovician platform margin
6. Lower Ordovician evaporite belt
7. Middle Ordovician evaporite belt
8. Cambro-Ordovician hinge line
9. Ordovician–Silurian platform margin
10. Lower Silurian facies contact
11. Silurian carbonate buildup belt
12. Southern boundary of the Ellesmere–Greenland fold belt
13. Amphibolite-facies metasediments
14. Lake Hazen – Harder Fjord fracture line
15. Judge Daly – Nyeboe Land fracture line
16. Judge Daly – Wulff Land anticlinal zone
17. Region of distinct magnetic character
18. Kane Basin magnetic anomaly
19. Prominent steep gravity gradient
20. Earthquake epicentre zone

Table 2. Main regional features that cross Nares Strait, and their tectonic inference

Reference letter	Type of feature	Tectonic inference	Area of Nares Strait	Reference this volume
A	General distribution and disposition of the Precambrian and Palaeozoic stratigraphic–structural provinces in Ellesmere Island and Greenland.	Ellesmere Island and Greenland developed as adjacent parts of the same plate	Baffin Bay to the Arctic Ocean	This paper Fig. 2
B	Nature and altitude of the Proterozoic erosion surface of the Shield at 79°N.	No major differential vertical movement between Ellesmere Island and Greenland	Kane Basin	This paper Fig. 6, Dawes et al., Peel et al.
C	Proterozoic Thule Group sediments and volcanics; stratigraphy and isopach pattern.	Consistent with little or no Phanerozoic lateral displacement	Northern Baffin Bay and Smith Sound	Dawes et al., Peel et al.
D	Basal Cambrian platform stratigraphy; facies and isopach pattern.	Consistent with little or no post-Cambrian lateral displacement	Smith Sound – Kane Basin	Peel et al.
E	Cambro-Ordovician platform and trough stratigraphy; facies and isopach pattern.	Consistent with little or no post-Ordovician lateral displacement	Kane Basin and Kennedy Channel	Peel & Christie
F	Upper Ordovician and Silurian platform–trough stratigraphy; facies and isopach pattern.	Consistent with little or no post-Silurian lateral displacement	Kennedy Channel to Lincoln Sea	Dawes, Hurst & Kerr a
G	Silurian deep-water basin; turbidite axial flow across North Greenland continued westward along the axis of the Hazen (Ellesmere) trough.	Consistent with little or no post-Silurian lateral displacement	Kennedy Channel to Lincoln Sea	Surlyk
H	Structural and metamorphic pattern of the Ellesmere–Greenland fold belt; zones of fold style, deformation intensity and metamorphic grade.	Consistent with little or no post-Devonian lateral displacement	Kane Basin to Lincoln Sea	Higgins et al.
I	Postglacial emergence zones (isobases) for 8000, 7000 and 6000 B.P.	Glacio-isostatic deformation of the lithosphere has taken place without influence of faulting	Hall Basin and Robeson Channel	England

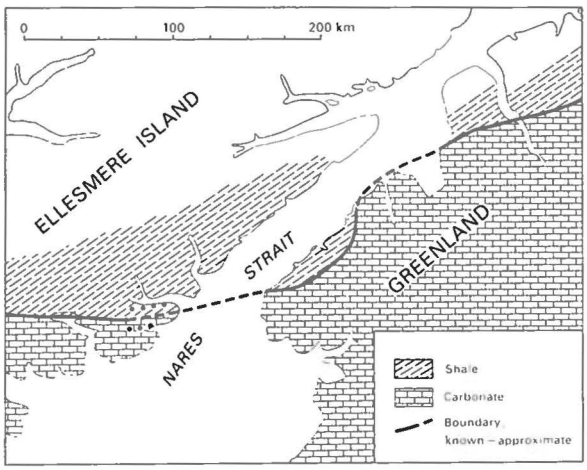


Fig. 4. Sketch showing by solid line the position of the Lower Silurian (middle Llandovery) carbonate-shale facies boundary in the Kennedy Channel area (marker 10 in Table 1). On eastern Ellesmere Island there is a gap in exposure of the actual boundary; here the most likely position is shown by a dashed line and the outer limits of its position are shown by dotted lines.

The most precise of these is the Lower Silurian (middle Llandovery, about 430 m.y.) facies front (marker 10) which is a near vertical contact between two very distinctive rock facies, precisely dated on faunal evidence and crossing Nares Strait at a relatively high angle in one of the narrowest parts of the seaway — Kennedy Channel (see Hurst & Kerr a, this volume). The marker on the Greenland side is precisely located in the high sea-cliffs of Washington Land; on the Ellesmere Island coast it is less well exposed but can be located within a narrow zone on strike across the Strait (Fig. 4). A median line drawn between the two possible extreme positions of the boundary in Ellesmere Island indicates that a net sinistral movement in the range of 0–25 km is compatible with the trend of this marker. This feature can be considered a fine adjustment to the approximate relative positioning of Ellesmere Island and Greenland obtained from other data from the Strait.

The markers and regional geological features summarised in Tables 1 and 2 do not demand any strike-slip motion along Nares Strait. We stress that all are entirely compatible with there having been little or no strike-slip movement along the Strait. Using a combination of all markers, we conclude that any *net* strike-slip displacement has been minor, possibly near zero and not greater than 25 km.

2. Nares Strait as a major lithospheric fracture: the problems

The present-day geological continuity between Ellesmere Island and Greenland discussed above is conclusive evidence that Nares Strait has not had a large net displacement. However, it is not conclusive evidence against movement back and forth, both dextral and sinistral cancelling each other out to give a net displacement of only a short distance or zero. It has been mentioned elsewhere in this volume (e.g. Keen & Peirce) and by Christie et al. (1981) that perhaps a tantalising solution to the Nares Strait debate might be sought in a complex tectonic history in which there has been major dextral displacement along Nares Strait sometime after the mid-Palaeozoic Ellesmerian orogeny, followed by a sinistral movement of Greenland in response to the opening of the surrounding oceans. Such a tectonic history would meet the evidence for present-day geological continuity as well as the fundamental requirements of theories in which Baffin Bay opened by substantial sea-floor spreading and in which Nares Strait acted as a plate boundary.

However, such a speculative scenario is controversial and involves an extreme coincidence; we can find no direct geological evidence in the Nares Strait – Baffin Bay region that suggests major dextral movement of Greenland in the Late Palaeozoic or Mesozoic. On the contrary, three lines of evidence strongly suggest that Nares Strait does not represent a plate boundary that has been involved in a long history of crustal block motion of continental proportions.

Seismic evidence

The boundaries of present-day plates are defined by their seismicity; Nares Strait is aseismic even down to small magnitude levels (Wetmiller & Forsyth, this volume). These authors point out that this lack of seismicity is even more remarkable when compared to areas of seismicity in adjacent Canada and their relationships to known geological structures. Thus, structures like the Boothia Uplift, and structures and arches of the Sverdrup Basin, which have been tectonically active in at least Palaeozoic and/or Mesozoic eras, also show residual seismic expression today. In addition, 'ancient' lineaments of great strike-slip motion in the North Atlantic region, such as for example the Great Glen Fault of Scotland, in which much wrench faulting took place in the late Palaeozoic, also remain seismically active today, indicating a response to the stresses in the evolving North Atlantic. Thus, if Nares Strait had been substantially active for a long period at the same time — and particularly if more recently than such structures, as has been suggested in many plate tectonic sea-floor spreading models — then it would be expected to show at least some sort of seismic activity. That it does not indicates that Nares Strait is not the site of any significant dynamic tectonism and it most likely does not represent a crustal or lithospheric fracture zone that has had a long and complex history.

Magnetic and gravity evidence

Available magnetic and gravity data from Arctic Canada and northern Greenland do not show any indication that Nares Strait represents a fundamental fracture of the crust or lithosphere (see Riddihough et al. 1973, Coles et al. 1976, Sobczak 1978, this volume, Grant, this volume). The major trends of the magnetic field and gravity anomalies can be traced directly across Nares Strait. Furthermore, there are no pronounced magnetic and gravity trends or expressions along the Strait. Indeed, the major magnetic and gravity trends defined by Riddihough et al. (1973) and Sobczak & Stephens (1974) in the northern Nares Strait region are parallel to and coincide with known geological structures which can be traced across Robeson Channel without offset (see Dawes, this volume).

Structural evidence

There is no evidence along the entire shores of Nares Strait from Baffin Bay to Lincoln Sea that indicates a long and complex history of fault movements along the Strait. Major strike-slip faults normally are complex features composed of a zone of dislocation planes, often with splay-out faults and often many kilometres wide. Many, for instance three of the most renowned wrench faults — the San Andreas Fault, the Great Glen Fault and the Alpine Fault of New Zealand — are part of fault systems that far exceed the width of Nares Strait. Despite the remarkably good exposures on the coasts of Nares Strait, no faults, crush zones, shear zones or lineaments parallel to the Strait have been mapped on the Greenland side, and none appear to be present on the islands in the Strait (J. M. Hurst, pers. comm. 1981). Minor shear zones at high angles to the Strait on Greenland (Hurst & Kerr b, this volume) have been considered to be compatible with minor rotation of Greenland. The only dislocations reported in Ellesmere Island are the structures described by Mayr & de Vries (this volume) between Hall Basin and Kane Basin. These authors tentatively suggest that a sinistral strike-slip movement of 19 km has occurred along the Judge Daly fault zone. Christie (1967, 1974) suggested that fault movements in this fault zone were not large-scale movements of the type to be expected when continental blocks have moved in relation to each other. In addition, the Judge Daly fault zone may well be the western part of a regional fracture line which is oblique to Nares Strait; thus any movement along it may not necessarily have resulted in the separation of Ellesmere Island and Greenland as separate plates (Dawes, this volume).

Faults linking Baffin Bay and Lincoln Sea presumably are present offshore in Nares Strait; faults are suggested by geophysical data in southern Nares Strait (Newman b, this volume) and others might be expected to be demarcated as geophysical surveys are carried out farther north. Even so, the presently mapped onshore faults do not support the concept of a complex fracture zone parallel to the Strait, along which there has been major dextral and sinistral movements.

3. The 250 km dextral reconstruction: the problems

This section adopts one of the common palaeogeographic reconstructions in which Greenland has undergone substantial dextral motion along Nares Strait. Evidence for major strike-slip movement along Nares Strait in late Phanerozoic time that fits generally with conventional plate tectonic reconstructions is summarised by Johnson & Srivastava (this volume). The estimates of sinistral wrench displacement vary from 200 to 320 km. Fig. 5 shows a typical pre-drift reconstruction based on a displacement of 250 km in which Baffin Bay has been closed by approximately juxtaposing the shelf edges at about the 1000 m isobaths. Similar (although not identical) reconstructions have been presented in this volume (e.g. Keen & Peirce, Menzies, Peirce, Srivastava & Falconer) and are common in the literature (e.g. Keen et al. 1972, Feden et al. 1979).

The following considerations show that such a reconstruction cannot represent the relative positions of Ellesmere Island and Greenland in Precambrian to mid-Palaeozoic time.

Disruption of the regional facies belts and geological–geophysical markers

The 250 km dextral reconstruction violates 26 out of the 29 lines of evidence quoted in Tables 1 and 2, and it is strongly argued against by the seismic, gravity and structural data discussed in section 2 above. The evidence that is not in obvious disagreement is the magnetic data (marker 18, Table 1), the postglacial isobase data (I, Table 2) and the earthquake epicentre zone (marker 20, Table 1). The latter two are not diagnostic because of age criteria, since they reflect geological events post-dating the suggested major late Phanerozoic sinistral displacement. Several magnetic anomalies occur in the Nares Strait region including magnetic highs which bridge the Strait without apparent offset (Riddihough et al. 1973). However, as noted by Peirce (this volume), the prominent magnetic high that crosses Kane Basin (marker 18, Table 1) could *theoretically* be matched otherwise in a reconstruction involving 250 km of displacement.

Juxtaposition of Precambrian and Palaeozoic rock units

The reconstruction restoring 250 km of sinistral displacement results in the juxtaposition of Precambrian

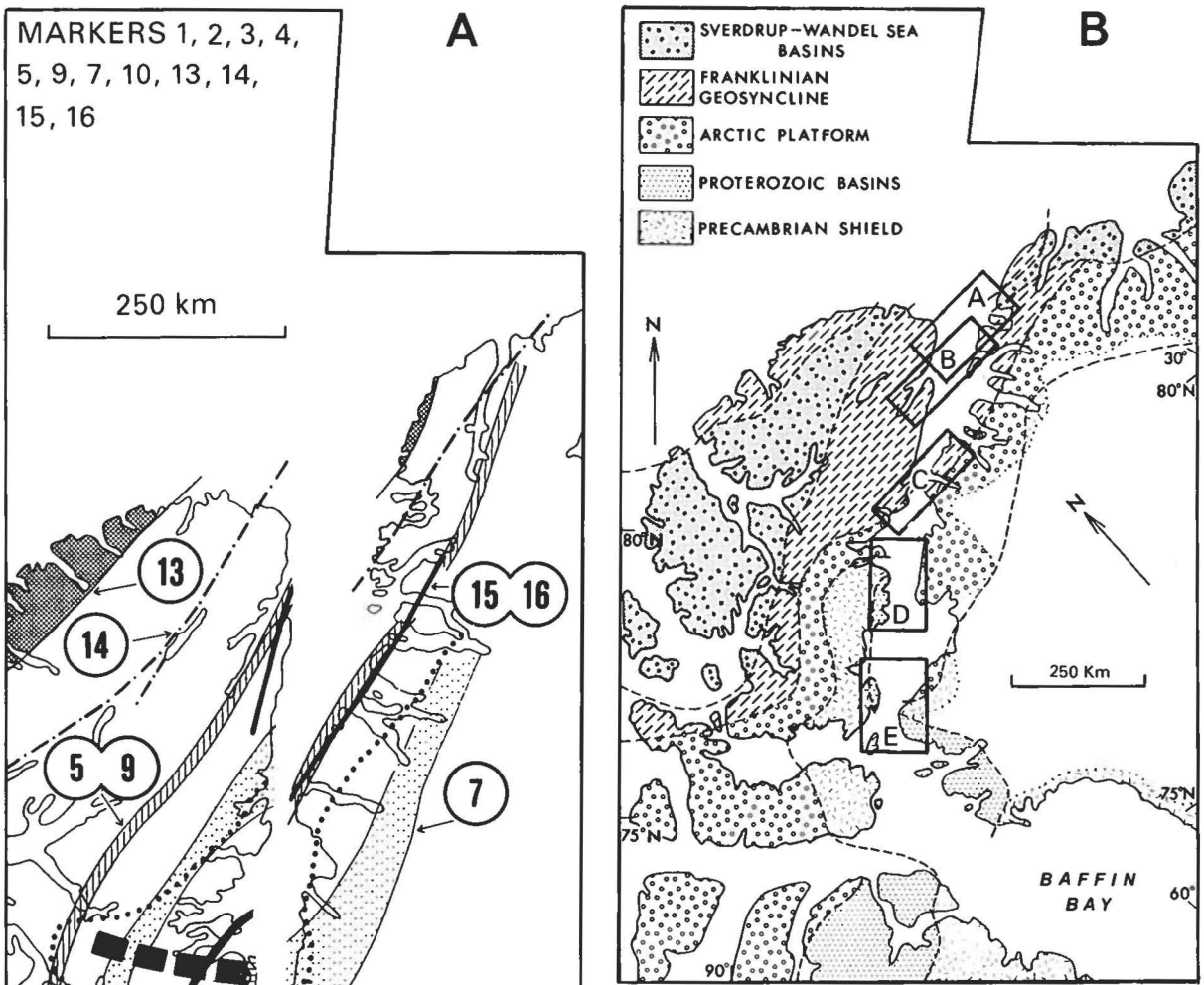


Fig. 5. Maps showing the effects of 250 km dextral displacement on a selection of geological markers (A) and on the main stratigraphic–structural units (B) of the Nares Strait region. The position of the markers and stratigraphic–structural units in present-day geography is given in Figs 3 and 2, respectively; the markers are described in Table 1. Marker 12, the southern boundary of the Ellesmere–Greenland fold belt, which has been the most discussed marker with regard to strike-slip displacement along Nares Strait, is not featured here but in Figs 8 and 9. The five frames lettered A to E locate areas of juxtaposed rock associations that are discussed in the text.

and Palaeozoic rock units which, because of their contrasting sites of formation, are in positions that are either incompatible or extremely unlikely. The most critical geological complications introduced by the reconstruction are discussed below — listed from north to south and grouped in five areas marked on Fig. 5B.

A. Unmetamorphosed Silurian–Devonian turbidites of the Imina Formation forming both sides of present-day Robeson Channel (Trettin 1979, Dawes, this volume, Hurst & Kerr a, this volume, Surlyk, this volume) are now separated and the amphibolite-facies Proterozoic and Cambrian metasediments of

northern Peary Land are positioned on regional strike with unmetamorphosed clastic rocks of the Hazen Plateau (see Higgins et al., this volume).

B. Both the Cambrian–Ordovician and Ordovician–Silurian platform margins in the Judge Daly Promontory area in Ellesmere Island (markers 5 and 8, Table 1) are separated from their counterparts in Greenland and placed so that they strike towards the axis of the North Greenland trough.

C. The configuration of the Ellesmere Island – Greenland Silurian turbidite basin (Surlyk, this volume) is completely disrupted and the North Greenland turbidites in Hall Land and Nyeboe Land strike towards the Proterozoic to Ordovician platform strata and the Precambrian Bache Peninsula arch.

The platform succession in Ellesmere Island contains Lower and Middle Ordovician evaporites of lagoonal and sabkha environments (markers 6 and 7, Table 1). These are separated from the equivalent evaporite-bearing formations in Washington Land, Greenland, and placed on strike with clastic strata of a slope to trough environment.

D. The unmetamorphosed Lower Palaeozoic strata of Washington Land (Hurst & Kerr a, this volume, Peel & Christie, this volume) are positioned against the Precambrian Shield with its Proterozoic cover sequence. Thus, the Greenland Silurian reef complex and interdigitating shale facies, which developed at the hinge between platform and basin slope, become separated from its natural counterpart in Canada and juxtaposed with Precambrian crystalline rocks on its basin slope side.

The thick Proterozoic Thule Group sediments and igneous rocks on south-eastern Ellesmere Island and their thinner platform correlatives in Bache Peninsula (Dawes et al., this volume, Peel et al., this volume) are left isolated as a small intracratonic basin with no remnant counterparts in Greenland, and placed in juxtaposition with platform rocks as young as Silurian.

E. The Proterozoic Thule Basin strata and the thin platform sequence over the Bache Peninsula arch in Inglefield Land are displaced from their obvious counterparts in Ellesmere Island, to become a separate intracratonic basin–arch–platform system, also without correlatives across Nares Strait. The thick Proterozoic volcanic–sedimentary succession at Clarence Head, Ellesmere Island — the detailed correlative of the Northumberland Ø succession in Greenland (Dawes et al., this volume) — is placed opposite the conspicuously thin Proterozoic–Cambrian platform rocks of Inglefield Land, a relationship that suggests a northerly-trending basin margin feature in the actual Strait area.

The reconstructed rock associations described above violate the well-established tectonic and stratigraphic framework, and serious problems arise in the interpretation of the geology in terms of universally accepted stratigraphic principles. In general terms, this is based on an ancient continental margin model — a Precambrian craton flanked to the north by a major sedimentary linear basin (geosyncline). There is a deep-water basin in Ellesmere Island (Trettin 1971, 1979) and there is a similar basin with very similar successions in northern Greenland (Dawes 1976, Surlyk et al. 1980). They are fringed to the south by a shallow-water carbonate platform. This couplet is a continuous feature across northern Ellesmere Island and northern Greenland and represents the Franklinian geosyncline of Schuchert (1923). This model also involves a Precambrian Shield which is part of the North Atlantic craton with its intracratonic Proterozoic basins that outcrop to the south of the Proterozoic–Devonian platform–basin couplet. The linear basin was deformed by the mid-Palaeozoic Ellesmerian (Silurian–Devonian) orogeny. Present-day rock distribution on either side of Nares Strait is entirely consistent with such a well-established model. The disruption of straightforward relationships in Precambrian and Palaeozoic rocks, which results from a reconstruction removing 250 km of sinistral displacement, challenges the clear understanding of the regional stratigraphy as well as principles that are accepted world-wide.

Attempts to explain the reconstructed rock associations by differential vertical movements of Ellesmere Island and Greenland are inadequate. For example, the juxtaposition of the Silurian carbonate and shale sections in Greenland close to the Precambrian Shield of Ellesmere Island (Fig. 5) cannot be explained as a product of uplift and erosion of Ellesmere Island, as suggested by Newman & Falconer (1978) and Newman (a, this volume). About midway along Nares Strait at latitude 79°N, the Precambrian Shield erosion surface is delimited precisely by the overlying Proterozoic and younger strata. This erosion surface is at a comparable height above sea level on the facing coasts of Bache Peninsula and Inglefield Land (Fig. 6). Thus, no appreciable difference in net vertical displacement has occurred, at least not in this part of Nares Strait. This is supported by the glacial rebound data from northern Nares Strait (England, this volume), which show no indication that fault movements along Nares Strait have affected the normal isostatic uplift of Greenland and Ellesmere Island.

4. A substantially wider Nares Strait: the problems

Inherent in many conventional reconstructions — those featured in this volume (e.g. Peirce, Srivastava & Falconer) and those published elsewhere (e.g. Bullard et al. 1965, Keen et al. 1972, Sclater et al. 1977, Sweeney et al. 1978, Srivastava 1978) — is the prediction that, *as well as* major sinistral strike-slip displacement along

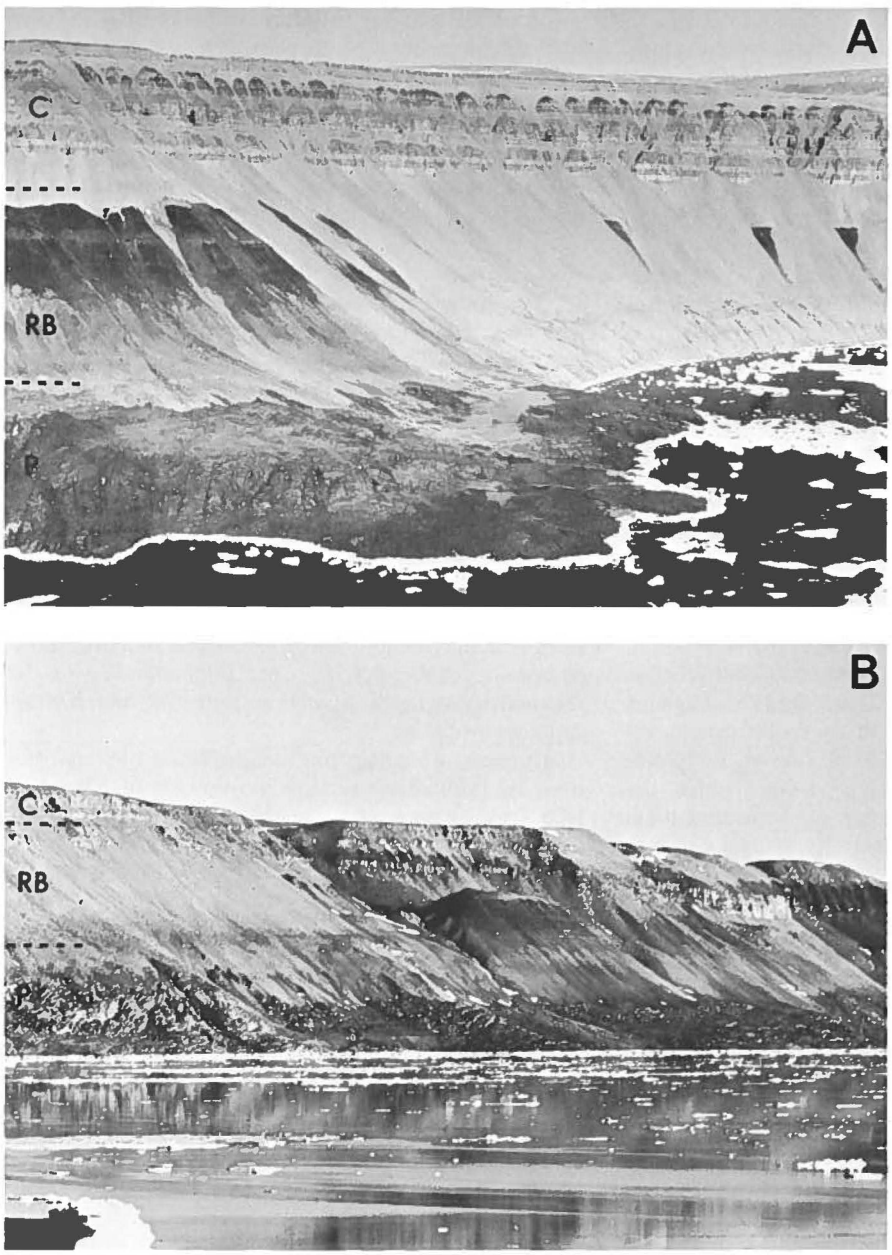


Fig. 6. Views of the Ellesmere Island (A) and the Greenland (B) shores of Nares Strait at about latitude 79°N showing the unconformity between the Precambrian Shield and the overlying flat-lying Proterozoic–Cambrian strata. A = Cape Camperdown, Bache Peninsula, B = east of Force Bugt, Inglefield Land — for locations, see in Figs 1 and 7. P = Precambrian Shield, RB = Proterozoic Rensselaer Bay Formation with two basic sills (dark), C = Cambrian strata with light-coloured clastics of the Dallas Bugt Formation overlain by dolomites of the Cape Leiper and Cape Ingersoll Formations. The Cape Camperdown coastal succession reaches into the Middle Cambrian — the comparable sequence is preserved along the Inglefield Land coast, north-east of Force Bugt (see Peel & Christie, this volume). Note the comparable height above sea level of the erosion surface of the Shield (~ 100 m) and the undeformed nature of the overlying cover. Photographs, A: P. Schledermann, B: J. S. Peel.

Nares Strait, initial separation of Greenland and Ellesmere Island was along a substantially wider Nares Strait. For example, Srivastava (1978) indicates a gap of 220 km between the two landmasses in a reconstruction for late Cretaceous time (Fig. 7), and other models show even greater oblique displacement (e.g. Bullard et al. 1965, Sclater et al. 1977, Sweeney et al. 1978).

The geological evidence against major strike-slip displacement already summarised in the preceding sections, argues against the parts of these reconstructions involving strike-slip. Additional evidence argues against the other feature of these reconstructions, wide separation. The most significant implication of reconstructions suggesting original wide separation is that a substantial volume of crustal material must have disappeared by transformation or subduction or was tectonically shortened by folding and faulting across Nares Strait as Greenland approached Ellesmere Island. There is no evidence on the shores of Nares Strait to suggest that such processes took place on the scale necessitated by the reconstructions. On the contrary, the coastal geology indicates that such dynamic tectonism did not occur. In addition, we have already quoted the data which strongly suggest that Nares Strait does not rep-

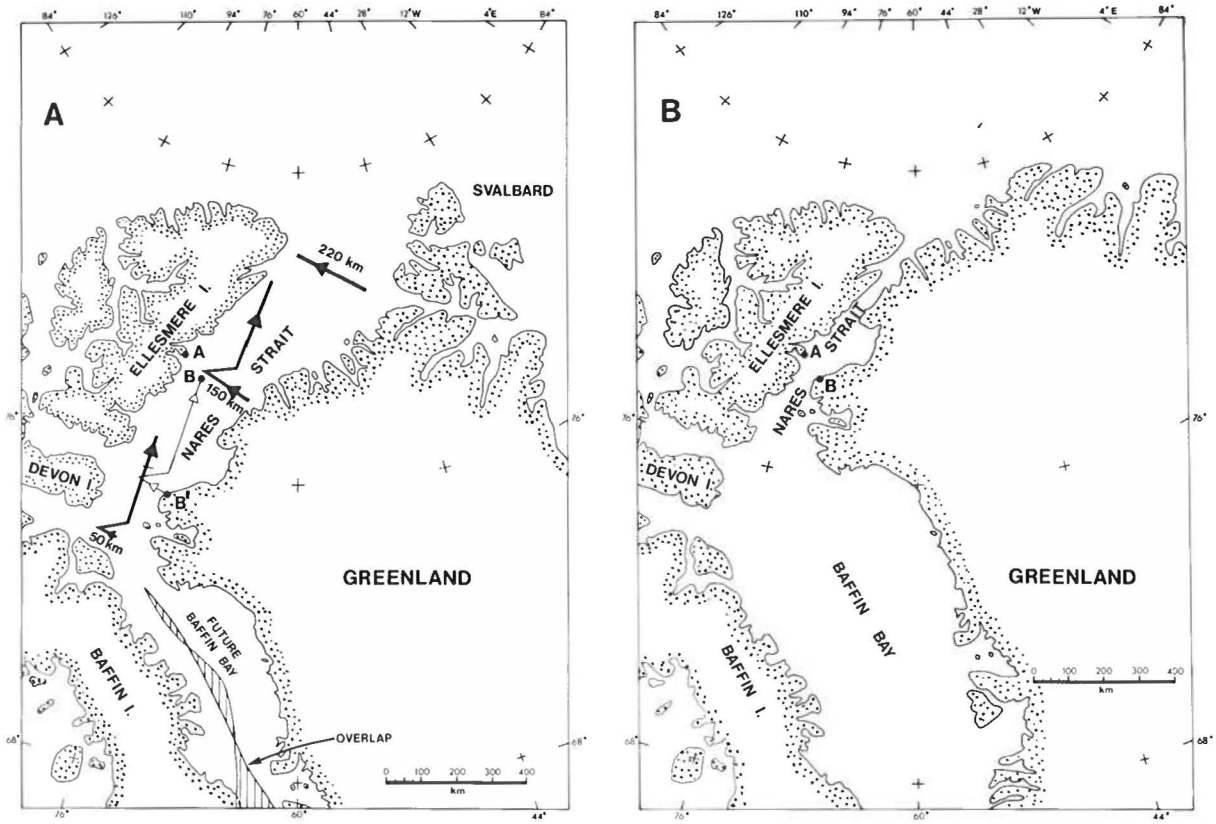


Fig. 7. Transverse drift of Greenland that is implicit in many reconstructions of the Baffin Bay region. A: Pre-drift position of Greenland in the late Cretaceous (Campanian, anomaly 32), with heavy arrows indicating movement directions. If such movements occurred, then a location east of Force Bugt, Inglefield Land (see Fig. 6B) would have originated at point B′, and would have travelled the path shown by the light arrow, to reach its present location at B. B: Present-day geography showing that the location east of Force Bugt, and point A (Cape Camperdown, Bache Peninsula, see Fig. 6A) are now only about 85 km apart. If they originated about 400 km apart as shown on the left, their transverse approach *alone* requires the disappearance (presumably by subduction) of at least 150 km of continental crust. The flat-lying undisturbed nature of the Proterozoic–Cambrian columns at A and B (together with similarly homoclinal Ordovician and Silurian strata) precludes any substantial transverse movement of those points toward each other.

resent a fundamental fracture of the lithosphere — an ancient plate boundary.

The main problem is schematically represented in Fig. 7 that illustrates the reconstruction by Srivastava (1978) and Srivastava & Falconer (this volume) in which a two-phase episode of movement is suggested. Other authors have suggested similar drift paths of Greenland (e.g. Sclater et al. 1977). Bache Peninsula and south-western Inglefield Land are geographic locations that today are about 75 km apart and midway along the Strait (Figs 1 and 7). The close geological correlation between the Precambrian Shield rocks on the shores of Smith Sound has been described in this volume (Frisch & Dawes), and there seems little doubt that in Precambrian and Palaeozoic time the entire region was underlain by continental crust. Thus, since Bache Peninsula and south-western Inglefield Land are several hundred kilometres apart in the reconstruction (Fig. 7), an appreciable segment of continental crust has in some way been dispensed with between the two points during the proposed travel path of Greenland. One would expect such a dynamic geological process — subduction or tectonic shortening or a combination — to leave some visible evidence on the shores of the Strait. Even if one speculated that Nares Strait was formerly the site of oceanic or attenuated crust, similar tectonic problems remain.

The Shield of Bache Peninsula and south-western Inglefield Land shows a well-developed erosion surface, on which flat-lying to shallow-dipping Proterozoic and basal Cambrian strata rest (Fig. 6). These platform successions form spectacular cliff sections on the shores of Nares Strait; they are undeformed, contain unstrained fossils and show no evidence of having been subjected to dynamic forces of the nature necessary to explain the disappearance between them of an extensive segment of continental crust. Furthermore, a guide to the thermal history of this part of the Nares Strait coast has been

obtained from the colour index (CAI) of Ordovician and Silurian conodonts which indicates a post-Silurian temperature in southern coastal Washington Land, Greenland, of less than 90°C (Aldridge 1980, R. J. Aldridge and J. S. Peel, pers. comm. 1981); a thermal condition difficult to explain if indeed Nares Strait were the site of a subduction zone.

It is also noteworthy that no late Phanerozoic igneous outcrops are known along Nares Strait that might be related to such a closing of an initial wide gap between Ellesmere Island and Greenland. Volcanic fragments and pebbles, mostly of basalt and of southerly provenance, occur in some clastic beds of the Paleocene Eureka Sound Formation in eastern Ellesmere Island (Miall 1981, 1982). This material is generally considered indicative of an episode of late Phanerozoic magmatism connected to the Nares Strait rift — a point also mentioned in this volume by Peirce. However, the age of the igneous clast material is unknown (A. D. Miall, pers. comm. 1981). Incidentally, the Kane Basin – Smith Sound region is part of an important Proterozoic igneous province which might also be considered a possible source for such basaltic material.

5. Marker evidence for the 250 km reconstruction: the problems

Three lines of evidence based on 'geological–geophysical markers' have been presented in this volume for major sinistral strike-slip displacement of Greenland. These, summarised by Johnson & Srivastava (this volume), are given below.

a) Correlation of the southern boundary of folded and faulted strata of the mid-Palaeozoic Ellesmere–Greenland fold belt (Keen & Peirce, Newman a, Srivastava & Falconer).
b) Correlation of the Proterozoic Thule Group of Greenland with the Proterozoic deposits of the Borden Basin of northern Baffin Island and Bylot Island (Newman a).
c) Correlation of two magnetic high anomalies in the Lincoln Sea (Kovacs).

a) The southern boundary of the Ellesmere–Greenland fold belt with its apparent mis-alignment (Fig. 2) has been the most discussed geological marker in the Nares Strait debate since Wegener (1915, 1924) first drew attention to it in a displacement context (e.g. Kerr 1967a, Trettin et al. 1972, Dawes 1973, Newman 1977, Srivastava 1978). There is no doubt that the southern boundary of the fold belt intersects Nares Strait at different latitudes on Ellesmere Island and Greenland today, and the points of intersection are separated by about 200 km (Figs 2 and 8). A cursory look at this situation might suggest that this is a marker which is compatible with a restoration of 200 km. However, as noted by Higgins et al. (this volume), such restoration introduces a fundamental structural discordance, because the fold trends in the Hall Land – Nyeboe Land region of Greenland trend at a steep angle to the coast, whereas the folds inland on Ellesmere Island have a

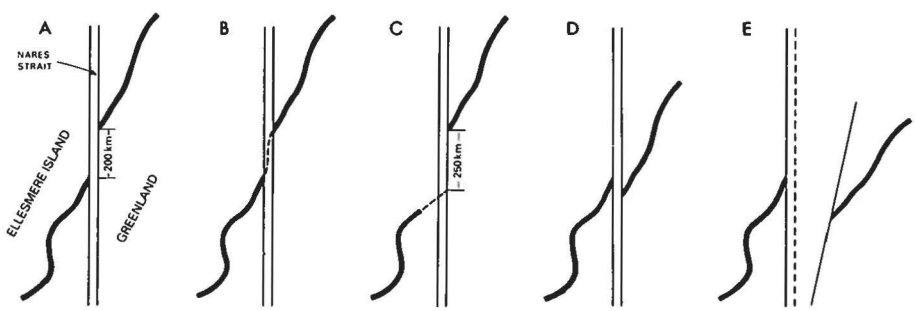


Fig. 8. Diagrammatic representation of Nares Strait and the interpretations placed on the southern boundary of the Ellesmere–Greenland fold belt (marker 12, Table 1).
A: Present-day configuration showing marker 12 intersecting the coasts of Ellesmere Island and Greenland at different latitudes.
B: Interpretation by geologists who have studied the fold belt in the Nares Strait region; the apparent mis-alignment of marker 12 across Nares Strait is due to an original feature of the fold belt margin, entirely consistent with the form of the margin elsewhere in Canada and Greenland. Any sinistral strike-slip faulting along Nares Strait has only accentuated the apparent mis-alignment.
C: Interpretation placed on marker 12 by proponents of conventional plate tectonic reconstructions of the Arctic. The fold belt boundary is interpreted as intersecting the Ellesmere Island coast at a high angle of intersection and the widely separated points of intersection of marker 12 on opposite sides of the Strait are due to about 250 km of sinistral strike-slip displacement of Greenland.
D and E: Diagrammatic reconstructions according to conventional plate tectonic models to illustrate the resultant structural discordancies in the fold belt margin. D: 250 km dextral movement of Greenland, E: 250 km dextral movement, as well as oblique separation, of Greenland and Ellesmere Island to a pre-drift position at the time of initial opening of the Labrador Sea. This corresponds to the reconstruction at anomaly 32 time (late Cretaceous) of Srivastava (1978).

similar trend, but curve near the coast where they strike parallel to the Strait (Fig. 9). As in many fold belts of the world, major fold trends are not straight, and arcuate or sinuous segments are common. The apparent mis-alignment of the Ellesmere–Greenland fold belt margin is the result of a bend or salient in the edge of the fold belt — presumably controlled by some irregularities in the geosyncline margin — that existed before the Strait was formed. A clear remnant of the bend is preserved on Ellesmere Island on the shore of Kane Basin (Kerr 1973a, b) and the site and trend of Nares Strait may well be controlled in some way by the tectonic fabric of this older salient.

There is another salient in the southern edge of the fold belt in south-western Ellesmere Island and Devon Island (Fig. 9) which is accurately mapped on land. To illustrate the suggestion that Nares Strait crosses a similar salient, we show in Fig. 9 a hypothetical seaway identical to Nares Strait in shape, size and orientation. It differs only in location. This hypothetical seaway truncates the fold belt margin at much the same angle as Nares Strait truncates the fold belt margin that now exists in the Nares Strait region. On the hypothetical seaway there is an apparent mis-alignment of the southern boundary of the fold belt by about 300 km and, incidentally, many of the markers shown in Table 1 and Fig. 3 would also show an apparent offset.

We concur to the view expressed in this volume by Higgins et al. that the southern limit of the Ellesmere–Greenland fold belt, on close examination of its structure including main trend of individual folds, supports the case against major strike-slip displacement along the Strait.

In a 250 km reconstruction, the prominent faults in northern Nyeboe Land, Greenland, have been brought into close proximity to the system of faults on the Ellesmere shore of Kane Basin (Newman & Falconer 1978, Newman a, this volume). However, the Nyeboe Land fault zone is a high-angle reverse fault system that coincides with the early Palaeozoic platform margin of the Franklinian geosyncline (Dawes, this volume); the Parrish Glacier thrust in Ellesmere Island is a low-angle thrust system of post-Paleocene age (Mayr & de Vries, this volume). There may well be a tectonic connection between these fault systems since both are situated near and parallel to the southern boundary of the Ellesmere–Greenland fold belt. However, we cannot support such a direct correlation of a low-angle thrust with a high-angle reverse fault as evidence indicating 250 km of strike-slip displacement, particularly since the more obvious continuation of the Nyeboe Land fault zone is the system of high-angle faults on Judge Daly Promontory which also coincide with the early Palaeozoic platform margin.

b) The similarity of the Thule Group strata in Greenland with the Proterozoic rocks of northern Baffin Island and Bylot Island has been suggested as evidence of original close proximity (Newman a, this volume). The detailed correlation of the Proterozoic successions between Greenland and Ellesmere Island in the Smith Sound region has been outlined in this volume (Dawes et al., Peel et al.). We regard it as geologically unsound to reposition together relatively flat-lying successions of general stratigraphic similarity and age which are now about 400 km apart, and abandon a detailed correlation of successions only tens of kilometres apart, including a major basin margin feature — Bache Peninsula arch.

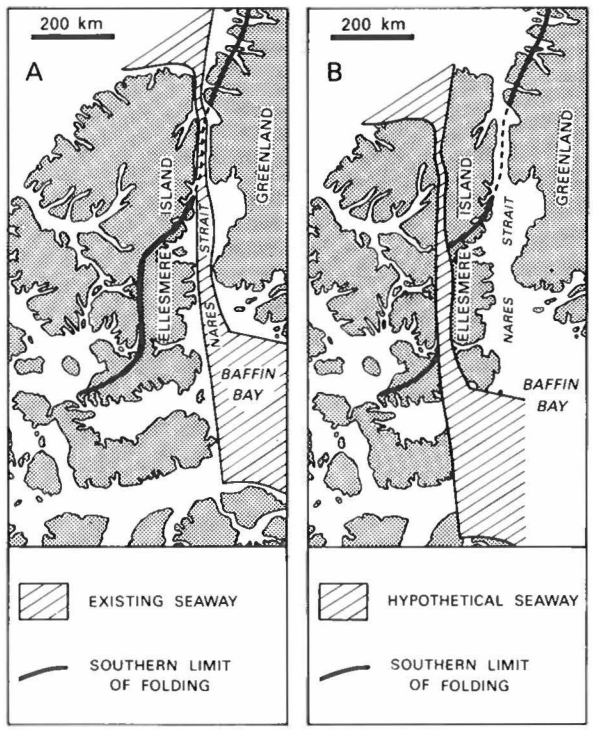


Fig. 9. Maps showing the southern boundary of the Ellesmere–Greenland fold belt (marker 12, Table 1) to illustrate the interpretation that the apparent mis-alignment in this marker along Nares Strait is due to a salient in the fold belt margin, of the same form as present on land in Ellesmere Island.

A: Present-day geography with the Baffin Bay – Nares Strait seaway outlined.

B: Hypothetical case where a seaway of the same shape and size as Nares Strait is located about 150 km to the west resulting in an apparent sinistral offset of marker 12 that simulates the present-day situation at Nares Strait.

c) The only 'direct' calculation of strike-slip displacement along Nares Strait based on a geophysical marker is discussed by Kovacs (this volume), who suggests 240 km displacement from aeromagnetic data from the Arctic Ocean. This displacement is based on a pair of magnetic high anomalies described as being separated and offset by the northern continuation of the Wegener fault in Nares Strait. However, the magnetic anomalies used in the displacement analysis in fact occur on the

same (north-west) side of the proposed site of the Wegener fault, albeit on either side of a suggested second fault line — the Ellesmere fault zone — that strikes toward Ellesmere Island. Thus, the proposed offset of the two magnetic anomalies bears little relationship to strike-slip displacement between Ellesmere Island and Greenland.

Discussion

The available data from the immediate Nares Strait region have been summarised, and we can find no evidence, neither geological nor geophysical, suggesting that strike-slip displacement of the magnitude necessary to produce major changes in the relative positions of Ellesmere Island and Greenland occurred along Nares Strait. We conclude that any case for major displacement must rest on evidence accumulated elsewhere, particularly in the Labrador Sea, Baffin Bay, the Arctic Ocean and the North Atlantic Ocean systems — as well as globally. Johnson & Srivastava (this volume) have shown that such evidence is mainly geophysical and comes from oceanic areas at some distance from the Strait. This evidence predicts great left-lateral displacement of Greenland during plate tectonic movements in late Phanerozoic time.

The foremost argument in support of great displacement along Nares Strait has been that without such displacement the structure and oceanic crust of Baffin Bay and Labrador Sea cannot be explained by the generally accepted plate tectonic model involving substantial sea-floor spreading. Keen & Peirce (this volume) and Srivastava & Falconer (this volume) have elucidated the various geophysical implications and problems which arise if only minor motion along Nares Strait is accepted. However, there has been a steady accumulation of data, in addition to the Nares Strait evidence, that are difficult to reconcile with a model based on rigid plates and substantial opening of Baffin Bay and Labrador Sea by sea-floor spreading and modifications and alternative interpretations have been suggested (e.g. Kerr 1967b, 1981a, b, Grant 1975, 1980, this volume, van der Linden 1975, 1977, Umpleby 1979). In view of the restraint placed on displacement on the Nares Strait lineament, alternatives based on the data and ideas in such papers must be particularly relevant. Linear magnetic anomalies of the type delimited in the *outer* margins of the Labrador Sea and in Baffin Bay (Srivastava 1978, Jackson et al. 1979) may have a number of possible origins and are not necessarily indicative of the generally accepted model of sea-floor spreading, while a gravity low in the centre of Baffin Bay represents a structure which is not necessarily a spreading ridge. We regard the geophysical interpretations of these specific features as suspect and indicative of a tendency to regard oceans as the products of an oversimplified model of sea-floor spreading, regardless of the restraints placed on such models by onshore geological data.

Monahan & Johnson (this volume) present an illuminating comparison between the physiography of Nares Strait and other major structural lineaments that also have been the centre of long controversies about fault displacement. They conclude that such onland lineaments as the San Andreas Fault and the Great Glen Fault are considerably more amenable to analysis of the amount of strike-slip displacement than is Nares Strait. Despite the fact that the land surrounding Nares Strait is partly ice covered, and the coasts are separated by more than 20 kilometres, we doubt this conclusion for the following reasons:

1) The regional geology of Nares Strait is relatively simple, being composed of rocks and structures of various ages, forming what is essentially a parallel system of stratigraphic–structural units.
2) The regional trend of the stratigraphic–structural units is persistent and is part of a system extending for 3000 km from the Beaufort Sea to the Greenland Sea. In the region of the Strait, the trend is not complicated by tectonic, metamorphic, magmatic or metasomatic processes.
3) Nares Strait cuts obliquely across all the regional stratigraphic–structural units in which a number of uncontroversial geological markers of known age occur.
4) Because of its high arctic location, Nares Strait geology is beautifully exposed: rocks are not covered by vegetation and since it is the site of a seaway, the sea-cliffs form excellent exposures, often continuous for tens of kilometres.
5) Regions of great strike-slip normally have associated structural complications over many, sometimes tens of kilometres from the main fault plane, which often make correlation across the main fault problematical. The coasts of Nares Strait do not show such complexities.

The science of geology has now reached a stage at which acceptance of basic stratigraphic and structural principles allows regional correlations to be made across all the major oceans of the world. In the last decade many papers put forward pre-drift reconstructions based on such rock correlations. This trend is well illustrated by, for example, Schenk's (1981) account of the Meguma Zone of Nova Scotia and the search for a suitable source continent, for which north-western Africa (Morocco) — some 4500 km away in present-day geography — is a prime candidate. Our present understanding of the origin of the Atlantic Ocean system is partly based on such long-range rock correlations and, indeed, Wilson's (1966) initial evidence concerning the closure and the reopening of the Atlantic was based on rock correlations. Recognition of exotic rock segments on the edges of the Atlantic is not at variance with the general thesis of the opening cycle of the Atlantic and as such they are understandably given serious appraisal.

Nares Strait is less than 25 km across in its narrowest part. Several islands in Kennedy Channel composed of various Silurian carbonate facies, which correlate with rocks on the adjacent Greenland coast (J. M. Hurst, pers. comm. 1981), reduce the actual distance across which correlation is made, to about 15 km. It is interesting that many earth scientists have found it difficult to accept correlation of near identical rock successions across such a narrow stretch of water as indicating that the rocks were formed in their present relative positions, when the majority are willing to accept, and to use, the general similarity of rock columns thousands of kilometres apart as evidence that those rocks were formed in close proximity. This is all the more surprising as rock correlations across Nares Strait are not problematical and in no way in conflict with any geological or geophysical data from the Strait area.

Conclusion

A summary of the present geological and geophysical knowledge of the Nares Strait region fails to reveal any feature indicating strike-slip displacement between Ellesmere Island and Greenland. We conclude that the evidence presented at the symposium "Did Greenland drift along Nares Strait?" and in this volume demonstrates that any net strike-slip displacement along Nares Strait has been minor, i.e. in the range of 0 to 25 km. Furthermore, there is no direct evidence to support a tectonic model involving a complicated history of motion along the Strait, with major strike-slip, vertical or oblique movements, that might be accommodated in some suitable scenario and connected to the suggested conventional opening of the surrounding oceans that predicts great sinistral motion of Greenland in late Phanerozoic time. If the evidence indicating geological continuity between Ellesmere Island and Greenland is accepted, modifications should be considered to the widely accepted ideas for the opening of the Labrador Sea and Baffin Bay, which are based on a conventional, substantial sea-floor spreading, rigid-plate model. If the restraints on motion along Nares Strait are ignored, Greenland will continue to be allowed to drift around in response to refinements of such plate tectonic models that are suspect, since they do not incorporate the available data from the immediate onshore regions.

Acknowledgements

We thank numerous colleagues at the Geological Survey of Greenland and the Geological Survey of Canada for co-operation and discussion over a number of years of the geology of the Nares Strait region, in particular R. L. Christie, W. E. Davies, J. M. Hurst, U. Mayr, W. W. Nassichuk, J. S. Peel, R. Thorsteinsson, H. P. Trettin and A. Weidick. A. K. Higgins and J. M. Hurst are thanked for critical reading of the text and we are grateful to Esben Glendal for valuable comments and technical assistance in the preparation of the manuscript. P. Schledermann, Arctic Institute of North America and J. S. Peel, Geological Survey of Greenland, are thanked for the photographs used in Fig. 6. The paper is published with the permission of the Director of the Geological Survey of Greenland.

References

Aldridge, R. J. 1980. Conodont colour as a guide to thermal history: variation in the upper Llandovery strata of North Greenland. – Unpubl. rep. Grønlands geol. Unders.: 4 pp.

Bullard, E., Everett, J. E. & Smith, A. G. 1965. The fit of the continents around the Atlantic. – In: Blackett, P. M. S., Bullard, E. & Runcorn, S. K., A symposium on continental drift. – Phil. Trans. Roy. Soc. Lond. 258A: 41–51.

Christie, R. L. 1967. Operation Grant Land (1966), northern Ellesmere Island. – Pap. geol. Surv. Can. 67–1: 2–3.

Christie, R. L. 1974. Northeastern Ellesmere Island: Lake Hazen region and Judge Daly Promontory. – Pap. geol. Surv. Can. 74–1A: 297–299.

Christie, R. L., Dawes, P. R., Frisch, T., Higgins, A. K., Hurst, J. M., Kerr, J. W. & Peel, J. S. 1981. Geological evidence against major displacement in the Nares Strait. – Nature, Lond. 291: 478–480.

Coles, R. L., Haines, G. V. & Hannaford, W. 1976. Large scale magnetic anomalies over western Canada and the Arctic: a discussion. – Can. J. Earth Sci. 13: 790–802.

Dawes, P. R. 1973. The North Greenland fold belt: a clue to the history of the Arctic Ocean basin and Nares Strait lineament. – In: Tarling, D. H. & Runcorn, S. K. (eds), Implications of continental drift to the earth sciences 2: 925–948. – Academic Press, London & New York.

Dawes, P. R. 1976. Precambrian to Tertiary of northern Greenland. – In: Escher, A. & Watt, W. S. (eds), Geology of Greenland: 248–303. – Geol. Surv. Greenland, Copenhagen.

Dawes, P. R. 1982. The Nyeboe Land fault zone: a major dislocation on the Greenland coast along northern Nares Strait. – This volume.

Dawes, P. R., Frisch, T. & Christie, R. L. 1982. The Proterozoic Thule Basin, Greenland and Ellesmere Island: importance to the Nares Strait debate. – This volume.

England, J. 1982. Postglacial emergence along northern Nares Strait. – This volume.

Feden, R. H., Vogt, P. R. & Fleming, H. S. 1979. Magnetic and bathymetric evidence for the "Yermak hot spot" northwest of Svalbard in the Arctic Basin. – Earth planet. Sci. Lett. 44: 18–38.

Frisch, T. & Dawes, P. R. 1982. The Precambrian Shield of northernmost Baffin Bay: correlation across Nares Strait. – This volume.

Grant, A. C. 1975. Geophysical results from the continental margin off southern Baffin Island. – In: Yorath, C. J., Parker, E. R. & Glass, D. J. (eds), Canada's continental margin and offshore petroleum exploration. – Mem. Can. Petrol. Geol. 4: 411–431.

Grant, A. C. 1980. Problems with plate tectonics: the Labrador Sea. – Bull. Can. Petrol. Geol. 28: 252–278.

Grant, A. C. 1982. Problems with plate tectonic models for Baffin Bay – Nares Strait: evidence from the Labrador Sea. – This volume.

Higgins, A. K., Mayr, U. & Soper, N. J. 1982. Fold belts and metamorphic zones of northern Ellesmere Island and North Greenland. – This volume.

Hurst, J. M. & Kerr, J. W. 1982a. Upper Ordovician to Silurian facies patterns in eastern Ellesmere Island and western North Greenland and their bearing on the Nares Strait lineament. – This volume.

Hurst, J. M. & Kerr, J. W. 1982b. Brecciated lineaments in Washington Land, North Greenland, and their relation to Nares Strait. – This volume.

Jackson, H. R., Keen, C. E., Falconer, R. K. H. & Appleton, K. P. 1979. New geophysical evidence for sea-floor

spreading in central Baffin Bay. – Can. J. Earth Sci. 16: 2122–2135.
Johnson, G. L. & Srivastava, S. P. 1982. The case for major displacement along Nares Strait. – This volume.
Keen, C. E. & Peirce, J. W. 1982. The geophysical implications of minimal Tertiary motion along Nares Strait. – This volume.
Keen, C. E., Barrett, D. L., Manchester, K. S. & Ross, D. I. 1972. Geophysical studies in Baffin Bay and some tectonic implications. – Can. J. Earth Sci. 9: 239–256.
Kerr, J. W. 1967a. Nares submarine rift valley and the relative rotation of north Greenland. – Bull. Can. Petrol. Geol. 15: 483–520.
Kerr, J. W. 1967b. A submerged continental remnant beneath the Labrador Sea. – Earth planet. Sci. Lett. 2: 283–289.
Kerr, J. W. 1973a. Geology, Dobbin Bay, District of Franklin. 1: 250,000. – Map geol. Surv. Can. 1358A.
Kerr, J. W. 1973b. Geology, Kennedy Channel and Lady Franklin Bay, District of Franklin. 1: 250,000. – Map geol. Surv. Can. 1359A.
Kerr, J. W. 1980. Did Greenland drift along Nares Strait? – Bull. Can. Soc. Petrol. Geol. 24: 279–289.
Kerr, J. W. 1981a. Evolution of the Canadian Arctic Islands: a transition between the Atlantic and Arctic Oceans. – In: Nairn, A. E. M., Churkin, M. & Stehli, F. G. (eds), The ocean basins and margins 5, The Arctic Ocean: 105–199. – Plenum Press, New York & London.
Kerr, J. W. 1981b. Stretching of the North American plate by a now dormant Atlantic spreading centre. – In: Kerr, J. W. & Fergusson, A. J. (eds), Geology of the North Atlantic borderlands. – Mem. Can. Soc. Petrol. Geol. 7: 245–278.
Kerr, J. W. 1982. History and implications of the Nares Strait conflict. – This volume.
Kovacs, L. C. 1982. Motion along Nares Strait recorded in the Lincoln Sea: aeromagnetic evidence. – This volume.
Mayr, U. & de Vries, C. D. S. 1982. Reconnaissance of Tertiary structures along Nares Strait, Ellesmere Island, Canadian Arctic Archipelago. – This volume.
Menzies, A. W. 1982. Crustal history and basin development of Baffin Bay. – This volume.
Miall, A. D. 1981. Late Cretaceous and Paleogene sedimentation and tectonics in the Canadian Arctic Islands. – In: Miall, A. D. (ed.), Sedimentation and tectonics in alluvial basins. – Spec. Pap. geol. Ass. Can. 23: 221–272.
Miall, A. D. 1982. Tertiary sedimentation and tectonics in the Judge Daly Basin, northeast Ellesmere Island, Arctic Canada. – Pap. geol. Surv. Can. 80–30: 17 pp.
Monahan, D. & Johnson, G. L. 1982. Physiography of Nares Strait: importance to the origin of the Wegener Fault. – This volume.
Newman, P. H. 1977. The offshore and onshore geophysics and geology of the Nares Strait region: its tectonic history and significance in regional tectonics. – Unpubl. M. Sc. thesis, Dalhousie Univ., Canada: 153 pp.
Newman, P. H. 1982a. A geological case for movement between Canada and Greenland along Nares Strait. – This volume.
Newman, P. H. 1982b. Marine geophysical study of southern Nares Strait. – This volume.
Newman, P. H. & Falconer, R. K. H. 1978. Evidence for movement between Greenland and Canada along Nares Strait. – Geol. Soc. Am. Abs. with Prog. 10: 463 only.
Peel, J. S. & Christie, R. L. 1982. Cambrian–Ordovician platform stratigraphy: correlations around Kane Basin. – This volume.
Peel, J. S., Dawes, P. R., Collinson, J. D. & Christie, R. L. 1982. Proterozoic – basal Cambrian stratigraphy across Nares Strait: correlation between Inglefield Land and Bache Peninsula. – This volume.
Peirce, J. W. 1982. The evolution of the Nares Strait lineament and its relation to the Eurekan orogeny. – This volume.
Riddihough, R. P., Haines, G. V. & Hannaford, W. 1973. Regional magnetic anomalies of the Canadian Arctic. – Can. J. Earth Sci. 10: 157–163.
Schenk, P. E. 1981. The Meguma Zone of Nova Scotia – a remnant of Western Europe, South America, or Africa? – In: Kerr, J. W. & Fergusson, A. J. (eds), Geology of the North Atlantic borderlands. – Mem. Can. Soc. Petrol. Geol. 7: 119–148.
Schuchert, C. 1923. Sites and nature of the North American geosynclines. – Bull. geol. Soc. Am. 34: 151–230.
Sclater, J. G., Hellinger, S. & Tapscott, C. 1977. The paleobathymetry of the Atlantic Ocean from the Jurassic to the present. – J. Geol. 85: 509–552.
Sobczak, L. W. 1978. Gravity from 60°N to the North Pole. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 67–74.
Sobczak, L. W. 1982. Fragmentation of the Canadian Arctic Archipelago, Greenland and the surrounding oceans. – This volume.
Sobczak, L. W. & Stephens, L. E. 1974. The gravity field of northeastern Ellesmere Island, part of northern Greenland and Lincoln Sea with map. Lincoln Sea 1: 500 000. – Earth Physics Branch, Gravity Map Series 114: 9 pp.
Srivastava, S. P. 1978. Evolution of the Labrador Sea and its bearing on the early evolution of the North Atlantic. – Geophys. J. Roy. astr. Soc. 52: 313–357.
Srivastava, S. P. & Falconer, R. K. H. 1982. Nares Strait: a conflict between plate tectonic predictions and geological interpretation. – This volume.
Surlyk, F. 1982. Nares Strait and the down-current termination of the Silurian turbidite basin of North Greenland. – This volume.
Surlyk, F., Hurst, J. M. & Bjerreskov, M. 1980. First age-diagnostic fossils from the central part of the North Greenland foldbelt. – Nature, Lond. 286: 800–803.
Sweeney, J. F., Irving, E. & Geuer, J. W. 1978. Evolution of the Arctic Basin. – In: Sweeney, J. F. (ed.), Arctic geophysical review. – Publ. Earth Physics Branch 45(4): 91–100.
Taylor, F. B. 1910. Bearing of the Tertiary mountain belt on the origin of the earth's plan. – Bull. geol. Soc. Am. 21: 179–226.
Trettin, H. P. 1971. Geology of the lower Paleozoic formations, Hazen Plateau and southern Grant Land mountains, Ellesmere Island, Arctic Archipelago. – Bull. geol. Surv. Can. 203: 134 pp.
Trettin, H. P. 1979. Middle Ordovician to lower Silurian deep-water succession at southeastern margin of Hazen trough, Cañon Fjord, Ellesmere Island. – Bull. geol. Surv. Can. 272: 84 pp.
Trettin, H. P., Frisch, T. O., Sobczak, L. W., Weber, J. R., Niblett, E. R., Law, L. K., DeLaurier, J. M. & Whitham, K. 1972. The Innuitian province. – In: Price, R. A. & Douglas, R. J. W. (eds), Variations in tectonic styles in Canada. – Spec. Pap. geol. Ass. Can. 11: 83–179.
Umpleby, D. C. 1979. Geology of the Labrador Shelf. – Pap. geol. Surv. Can. 79–13: 34 pp.
van der Linden, W. J. M. 1975. Crustal attenuation and sea-floor spreading in the Labrador Sea. – Earth planet. Sci. Lett. 27: 409–423.
van der Linden, W. J. M. 1977. How much continent under the ocean? – Mar. geophys. Res. 3: 209–224.
Wegener, A. 1915. Die Entstehung der Kontinente und Ozeane. – Friedr. Vieweg & Sohn, Braunschweig: 94 pp.
Wegener, A. 1924. The origin of continents and oceans. – Methuen & Co., London: 212 pp. (Transl. 3rd edit. by J. G. A. Skerl).
Wetmiller, R. J. & Forsyth, D. A. 1982. Review of seismicity and other geophysical data near Nares Strait. – This volume.
Wilson, J. T. 1966. Did the Atlantic close and then re-open? – Nature, Lond. 211: 676–681.

Appendix

Note on the magnetic polarity time scales used in the Nares Strait volume

The Nares Strait volume focuses on a debate concerning various problems of correlating the onshore geological record of the Nares Strait region with large-scale plate tectonic movements — in particular the spreading history of the surrounding oceans. Complete integration of the onshore and marine geological and geophysical data must be a prime target for any understanding of the Earth's crust. Nares Strait is one region among many where much work remains to be done in this direction. To effect a synthesis between the onland geological data and those of the ocean basins, a common geochronological framework is required; the geomagnetic polarity time scale has provided this standard for correlation.

Since the 'tape-recorder' concept of linear magnetic anomalies, developed by Vine & Matthews and others, was published in 1963, the interpretation of the magnetic anomaly patterns of the oceans has been an extremely valuable tool in sea-floor spreading/plate tectonic research. The Heirtzler et al. (1968) geomagnetic time scale, derived from the idea that the linear magnetic anomaly pattern associated with oceanic ridges is a manifestation of previous geomagnetic field reversals, is based on a particular magnetic profile in the South Atlantic; dates were assigned by extrapolation from a date of 3.35 m.y. that marks the boundary of the Gauss and Gilbert polarity epochs. It was further assumed that spreading rates prior to the Gilbert epoch (pre-anomaly 3) were unchanged at 1.9 cm/yr^{-1}. Absolute ages of the reversals to produce the time scale were obtained by incorporating palaeomagnetic and palaeontological data.

Since the Heirtzler time scale was introduced, many studies from various branches of geology, geophysics and geochemistry have directly and indirectly provided data which have been used in correction and refinement of the scale. The Deep Sea Drilling Project has played a major role in this and it has been directed to many aspects of linking together the magnetic and biostratigraphic records by detailed investigation of the offshore sediments and their relationships to the sites of identified magnetic anomalies. For example, an early modification of the Heirtzler scale from bore-hole data was made by Sclater et al. (1974), who demonstrated that the Cretaceous–Tertiary boundary in the Indian Ocean (site 216), dated radiometrically at 64.1 ± 1 m.y., is located between magnetic anomalies 29 and 30 and not with anomaly 26 as inferred by the Heirtzler scale. The dating of Palaeogene anomalies was also refined by using a spreading rate in the South Atlantic of 2.04 cm/yr. There has been continual revision of the Cretaceous–Cenozoic magnetic polarity time scale as illustrated by the versions published by, for example, Tarling (1971), Sclater et al. (1974), Tarling & Mitchell (1976), La Brecque et al. (1977), Berggren et al. (1978) and Hailwood et al. (1979).

Many of the revised time scales show serious discrepancies with the original Heirtzler scale, and also between each other. This can be illustrated for example by the ages assigned to anomaly 24; an anomaly that holds a critical place in the North Atlantic sea-floor spreading history, being generally interpreted as the oldest recognisable linear anomaly between Greenland and Europe. On the Heirtzler scale this anomaly is assigned an age of 60 m.y. corresponding to mid-Paleocene time, on the Sclater et al. and LaBrecque et al. scales it is designated as 56 m.y., late Paleocene, while on the Tarling & Mitchell scale it is assigned 49 m.y. at about the Paleocene–Eocene boundary. On more recent scales, anomaly 24 is assigned 54 m.y., Paleocene–Eocene boundary (Berggren et al.) and 52 m.y., early Eocene (Hailwood et al.). A main feature of the revision has been a general expansion of the polarity epochs in Paleocene time and it is now apparent that the original geomagnetic time scale assigns ages to the late Cretaceous and early Tertiary magnetic reversals that are at least 10 per cent too high.

Several papers in this volume that deal with the spreading history of the oceanic regions surrounding Nares Strait refer to the magnetic polarity time scale. A main topic discussed is the relationship between the sea-floor spreading history of the North Atlantic, and the Tertiary (Eurekan) orogeny which affected large parts of Arctic Canada and northern Greenland adjacent to Nares Strait. In such instances a common time scale becomes very relevant, so much so that geological correlations not apparent with reference to one time scale may become so by use of a revised scale.

In order to effect an easier comparison in this volume of the onshore geological events to sea-floor spreading history it was hoped that a 'standard' polarity time scale could be adopted. However, in the editing of this volume such an attempt to single out a particular time scale as 'standard' was given up as impracticable. Hence as many as five different polarity time scales are used in the contributions. To ease the difficulties that the reader might have in equating magnetic anomaly 'time' with biostratigraphic and radiometric 'time', the time scales referred to in this volume are summarised in Fig. 1. It

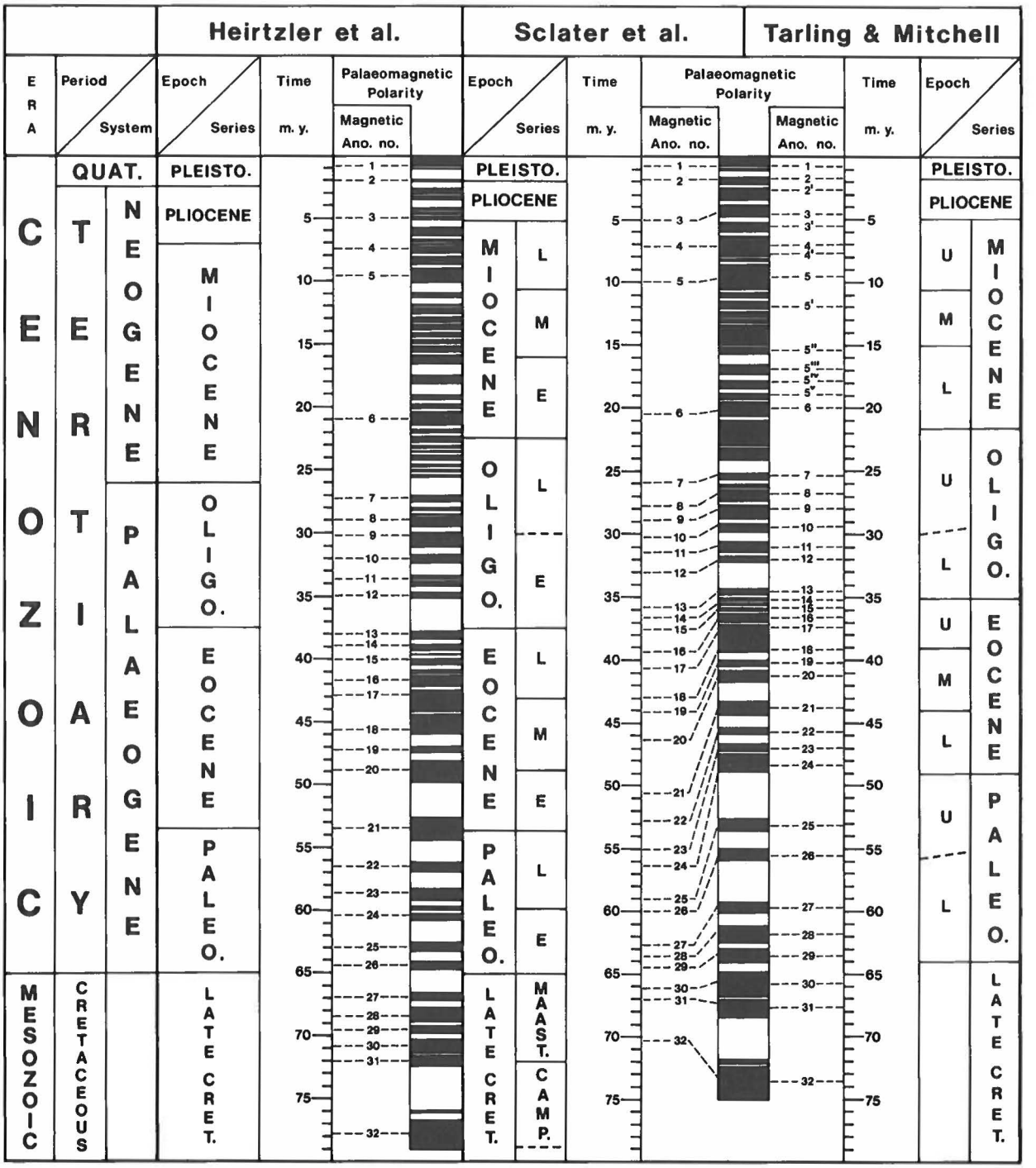
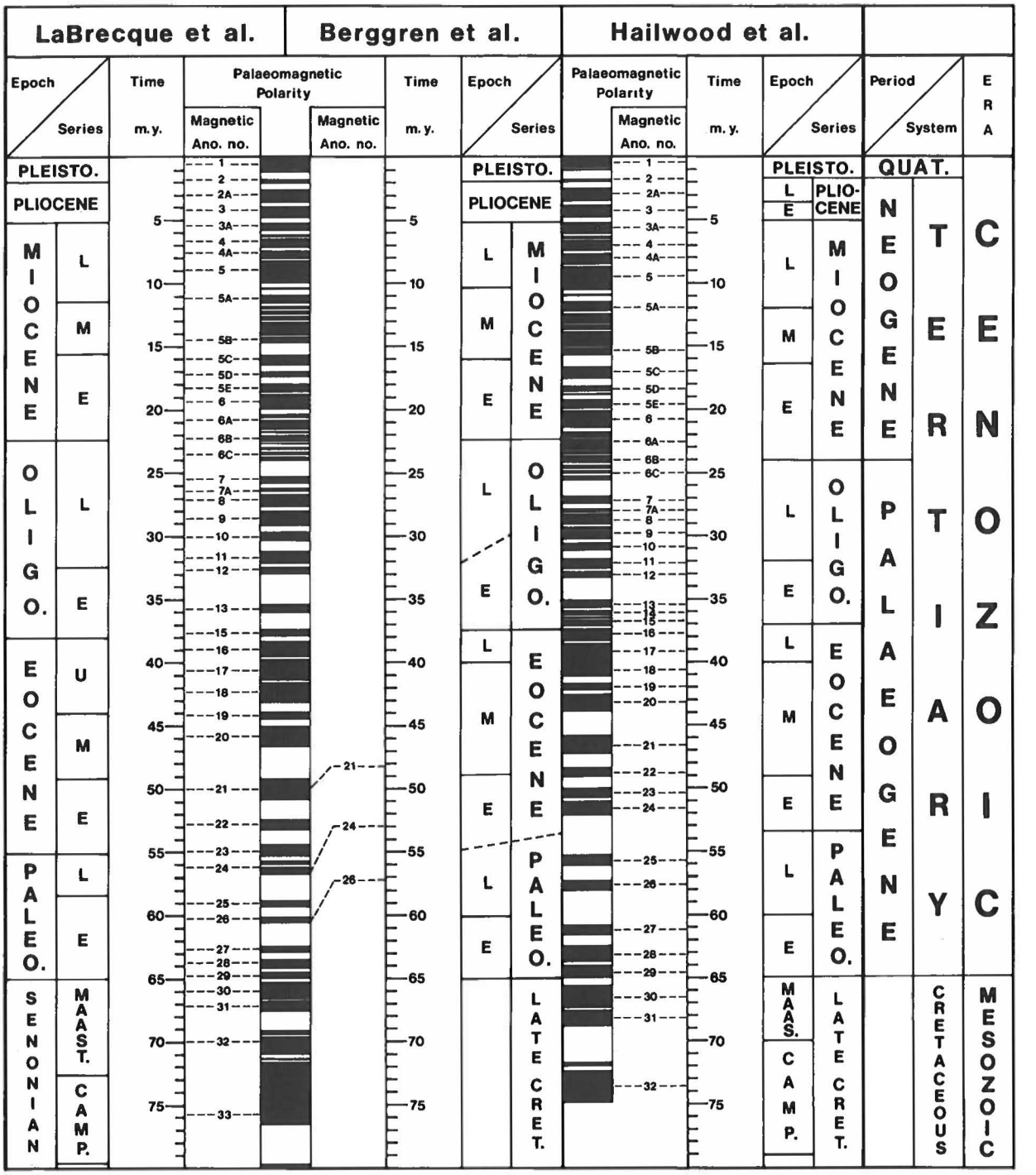


Fig. 1. Comparison of various magnetic polarity and geologic time scales redrawn from Heirtzler et al. (1968), Sclater et al. (1974), Tarling & Mitchell (1976), LaBrecque et al. (1977), Berggren et al. (1978) and Hailwood et al. (1979). The Neogene part of the geologic scale used for Berggren et al. is taken from Berggren (1972). The magnetic polarity scales used in the two central columns are taken from Tarling & Mitchell and LaBrecque et al. so that the magnetic anomaly numbers for Sclater et al. and Berggren et al. are adjusted to the radiometric time scale (m.y.) accordingly.


LaBrecque et al.
Berggren et al.
Hailwood et al.
Epoch
Series
Time
m.y.
Palaeomagnetic Polarity
Magnetic Ano. no.
Period
System
ERA
PLEISTO.
PLIOCENE
MIOCENE
OLIGO.
EOCENE
PALEO.
SENONIAN
L
M
E
U
MAAST.
CAMP.
PLIO-CENE
MIOCENE
OLIGO.
EOCENE
PALEO.
LATE CRET.
QUAT.
NEOGENE
PALAEOGENE
TERTIARY
CRETACEOUS
CENOZOIC
MESOZOIC

should be noted that since the preparation of Fig. 1, the use of the Tarling & Mitchell (1976) time scale in one paper has been discontinued; however, reference to this scale has been retained in the diagram for convenience.

References

Berggren, W. A. 1972. A Cenozoic time-scale – some implications for regional geology and paleobiogeography. – Lethaia 5: 195–215.

Berggren, W. A., McKenna, M. C., Hardenbol, J. & Obradovich, J. D. 1978. Revised Paleogene polarity time scale. – J. Geol. 86: 67–81.

Hailwood, E. A., Bock, W., Costa, L., Dupeuble, P. A., Müller, C. & Schnitker, D. 1979. Chronology and biostratigraphy of Northeast Atlantic sediments, DSDP Leg 48. – Initial Rep. Deep Sea Drilling Proj. 48: 1119–1141. – U.S. Gov. Print. Off., Washington, D.C.

Heirtzler, J. R., Dickson, G. O., Herron, E. C., Pitman, W. C. & Le Pichon, X. 1968. Marine magnetic anomalies, geomagnetic field reversals, and motions of the ocean floor and continents. – J. geophys. Res. 73: 2119–2136.

LaBrecque, J. L., Kent, D. V. & Cande, S. C. 1977. Revised magnetic polarity time scale for Late Cretaceous and Cenozoic time. – Geology 5: 330–335.

Sclater, J. G., Jarrard, R. D., McGowran, B. & Gartner, S. 1974. Comparison of the magnetic and biostratigraphic time scales since the Late Cretaceous. – Initial Rep. Deep Sea Drilling Proj. 22: 381–386. – U.S. Gov. Print. Off., Washington, D.C.

Tarling, D. H. 1971. Principles and applications of palaeomagnetism. – Chapman & Hall, London: 164 pp.

Tarling, D. H. & Mitchell, J. G. 1976. Revised Cenozoic polarity time scale. – Geology 4: 133–136.

Vine, F. J. & Matthews, D. H. 1963. Magnetic anomalies over oceanic ridges. – Nature, Lond. 199: 947–949.

P. R. Dawes

Meddelelser om Grønland

The series *Meddelelser om Grønland* was started in 1879 and has since then published results from all fields of research in Greenland. In 1979 it was split into three separate series:

Geoscience
Bioscience
Man & Society

The series should be registered as *Meddelelser om Grønland, Geoscience (Bioscience, Man & Society)* followed by the number of the paper. Example: *Meddr Grønland, Geosci.* 1, 1979.

The new series are issued by Kommissionen for videnskabelige Undersøgelser i Grønland (The Commission for Scientific Research in Greenland).

Correspondence
All correspondence and manuscripts should be sent to:

The Secretary
Kommissionen for videnskabelige Undersøgelser i Grønland
Øster Voldgade 10
DK-1350 Copenhagen K.

Questions concerning subscription to any or all of the series should be directed to the agent.

Agent
Nyt Nordisk Forlag – Arnold Busck A/S, Købmagergade 49, DK-1150 Copenhagen K. Tlf. +45.1.122453.

Meddelelser om Grønland, Geoscience

Meddelelser om Grønland, Geoscience invites papers that contribute significantly to studies in Greenland within any of the fields of geoscience (physical geography, oceanography, glaciology, general geology, sedimentology, mineralogy, petrology, palaeontology, stratigraphy, tectonics, geophysics, geochemistry). Papers primarily concerned with other areas in the Arctic or Atlantic region may be accepted, if the work also covers Greenland or is of direct importance to continued research in Greenland. Papers dealing with environmental problems and other interdisciplinary studies may be referred to either *Geoscience* or *Bioscience*, according to emphasis and editorial policy.

Editor
T. C. R. Pulvertaft, Institute of General Geology, Øster Voldgade 10, DK-1350 Copenhagen K. Tlf. +45.1. 112232. Telegr. Unigeol.

© 1982 Kommissionen for videnskabelige Undersøgelser i Grønland. All rights reserved. No part of this publication may be reproduced in any form without the written permission of the copyright owner.

N.B. Copies of this volume "Nares Strait and the drift of Greenland: a conflict in plate tectonics" (Geoscience 8) can be purchased either from the agent in Copenhagen or from Serv-West Book Distribution Ltd., 1506 Meridan Road N.E., Calgary, Alberta, Canada T2A 2N9. Tlf. 248-4121.

Meddelelser om Grønland

Bioscience
Geoscience
Man & Society

Published by
The Commission
for Scientific
Research
in Greenland

AiO Print as, Odense, Denmark