This document was produced by scanning the original publication.

Ce document est le produit d'une numérisation par balayage de la publication originale.

GEOLOGICAL SURVEY OF CANADA

DEPARTMENT OF ENERGY, MINES AND RESOURCES

BULLETIN 229

# METAMORPHIC AND PLUTONIC ROCKS OF NORTHERNMOST ELLESMERE ISLAND, CANADIAN ARCTIC ARCHIPELAGO

Thomas Frisch

ce, $4.00

Ottawa
Canada
1974

# METAMORPHIC AND PLUTONIC ROCKS OF NORTHERNMOST ELLESMERE ISLAND, CANADIAN ARCTIC ARCHIPELAGO

*1,400–1973–6358-11*

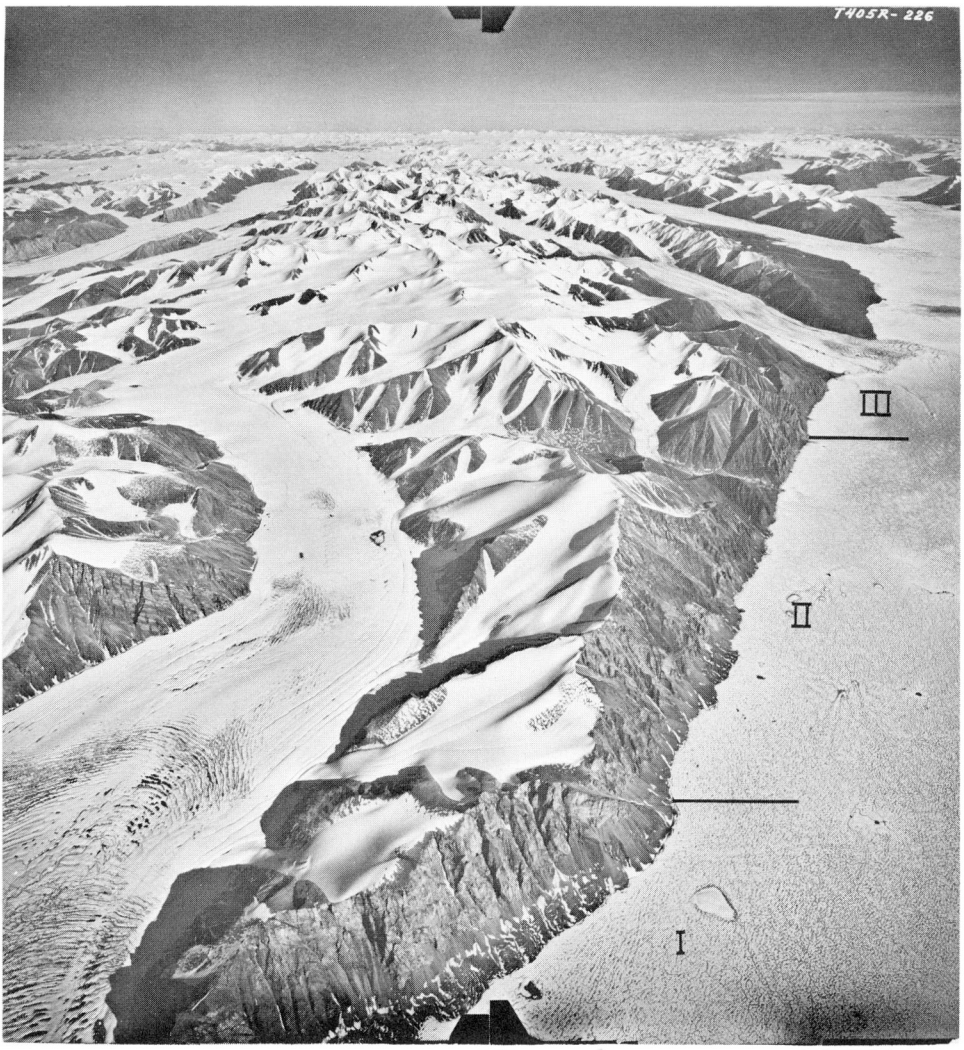
Technical Editor
E. J. W. IRISH

Critical Readers
E. W. REINHARDT
R. L. CHRISTIE
H. P. TRETTIN

Editor
MARGUERITE RAFUSE

Text printed on Georgian offset smooth (brilliant white)
Set in Times Roman with
captions in 20th Century, medium

Artwork by CARTOGRAPHIC UNIT, GSC




The eastern coastline of Yelverton Bay and Fiord, which affords the finest exposures known of metamorphic rocks of the Franklinian Geosyncline. Roman numerals refer to three subdivisions of the rocks: I – leucocratic granitic gneisses; II – mixed leucocratic and mafic gneisses and amphibolite; III – gneisses and schists, largely of metamorphic grade lower than the rocks of I and II. (Oblique air photograph) RCAF T405 R-226, July 15, 1950.

GEOLOGICAL SURVEY
OF CANADA

*BULLETIN 229*

# METAMORPHIC AND PLUTONIC ROCKS OF NORTHERNMOST ELLESMERE ISLAND, CANADIAN ARCTIC ARCHIPELAGO

By
Thomas Frisch

DEPARTMENT OF
ENERGY, MINES AND RESOURCES
CANADA

© Crown Copyrights reserved

Available by mail from Information Canada, Ottawa,
from Geological Survey of Canada, 601 Booth St., Ottawa,
and at the following Information Canada bookshops:

HALIFAX
1683 Barrington Street

MONTREAL
640 St. Catherine Street West

OTTAWA
171 Slater Street

TORONTO
221 Yonge Street

WINNIPEG
393 Portage Avenue

VANCOUVER
800 Granville Street

or through your bookseller

A deposit copy of this publication is also available for reference in
public libraries across Canada

Price: $4.00 Catalogue No. M42-229

Price subject to change without notice

Information Canada
Ottawa, 1974

Southam Murray
02KX-23387-6455

# PREFACE

A principal aim of the Geological Survey is the estimation of the potential abundance and probable distribution of mineral and fuel resources available to Canada. Such estimates depend on the availability of information concerning the geological framework.

The last two decades have seen a great increase in the geological data available for the Arctic Archipelago, much of this being derived from reconnaissance and detailed studies made by the Geological Survey. This report presents results of the first study dealing specifically with the metamorphic and plutonic rocks of the Franklinian Geosyncline. Crystalline rocks underlie areas larger than previously realized and have been involved in metamorphism and tectonism subsequent to their initial recrystallization. Isotopic evidence suggests that plutonic activity was particularly intense in Early Devonian time and, therefore, coincided with the Caledonian earth movements that previously were known to have affected only northwestern Ellesmere Island and central parts of the Canadian Arctic Archipelago.

D. J. MCLAREN,
*Director, Geological Survey of Canada*

OTTAWA, August 1973

# CONTENTS

## CHAPTER I


PAGE

INTRODUCTION ..... 1
- Location ..... 1
- Physiography ..... 1
- Climate ..... 1
- Accessibility and methods of travel ..... 2
- Geological setting of the metamorphic and plutonic rocks ..... 3
- Previous work ..... 3
- Present work ..... 4
- Acknowledgments ..... 5


## CHAPTER II


METAMORPHIC ROCKS ..... 6
- Cape Columbia Complex ..... 6
  - The gneissic rocks of the coastal exposures ..... 6
    - Banded garnet-biotite gneisses ..... 6
    - Hornblende-bearing rocks ..... 16
    - Miscellaneous metamorphic rocks ..... 19
    - Granitic and pegmatitic dykes ..... 20
    - Structural features ..... 20
  - Muscovite-rich schists south of Capes Aldrich and Columbia ..... 20
  - Metamorphic rocks of Markham Fiord ..... 23
  - Ward Hunt Island ..... 23
  - Origin and metamorphic facies of the Cape Columbia Complex ..... 25
  - Age of the Cape Columbia Complex ..... 26
- Upper M'Clintock Inlet ..... 27
  - Petrography ..... 27
  - Structural relationships and age ..... 28
- Eastern and southern Ayles Fiord ..... 29
  - The northeastern shore ..... 29
    - Schistose metamorphic rocks ..... 29
    - Granitic gneiss ..... 30
  - The head of Ayles Fiord ..... 32
  - Structural features of the Ayles Fiord region ..... 33

PAGE

Eastern and southern Ayles Fiord (*conc.*)
Correlation with other parts of the northern Ellesmere metamorphic complex ..... 33
The coastal area, western Ayles Fiord to Petersen Bay ..... 33
Northern granitic gneiss terrain ..... 33
Petersen Bay ..... 35
Amphibolites ..... 36
Mica schists ..... 36
Yelverton Bay and Fiord ..... 38
The eastern coastline ..... 38
The northern low-grade sequence ..... 38
The high- to low-grade sequence ..... 39
Zone I ..... 39
Zone II ..... 41
Zone III ..... 45
Minor intrusive rocks ..... 46
The western shore of Yelverton Bay ..... 46
Age of the metamorphic rocks of the Yelverton Fiord region ..... 47
Other areas of regional metamorphism in northern Ellesmere Island ..... 47
Metamorphic facies series in northern Ellesmere Island ..... 48

CHAPTER III

PLUTONIC ROCKS ..... 49
Mafic and ultramafic rocks ..... 49
Cape Fanshawe Martin intrusion ..... 49
Petrography and structure ..... 51
Age ..... 56
Ultramafic massifs on M'Clintock Inlet ..... 56
The M'Clintock West ultramafic massif ..... 56
The M'Clintock East ultramafic massif ..... 63
Petrogenesis ..... 64
Age ..... 65
Other ultramafic and mafic bodies ..... 66
Cape Columbia ..... 66
Bromley Island and Kleybolte Peninsula ..... 67
Felsic rocks ..... 67
Cape Richards intrusive complex ..... 67
Petrography ..... 67
Petrogenesis ..... 73
The country rocks and contact zone ..... 73
Age ..... 74
The Phillips Inlet intrusion and related rocks ..... 74
Petrography and structure ..... 74
Other intrusions in the Phillips Inlet area ..... 78
Petrogenesis of the intrusions in the Phillips Inlet area ..... 79
Age of the intrusions in the Phillips Inlet area ..... 79
Other felsic intrusions ..... 80

PAGE

REFERENCES ..... 81

Appendix 1. Terminology ..... 85
Appendix 2. Chemical analytical methods ..... 85
Appendix 3. Mineral assemblages ..... 86

INDEX ..... 87

Table I. Modal and chemical analyses of quartzo-feldspathic gneisses from the type area of the Cape Columbia Complex ..... 10
II. Modal and chemical analyses of hornblende-bearing rocks from the type area of the Cape Columbia Complex ..... 17
III. Modal analyses of muscovite-rich schists from south of Capes Aldrich and Columbia ..... 22
IV. Modal analyses of granitic gneisses from the metamorphic complex between Ayles Fiord and Petersen Bay ..... 35
V. Modal analyses of staurolite-bearing schists from Petersen Bay ..... 37
VI. Modal and chemical analyses of metamorphic rocks from the Yelverton Bay and Fiord area ..... 40
VII. Modal and chemical analyses of leucocratic rocks from the M'Clintock West massif ..... 60
VIII. Modal and chemical analyses of rocks from the Cape Richards complex ..... 70
IX. Modal and chemical analyses of rocks from the Phillips Inlet intrusion ..... 75

## Illustrations

Eastern coastline of Yelverton Bay and Fiord ..... *Frontispiece*

Figure 1. Index map showing location of study-area ..... 2
2. Banded garnet-biotite-feldspar-quartz gneiss of the Cape Columbia Complex (field photograph) ..... 7
3. "Giant augen" gneiss from the Cape Columbia Complex (field photograph) ..... 9
4. Deformed augen gneiss of the Cape Columbia Complex (photograph of hand specimen) ..... 9
5. Classification of granitic rocks ..... 9
6. Modal compositions of quartzo-feldspathic gneisses from the type area of the Cape Columbia Complex ..... 13
7A. Granular epidote in a plagioclase porphyroblast from the Cape Columbia Complex (photomicrograph) ..... 13

PAGE

Figure 7B. Muscovite flakes in a plagioclase porphyroblast from the Cape Columbia Complex (photomicrograph) ..... 13
8. Texture of garnet-biotite-feldspar-quartz gneiss from the Cape Columbia Complex (photomicrograph) ..... 15
9. Boudinaged pegmatite dykes in the Cape Columbia Complex (field photograph) ..... 21
10. Truncated minor fold in the Cape Columbia Complex (field photograph) ..... 21
11. Metamorphic and Paleozoic sedimentary rocks on the western shore of M'Clintock Inlet (field photograph) ..... 29
12A,B. High- and low-grade metamorphic rocks on eastern Ayles Fiord (field photographs) ..... 31
13. Modal compositions of granitic gneiss from between Ayles Fiord and Petersen Bay ..... 34
14. Modal composition of rocks from eastern Yelverton Fiord ..... 40
15. Gneisses and amphibolites on the eastern shore of Yelverton Fiord (field photograph) ..... 42
16. Quartzo-feldspathic lenses in gneiss, eastern shore of Yelverton Fiord (field photograph) ..... 43
17. Amphibolite layers in gneisses, eastern shore of Yelverton Fiord (field photograph) ..... 43
18. Granitic cataclasite body, eastern shore of Yelverton Fiord (field photograph) ..... 44
19. The Cape Fanshawe Martin intrusion (oblique air photograph) ..... 50
20. Classification of mafic and ultramafic rocks ..... 51
21. Ultramafic and metamorphic rocks on M'Clintock Inlet (oblique air photograph) ..... 57
22. Ternary quartz-feldspars plot of modes of leucocratic rocks from the M'Clintock West massif ..... 61
23. Ternary mafic minerals-feldspars and quartz plot of modes of leucocratic rocks from the M'Clintock West massif ..... 61
24. Harker diagram of leucocratic rocks from the M'Clintock West massif ..... 63
25. The Cape Richards complex (field photograph) ..... 68
26. Ternary quartz-feldspars plot of modes of rocks from the Cape Richards complex ..... 71
27A,B. Late albite in syenite of the Cape Richards complex (photomicrographs) ..... 71
28. Ternary quartz-feldspars plot of modes of rocks from the Phillips Inlet intrusion ..... 76
29. Geology of metamorphic and plutonic rocks, northernmost Ellesmere Island ..... *In pocket*

# METAMORPHIC AND PLUTONIC ROCKS OF NORTHERNMOST ELLESMERE ISLAND, CANADIAN ARCTIC ARCHIPELAGO


## *Abstract*

Information obtained from reconnaissance mapping between Cape Aldrich and Phillips Inlet has resulted in an extension of previously defined limits of areas underlain by metamorphic and plutonic rocks in the Northern Ellesmere Fold Belt of the Franklinian Geosyncline. Metamorphic rocks are most abundant between M'Clintock and Phillips Inlets; they occur in lesser amount between Cape Aldrich and M'Clintock Inlet. Amphibolite facies gneisses and schists tend to be restricted to the northernmost coasts, whereas greenschist facies rocks form extensive terrains inland. Metamorphism appears to have been of Barrovian type and the highest grade exposed is that of the kyanite zone. Retrogressive metamorphism, promoted by cataclasis, was particularly pronounced in the Cape Columbia area.

Three new K/Ar age determinations on metamorphic rocks from the Cape Columbia area yield Middle Ordovician to Early Devonian dates but field evidence indicates that at least part of the northern Ellesmere metamorphic complex is of pre-Middle Ordovician age. Possible remobilization of metamorphic basement on Ayles Fiord in mid-Paleozoic time is suggested by a new K/Ar age determination.

Plutonic activity appears to have been particularly active in Early Devonian time. A deformed, partly layered gabbro-peridotite body of tholeiitic affinity has intruded metamorphic rocks at Cape Fanshawe Martin. A K/Ar age determination on biotite from the intrusion indicates 376 $\pm$ 16 m.y. as a minimum age of crystallization. Two alpine-type serpentinite massifs occur near the head of M'Clintock Inlet and were intruded by small felsic sheets and plugs, one of which has been dated (hornblende K/Ar) at 390 $\pm$ 18 m.y. South of Cape Richards, a zoned complex of alkalic syenite bordered by quartz monzonite has intruded pre-Middle Ordovician metamorphic rocks. A hornblende K/Ar date of 390 $\pm$ 18 m.y. for the quartz monzonite probably gives a reliable minimum age of crystallization. A granitic pluton (biotite K/Ar age 345 $\pm$ 15 m.y.) at the mouth of Phillips Inlet is bordered by a narrow, staurolite-kyanite-sillimanite-bearing contact aureole and is probably related to nearby smaller felsic plutons (biotite K/Ar age: 325 $\pm$ 14 m.y.), which earlier have been interpreted to be mesozonal to epizonal intrusions emplaced during a Middle Devonian to Mississippian orogeny.

*Résumé*

Les données recueillies au cours des travaux de cartographie de reconnaissance entre le cap Aldrich et l'inlet Phillips ont pour résultat de reculer les limites connues des régions qui reposent sur des roches métamorphiques et plutoniennes dans la zone de plissements du nord d'Ellesmere du géosynclinal franklinien. Les roches métamorphiques se retrouvent en abondance entre les inlets M'Clintock et Phillips; elles sont moins nombreuses entre le cap Aldrich et l'inlet M'Clintock. Les gneiss et schistes à faciès à amphibolite semblent se limiter aux côtes les plus au nord alors que les roches à faciès de schiste vert couvrent de vastes étendues à l'intérieur des terres. Le métamorphisme semble avoir été de type barrovien et le plus intense qu'on puisse observer est celui de la zone de cyanite. Le métamorphisme régressif, causé par cataclase, a été particulièrement prononcé dans la région du cap Columbia.

Trois nouvelles datations au potassium argon de roches métamorphiques de la région du cap Columbia en font remonter la formation entre l'Ordovicien moyen et le début du Dévonien, mais des indices recueillis sur le terrain attestent qu'au moins une partie du complexe métamorphique du nord d'Ellesmere est antérieure à l'Ordovicien moyen.

Une nouvelle datation au potassium-argon suggère la possibilité d'un déplacement du socle métamorphique du fiord Ayles, au milieu de l'ère paléozoïque.

L'activité plutonique semble avoir été particulièrement grande au début du Dévonien. Un massif déformé et partiellement stratifié, composé de péridotite à gabbro de caractère tholéiitique a pénétré des roches métamorphiques à cap Fanshawe Martin. Une datation au potassium–argon de biotite qui provient de l'incusion donne 376 ± 16 millions d'années comme date minimale de cristallisation. Deux massifs de serpentine de type alpin se dressent près de la pointe de l'inlet M'Clintock et ils ont été pénétrés par des petits lits et bouchons de felsite et on a établi l'âge de l'un d'eux (hornblende potassium–argon) à 390 ± 18 millions d'années. Au sud du cap Richards, un complexe de syénite alcaline entouré de monzonite à quartz a pénétré des roches métamorphiques antérieures à l'Ordovicien moyen. L'âge de la monzonite à quartz qu'on a fixé à 390 ± 18 millions d'années par la datation au potassium–argon de la hornblende donne une idée précise de la date minimale de cristallisation. Un pluton granitique (âge de la biotite fixé à 345 ± 15 millions d'années par datation au potassium–argon) à l'entrée de l'inlet Phillips est bordé d'une étroite auréole de contact à staurolite et à sillimanite, et ce pluton est probablement apparenté à des plutons de felsite, voisins et plus petits (âge de la biotite par datation au potassium–argon: 325 ± 14 millions d'années) qu'on croyait être des inclusions mésozonales ou épizonales mises en place pendant une orogénèse qui eut lieu du Dévonien moyen jusqu'au Mississippien.

# *Chapter I*

# INTRODUCTION

## Location

Ellesmere Island is the largest and most northerly of the Queen Elizabeth Islands, District of Franklin, Northwest Territories, Canada. This report describes part of the geology of the island's northern coast between, approximately, latitude 83°N and longitude 70°W, and latitude 82°N and longitude 87°W (Fig. 1).

## Physiography

The northern coastal region of the island is a rugged, fiord-dissected mountain terrain of moderate relief. On the coast itself, some mountains attain elevations of about 3,500 feet but most are from 2,000 to 3,000 feet high. The longer fiords reach the edge of the United States Range, which is largely obscured by a snowfield and whose highest peaks lie between 8,000 and 10,000 feet. Mountains flanking the range rise to more than 5,000 feet and, in general, gradually diminish in height northward toward the coast.

Although generally moderate, the relief locally may be considerable as, for example, in trunk glacial valleys draining northward from the ice-cap of the United States Range. Numerous lower valleys are occupied by glaciers or by outwash and braided streams. Higher valleys and gullies are almost invariably filled with permanent ice and snow. Most of the ice-free parts of the mountains are covered by talus and felsenmeer.

## Climate

The climate appears to be variable over the north coast of the island; it is certainly not so severe as that much farther south on the northern Canadian mainland. Snowfall, although greater than in other parts of the northern Arctic Archipelago, is relatively light even in winter (probably much less than 12 inches precipitation per year) and is rare in summer, when maritime conditions frequently prevail and give rise to rain and fog over the coastal region.

The melt season begins in early June and conditions for air-supported geological traversing are suitable from late June or early July until about the middle of August, when poor weather and low sun often hinder flying. Melting may start inland well before it does on the coast, where north-facing slopes may remain snow-covered into July.

Daylight is, of course, continuous throughout the spring and summer. Temperatures during the field season seldom fall below freezing in July and early August. Mean temperature is 40°–45°F but temperatures as high as 70°F were recorded inland near M'Clintock Inlet in 1966.

Original MS. submitted February 19, 1970.
Final version approved for publication August 1973.

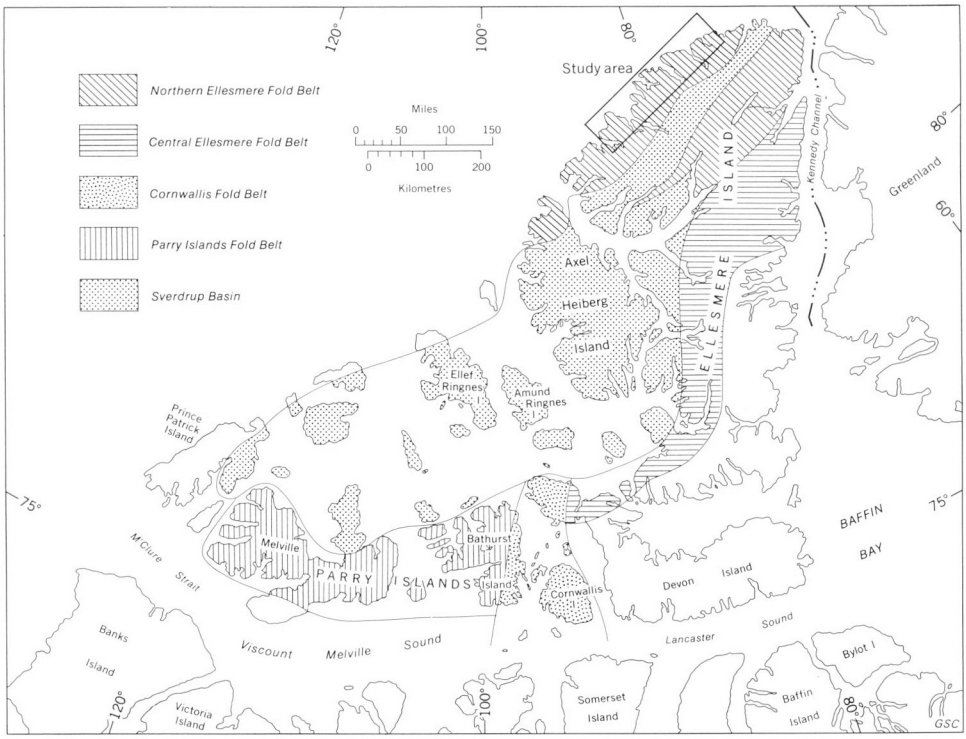



FIGURE 1. Index map showing location of the study-area, and the four fold belts of the Franklinian Geosyncline and Sverdrup Basin (adapted from Douglas *et al.*, 1963, Fig. 2).

## Accessibility and Methods of Travel

The field area may be reached by air from Alert, a weather station on the northeastern extremity of Ellesmere Island, where an airstrip is maintained the year round. A similar airstrip at Eureka weather station, 150 to 300 miles south of the field area, also is accessible by ship most years. The nearest airport with commercial air transport is Resolute, 600 miles to the southwest of the area.

Alert and Eureka weather stations are the only permanently occupied settlements in the northern parts of Ellesmere Island. Several small scientific field stations have been built both within and near the field area, and are intermittently occupied by parties of the Defence Research Board of Canada. These stations, semi-permanent huts with prepared or marked-out airstrips, are of great value in this mountainous region as landing points for light, fixed-wing aircraft. One has been built on Ward Hunt Island, and another at Lake Hazen; a third, at the head of Tanquary Fiord, is accessible by ship most years.

Geological work within the field area is best supported by helicopter, but a STOL (short take-off and landing) light aircraft such as the Piper Super Cub can be used in many places. Such an aircraft should be specially adapted for landings on unprepared ground—that is, equipped with a strengthened undercarriage and oversize tires and brakes.

The ground is generally firm during most of July and August and this, combined with lack of vegetation more than a few inches high, makes walking easy. It can be particularly

good on the sea ice (fiords remain frozen the year round), but in places shore leads inhibit access to the ice from the land. Nevertheless, ruggedness of the terrain and the abundance of permanent snow and ice render many areas, for practical purposes, inaccessible for travel on foot in the time usually available.

## Geological Setting of the Metamorphic and Plutonic Rocks

The main points of the geologic history of the Canadian Arctic have been summarized by Thorsteinsson and Tozer (1970). The following is a condensation of their and other work on the geology of northern Ellesmere Island.

The Franklinian Geosyncline, originally recognized by Schuchert (1923), was the site of sedimentation between late Precambrian and Late Devonian times. Its exposed part forms a reverse "L", which trends easterly through the Parry Islands and then northeasterly through Ellesmere Island (Fig. 1). The geosyncline may be divided into a northwestern eugeosynclinal belt and a miogeosyncline to the south and east.

Four fold belts constitute the geosyncline: the Parry Islands, Cornwallis, Central Ellesmere, and Northern Ellesmere belts (Fig. 1, and Douglas *et al.*, 1963, p. 10). With the exception of widespread hypabyssal mafic intrusions, metamorphic and plutonic rocks are known only in the northern part of the Northern Ellesmere Fold Belt. Also restricted to this belt are sedimentary and volcanic rocks of eugeosynclinal character.

An early Paleozoic or older event produced the Cape Columbia Complex of metamorphic rocks in its type area. Unmetamorphosed Silurian, Ordovician (and older?) clastic, carbonate and volcanic rocks and weakly metamorphosed equivalents were tightly folded in mid- and/or early Paleozoic time. A Middle Ordovician or older event is suggested by a marked angular unconformity between low-grade metamorphic rocks and Middle Ordovician sediments in the M'Clintock Inlet area (Trettin, 1969b).

A Late Silurian to Early Devonian (Caledonian) event has been recognized in northern Axel Heiberg Island (Trettin, 1969a), the region of the Cornwallis Fold Belt and Boothia Uplift (Kerr and Christie, 1965), and in East Greenland (Haller and Kulp, 1962). Although no direct evidence of such an event has been found in northern Ellesmere, the youngest pre-Carboniferous strata known there being of Ludlovian age, it is entirely possible that this region was affected also by a Caledonian orogeny. Most isotopic ages from northern Ellesmere fall into this time interval.

A Middle Devonian to Early Mississippian orogeny, accompanied by plutonism and, in the Phillips Inlet area, by regional metamorphism, has been inferred in northern Axel Heiberg Island and northwestern Ellesmere by Trettin (1969a).

The highly deformed lower Paleozoic and older(?) rocks are overlain with pronounced angular unconformity by Pennsylvanian, Permian, and Mesozoic strata of the Sverdrup Basin. The upper Paleozoic and Mesozoic beds were deformed mostly by faulting, probably in Tertiary time.

Unmetamorphosed plutonic rocks have been found to intrude only metamorphosed rocks and lower Paleozoic strata, and it seems likely that they are all of early to mid-Paleozoic age. Isotopic age determinations reported here bear out this interpretation.

## Previous Work

The first mention of metamorphic rocks on northern Ellesmere Island appears in the geological report of the Nares Expedition (Feilden and de Rance, 1878). Their accompanying

geological map shows "mica schists and altered rocks" in the region of Cape Columbia but the rocks are not described in the report. Feilden and de Rance (1878, p. 559) also noted "green slates and metamorphic rocks belonging to the Cape Rawson beds" on Feilden Peninsula.

Peary (1907, p. 231) mentions a "granite erratic" on Cape Aldrich. It was not until 1953 that a geological survey of the northern coast of Ellesmere Island was made. This was a coastline reconnaissance between Alert and Markham Fiord by R. G. Blackadar (1954) of the Geological Survey of Canada, who confirmed the presence of metamorphic rocks in the Cape Aldrich–Markham Fiord area. He briefly described various types of garnet, biotite, and granitic gneisses and named them the Cape Columbia Group. One of these rocks was later radiometrically dated (Blackadar, 1960).

In 1954, R. L. Christie (1957, 1964), also of the Geological Survey of Canada, extended the reconnaisance westward. He discovered, in addition to further outcrops of metamorphic rocks, intrusions of granitic and ultramafic-mafic rocks at various points along the coast. Christie's work, like Blackadar's, was a coastline reconnaissance by dog-sled of a very large area in the space of a spring and summer.

The United States Air Force Cambridge Research Laboratories sponsored a study of the Ward Hunt Ice Shelf in 1960 and, as part of this program, J. B. Lyons and F. G. Leavitt (1961) investigated the geology of Ward Hunt Island. Their map shows schists and amphibolite, which they assigned to the Cape Columbia Group, and gneissic syenite. These rocks are not described in their report.

During his study of the lower Paleozoic eugeosynclinal rocks in northwestern Ellesmere Island, H. P. Trettin (1964, 1969a, and 1971), of the Geological Survey of Canada, found isolated areas of metamorphic rocks and various intrusive rocks. In 1965 and 1966, Trettin (1966, 1967, 1969b, and 1971) mapped granitic, mafic, and ultramafic intrusions in the Phillips Inlet and M'Clintock Inlet areas.

## Present Work

The investigation with which this report is concerned was part of the Geological Survey of Canada's "Operation Grant Land," a reconnaissance study of the geology of northern Ellesmere Island directed by R. L. Christie. The writer was assigned responsibility for the metamorphic and plutonic rocks in the field area.

Field work in 1965 began on June 21 and ended on August 11. Base camp was established at Alert and field work was carried out from one- or two-man fly camps in the area between Cape Aldrich and Ayles Fiord. Camp moves were made by helicopter and Piper Super Cub. Traversing was mostly done on foot, but a few reconnaissance flights, with occasional sampling stops, were made by helicopter.

In 1966 the method of field work was similar, but with the base camp located near the northern end of Lake Hazen. Field work began on June 27 and ended on August 10, and was confined to the region between M'Clintock Inlet and Phillips Inlet. A helicopter was used for all camp moves.

In 1967 the author was able to spend a week near the head of M'Clintock Inlet, aerial transport from Alert being provided by the Polar Continental Shelf Project.

Laboratory studies by the writer involved the examination of about 300 thin sections and the determination of optical properties of selected minerals. X-ray and isotopic age determinations, chemical analyses and photomicrographs were made by staff members of the Geological Survey of Canada.

Modal analyses were carried out with a Swift electric point counter, 1,000 to 2,000 points being counted in a given thin section.

Compositions of all plagioclase in the metamorphic rocks, and of plagioclase less calcic than $An_{50}$ in the plutonic rocks, were determined on the universal stage, using low-temperature curves for $n'_{\alpha} \wedge (010) \perp 100$ given by Burri, Parker and Wenk (1967). In order to confirm that the plagioclase was of the low-temperature variety, spot checks of structural state were made by the method of Rittmann and El-Hinnawi (1961). Turner's curves as modified by Slemmons (1962) and Uruno's (1963) curves were employed to determine composition and structural state of plagioclase more calcic than $An_{50}$. The determinative curves of Tröger (1959) were frequently used as supplementary aids.

Measurements of optic angle were made by conoscopic methods on the universal stage and are correct to $\pm 2°$. Refractive indices were measured on crushed grains in immersion oils and are correct to $\pm 0.001$.

Forsterite content (in mol. per cent) of olivine was determined with a small-diameter X-ray powder camera by measurement of $d_{174}$ spacings (Jambor and Smith, 1964).

Details of chemical analysis are given in Appendix 2.

## Acknowledgments

The results of this study were presented in a doctoral dissertation at the University of California, Santa Barbara. Facilities for laboratory work were provided by the Geological Survey of Canada, Ottawa, and by the university. Thanks are due A. C. Waters, C. A. Hopson, and W. S. Wise, who served on the thesis committee.

Competent help and cheerful companionship in the field were provided by John Taylor and Robert Wilson in 1965, and by Robert Keith and Pierre LaBreque in 1966.

In 1965, J. Cline piloted a Hiller 12E helicopter of Highland Helicopters, Ltd., and J. Strickland a Piper Super Cub of Bradley Air Services, Ltd. In 1966, L. W. Myers flew a Bell G-4 helicopter of Autair, Ltd. and A. Mackenzie and S. Stanish a Super Cub and De Havilland Beaver, respectively, of Bradley Air Services, Ltd. All flew their aircraft skillfully, often in adverse conditions.

Field work in 1967 was made possible through the co-operation of E. F. Roots, then co-ordinator of the Polar Continental Shelf Project, and his team at Alert in providing transportation and accommodation in the field.

## *Chapter II*

# METAMORPHIC ROCKS

The metamorphic rocks described here occur in three main regions separated by areas of unmetamorphosed younger strata: (a) Cape Aldrich to Ward Hunt Island; (b) upper M'Clintock Inlet; and (c) a large area extending from Ayles Fiord to Phillips Inlet. These three areas include the bulk of the strongly metamorphosed rocks of the Franklinian Geosyncline.

First described is the Cape Columbia Complex, the type area for which lies within the Cape Aldrich to Ward Hunt Island metamorphic region. Other metamorphic rocks of the north coast of Ellesmere Island are described under "locality" or "area" headings.

A list of mineral assemblages found in prograde regionally metamorphosed rocks of northern Ellesmere Island is given in Appendix 3.

## Cape Columbia Complex

(Fig. 29A and B; map-units 1a, 1b)

The use of the name Cape Columbia Complex is restricted to the type locality and environs (*see* Fig. 29A, B). This avoids uncertain correlations with the wide variety of metamorphic rocks on the north coast of Ellesmere Island. First described are the coastal, or type exposures of the Cape Columbia Complex; second, adjacent rocks assumed to be part of the Complex including schists exposed south of Capes Aldrich and Columbia; third, metamorphic rocks of Markham Fiord; and fourth, the crystalline rocks of Ward Hunt Island.

### The Gneissic Rocks of the Coastal Exposures

(Map-unit 1a)

These rocks constitute the Cape Columbia Group as first described by Blackadar (1954). Trettin (1969a, p. 28) later redefined the unit as the Cape Columbia Complex.

The sea-cliffs of the Cape Aldrich–Cape Columbia region afford excellent outcrops, whereas mountain-tops and slopes are heavily mantled by felsenmeer and talus. Frost-heaving is intense (Fig. 2), and the study of minor structural elements is hindered by numerous dislodged blocks of rock.

A wide variety of metamorphic rocks is present. Foliation strikes generally northwest and dips vertically or steeply to the southwest. Gneisses predominate over schists; quartzite and marble are comparatively rare; and pegmatite dykes are common (for definitions of terms, *see* Appendix 1). Two ubiquitous features of the rocks are marked post-crystalline deformation and significant retrogressive metamorphism.

#### *Banded Garnet–Biotite Gneisses*

Banded, garnetiferous, biotite-rich gneiss appears to be the most abundant rock in the area. The banding is an expression of alternating leucocratic quartzo-feldspathic and darker

biotite-quartz-feldspar layers. Texture and structure vary considerably, even in a single hand specimen, as is shown by Figure 2. Grain size (excluding that of porphyroblasts) ranges from fine to medium (these terms defined in Appendix 1). The grain size generally is dependent on the amount of crushing the rock has undergone. In addition to vertical variations, changes in fabric occur laterally within a given layer: thickening and thinning, lensing, and pinch-and-swell structures are very common.

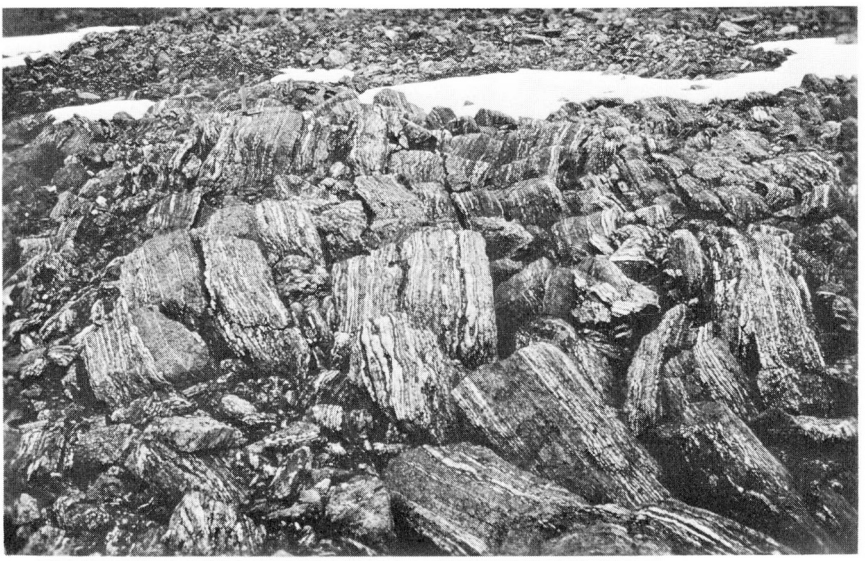
Most of the banded gneisses display feldspar porphyroblasts; their size depends largely on the amount of deformation and recrystallization. Undeformed porphyroblasts are commonly 0.5 to 10 mm long, but within the typical gneisses are numerous intercalations of spectacular rocks containing porphyroblasts commonly 3 to 4 inches but in places up to 7 inches long (Fig. 3). These coarsely porphyroblastic or "giant augen" gneisses were not studied in detail but some general conclusions as to their nature can be drawn from the available data.

There are at least two generations of feldspar porphyroblasts: one formed before major deformation, another during or after deformation. The actual age difference between the two may be slight.

The earlier generation is represented by augen that are lensoid or ovoid. The foliation, defined by quartzo-feldspathic and biotite-rich layers, wraps around the augen. Many crystals have been rotated and broken and subsequently were partly replaced by groundmass minerals. That these large crystals are true metamorphic growths (porphyroblasts) and not phenocrysts inherited from a metamorphosed porphyritic intrusive rock is suggested by:

(1) their restriction to layers within finer grained gneisses, into which the layers grade; and

(2) the uneven distribution and size of the augen within the layers (Fig. 3).



202022

FIGURE 2. Typical bedrock exposure of banded garnet-biotite-feldspar-quartz gneiss between Capes Aldrich and Columbia, the type area of the Cape Columbia Complex. Note the irregular shape and distribution of the leucocratic layers, and the intense frost-heaving. Hammer for scale, in top of picture. (Field photograph.)

Porphyroblasts of a later generation are not as common. These are generally extremely large and lack fractures and other signs of significant deformation; foliation has not been displaced around them. Their growth may have kept pace with, and even outstripped the movement that created the foliation, so that the porphyroblasts grew essentially without disturbance in an active environment.

Other examples of syntectonic porphyroblastesis are euhedral crystals elongated parallel to the foliation. These crystals are optically homogeneous (i.e., extinction is uniform) and contain drop-like inclusions of unstrained quartz. Quartz outside the porphyroblasts is commonly highly strained.

Late cataclasis locally has affected all porphyroblastic rocks and tends to obscure age relationships of porphyroblasts. Both augen and groundmass have been granulated. Cataclasis of porphyroblasts is most commonly shown by granulated crystal margins but in places whole crystals have been crushed and elongated and "tails" with mortar texture have developed.

Certain layers within the augen gneiss have been intensely deformed. Figure 4 shows the resulting rock, which consists of stretched and folded lenses of recrystallized plagioclase, potassium feldspar, and quartz. Because the augen gneisses grade into this rock, at least some of the quartzo-feldspathic lenses of the latter are considered to be recrystallization products of highly deformed feldspar augen.

Most, and perhaps all, of the "giant augen" are microcline perthite, according to field, staining, and microscopic studies. The groundmass of the "giant augen" gneisses is poor in potassium feldspar and is similar to that of the much more abundant plagioclase-porphyroblastic gneiss, described below.

Invariably accompanying the large potassium feldspar porphyroblasts are plagioclase porphyroblasts of smaller size (1 inch or less). They appear to have formed prior to deformation and, in thin section, most appear to be porphyroclasts rather than porphyroblasts. These crystals are cracked and minute grains of recrystallized plagioclase fill the fissures.

### *Prograde Plagioclase-Porphyroblastic Gneisses*

Few of these rocks are completely free from the effects of cataclasis and attendant retrogressive metamorphism, but pre-cataclasis mineralogy and texture are preserved in some specimens. A modal analysis of the latter is given in Table I (no. 1). According to the classification of granitic[1] rocks used in this report (Fig. 5), the rock is of quartz dioritic composition (Fig. 6).

*Plagioclase* in porphyroblasts and groundmass is of identical composition: oligoclase ($An_{22-26}$); rarely, compositions as calcic as $An_{32}$ were noted. Zoning is fairly common in the porphyroblasts but absent in groundmass crystals. The zoning is generally normal and compositional ranges are narrow (a difference of 3% to 4% An between core and rim). Instances of reverse zoning were found, but these are rare. Replacement anti-perthite texture was noted in one porphyroblast and may be a common feature of the larger crystals; however, their microscopic study is hindered by their great size. Sericitization or other alteration is slight to non-existent, except in more deformed rocks, wherein plagioclase porphyroblasts have been muscovitized (*see* below). The plagioclase of leucocratic layers in the banded gneiss is discussed in a later section.

*Potassium feldspar* forms porphyroblasts in some of the banded gneisses but is generally confined to the groundmass. Porphyroblasts are perthitic microcline, with fine cross-hatched

[1]For definition of this term, *see* Appendix 1.

202022-A

FIGURE 3

Boulder of "giant augen" gneiss from the Cape Columbia Complex, west of Cape Aldrich. Most, if not all, of the large augen are potassium feldspar; most of the smaller ones are plagioclase. (Field photograph.)

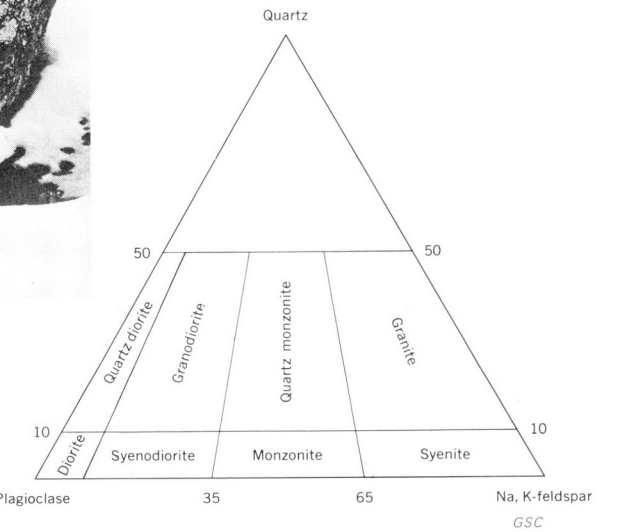
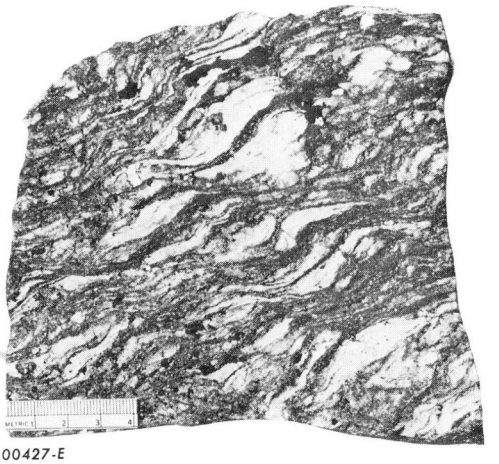



FIGURE 5

Classification of granitic rocks used in this study (*see also* Appendix 1).



200427-E

FIGURE 4

Intensely deformed augen gneiss, a common rock in the Cape Columbia Complex between Capes Aldrich and Columbia. The original feldspar augen have been compressed, stretched, and recrystallized into lenses of granular plagioclase, potassium feldspar, and quartz. (Photograph of hand specimen.)

TABLE I

*Modal and chemical analyses of quartzo-feldspathic gneisses from the type area of the Cape Columbia Complex.*

| | 1 | IV | 2 | 3 | 4 | 4V | 5 |
|---|---|---|---|---|---|---|---|
| Plagioclase | 40.1 | 52.0 | 35.4 | 35.4 | 42.4 | 60.5 | 27.1 |
| K-feldspar | – | – | – | 2.2 | 4.5 | 7.8 | 35.2 |
| Quartz | 26.1 | 46.6 | 28.6 | 34.4 | 21.6 | 29.4 | 22.7 |
| Garnet | 7.2 | – | 10.4 | 7.1 | 7.1 | – | 4.0 |
| Biotite | 24.8 | 1.4 | 7.7 | tr | 16.8 | 2.3 | 6.4 |
| Muscovite | – | – | 17.3 | 3.3 | – | – | 3.6 |
| Chlorite | – | – | – | 12.7* | – | – | – |
| Epidote | – | – | tr | 2.8 | 4.8 | – | – |
| Apatite | tr | tr | 0.6 | – | 1.0 | tr | tr |
| Zircon | tr | – | tr | – | tr | – | – |
| Carbonate | – | – | – | 2.1 | – | – | – |
| Allanite | tr | – | – | – | – | – | 1.0 |
| Sphene | 1.8 | – | tr | – | 1.9 | – | tr |
| Black Fe oxide | tr | – | tr | tr | tr | – | – |
| *after biotite | | | | | | | |
| Points counted | 1986 | 1704 | 2003 | 1990 | 1952 | 1730 | 2017 |
| Plagioclase comp. | $An_{24}$ | $An_{24}$ | $An_{18}$ | $An_{20}$ | $An_{19}$ | $An_{19}$ | $An_{20}$ |
| | | | | | | | |
| $SiO_2$ | | | 64.4 | 73.4 | 65.3 | | |
| $TiO_2$ | | | 0.98 | 0.08 | 0.72 | | |
| $Al_2O_3$ | | | 18.1 | 14.8 | 15.6 | | |
| Tot. Fe as $Fe_2O_3$ | | | 8.2 | 5.0 | 5.8 | | |
| MnO | | | 0.14 | 0.22 | 0.10 | | |
| MgO | | | 2.3 | 0.5 | 2.1 | | |
| CaO | | | 1.2 | 1.4 | 2.7 | | |
| $Na_2O$ | | | 2.0 | 2.3 | 3.7 | | |
| $K_2O$ | | | 4.6 | 4.6 | 4.3 | | |
| Total | | | 101.9 | 102.3 | 100.3 | | |

Analysts: Rapid Methods Group, Geological Survey of Canada.

1. Prograde banded garnet-biotite gneiss.
IV. Leucocratic vein in 1.
2,3,4. Retrograded banded garnet-biotite gneisses.
4V. Leucocratic vein in 4.
5. Quartz-plagioclase microcline gneiss.

*GSC*

twinning and angles 2V of 68° to 74°. The perthite is of the stringlet or bead type (Deer, Howie, and Zussman, 1963b, p. 68). Many porphyroblasts are undeformed and probably represent relatively late potassium metasomatism. Even in these rocks, potassium feldspar is a minor constituent of the groundmass. Here also it appears to be a late growth, being generally restricted to discrete leucocratic layers, where it is intricately intergrown with plagioclase and quartz. Twinning is lacking and myrmekite is not seen in these fine-grained layers.

Some of the plagioclase-porphyroblastic gneisses contain no potassium feldspar of any kind. The very irregular distribution of potassium feldspar—ranging from large porphyroblasts in the "giant augen" gneisses to trace amounts in the groundmass of other rocks—suggests that the feldspar originated from granitizing solutions and did not simply grow from original rock material in a closed system.

*Quartz* is always a major mineral in the rock and appears to have been very sensitive to deformation. Even when all the other minerals appear to be undeformed, quartz grains are more or less strained, as shown by undulatory extinction. In some specimens, quartz has tended to aggregate in clusters, which probably represent recrystallization aggregations formed during metamorphism. Unstrained, drop-like quartz is a common inclusion in feldspar porphyroblasts.

Most *biotite* has X = pale yellow and Z = deep reddish brown, which suggests relatively high $TiO_2$ and low $Fe_2O_3$ contents (Deer, Howie, and Zussman, 1962b, p. 71), and a preferred orientation parallel to the foliation.

*Muscovite*, where present, is invariably subordinate in amount to biotite and younger. Most of it occurs intergrown with biotite and lacks the preferred orientation of the latter, which suggests that muscovite grew at a later time.

*Garnet* is a ubiquitous mineral in this group of rocks. Unit cell edges and refractive indices of garnets from two rock specimens are 11.611Å, 1.774 and 11.619Å, 1.782, respectively. According to Sastri's (1962, Pl. 35) diagram relating these parameters to composition, the garnets are almandine, consisting chiefly of the almandine ($>50\%$) and grossular ($\sim 30\%$) molecules. Grain size within a single rock sample varies widely but subhedral porphyroblasts (2–3 mm in diameter, on the average; maximum 3 cm) are always present. Most crystals show deformation features and have clearly formed before or during cataclastic deformation. In several rocks, "snowball" or "pinwheel" garnet, with well-developed spiral inclusion trains, attests to syntectonic growth of the mineral. In contrast, post-tectonic growth is suggested by inclusion-free rims around sieved cores of garnets. Possible further evidence of a complex history is provided by garnets in which anisotropism varies regularly from core to rim; this may be an expression of chemical zoning.

Among the accessory constituents, *iron opaque* minerals are rare; a typical thin section contains only a few scattered specks of Fe-Ti oxide and hematite.

*Epidote* is generally present, and *sphene* and *zircon*, associated with biotite, are ubiquitous. The epidote is generally restricted to an overgrowth on brown allanite. Some of this "epidote" may actually be clinozoisite, because of yellow-blue interference colour and positive sign (Deer, Howie, and Zussman, 1962a, p. 184), but other grains with a similar interference colour are optically negative. Highly birefringent grains are invariably negative and, therefore, are definitely epidote.

Dark brown, nearly isotropic *allanite*, with frayed edges and no epidote rims, is present in some rocks but bleached allanite, mantled by epidote (clinozoisite?), is more common.

It is possible that the allanite may be of significance as regards the origin of its host rock. Deer, Howie, and Zussman (1962a, pp. 217–18) imply that allanite is most abundant in granitic and pegmatitic rocks. Rein (1953) made a close study of allanite in the gneisses of the Black Forest and concluded that its development was largely a metamorphic process during granitization. If allanite in the Cape Columbia Complex is a primary (pre-metamorphic) mineral, it may indicate that the host rock was granitic before metamorphism; or it may simply be a relict detrital mineral. Formation of allanite during metamorphism is not considered likely in the present instance—its distribution is too haphazard and fresh crystals are absent.

### *Retrograde Plagioclase-Porphyroblastic Gneisses*

Among the rocks of the Cape Aldrich–Cape Columbia area, distinct retrogressive metamorphic features are the rule rather than the exception. Almost every specimen of banded gneiss studied shows at least minor retrograde effects. With increasing retrogressive metamorphism, the minor effects are enhanced and new ones appear. Relations between texture and mineralogy make it obvious that cataclastic deformation greatly favoured retrograde mineral reaction. Some of the more significant features are described below.

In Table I modes of progressively and retrogressively metamorphosed rocks of similar original (metamorphic) mineralogy are listed. Figure 6 shows their position in the quartz-Na,K-feldspar-plagioclase diagram. The major differences between the two rock groups are as follows.

The anorthite content of *plagioclase* decreases with retrogressive metamorphism. Porphyroblasts and finer grained, presumably recrystallized, groundmass crystals range from $An_{17}$ to $An_{22}$; again, there is no detectable compositional difference between porphyroblast and groundmass plagioclase. Porphyroblasts are commonly riddled with grains of epidote (Fig. 7A). This feature is so characteristic that accumulations of tiny grains of epidote in aggregates of fine-grained quartz and feldspar are interpreted as marking the sites of former plagioclase porphyroblasts, now recrystallized. The formation of epidote is surely related to the decalcification of plagioclase: calcium was expelled and the other elements plus water, necessary for epidote to form, were doubtless available as a result of other retrograde reactions. In contrast to the porphyroblasts, the groundmass plagioclase is fresh.

A second striking feature of the retrogressive metamorphism is the appearance of *muscovite* as a major mineral in some rocks. Muscovite has commonly formed by replacement of red-brown biotite. Muscovite flakes are commonly three to four times larger than biotite flakes and fresher looking; many lie across the foliation, whereas biotite shows strong preferred orientation parallel to the foliation. The replacement of biotite by muscovite may be an alteration reaction. Biotite in contact with muscovite is bleached in places and black iron oxides are associated. These features indicate that mafic elements have been removed during the biotite-to-muscovite transformation. Müller (1966) has assembled impressive petrographic and chemical evidence that biotite changes deuterically into muscovite in granitic rocks. This supports the theory of numerous workers that, in many rocks, muscovite has developed secondarily from biotite.

In the more deformed gneisses, muscovite also occurs as newly formed, minute flakes lying between grains of recrystallized quartz and feldspar and along crystallographic planes in plagioclase porphyroblasts (Fig. 7B).

*Garnets* occur in all stages of break-up, from rounded porphyroblasts cut by a few fractures to finely granular remains of larger crystals strung out along foliation planes.

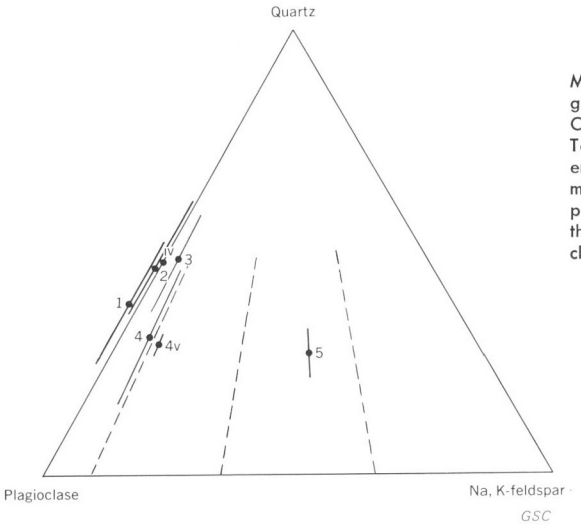
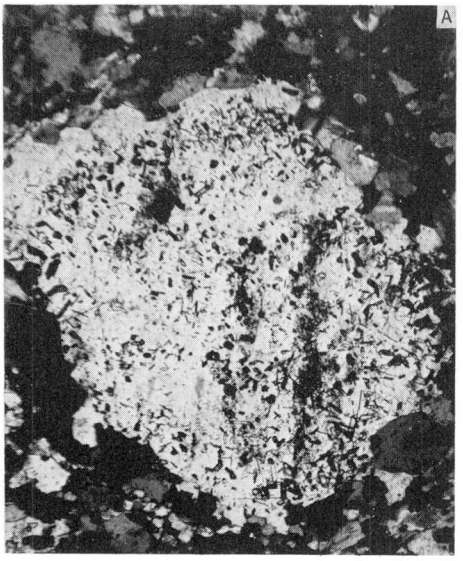
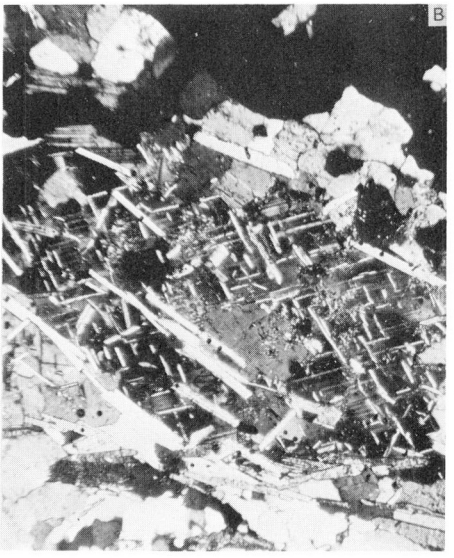



FIGURE 6

Modal compositions of quartzo-feldspathic gneisses from the type area of the Cape Columbia Complex. Numbers are those in Table I. The vertical distance between the ends of each line gives the percentage of mafic minerals. Points plotting on the quartz-plagioclase sideline have been shifted to the left for legibility. Broken lines denote classification (Fig. 5).




202022-B

202022-C

FIGURE 7. Retrograde effects in plagioclase from gneisses of the Cape Columbia Complex.

A. Granular epidote in a plagioclase porphyroblast. The epidote formed during recrystallization of the plagioclase from a composition of probably $An_{25}$ to $An_{19}$. Crossed polarizers x23. (Photomicrograph)

B. Oriented, newly formed muscovite flakes in a plagioclase porphyroblast. Crossed polarizers x70. (Photomicrograph)

Biotite and quartz are the chief minerals replacing garnet. Chlorite, the common alteration product of garnet, is extremely rare, and this may be evidence that the alteration took place at higher temperatures than those at which chlorite is formed. The following observations support this inference:

(1) The biotite that replaces garnet is of the red-brown variety. In one rock, garnet has been replaced by red-brown biotite, which in turn is partly altered to greenish brown biotite. If, as seems likely, the biotite colour is due to high titania and the concentration of Ti in biotite increases with increasing metamorphic grade (a common, albeit not universal, finding, e.g. Guidotti (1970)), then the red-brown biotite is a relatively high-temperature mineral.

(2) Biotite that is apparently unrelated to garnet in retrograde rocks is not visibly different from that in prograde rocks.

The replacement of garnet by red-brown biotite is particularly noticeable in cataclastic rocks and probably is due largely to deformation (at high temperature?) rather than to hydrothermal activity (at lower temperature?). Except for loss of calcium, which apparently is relatively abundant in the almandine garnet (*see* above), and addition of alkalies and water, the almandine-to-biotite alteration probably involved little loss or gain of material. The platy biotite formed because it was physically more stable during cataclasis than the granular garnet.

In those rocks wherein much garnet has been replaced by biotite, *epidote* is a common constituent. It is invariably associated with biotite, which has overgrown it in many places and so cannot be parent to the epidote. Development of the epidote is evidently connected with the alteration of garnet. Perhaps it formed to accommodate calcium released in the garnet-to-biotite alteration.

In incompletely altered rocks, relict textures indicate that *chlorite* (generally penninite) has formed from the biotite remaining after muscovitization. In highly cataclastic rocks, where alteration has been more intense, chlorite has been abundantly developed and biotite may be absent altogether. Also present is *low-iron epidote* (negative sign, low birefringence), in prisms 1 to 2 mm long lying parallel to the foliation; it appears to have formed syntectonically. No doubt cataclasis favoured retrograde reactions and access of water.

*Potassium feldspar* is a common minor constituent of the retrogressively metamorphosed gneisses. It occurs in intergrowths with fine-grained, recrystallized plagioclase and quartz and is probably a product of one or more retrograde reactions. Where biotite has been extensively chloritized (e.g., Table I, no. 3), formation of potassium feldspar may be related to this alteration, as suggested by Chayes (1955).

Cataclastic textures in the retrograde garnet-biotite gneisses are well developed. Foliation is defined not only by alignment of platy minerals but also by layers of recrystallized quartz and/or feldspar (Fig. 8). "Plättung"-quartz (Turner and Weiss, 1963, p. 452) is conspicuous in many rocks. Even some of the cross-muscovite is kinked, which indicates that movement continued past the late stages of recrystallization.

### *Leucocratic Layers*

The leucocratic parts of the gneisses are concordant, vein-like masses that thicken and thin along strike (Fig. 2). With few exceptions, they are separated from the host rock by zones of biotite-rich gneiss, whose width is about one-quarter that of the vein.

202022-D

FIGURE 8

Texture of a typical garnet-biotite-feldspar-quartz gneiss from the Cape Columbia Complex. Note the strong cataclasis, marked by the flattened quartz grains and lenses and the granulated garnet. Crossed polarizers x23. (Photomicrograph.)

Modal analyses of two typical veins and their host rocks are listed in Table I and plotted on the quartz-Na,K-feldspar-plagioclase triangle of Figure 6. The veins represent concentrations of the light coloured minerals in the host and are essentially free of ferromagnesian constituents.

Plagioclase compositions in the two rock types are the same and there are no perceptible differences in optical properties between other minerals that occur in both. Grain size, however, does differ: grains in the veins are several times larger than in the host (1–2 mm as against 0.2–0.5 mm).

It seems likely that the veins are composed of material extracted from the host rock and do not represent infusions from afar. Furthermore, they were probably formed by solid diffusion rather than by partial melting of country rock, whereby more sodic plagioclase would have been produced. The mafic rims represent host rock from which the leucocratic material was removed.

This interpretation is supported by detailed petrographic and chemical work by Kretz (1966) on rocks very similar to those described here. Kretz's samples were plagioclase-quartz veins in a garnet-biotite-plagioclase-quartz gneiss; potassium feldspar was a minor constituent. Compositional similarity between minerals of host and vein and loss-gain calculations for host versus vein strongly indicate that the vein material was derived from the host.

### *Potassium Feldspar-bearing Gneisses*

Among the plagioclase-rich garnet-biotite gneisses are layers of more leucocratic, non-porphyroblastic gneiss with abundant potassium feldspar. They are distinguished from their porphyroblastic, strikingly banded associates by a pink or grey colour and a more homogeneous appearance.

A modal analysis of a typical specimen is given in Table I (no. 5) and plotted in Figure 6. The potassium feldspar (perthitic microcline) is as abundant in prograde rocks as it is in retrograde ones, so is clearly not related to any enrichment of potassium during late recrystallization. In other respects, these gneisses closely resemble mineralogically the plagioclase-rich varieties. Plagioclase is oligoclase, biotite is reddish brown, and garnet is pink

almandine. In cataclastic, retrogressively metamorphosed rocks, biotite has been replaced by muscovite or chlorite and garnet has altered to biotite and muscovite. In general, retrograde effects have been just as marked as in the plagioclase gneisses.

## *Hornblende-bearing Rocks*

Rocks containing hornblende constitute a second distinct petrographic group in the exposures bordering the ocean between Capes Aldrich and Columbia.

Except for quartzo-feldspathic layers, which are irregularly developed, the hornblende-bearing rocks are dark and well foliated. Locally, the greatly contrasting white quartzo-feldspathic and dark green amphibolitic layers also bring out structural features such as folding and boudinage.

Porphyroblastesis of feldspar is not as spectacularly developed as in the garnet-biotite-plagioclase gneisses. Garnet, however, has grown to moderate size (up to 1 inch) in many of the rocks.

For descriptive purposes, the hornblende-bearing rocks may be divided into three groups:

(1) amphibolites (with less than 50% feldspar plus quartz)

(2) hornblende gneisses (with more than 50% feldspar plus quartz)

(3) retrogressively metamorphosed amphibolites and gneisses.

Modes of representative rocks are given in Table II.

### *Amphibolites*

The *hornblende* of these rocks has the pleochroic formula X = pale yellowish green, Y = brownish green, and Z = dark green. Intergrown with it in some rocks is a deep red-brown *biotite*. The latter appears to be a primary mineral, as there is no evidence that it is an alteration product of the hornblende. Where chloritization of biotite has occurred, sphene has apparently resulted and has crystallized along hornblende grain boundaries. In rocks with fresh red-brown biotite but otherwise similar mineralogy, sphene is rare. This, along with the colour, suggests that the biotite is rich in titanium.

*Garnet* is generally abundant in rounded grains, 1 to 2 mm in diameter, sieved with quartz inclusions. More rarely, it forms porphyroblasts 2 to 3 cm across. Unit cell edge and refractive index of a typical garnet from an amphibolite are 11.620Å and 1.795, respectively. These data place the garnet in the field of almandine containing andradite molecule + grossular molecule $> 30$, in Sastri's (1962, Pl. 35) classification. Conspicuous quartz-rich pressure shadows form "tails" to many of the larger porphyroblasts.

*Plagioclase* is another mineral that may or may not be porphyroblastic in the amphibolites. Porphyroblasts appear to have developed rather late, as deformational textures are lacking. The crystals tend to occur in discrete layers at irregular intervals (both laterally and vertically) and may be joined by thin feldspathic veins, giving rise to a "beaded" structure.

Porphyroblasts and groundmass plagioclase is oligoclase ($An_{25-27}$). Zoning is slight and normal in porphyroblasts and was not detected in groundmass crystals.

*Scapolite* is an accessory mineral in some amphibolites. It is invariably associated with plagioclase and forms anhedral patches 0.1 to 0.2 mm in size. It occurs in other hornblende-bearing rocks (*see* below), but has not been found elsewhere in the region. Thus,

TABLE II

*Modal and chemical analyses of hornblende-bearing rocks from the type area of the Cape Columbia Complex.*

| | 1 | IV | 2 | 3 | 4 | 4V | 5 | 6 | 7 | 8 |
|---|---|---|---|---|---|---|---|---|---|---|
| Plagioclase | 16.4 | 50.0 | 27.4 | 35.3 | 42.1 | 63.0 | 29.3 | 35.2 | | |
| Quartz | 9.1 | 46.3 | 15.5 | 18.1 | 14.5 | 31.4 | 21.7 | 10.1 | 12.2 | |
| Hornblende | 60.5 | 2.6 | 47.1 | 26.0 | 38.6 | 5.6 | 35.4 | 22.3 | 29.9 | |
| Garnet | 7.1 | 1.1 | 3.8 | – | 0.8 | – | 6.6 | – | 5.3 | |
| Biotite | 2.1 x | – | 4.2 | 9.6 | – | – | – | 9.6 | 2.9 | |
| Epidote + | – | – | – | 2.9 | 4.0 | tr | 3.9 | 19.8 | 2.3 | |
| Chlorite | tr | – | – | – | – | – | – | 0.9 | 12.4* | |
| Scapolite | – | – | 2.0 | 7.2 | – | – | 2.0 | 2.1 | – | |
| Apatite | 0.8 | tr | tr | 0.9 | tr | tr | 0.9 | tr | tr | |
| Zircon | – | – | – | – | – | – | – | tr | tr | |
| Carbonate | – | – | – | tr | – | – | – | – | 2.1 | |
| Rutile | – | – | – | – | – | – | tr | – | – | |
| Sphene | 3.0 | – | tr | – | – | – | – | – | 1.9 | |
| Black Fe oxide | 1.0 | – | – | – | tr | – | tr | tr | tr | |
| x chloritized | | | + includes clinozoisite | | | | | *after biotite | | |
| Points counted | 2006 | 1684 | 2010 | 1990 | 2003 | 1686 | 1986 | 2036 | 2009 | |
| Plag. comp. | $An_{27}$ | $An_{27}$ | $An_{26}$ | $An_{32}$ | $An_{23}$ | $An_{23}$ | $An_{26}$ | $An_{25}$ | $An_{27}$ | |
| | | | | | | | | | | |
| $SiO_2$ | 48.6 | | | | | | 56.0 | | 49.5 | 50.3 |
| $TiO_2$ | 1.35 | | | | | | 0.68 | | 1.20 | 1.6 |
| $Al_2O_3$ | 14.4 | | | | | | 14.7 | | 12.3 | 15.7 |
| Tot. Fe as $Fe_2O_3$ | 14.9 | | | | | | 7.3 | | 11.7 | 12.3 |
| MnO | 0.19 | | | | | | 0.15 | | 0.19 | 0.2 |
| MgO | 7.7 | | | | | | 3.9 | | 5.3 | 7.0 |
| CaO | 9.7 | | | | | | 5.1 | | 11.1 | 9.5 |
| $Na_2O$ | 3.3 | | | | | | 3.6 | | 2.3 | 2.9 |
| $K_2O$ | 0.8 | | | | | | 2.3 | | 1.3 | 1.1 |
| Total | 100.9 | | | | | | 93.7 | | 94.9 | |

Analysts (1,5,7): Rapid Methods Group, Geological Survey of Canada

1,2. Amphibolite
IV. Leucocratic vein in 1
3,4,5. Prograde hornblende gneisses
4V. Leucocratic vein in 4
6,7. Retrograded hornblende-bearing rocks
8. Average of chemical analyses of 200 amphibolites (Poldervaart, 1955, Table 21, p. 136)

*GSC*

its formation may depend on the presence of hornblende. Both minerals may contain significant amounts of fluorine or chlorine and perhaps the hornblende combined with plagioclase to provide the material for producing scapolite. Yet both the hornblende and the plagioclase are perfectly fresh and there is no sign of significant metasomatism, the process usually considered responsible for formation of scapolite.

### *Hornblende Gneisses*

These rocks are distinguished from the amphibolites chiefly on the basis of their high content of feldspar and quartz. Foliation in the gneisses, while excellent on the scale of a large hand specimen or outcrop, is not so well defined as in the amphibolites in thin section.

Compared with its appearance in the amphibolites, the Z colour of the *hornblende* is a lighter green, in some rocks with a bluish tinge. *Biotite* is dark brown rather than red-brown. *Plagioclase*, while unzoned, varies considerably in composition and values of $An_{21}$ to $An_{44}$ were recorded; the most common is $An_{25}$. *Scapolite* is coarser (maximum size 1.5 mm) and more abundant than in the amphibolites. In one specimen, it is confined to 1-mm-wide layers, associated with plagioclase and minor hornblende, parallel to the foliation. In this specimen, scapolite may well have resulted from the introduction of metasomatic fluids.

A common accessory mineral is zoned *clinozoisite-epidote*. Crystals, 0.5 to 1 mm across, consist of cores of clinozoisite (itself zoned) and rims of epidote. $Fe^{3+}$ content of epidote usually decreases with progressive metamorphism according to Miyashiro and Seki (1958, quoted in Deer, Howie, and Zussman, 1962a, p. 207). The reverse zoning of epidote, as seen here, would thus indicate retrogressive metamorphism and, indeed, as mentioned below, clinozoisite is rare or lacking altogether in more retrograded hornblende gneisses, only epidote being present.

### *Retrograde Amphibolites and Hornblende Gneisses*

Retrograde effects in the hornblende-bearing rocks are notably less marked than in the granitic gneisses. Significantly, cataclastic textures are rarer. Yet both groups were presumably subjected to the same tectonic stresses and metamorphic conditions at any given time. As the degree of retrogression seems clearly related to intensity of penetrative deformation, the lesser retrogression of the mafic rocks may be due to their possessing a physical character different from that of the felsic rocks, leading to a different response to the same stresses.

Retrograde features in the hornblende-bearing rocks are:

(1) Abundance of epidote
(2) Chloritization of biotite
(3) Raggedness of hornblende grains due to replacement by quartz and feldspar
(4) Notable content of sphene and carbonate
(5) Sericitization of plagioclase
(6) Deformation textures, such as granulation of garnet crystals and straining of plagioclase and quartz.

Clinozoisite is a very minor constituent of these rocks. When present, it forms small cores of epidote crystals. From observations in the other hornblende-bearing rocks it is concluded that original clinozoisite has been converted to its more iron-rich relative, epidote, by the replacement of Al by $Fe^{3+}$, as may be expected in retrogressive metamorphism (*see* above).

### *Leucocratic Layers in the Hornblende-rich Rocks*

The light coloured layers or veins that contrast so markedly with the enclosing dark, hornblende-rich rock vary in thickness and extent. Some are only a few centimetres wide, others a foot or more. Both laterally extensive and discontinuous lensoid types were recognized. A few cut across the foliation but most are concordant. Many are thinly bordered by country rock enriched in hornblende.

Modes of two specimens of leucocratic veins and their host rocks (one an amphibolite, the other a hornblende gneiss) are given in Table II. The veins consist of plagioclase and quartz, coarser grained than these same minerals in the host rock, but otherwise identical, and carry very little ferromagnesian material. They are thus analogous to the veins in the granitic gneisses and are probably also "sweated out" from the country rock. Again, this interpretation is supported by exhaustive studies by Kretz (1966), who found only very small chemical differences between host-hornblende and vein-hornblende and none in plagioclase composition.

Segregation of mafic and felsic material as a result of deformation during metamorphism has been documented by Bowes and Park (1966) in Lewisian amphibolites remarkably similar to the Cape Columbia rocks.

Such "metamorphic differentiation," carried to an advanced stage (i.e., a multitude of closely spaced zones of differentiation), might result in the eventual formation of the feldspar-rich hornblende gneisses described above. The fact that considerable redistribution of material appears to take place during metamorphism producing hornblende-rich rocks may invalidate the opinion, held by many geologists, that amphibolites with abundant quartz are of sedimentary origin.

## *Miscellaneous Metamorphic Rocks*

Clearly metasedimentary rocks are of very minor occurrence in the type area of the Cape Columbia Complex, as originally defined by Blackadar (1954). Some marble beds outcrop among granitic gneisses west of Cape Columbia. They are generally impure calcium carbonate rocks with abundant phlogopitic mica and quartz. Carbonate (1–2 mm) and quartz grains are highly strained.

Undoubted igneous rocks also were involved in metamorphism. The rock of a small outcrop on the slopes of Mount Cooper Key, west of Cape Aldrich, is equigranular and non-foliated. Outcrop was insufficient to identify it as a former dyke, sill, or flow but a thin section shows that the rock was probably gabbroic before metamorphism.

Relict pale green clinopyroxene (heavily dusted with black opaque grains) is mantled by green hornblende, which, in turn, is surrounded by granular garnet. Coronas consisting of red-brown biotite rimmed by garnet surround most of the larger, relict "iron ore" grains. Between the mafic mineral–corona aggregates lie granular plagioclase and quartz, minute needles of epidote or clinozoisite and a little biotite. The plagioclase (0.2 mm) has the composition $An_{27-30}$ and is interpreted to be recrystallized original, more calcic plagioclase. Calcium released in the recrystallization has gone to form epidote (or clinozoisite). The quartz probably is also a newly formed mineral.

The coronas attest to the instability of pyroxene in this particular metamorphic environment. They are considered to have developed in the solid (gabbroic) rock as a result of regional metamorphism. Similar textures in metagabbros of the Adirondack Highlands have been similarly interpreted by Buddington (1939, p. 296–297).

### *Granitic and Pegmatitic Dykes*

Numerous leucocratic dykes cut the gneissic rocks of the area. Some are highly discordant to the foliation, while others are more concordant. Variations in proportions of quartz, feldspar, and mica (mainly muscovite), and the presence or absence of garnet result in different rock types.

An important feature is that all dykes studied exhibit effects of deformation. Intensity of deformation varies greatly, which perhaps indicates more than one period of dyke intrusion. Signs of contact metamorphism and chilling were not observed, which suggests that the dykes were emplaced while the country rocks were still hot. Most dykes are much less than a foot wide; small size would be expected in dykes formed by exudation of felsic material during metamorphism. For these various reasons the writer feels that the dykes studied, particularly the one sampled for age determination (*see* below), are syn- or even pre-tectonic. Further comments on deformation of dykes follow in the next section.

### *Structural Features*

No attempt was made at detailed study of the structural geology of the Cape Columbia Complex, either in the type area or elsewhere. Structures are obviously complex and unravelling them would be a major undertaking in itself. Brief comments on the main features observed follow.

Attitude of the foliation in the rocks exposed along the coast is rather constant, on a broad scale, the strike being northwest, except on the headland west of Cape Columbia where it is east–west. Dip varies from 50°SW to vertical and, in places, overturning was recognized.

Intricate folding is common in the rocks. Isoclinal folds with steep to vertical axes and recumbent folds were found all through the area. Marble beds are flow-folded in places.

Boudinage is a feature of the hornblende-bearing rocks. A particularly fine example is pictured in Figure 9, which shows two boudinaged pegmatite dykes in hornblende gneiss. Some of the boudins have been widely separated, rotated, and country rock has "flowed" between them. Two periods of deformation are indicated—one produced the foliation that is truncated by the dykes; the other, the boudinage.

What appears to be the reverse has occurred in nearby hornblende-bearing rocks. There, the boudins are hornblende gneiss and amphibolite in a medium-grained quartz-feldspar gneiss host. The relative competencies of the two rock types are the opposite of those in the first-mentioned example due, perhaps, to different conditions of deformation.

Figure 10 shows the intense shearing and folding that is locally visible in the hornblende-bearing rocks. The truncation of the fold is evidence that shearing followed folding and exemplifies the post-crystalline deformation that has left so clear a mark on the Cape Columbia Complex.

## Muscovite-rich Schists South of Capes Aldrich and Columbia
(Map-unit 1b)

Separated from most of the coastal exposures of gneiss by a snowfield, a belt of muscovite-rich schists extends from the western shore of Parr Bay to the eastern shore of Markham Fiord. At both extremities of the belt, the schists are in exposed contact with the gneissic rocks described in the preceding pages. It was at these localities that they were studied in

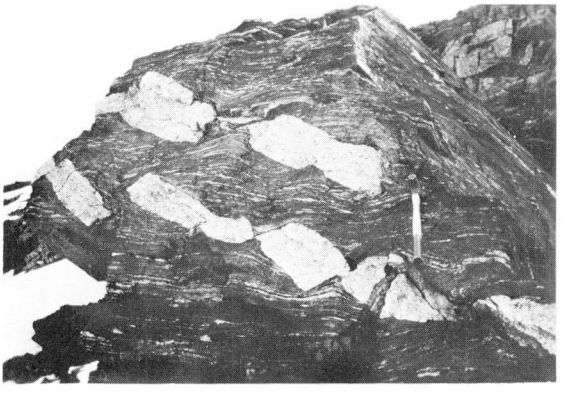


FIGURE 9

Two boudinaged pegmatite dykes in hornblende gneiss of the Cape Columbia Complex. Note that the gneissic foliation both terminates abruptly against, and flows around, the boudins. At least two periods of deformation are indicated. (Field photograph.)

202022-E

FIGURE 10

A minor fold in a migmatitic gneiss of the Cape Columbia Complex has been cut off by later shearing. Outcrop west of Cape Columbia. (Field photograph.)

202022-F

1965 by the writer for the first time. Because of their proximity to the gneisses and their mineralogical affinities, the schists are considered to be a part of the type Cape Columbia Complex.

The schists commonly display a lustrous, silvery sheen, due to their abundant muscovite. They are knotted, with porphyroblastic garnet, feldspar and/or quartz, and are crenulated by a late strain-slip cleavage. Some varieties have randomly oriented ("feather"), chloritized amphibole prisms lying in the foliation planes.

The relatively low-grade appearance of these rocks in outcrop is confirmed by microscopic study. Modes of four representative specimens are listed in Table III.

*Plagioclase* occurs in most rocks as 0.2-mm grains in the groundmass. One specimen studied, however, has, in addition, 1- to 2-mm porphyroclasts of unzoned albite ($An_5$), which appear to have been rotated out of alignment with the foliation. Groundmass plagioclase is albitic in all rocks examined.

*Quartz* is extremely abundant in all the schists. It is commonly intergrown with garnet and may in part represent replacement of the latter.

Pink *garnet* is the common porphyroblastic mineral in the schists. Signs of a complex history of growth and movement are:

(1) Cores are filled with inclusions, while rims are clear.

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| Plagioclase | 9.2 | 8.8 | 30.1 | 26.3 |
| Quartz | 32.3 | 47.3 | 37.1 | 50.1 |
| Amphibole# | – | – | 7.7 | 2.0 |
| Garnet | 9.8 | – | 10.4 | 1.7 |
| Biotite | 1.2 | – | – | 7.6 |
| Muscovite | 44.5 | 38.9 | 14.7 | 7.5 |
| Epidote | – | – | – | 3.8 |
| Chlorite* | 3.0 | 2.2 | – | 1.0 |
| Apatite | – | – | tr | tr |
| Zircon | – | – | tr | tr |
| Carbonate | – | 2.8 | – | – |
| Black Fe oxide | tr | tr | tr | tr |
| # largely to completely pseudomorphs of chlorite<br>* does not include chlorite pseudomorphs after amphibole | | | | |
| Points counted | 1786 | 1892 | 2003 | 1990 |
| Plagioclase comp. | $An_5$ | $An_{10}$ | $An_6$ | $An_9$ |

1. Garnet-muscovite schist.
2. Muscovite schist.
3. Amphibole (chloritized)-garnet-muscovite schist.
4. Garnet-amphibole (chloritized)-mica schist.

GSC

**TABLE III**

*Modal analyses of muscovite-rich schists from south of Capes Aldrich and Columbia.*

(2) Lines of inclusions are strongly sigmoidal or helical ("snowball" structure).

(3) Straight lines of inclusions parallel the elongation direction of the crystal, which is not parallel to the foliation.

(4) In the same specimen, some crystals are "spongy" (ragged edges, highly sieved), while others are euhedral to subhedral.

(5) Quartz-rich "pressure shadows" occur as "tails" to crystals.

Observations (1) and (2) suggest two periods of garnet growth. Observation (2) provides evidence that growth and movement of garnet were concurrent in some rocks, whereas in others (observation 3) the garnets had largely completed growth before being rotated across the foliation. Even in the latter case, some post-rotational growth is suggested by the apparent indentation of garnet rims against the matrix.

*Muscovite* is the major platy mineral. It is notably coarse grained in comparison to the other minerals and in places has grown across the schistosity.

Brown or greenish brown *biotite* is present in some specimens. One rock (Table III, no. 1) contains a relict reddish brown variety, intergrown with chlorite and muscovite. All the biotite is of very fine grain size, ragged and partly altered to penninite, which suggests that it was unstable in these rocks.

*Amphibole* was clearly an unstable mineral in the few rocks in which it occurs. "Feather" amphibole grains are partly to completely pseudomorphs of fine-grained chlorite, and crystals are highly sieved with inclusions of quartz and feldspar.

*Epidote* is abundant in only one rock studied, wherein most of the garnet has been replaced by quartz "eyes." Perhaps the epidote was produced by this breakdown.

Yellowish green *tourmaline* is an accessory mineral and *carbonate* is abundant in some of the schists.

## Metamorphic Rocks of Markham Fiord
(Fig. 29A)

### *The Eastern Shore*

A sequence of low-grade metamorphic rocks (map-unit 1c) outcrops south of the muscovite-rich schists described in the previous section. The exposures extend about 8 miles up the shore of the fiord.

Grey-green to dark green phyllitic schists, phyllites, marbles, micaceous marbles and quartzites, and epidote-rich (metavolcanic?) rocks are interbedded, with phyllitic rocks predominating. The latter are generally crenulated; marbles are flow-folded; and contorted quartz veins and lenses are abundant in most rocks. Major folds with east–west axes are common.

The rocks strike approximately east–west and dip moderately, and thus are broadly concordant with the higher grade rocks to the north.

### *The Western Shore*

The western side of the entrance to Markham Fiord is marked by Cape Nares, where an angular unconformity between the Cape Columbia Complex and the Permo-Carboniferous strata is magnificently displayed.

At the base of Cape Nares, dark, altered biotite gneisses, in places garnetiferous, and garnet-hornblende gneisses (map-unit 1a) dip steeply to the northwest. Small (2–3 mm) white augen present in many rocks are porphyroclastic plagioclase. Retrogressive metamorphic and cataclastic effects are distinct and finely disseminated hematite colours much of the rock.

Above the gneisses lie moderately dipping red conglomerate and sandstone of Permo-Carboniferous age (Christie, 1964, p. 36–37). The conglomerate contains pebbles and fragments of rocks of the Cape Columbia Complex. Overlying the red beds, which are about 150 feet thick, is at least 500 feet of fossiliferous, well-bedded limestone.

Post-Permian (probably Tertiary) faulting has affected the rocks and a fault slice of red beds is interleaved with the gneisses.

South of the cape, greenish phyllitic schist, some of it garnetiferous, and phyllite are the major rock types (map-unit 1c). Minor quartzite beds are interlayered with the phyllites. All rocks are extremely contorted. Six miles south of the cape, banded, brown and white marble is in contact with Permo-Carboniferous strata. There, red beds are lacking and the contact, which is steeply dipping, may be a fault.

## Ward Hunt Island
(Fig. 29B)

Ward Hunt Island, with an area of 8 square miles, lies off the mouth of Disraeli Fiord. The island may be divided geologically into two parts, a northern belt of Permo-Carbon-

iferous sediments and, underlying these and exposed to the south, schists (map-unit 1c) and metaplutonic rocks (map-unit 1a).

The geology of the island has been studied by Christie (1957, 1964), Lyons and Leavitt (1961), and the writer, who concerned himself mainly with the crystalline rocks.

The western tip of the island is formed by well-bedded Permo-Carboniferous limestone, dipping moderately northward. A little inland, the underlying schists are exposed and the sedimentary rocks are represented only by felsenmeer, but nevertheless are doubtless in situ. Foliation in the schists dips steeply to the north. The red conglomeratic beds of the basal Permo-Carboniferous section on the coast of Ellesmere Island to the south (Lyons and Leavitt, 1961, p. 4) are absent there. On this basis, Lyons and Leavitt suggested that the limestone-schist contact is a fault, rather than a simple unconformity. In 1965, the writer found Permo-Carboniferous red sandstone resting on crystalline rocks in a small exposure east of the lake in the central part of the island. In general, however, the conspicuous basal red-bed sequence is lacking in the Permo-Carboniferous section on Ward Hunt Island and the lower contact is probably a faulted unconformity.

The metamorphic and plutonic rocks that form a basement complex on Ward Hunt Island fall into three groups:

(1) Mica and chlorite schists; phyllite, and minor amphibolite.

(2) Intrusive syenite.

(3) Hybrid rock (schist impregnated with syenite).

The metamorphic rocks are chiefly of low grade (greenschist or albite-epidote-amphibolite facies). Some varieties of crenulated green phyllite are identical to rocks of the Markham Fiord area. Pale yellow to brown or greenish brown biotite, muscovite, chlorite (commonly after biotite), epidote, albitic plagioclase ($An_{5-13}$), and quartz are the main constituents. Cataclasis has been so intense in places that the affected rocks might be termed mylonites.

Amphibolite forms minor interlayers in the schist. Much of the original green hornblende has been altered to greenish brown biotite and the plagioclase is sodic oligoclase ($An_{15-16}$). Feldspar and quartz, both showing granulation and strain, make up 30 to 40 per cent of the rock.

A dearth of good outcrop hinders mapping but, as Lyons and Leavitt (1961) recognized, syenite has intruded schist in an interfingering pattern over much of the south half of the island. The syenite is pink, medium to coarse grained, and consists of microcline perthite, interstitial quartz (estimated to be 5–7% by volume) and mafic minerals, which are now highly altered to green biotite and black opaque material. Most of the syenite has been crushed and rendered gneissic. However, the appearance of the undeformed rock and the manner in which it veins the schist, both on the scale of the hand specimen and in outcrop, leave no doubt that the syenite is an intrusive rock younger than the schist. The syenite is considered pre-Permo-Carboniferous because (1) it is highly deformed and the limestone is not, and (2) it does not intrude the limestone. In this study, it is included in the Cape Columbia Complex.

Where syenite has been injected into schist and cataclasis has occurred, a peculiar hybrid gneissic rock has resulted. Angular grains of albite and microcline perthite, which appear to be fragments of larger grains, lie in a granular matrix of quartz and feldspar. Chlorite is generally abundant and, with muscovite, defines an excellent foliation. The perthite is thought to have been contributed by the syenite, and the plagioclase by the schist.

The youngest rocks on the island form dykes that intrude the basement and its sedimentary cover. The major variety is a medium-grained gabbro, consisting largely of altered augite, labradorite, and "iron ore." Also found was a porphyritic to aphanitic dyke of probable andesitic composition.

## Origin and Metamorphic Facies of the Cape Columbia Complex

Most of the rocks in the type area of the Cape Columbia Complex—the coast between Cape Aldrich and the entrance to Markham Fiord—have been so thoroughly recrystallized and deformed that any considerations of their original nature are highly speculative. No unequivocal sedimentary structures or igneous features were observed but the marble and metagabbro indicate that both sedimentary and igneous rocks were originally present.

Unfortunately, because of their small size, the few available samples of typical garnet-biotite gneiss without significant retrogressive metamorphism are unsuitable for chemical analysis. Two specimens of moderately retrograded gneisses (Table I, nos. 2, 4) were analyzed (for details of analytical methods used, *see* Appendix 2). Of these, no. 4 is less altered—garnet has been partly replaced by biotite and epidote is conspicuous, but the rock is otherwise fresh. A large proportion of the biotite in no. 2 has been replaced by muscovite.

The persistent compositional layering of the banded gneisses probably reflects original bedding. Comparison of the three chemical analyses (Table I) with analyses of typical psammitic and pelitic sediments points up the following facts concerning possible sedimentary parent rocks for the gneisses. Silica is lower and iron is higher than in arkose. As in shale, potash exceeds soda, whereas the reverse is true for greywacke. However, the alkali:alumina ratio and silica are higher than in shale. Volcanic sediments such as dacite-rhyodacite tuffs are an unlikely parent because of their high soda:potash ratio. Thus the nature of the original rocks of the banded gneiss sequence remains uncertain, at least on the basis of the data available. In any event, metasomatism may have changed the original composition beyond recognition.

In the hornblende-bearing rocks, chemical analysis does aid in understanding their origin. The three specimens analyzed (Table II) all display high Fe:Mg ratios, which is typical of igneous rocks, as dolomite-pelite mixtures are low in iron. Furthermore, the low potash (Niggli $k$ number $=0.14$) and moderately high $TiO_2$ contents of the amphibolite (no. 1) also suggest an igneous parent (Leake, 1964, p. 239).

Throughout this report, the metamorphic facies classification of Fyfe and Turner (1966) is followed. Their simple scheme, essentially that of Turner and Verhoogen (1960) without any division into subfacies, is well suited to reconnaissance work.

The gneisses and amphibolites of the Cape Columbia Complex type area belong to the amphibolite facies of Fyfe and Turner (1966), which is equivalent to the almandine-amphibolite facies of Turner and Verhoogen (1960). Two mineral assemblages diagnostic of this facies are present: quartz-microcline-plagioclase-biotite in quartzo-feldspathic rocks and hornblende-medium plagioclase-almandine-epidote in metamafic rocks. Two other features typical of the amphibolite facies are:

(1) major potassium feldspar porphyroblastesis

(2) a reddish brown Z colour of biotite.

The mineralogy of the muscovite-rich schists is of the greenschist or albite-epidote-amphibolite facies, depending on whether the amphibole is (or was) actinolite or hornblende (Fyfe and Turner, 1966, p. 362). The reddish brown biotite was found only in rocks at the

contact with undoubted amphibolite facies gneisses and these schists may be retrograded gneisses. It might be argued that the muscovite-rich schists are equivalent in metamorphic grade to the gneisses of the northern coastline but differ from these in bulk composition. However, minerals diagnostic of amphibolite facies pelites, such as staurolite and kyanite, are absent from the schists (but are present elsewhere on the north coast of Ellesmere Island). Furthermore, a general decrease in metamorphic grade from north to south is apparent in many areas of the northern Ellesmere Island coastline. For these reasons, the muscovite-rich schists are considered to have formed under lower pressure and temperature conditions than the gneisses. The original rocks were doubtless argillaceous sediments.

## Age of the Cape Columbia Complex

In the type area of the Cape Columbia Complex, the metamorphic rocks are not in contact with rocks of known age older than Permo-Carboniferous. Thus radiometric dating must be used in an attempt to determine the age of the Complex. Before discussing the isotopic results, it would be well to clarify what is meant by the "age" of a metamorphic terrain.

Even today, many geologists tacitly equate metamorphism with an "orogeny" that occupied only an instant of geologic time. Recent petrographical, structural, and isotopic investigations of metamorphic terrains have invalidated this over-simple concept. The histories of many, and probably most, metamorphic belts appear to involve two or more major periods of metamorphism, each of which may comprise a succession of phases of deformation and recrystallization. The whole process may require 200 to 1,000 million years (Sutton, 1965). Metamorphism, then, has no exact age—it normally spans a considerable period of time.

Added to these difficulties in determining the time of metamorphism is the non-coincidence of the peak of a metamorphic event and the point of time at which a radiogenic isotope is "frozen" in a mineral. To speak of a K-Ar date on a metamorphic mineral as a reliable measure of the "age" of even the latest period of metamorphism is certainly incorrect. All that such a determination gives us is an approximate time when the rock system cooled through some critical isotherm. Also, temperature may not be the sole factor: pressure, shear stresses, presence of volatiles and chemical composition of the system all undoubtedly control diffusion of radiogenic argon. Thus a K-Ar date may reflect the time of uplift and cooling of a rock rather than the time of its metamorphism.

The first K-Ar age determination on the Cape Columbia Complex, on biotite from a gneiss, yielded $545 \pm 35$ m.y. and was published by Blackadar (1960). Subsequent re-analysis of this sample in 1970 resulted in a date of $465 \pm 19$ m.y., which is considered the preferred value (R.D. Stevens, pers. com., 1971).

Three other K-Ar determinations were an outcome of the present study:

$445 \pm 18$ m.y. on biotite from a garnet-mica-feldspar augen gneiss;

$403 \pm 17$ m.y. on muscovite from a pegmatite cutting gneisses;

$389 \pm 21$ m.y. on hornblende from a garnet-hornblende gneiss (Table II, no. 5).

All four rocks occur within a few miles of each other along the coastline between Capes Aldrich and Columbia (*see* Fig. 29). The biotite-bearing gneisses, which were sampled at locations less than a mile apart, show definite evidence of retrogressive metamorphism: some of the biotite appears to be an alteration of garnet. The pegmatite occurs as a narrow dyke, slightly discordant to the foliation of the enclosing gneisses and, although intrusive,

was interpreted in the field to be essentially contemporaneous with the metamorphism that produced the country rocks.

The date originally obtained on Blackadar's sample—545 ± 35 m.y.—was interpreted as an upper limit to the time of metamorphism of the Cape Columbia Complex (Blackadar, 1960; Haller and Kulp, 1962, p. 29). According to the Geological Society of London time-scale (1964, p. 260–262), this limit would lie on the Middle Cambrian–Lower Cambrian boundary. The implication was that metamorphism took place in Early Cambrian or Precambrian time. On this basis, Haller and Kulp (1962) further suggested that the Cape Columbia Complex was an extension of the Precambrian Carolinidian fold belt of northeastern Greenland. Because of the younger date obtained in the redetermination, Haller and Kulp's suggested correlation no longer has a basis in isotopic age but may still be correct, as the Cape Columbia Complex is almost certainly lower Paleozoic or even Precambrian (*see* below).

The 465 m.y. and 445 m.y. age determinations are essentially in agreement within error limits, i.e., Lower to Middle Ordovician. The other two dates are younger still: Late Silurian and Early Devonian. All these apparent ages fall in the range obtained by Haller and Kulp (1962, Table I) in East Greenland and termed by them "Caledonian."

The Ordovician to Devonian dates from Ellesmere Island may be the result of one or more geological events, or a combination of these. First, they may give the times of uplift and cooling following metamorphism of the Cape Columbia Complex. Secondly, the retrogressive metamorphism and cataclasis that were so pronounced in the type area could well have caused argon loss, leading to lower apparent ages. A third possibility is a "resetting of the 'argon clock' " as a result of reheating of the complex by the widespread plutonism that, according to isotopic evidence from igneous rocks, occurred in Caledonian time.

The large spread in the isotopic ages from a restricted area of undoubtedly contemporaneous rocks illustrates the limitations of the potassium-argon dating technique when applied to basement rocks of an orogenic belt with a complex history. Of far greater value in setting a minimum age for the Cape Columbia Complex are the following observations made in the field: (1) unmetamorphosed Middle Ordovician (Wildernessian) strata rest with angular unconformity on chlorite-muscovite schists west of the mouth of M'Clintock Inlet (Trettin, 1969b, p. 7); (2) thick sequences of Upper Ordovician rocks on M'Clintock Inlet are unmetamorphosed. Although it is not certain that the schists are correlative with the Cape Columbia Complex, it seems safe to conclude that the latter is of pre-Middle Ordovician age.

## Upper M'Clintock Inlet

(Fig. 29C)

On the western side of upper M'Clintock Inlet, metamorphic rocks are exposed in the valley of Ooblooyah Creek and at several places on both shores to the south. They are in contact with lower Paleozoic (map-unit 2), Permo-Carboniferous (map-unit 3), and ultramafic rocks (map-unit A).

### Petrography

The two most abundant metamorphic rocks in the area are quartz-rich biotite schists and marble (map-unit 1b). Concordant and discordant veins of quartz, in places beaded or boudinaged, as well as lenses or pods of quartz, are conspicuous in most of the rocks. Quartz segregations are found to a lesser extent in the marbles. Foliation is extremely well

developed, partly due to the intense deformation that has affected all the rocks. Marbles are flow-folded and schists are highly cataclastic.

Deformation and retrogressive metamorphism have obscured the original mineralogy of most of the rocks and have given them a deceptively low-grade appearance. The lowest grade rocks, chlorite-quartz schists of the greenschist facies (map-unit 1c), occur in the valley of Ooblooyah Creek. Elsewhere, biotite schists with amphibolite facies mineralogy and interlayered marbles predominate.

Cataclastic textures of the biotite schists, best seen in thin section, attest to a deformation that must have been as intense as any in the metamorphics of northern Ellesmere. Recrystallization both preceded and followed deformation, which may itself have been multi-phase. *Plagioclase* ($An_{21-25}$) occurs as cracked and rotated porphyroclasts, 1 to 2 mm in size. Colourless to pale reddish brown *biotite* has varying orientation, many grains lying parallel to the foliation, others perpendicular or diagonal to it. Most cross-biotite is kinked. There is no clear textural evidence (e.g., grain size of all biotite is essentially uniform) that the cross-biotite necessarily grew in that orientation, rather it appears to have been rotated out of alignment with the foliation. Much of the *muscovite* is secondary after biotite and feldspar. *Quartz* is aggregated in lenses with mortar texture and in "Plättung" layers. *Garnet* invariably shows bent inclusion trails at angles to the foliation, pressure shadows, and ragged margins where replaced by quartz and feldspar. One rock contains *sillimanite* in fibrolite sheaves intergrown with biotite. Another has *staurolite* crystals (2–3 mm), which are deformed and largely replaced by chlorite. A little *potassium feldspar* is present locally. *Tourmaline* is a common accessory but far more abundant is dusty *graphite*, which occurs throughout the rock but is especially concentrated in micaceous layers and in clots that may represent sites of former porphyroblasts. The graphite particles are so dense in places that useful point counting is not feasible.

Marble layers vary in composition from relatively pure calcium carbonate to phlogopite-scapolite-quartz-calcite rocks.

## Structural Relationships and Age

On the northern valley slopes of Ooblooyah Creek, quartz-chlorite schists (map-unit 1c) are in fault contact with the pre-Pennsylvanian M'Clintock West ultramafic massif (map-unit A, *see* Chap. III). Near the mouth of the stream, similar metamorphic rocks are unconformably overlain by Pennsylvanian red beds. Block faulting has locally juxtaposed metamorphics and Permo-Carboniferous sediments.

Biotite schists and marbles, overlain by unmetamorphosed Ordovician and Permo-Carboniferous strata (map-unit 3), are exposed along the shore of M'Clintock Inlet south of Ooblooyah Creek (Fig. 11). Similar metamorphic rocks form the two small islands off the western shore of the fiord, and are capped by Ordovician and Silurian(?) carbonate rocks (map-unit 2) on the eastern shore.

The appearance of the basement exposures and the overlying sediments suggests an unconformable relationship, particularly on the eastern shore at the head of the inlet. Horizontal to gently folded, well-bedded carbonate rocks are apparently draped over southward-dipping metamorphic rocks. The sedimentary sequence was assumed by Christie (1964) to be Permo-Carboniferous, but work during "Operation Grant Land" showed that it was, in fact, Upper Ordovician (Richmondian) on the west side of the inlet, and Silurian and Upper Ordovician on the east side (*see* Trettin, 1969b, Fig. 7, which shows the eastern shore of upper M'Clintock Inlet, not the west side as stated in the caption). Capping them are flat-lying Pennsylvanian red beds.

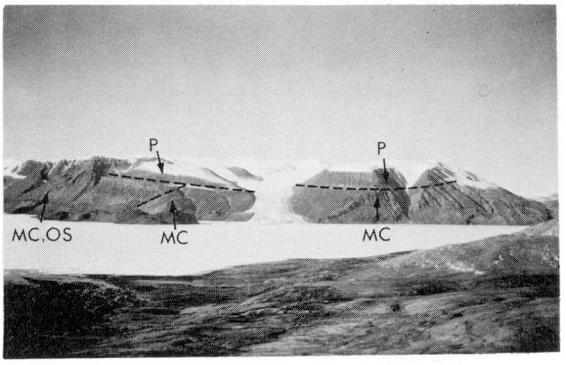



FIGURE 11

Metamorphic and Paleozoic sedimentary rocks on the western shore of upper M'Clintock Inlet. Flat-lying Pennsylvanian red beds (P) overlie Ordovician and (?) Silurian carbonates (OS) and south-dipping metamorphic rocks (MC) of probable pre-Middle Ordovician age. The Ordovician–Silurian strata and the metamorphics are everywhere in fault contact. The valley of Ooblooyah Creek is at the extreme right of the picture (*see also* Fig. 21) (Field photograph, August 1965).

*202022-N*

H. P. Trettin and the writer examined these exposures in detail in 1967. In places where the sedimentary strata, from a distance, appear to be domed up over the basement, it was found that part of the sedimentary sequence is at the same level as the metamorphic rocks and strikes into them. Across the fiord, west of the island, low-angle fault slices of metamorphic and sedimentary rock interfinger. Ordovician carbonate strata wedge out against basement rocks along these faults, which pre-date Pennsylvanian deposition, as the red beds are not displaced. There is no sign of a basal conglomerate in the Ordovician section. Furthermore, at least 9,000 feet of Middle and Upper Ordovician volcanic and sedimentary rocks that normally underlies the Richmondian carbonate strata (Trettin, 1969b) is not observable here.

In summary, all the evidence points to a fault contact rather than an unconformity between the metamorphics and the Upper Ordovician strata. As these strata are unmetamorphosed, a minimum Late Ordovician age can be assigned to the metamorphic rocks. If, as seems likely, the latter are correlative with the pre-Middle Ordovician chlorite-muscovite schists west of the mouth of M'Clintock Inlet (Trettin, 1969b), their minimum age may be lowered to pre-Middle Ordovician.

## Eastern and Southern Ayles Fiord

(Fig. 29D)

Examination of the metamorphic rocks of Ayles Fiord was largely confined to the northeastern shore and the southwestern inlet at the head of the Fiord. Exposures on the northwestern shore are described in a later section.

### The Northeastern Shore

#### *Schistose Metamorphic Rocks*

Sea-cliffs provide excellent exposures of a thick metasedimentary sequence (map-unit 1c) that has been affected by igneous intrusion (Fig. 19). The extreme northeastern part of the coast is occupied by the Cape Fanshawe Martin mafic-ultramafic pluton (map-unit C), which has intruded the metasedimentary rocks (Chap. III). The immediate country rock is banded calcite marble, dipping steeply to the south. Southward along the cliffs are exposed quartzite, marble, mica (with fibrolite) and chlorite schists, and phyllite, the rocks appearing

in this general order but with repetition of units in the section. Strong colour contrasts between many of the units produce a striking banded appearance.

The mica schists are generally fine grained and have been cataclastically deformed. Chloritized brown and green biotite, muscovite, epidote, sodic plagioclase (R.I. less than quartz) and quartz are the chief minerals. The presence of relict brown biotite and green hornblende suggests that the schists have been downgraded from a higher rank. The same may be true of the chlorite schist and phyllite, but there is no mineralogical evidence of retrogressive metamorphism.

Sillimanite occurs in mica-chlorite schists, some of which carry hornblende, calcite and/or epidote. Sillimanite and hornblende normally being incompatible, these two minerals must be out of equilibrium. The hornblende is in small, ragged flakes, so is probably relict. All the sillimanite occurs as fibrolite and appears to have formed from brown biotite, with which it is intergrown. Much of the biotite and its attendant fibrolite was later replaced by muscovite, which tends to be relatively coarse and randomly oriented. Chlorite was a still later alteration of biotite during retrogressive metamorphism.

The sillimanite-bearing schist lies about half a mile from the contact of the Cape Fanshawe Martin pluton and is the nearest pelitic rock (intervening rocks are remarkably pure quartzites and marbles). The pluton, which must have been extremely hot when it intruded, probably provided the heat and fluids needed to produce sillimanite at elevated temperature and muscovite at waning temperature. A similar occurrence in Ireland of late sillimanite and muscovite has been described by Tozer (1955), who concluded that the minerals are the result of a nearby granitic intrusion. It is evident that such sillimanite cannot be used as an index of grade of regional metamorphism.

South of the small inlet on the eastern shore of Ayles Fiord (*see* Fig. 19) (hereafter in this section referred to informally as "the northeast inlet"), metasedimentary and meta-igneous rocks appear to diminish in metamorphic grade southward. The sequence is characterized by great colour variation, fine grain size, and conspicuous banding. Numerous beds are only a few inches thick. The main rock types are phyllite, slate, and quartzite. Marble, amphibolite, and schist occur in minor amounts.

A typical phyllite in thin section shows banding of micaceous and quartzo-feldspathic laminae, which are greatly crenulated. A second S-plane at an angle of 30 to 50 degrees to the laminae (measured in the plane of the thin section) is defined by the orientation of the grains in the laminae. Garnet has formed in one leucocratic lamina, the crystals being aligned parallel to the banding. Clearly, the laminae represent original bedding of rocks of different compositions, and a vague foliation has developed at an angle to it.

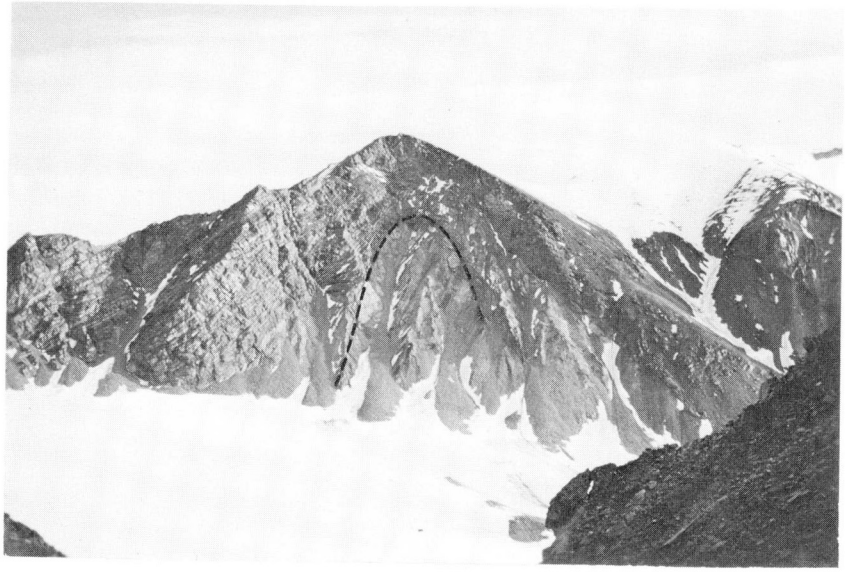
In the amphibolite, hornblende (X = yellowish green, Y = green, Z = bluish green) has a near-decussate texture and is intergrown with abundant clinozoisite, which forms radial aggregates. Black iron opaque grains are a major accessory constituent. It seems likely that this is a metamorphosed igneous rock.

### *Granitic Gneiss*
(Map-unit 1a)

Granitic gneiss outcrops in the area of the "northeast inlet" on the northeastern shore of Ayles Fiord.

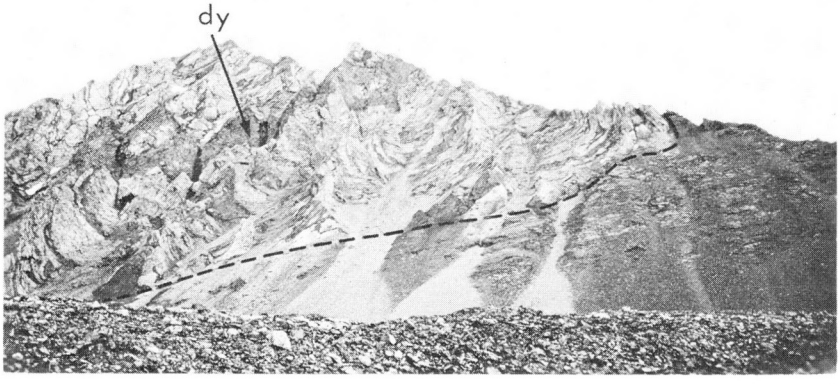
The gneiss is a leucocratic rock of quartz monzonitic composition, in places augen textured and well foliated. The augen consist of perthitic microcline, which continued to grow after the groundmass minerals were formed. Inclusions of the latter in the augen are

common, particularly sericitized and epidotized plagioclase of composition $An_{28-30}$. Exsolved albite, in numerous well-twinned patches, is $An_5$. In highly deformed rocks, the augen, as well as the groundmass, have been strained or even cracked but in most rocks they are little deformed, although the groundmass is commonly cataclastic. Groundmass minerals, besides plagioclase, are microcline, dark brown biotite, muscovite, minor epidote and green hornblende, and abundant quartz.



202022-K

FIGURE 12A. Metamorphic rocks of greenschist facies-grade domed up over granitic gneiss, which may be a remobilized higher grade part of the pre-Middle Ordovician basement complex. The contact is delineated by a broken line. Two miles south of "the northeast inlet" on eastern Ayles Fiord (see Fig. 19). (Field photograph, July 1966.)




202022-L

FIGURE 12B. Granitic gneiss, similar to that shown in Figure 12A and containing large inclusions of amphibolite, thrust(?) over metamorphic rocks of greenschist facies-grade, immediately east of the locality pictured in Figure 12A. An undeformed, unmetamorphosed diabase dyke (dy) cuts the gneisses. (Field photograph.)

An apparently concordant mass of granitic augen-gneiss occurs in phyllite north of the northeast inlet. The gneiss exposure extends only a short distance inland and is there represented chiefly by felsenmeer. No intrusive features were observed at its borders. The occurrence of such a rock in a metamorphic sequence of low regional grade is anomalous. For these various reasons, the mass is considered to be a fault slice brought up from a lower crustal level.

In the area of the small bay south of the northeast inlet, granitic gneiss is well exposed and shows complex relationships with the banded metamorphic rocks. Just inland from the head of the bay, the banded sequence is domed up over the gneiss, which is non-porphyroblastic, poorly foliated and highly cataclastic (Fig. 12A). One mile to the west, granitic gneiss, containing innumerable shredded amphibolite masses, has apparently been thrust over the banded rocks (Fig. 12B). Still farther west, this gneiss body, which is locally augen-textured, well foliated, and cataclastic, is exposed on the shore of the fiord, where it is concordantly bordered by chloritic quartzite and fine-grained mica schists. Vertical, undeformed diabase dykes, up to 10 feet wide, have intruded the gneiss but were not observed in the banded rocks. Within the gneiss are small muscovite pegmatite dykes and segregations; pegmatite was not seen in the bordering rocks.

It might be argued that the granitic gneiss is a deformed intrusion. This is considered unlikely, however, on the grounds that no intrusive relationship between gneiss and adjacent rock was seen anywhere. Furthermore, the association of granitic gneiss with deformed amphibolite lenses (relict dykes?) and unmetamorphosed diabase dykes is characteristic of the metamorphic complex west of Ayles Fiord.

A potassium-argon age determination on muscovite from a pegmatite in the coastal gneiss body gave a value of $354 \pm 15$ m.y. Because pegmatite occurs in segregations having gradational contacts with the gneiss as well as in clear-cut dykes in it, but not in the surrounding rocks, it seems safe to conclude that the pegmatite is essentially contemporaneous with the gneiss. The age determination should then be representative of the gneiss. If so, it suggests that the gneiss was involved in a thermal and/or tectonic event in, or more likely prior to, the Late Devonian. If the interpretation that the gneiss is remobilized basement is correct, an early Paleozoic or older age would hardly be expected. Of course, the radiometric age determined certainly does not exclude the possibility that the granitic gneiss represents a magmatic intrusion that was subsequently deformed. In any event, the age is comparable to that obtained on biotite from the Cape Richards intrusion (Chap. III), $347 \pm 15$ m.y. Coexisting hornblende yields $390 \pm 18$ m.y. and strengthens the conclusion, reached elsewhere in this report, that a major thermal event involving igneous intrusion and possibly remobilization of basement took place in northern Ellesmere Island in Early Devonian (Caledonian) time.

## The Head of Ayles Fiord

Low-grade metamorphic rocks (map-unit 1c) are extensively exposed in high mountains south of the fiord. They include tightly folded and crenulated pyritiferous slate, phyllite, and weakly recrystallized limestone. It appears that these rocks pass northward into the banded sequence of the northeastern shore (described in a preceding section). Southward they apparently extend to the edge of the inland snowfield, about 10 miles from the head of the western arm of Ayles Fiord.

### Structural Features of the Ayles Fiord Region

Mention has already been made of some of the intense deformation in the Ayles Fiord area. Other indications of deformation are isoclinal folding and mullion structure of quartzite beds and flow-folding of marble. Fold axes in quartzite plunge steeply to the east. Axes of mullions, which are of the "bedding" type (Whitten, 1966, p. 315), parallel the fold axes.

A large recumbent fold occurs in marbles and amphibolites on the northern shore of the northeast inlet.

A marked change in the trend of layering in the metamorphic rocks constitutes a major structural feature of the Ayles Fiord region. South of the fiord, the general trend of bedding is north-south. At the southwestern head of the fiord, the beds bend to the northeast and dip steeply to the northwest or vertically. North of the northeast inlet, the beds strike east-west, dip steeply to the south and, judging by the attitude of crossbedding in quartzites, are right side up. If, on the basis of their higher metamorphism, the northern rocks are considered to be older than the southern rocks, then the flexure is synclinal, plunging southward.

This change in trend of layering can be followed also on the eastern shore of Milne Fiord, to the west.

### Correlation with Other Parts of the Northern Ellesmere Metamorphic Complex

Most rocks in the Ayles Fiord area are of very low metamorphic grade. Although a northward increase is observable, a continuous transition to high-grade metamorphic rocks is not found along the eastern shore of Ayles Fiord (but may exist on the western shore, *see* below). Rocks that are possibly correlative with the type Cape Columbia Complex appear to be fault slices in the low-grade sequence. Too little is known of the low-grade metamorphic rocks of this and other regions of northern Ellesmere Island to enable one to say whether they are the lower grade equivalents of the Cape Columbia Complex or younger rocks that were metamorphosed during a later orogeny.

## The Coastal Area, Western Ayles Fiord to Petersen Bay[1]

(Fig. 29D)

The region of western Ayles Fiord and Milne Fiord is mainly a gneiss terrain. Exposures along the shores of Milne Fiord were examined only briefly during a helicopter traverse. Most of this section deals with the rocks of northwestern Ayles Fiord and the eastern shore of Petersen Bay, the bay between Milne Fiord and Yelverton Bay.

### Northern Granitic Gneiss Terrain

The northernmost exposures are predominantly of a leucocratic mica-feldspar-quartz gneiss, quartz dioritic to quartz monzonitic in composition (map-unit 1a). Modes of four specimens are given in Table IV and plotted in the quartz-potassium feldspar-plagioclase triangle of Figure 13. However, the specimens are not necessarily representative of this gneiss terrain, which is very extensive and clearly heterogeneous.

*Plagioclase* composition ranges from $An_{18}$ to $An_{22}$, zoning being slight to absent. *Potassium feldspar* occurs as microcline perthite, well twinned and with exsolved plagioclase of stringlet type (Deer, Howie, and Zussman, 1963b, p. 68). *Quartz* is evenly distributed

[1]Name approved by the Canadian Permanent Committee on Geographic Names on May 10, 1973.

in some specimens; in others, particularly cataclastic ones, it tends to form aggregates of recrystallized grains. *Biotite* has X = pale yellow and Y = deep brown and in cataclastic gneisses has been replaced largely by *penninite*. *Muscovite* is generally intergrown with biotite and becomes abundant in deformed biotite-bearing rocks. Both micas in such rocks are confined chiefly to clots or sheaves among the quartz and feldspar grains.

Cataclastic texture is common, particularly in the gneisses of western Ayles Fiord. Grain size varies from fine to medium and layers of augen gneiss occur within more even-grained rock. A specimen (Table IV, no. 3; collected by R. L. Christie in 1954) from the western shore of Milne Fiord is of interest in that, while foliated and vaguely augen-textured, its texture under the microscope turns out to be hypidiomorphic-granular, with no hint of cataclasis. The augen (0.5 to 10 mm long) are well-twinned microcline perthite and quartz and are rimmed by biotite flakes, which, with muscovite of the groundmass, define a wavy foliation. Whatever was its nature (plutonic or metamorphic) before the latest metamorphism, the rock appears to have been extensively recrystallized under comparatively static conditions.

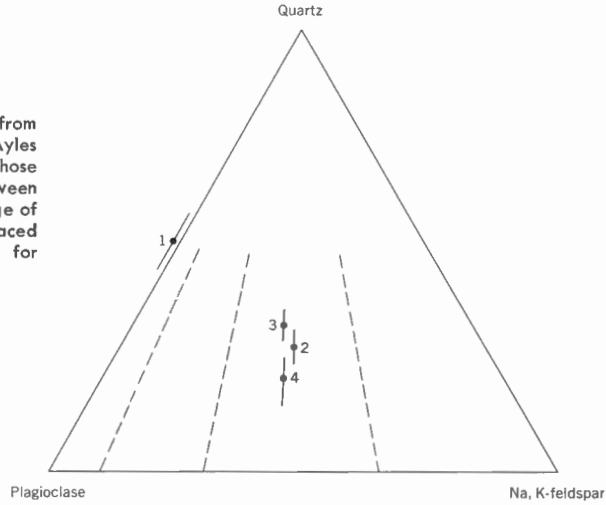
The granitic gneiss on the western shore of Ayles Fiord is rich in dark, schistose layers that parallel the gneissic foliation. They vary in thickness from a few inches to a foot and occur at irregular intervals. Due to incomplete exposure, their maximum lateral extent is not known but some are continuous over at least several hundred feet.

One such layer is composed of poorly foliated, brown biotite intergrown with epidote. Quartz and feldspar? are present in very minor amounts. Calcite has formed late in aggregates replacing other minerals and in crosscutting veinlets. The scarcity of felsic minerals suggests a meta-igneous origin but the abundance of biotite makes this seem unlikely. The rock may be a metamorphic differentiate.

The other three specimens appear to be of metasedimentary or metavolcanic rocks. They are garnetiferous gneisses with 60 to 70 per cent plagioclase and quartz. Most of the feldspar is finely crushed but heavily sericitized porphyroclasts of oligoclase are present. Two specimens are biotite gneisses in which the biotite is identical to that in the enclosing granitic gneiss (Table IV, no. 1). Epidote, chlorite, and minute muscovite flakes are common and indicate retrogressive metamorphism. The third specimen has chlorite but no biotite and is granulated also.

FIGURE 13

Modal compositions of granitic gneiss from the northern gneiss terrain between Ayles Fiord and Petersen Bay. Numbers are those in Table IV. The vertical distance between the ends of each line gives the percentage of mafic minerals. No. 1 has been displaced from the quartz-plagioclase sideline for legibility.




GSC

| | 1 | 2 | 3 | 4 |
|---|---|---|---|---|
| Plagioclase | 35.4 | 33.1 | 33.5 | 36.3 |
| K-feldspar | – | 30.1 | 27.7 | 30.8 |
| Quartz | 38.6 | 24.0 | 30.6 | 18.2 |
| Biotite | 10.4 | – | 6.1 | 9.4 |
| Muscovite | 13.9 | 3.8 | 2.1 | 3.6 |
| Epidote | 0.8 | 1.1 | tr | 0.3 |
| Chlorite | – | 7.9 | tr | 1.4 |
| Apatite | tr | tr | tr | tr |
| Zircon | – | tr | tr | tr |
| Carbonate | 0.9 | – | – | – |
| Allanite | tr | – | – | – |
| Black Fe oxide | – | – | tr | tr |
| Points counted | 2006 | 1986 | 2003 | 1752 |
| Plagioclase comp. | $An_{22}$ | $An_{19}$ | $An_{21}$ | $An_{21}$ |

1, 2. Northwestern shore of Ayles Fiord.
3. Northwestern shore of Milne Fiord.
4. Northern shore of Petersen Bay.

*GSC*

**TABLE IV**

***Modal analyses of granitic gneisses from the metamorphic complex between Ayles Fiord and Petersen Bay.***

These rocks are interpreted as relics of sedimentary and/or igneous origin (siliceous tuffs?) that survived granitization but underwent metamorphic differentiation. Subsequent deformation imposed cataclastic texture on both the relics and the granitic gneiss. Retrogressive metamorphism accompanied the deformation.

If the granitic gneiss did originate through granitization of pre-existing rocks, the gneiss terrain would have to be considered a culmination zone of high-grade regional metamorphism, which typically produces large volumes of granitic rocks. In that event, the gneiss should be bordered by other gneisses and schists that diminish in metamorphic grade away from the contact.

In northwestern Ayles Fiord, a mile-wide glacier separates the southernmost exposures of the granitic gneiss from marble. The latter rock is part of the unit that outcrops within the low-grade, banded sequence on the opposite side of the fiord. Aerial reconnaissance in northeastern Milne Fiord revealed various gneisses, schists, and amphibolites between the granitic gneiss and the lower grade rocks of the banded sequence.

## Petersen Bay

More is known of relationships in Petersen Bay, where the granitic gneiss extends along the northern shore almost to the head of the bay. A belt of steeply dipping amphibolites and schists (map-unit 1b), half a mile wide, lies between the gneiss and low-grade rocks that are tentatively correlated with the banded sequence of Milne and Ayles Fiords. From the gneiss contact to the margin of the low-grade sequence, i.e., from northwest to southeast, the succession is as follows:

(1) Amphibolite, rarely garnetiferous; commonly banded with quartzo-feldspathic layers; includes feather amphibolite. Mafic dykes cut the amphibolites.
(2) Mica schists including garnetiferous muscovite schist and kyanite-staurolite-garnet schist.
(3) Biotite-chlorite-muscovite schist.

### *Amphibolites*

These rocks are particularly interesting because they exhibit relict igneous features. Their contact with the granitic gneiss is structurally concordant and very sharp. Most amphibolites are foliated but some contain large proportions of randomly oriented amphibolite prisms, whose abundance varies from layer to layer of a single rock unit. Grain size ranges from fine to medium even in a hand specimen. Except for minor varieties with thin quartzofeldspathic layers, the amphibolites are dark green and notably lacking in light-coloured minerals.

Fine-grained, wholly granoblastic amphibolites are present also and in places contain irregularly shaped inclusions, typically 6 inches across, of "augen" amphibolite. The latter is composed of numerous amygdule-like, leucocratic "augen," approximately 3 mm long, in a matrix of granoblastic hornblende slightly coarser grained than the host amphibolite. Under the microscope, the "augen" appear as granular aggregates of anhedral, 0.2-mm plagioclase grains. The plagioclase is commonly zoned: $An_{43}$ (core) to $An_{31}$ (rim). Apatite is a minor constituent of some "augen" and here and there a hornblende needle of the matrix projects into an "auge." The hornblende of the matrix is very fresh, not markedly pleochroic (X = very pale green, Y = pale green, Z = slightly darker green) and the only constituent of the matrix, except for a little plagioclase (whose composition could not be determined), sphene, and hematite. The amphibole of the host material is paler green, has a smaller $Z\wedge c$ angle (13° as against 18°) and resembles actinolite. The amphibole prisms have ragged terminations and are intergrown in a decussate texture. Scattered black opaque grains are accessory. Feldspar and quartz are conspicuously absent.

The plagioclase "augen" are interpreted to be either recrystallized phenocrysts or glomerophenocrysts in a metavolcanic rock. Inclusions of this rock are too sparse for the host amphibolite to be a metavolcanic breccia. Besides, no other types of inclusion were found. The mineralogy of the host suggests that it is meta-igneous, perhaps originally a lava flow in which fragments of glomeroporphyritic or porphyritic rock were incorporated.

Amphibolites with compositional layering show considerable variation in mineralogy and texture. Hornblende generally is both parallel to the layering and diagonal to it in euhedral to subhedral prisms with helicitic inclusion trails of quartz. The lines of inclusions parallel the foliation, which suggests that the host hornblende has grown under largely static conditions. Brown biotite flakes lying between grains of the matrix (plagioclase $An_{27}$ and quartz) give the rock a foliation parallel to the compositional layering. The biotite flakes terminate abruptly against hornblende margins. Cross-biotite is ubiquitous but quantitatively unimportant, most of the biotite probably having formed under directed stress prior to much of the hornblende.

Hornblende-bearing layers with significant content of plagioclase and quartz are minor in the layered amphibolites, most of which show variation in amounts of mafic minerals associated with hornblende, such as epidote, biotite, and sphene. Layers of epidote amphibolite, biotite amphibolite, and pure amphibolite occur in a single thin section.

### *Mica Schists*

These schists are generally silvery grey, are knotted and crenulated and have variable feldspar and biotite content. Three varieties were studied microscopically. Two are rich in feldspar and biotite and contain kyanite and staurolite. Megascopic examination suggests that the more biotite- and feldspar-poor rocks have little to no kyanite or staurolite.

Like the amphibolites, the schists show multi-phase evolution of their mineralogy. Modes of the staurolite-bearing schists are given in Table V, and descriptions of these rocks follow.

Porphyroblasts (1–2 mm in diameter) of *garnet* and *kyanite* have obviously grown before or during the main deformation. In both minerals there is an internal schistosity ($s_{int}$), defined by quartz inclusions, that is out of alignment with the external schistosity ($s_{ext}$). Kyanite is commonly strained and ragged, having been partly replaced by muscovite, quartz, and plagioclase. Garnet is always associated with biotite. *Staurolite* porphyroblasts (1–2 mm in diameter) are of similar size, euhedral, and rich in quartz inclusions. In the muscovite-poor schist, staurolite has grown across deformed trains of graphite particles, which doubtless represent an earlier, deformed schistosity (bedding?). Staurolite in the muscovite-rich schist apparently has been formed during progressive metamorphism, as it has been replaced partly by groundmass minerals and is deformed. Almost all *muscovite* is parallel to the foliation, i.e., is pre- or syntectonic, in contrast to *biotite* (X=colourless, Z=brown), most of which lies across the foliation and post-dates the main deformation ($s_{int}//s_{ext}$). The plagioclase ($An_{25-28}$) is moderately well twinned and unzoned. *Tourmaline* and *graphite* are the common accessories. Late strain-slip cleavage ("crenulation foliation" of Whitten, 1966, p. 230) has affected all minerals of the muscovite-rich schist and forms a distinct secondary s-surface at an angle to the main foliation.

According to Hoschek (1967a, p. 135), staurolite most commonly forms in lower amphibolite facies pelitic schists that contain more alumina, less alkalies and calcium, and have a Mg:Fe ratio slightly less than the average argillaceous sediment.

Between the staurolite-bearing schists and highly sheared and flow-folded marble lies biotite-chlorite-muscovite schist, which also is sheared and crenulated. *Biotite* (X=colourless, Z=brown) has formed randomly oriented porphyroblasts (0.3–1 mm) with helicitic quartz inclusions forming a $s_{int}//s_{ext}$. Most biotite porphyroblasts are partly replaced by

| | 1 | 2 |
|---|---|---|
| Plagioclase | 31.3 | 33.6 |
| Quartz | 33.2 | 34.7 |
| Garnet | 1.9 | 0.7 |
| Kyanite | 1.8 | – |
| Staurolite | 3.1 | 9.8 |
| Biotite | 10.7 | 17.0 |
| Muscovite | 16.8 | tr |
| Chlorite | – | 2.3 |
| Apatite | tr | tr |
| Zircon | tr | tr |
| Tourmaline | tr | 1.1 |
| Graphite | 1.2 | 0.9 |
| Black Fe oxide | tr | tr |
| Points counted | 2004 | 2001 |
| Plagioclase comp. | $An_{26-28}$ | $An_{25-27}$ |

1. Kyanite-garnet-staurolite-mica schist.
2. Garnet-staurolite-biotite schist.

*GSC*

TABLE V

*Modal analyses of staurolite-bearing schists from Petersen Bay.*

very fine grained quartz and muscovite of the matrix and partly altered to *penninite*. Late crenulation has crinkled the biotite and caused the formation of quartz "pressure shadows" at its margins. As in the kyanite-staurolite-muscovite schist, *muscovite* defines the foliation and clearly formed earlier than biotite. Muscovite and *quartz*, which make up separate lenses, as well as a large part of the matrix, are the most abundant minerals. Platelets of Fe-Ti oxide lying across the foliation are bordered by chlorite "pressure shadows." Sodic *plagioclase* (untwinned, R.I. less than quartz) and *epidote* are minor constituents. The rock is certainly of lower grade than the staurolite-bearing schists; whether this is due to retrogressive metamorphism or is an original feature cannot be said.

The amphibolite-schist sequence is considered to be part of a conformable metamorphic envelope around, and related to, the granitic gneiss complex, which is a culmination zone of metamorphism. The lower grade rocks south of the schists, about which little is known, may be an integral part of the envelope. Whether the envelope is complete or not is questionable. On the eastern shore of Yelverton Bay, as is described in the next section, a quartz monzonitic gneiss complex is flanked by a mixed gneiss, amphibolite and schist sequence many times thicker than the one in Petersen Bay. Still, the Petersen Bay schists are important because their kyanite and staurolite are useful indicators of physical conditions of metamorphism and the nature of the rocks before metamorphism.

## Yelverton Bay and Fiord

(Fig. 29D)

The Yelverton Bay region offers the finest known exposures of metamorphic rocks in northern Ellesmere Island. A variety of rocks of high to low metamorphic grade is exposed in sea-cliffs that rise more than 3,000 feet and extend 30 miles up the fiord.

The northernmost exposures of both sides of the bay consist of low-grade metamorphic rocks, not necessarily correlative. On the eastern side, these rocks are in contact with gneisses and amphibolites that decrease in metamorphic grade southward, passing into schists, phyllite, slates, and marble. At the extreme head of Yelverton Fiord (which opens into Yelverton Bay), Permo-Carboniferous sediments overlie the metamorphic rocks (not shown in Fig. 29). Near the western shore of the bay, bedded, unmetamorphosed volcanic strata are in fault contact with granitic gneisses.

The studies described here are of (a) a 12-mile stretch of the eastern coastline of the bay; and (b) the area adjacent to the volcanic rocks.

### The Eastern Coastline

#### *The Northern Low-grade Sequence*

The northern extremity of the eastern coastline is underlain by low-grade metamorphic rocks (map-unit 1c), many of which are reminiscent of the banded metamorphic sequence of northeastern Ayles Fiord. The Yelverton Bay rocks weather dark and commonly rusty, which distinguishes them from the leucocratic, fresh granitic gneisses to the south. This locality was Christie's only stop in northeastern Yelverton Bay on his 1954 sledging traverse (Christie, 1957). What he found supported his impression, gained from other areas, that quartz-mica schists and quartzite predominate in the metamorphic terrain between Ayles Fiord and Phillips Inlet (Christie, 1957, p. 10). The present work has shown that there is, in fact, more high-grade metamorphic rock west of Ayles Fiord than east of it.

The low-grade sequence comprises knotted garnet-chlorite-mica schist, largely of pelitic composition but containing thin psammitic layers, quartzite, and minor 1-foot-wide marble beds. Minor folds in quartzite and marble and foliation crenulation in schist are common.

The schists and impure quartzites show clear signs of retrogressive metamorphism. Garnet porphyroblasts in the schists have been rotated after growth, are bordered by "pressure shadows," and are partly to wholly altered to chlorite. Reddish brown biotite has been almost completely replaced by chlorite. Some specimens contain, in addition, a later, fresher looking biotite having X = brownish green and Z = dark greenish brown. Muscovite occurs both as late cross-mica in some rocks and as an older, deformed mineral in others. Plagioclase is invariably heavily sericitized and quartz is strained. Certain schists, particularly those near the granitic gneiss, are similar to the muscovite-rich schists of the Cape Columbia region.

### *The High- to Low-grade Sequence*

Southeast of the low-grade sequence just described lies a succession of rocks that range from high to low metamorphic grade. The superb exposures of these rocks are evident from the *Frontispiece*.

The succession may be divided into three zones, I, II, and III, which are marked on the *Frontispiece* and Figure 29. Zone I (map-unit 1a) includes granitic gneisses of generally uniform appearance, with only minor amphibolite layers. Zone II (map-unit 1a) contains a more heterogeneous assemblage of interlayered gneisses of varied composition, amphibolites and schists, as well as a probable granitic meta-plutonite. Zone III (map-unit 1c) consists of medium- to low-grade schists and marble. Modal analyses of representative rocks of each zone are given in Table VI and granitic varieties are plotted on the quartz-Na,K-feldspar-plagioclase diagram of Figure 14.

#### *Zone I*

Seen from a distance, Zone I appears as a rather uniform mass of leucocratic gneiss, cut by numerous deformed and less numerous undeformed mafic dykes. In detail, however, there is plenty of variation in texture, if not in mineralogy.

The typical rock is a (garnet-) biotite-muscovite-plagioclase-perthite-quartz gneiss. Common varieties are:

(1) Medium-grained; banded, with 2- to 3-mm biotite—and/or muscovite-rich layers (e.g., Table VI, no. 1).

(2) Medium-grained; banded, with 1- to 3-cm layers or lenses of quartzo-feldspathic material rimmed by biotite, alternating with well-foliated biotite gneiss (Table VI, no. 2).

(3) Augen-textured; augen 1- to 2-cm long of feldspar in a fine- to medium-grained, well-foliated biotite gneiss (Table VI, no. 3).

(4) Porphyroblastic to augen-textured; euhedral porphyroblasts of feldspar accompanied by augen of feldspar; both types of crystals are 3 to 4 inches long; groundmass is medium- to coarse-grained, poorly foliated biotite gneiss.

Hornblende-bearing varieties occur but are rare. These are associated with thin amphibolite layers and may represent granitized amphibolites.

Differences in grain size are in many instances attributable to different degrees of post-crystalline deformation, which has been intense in places (e.g., the contact with the low-grade rocks to the north). Yet, on the whole, cataclasis was not as pervasive in the Yelverton Bay

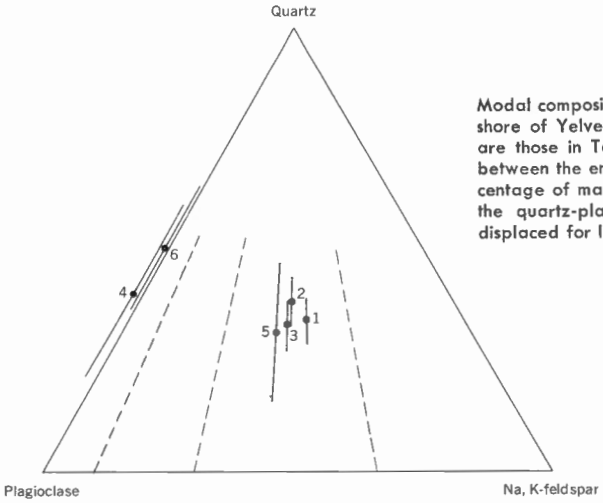
**TABLE VI** *Modal and chemical analyses of metamorphic rocks from the Yelverton Bay and Fiord area.*

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 | 13 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Plagioclase | 27.2 | 25.9 | 30.6 | 37.8 | 26.5 | 39.4 | 39.2 | 32.0 | 10.4 | – | 16.6 | | |
| K-feldspar | 29.9 | 24.5 | 27.8 | – | 21.4 | – | – | – | – | – | – | | |
| Quartz | 31.4 | 31.6 | 28.5 | 24.8 | 22.6 | 33.9 | 4.9 | 27.1 | 35.8 | 46.2 | 42.3 | | |
| Hornblende | – | – | – | – | 5.2 | 22.8 | 44.3 | 18.4 | – | – | 8.1 | | |
| Garnet | 0.9 | 2.9 | 2.8 | tr | – | – | – | 3.9 | – | – | – | | |
| Biotite | 6.0 | 8.8 | 7.1 | 36.4 | 12.4 | 2.9 | 6.9 | 7.7 | – | 20.1 | 19.7 | | |
| Muscovite | 3.8 | 6.3 | 1.8 | – | – | – | – | – | 7.9 | 4.1 | – | | |
| Epidote | 0.8 | tr | 0.8 | 1.0 | 1.3 | tr | 1.8 | tr | – | – | – | | |
| Chlorite | – | – | – | – | 8.4* | – | – | 1.9 | 15.9 | – | tr | | |
| Apatite | tr | tr | 0.6 | tr | tr | tr | tr | 1.8 | tr | tr | tr | | |
| Zircon | tr | tr | tr | tr | tr | tr | tr | – | – | tr | tr | | |
| Carbonate | – | – | – | – | – | – | – | 4.3 | 27.3 | 29.6 | 13.3 | | |
| Allanite | – | – | – | – | – | – | – | – | – | – | – | | |
| Sphene | tr | tr | tr | tr | 2.2 | 1.0 | 2.9 | – | – | tr | tr | | |
| Black Fe oxide | – | tr | tr | tr | tr | tr | tr | 2.9 | 2.7 | tr | tr | | |
| *after biotite | | | | | | | | | | | | | |
| Points counted | 1892 | 2006 | 1906 | 1852 | 1972 | 2032 | 1833 | 1955 | 1531 | 1601 | 1545 | | |
| Plag. comp. | $An_{25}$ | $An_{25}$ | $An_{23}$ | $An_{25}$ | $An_{27}$ | $An_{26}$ | $An_{40-32}$ | $An_{11}$ | n.d. | – | n.d. | | |
| | | | | | | | | | | | | | |
| $SiO_2$ | 70.7 | 74.9 | | | | 70.0 | 43.5 | | | | | 69.7 | 75.80 |
| $TiO_2$ | 0.52 | 0.36 | | | | 0.66 | 2.42 | | | | | 0.6 | 0.15 |
| $Al_2O_3$ | 14.0 | 13.2 | | | | 14.7 | 13.1 | | | | | 14.3 | 11.74 |
| Tot. Fe as $Fe_2O_3$ | 4.6 | 2.6 | | | | 5.4 | 18.2 | | | | | 3.8 | 2.63 |
| MnO | 0.07 | 0.03 | | | | 0.09 | 0.27 | | | | | 0.1 | 0.05 |
| MgO | 1.1 | 0.5 | | | | 2.1 | 7.5 | | | | | 1.2 | 0.54 |
| CaO | 2.6 | 1.1 | | | | 7.5 | 9.5 | | | | | 1.9 | 1.41 |
| $Na_2O$ | 3.7 | 3.8 | | | | 2.3 | 2.8 | | | | | 3.5 | 2.40 |
| $K_2O$ | 2.9 | 4.6 | | | | 0.4 | 0.7 | | | | | 2.4 | 4.51 |
| Total | 100.2 | 101.1 | | | | 103.2 | 98.0 | | | | | | |

Analysts (1,2,6,7): Rapid Methods Group, Geological Survey of Canada.

1. Zone I granitic gneiss.
2. Zone I garnet-biotite gneiss.
3. Zone I augen gneiss.
4. Zone II biotite gneiss.
5,6. Zone II hornblende-bearing gneisses.
7. Zone II amphibolite.
8. Zone III garnet-hornblende-quartz-feldspar schist.
9. Zone III chlorite-carbonate-quartz schist.
10. Zone III micaceous calcareous metasandstone.
11. Zone III hornblende-bearing calcareous metasandstone.
12. Average of 14 New Zealand greywackes (after Reed, 1957, quoted in Pettijohn, 1963, Table 7, col. C, p. S7).
13. Torridonian arkose from Scotland (Kennedy, 1951, quoted in Pettijohn 1963, Table 8, col. B, p. S9).

*GSC*




*GSC*

**FIGURE 14**

**Modal compositions of rocks from the eastern shore of Yelverton Bay and Fiord. Numbers are those in Table VI. The vertical distance between the ends of each line gives the percentage of mafic minerals. Points plotting on the quartz-plagioclase sideline have been displaced for legibility.**

region as it was in the Cape Aldrich–Cape Columbia area. Many Zone I rocks in thin section exhibit a texture more granitic than gneissic.

A striking feature of rocks of this region, as compared with those of the type area of the Cape Columbia Complex, is the abundance of *potassium feldspar*, in the form of microcline perthite. It forms both large porphyroblastic crystals and part of the groundmass, occurring in all specimens of Zone I rocks studied under the microscope. Perthite of stringlet type is homogeneously distributed through most grains and is probably of exsolution origin. Other grains have marginal flame perthite that tapers inward, suggestive of replacement. The large porphyroblasts have grown later than most of the groundmass grains, as their margins commonly include some of the latter. Most porphyroblasts show signs of post-crystalline deformation—cracks are filled with fine-grained recrystallized feldspar and margins are granulated—but, except for sporadic minor sericitization, are very fresh. Rocks with *plagioclase* ($An_{25}$) porphyroblasts (invariably accompanied by microcline porphyroblasts) are uncommon. The plagioclase is riddled with muscovite flakes along cleavage cracks and epidote particles. The difference in degree of alteration of the two feldspars may be due to the greater susceptibility of plagioclase to alteration. On the other hand, the marked contrast between the fresh microcline and the conspicuously altered plagioclase suggests that the latter is an earlier mineral.

Late potassium feldspar porphyroblastesis is a common feature of metamorphic terrains. It is generally associated with a temperature rise, which would support the hypothesis that Zone I is a core zone of high-grade metamorphism.

Except in their smaller grain size, groundmass plagioclase and potassium feldspar are identical to their porphyroblastic counterparts. Neither are there visible differences between feldspar of the leucocratic, biotite-rimmed layers and that of the enclosing rock in the gneiss of variety (2).

*Biotite* has X = pale yellow and Z = reddish or dark brown and is fresh, except in the cataclastic rocks, where it is partly to completely chloritized. *Muscovite* is generally intergrown at random with biotite but is less abundant. Pink *garnet* occurs in minor amounts in five of the ten Zone I specimens studied. *Epidote* (in places cored by allanite) and *sphene* are common accessory minerals associated with biotite.

Hornblende-bearing gneisses and amphibolite layers are similar to those in Zone II, where they are better developed, as described below. Post-metamorphic diabase dykes are common in Zone I and are described in a later section, under the heading "Minor intrusive rocks."

## *Zone II*

The gneisses of Zone II are characterized by heterogeneity and an abundance of hornblende (Fig. 15). However, the line between Zones I and II is rather arbitrarily drawn and large volumes of granitic gneiss typical of Zone I occur in the northern part of Zone II. The rocks of Zone II may be divided into three groups: (1) biotite-gneisses, (2) amphibolites and hornblende-bearing gneisses, and (3) a granitic cataclasite body.

### *Biotite Gneisses*

These are (garnet-) biotite gneisses of fine to medium grain; some are so rich in biotite that they might be termed schists (Table VI, no. 4). Rocks with little or no potassium feldspar are common but otherwise the hornblende-free gneisses differ little in mineralogy from those of Zone I. The Z colour of biotite is reddish brown but here and there it is green at

ends of flakes, which are associated with epidote. Such indications of retrogressive metamorphism are minor, however. Cataclasis is only locally conspicuous.

As with some of the rocks of Zone I, much of the biotite gneiss of Zone II contains segregations of leucocratic minerals. These are discrete lenses, rarely more than 2 feet long, made up of microcline perthite, plagioclase, and quartz (Fig. 16). The same minerals are found in the surrounding gneiss, which is separated from the lens by a narrow biotite-rich zone. It is clear from the photograph that the leucocratic material of the lenses could not have been injected into the gneiss from some source of granitic melt but must have formed in situ as neosome[1] derived from biotite gneiss paleosome[2]. Less common are granitic pods with ptygmatically folded "tails," irregularly distributed in the gneiss; such veined rocks might be termed arterite (Mehnert, 1968, p. 353) and the granitic component may be an injection from an external source.

The hornblende-free, quartzo-feldspathic gneisses probably represent metamorphosed sediments. Chemical analyses of two typical rocks (from Zone I) are presented in Table VI and compared with greywacke and arkose analyses given by Pettijohn (1963). Sandstones are suggested parent rocks because of high silica and moderate alumina in the gneisses.

### *Amphibolites and Hornblende-bearing Gneisses*

Amphibolite layers range in thickness from a few inches (Fig. 17) to 30 feet. They are generally concordant with the foliation but locally cut it. Discordant varieties are particularly evident among the thinner amphibolite layers. Intense deformation is signified not only by numerous irregular masses of amphibolite enclosed by gneiss but also by folded quartzo-feldspathic veins and lenses in the amphibolites. Many of the thinner amphibolites are boudinaged (Fig. 15).

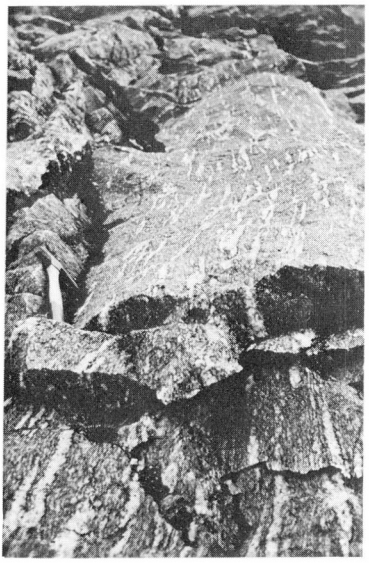
Mineralogically, the amphibolites are characterized by *hornblende* with X = yellowish green, Y = green, and Z = grass green with bluish tinge. *Plagioclase* is calcic oligoclase or andesine. In some rocks, normal zoning in plagioclase is conspicuous ($An_{40-32}$), which may be indicative of an igneous parent rock. Red-brown *biotite* is commonly present as randomly oriented flakes and has been intensively *chloritized* in places. *Epidote* and *sphene* occur in all specimens studied and are abundant in some. Significantly, *quartz* is either absent or

[1]Neosome = the newly formed part of a migmatite (Mehnert, 1968, p. 8).
[2]Paleosome = the unaltered or only slightly modified part of a migmatite (Mehnert, 1968, p. 7).

FIGURE 15

Gneisses and amphibolites in Zone II, eastern shore of Yelverton Fiord. Note the boudinage of some of the amphibolite layers. The vertical diabase dyke at right is undeformed and unmetamorphosed. (Field photograph, July 1966.)

202022-G



202022-H

FIGURE 16

Quartzo-feldspathic lenses in biotite gneiss of Zone II, eastern shore of Yelverton Fiord. Note the dark borders of the lenses, where biotite is concentrated. (Field photograph.)

202022-I

FIGURE 17

Largely concordant amphibolite layers in Quartzo-feldspathic gneisses of Zone II, eastern shore of Yelverton Fiord. (Field photograph.)

present in only accessory amounts, which favours igneous parentage for these rocks. *Garnet* and *scapolite* are rare. Number 7 in Table VI is a mode of a typical amphibolite.

As might be expected in a zone of such intense metamorphism, relict textures diagnostic of a particular origin of the amphibolites are lacking. The crosscutting nature of some layers (Fig. 15) is suggestive of an original intrusive relationship. Dolomitic rocks, which might have participated in formation of amphibolite, were not found in Zone II (but marbles are common in Zone III). Many amphibolites are associated with hornblende gneisses but others lie in rocks completely free of hornblende. Metamorphic differentiation could hardly be responsible for the formation of 30-foot-thick amphibolite layers having sharp contacts with hornblende-free gneisses. Thus a meta-igneous (volcanic or hypabyssal) origin appears likely for at least some amphibolites.

There is complete gradation between amphibolites with little plagioclase and no quartz to feldspar-quartz gneisses with scattered hornblende grains. Indications of this gradation may be seen in marginal parts of thin (6-inch) amphibolite layers. The centre of the layer is highly mafic, whereas the margins are comparatively rich in plagioclase. The bordering leucocratic gneiss may or may not be hornblende-bearing. When hornblende is present, the amount is far less than in the amphibolite. Such amphibolite-hornblende gneiss associations may be examples of metamorphic differentiation.

The chemical composition of an amphibolite (Table VI, no. 7) from a one-foot-wide layer in hornblende-free granitic gneiss in Zone I indicates a meta-igneous origin. Titania content is high (2.42%); potash is low (0.7%), leading to a low Niggli $k$ number (0.13). These are reliable criteria of an igneous parent rock (Leake, 1964, p. 239).

The chemical analysis of a hornblende gneiss (Table VI, no. 6) associated with amphibolite suggests that it may be an orthogneiss, because of very low $K_2O$. Perhaps redistribution of material during metamorphism has resulted in a mafic rock becoming more felsic, yet preserving its original low potash content.

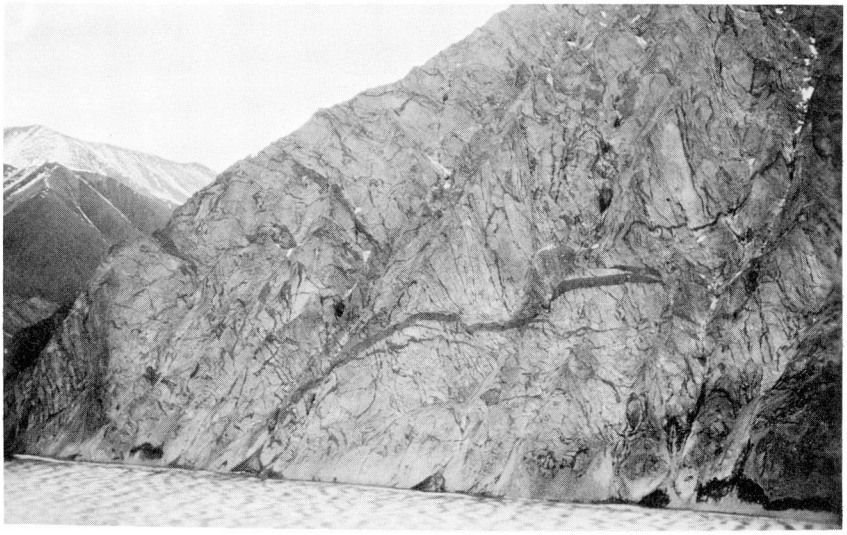
Varieties of hornblende-bearing gneisses containing potassium feldspar (Table VI, no. 5) may represent products of granitization. That the potassium feldspar was introduced late in the formation of the rock (probably by metasomatic processes) is suggested by its confinement to discrete, thin, hornblende-free layers, between which lie layers of hornblende, plagioclase, and quartz.

It seems very probable that the amphibolites and hornblende-rich gneisses of Zone II formed in more than one way. However, one possibility that should be kept in mind is that these rocks together constitute a basic "behind," composed of mafic elements expelled during the granitization that presumably formed Zone I, which is remarkably poor in metamafic rocks.

### *The Granitic Cataclasite*

The line dividing Zones II and III is drawn at the southern border of a granitic cataclasite, (*see* Fig. 18). The mass is immediately distinguished from the bordering rocks by its lack of foliation, abundance of contorted mafic lenses and homogeneous, leucocratic appearance. Its exposure along the shore is approximately a mile long.

Severe cataclasis of the granitic rock has been partly obscured by recrystallization, particularly of quartz, which has formed aggregates of grains 0.5 to 2 mm in size and is



202022-J

FIGURE 18. Granitic cataclasite body at the boundary between Zones II and III, eastern shore of Yelverton Fiord. The dark stringers are stretched and shredded amphibolite masses. A large unmetamorphosed diabase dyke has intruded the body. Compare the lack of structure within the main granitic mass with the excellent foliation of the border rocks on the left. Pseudotachylite(?) occurs in the contact zone between the gneisses and the cataclasite. (Field photograph, July 1966.)

only moderately strained. The other medium-grained constituent is microcline perthite as porphyroclasts up to an inch square. Numerous cracks in the crystals are filled with fine recrystallized microcline and quartz. Among the quartz aggregates lie finely crushed feldspar and quartz and fine-grained muscovite; the latter has probably resulted from alteration of feldspar. Most of the groundmass feldspar is well-twinned microcline; what plagioclase there is has been heavily sericitized. Scattered patches of brown biotite, largely altered to a greenish brown variety, are smeared out between grains of feldspar and quartz. Garnet was noted in hand specimen but not found in thin section.

The border rocks are foliated but of similar mineralogy, and cataclastic features are as marked as in the non-foliated rocks. Patches and veinlets of dark material with inclusions of fragmental feldspar and quartz are isotropic and may be pseudotachylite. These foliated border rocks either are a part of the cataclasite mass or belong to Zone II gneisses that were affected by the cataclasis. The change from foliated granitic gneiss to non-foliated granitic cataclasite is very abrupt, which suggests that the two rock types are separate entities. Yet their mineralogical similarity is so strong that the foliated gneisses may represent an original foliated border zone of a former pluton that underwent severe post-crystalline deformation.

On closer inspection of outcrop, mafic inclusions are seen to be even more common than a distant view indicates. All inclusions studied are poorly foliated amphibolite, in all shapes and sizes, from shreds a few inches long to large, highly contorted dyke-like bodies in completely random orientation. These may be former intrusive bodies. Late, intrusive diabase dykes, which are unmetamorphosed, cut all rocks of the cataclasite mass.

### *Zone III*

South of the cataclasite body, gneisses are absent and schists steadily diminish in metamorphic grade southward. In this direction, the rock succession of Zone III, as far as it was studied, is as follows.

(1) Garnet-chlorite schist. This rock is identical to rocks found in the low-grade sequence north of Zone I, in Markham Fiord and elsewhere.

(2) Hornblende-chlorite-plagioclase-carbonate-quartz schist, with or without garnet and biotite (Table VI, no. 8). Ragged porphyroblasts and smaller grains of bluish green hornblende lie in a matrix of highly sericitized plagioclase and quartz, with chlorite-epidote and carbonate patches. Garnet occurs in pink porphyroblasts up to 4 cm across, which are ragged and highly sieved with inclusions of groundmass minerals. Biotite has X = pale yellow with greenish tinge and Z = dark brown with greenish tinge. Muscovite is a constituent of some specimens. The ragged garnet, greenish biotite, and extremely sericitized plagioclase indicate retrogressive effects.

(3) Chlorite-carbonate-quartz schist with pyrite cubes (Table VI, no. 9). The chlorite is probably a retrograde mineral but no relict parent remains. Deformation is indicated by folded psammitic layers.

(4) Marble, partly dolomite.

(5) Micaceous calcareous sandstone (Table VI, nos. 10, 11). This is the lowest grade metamorphic rock found on the 12-mile stretch of coast studied. Beds, a few inches thick, of fine-grained, mica-poor sandstone alternate with slightly coarser grained, mica-rich sandstone. The grain size difference is probably relict graded bedding. Carbonate content is variable. Quartzite interbeds show pinch-and-swell structure, which is likely to be an original sedimentary feature. Had it been formed during deformation accompanying metamorphism, similar structures and boudinage should have been produced elsewhere in Zone III but such

were not seen. On the other hand, the micaceous sandstones are tightly folded and the quartzite beds are approximately parallel to axial planes of isoclinal folds, so may owe their shape to stretching during folding.

Within Zone III, calcareous sandstone shows prograde metamorphism from south to north. Three specimens were studied in thin section. The southernmost specimen shows the beginnings of phlogopitic biotite growth in a very fine grained carbonate-quartz sandstone. A little farther north, in an apparently similar rock, phlogopitic biotite porphyroblasts are of greater abundance and size but still without preferred orientation. Still farther north, in the most metamorphosed rock, biotite is more iron-rich (X = yellow-brown, Z = dark brown) and defines a foliation. Carbonate is less abundant than in the other two specimens and epidote and sphene have appeared. Quartz and feldspar (untwinned, R.I. less than quartz) make up the remainder of the groundmass.

One specimen of a schistose sandstone (Table VI, no. 11), rich in feldspar, carries highly sieved, pale green amphibole porphyroblasts in apparently early stages of growth, associated with phlogopitic biotite. Groundmass consists of feldspar, quartz, carbonate, and a little chlorite. The original sediment was doubtless a calcareous sandstone of a composition different from the one described above.

### *Minor Intrusive Rocks*

Rocks of the northeastern coastline of Yelverton Bay are intruded by diabase dykes that post-date the metamorphism. The dykes possess chilled borders and igneous textures, and are clearly discordant to the country rocks (Fig. 15). Specimens studied consist of augite, labradorite, and black iron "ore." Dykes were observed to be abundant in the gneisses and higher grade schists of Zones I and II, but none was seen in the southern part of Zone III. These observations agree with others made of the metamorphic complex of northern Ellesmere Island. It appears that abundance of post-metamorphic mafic dykes is a characteristic feature of the higher grade parts of the northern Ellesmere metamorphic complex with, however, the notable exception of the Cape Columbia area.

Pegmatite dykes and sills also are common in Zones I and II. They are of two types: one is a generally concordant, deformed variety, 4 to 6 inches thick; the other is discordant, rather straight and up to several feet thick. The first type is considered to be an exudation or "sweat-out" from granitic country rock and the second is probably a later, post-tectonic, magmatic intrusion.

## The Western Shore of Yelverton Bay

From aerial reconnaissance it appears that granitic gneiss, probably equivalent to Zones I and/or II of the eastern shore of the bay, forms a continuous belt of high mountains extending from west of Yelverton Bay to north of the head of Phillips Inlet.

Two miles inland from the western shore and directly west-southwest of the tip of the peninsula west of Yelverton Fiord, the granitic gneiss is bordered by a narrow belt of volcanic rock of unknown age. The volcanics form cliffs of alternating dark red and green flows and breccia, which are altered and devitrified but apparently unmetamorphosed. They appear to be of intermediate composition. The volcanics are steeply tilted at their contact with the gneisses, but elsewhere are horizontally bedded. The adjacent metamorphic rock is a breccia composed of quartz, microcline, and plagioclase fragments, with mica aggregates, in a mylonitic, quartzo-feldspathic groundmass. A little farther from the contact, the rock is a cataclastic granitic gneiss with abundant, fractured microcline perthite porphyroclasts.

The gneisses have a southward dip similar to that on the eastern shore of Yelverton Fiord and the bulk of them occur at a considerably higher stratigraphic level than the volcanics. Evidently, the gneisses have been thrust over the volcanics, indicating northward thrusting in the Yelverton region. Such thrusting was suspected at the contact between Zone I and the low-grade sequence on the northeastern coast.

The age of the volcanic rocks is not known but they may be correlative with the "basaltic conglomerate" found by Christie (1957, p. 22) at Hansen Point, at the mouth of Petersen Bay, and suggested by him to be of Mesozoic or Cenozoic age, because of its fresh appearance.

### Age of the Metamorphic Rocks of the Yelverton Fiord Region

The metamorphic rocks of Zones I and II on the eastern shore of Yelverton Bay form part of a continuous belt of gneisses and related rocks of amphibolite facies grade extending from the western shore of Ayles Fiord to the head of Phillips Inlet. Presumably, all these rocks are contemporaneous.

K-Ar age measurements have been made on micas from a biotite-muscovite-feldspar-quartz gneiss from Zone I (*see* Fig. 29D). Two determinations made at different times on different fractions of a biotite concentrate gave apparent ages of crystallization of $189 \pm 9$ m.y. and $177 \pm 8$ m.y., i.e., Early Jurassic, according to the Geological Society of London time-scale (1964, p. 262). These young ages are unlikely to have geological significance and illustrate the hazards inherent in the use of the K-Ar technique for dating metamorphic rocks from an orogenic belt with a complex history. Thus the age of metamorphism of the Ayles Fiord–Phillips Inlet gneissic belt and its relation to the Cape Columbia Complex remain unknown.

Similarly, uncertainties exist concerning the age of the lower grade rocks of the Yelverton region, e.g., Zone III. The latter may be contemporaneous with Zones I and II, indicating increasing metamorphic grade from southeast to northwest. On the other hand, Zone III may be younger than Zones I and II and was brought to the same level as the latter by differential uplift of the high-grade rocks. The calcareous sandstone in Zone III, described above, is comparable in lithology to, and lies on strike with, the Silurian and (?)Upper Ordovician Imina Formation of northwestern Ellesmere Island (H. P. Trettin, 1968, pers. com., and 1971).

## Other Areas of Regional Metamorphism in Northern Ellesmere Island

In addition to the areas described in the preceding pages, several other terrains showing regional metamorphism of largely greenschist facies grade either have not been visited by the writer or have not been studied in detail.

On Parr Bay, immediately east of the type area of the Cape Columbia Complex (Fig. 29A) strata of the Middle Ordovician(?) or older Mount Disraeli Group have been metamorphosed to slates and phyllites, which appear to grade into the lower grade rocks of the Cape Columbia Complex (Trettin, 1966).

An extensive terrain between Yelverton Fiord and Henson Bay shows regional metamorphism that is apparently related to granitic plutons of probably Middle to Late Devonian age (*see* Chap. III; Trettin, 1971). In the vicinity of some of the intrusions, metasedimentary rocks, some of which are as young as Late Silurian, have been regionally metamorphosed to garnet grade (Trettin, 1971). This region is especially significant because it provides evidence of a regional metamorphic phase in northwestern Ellesmere Island, probably during Devonian time.

## Metamorphic Facies Series in Northern Ellesmere Island

Miyashiro (1961) introduced the useful concept of the metamorphic facies series to denote the sequence of facies in a metamorphic terrain. Taking into account world-wide similarities in metamorphic zoning, he proposed five facies series. Because pressure-temperature gradients for the various facies series are now known to good approximation, identification of the facies series to which a particular metamorphic terrain belongs gives some idea of the physical conditions of its formation.

Our limited knowledge of the distribution and age of metamorphism in northern Ellesmere Island requires that caution be exercised in making any statements implying that the various metamorphic terrains constitute a single entity. Type and grade of metamorphism have varied and there is clear evidence of regional metamorphism at different times in different places.

Nevertheless, despite the fact that not all the metamorphic rocks are contemporaneous, sufficient similarities exist between them that a few generalizations may be made.

The regionally metamorphosed areas of northern Ellesmere Island are characterized by:

(1) Extensive development of greenschist and amphibolite facies; the intermediate albite-epidote-amphibolite facies is less well developed.
(2) Presence of kyanite and staurolite.
(3) Absence of andalusite and cordierite. Sillimanite is very rare in the regionally metamorphosed rocks.

These features indicate that metamorphism was of kyanite-sillimanite (Miyashiro, 1961) or Barrovian (Winkler, 1965, p. 70) type. The fact that kyanite, not andalusite or sillimanite, is the aluminosilicate polymorph developed suggests relatively steep pressure-temperature gradients.

## ADDENDUM (March 1974)

After this report was submitted, whole-rock Rb/Sr isochron age investigations were initiated by A. K. Sinha (Virginia Polytechnic Institute and State University, Blacksburg, Va.) on rocks collected by the author from the Cape Columbia Complex and the gneiss terrain between Ayles and Yelverton Fiords. These studies have shown conclusively that the gneisses are of Precambrian age. Full details will be given in a later publication.

# *Chapter III*

# PLUTONIC ROCKS

## Mafic and Ultramafic Rocks

The main members of this group in northern Ellesmere Island are:

(1) The Cape Fanshawe Martin intrusion, a layered gabbro-peridotite mass, which is of Early Devonian age according to a radiometric determination.

(2) Two large, alpine-type serpentinite bodies near the head of M'Clintock Inlet. They are Early Devonian or older, on the basis of radiometric dating.

(3) Altered mafic and ultramafic minor intrusions on Bromley Island, east of Cape Fanshawe Martin. These have been described by Trettin (1969b, p. 58), who ascertained that they are of pre-Middle Ordovician age. Unaltered, post-Middle Ordovician mafic intrusions are present also in this area and on M'Clintock Inlet.

(4) A small, partly metamorphosed ultramafic body in the Cape Columbia Complex. This mass was probably emplaced in either early Paleozoic or Precambrian time.

### Cape Fanshawe Martin Intrusion
(Fig. 29D; map-unit C)

The coast on the eastern side of the mouth of Ayles Fiord and around Cape Fanshawe Martin is underlain by an ultramafic-mafic pluton, first recognized by Christie (1957). The intrusion was included with the "plutonite of Cape Richards ... a complex association of ... granite and ... norite and ... peridotite" and it was noted that the peridotitic rocks are cut by the "granite" (Christie, 1957, p. 24). The present work has shown that granitic rocks are, in fact, not abundant in the ultramafic-mafic intrusion, which although spatially associated with a major granitic complex at Cape Richards (the next cape east) is actually a separate body.

The mafic complex of Cape Fanshawe Martin is about 7 miles long and 4 miles wide; a considerable part of it is glacier-covered. Its northeastern extremity is under ice, and its middle part is cut by a 2-mile-wide glacier. Exposure is further diminished by extensive frost heaving, particularly on the inland slopes, but good outcrops are found in scattered areas and in sea-cliffs.

Well-exposed in the sea-cliff on the northeastern shore of Ayles Fiord, the southwestern end of the intrusion is in intrusive contact with marbles of the metamorphic "banded sequence" (map-unit 1c; Fig. 19). Xenoliths of marble are present in the marginal intrusive rocks. Viewed from a distance, the contact between the igneous and the country rocks to the north appears to be similarly sharp, with dark igneous-looking (gabbroic?) rocks to the west and light marbles to the east.

At the contact, the marbles show intense cataclasis and recrystallization. Part of the cataclasis might be attributed to deformation that affected all the rocks of the area, but the marble farther from the contact is much less deformed. There is no evidence of extensive calc-silicate development; this is probably due to the purity of the calcium carbonate marble of the wall-rock. The gabbro at the contact is considerably finer grained than its kindred farther into the intrusion—doubtless the result of chilling—but otherwise is similar.

The Cape Fanshawe Martin intrusion was found to be distinctly zoned (Fig. 19). Gabbroic rocks form the bulk of the exposures but pass into peridotitic rocks near the cape, which appears to be in the central part of the body. This progression from mafic to ultramafic composition inward from the sides is reflected also to some degree in cryptic zoning of olivine and plagioclase, described below.

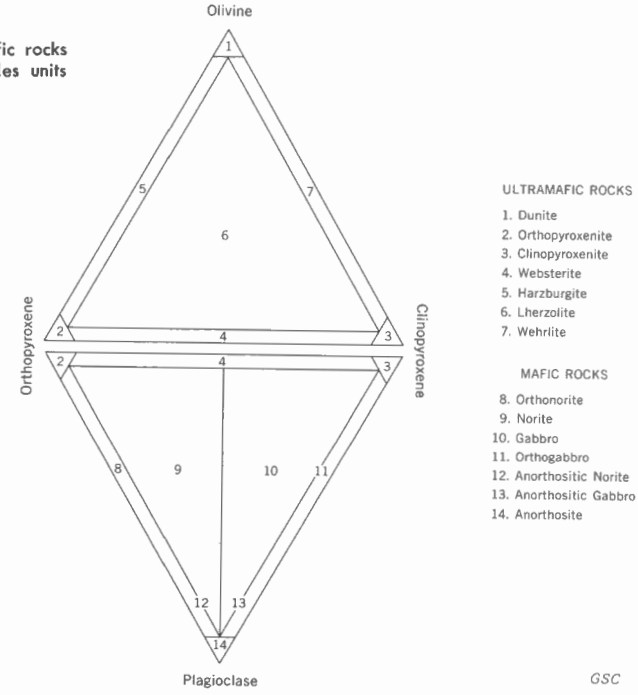
FIGURE 19. The Cape Fanshawe Martin gabbro-peridotite intrusion at the entrance to Ayles Fiord; view to the east. The contact between the dark intrusion and metamorphic rocks of greenschist facies-grade is marked by a broken line. The approximate boundary between the "inner zone" (IZ), much of which is ultramafic, and the "outer zone" (OZ), chiefly gabbroic, of the intrusion is shown by a dotted line. Primary flow structure (igneous lamination) is symbolized by ⊸. Farther east, the Cape Richards felsic intrusive complex (CR) and, in the distance, M'Clintock Inlet and Ward Hunt Island, are visible. (Oblique air photograph) RCAF T407 R-217, July 15, 1950.

The rocks at the southern end of the intrusion, near the contact with the metasedimentary rocks, comprise normal gabbros, olivine gabbros, and norite. About 5 miles to the north and a mile in from the contact, plagioclase-rich gabbros, some of which carry significant amounts of biotite, are common.

The intrusion may be divided into a mafic outer zone and an ultramafic inner zone; the boundary between them, however, is rather arbitrarily drawn (Fig. 19). Rock nomenclature follows the classification given in Figure 20. Peridotite, as used in this report, includes units 5, 6, and 7 in Figure 20.

FIGURE 20

**Classification of mafic and ultramafic rocks used in this report. Peridotite includes units 5, 6, and 7.**




## *Petrography and Structure*

### *Outer Zone*

Rocks of the marginal zone were investigated mainly at the southern end of the intrusion. There they differ from rocks of the inner zone by (1) being poorer in olivine and richer in feldspar, (2) commonly having minor quantities of brown hornblende and biotite, and (3) being more altered. Layering appears to be less common than in the inner zone and only in the outer zone were granitic rocks found, but these differences may be due to inadequate sampling and therefore may not be real.

The outer zone consists of gabbros, both with and without orthopyroxene, and anorthositic gabbros, anorthosites, and olivine gabbros. No zonation of rock types was recognized. Textures are generally hypidiomorphic-granular or subophitic to ophitic.

Of eleven outer zone rocks studied microscopically, two contained fresh *olivine* and two contained carbonate pseudomorphs after olivine. X-ray determinations ($d_{174}$ spacings) of the composition of olivine gave 69 and 74 mol. per cent forsterite. The $Fo_{69}$ olivine coexists with plagioclase of $An_{76-69}$, against which it shows a kelyphitic rim of tremolite?, uralitized clinopyroxene, some brown to green hornblende, and a little biotite. The $Fo_{74}$ olivine is in a more mafic, ophitic-textured rock, with fresh clinopyroxene and minor orthopyroxene, plagioclase of $An_{83-78}$, and minor phlogopitic biotite.

*Orthopyroxene* is rarely noticeably pleochroic and generally shows a very fine lamellar structure under crossed nicols. The lamellae may have developed as a consequence of strain, rather than from exsolution of pyroxene of different composition. Bending and wavy extinction of much of the orthopyroxene supports such an interpretation. On the other hand, irregular blebs of pyroxene in some grains probably did form through exsolution. Zoning is rare (occasional thin rims of different birefringence than core) or entirely lacking. The optical data, $2V_\alpha = 69°$, $n_\gamma = 1.697$, indicate a composition $En_{76}Fs_{24}$ (Deer, Howie, and Zussman, 1963a, p. 28).

Most of the *clinopyroxene* is zoned, showing a colourless or neutral core and a brownish or pink rim. Clinopyroxene not visibly zoned is mostly brown or pink and, judging from the local intensity of the latter colour, may be markedly titaniferous. Optical data are as follows: $2V_\gamma$ (colourless or neutral) $= 46°$, $2V_\gamma$ (pink) $= 41°$, $n_\beta$ (pink) $= 1.701$. These data place the clinopyroxene in the augite field of the pyroxene quadrilateral (Deer, Howie, and Zussman, 1963a, p. 1, 132).

*Plagioclase* is generally zoned but rarely displays a wide range of composition within any one rock. Compositions of cores are $An_{70}$ to $An_{80}$, on the average, but rims in the more differentiated gabbros are andesine (one is as sodic as $An_{46}$, where the core is $An_{61}$). Determinations in most cases fell on or near the low-temperature curves of Slemmons (1962) and Uruno (1963). Values falling half-way between the low- and high-temperature curves are not considered optically significant. The author follows Burri, Parker, and Wenk (1967, p. 191) in believing that these points resulted from either measurement errors or imprecisely located curves or both. Disordered plagioclase was not found.

*Hornblende* (excluding uralite) and *biotite* are strictly subordinate in all rocks but appear to be pyrogenic. Both minerals are reddish brown and are obviously later than pyroxene. While hornblende invariably replaces pyroxene, biotite seems to do no more than fill interstices. In the more altered and differentiated gabbros, hornblende may show greenish rims. *Biotite* is Mg-rich (phlogopitic; Z = pale yellow-brown) in the less differentiated gabbros, and more Fe-rich (Z = deep reddish brown) in the more differentiated rocks. Uralitic hornblende is described in the discussion of alteration effects, below.

Minor amounts of interstitial micropegmatite are present in some of the more differentiated gabbros.

Alteration of the gabbros is much more intense in the margins of the intrusion than in the central parts. For reasons given below, the alteration does not appear to be the result of incipient metamorphism but rather of the usual deuteric effects associated with magmatic intrusion. There are three main manifestations of this alteration: carbonate replacement of olivine, uralitization and chloritization of pyroxene, and saussuritization of plagioclase. As olivine in the more mafic inner part of the intrusion is altered to serpentine, never to carbonate, the olivine→carbonate reaction in the marginal rocks can be ascribed to access of $CO_2$ from the country rock marbles.

Uralitization of pyroxene does not seem to depend on its composition, because instances of orthopyroxene alone, clinopyroxene alone, or both being uralitized were found among rocks containing two pyroxenes. Some specimens in thin section show clearly the progression pyroxene→uralite→brown hornblende→green hornblende, indicating that the uralitization was a relatively early phenomenon in the crystallization of the magma. It also lends credence to the hypothesis that uralitization is a deuteric alteration, in the sense of Sederholm (1916), who originally defined the latter as metasomatic changes taking place "in direct continuation of the consolidation of the magma of the rock itself."

Saussuritized plagioclase is of common but irregular occurrence in the intrusion. It is especially prevalent in some of the marginal gabbros but appears to be absent from feldspar-bearing central peridotites. The usual direct correlation of intensity of saussuritization with anorthite content is not apparent. Furthermore, one uralitized gabbro from near the intrusive contact contains low-iron zoisite (abnormal blue interference colours, parallel extinction) in 0.5- to 1-mm grains and clusters, invariably associated only with plagioclase, which is otherwise fresh.

In summary, there is evidence that fluids were localized in the marginal zones but varied in their distribution and effect.

A striking feature of the southern end of the intrusion is the sporadic occurrence of granitic rock intrusive into gabbros. Angular fragments of dark gabbro, ranging in size from a few inches to more than a foot, may commonly be observed enclosed in a quartz monzonite alaskite. The alaskite is fresh and consists of microcline perthite, oligoclase ($An_{20-25}$), slightly chloritized biotite (Z = deep reddish brown, which indicates a high $TiO_2$ content, according to Deer, Howie, and Zussman, 1962b, p. 71), and quartz, with a little allanite among the accessory minerals. This rock in places has a knife-sharp contact with the gabbro, which is generally uralitized, but elsewhere alaskite is separated from gabbro by a reaction zone of hornblende-bearing leucocratic rock. It is unfortunate that the gabbro-felsite association could be observed only in loose blocks on frost-heaved surfaces, commonly only a few square yards in area. Christie (1957, p. 24) states that "granite, variations of which are aplite, cuts the peridotitic rocks." The present writer did not find such relationships but did recognize syenitic dykelets and veins in the more ultramafic inner zone.

Chilled margins were nowhere found in the gabbroic and granitic rocks, which indicates that they came together when both were hot. The reaction zone of intermediate composition between some of the felsic and mafic rocks also attests to a similarity in temperature of the rocks, as it is unlikely that a relatively refractory gabbro would be much affected by a lower temperature granitic melt. Thus one may infer that, although the felsic rock is slightly younger than the mafic rock, the two are essentially contemporaneous.

The granitic melt may have been (1) a differentiate of the gabbro, or (2) derived from an unrelated source. Evidence for (1) is provided by: (a) the apparent contemporaneity of the felsic and mafic rocks; (b) the fact that the parent magma of the Cape Fanshawe Martin intrusion was tholeiitic (hypersthene is common and interstitial granophyre is present); and (c) the very small amount of granite and the very large quantity of mafic rock.

A rock similar to the one described here forms veins in the Freetown layered gabbroic complex of Sierra Leone (Wells, 1962, p. 95–97). Wells cites evidence that the granite is genetically related to late basic dyking in the complex. In a valuable review of the felsic-mafic rock association, King (1963) summarizes the abundant evidence for the development of a minor granitic residuum from a tholeiitic magma.

Seemingly unfavourable to this differentiation hypothesis is the lack of rocks intermediate in composition between gabbroic and granitic in the Cape Fanshawe Martin in-

trusion. Bowen (1928, p. 90–91) accounted for this by pointing to the high fractionation of plagioclase, which results in the dioritic phase being represented by certain zones in the plagioclase crystals and not by actual bodies of rock. The sodic labradorite to andesine zones in some of the marginal rocks lend support to this idea. Results of experimental petrology have led Wyllie (1963) to postulate that the liquidus path for magma of intermediate composition, plotted against temperature, is rather shallow, whereas it is steep for mafic and felsic magmas. Cooling through a small temperature interval suffices to change a mafic liquid to a felsic one without an intermediate stage of differentiation. Rapid cooling would be expected in a moderate-sized body such as the Cape Fanshawe Martin intrusion. Wyllie (1963, p. 210) agrees with Bowen that the intermediate stage is represented by zones of intermediate composition in plagioclase crystals.

A separate source of the granitic melt might be sought in the "granite" pluton at nearby Cape Richards or even in granitic rock of the basement, which surely underlies the Cape Fanshawe Martin intrusion and is actually exposed to the south and west. One cannot deny the possibility of such an origin, but derivation from the mafic parent magma of the intrusion seems more plausible.

The scarcity of good continuous outcrop hinders recognition of internal structures in the gabbros. Alternation of 3- to 4-inch layers of mafic gabbro and gabbroic anorthosite, presumably an example of rhythmic layering, was observed in one outcrop. Igneous lamination of platy minerals was not seen. Cumulate textures are rare in the outer zone but one rock examined has plagioclase laths, a cumulus phase, enclosed in large (1–2 cm), unzoned clinopyroxene crystals. This is probably a heteradcumulate (Wager, Brown, and Wadsworth, 1960, p. 18).

### *Inner Zone*

Most rocks of the inner zone are more mafic than those in the outer zone and the layering characteristically associated with such rocks is well developed and common.

Peridotite, troctolite, olivine gabbro, olivine-gabbroic anorthosite, and pyroxenite are the main varieties. The pyroxenites occur in layers 1 to 2 inches thick in peridotites and troctolites. An example studied in thin section consists of neutral-coloured clinopyroxene (average grain size 3 mm, maximum 10 mm) that is slightly uralitized, a little serpentinized olivine, and scattered flecks of phlogopitic biotite. The contact of this layer with the enclosing troctolite is irregular and interfingering. The significance of this is discussed below.

All gradations may be found from olivine gabbros through peridotites and troctolites to dunites, but dunites are rare. Pegmatitic segregations, with pyroxene in euhedral grains up to 3 inches long, occur in the gabbros.

*Olivine* in the olivine gabbros ranges from $Fo_{71}$ to $Fo_{84}$ but averages $Fo_{75}$. It is generally fresh. *Orthopyroxene* is much less common than clinopyroxene. It was found in only one of the seven gabbroic rocks studied microscopically, where it is clearly later than the clinopyroxene. It is markedly pleochroic (X = pale pink, Y = very pale brown, and Z = pale green), has a $2V_\alpha = 55°$ and $n_\gamma = 1.710$ and so has the composition $En_{63}$ $Fs_{37}$ (Deer, Howie, and Zussman, 1963a, p. 28).

In contrast to its colour in the outer zone, the *clinopyroxene* of the inner zone rocks is neutral or very pale brownish. Optical data are: $2V_\gamma = 46°–52°$ and $n_\beta = 1.679–1.688$. These are properties of Ca-Mg clinopyroxenes, which typically are formed early in the differentiation of basaltic magma. Numerous grains, particularly in their cores, show platelets of a black opaque mineral (presumably magnetite) exsolved in a schiller structure. Such a structure

can be produced quite simply in the laboratory by heating augite crystals (Wells, 1962, p. 46) (in air, presumably); although, admittedly, conditions might be somewhat different in natural rocks, the exsolved opaques might be regarded as indicative of thermal metamorphism, perhaps by a later injection of magma of the main intrusion, or of a dyke. Alteration of the pyroxenes is generally slight but some of the pyroxenites are considerably uralitized. *Plagioclase* is commonly fresh and slightly zoned (less than 10 per cent difference in An content between core and rim) and is in the low-temperature form. Its composition, labradorite to bytownite, varies roughly with the forsterite content of coexisting olivine—the more forsteritic the olivine, the more calcic the plagioclase. This is to be expected, as the plagioclase in these rocks appears to be a cumulus, rather than an interstitial, phase. Brown *hornblende* and *biotite* are found in several rocks as minor, late pyrogenic minerals. Interstitial *micropegmatite* is present in one uralitized gabbro.

The olivine-rich rocks of the troctolite and peridotite groups are particularly abundant on the seaward side of the inner zone. From Cape Fanshawe Martin inland the ultramafic rocks grade into feldspathic gabbros. The peridotites and near-dunites (olivinites with a little plagioclase) weather brown when unserpentinized and black when serpentinized. On a freshly broken surface they are green or black. Serpentinization is seldom intense and is usually restricted to thin layers between fresh olivine. There are, however, outcrops of completely serpentinized rock consisting of alternating layers of green and black serpentine.

*Olivines* from two peridotites, a troctolite, and a dunite, yield X-ray compositions of 82 to 86 mol. per cent forsterite. The ranges of *plagioclase* compositions is $An_{70}$ to $An_{89}$ (low structural state). The most sodic and most calcic plagioclases coexist with the most forsteritic and most fayalitic olivines, respectively, in rocks texturally similar in which olivine is a cumulus phase and plagioclase an intercumulus phase (terminology from Wager, Brown, and Wadsworth, 1960). The more sodic plagioclase may have crystallized under conditions that permitted diffusion between the interstitial (plagioclase) melt and overlying magma, resulting in a more differentiated plagioclase. The highly calcic plagioclase may be the product of crystallization of a melt completely entrapped by the olivine grains. Unlike the other plagioclases of this group of rocks, this plagioclase is conspicuously zoned, suggestive of a lack of diffusion. *Biotite* occurs in trace amounts. Black *iron "ore"* is a particularly common accessory mineral in the olivine-rich rocks. With the exception of local intense serpentinization, alteration of this group of rocks is minor.

Fluxion and cumulate textures are common in rocks of the inner zone. Examples of the fluxion texture are aligned plagioclase laths and, more rarely, elongate olivine grains in olivine gabbros. Cumulate texture is presented by dunitic rock consisting of cumulus olivine and very minor intercumulus plagioclase in the central part of the inner zone. Heteradcumulate texture identical to that of the outer zone rocks is of much more frequent occurrence in the inner zone.

The nature of the layering of different rock units in the inner zone deserves mention. Many wavy and irregular contacts between layers were observed; the irregularities are discernible also on a microscopic scale. Non-quiescent conditions of magmatic intrusion and crystallization are indicated. Perhaps there was more than one injection of magma and, indeed, one aspect of the internal structure of the intrusion tends to support this: some peridotite bodies appear to cut across other rocks in dyke-like fashion. These crosscutting relationships may have arisen through later deformation but the small-scale folding of igneous contacts probably took place during intrusion. The minerals on both sides of the contact are fresh, interfinger across the contact, and show no sign of cataclasis.

Vertical igneous layering is widespread in the inner zone and is difficult to explain by invoking any process other than deformation after solidification of a horizontally layered intrusion. Assuming that the original intrusion was floored by ultramafic rocks (now the inner zone) and overlain by the metamorphic rocks, one may postulate that the whole region has been tipped eastward.

*Minor Intrusions*

Dykes occur throughout the intrusion. Most of the mafic ones are diabases (dolerites) but one dyke observed is porphyritic with an intersertal groundmass. The phenocrysts are chiefly plagioclase ($An_{64-42}$), some of which contain clinopyroxene grains in their cores. Microphenocrysts of clinopyroxene and a glomerophenocryst of plagioclase and clinopyroxene are present. The groundmass consists of plagioclase microlites ($An_{55-20}$) and a large amount of fine-grained iron oxide and carbonate. The rock might be named a basaltic andesite.

A number of altered syenite dykelets cut gabbros of the inner zone. In thin section, plagioclase and perthite can be recognized but are highly altered. Epidote fills veins in the rock and occurs at contacts with the gabbro.

### *Age*

A K-Ar age determination on biotite from a norite of the outer zone yields a figure of $376 \pm 16$ m.y. or Early Devonian, according to the Geological Society of London time-scale (1964). The mineral concentrate is described by the Geochronology Section of the Geological Survey of Canada as "clean, relatively unaltered pale reddish brown biotite with less than 1% hornblende impurity." This contrasts with some of the clinopyroxene in the same rock, which is heavily uralitized. The uralitization is extremely spotty, typical of the intrusion as a whole, so it seems unlikely that the rock has undergone metamorphism. The biotite is apparently not an alteration product: it occurs in interstices and has not overgrown any minerals. This and its reddish brown colour indicate that the biotite is pyrogenic. Biotite from unmetamorphosed mafic intrusions has, in general, given reliable ages of crystallization (Hamilton, 1965, p. 72–73). The age obtained also agrees with several other recent determinations on metamorphic and plutonic rocks in northern Ellesmere Island, discussed elsewhere in this report. On the other hand, it may be no more than a minimum age of intrusion. This subject will be taken up again after other occurrences of mafic rocks have been described.

## Ultramafic Massifs on M'Clintock Inlet
(Fig. 29C; map-units A, B)

Two large ultramafic bodies are exposed east and west of the southern part of M'Clintock Inlet (Fig. 21). The western body, discovered by H. P. Trettin in 1965, forms a range of mountains that rise to a maximum height of more than 4,000 feet, north of Ooblooyah Creek. Sixteen miles away to the east of the inlet a similar mass was found by the writer in 1967. Because of their large size and uncertain mode of emplacement, the two bodies are herein termed the M'Clintock West and M'Clintock East ultramafic massifs (map-unit A).

### *The M'Clintock West Ultramafic Massif*

*General Description*

The massif is approximately rectangular, 12 miles long and 3 to 4 miles wide. It is largely ice and snow covered and much of its southern side is bordered by a large glacier,

FIGURE 21. Ultramafic and metamorphic rocks in the upper reaches of M'Clintock Inlet; view to the east. The rocks flanking the M'Clintock West serpentinite massif to the south are chlorite schists and marbles of probable pre-Middle Ordovician age (MC) and Pennsylvanian red beds (P). On the far side of the fiord, the metamorphics are in fault contact with Ordovician and (?)Silurian carbonates (see *also* Fig. 11). (Oblique air photograph) RCAF T409 R-224, July 16, 1950.

which feeds Ooblooyah Creek. Only the eastern end of the body, where it is well exposed, was studied in detail by H. P. Trettin in 1965 and by the writer in 1966. Trettin's work was chiefly on the northern side, while the writer concerned himself with the southern side of the massif, along Ooblooyah Creek, and with the eastern margin, which Trettin also reconnoitered. In addition, the writer made a traverse by helicopter along the crest of the mountains. Two sampling stops were made on the helicopter traverse. At both localities, greatly sheared serpentinite, weathering grey-green with a rusty stain, was found. At one locality, a clear foliation, imparted by parallel shear planes, is present. Interlayered with the serpentinite is a much-altered gabbro, which, both in hand specimen and in thin section, shows evidence of shearing. Pyroxene and plagioclase crystals are bent and broken and undulatory extinction is common.

Despite heavy alteration, the original mineralogy of the gabbro is microscopically still discernible. *Clinopyroxene* is in 1- to 3-mm subhedral grains, more or less altered to pale green penninite and finely granular sphene in some rocks, in others to uralite along its edges. *Plagioclase* is invariably saussuritized to such an extent that optical determination of its anorthite content is not feasible. Grain size averages from 2 to 3 mm. Two specimens carry interstitial micropegmatite and one of these a little strained *quartz* in aggregates associated with chlorite, which is probably an alteration mineral. The remainder typically consists of serpentine and talc, penninite, sphene, and black iron opaque grains. The texture of all specimens is hypidiomorphic-granular.

Nearer the northern and eastern margins of the massif, the rocks are fresher and fairly good outcrops exist east of the northwest-flowing glacier (Fig. 21). They occur in stream-cuts on the northern slopes of the mountains and, farther north, in the low area bordering the glacier. On the mountain slopes, the rock is a black serpentinite but in the low area a gabbroic-textured serpentinized peridotite, as well as highly sheared serpentinite, is found. Minor granitic intrusions are common and are described in a later section.

The gabbroic-textured peridotite, more specifically wehrlite, outcrops in isolated small bodies, between which no other bedrock exposures were observed. It is assumed that the whole of the low area is underlain by this rock (and by minor granitic intrusions). Partly to completely serpentinized *olivine* (45.8 vol. per cent) forms a groundmass enclosing clinopyroxene grains. Finely disseminated magnetite in cracks in the serpentinite brings out the mesh structure. According to X-ray measurements, the olivine is $Fo_{79-82}$. The *clinopyroxene* crystals (52.4%) stand out well in hand specimen and may be 5 mm across but average 1 to 2. It is neutral in colour, has a $2V_{\gamma}=60°$ and $n_{\beta}=1.704$, suggesting an iron-rich salite (Deer, Howie, and Zussman, 1963a, p. 1, 132). Inclusions of serpentinized olivine are common, indicating early serpentinization. Phlogopitic *biotite* (1.0%) is a ubiquitous accessory. Primary *magnetite* (0.8%) is common in all specimens and many of the grains in one rock have cores of green spinel (pleonaste or hercynite). Iron-rich spinel is characteristic of orthopyroxene-free, clinopyroxene-rich peridotites and is probably a reflection of an original silica-poor parent magma (Irvine, 1967, p. 95).

### *Contact Relationships*

According to Trettin (1969b, p. 57) the northern side of the massif is in fault contact with the Ayles and M'Clintock Formations of Ordovician age (map-unit 2). The contrast between the buff weathering carbonate rocks of the Ayles Formation and the dark ultramafic rocks makes the contact easily recognizable both on the ground and on aerial photographs. Reconnaissance from the air suggests faulted contacts at the massif's western end, where the overlying sedimentary rocks dip steeply to the west. The eastern extremity of the massif is unconformably overlain by Pennsylvanian strata (map-unit 3).

The southeastern contact, exposed in the valleys of the easternmost two northern tributaries of Ooblooyah Creek (Fig. 21), was examined by the writer. Talus obscures the actual contact, which is between serpentinite and chlorite-quartz schists (map-unit 1c). Both rock-types are badly altered and sheared. The cataclastic texture of the serpentinite is not readily apparent in the field but is immediately visible on a sawn surface. Thin-section study confirms that the serpentinite is highly cataclastic and shows that very fine grained carbonate is abundant. Fractured brownish grains of relict clinopyroxene are present. No inclusions of schist were observed in the serpentinite. There is no marked change in appearance of the schist as one approaches the contact with the ultramafic rock.

The nature of the contact between schist and serpentinite remains problematical. Lack of thermal metamorphism in the schist does not necessarily support a non-intrusive contact, as "cold" intrusions are common in the orogenic serpentinite suite. Similarly, absence of xenoliths of country rock is expected in a "solid" intrusion. The cataclasis may be the result of the intrusion of a cold, solid body rather than of later deformation. On the other hand, as described below, post-Permo-Carboniferous faulting has been intense at the margins of the M'Clintock East massif. Applied to the similar M'Clintock West massif, this evidence favours a fault contact.

### *Associated Felsic Intrusions (map-unit B)*

Small bodies of unmetamorphosed granitic rocks intrusive into the eastern end of the M'Clintock West massif were first reported by Trettin (1966, p. 10; 1969b, p. 59ff). The writer visited these occurrences in 1966 and discovered another one to the west.

The intrusions appear to be sheets and plugs. On low, weathered slopes, felsic talus provides the only indication of an underlying felsic body. In the best exposures of granitic rock found, in a stream-cut on the northern slope of the massif, the felsic intrusion is about 30 feet thick. It is homogeneous, except for concentrations of epidote at the serpentinite contact. Angular fragments of serpentinite are common in the granitic rock. The ultramafic rock at the contact is not noticeably altered.

Although the rocks of the different intrusions are generally similar, varieties of rock occur, ranging from dark pegmatitic diorite, rich in hornblende prisms 3 inches long, to fine-grained albite-bearing leucogranodiorite. The most abundant rock is a fine- to medium-grained hornblende diorite, with or without quartz. Table VII includes typical modes of this rock (nos. 2 and 3) and of other members of the association. All these rocks are plotted on quartz-Na, K-feldspar-plagioclase and mafic minerals-K-feldspar+quartz ternary diagrams (Figs. 22, 23). It is evident from the latter diagram that a continuous series from mafic to felsic rock exists, but this was not observed in any single intrusion. The hornblende diorite occupies an intermediate position in the series.

The petrography of the felsic rocks is conveniently described under the headings (a) diorite, and (b) quartz diorite-granodiorite-syenodiorite association.

(a) The diorites: More than half of the diorites examined contain *clinopyroxene* but it is usually greatly subordinate to hornblende. It is very pale green and has $2V_\gamma = 56°$ and $n_\beta = 1.721$, corresponding to a Ca-rich ferroaugite (Deer, Howie, and Zussman, 1963a, p. 1, 132). Invariably, it shows at least incipient replacement by green hornblende. *Hornblende* (X = yellow with greenish tinge, Y = olive-green, and Z = grass green) is always the dominant mafic mineral and appears to have formed both by replacement of pyroxene and independently by magmatic crystallization. Hornblende crystals with randomly oriented, sharply bordered inclusions of clinopyroxene represent overgrowths on pre-existing pyroxene. Some of this hornblende is zoned. Regularly oriented, diffusely bordered clinopyroxene, chiefly in cores of hornblende crystals, indicates replacement of pyroxene by hornblende. Independent growth of the two minerals may be a consequence of the partial overlap of their stability fields. In altered varieties of diorite, hornblende is partly replaced by *chlorite*. In contrast to hornblende, which is generally fresh, *plagioclase* is always altered to sericite, epidote and/or carbonate, commonly obscuring the twin lamellae. In some specimens, cores of crystals are riddled with epidote grains whereas rims are altered to sericite or carbonate. The normal zoning indicated is confirmed by optical measurements on suitably fresh grains: core of $An_{43-36}$, rim of $An_{23-20}$. *Microcline*, which is finely perthitic and shows poor grid twinning, most commonly fills a few small interstices, but in one leucodiorite it has formed

TABLE VII

*Modal and chemical analyses of leucocratic rocks from the M'Clintock West massif.*

| | 1 | 2 | 3 | 4 | 5 | 6 |
|---|---|---|---|---|---|---|
| Quartz | 0.8 | tr | 1.5 | 4.1 | 22.8 | 22.7 |
| K-feldspar | – | tr | – | 19.8 | 5.9 | 11.9 |
| (perthite) | | | | | | |
| Plagioclase | 20.8 | 32.1 | 68.1 | 60.1 | 62.2 | 58.8 |
| Clinopyroxene | 31.3 | 1.3 | – | – | – | – |
| Hornblende | 45.9 | 60.9 | 23.0 | 6.8 | 3.1 | – |
| Biotite | – | – | tr | 0.7 | 0.4 | 0.3 |
| Muscovite | – | – | 2.2 | – | – | – |
| Epidote | 0.2 | 5.0 | 2.7 | – | 1.0 | – |
| Chlorite | – | tr | – | – | 1.6 | – |
| Actinolite | – | – | – | 5.0 | – | 3.4 |
| Hematite | tr | – | – | 3.5 | – | 0.8 |
| Calcite | – | – | tr | – | 0.8 | – |
| Light accessories | 0.8 | 0.4 | tr | tr | 0.2 | 1.0 |
| (apatite, zircon) | | | | | | |
| Dark accessories | 0.2 | 0.1 | 2.5 | – | 2.0 | 1.1 |
| (sphene, opaques) | | | | | | |
| Points counted | 1745 | 1468 | 1508 | 1899 | 1655 | 1886 |
| Plagioclase comp. | $An_{43-23}$ | $An_{38-23}$ | $An_{40-22}$ | $An_{27-18}$ | $An_{30-18}$ | $An_{18-9}$ |
| | | | | | | |
| $SiO_2$ | 49.3 | | 51.5 | | | 65.7 |
| $TiO_2$ | 0.49 | | 0.46 | | | 0.22 |
| $Al_2O_3$ | 10.3 | | 17.4 | | | 14.6 |
| Tot. Fe as $Fe_2O_3$ | 7.6 | | 6.8 | | | 1.6 |
| MnO | 0.14 | | 0.35 | | | 0.13 |
| MgO | 12.4 | | 6.5 | | | <0.5 |
| CaO | 14.4 | | 4.9 | | | 1.4 |
| $Na_2O$ | 2.4 | | 5.0 | | | 5.8 |
| $K_2O$ | 0.6 | | 2.1 | | | 2.6 |
| Total | 97.6 | | 95.0 | | | 92.1 |

Analysts: Rapid Methods Group, Geological Survey of Canada

1. Clinopyroxene-hornblende diorite
2,3. Hornblende diorite
4. Syenodiorite
5. Quartz diorite
6. Albite-bearing granodiorite

*GSC*

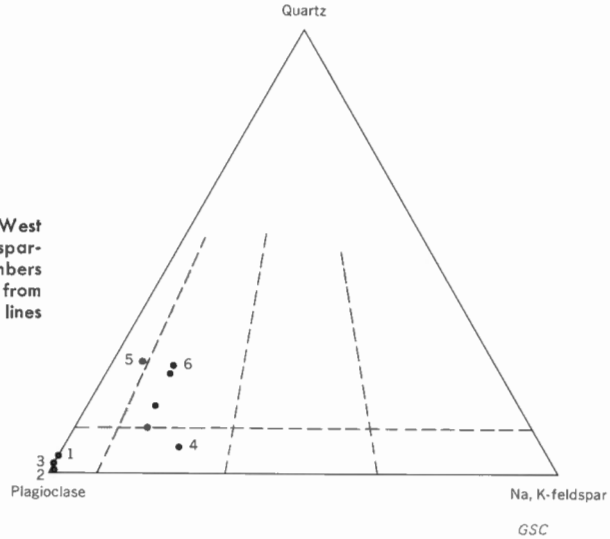



FIGURE 22

Leucocratic rocks of the M'Clintock West massif plotted in the quartz-Na, K-feldspar-plagioclase ternary classification. Numbers are those in Table VII; remainder are from Trettin (1969b, Table VIII). The broken lines denote boundaries as in Figure 5.

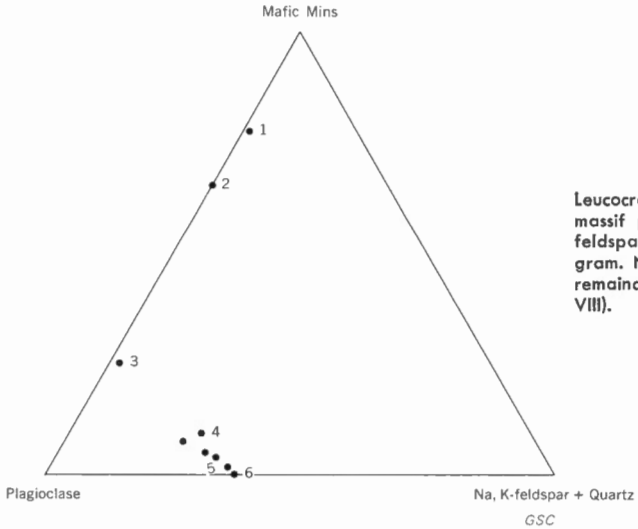



FIGURE 23

Leucocratic rocks of the M'Clintock West massif plotted on a mafic minerals-Na, K-feldspar + quartz-plagioclase ternary diagram. Numbers refer to rocks in Table VII; remainder are from Trettin (1969b, Table VIII).

3- to 6-mm equant tablets, poikilitically enclosing hornblende and plagioclase grains. *Quartz* is less abundant than microcline. Also present are accessory brown biotite, sphene, and apatite. In the more altered rocks, carbonate, together with epidote and muscovite, fills veinlets and cracks.

Texture of the diorites varies from coarse to medium grained hypidiomorphic-granular. Undulatory extinction is common in interstitial microcline and quartz but only one specimen shows cataclasis, whereby plagioclase has been bent or broken and numerous cracks (now filled by carbonate) transect the various minerals. Thus, compared with the highly sheared serpentinites, the diorites are little deformed.

(b) The quartz diorite-granodiorite-syenodiorite association: Rocks of this suite vary in mineralogy but show strong affinities to each other as well as to the diorites, with which they are associated in some of the felsic intrusions. Clinopyroxene is absent and biotite is a varietal mineral in several rocks.

Green *hornblende* is of the same variety as is found in the diorites but a brown hornblende was observed in trace amount in one granodiorite. The brown variety may be a relic of an earlier, higher temperature generation of amphibole. The *biotite* of most rocks is heavily to completely chloritized to penninite, usually associated with epidote. When fresh, it is straw yellow to brown, which may be indicative of a high $TiO_2$ content and a low $Fe_2O_3$/($Fe_2O_3$ + FeO) ratio (Deer, Howie, and Zussman, 1962b, p. 71). *Plagioclase* is always zoned and ranges in composition, as far as could be determined, from $An_{30}$ (cores of grains in quartz diorite) to $An_9$ (rims of grains in leucogranodiorite). Commonly superimposed on this normal zoning is a well-developed oscillatory one. As in the diorites, the plagioclase has been selectively sericitized so that cores are always more altered than rims, which may even be perfectly fresh. *Microcline* is considerably more abundant than in the typical diorite. Clear cases of microcline replacing plagioclase were noticed in many thin sections and probably represent an advanced stage of K-feldspar growth, as most of the microcline has grown interstitially to include, but not apparently to replace, plagioclase. Stringlet- and string-type perthite (Deer, Howie, and Zussman, 1963b, p. 68) texture in the microcline is ubiquitous but irregularly developed, both within any single grain and from rock type to rock type. *Quartz* is always strained and many grains are cracked. *Hematite* commonly fills these cracks, and growing out from them is fibrous *actinolite* in a dendritic or radial pattern. Both minerals probably were formed by deuteric processes.

Three of the eight specimens of this suite studied are cataclastic in thin section. Plagioclase laths are bent, broken and microfaulted, and quartz, as mentioned above, is cut by numerous fractures. The remainder of the suite possesses a hypidiomorphic-granular texture. Deformation was certainly not as intense as in the serpentinite.

Several lines of evidence indicate that the entire leucocratic suite is of magmatic origin. Textures of the diorites show particularly well the following, typically magmatic crystallization sequence, beginning with clinopyroxene and ending with quartz:

clinopyroxene ————
hornblende ————
plagioclase ————
potash feldspar ————
quartz ————

Zoning is conspicuous in plagioclase and in some of the larger hornblende crystals throughout the felsic suite.

The magmatic differentiation trend suggested by Figure 23 is reflected in the chemistry of the rocks. Chemical compositions of clinopyroxene-hornblende diorite, hornblende diorite, and albite-bearing leucogranodiorite are listed in Table VII (nos. 1, 3, and 6) and plotted on a silica variation (Harker) diagram in Figure 24. As silica increases, lime, iron, and magnesia contents fall, and alkali content rises. Alumina variation is more erratic, the curve rising at first, then falling. MnO remains rather constant. All these features are characteristic of normal igneous differentiation.

Effects of volatile components are obvious in the felsic rocks. The pegmatitic hornblende diorites probably were formed from magma rich in water vapour, which lowered the temperature of crystallization and the viscosity, permitting minerals to grow to a large

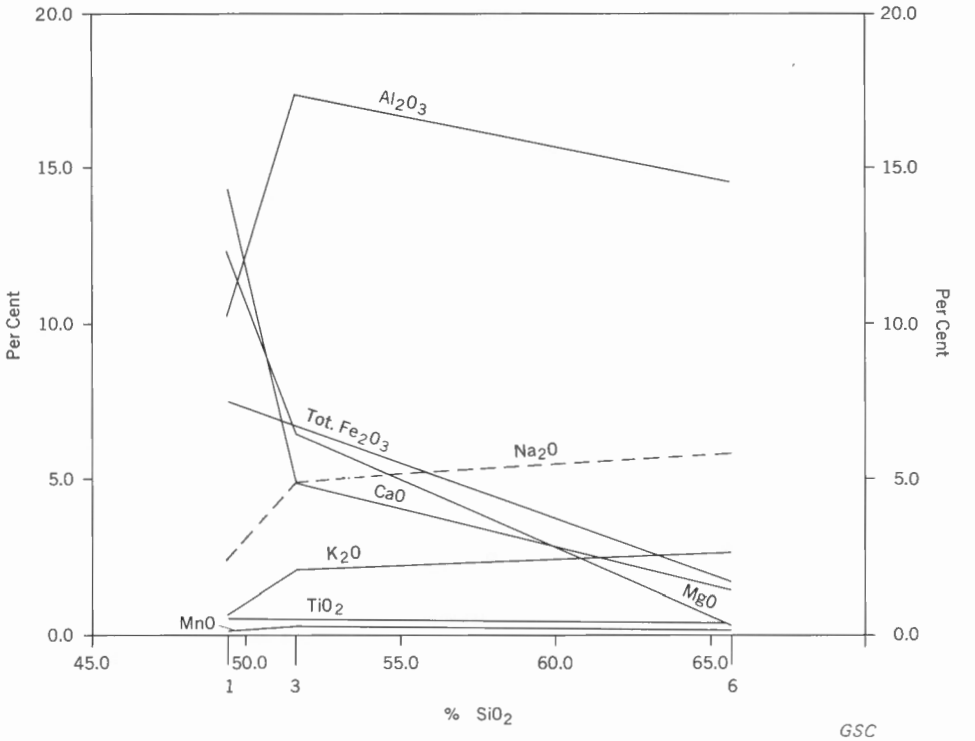



FIGURE 24. Silica variation (Harker) diagram for the diorite-leucogranodiorite association of the M'Clintock West massif. Numbers are those in Table VII; 1 – clinopyroxene-hornblende diorite, 3 – hornblende diorite, 6 – albite-bearing granodiorite.

size. The ubiquitous partial to complete alteration of pyroxene to hornblende may be evidence of increasing $P_{H_2O}$ and/or decreasing temperature as crystallization proceeded. $CO_2$ also must have been significant in volume because carbonate veinlets and patches and carbonate alteration of plagioclase are common.

## *The M'Clintock East Ultramafic Massif*

Only cursory examination of the western end of this body has been made and not even its outlines are fully known. Aerial reconnaissance indicates that the massif trends northeast and is probably 8 miles long and 3 miles wide. Like the M'Clintock West body, it forms a ridge of mountains, the higher parts of which are extensively snow-covered (Fig. 21).

All ultramafic rocks examined are dark green to black peridotite, which is serpentinized to greater or lesser extent. Two specimens were studied in thin section. One is a serpentinized wehrlite, analogous to the wehrlite of the M'Clintock West massif, though richer in serpentine (after olivine). The other is a serpentinized olivine clinopyroxenite, the clinopyroxene of which is similar to that of the wehrlite ($2V_\gamma = 60°$). About 5 per cent of the rock is serpentinized olivine. Both rocks contain abundant magnetite and a little phlogopite.

The talus includes boulders of pink and grey syenitic rock. One variety consists mainly of decussate plagioclase laths (2 mm), which have been bent and are partly altered to sericite and epidote. The *plagioclase* is zoned and ranges in composition from $An_{30}$ to $An_{25}$. Patches

of *potassium feldspar* (poorly twinned microcline?) *perthite* enclose and partly replace much of the plagioclase. At contacts between the two feldspars untwinned albite has formed. Small clusters of ragged and kinked *biotite* (X = pale greenish brown, Z = dark brown) and granular *epidote* are common.

Another rock studied appears to be an altered syenite. Most of the feldspar (perthite as well as plagioclase) has been considerably altered to sericite and fine-grained carbonate. Coarser grained muscovite (0.75 mm) and carbonate are irregularly distributed throughout the rock. Apatite, in grains as long as a millimetre, is a common accessory. Alteration of a syenite, similar to the one described above, by late volatile-rich solutions seems likely.

No bedrock exposures of leucocratic rock were found in a brief search. Again it is emphasized that no granitic rock was seen as either boulders or bedrock outside the limits of the ultramafic body; this strengthens the hypothesis that the two rock types are genetically related.

The extreme eastern end of the massif is abutted by a glacier. Its northern contact, with Permo-Carboniferous sedimentary rocks (map-unit 3), is well exposed in a deep stream valley at the eastern end of the body. Post-Permo-Carboniferous faulting in places has carried ultramafic rock over sediments. The peridotite is highly sheared and the Permo-Carboniferous sandstones (including Pennsylvanian red beds) and carbonate rocks are strongly contorted. Seen from a distance, the sediments elsewhere appear to overlie the peridotite.

The strike of the sedimentary strata parallels the long axis of the massif. No contact metamorphic effects were observed.

### *Petrogenesis*

According to Turner and Verhoogen (1960, p. 307ff), ultramafic bodies of orogenic belts have the following characteristics:

(1) Occurrence in folded geosynclinal rocks.
(2) Principal rock types are peridotite and serpentinite; common minor associates are dioritic to granitic rocks.
(3) Broad concordance between direction of elongation of the ultramafic bodies and the trend of foliation or bedding of the enclosing rocks.
(4) Location in fault zones.

Thayer (1963, p. C85) maintains that gabbro is also an important associate of orogenic ultramafic complexes and that peridotite, gabbro, and dioritic rocks constitute an "alpine mafic magma stem."

As all these features are exhibited by the M'Clintock Inlet massifs, they are considered to be of alpine type. However, so far as is known, orthopyroxene is absent from these bodies. According to some workers (e.g., Irvine, 1967, p. 92), this would place them in a group distinct from the typical alpine-type complexes. But it should be emphasized that further work may reveal the existence of orthopyroxene-bearing rocks in the M'Clintock bodies.

Major examples of orthopyroxene-free ultramafic rocks are the southeastern Alaska complexes (Taylor and Noble, 1960). Although not comparable in detail to the M'Clintock Inlet bodies, the Alaskan complexes also have an olivine-clinopyroxene rock as a major unit. The olivine has the composition $Fo_{85-95}$ and the clinopyroxene is diopsidic augite (Taylor and Noble, 1960, p. 176), similar to the minerals of the M'Clintock West massif. Presnall (1966) has shown experimentally that such a rock can be produced by fractional crystallization of an ultramafic magma consisting of 71 weight per cent diopside, 16 per

cent forsterite, and 13 per cent total Fe as FeO, at constant pressure under conditions that allow free exchange of oxygen between the magma and its surroundings. His data indicate that such a peridotite magma can exist at temperatures down to 1250°C (Presnall, 1966, p. 788). If such an origin is applicable to the M'Clintock bodies, the lack of contact metamorphism around them must mean that the bodies have been displaced from their original wall-rocks.

Also noteworthy are the leucocratic intrusions into the ultramafic bodies. Circumstantial but strong evidence suggests that they are genetically related to the ultramafic rocks:

(1) The felsic intrusions spatially are related closely to the ultramafic bodies. Only one intrusion of broadly similar type, an altered, albitized diorite dyke, has been found in the area outside the M'Clintock Inlet massifs (Trettin, 1969b, p. 59).
(2) The volume of felsic rock is insignificant in comparison to that of the ultramafic rock. Such a volume relationship would favour the derivation of granitic from the ultramafic rock.
(3) Diorite and albite-bearing rocks are common associates of alpine-type ultramafic rocks (Turner and Verhoogen, 1960, p. 311; Thayer, 1963, p. C85).

It is difficult to ascertain the length of time that elapsed between solidification of the ultramafic rock and the crystallization of the felsic rocks. The angular fragments of the former in the latter mean that the ultramafic rock must have been solid when the felsite intruded. Chilled zones of felsic against ultramafic rock were not observed but they may be obscured by the extensive epidotization and other alteration at the contact. The scarcity of marked cataclasis in the felsic rocks favours the hypothesis that the felsic rocks post-date the emplacement of the ultramafic rocks, which are sheared and foliated.

The field evidence, textures, mineralogy, and chemistry of the felsic rocks indicate that they are middle- to late-stage magmatic differentiates. Their immediate parent may be represented by the gabbroic rocks found on the helicopter traverse. However, a major obstacle to derivation of the quartz-bearing members of the leucocratic suite from the ultramafic rocks is that orthopyroxene-free, clinopyroxene-rich peridotites must have crystallized from a silica-poor magma. It seems likely that there was addition of, at least, silica and water from an outside source.

Little has been written on the origin of felsic rocks in orogenic ultramafic complexes. Olsen (1961) concluded that felsic rocks (certain types of which resemble those described here) associated with serpentinites in Quebec, are high-temperature, metasomatically altered gabbro bodies. For the M'Clintock rocks, however, relatively low temperatures are indicated. Thayer (1963) has described briefly quartz diorite and albite granite associated with a peridotite-gabbro complex in Oregon. He states that the granites are metasomatically altered (albitized) quartz diorites but gives no supporting evidence. The M'Clintock albite-bearing rocks do not appear to have been produced by albitization: albite occurs as rims on oligoclase, indicating normal magmatic differentiation (this is supported by the chemical variation trends of Figure 24). A recently described ophiolite complex in Greece comprises 75 to 85 per cent ultramafic rocks and 15 to 25 per cent gabbro, diorite, quartz diorite, and partly pillowed lava, which appear to represent a differentiation sequence (Moores, 1969).

### *Age*

As Pennsylvanian rocks unconformably overlie the M'Clintock West massif, it is known to be pre-Pennsylvanian. Furthermore, hornblende from a syenodiorite intrusive into the massif yields a K-Ar age of 390 ± 20 m.y., or Early Devonian, according to the Geological

Society of London time-scale (1964). This date sets an upper limit to the age of the ultramafic rock and, if the syenodiorite and the ultramafics are contemporaneous, may even give an approximate time of emplacement of the latter.

On the basis of its proximity and similarity to the western body, the M'Clintock East massif is considered to be of the same age. Together, the two bodies form a major zone of ultramafic rock emplacement (*see* Trettin, 1969b, p. 71).

## Other Ultramafic and Mafic Bodies

### *Cape Columbia*

(Fig. 29A; map-unit A)

In 1965, the writer discovered a mass of partly metamorphosed ultramafic rock a mile west of Cape Columbia. The outcrop is ridge-like in form, about 500 feet long and 30 feet high. Extensive yellow-green serpentinization makes the mass stand out from the surrounding darker scree of metamorphic rock. The country rock (chiefly hornblende gneiss and amphibolite) of the immediate area is in the amphibolite facies of regional metamorphism but was not found in contact with the ultramafic body. There is little doubt, however, that the scree is essentially in situ. The long axis of the ultramafic body lies approximately east-west, which parallels the overall trend of foliation in the country rock.

Microscopic study reveals that part of the mass has been completely metamorphosed. Tight folding of thin epidote-rich layers is visible on the sawed surface of one specimen. Other samples are extensively serpentinized but only incipiently metamorphosed.

The weakly metamorphosed rock is serpentinized peridotite, consisting chiefly of clinopyroxene and serpentinized olivine. The clinopyroxene is in rounded grains (1–3 mm) with ragged edges and heavily dusted with black iron oxide material. Very fine magnetite has exsolved from the clinopyroxene. Some grains have been extensively replaced by serpentine, which has grown inwards from cracks in the pyroxene crystal. A few shreds of orthopyroxene (optically positive, $2V_\alpha = 90°$, suggesting enstatite) have apparently survived serpentinization. The groundmass is made up of serpentine and relict *olivine* in mesh texture, with accessory very pale green, Mg-rich hornblende (described below), magnetite, rutile, epidote, and sphene. The serpentine clearly is derived largely from olivine but some may represent serpentinized orthopyroxene.

The metamorphosed ultramafic rock comprises magnesian hornblende and epidote. The *amphibole* occurs in prisms (1–2 mm), generally parallel to layering in the rock. Optical properties are as follows: pleochroic formula X = pale greenish yellow, Y = pale green with brownish tinge; $Z \wedge c = 18°$; optically positive; $2V_\alpha = 60°$. These data suggest a magnesian hornblende, probably pargasite. The other major mineral is *epidote*, most of which is concentrated in bands, a few millimetres wide, between hornblende-rich layers. Grains are 0.1 to 0.2 mm in size and in places are included in hornblende. Their interference colour is generally an anomalous blue-yellow and their optic sign is negative.

Epidote is certainly not a stable, primary mineral of ultramafic rocks. The fact that the amphibole is invariably associated with epidote suggests that it, too, was not an original mineral of the peridotite and that both minerals are related. They probably formed by metamorphism and metasomatism of olivine and pyroxene. The fact that the ultramafic rock has been metamorphosed, rather than merely serpentinized, suggests that it was em-

placed before metamorphism of the Cape Columbia Complex, and was subsequently metamorphosed along with the surrounding rocks. No other occurrences of metamorphosed ultramafic rocks are known in Ellesmere Island.

### *Bromley Island and Kleybolte Peninsula*

Exposures of mafic and ultramafic rocks in northern Ellesmere Island not studied by the writer include altered basic to ultrabasic intrusions on Bromley Island, dated, on the basis of stratigraphic evidence, as Middle Ordovician or older (Trettin, 1969b, p. 58) and a serpentinized dunite dyke on Kleybolte Peninsula which is Tertiary or older (Trettin, 1969a, p. 45–46).

## Felsic Rocks

In this section the major granitic and syenitic intrusions known in northern Ellesmere Island are described and their place in the tectonic history of the geosyncline is discussed.

All the major felsic plutons post-date the regional metamorphism that took place in pre-Middle Ordovician time and some are known to be younger than Early Silurian. Potassium-argon dating suggests apparent ages of crystallization ranging from Early Devonian to Late Mississippian.

### Cape Richards Intrusive Complex
(Fig. 29D; map-unit D)

In the course of his coastal reconnaissance, Christie (1957, p. 24) noted "granite" near Cape Richards. Further work by the present author showed that the granitic rock near the cape is only part of a larger, zoned mass that also includes syenites and diorites.

Figure 25 is an aerial view of the complex. Exposure of the intrusion is largely confined to a narrow belt of coastal mountains, whose landward back slopes are scree-mantled and partly covered by ice and snowfields. Field work was limited to study of the sea-cliffs, which apparently provide a good section through the intrusion.

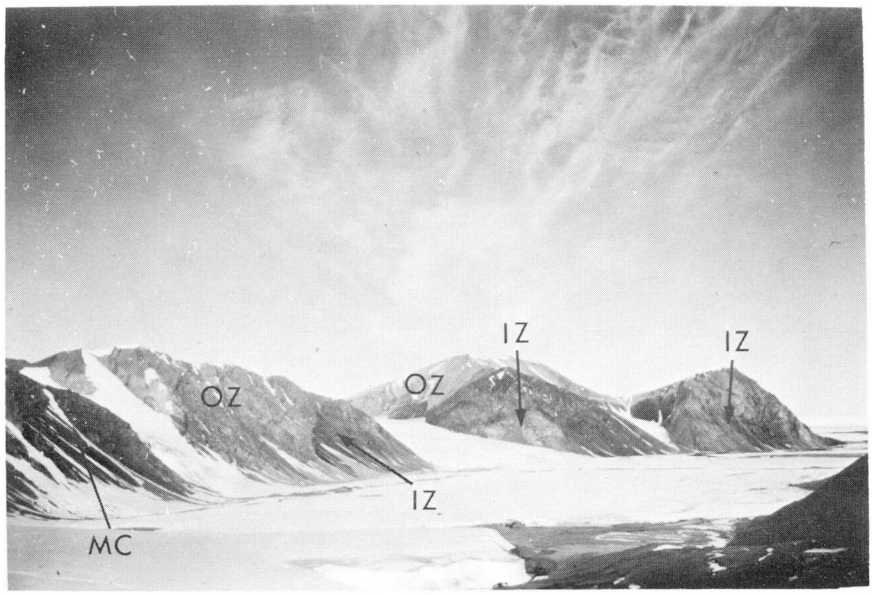
The Cape Richards complex can be divided into an outer zone of quartz monzonitic rock, which forms the northern and southern ends of the complex, and an inner zone of chiefly syenitic rock. It is not known if the granitic rock surrounds the inner zone inland from the coast. To the east, on the other side of the fiord, metasedimentary rocks similar to the country rock (map-unit 1c) around the intrusion are unconformably overlain by Middle Ordovician strata (Trettin, 1969b, p. 7). The complex, therefore, is at least 2 miles, but no more than 3 miles, wide and 6 miles long.

#### *Petrography*

*Outer Zone*

The chief rock in both the northern and the southern parts of the outer zone is a light weathering, medium-grained quartz monzonite, which has a composition near that of granodiorite. The rock is well exposed on precipitous slopes that rise to more than 1,500 feet above sea level at both ends of the intrusive complex.

Though not obvious in hand specimen, a cataclastic texture is easily recognizable in thin section. Abundant epidote coating shear surfaces in the rock may be related to this




202022-M

FIGURE 25. The Cape Richards intrusive complex, viewed to the northwest. Light-coloured quartz monzonite of the "outer zone" (OZ) of the complex, cut by mafic dykes, surrounds darker quartz diorite and syenites of the "inner zone" (IZ). The dark rocks on the extreme left (MC) are greenschist facies metamorphics of pre-Middle Ordovician age. (Field photograph, July 1966.)

texture, perhaps having been deposited by solutions whose entry and mobility was facilitated by cataclasis. Significantly, epidote is only a minor constituent of the non-cataclastic inner zone rocks.

Numerous mafic dykes crisscross the granitic rock. Dykes are less abundant in the inner zone.

A modal analysis that is the average of the modes of two almost identical rocks from both parts of the outer zone is given in Table VIII and is plotted on a quartz-K-feldspar-plagioclase triangle in Figure 26. All the outer zone granitic rocks are more or less altered and original mafic minerals are not everywhere preserved. *Hornblende* is present in both analyzed rocks and has X = greenish yellow, Y = olive, and Z = grass green. Grains are generally ragged and partly altered to epidote, sphene, and/or chlorite. In one rock, incipient replacement of hornblende by brown biotite was observed. *Biotite* accompanies hornblende and is partly to completely chloritized. Its pleochroic formula is X = pale yellowish brown, Y = greenish brown, and Z = dark brown, suggesting intermediate proportions of Ti and $Fe^{3+}$ (Deer, Howie, and Zussman, 1962b, p. 71). *Plagioclase* is commonly intensely sericitized and composition determinations are difficult. In fresher material, oscillatory zoning with a normal trend is generally visible and measured compositions range from calcic oligoclase to calcic albite. Plagioclase is the only mineral that is consistently euhedral to subhedral in these quartz monzonites. *Microcline perthite* is clearly later than plagioclase, large anhedral crystals being riddled with sericitized plagioclase inclusions. The perthite commonly shows albite stringlets predominating at crystal margins and extending inward from them. This indicates that the albite is of replacement, rather than exsolution, origin. Undulatory extinc-

tion is marked in most of the perthite. *Quartz* also shows definite signs of deformation, tending to occur as clusters of minute grains with sutured borders and highly undulatory extinction. Epidote, sphene, apatite, and zircon are accessory minerals. Specular hematite was seen in several hand specimens but is never abundant.

The common texture of the outer zone rocks is hypidiomorphic-granular modified by cataclastic deformation. As strain features are confined almost solely to potassium feldspar and quartz, which are late crystallizing minerals, deformation must have been protoclastic, as might be expected in an intrusion that was forcefully emplaced.

The outer zone rock at the contact with the metasedimentary country rocks at the southern end of the complex is porphyritic, with phenocrysts of plagioclase (0.5–1 mm long), hornblende and chloritized biotite in an extremely fine grained quartzo-feldspathic groundmass. The fine-grained groundmass apparently is not due to cataclasis and the rock may be a chilled border phase. This porphyritic rock grades into the normal quartz monzonite described above.

### *Inner Zone*

In the northern part of the complex, the inner zone is separated from the outer zone by a septum of metasedimentary country rock. But to the south, the two zones are in contact and can be distinguished by contrasting weathering: outer zone rocks weather light grey, inner zone rocks are brown. The nature of the contact could not be established due to thick talus cover and a mafic dyke that appears to run along the contact. Figure 25 shows how the outer zone rock tends to wrap around and even overlie the inner zone rock. Thus it appears that the rock of the inner zone has intruded that of the outer zone.

The dominant rock of the inner zone is a quartz-bearing syenite with a variable content of mafic minerals. At the southern margin of the zone, quartz diorite is exposed. At the northern limit of the zone, syenite is in contact with metasedimentary rocks.

The mode of the quartz diorite is given in Table VIII (no. 2) and plotted in Figure 26. Relict, very pale green *clinopyroxene* forms the cores of many hornblende crystals. The *hornblende* has X = pale brownish green, Y = brownish green, and Z = darker green with brownish tinge, and is partly altered to epidote, sphene, and/or penninite. The *plagioclase* is in euhedral to subhedral laths, so sericitized that optical determinations are not feasible. A little interstitial, altered *perthite* is present and, where in contact with plagioclase, shows a thin rim of clear albite, which may be the product of a decalcification reaction between the potassium and plagioclase feldspars. Unstrained *quartz* fills interstices.

Modal data on the main varieties of syenite in the intrusion may be found in Table VIII and Figure 26. The inner zone is symmetrically zoned with respect to mafic mineral (pyroxene and amphibole) content: the rocks become more mafic towards the centre of the intrusion, both from the north and from the south.

Unaltered *clinopyroxene* was found only in the central part of the inner zone. It is zoned, having a pale pink core ($2V_\gamma = 48°$) and a pale green rim ($2V_\gamma = 57°$, $n_\beta = 1.711$), suggesting a titaniferous augite to ferrosalite crystallization sequence (Deer, Howie, and Zussman, 1963a, p. 1, 132). In most inner zone rocks, clinopyroxene is restricted to cores of amphibole grains. It appears to be of the same variety as that bordering the titaniferous augite, on the basis of its pale green colour and birefringence. Where separate from hornblende, clinopyroxene tends to be poikilitically enclosed by perthite. The *amphibole* shows little variation in optical properties and is of a more intense green than the outer zone variety. Its pleochroic formula is X = brownish green, Y = dark brownish green, and Z = dark grass green or bluish green, and $2V_\alpha$ and $n_\gamma$, respectively, range from 54° and 1.688 to 48° and

TABLE VIII | *Modal and chemical analyses of rocks from the Cape Richards complex.*

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
|---|---|---|---|---|---|---|---|
| Quartz | 18.0 | 9.1 | 8.0 | 2.5 | 0.2 | 18.2 | |
| K-feldspar (perthite) | 24.2 | 5.2 | 84.3 | 93.3 | 85.7 | 71.6# | |
| Plagioclase | 45.1 | 64.7 | – | tr | 2.4 | | |
| Clinopyroxene | | 1.5 | – | 0.7 | 7.9 | – | |
| Hornblende | 4.1 | 11.5 | – | 1.7 | 0.2 | 10.2+ | |
| Biotite | 2.5 | – | – | 0.7 | – | – | |
| Epidote | tr | 1.3 | – | – | – | – | |
| Chlorite | 4.1* | 5.7 | 4.7 | – | – | – | |
| Hematite and limonite | – | – | 3.3 | 0.8 | 2.9 | – | |
| Light accessories | 1.5 | 0.2 | – | tr | – | tr | |
| (apatite, zircon) | | | | | | | |
| Dark accessories | 0.5 | 0.8 | – | 0.2 | 0.7 | – | |
| (sphene, opaques) | | | | | | | |
| *after biotite<br>#includes 4.1% phenocrystic K-feldspar perthite and 67.5% micropegmatitic K-feldspar + plagioclase<br>+ riebeckite | | | | | | | |
| Points counted | 1804 | 1566 | 1504 | 1510 | 1511 | 1505 | |
| Plagioclase comp. | $An_{25-10}$ | n.d. | – | n.d. | $An_{10}$ | n.d. | |
| | | | | | | | |
| $SiO_2$ | 66.1 | | 61.4 | 60.8 | 61.7 | 76.3 | 70.47 |
| $TiO_2$ | 0.31 | | 0.46 | 0.38 | 0.39 | 0.19 | 0.30 |
| $Al_2O_3$ | 15.6 | | 16.9 | 15.2 | 16.6 | 11.4 | 15.50 |
| Tot. Fe as $Fe_2O_3$ | 2.0 | | 4.0 | 3.9 | 6.4 | 4.0 | 2.75 |
| MnO | 0.04 | | 0.08 | 0.14 | 0.23 | 0.05 | 0.03 |
| MgO | 0.6 | | 0.5 | < 0.5 | < 0.5 | < 0.5 | 0.65 |
| CaO | 2.5 | | 1.6 | 1.2 | 1.4 | 0.2 | 1.91 |
| $Na_2O$ | 4.6 | | 5.4 | 5.6 | 5.3 | 4.4 | 4.12 |
| $K_2O$ | 3.4 | | 5.7 | 5.4 | 5.8 | 4.2 | 3.59 |
| Total | 95.2 | | 96.0 | 92.6 | 97.8 | 100.7 | |

Analysts: Rapid Methods Group, Geological Survey of Canada.

1. Quartz monzonite of outer zone (mode listed is average of two).
2. Quartz diorite of inner zone.
3. Quartz syenite from outer part of inner zone.
4. Syenite from inner zone.
5. Clinopyroxene syenite from inner zone.
6. Riebeckite granite dyke.
7. Average muscovite-biotite granodiorite (Nockolds, 1954, p. 1014).

*GSC*

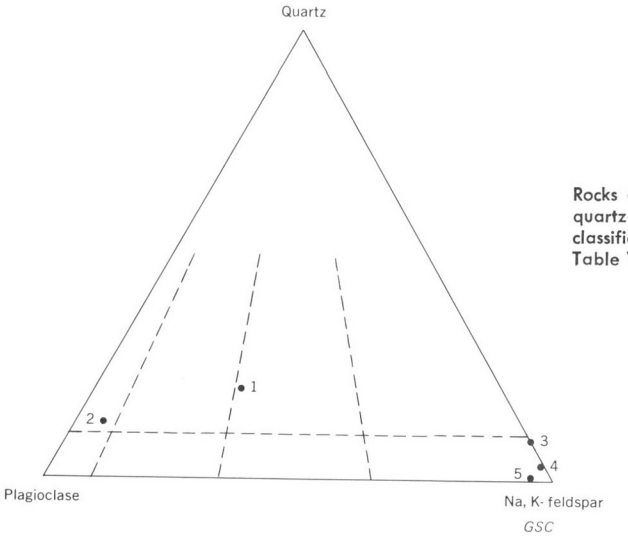
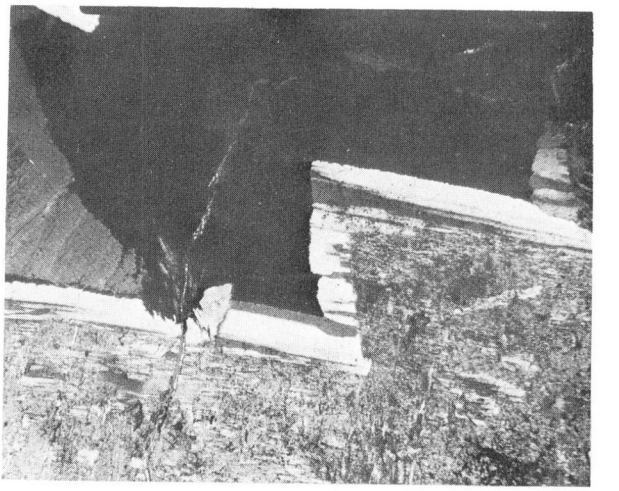



FIGURE 26

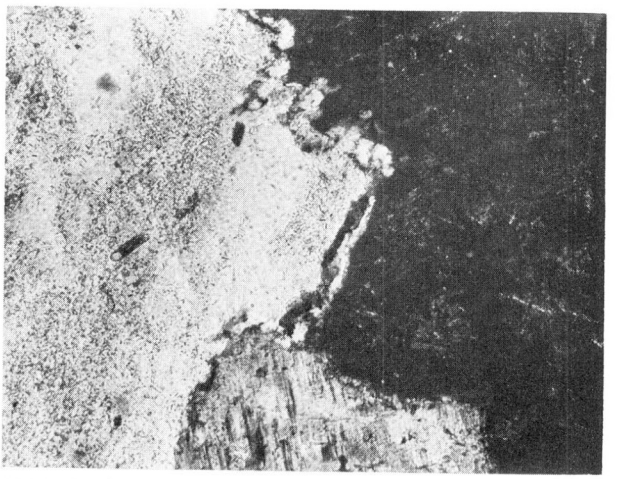
Rocks of the Cape Richards complex in the quartz-Na, K-feldspar-plagioclase ternary classification. Numbers refer to rocks in Table VIII.



200426-R

FIGURE 27A

Late albite in syenite of the inner zone of the Cape Richards complex. Albite rimming perthite. Crossed polarizers x40. (Photomicrograph.)



200426-Q

FIGURE 27B

Late albite in syenite of the inner zone of the Cape Richards complex. Intergrowth of albite rims of two perthite grains, one of which is in extinction. Note the optical continuity of the albite rims with their respective perthite hosts. The albite was probably formed during late, minor recrystallization. Crossed polarizers x138. (Photomicrograph.)

1.701, suggesting an iron-rich variety of hornblende (Deer, Howie, and Zussman, 1963a, p. 296, 298). Hornblende encloses quartz and fills interstices, providing evidence that at least some of the hornblende crystallized late. *Biotite* (X = pale yellow, Z = very deep brown) is always associated with the other mafic minerals. The major mineral in every rock is *perthite*, usually of a type in which blocky, well-twinned albite ($An_5$) is enclosed by microcline. Intergrowths of the two feldspars are more irregular in some specimens and alteration further obscures relationships. A common feature is a thin, discontinuous rim of albitic ($An_{10}$) plagioclase fringing perthite grains (Fig. 27A). Also, the contacts between many perthite grains are marked by a zone of albite, which is an intergrowth of the albite rims of adjoining grains (Fig. 27B). Thus, there is evidence both of late crystallization of albite and of migration of albite during minor recrystallization. *Quartz* was seen in every rock examined microscopically but is unquestionably more abundant in the marginal than in the central parts of the inner zone. It invariably crystallized late and fills interstices, but many grains show strain, which may be related to the late recrystallization that caused the interpenetrating contacts between albite-rimmed perthite grains. Common accessory minerals are apatite, zircon, pyrite, and other black iron "ore."

Alteration of feldspar and the mafic minerals is a ubiquitous feature of the inner zone. Hornblende and biotite in some rocks are replaced by hematite or limonite, while in other rocks, chloritization is incipient to complete.

The texture of the inner zone rocks is typical of quartz-bearing alkali-feldspar syenites. The perthite grains (2–3 mm) are irregular in detail but overall give an impression of being subhedral tabular. Perhaps they were euhedral up to a late stage in their crystallization, when minor recrystallization took place at crystal interfaces. In places this was prevented by interstitial silica-rich melt, and euhedral perthite crystal faces bordered by quartz resulted. The albite rims around perthite probably formed late enough to avoid distortion by recrystallization.

### *Dyke Rocks of the Inner Zone*

The dyke rocks observed in the inner zone are of two groups: (1) varieties of the main intrusive rocks; and (2) riebeckite granite.

The first group is the largest and its members are finer grained equivalents of the more felsic varieties of inner zone rock. Porphyritic varieties in this group carry phenocrysts of either perthite or plagioclase ($An_{25-30}$) in a matrix of perthite and interstitial quartz. One aphyric specimen studied is simply a fine-grained (1–2 mm) hornblende-quartz syenite and contains an inclusion of the coarser grained, but otherwise identical, syenite it has intruded.

At the northern margin of the inner zone, riebeckite-bearing dykes cut the leucocratic quartz syenites. These dykes are unlike the rocks they intrude: they are medium to dark grey, porphyritic and fine grained (phenocrysts rarely exceed 2 mm in size and the matrix averages 0.2 mm). Two varieties were examined under the microscope; both contain perthite phenocrysts in a matrix of perthite and quartz. The strongly pleochroic amphibole present is riebeckite, on the basis of its parallel extinction, large 2V ($>60°$), and pleochroic formula of X = deep blue, Y = yellowish with green-brown tinge, and Z = deep blue (almost black). In one rock, riebeckite forms a mat of flow-aligned needles, almost completely obscuring the groundmass feldspar and quartz.

Dyke widths observed range from a few inches to 5 feet, and crosscutting relationships show that they were not all intruded at the same time. Yet the mineralogy and occurrence of the dykes indicate that they represent highly fluid, late-stage differentiates of the inner zone intrusion. Unlike the aphyric inner syenites, the dykes, because of their feldsparphyric texture, were probably below their liquidus temperature prior to emplacement.

### *Petrogenesis*

The alkalic nature of the inner zone is evident from petrography alone. This contrasts with the quartz-rich outer zone. Chemical analyses of representative rocks are given in Table VIII and confirm that the inner zone is alkalic. The composition of the outer zone quartz monzonite resembles Nockolds' (1954) average muscovite-biotite granodiorite (Table VIII, no. 7). This might be expected from the relative paucity of potassium feldspar in the quartz monzonite (*see* Fig. 26). The rock is richer in silica but poorer in iron and alkalies than the inner zone syenites. Petrographically, too, the inner and outer zones are distinct and their genetic relationship, if any, remains uncertain. Yet a relationship does seem likely, if only because of the close spatial association of the rocks. Similarly, data available do not permit conclusions on the possible connection between the quartz diorite and syenite of the inner zone. However, there are obvious similarities in petrography between the two, notably in the type of clinopyroxene and the form of occurrence of quartz; the possibility of a common magmatic parentage seems good.

The highly perthitic syenites of the inner zone, because they lack plagioclase except as a component of perthite, may be termed hypersolvus (Tuttle and Bowen, 1958, p. 129). (This terminology can only be applied to rocks whose compositions fall in or near "petrogeny's residua system," but it seems safe to assume that the rocks under consideration belong to this group.) Experimental studies have shown that such rocks must have crystallized at high temperatures (above 660°C), either very rapidly or in a dry environment (Tuttle and Bowen, 1958, p. 128). The fact that hornblende and biotite occur in fair abundance in many of the syenites argues against a dry environment. The minor late albite rimming perthite grains may indicate some lowering of temperature to promote unmixing (which would lead to a subsolvus rock).

The analyzed riebeckite granite dyke (Table VIII, no. 6), by virtue of its occurrence, mineralogy, and chemistry, represents an end-stage in the evolution of the inner zone. With continued cooling, $P_{H_2O}$ and $P_{O_2}$ rose and volatile-rich melts were injected into the syenites to form dykes. From Ernst's (1962, p. 734) experiments, it is known that pure hydroxyl riebeckite is stable at relatively low temperatures and high $P_{O_2}$ in rocks with high soda and iron and low lime, magnesia, and alumina. Such a composition is exemplified by the riebeckite-bearing dyke.

### *The Country Rocks and Contact Zone*

The contact of the intrusive complex with the country rock was examined (1) in its northern part, where a septum of country rock separates the granitic rock near Cape Richards from the rest of the intrusion to the south, and (2) at the southern border of the intrusion.

The country rock at both localities comprises grey, micaceous marble, graphite-muscovite-quartz schist, and albite-biotite-quartz schist (map-unit 1c). Some of the schistose rocks have been intensely sheared. The one specimen studied in thin section has a blastomylonitic texture, with highly strained quartz grains and streaky, mortared quartz lenses. These rocks are doubtless correlative with chlorite-muscovite schists on the other side of the bay from Cape Richards that are known to be of pre-Middle Ordovician age (Trettin, 1969b).

The igneous contact is sharp and can be seen even from a distance, due to the contrast between light intrusive rock and the relatively darker country rock, and to associated extensive rusty zones. Xenolithic blocks of marble were observed well inside the intrusion.

Crush texture is displayed also by the igneous rocks of the contact zone but was not seen in the central part of the complex. The cataclastic texture of the rocks inside and outside the contact, the xenoliths, and the sharpness of the contact all point to forceful emplacement of at least the outer zone of the intrusion.

### *Age*

Potassium-argon age determinations have been made on coexisting hornblende and biotite from a quartz monzonite of the northern outer zone. As is usual, the biotite age, 347 ± 15 m.y., is younger than the hornblende age, 390 ± 18 m.y. Neither mineral concentrate was particularly pure, according to the Isotope Geology Section of the Geological Survey of Canada, but the hornblende age is probably a reliable minimum age of intrusion of the rock. The large discrepancy between the two values is probably due to the cataclasis undergone by the outer zone rocks and the lower argon retentivity of biotite with respect to hornblende. The hornblende age coincides with that determined on hornblende from the syenodiorite intrusion in the M'Clintock West massif and agrees, within the limits of experimental error, with the biotite age from the Cape Fanshawe Martin intrusion. Clearly, a good case can be made for a major period of intrusive activity in Early Devonian (Caledonian) time in northern Ellesmere Island.

## The Phillips Inlet Intrusion and Related Rocks
(Fig. 29E; map-unit E)

A granitic pluton, first described by Christie (1957), underlies the region inland from Cape Woods, at the entrance to Phillips Inlet. It is the largest of several intrusions known in the area and is here referred to as the Phillips Inlet intrusion. Other smaller granitic bodies to the east and south have been mapped by Trettin (1969a, 1971), whose findings are summarized in a later section of this report. Because the various granitic intrusions in the Phillips Inlet area appear to be related, their origins and ages are discussed together.

The Phillips Inlet intrusion covers an area of 64 square miles, of which approximately half is under ice or drift. However, exposure is excellent on the south-facing, glacially eroded slopes of mountains that are up to 3,500 feet high.

Mineralogically, the intrusion appears to be homogeneous, consisting of biotite quartz monzonite to granodiorite, veined by minor aplite and pegmatite (*see* Table IX for modal analyses and Fig. 28). Texturally, the main intrusive rock is variable: it changes from equigranular to porphyritic and back to equigranular over a distance of a few feet, and at the contact it is relatively fine grained and cataclastic.

### *Petrography and Structure*

#### *Central Zone*

The typical rock of the intrusion is a fresh-looking biotite quartz monzonite or granodiorite. Megacrysts of potassium feldspar, 1 to 4 cm long, white to pale pink, are abundant. The groundmass (grain size 1–3 mm) consists of plagioclase, potassium feldspar, biotite, and quartz. Foliation is generally absent but in places a vague alignment of biotite flakes is discernible. A well-developed, east-dipping joint set is discordant to the foliation.

Plagioclase is in euhedral to subhedral tablets and displays oscillatory zoning with a superimposed normal trend. However, whereas the most calcic composition lies in the core,

| | 1 | 2 |
|---|---|---|
| Quartz | 29.2 | 33.2 |
| K-feldspar (perthite) | 21.4 | 18.1 |
| Plagioclase | 34.8 | 37.5 |
| Biotite | 12.4 | 9.1 |
| Muscovite | 0.1 | – |
| (after plag.) | | |
| Light accessories | 1.3 | 1.2 |
| (apatite, zircon) | | |
| Dark accessories | 0.8 | 0.9 |
| (allanite, opaques) | | |
| Points counted | 2026 | 2005 |
| Plagioclase comp. | $An_{26-20}$ | $An_{25-17}$ |
| | | |
| $SiO_2$ | | 73.2 |
| $TiO_2$ | | 0.30 |
| $Al_2O_3$ | | 12.8 |
| Tot. Fe as $Fe_2O_3$ | | 1.8 |
| MnO | | 0.05 |
| MgO | | <0.5 |
| CaO | | 1.6 |
| $Na_2O$ | | 3.8 |
| $K_2O$ | | 2.8 |
| Total | | 96.4 |

Analysts: Rapid Methods Group, Geol. Survey of Canada.

1. Quartz monzonite from central zone.
2. Granodiorite from central zone.

*GSC*

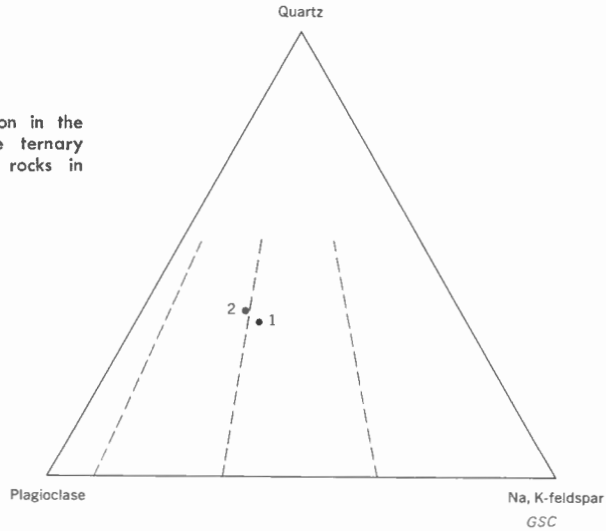
**TABLE IX**

***Modal and chemical analyses of rocks from the Phillips Inlet intrusion.***

the outermost rim may not always be the most sodic part of the crystal. Compositions range from $An_{26}$ to $An_{17}$. Selective sericitization has affected the calcic zones of most grains. The potassium feldspar is *microcline perthite*, which contains numerous inclusions of 0.5-mm plagioclase grains and, more rarely, biotite flakes and quartz. The microcline evidently continued to grow past the stage of groundmass crystallization. Myrmekite is abundantly developed in margins of grains adjacent to plagioclase. *Biotite* has X = straw yellow and Z = deep brown, and is fresh. *Quartz* tends to cluster in aggregates of anhedral, slightly strained grains. Accessory minerals are allanite, zircon, apatite, and opaque iron "ore." The texture is hypidiomorphic-granular, characteristic of undeformed granitic rocks.

FIGURE 28

Rocks of the Phillips Inlet intrusion in the quartz-Na, K-feldspar-plagioclase ternary classification. Numbers refer to rocks in Table IX.




The typical rock described above commonly grades into non-porphyritic but otherwise similar types, or into even more felsic quartz monzonite.

Aplite[1] and pegmatite dykes, veins and patches are extremely common in the central zone. Numerous examples of aplite and pegmatite side-by-side also were found. This association was not observed near the contact but aplite dykes and veins abound throughout the intrusion.

The aplites are pink or pale grey and consist almost solely of plagioclase ($An_{15}$), microcline perthite, and quartz; biotite and muscovite are minor accessory minerals. The pegmatite is formed of pink potassium feldspar, quartz (commonly graphically intergrown with the feldspar), and biotite or pale greenish muscovite.

There is nothing unusual about aplite or pegmatite occurring alone but the intimate coexistence of the two is of interest. Veins of coexisting aplite and pegmatite are about 6 inches wide and of several different configurations: aplite bordered on both sides by pegmatite; pegmatite bordered by aplite; and alternating aplite and pegmatite, resulting in pegmatite layers bordering aplite, in the centre of which is a pegmatite layer. Inclusions of pegmatite in aplite were not found, suggesting that aplite did not intrude pegmatite. The veinlets of pegmatite cutting aplite and rare aplite inclusions in pegmatite support an intrusive origin of the pegmatite. However, chilled margins are absent in the pegmatite (and in the aplite), indicating that intrusion of pegmatite into relatively cool, solidified aplite is unlikely. It seems hardly plausible that the two varieties of rock were formed by simultaneous injection of volatile-rich and volatile-poor liquids; the liquids would mix too readily within such a small space. Some sort of recrystallization of aplite to pegmatite may explain the association better than a multiple injection that produced composite dykes.

Reitan (1965) has proposed the mechanism of secondary recrystallization to account for aplite-pegmatite veins such as those described here. In metals it is commonly observed that a few grains have grown at the expense of many primary grains in a matrix that is energetically nearly stable, but not completely. A primary grain of low interfacial energy relative to the neighbouring grains would tend to enlarge by consuming the surrounding

[1]Aplite, as used here, denotes any kind of fine-grained, sugary textured granitic rock.

grains, leading to a more stable configuration. The uniform, fine-grained texture of the aplite may fulfil the requirements for secondary recrystallization. Grains suitably oriented for secondary recrystallization may have developed in certain zones of the aplite veins, due to dynamic and/or thermal differences across the vein as it intruded. These zones or loci are now reflected in the various arrangements of pegmatite layers. Veins composed wholly of pegmatite may be the products of complete secondary recrystallization but, in many instances, recrystallization was apparently only partial, as aplite is still present. According to Reitan (1965, p. 34), metallurgists have shown that secondary recrystallization is minimal at low temperatures. This may explain the absence of pegmatite in the marginal zone of the intrusion, which presumably cooled more quickly than did the central zone.

### *Marginal Zone*

The intrusive rocks near the igneous contact are mineralogically similar to those of the central zone, but differ markedly in texture. They are foliated and, in thin section, exhibit a cataclastic texture, which is most intense at the contact. Grain size and mafic mineral content decrease toward the contact.

The undeformed quartz monzonite of the central zone is cataclastically deformed in the marginal zone. In the latter, plagioclase, microcline and quartz all show strain and the biotite is largely chloritized. Cataclasis becomes even more marked in the rock at the contact. Foliation, visible even in hand specimen, is manifested by parallel, discontinuous lenses of feldspar and quartz alternating with finer grained, crushed quartzo-feldspathic material.

This evidence for deformation in the border zone of the intrusion suggests that the Phillips Inlet pluton was emplaced forcefully.

### *The Country Rocks and Contact Zone*

With the exception of the contact zone, the rocks around the intrusion are of greenschist facies grade (map-unit 1c). At the southern end of the intrusion, pelitic rocks occurring along the contact include sillimanite-kyanite-staurolite schist, which is described below. Farther from the contact and overlying the schist are quartzite and marble that have been flow-folded. To the east, a great thickness of unfossiliferous, weakly metamorphosed Lower Silurian (Trettin, 1967, p. 16) limestone follows and the talus contains abundant slaty rock. Country rocks at the northern end of the intrusion are sericitic, conglomeratic quartzite and marble (R.L. Christie, 1967, pers. com.).

In the sillimanite-kyanite-staurolite schist of the contact zone, *staurolite* is abundantly developed in 0.5- to 1-mm grains. It is embayed and partly replaced by quartz of the matrix and commonly charged with graphite dust. *Kyanite* (1 mm) is more altered than staurolite and in thin section shows evidence of having been rotated during growth. All kyanite grains are fractured and partly replaced by muscovite. Of the three key minerals, only *sillimanite* shows a preferred orientation, parallel to the foliation; it is in the form of fibrolite, replacing reddish brown biotite, which defines the foliation. The sillimanite is restricted to a few narrow bands in thin section, suggesting that it is of metasomatic origin, i.e., deposited by alumina-rich solutions (derived, perhaps, from the breakdown of kyanite) on biotite, which acted as a nucleating agent, as has been proposed by Chinner (1961). In any event, there is no evidence that the sillimanite formed by replacement of another aluminosilicate polymorph. Muscovite appears as relatively broad, clean laths lying across the foliation and replacing kyanite, and also is intergrown with biotite. The cross-muscovite, and perhaps that associated with biotite, probably represents post-movement growth from materials supplied by late fluids from the granitic intrusion. Greenish, zoned tourmaline is a common accessory.

Although the schist has all the aspects of a regionally metamorphosed rock, it is considered to have developed entirely as a result of the granitic intrusion. In other words, it is a contact metamorphic rock because (1) no rocks even approaching this grade of metamorphism occur in the area of the Phillips Inlet intrusion, and (2) the schist is found immediately adjacent to the intrusive rock.

Deer, Howie, and Zussman (1962a, p. 159) contend that staurolite is unstable in the environment of thermal metamorphism and that its occurrence in contact aureoles is exceptional. The facts, however, do not bear this out. Pitcher and Read (1963, p. 293) describe contact rocks from the Donegal granites of Ireland that are similar to the rock discussed here and list several additional examples to show that staurolite is a common mineral in contact aureoles. Hoschek (1967a, b) gives a number of other examples and cites results of his own experiments, which indicate that staurolite begins to form at around 550°C at 4 to 8 kb. Experimental work by Richardson (1968) has shown that Fe-staurolite is stable at low pressure and may persist up to granulite facies temperatures.

In the Phillips Inlet intrusion, the "regional" metamorphic aspect of the sillimanite-kyanite-staurolite schist in the contact aureole was probably due to the heat and stress of forceful emplacement of the igneous rock.

The southern contact of the intrusion is sharp and straight, parallel to a major northeast-trending joint plane in the granitic rock. A vague foliation visible in the marginal intrusive rock is discordant to the attitude of the metasedimentary rocks, which dip steeply to the southeast. Xenoliths of schist are present in the granitic rock but generally lie concordant to the attitude of the country rocks. Small veins of granitic rock cut the metasedimentary rocks.

### *Mafic Dykes*

Numerous dark dykes cut the intrusion and the country rocks. Study of aerial photographs reveals two main trends of dykes: one northeast, and the other northwest, with generally vertical dips. Those recognizable on the photographs are shown on the geological map. A few were examined both in the field and microscopically.

Most dykes are diabase, several feet wide, with chilled, aphyric, intergranular texture, and fine to medium grain size. The chief minerals are labradorite, zoned to andesine at crystal margins, and pale brownish clinopyroxene, slightly uralitized at the edges. Black iron oxide with a little associated biotite is a common accessory mineral. None of the primary minerals is significantly altered but interstices are usually filled with clay and carbonate minerals.

One dyke examined may be a lamprophyre that has been considerably altered. This dyke is less than a foot wide and microporphyritic. In thin section, the "microphenocrysts" are seen to be (1) carbonate pseudomorphs of skeletal olivine(?) and perhaps other minerals, (2) ocelli(?) of carbonate rimmed by chlorite and, more rarely, (3) highly altered plagioclase laths. The groundmass is a dense aggregate of carbonate(?) and innumerable minute needles of black "ore" arranged in a grid pattern. Abundant carbonate pseudomorphs of mafic minerals, ocellar-like texture, and the well-known association of granitic and lamprophyric rocks are evidence that the dyke is a lamprophyre.

## *Other Intrusions in the Phillips Inlet Area*

Trettin (1969a, 1971) has studied several smaller granitic plutons south and east of the Phillips Inlet intrusion. They are mostly concordant, in part foliated, and vary in composition

from quartz diorite to quartz monzonite. The foliate texture, abundance of quartz and occurrence of garnet in some of these bodies suggest that such rocks are not truly igneous but metamorphic or metasomatic (Trettin, 1971). Local metamorphic terrains accompany and appear to be related to the intrusions (*see* Chap. II). The country rocks, except those in the immediate vicinity of the intrusions, are of low metamorphic grade; Silurian and older strata are represented among them (Trettin, 1971).

### *Petrogenesis of the Intrusions in the Phillips Inlet Area*

Several classifications of granitic rock have been proposed and recently discussed. Two of them may be used to advantage in the interpretation of the granitic intrusions around Phillips Inlet. Buddington (1959) categorized granitic rocks according to their level of emplacement in the earth's crust: epizone, mesozone, and catazone. Marmo (1962) has elaborated on Eskola's tectonic classification: synkinematic, late-kinematic, and post-kinematic intrusions.

The weight of evidence indicates epizonal emplacement of the main Phillips Inlet intrusion. The inner part of the body, where studied, is structureless. The country rocks, except at the contact, are weakly metamorphosed (greenschist facies). A sillimanite-bearing contact aureole has been developed. Unfavourable to the hypothesis of epizonal emplacement is the broad concordance of the southeastern margin of the intrusion with the country rock, but there is discordance on the scale of an outcrop.

The intrusive rocks vary from quartz monzonite to granodiorite; the potassium feldspar is microcline, which continued to grow at a late stage of crystallization of the magma; and the plagioclase is oligoclase. All these features are characteristic of synkinematic granitic intrusions (Marmo, 1962). Chemical analysis of a microcline-rich granodiorite (Table IX, no. 2) shows a low $K_2O$ content, 2.8 per cent. This is the average potash content of Svecofennidian synkinematic granitic rocks (Simonen, 1948, Table II, no. 5).

The smaller plutons described by Trettin (1971), which are largely concordant and foliated and emplaced in rocks that are generally of low metamorphic grade but include some migmatite, may be classified as mesozonal and synkinematic. Radiometric age determinations reported below suggest that they are lower level equivalents of the Phillips Inlet intrusion.

### *Age of the Intrusions in the Phillips Inlet Area*

Potassium-argon age determinations on biotite are available from two intrusions. One was made on coarse-grained quartz monzonite from the central zone of the Phillips Inlet intrusion and yielded a date of $345 \pm 15$ m.y. The other, made on gneissic quartz monzonite from a smaller intrusion 8 miles to the southeast, gave $325 \pm 14$ m.y. (Trettin, 1971). Thus, the two dates agree within the given error limits and lie around the Devonian–Mississippian boundary, according to the Geological Society of London time-scale (1964, p. 262).

The radiometric ages are undoubtedly minimum values and suggest intrusion in mid-Paleozoic time. As some of the smaller plutons have intruded Silurian rocks, they can be no older than Silurian. Accepting that metamorphism of regional extent accompanied intrusion, one may conclude that, in the Phillips Inlet area, granitic intrusion and metamorphism were an expression of the Middle Devonian to Mississippian orogeny in the Franklinian Geosyncline.

## Other Felsic Intrusions

Several intrusions of quartz diorite to quartz monzonite composition occur between Phillips Inlet and Kleybolte Peninsula. The plutons are small (generally less than 1.5 miles in diameter), massive, discordant, and intrude rocks that have been slightly regionally metamorphosed. A granitic stock east of Henson Bay (Trettin, 1969a, 1971) is of quartz monzonite metamorphosed in the greenschist facies; it resembles the smaller granitic bodies around Phillips Inlet.

The plutons on Kleybolte Peninsula are similar in lithology and geologic setting to granitic bodies in northern Axel Heiberg Island (Trettin, 1969a, p. 27). A biotite K-Ar age determination on one of the latter gave $360 \pm 25$ m.y., which suggests that the Kleybolte Peninsula intrusions also are related to the Middle Devonian to Mississippian orogeny (Trettin, 1969a, p. 45).

Biotite from the pluton near Henson Bay has an apparent K-Ar age of $261 \pm 12$ m.y. but this low value is considered anomalous and probably resulted from argon loss during Pennsylvanian or later fault movements (Trettin, 1969a, p. 45).

# REFERENCES


Blackadar, R. G.
1954: Geological reconnaissance, north coast of Ellesmere Island, Arctic Archipelago, Northwest Territories; Geol. Surv. Can., Paper 53-10.
1960: The age of the metamorphic complex of northernmost Ellesmere Island; Arctic, v. 13, p. 51–52.

Bowen, N. L.
1928: The evolution of the igneous rocks; Dover Publs., Inc., New York, 1956, 334 p.

Bowes, D. R. and Park, R. G.
1966: Metamorphic segregation banding in the Loch Kerry basite sheet, from the Lewisian of Gairloch, Ross-shire, Scotland; J. Petrol., v. 7, p. 306–330.

Buddington, A. F.
1939: Adirondack igneous rocks and their metamorphism; Geol. Soc. Amer., Mem. 7.
1959: Granite emplacement with special reference to North America; Bull. Geol. Soc. Amer., v. 70, p. 671–747.

Burri, C., Parker, R. L., and Wenk, E.
1967: Die optische Orientierung der Plagioklase; Birkhäuser Verlag, Basel, 334 p.

Chayes, F.
1955: Potash feldspar as a by-product of biotite-chlorite transformation; J. Geol., v. 63, p. 75–82.

Chinner, G. A.
1961: The origin of sillimanite in Glen Cova, Angus; J. Petrol., v. 2, p. 312–323.

Christie, R. L.
1957: Geological reconnaissance of the north coast of Ellesmere Island, District of Franklin, Northwest Territories; Geol. Surv. Can., Paper 56-9.
1964: Geological reconnaissance of northeastern Ellesmere Island, District of Franklin; Geol. Surv. Can., Mem. 331.

Deer, W. A., Howie, R. A., and Zussman, J.
1962a: Rock-forming minerals, Vol. 1: Ortho- and ring-silicates; Longmans, Green, London, 333 p.
1962b: Rock-forming minerals, Vol. 3: Sheet silicates; Longmans, Green, London, 270 p.
1963a: Rock-forming minerals, Vol. 2: Chain silicates; Longmans, Green, London, 379 p.
1963b: Rock-forming minerals, Vol. 4: Framework silicates; Longmans, Green, London, 435 p.

Douglas, R. J. W., Norris, D. K., Thorsteinsson, R., and Tozer, E. T.
1963: Geology and petroleum possibilities of northern Canada; Geol. Surv. Can., Paper 63-31.

Ernst, W. G.
1962: Synthesis, stability relations and occurrence of riebeckite and riebeckite-arfvedsonite solid solutions; J. Geol., v. 70, p. 689–736.

Feilden, H. W. and 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.; Quart. J. Geol. Soc. London, v. 34, p. 556–567.

Fyfe, W. S. and Turner, F. J.
1966: Reappraisal of the metamorphic facies concept; Contrib. Mineral. Petrol., v. 12, p. 354–364.

Geological Society of London
1964: Geological Society Phanerozoic time-scale; Quart. J. Geol. Soc. London, v. 120S, p. 206–262.

Guidotti, C. V.
1970: The mineralogy and petrology of the transition from the lower to upper sillimanite zone in the Oquossoc area, Maine; J. Petrol., v. 11, p. 277–366.

Haller, J. and Kulp, J. L.
1962: Absolute age determinations in East Greenland; Medd. om Grønland, Bd. 171, Nr. 1.

Hamilton, E. I.
1965: Applied geochronology; Academic Press, New York, 267 p.

Hoschek, G.
1967a: Untersuchungen zum Stabilitätsbereich von Chloritoid und Staurolith; Contrib. Mineral. Petrol., v. 14, p. 123–162.
1967b: Zur unteren Stabilitätsgrenze von Staurolith; Naturwiss., 54 Jg., p. 200.

Irvine, T. N.
1967: Chromian spinel as a petrogenetic indicator, Pt. 2: Petrologic applications; Can. J. Earth Sci., v. 4, p. 71–100.

Jambor, J. L. and Smith, C. H.
1964: Olivine composition determination with small-diameter X-ray powder cameras; Mineral. Mag., v. 33, p. 730–741.

Kerr, J. W. and Christie, R. L.
1965: Tectonic history of Boothia Uplift and Cornwallis Fold Belt, Arctic Canada; Bull. Amer. Assoc. Petrol. Geol., v. 49, p. 905–926.

Kerr, J. W. and Trettin, H. P.
1962: Mississippian rocks and the mid-Paleozoic earth-movements in the Canadian Arctic Archipelago; J. Alberta Soc. Petrol. Geol., v. 10, p. 247–256.

King, B. C.
1963: The nature of basic igneous rocks and their relations with associated acid rocks, Pt. II; Sci. Progress, v. 51, p. 259–275.

Kretz, R.
1966: Metamorphic differentiation at Einasleigh, northern Queensland; J. Geol. Soc. Australia, v. 13, p. 561–582.

Leake, B. E.
1964: The chemical distinction between ortho- and para-amphibolites; J. Petrol., v. 5, p. 238–254.

Lyons, J. B. and Leavitt, F. G.
1961: Structural-stratigraphic studies on the Ward Hunt Ice Shelf; USAF Cambridge Res. Labs. AFCRL-442.

Marmo, V.
1962: On granites; Comm. géol. Finlande, Bull. 201.

Mehnert, K. R.
1968: Migmatites and the origin of granitic rocks; Elsevier, New York, 393 p.

Miyashiro, A.
1961: Evolution of metamorphic belts; J. Petrol., v. 2, p. 277–311.

Moores, E. M.
1969: Petrology and structure of the Vourinos ophiolitic complex of northern Greece; Geol. Soc. Amer., Spec. Paper 118.

Müller, G.
1966: Die autometamorphe retrograde Umwandlung von Biotiten in Chlorite and Muskowite in sauren Tiefengesteinen; Contrib. Mineral. Petrol., v. 13, p. 295–365.

Nockolds, S. R.
1954: Average chemical composition of some igneous rocks; Bull. Geol. Soc. Amer., v. 65, p. 1007–1032.

Olsen, E. J.
1961: High temperature acid rocks associated with serpentine in eastern Quebec; Amer. J. Sci., v. 259, p. 329–347.

Peary, R. E.
1907: Nearest the Pole; Hutchinson, London, 410 p.

Pettijohn, F. J.
1963: Chemical composition of sandstones—excluding carbonate and volcanic sands; U.S. Geol. Surv., Prof. Paper 440-S.

Pitcher, W. S. and Read, H. H.
1963: Contact metamorphism in relation to manner of emplacement of the granites of Donegal, Ireland; J. Geol., v. 71, p. 261–296.

Poldervaart, A.
1955: Chemistry of the earth's crust; *in* Crust of the Earth, A. Poldervaart, ed., Geol. Soc. Amer., Spec. Paper 62, p. 119–144.

Presnall, D. C.
1966: The join forsterite-diopside-iron oxide and its bearing on the crystallization of basaltic and ultramafic magmas; Amer. J. Sci., vol. 264, p. 753–809.

Rein, G.
1953: Der Werdegang des Orthits in der magmatischen und metamorphen Abfolge des mittleren Schwarzwaldes; N. Jb. Mineral. Abh., Bd. 84, p. 365–435.

Reitan, P. H.
1965: Pegmatite veins and the surrounding rocks, V: Secondary recrystallization of aplite to form pegmatite; Norsk. Geol. Tidsskr., v. 45, p. 31–40.

Richardson, S. W.
1968: Staurolite stability in a part of the system Fe-Al-Si-O-H; J. Petrol., v. 9, p. 467–488.

Rittmann, A. and El-Hinnawi, E. E.
1961: The application of the zonal method for the distinction between low- and high-temperature plagioclase feldspars; Schweiz. Mineral. Petrog. Mitt., v. 41, p. 41–48.

Sastri, G. G. K.
1962: Determination of the end-member composition of garnets from their physical properties; Geol. Surv. India Records, v. 87, p. 757–780.

Schuchert, C.
1923: Sites and nature of North American geosynclines; Bull. Geol. Soc. Amer., v. 34, p. 151–299.

Sederholm, J. J.
1916: On synantetic minerals and related phenomena; Comm. géol. Finlande, Bull. 48.

Simonen, A.
1948: On the petrochemistry of the infracrustal rocks in the Svecofennidic territory of southwestern Finland; Comm. géol. Finlande, Bull. 141.

Slemmons, D. B.
1962: Determination of volcanic and plutonic plagioclases using a three- or four-axis universal stage; Geol. Soc. Amer., Spec. Paper 69.

Sutton, J.
1965: Some recent advances in our understanding of the controls of metamorphism; *in* Controls of metamorphism, W. S. Pitcher and G. Flinn, eds., p. 22–45; J. Wiley, New York, 368 p.

Taylor, H. P., Jr. and Noble, J. A.
1960: Origin of the ultramafic complexes in southeastern Alaska; Rept. XXI Int. Geol. Congr., Pt. 13, p. 175–187.

Thayer, T. P.
1963: The Canyon Mountain complex, Oregon, and the alpine mafic magma stem; U.S. Geol. Surv., Prof. Paper 475-C, p. C82–85.

Thorsteinsson, R. and Tozer, E. T.
1970: Geology of the Arctic Archipelago; *in* Geology and economic minerals of Canada, R. J. W. Douglas, *ed.*, p. 548–590; Geol. Surv. Can., Econ. Geol. Rept. No. 1.

Tozer, C. F.
1955: The mode of occurrence of sillimanite in the Glen district, Co. Donegal; Geol. Mag., v. 92, p. 310–320.

Trettin, H. P.
1964: Pre-Mississippian rocks of Nansen Sound area, District of Franklin; Geol. Surv. Can., Paper 64-26.
1966: Precambrian to Carboniferous rocks of M'Clintock Inlet region, northeastern Ellesmere Island; Geol. Surv. Can., Paper 66-1, p. 7–11.

Trettin, H. P.—*Concluded*
1967: Geology of pre-Mississippian "eugeosynclinal" rocks in selected areas of northern Ellesmere Island; Geol. Surv. Can., Paper 67-1, Pt. A, p. 13–18.
1969a: Pre-Mississippian geology of northern Axel Heiberg and northwestern Ellesmere Islands, Arctic Archipelago; Geol. Surv. Can., Bull. 171.
1969b: Geology of Ordovician to Pennsylvanian rocks, M'Clintock Inlet, north coast of Ellesmere Island, Canadian Arctic Archipelago; Geol. Surv. Can., Bull. 183.
1971: Reconnaissance of lower Paleozoic geology, Phillips Inlet region, north coast of Ellesmere Island; Geol. Surv. Can., Paper 71-12.

Tröger, W. E.
1959: Optische Bestimmung der gesteinsbildenden Minerale, Teil 1: Bestimmungstabellen; 3rd ed.; E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, 147 p.

Turner, F. J. and Verhoogen, J.
1960: Igneous and metamorphic petrology; 2nd ed.; McGraw-Hill, New York, 694 p.

Turner, F. J. and Weiss, L. E.
1963: Structural analysis of metamorphic tectonites; McGraw-Hill, New York, 545 p.

Tuttle, O. F. and Bowen, N. L.
1958: Origin of granite in the light of experimental studies in the system $NaAlSi_3O_8 - KAlSi_3O_8 - H_2O$; Geol. Soc. Amer., Mem. 74.

Uruno, K.
1963: Optical study of the ordering degree of plagioclase; Sci. Repts. Tohoku Univ., 3rd ser., v. 8, p. 171–220.

Wager, L. R., Brown, G. M., and Wadsworth, W. J.
1960: Types of igneous cumulates; J. Petrol., v. 1, p. 73–85.

Wells, M. K.
1962: Structure and petrology of the Freetown layered basic complex of Sierra Leone; Overseas Geol. Min. Res. Bull., Suppl. 4.

Whitten, E. H. T.
1966: Structural geology of folded rocks; Rand McNally, Chicago, 678 p.

Winkler, H. G. F.
1965: Petrogenesis of metamorphic rocks; Springer Verlag, New York, 220 p.

Wyllie, P. J.
1963: Effects of the changes in slope occurring on liquidus and solidus paths in the system diopside-anorthite-albite; *in* Mineral. Soc. Amer., Spec. Paper 1, p. 204–212.

## APPENDIX 1: Terminology

*Gneiss.* A metamorphic rock with abundant feldspar and/or quartz and with a poor to good foliation defined by platy or prismatic minerals, which are subordinate in amount to quartz and feldspar.

*Grain sizes.* Fine: less than 1 mm.
Medium: between 1 and 5 mm.
Coarse: greater than 5 mm.

*Granitic* rocks may include all the rocks shown in Figure 5, but is most commonly used in this report for those varieties containing more than 10 per cent quartz. *Granite* denotes only those falling in the granite field of Figure 5.

*Marble.* A metamorphic rock composed mainly of calcite and/or dolomite.

*Schist.* A metamorphic rock with abundant platy or prismatic grains and with a well-defined foliation (excluding amphibolite).

## APPENDIX 2: Chemical Analytical Methods

Chemical analyses reported here were made by the Analytical Chemistry Section of the Geological Survey of Canada in 1967. Si, Al, Fe, Ca, Mg, K, Ti, and Mn were determined by X-ray fluorescence spectroscopy, using an A.R.L. multi-channel vacuum spectrometer. Na was determined by flame photometry.

The X-ray fluorescence method used does not require fusion of the material but involves a large aliquot portion of the powdered sample. Quantitative results are obtained by referring the instrumental readings to working curves prepared for standards having compositions as closely related to the unknowns as possible. Each batch of samples is monitored by a series of laboratory standards that are run at the same time.

The estimated accuracy of the individual determinations is as follows, expressed as $\pm$ 1 standard deviation (or $1\sigma$):

$SiO_2$ (30–70%) $\pm$ 1.2%
$Al_2O_3$ (up to 20%) $\pm$ 0.7
Total Fe as $Fe_2O_3$ $\pm$ 0.5
(up to 15%)
CaO (up to 40%) $\pm$ 0.3
MgO (up to 40%) $\pm$ 1.0
$K_2O$ (up to 5%) $\pm$ 0.1
$TiO_2$ (up to 2%) $\pm$ 0.05
MnO (up to 1%) $\pm$ 0.02
$Na_2O$ (up to 10%) $\pm$ 0.15

# APPENDIX 3: Mineral Assemblages

Listed below are the mineral assemblages that have been found in predominantly prograde regionally metamorphosed rocks in northernmost Ellesmere Island. In each assemblage, minerals are listed in order of increasing abundance. Fe-Ti oxide minerals are not considered.

*Cape Columbia Complex*

1. Garnet-biotite-quartz-oligoclase.
2. Biotite-garnet-muscovite-quartz-oligoclase.
3. Muscovite-garnet-biotite-quartz-oligoclase-microcline.
4. Biotite-sphene-garnet-quartz-oligoclase-hornblende.
5. Scapolite-garnet-biotite-quartz-oligoclase-hornblende.
6. Epidote (incl. clinozoisite)-scapolite-biotite-quartz-hornblende-andesine.
7. Scapolite-epidote (incl. clinozoisite)-garnet-quartz-oligoclase-hornblende.
8. Biotite-chlorite-albite-garnet-quartz-muscovite.
9. Chlorite-albite-muscovite-quartz.

*M'Clintock Inlet and Cape Richards area*

10. Muscovite-biotite-quartz-microcline-oligoclase.
11. Biotite-sillimanite (fibrolite)-muscovite-plagioclase-quartz.
12. Tremolite-scapolite-phlogopite-quartz-carbonate.
13. Staurolite-garnet-muscovite-quartz.
14. Graphite-muscovite-quartz.
    Also 8 (less chlorite, garnet and muscovite), 9 (less albite).

*Ayles Fiord to Henson Bay*

15. Graphite-kyanite-garnet-staurolite-biotite-muscovite-oligoclase-quartz.
16. Garnet-tourmaline-chlorite-staurolite-biotite-oligoclase-quartz.
17. Epidote-sphene-hornblende-biotite-microcline-quartz-oligoclase.
18. Apatite-chlorite-garnet-carbonate-biotite-hornblende-quartz-oligoclase.
19. Biotite-hornblende-quartz-andesine.
20. Muscovite-plagioclase-chlorite-carbonate-quartz.
21. Muscovite-biotite-carbonate-quartz.
22. Feldspar-carbonate-garnet-biotite-muscovite-quartz.
    Also 1, 2, 3, 6 (less scapolite), 8 (with and without garnet), 9, 10.

# Index


PAGE

Amphibolite.....16, 19, 24, 25, 30, 32, 35, 36, 42–45
Andesite.....25, 56
Anorthosite..... 54
Aplite..... 76
Axel Heiberg Island.....3, 80

Boudinage.....20, 21, 27, 42

Caledonian Orogeny.....3, 27, 32, 74
Carolinidian Orogeny..... 27
Cataclasis.....7, 8, 12, 14, 15, 18, 28, 34, 39, 44, 46, 50, 57, 58, 67, 74, 77
Chemical composition, igneous rocks 62, 73, 79
Chemical composition, metamorphic rocks 25, 42–44
Corona..... 19

Diabase.....31, 46, 56, 78
Diorite.....59, 62, 65, 67
Dykes.....6, 19, 20, 25, 26, 32, 46, 55, 56, 68, 72, 73, 78

Feilden Peninsula..... 4

Gabbro.....19, 25, 50, 51, 53, 54, 56
Garnet.....11, 12, 14, 16, 21, 22, 23, 28, 34, 35, 37, 39, 41, 43, 45, 47, 79
Granodiorite.....62, 67, 73, 74, 79
Greenland.....3, 27

Henson Bay.....47, 80

K/Ar ages.....26, 27, 32, 33, 47, 56, 65, 74, 79, 80
Kleybolte Peninsula.....67, 80
Kyanite.....37, 48, 77

Lamprophyre..... 78

Marble.....6, 19, 23, 27, 28, 29, 39, 45, 49, 73, 77
Mid-Paleozoic Orogeny.....3, 47, 79, 80

PAGE

Olivine.....5, 52, 54, 55, 58, 66
Ordovician rocks.....3, 27, 28, 29, 47, 58, 67

Pegmatite.....6, 19, 21, 26, 32, 46, 76
Peridotite.....54, 55, 58, 63, 64
Peridotite, metamorphosed..... 66
Permo-Carboniferous rocks 23, 24, 28, 58, 64
Phyllite.....23, 24, 29, 30, 32, 47
Plagioclase, determinative methods..... 5
Porphyroblasts.....7, 8, 9, 12, 13, 16, 39, 40, 41
Pseudotachylite.....44, 45
Pyroxene.....19, 52, 53, 54, 58, 59, 63, 64, 66, 69
Pyroxenite.....54, 63

Quartz diorite, igneous.....62, 65, 69, 79, 80
Quartz diorite, metamorphic.....8, 33, 41–44
Quartz monzonite, igneous.....53, 67, 73, 74, 79, 80
Quartz monzonite, metamorphic..30, 33, 39–41
Quartzite.....6, 23, 29, 30, 33, 39, 45, 77

Rb/Sr ages..... 48
Retrogressive metamorphism.....12, 13, 14, 16, 18, 28, 34, 39
Riebeckite granite.....72, 73

Scapolite.....16, 18, 43
Schist, calcareous.....19, 45
Schist, pelitic.....20, 21, 22, 24, 25, 26, 28, 30, 34, 36, 37, 39, 77
Schist, psammitic.....39, 45, 46
Serpentinite.....28, 55, 57, 59, 64, 67
Sillimanite.....28, 30, 48, 77
Silurian rocks.....3, 28, 47
Staurolite.....28, 37, 77, 78
Syenite.....24, 56, 63, 64, 69, 73
Syenodiorite.....62, 65

Unconformity.....3, 23, 24, 27, 58, 67

Volcanic rock.....36, 46, 47

## BULLETINS

### Geological Survey of Canada

**Bulletins present the results of detailed scientific studies on geological or related subjects. Some recent titles are listed below (Information Canada No. in brackets):**

200 Part I: Biostratigraphy of some Early Middle Silurian Ostracoda, eastern Canada; Part II: Additional Silurian Arthropoda from arctic and eastern Canada, *by* M. J. Copeland, $1.50 (M42-200)

201 Archaeocyatha from the Mackenzie and Cassiar Mountains, Northwest Territories, *by* R. C. Handfield, $2.00 (M42-201)

202 Faunas of the Ordovician Red River Formation, Manitoba, *by* D. C. McGregor, *et al.*, $2.00 (M42-202)

203 Geology of lower Paleozoic formations, Hazen Plateau and southern Grant Land Mountains, Ellesmere Island, *by* H. P. Trettin, $3.00 (M42-203)

204 Brachiopods of the Detroit River Group (Devonian) from southwestern Ontario and adjacent areas of Michigan and Ohio, *by* J. A. Fagerstrom, $2.00 (M42-204)

205 Comparative study of the Castle River and other folds in the Eastern Cordillera of Canada, *by* D. K. Norris, $2.00 (M42-205)

206 Geomorphology and multiple glaciation in the Banff area, *by* N. W. Rutter, $2.00 (M42-206)

207 Geology of the resurgent cryptoexplosion crater at Mistastin Lake, Labrador, *by* K. L. Currie, $2.00 (M42-207)

208 The geology and origin of the Faro, Vangorda and Swim concordant zinc-lead deposits, Central Yukon Territory, *by* D. J. Tempelman-Kluit, $3.00 (M42-208)

209 Redescription of *Marrella splendens* (Trilobitoidea) from Burgess Shale, Middle Cambrian, British Columbia, *by* H. B. Whittington, $3.00 (M42-209)

210 Ordovician trilobites from the central volcanic mobile belt at New World Island, northeastern Newfoundland, *by* W. T. Dean, $2.00 (M42-210)

211 A Middle Ordovician fauna from Braeside, Ottawa Valley, Ontario, *by* H. Miriam Steele and G. Winston Sinclair, $2.00 (M42-211)

212 Lower Cambrian trilobites from the Sekwi Formation type section, Mackenzie Mountains, northwestern Canada, *by* W. H. Fritz, $4.00 (M42-212)

213 Sequence of glacial lakes in north-central Alberta, *by* D. A. St-Onge, $2.00 (M42-213)

214 Classification and description of copper deposits, Coppermine River area, District of Mackenzie, *by* E. D. Kindle, $4.00 (M42-214)

215 Brachiopods of the Arisaig Group (Silurian–Lower Devonian) of Nova Scotia, *by* Charles W. Harper, Jr., $5.00 (M42-215)

217 The geology and petrology of the alkaline carbonatite complex at Callander Bay, Ontario, *by* John Ferguson and K. L. Currie, $2.00 (M42-217)

218 Keeweenawan volcanic rocks of Michipicoten Island, Lake Superior, Ontario (41N): An eruptive centre of Proterozoic age, *by* R. N. Annells, $5.00 (M42-218)

219 Lower Cretaceous Bullhead Group, between Bullmoose Mountain and Tetsa River, Rocky Mountain Foothills, northeastern British Columbia, *by* D. F. Stott, $6.00 (M42-219)

221 Chitinozoa and Acritarcha of the Hamilton Formation (Middle Devonian), southwestern Ontario, *by* J. A. Legault, $4.00 (M42-221)

222 Contributions to Canadian Paleontology, *by* D. E. Jackson, *et al.*, $6.00 (M42-222)

224 Carboniferous and Permian stratigraphy of Axel Heiberg Island and western Ellesmere Island, Canadian Arctic Archipelago, *by* R. Thorsteinsson, $6.00 (M42-224)

229 Metamorphic and plutonic rocks of northernmost Ellesmere Island, Canadian Arctic Archipelago, *by* Thomas Frisch, $4.00 (M42-229)

230 Triassic rocks of the southern Canadian Rocky Mountains, *by* D. W. Gibson, $4.00 (M42-230)

231 *Yohoia* Walcott and *Plenocaris* n. gen. arthropods from the Burgess Shale, Middle Cambrian, British Columbia, *by* H. B. Whittington, $3.00 (M42-231)

232 Condodonts of the Waterways Formation (Upper Devonian) of northeastern and central Alberta, *by* T. T. Uyeno, $4.00 (M42-232)

234 Evolution of a Middle and Upper Devonian sequence from a clastic coastal plain–deltaic complex into overlying carbonate reefs and banks, Sturgeon–Mitsue area, Alberta, *by* L. F. Jansa and N. R. Fischbuch, $5.00 (M42-234)