National Library
of Canada

Bibliothèque nationale
du Canada

CANADIAN THESES
ON MICROFICHE

THÈSES CANADIENNES
SUR MICROFICHE

41662

NAME OF AUTHOR/NOM DE L'AUTEUR JOHN FRANCIS CHAPPELL

TITLE OF THESIS/TITRE DE LA THÈSE THE CLARE RIVER STRUCTURE AND ITS
TECTONIC SETTING

UNIVERSITY/UNIVERSITÉ CARLETON

DEGREE FOR WHICH THESIS WAS PRESENTED/
GRADE POUR LEQUEL CETTE THÈSE FUT PRÉSENTÉE Ph. D.

YEAR THIS DEGREE CONFERRED/ANNÉE D'OBTENTION DE CE DEGRÉ 1979

NAME OF SUPERVISOR/NOM DU DIRECTEUR DE THÈSE DR. R. L. BROWN

Permission is hereby granted to the NATIONAL LIBRARY OF
CANADA to microfilm this thesis and to lend or sell copies
of the film.

The author reserves other publication rights, and neither the
thesis nor extensive extracts from it may be printed or other-
wise reproduced without the author's written permission.

L'autorisation est, par la présente, accordée à la BIBLIOTHÈ-
QUE NATIONALE DU CANADA de microfilmer cette thèse et
de prêter ou de vendre des exemplaires du film.

L'auteur se réserve les autres droits de publication; ni la
thèse ni de longs extraits de celle-ci ne doivent être imprimés
ou autrement reproduits sans l'autorisation écrite de l'auteur.

DATED/DATE Nov 10/78 SIGNED/SIGNÉ J. F. Chappel

PERMANENT ADDRESS/RÉSIDENCE FIXE

5227 MARK BOROUGH DR. N.E.
CALGARY, ALTA
T2A 2Z8

National Library of Canada

Cataloguing Branch
Canadian Theses Division

Ottawa, Canada
K1A 0N4

Bibliothèque nationale du Canada

Direction du catalogage
Division des thèses canadiennes

NOTICE

The quality of this microfiche is heavily dependent upon the quality of the original thesis submitted for microfilming. Every effort has been made to ensure the highest quality of reproduction possible.

If pages are missing, contact the university which granted the degree.

Some pages may have indistinct print especially if the original pages were typed with a poor typewriter ribbon or if the university sent us a poor photocopy.

Previously copyrighted materials (journal articles, published tests, etc.) are not filmed.

Reproduction in full or in part of this film is governed by the Canadian Copyright Act, R.S.C. 1970, c. C-30. Please read the authorization forms which accompany this thesis.

**THIS DISSERTATION
HAS BEEN MICROFILMED
EXACTLY AS RECEIVED**

AVIS

La qualité de cette microfiche dépend grandement de la qualité de la thèse soumise au microfilmage. Nous avons tout fait pour assurer une qualité supérieure de reproduction.

S'il manque des pages, veuillez communiquer avec l'université qui a conféré le grade.

La qualité d'impression de certaines pages peut laisser à désirer, surtout si les pages originales ont été dactylographiées à l'aide d'un ruban usé ou si l'université nous a fait parvenir une photocopie de mauvaise qualité.

Les documents qui font déjà l'objet d'un droit d'auteur (articles de revue, examens publiés, etc.) ne sont pas microfilmés.

La reproduction, même partielle, de ce microfilm est soumise à la Loi canadienne sur le droit d'auteur, SRC 1970, c. C-30. Veuillez prendre connaissance des formules d'autorisation qui accompagnent cette thèse.

**LA THÈSE A ÉTÉ
MICROFILMÉE TELLE QUE
NOUS L'AVONS REÇUE**

THE CLARE RIVER STRUCTURE AND
ITS TECTONIC SETTING

by

JOHN FRANCIS CHAPPELL, B.Sc.

A Thesis submitted to the Faculty of
Graduate Studies and Research in partial fulfillment
of the requirements for the degree of
Doctor of Philosophy

Department of Geology

Carleton University

Ottawa, Ontario

June, 1978

© 1979 John Francis Chappell

The undersigned hereby recommend to the Faculty of Graduate
Studies and Research acceptance of this thesis, submitted by J. F.
Chappell, B.Sc., in partial fulfillment of the requirements for the
degree of Doctor of Philosophy.

W. B. Whipple
.....
Chairman, Department of Geology

Robert W. Whipple
.....
Supervisor

N. M. Stanley
.....
External Examiner

ABSTRACT

The Clare River Structure is one of several northeast-trending supracrustal belts of the Kaladar-Dalhousie Trough, and is separated from the adjacent supracrustal belts by metamorphosed plutonic rocks.

Five formations make up the supracrustal succession of the Clare River Structure. They are as follows, from oldest to youngest: the Otter Creek Formation - massive felsic metavolcanics, biotite-hornblende gneisses and schists, quartzofeldspathic psammite and pelite, rusty psammite, and calcareous marble; the Todd Lake Formation - dolomitic marble, volcanic pebble metaconglomerate, metachert, para-amphibolite, calc-silicate paragneiss; the Shovel Lake Formation - mafic, intermediate, and felsic metavolcanics, calc-silicate paragneisses; the Beatty Formation - pelitic schist, rusty psammite; the Bogart Formation - calcareous marble, dolomitic marble, and various calc-silicate paragneisses.

The upper two formations belong to the Flinton Group, as defined by Moore and Thompson (1972) and the lower three are members of the Hermon Group, defined by Lumbers (1967b).

The outcrop pattern of the Clare River Structure defines a type III interference pattern which can be explained by two major phases of deformation. D1 produced large reclined to recumbent northeast-trending folds which were reoriented by second phase upright, tight to isoclinal, northeast-trending folds. A pervasive mineral foliation and compositional layering (S1) was generated during D1. Long-limbed isoclinal folds (F1) occur locally, but a lineation (L1) is not developed. D2 resulted in a pronounced crenulation lineation (L2), widely spread F2 minor folds and a weak fracture or crenulation cleavage (S2). The entire structure lies in fault contact on the Addington Pluton. On the basis of stratigraphic correlation with adjacent areas, much of the Clare River succes-

ion is interpreted as being overturned.

Metamorphic grade in the Clare River Structure is confined to the amphibolite facies and increases from northeast to southwest. Three mineral zones can be defined in the field, staurolite-kyapite, staurolite-sillimanite, and sillimanite. The peak of metamorphism postdates D1 in the lower grade area, but coincides with the final stages of D1 in the higher grade areas. D2 postdates nearly all mineral growth in the Clare River Structure.

The metavolcanic rocks of the Shovel Lake Formation are sub-alkaline. The basalts show tholeiitic affinity, and the intermediate and felsic rocks show a poorly defined calc-alkalic trend. The chemistry of the basalts suggests that they formed in the early stages of the evolution of an island arc.

ACKNOWLEDGEMENTS

The writer is indebted to Dr. R.L. Brown and Dr. J. M. Moore, who provided encouragement, guidance and advice throughout the course of this study.

The writer is also indebted to Mr. John F. Psutka for his superlative work as a field assistant during the summer of 1975 and for his acting as a sounding board for the embryonic stage of many of the ideas presented in this thesis. Rodger Kidner also assisted in the field for two weeks in the summer of 1977.

Discussions with Dr. P. H. Thompson, Dr. R. L. Morton, Mr. B. C. Ricketts, Mr. R. W. Leatherbarrow, Mr. D. V. Lefebure, Mr. P. A. Goetz and Dr. T. Rivers were stimulating and informative. Dr. J. H. Bourne of the Geological Survey of Canada kindly provided chemical analyses of samples of the Shovel Lake Formation. Students in Carleton University's advanced field course produced maps of parts of the Kaladar Complex and the Clare River Structure that were of use to the writer. Leslie Whitmore drafted several maps, and Ron Conlon and Ross Taylor made countless thin sections. Mr. Jack Hogg's invaluable assistance with all technical aspects involved in this research is gratefully acknowledged.

Financial support in the form of a National Research Council Scholarship (1974-1977) is gratefully acknowledged. Financial support was also obtained from National Research Council Grants to Dr. R. L. Brown and Dr. J. M. Moore, and from Carleton University.

Finally, I wish to thank my wife, Lee, for typing, colouring, and especially for her patience and constant moral support throughout the duration of this study.

TABLE OF CONTENTS

	page
ABSTRACT	1
ACKNOWLEDGEMENTS	iii
TABLE OF CONTENTS	iv
LIST OF FIGURES	ix
LIST OF TABLES	xi
LIST OF PLATES	xii
1. INTRODUCTION	1
Purpose of Study	1
Location, Access, and Methodology	3
Previous Work	3
Definition of the Kaladar-Dalhousie Trough	5
2. DESCRIPTION OF LITHOLOGIC UNITS	6
Unit A	6
Hermon Group	6
Otter Creek Formation	6
Unit 1	6
Unit 2	9
Unit 3	9
Unit 4	10
Unit 5	10
Todd Lake Formation	11
Unit 6	11
Unit 7	12
Shovel Lake Formation	13
Unit 8	13
Unit 9	14
Unit 10	15

	page
Unit 11	16
Metamorphosed Plutonic Rocks	17
Unit 12	17
Unit 13	18
Unit 14	19
Flinton Group	20
Unit 15	20
Beatty Formation	21
Unit 16	21
Unit 17	23
Bogart Formation	23
Unit 18	23
Unit 19	24
Unit 20	25
Unit 21	26
Unit 22	26
Unit 23	27
Post Flinton	28
Unit 24	28
Unit 25	28
3. GEOMETRY OF THE CLARE RIVER STRUCTURE	32
Introduction	32
Mesoscopic Folds and Fabric Elements	33
Introduction	33
D1 Structures	34
D2 Structures	35
Later Structures	37
Detailed Descriptions of Subareas	37

	page
Introduction,	37
Subarea 1	39
Subarea 2	42
Subarea 3	43
Subarea 4	45
Subarea 5	46
Subarea 6	48
Subarea 7	49
Subarea 8	50
Conclusions	51
4. STRATIGRAPHY AND CORRELATION	55
Introduction	55
The Grenville-Hastings Controversy	55
The Flinton Group	58
Stratigraphy of the Clare River Structure	61
Introduction	61
Hermon Group	65
Otter Creek Formation	65
Todd Lake Formation	65
Shovel Lake Formation	66
Flinton Group Correlatives	67
Beatty Formation	68
Bogart Formation	69
Facing Direction of the Clare River Supracrustals	70
Correlation with the Flinton Group	71
Other Correlations	72
5. STRUCTURAL CORRELATION	76
Introduction	76

	page
Regional Correlation of Structures	76
Pre-D1 Structures	76
D1 Structures.	76
D2 Structures	80
D3 Structures	82
Polyphase versus Continuous Deformation	83
Conclusions	84
6. TIMING OF METAMORPHISM AND DEFORMATION	85
Introduction	85
Mineral Textures in the Clare River Structure	87
Introduction	87
Foliation (S1) Forming Minerals	87
Relations between Porphyroblasts and Foliation	88
D2 Textures	91
Grade of Metamorphism	92
Timing of Metamorphism and Deformation	94
Correlation with Adjacent Areas	94
7. GEOLOGY AND GEOCHEMISTRY OF THE SHOVEL LAKE FORMATION	97
Introduction	97
Distinction of Ortho- and Paragneisses	98
Field Criteria	98
Chemical Criteria	100
Metavolcanic Rocks of the Shovel Lake Formation	105
Field Characteristics	105
Geochemical Classification	108
Conclusions	122
8. TECTONIC EVOLUTION OF THE KALADAR-DALHOUSIE TROUGH	123
Introduction	123

	page
Paleogeographic Models	123
Plate Tectonic Models	125
The Role of Diapirism and the Emplacement of the Plutonic Bodies	127
A Model of Tectonic Evolution	130
9. SUMMARY OF CONCLUSIONS	134
REFERENCES	137
APPENDICES	143
I. Mineralogy of Thin Sections Examined	143
II. Construction of Cross Sections	146
III. Mineral Textures Observed in Thin Sections of the Pelitic Schist (unit 16)	148
IV. Chemical Tests to Distinguish between Ortho- and Paragneisses	156
V. List of 1974 Stations and Corresponding Samples	160
VI. Geology and Geochemistry of the Kaladar Complex	161

LIST OF FIGURES

	page
1. Location of the Central Metasedimentary Belt within the Grenville Province and of the Study Area within the Central Metasedimentary Belt	2
2. Subdivisions of a Part of the Central Metasedimentary Belt	4
3. Geology of the Clare River Structure	in pocket
4. Structural Geology of the Clare River Structure	in pocket
5. Formation of the Prominent L2 Lineation	36
6. Serial Sections through the Clare River Structure	in pocket
7. Maps of the Clare River Structure Contrasting the Interpretation of Burns (1951) and that of the Writer	40
8. Illustration of the Pattern of L2 Lineations Formed on the Divergent Limbs of an F1 Fold	43
9. Section Showing the Hypothetical Relationship of the Mayo and Hermon Groups	57
10. Distribution of the Flinton Group in a Portion of the Kaladar-Dalhousie Trough	59
11. Columnar Sections of the Flinton Group in a Part of the Kaladar-Dalhousie Trough	62
12. Stratigraphic Succession in the Clare River Structure	64
13. Correlation of the Major Structures in a Part of the Kaladar-Dalhousie Trough	77
14. Cross Section across the Flinton Syncline, Kaladar Belt and Clare River Structure	81
15. Hypothetical Examples of Pre-, Syn-, and Post-Tectonic Mineral Growth	86
16. Textural Relationships of the Pelitic Schist	89
17. Mineral Zones in the Kaladar-Dalhousie Trough	100
18. Location Map of Samples of the Shovel Lake Formation Submitted for Chemical Analyses	109
19. Alkali-Silica Plot for the Metavolcanics of the Shovel Lake Formation	111

	page
20. AFM Plot for the Metavolcanics of the Shovel Lake Formation	113
21. Plot of Al_2O_3 Versus Normative Plagioclase for the Metavolcanics of the Shovel Lake Formation	114
22. Plot of SiO_2 Versus FeO^*/MgO for the Metavolcanics of the Shovel Lake Formation	116
23. FeO^* Versus FeO^*/MgO for the Metavolcanics of the Shovel Lake Formation	117
24. Plot of TiO_2 Versus FeO^*/MgO for the Metavolcanics of the Shovel Lake Formation	118
25. Major Element Discriminant Function Plot for the Amphibolites of the Shovel Lake Formation	120
26. Major Element Discriminant Function Plot for the Amphibolites of the Shovel Lake Formation	121
27. Normative Quartz-Albite-Orthoclase Plot for the Intermediate and Felsic Gneisses of the Shovel Lake Formation	157
28. Station Location Map of the Clare River Structure	in pocket
29. Location and Distribution of Mafic Intrusions of the Early Diorite Group	162
30. Plutonic Events in a Part of the Ontario Segment of the Central Metasedimentary Belt	163
31. Geology of the Kaladar Complex	in pocket
33. AFM Plot for the Rocks of the Kaladar Complex	173
34. Alkali-Silica Plot for the Rocks of the Kaladar Complex	174

LIST OF TABLES

	page
1. Rock Units of the Clare River Structure	7
2. Structural History of the Clare River Structure	38
3. General Lithology of Formations of the Hermon and Mayo Groups	56
4. Major Element Analyses of Rocks of the Shovel Lake Formation	101
5. Trace Element Analyses of Rocks of the Shovel Lake Formation	102
6. Major Element Abundances in Volcanic Rocks of Island Arcs	103
7. Trace Element Abundances in Volcanic Rocks of Island Arcs	104
8. Tectonic Evolution of the Kaladar-Dalhousie Trough	133
9. Normative Mineralogy of Rocks of the Shovel Lake Formation	158
10. Results of Tests to Distinguish between Ortho- and Paragneisses	159
11. Major Element Analyses of Rocks of the Kaladar Complex	170
12. Trace Element Analyses of Rocks of the Kaladar Complex	171
13. Rock Units Corresponding to Chemical Analyses Listed in Tables 11 and 12	172

LIST OF PLATES

	page
1. Aerial Photograph of the Southwestern Margin of the Clare River Structure	29
2. Relict Textures of the Shovel Lake Formation	30
3. Primary Features Preserved in the Metasediments of the Flinton Group in the Clare River Structure	31
4. Minor Structures in the Clare River Structure	53
5. Minor Structures in the Clare River Structure	54
6. Metamorphic Textures of the Pelitic Schist	96
7. Field Photographs of the Kaladar Complex	184

1

1. INTRODUCTION

Purpose of Study

The purpose of this study is to further the understanding of the structural and stratigraphic evolution of the section of the Central Metasedimentary Belt of the Grenville Province known as the Kaladar-Dalhousie Trough (Figures 1 and 2). The Clare River Structure¹ is one of the major supracrustal belts in the Kaladar-Dalhousie Trough and was chosen as a study area because it lies adjacent to other supracrustal belts recently mapped in detail by Thompson (1972), Moore and Thompson (1972), Psutka (1976), and Rivers (1976). The specific aims in undertaking the study were:

- 1) To determine the structural geometry and stratigraphic succession of the Clare River Structure.
- 2) To determine the nature of the contact between the Clare River supracrustal rocks and the adjacent granite gneisses.
- 3) To evaluate the significance of two conglomerate horizons that outcrop within the Clare River Structure.
- 4) To discover the relative timing of metamorphism and deformation.
- 5) To correlate structure and stratigraphy of the Clare River Structure with the structure and stratigraphy of adjacent areas.
- 6) To evaluate the respective roles of gravity and lateral compression in the evolution of the structure of the Kaladar-Dalhousie Trough.

1. Historically, the Clare River Structure was referred to as the Clare River Syncline (Wilson, 1933; Burns, 1951) and more recently, as the Clare River Synform (Moore and Thompson, 1972). In this thesis the term Clare River Synform is used to describe a second phase fold, which together with the Hungerford Antiform, make up the Clare River Structure. The term Clare River Syncline is not used.

Figure 1. Location of the Central Metasedimentary Belt within the Grenville Province and of the study area within the Central Metasedimentary Belt. Boundary of the Central Metasedimentary Belt after Wynne-Edwards, 1972.

Location, Access, and Methodology

The Clare River Structure is situated in southeastern Ontario on the southwestern boundary of the Grenville Province (Figure 1). This thesis is based on the detailed mapping of the southwestern portion of the Clare River Structure from Highway 41 to the closure at Stoco Lake (Figures 2, 3).

With the exception of the extreme southwestern portion, the rocks in the map area are well exposed and access is facilitated by two major highways (7 and 41) and a network of gravel roads. Geological mapping was carried out on a scale of an inch to a quarter mile. In the southwestern portion of the map area, which is mainly open farming country, outcrops were easily located and plotted directly on air photo overlays in the field. In the northwestern portion of the area, which is heavily wooded, it was necessary to run compass traverses across strike at about 0.5 km intervals. All data are plotted on a quarter mile (1:15,840) base map that was enlarged from 1:50,000 scale federal topographic maps 31C/11 and 31C/6.

Previous Work

Maps of the southwestern portion of the Clare River Structure have been produced by Wilson (1933, 1940) and Burns (1951), and the north-eastern section of the writer's map area is included in maps of Kaladar and Kennebec Townships by Harding (1944) and Wolff (1978). Other workers (Church, 1954; Pearson, 1958; Venkitasubramanyan, 1969; Cermignani, 1971; Currie, 1972; and Schwerdtner et al., 1977) have studied various aspects of the geology of the Clare River Structure and students in Carleton University's advanced field course (1973-1977) have mapped small areas of it adjacent Highway 41. These studies will be discussed where they bear on the arguments presented in this thesis.

Figure 2. Subdivisions of a part of the Central Metasedimentary Belt, modified after Wynne-Edwards (1972), and Hewitt (1956). Geology simplified after MacDonald (1970). Detailed map areas, outlined in the Kaladar-Dalhousie Trough, are as follows: A) this study; B) Rivers, 1976; C) Psutka, 1976; D) Thompson, 1972.

Definition of the Kaladar-Dalhousie Trough

Hewitt (1956) first defined the Kaladar-Dalhousie Trough which he distinguished from the Hastings Basin to the west and the Frontenac Axis to the east. Wynne-Edwards, in his compilation of 1972, redefined the Hastings Basin to include the Hastings Basin and the Kaladar-Dalhousie Trough of Hewitt. In this thesis, Wynne-Edwards' subdivision of the Central Metasedimentary Belt of the Grenville Province is modified to include the Kaladar-Dalhousie Trough (Figure 2).

The reintroduction of the Kaladar-Dalhousie Trough as a subdivision of the Central Metasedimentary Belt is justified because its map pattern differs from that of adjacent areas. It comprises a series of northeast trending supracrustal belts that are separated by metaplutonic bodies (mainly granodiorite). The Elzevir Batholith and Weslemkoon Pluton mark its western boundary, the Hinchinbrooke Pluton¹ its eastern boundary.

The Hastings Basin, which is situated west of the Elzevir Batholith, is characterized by a weak north-south structural grain that is defined by belts of clastic sediments separated by large irregular areas of marble. These supracrustal rocks wrap around small subcircular (5 to 10 km) plutonic bodies. The Frontenac Axis, situated east of the Hinchinbrooke Pluton, is characterized by a pronounced northeast structural grain. Its plutons and supracrustal belts are smaller than those in the Kaladar-Dalhousie Trough.

1. In the literature, several of these bodies have been referred to as gneisses (e.g. Hinchinbrooke Gneiss, Cross Lake Gneiss, Addington Gneiss). There is little doubt that they are metamorphosed igneous intrusions and are therefore referred to as plutons in this work.

2. DESCRIPTION OF LITHOLOGIC UNITS

In this chapter, the rock units of the Clare River Structure will be described in the order of their occurrence in the legend of Figure 3, presented here as Table 1. Unless otherwise noted, all localities referred to are shown on Figure 3. The justification of this order and the arguments for stratigraphic correlation with adjacent areas are presented in Chapter 4, and the mineralogy, based on thin section examination of most of the units is tabulated in Appendix I.

Unit A outcrops as a uniformly thin septum (70 to 90 m) that separates the Mellon Lake and Addington Plutons along the southeast margin of the Clare River Structure. It was previously mapped as being continuous with marbles (units 6 and 18) exposed within the Structure (Wilson, 1933; Burns, 1951), but there is no outcrop in the critical area to support this. Aerial photographs (Plate 1) support the interpretation of the writer, presented in Figure 3.

Unit A consists of calc-silicate gneisses, hornblende-biotite gneisses, and marbles, all of which are cut by white, coarse grained microcline-plagioclase-quartz pegmatite. The pegmatite is ubiquitous and occurs on all scales, from irregular bodies occupying entire outcrops to concordant lenses 1 to 2 cm across.

Hermon Group

Otter Creek Formation

Unit 1. The massive felsic metavolcanics which constitute unit 1 underlie an area of high, sparsely-vegetated ground which extends from Highway 41 southwest for about 3 km. They consist of a fine grained mosaic of quartz and feldspar with minor biotite and muscovite. Outcrops

Table 1. Rock units of the Clare River Structure

Unit 25: Paleozoic Sandstone

Unconformity

GRENVILLE SUPERGROUP AND ASSOCIATED PLUTONIC ROCKS

Unit 24: Pegmatite

Intrusive Contact

FLINTON GROUP

Bogart Formation

Unit 23: Calc-silicate paragneiss and pegmatite

Unit 22: Uniform calc-silicate paragneiss

Unit 21: Dolomitic marble, rusty paragneiss

Unit 20: Calc-silicate paragneiss with quartzofeldspathic stringers

Unit 19: Rhythmically layered calc-silicate paragneiss

Unit 18: Calcareous marble

Beatty Formation

Unit 17: Rusty graphitic psammite

Unit 16: Pelitic schist

Unit 15: Granitic clast metaconglomerate

Unconformity

METAMORPHOSED PLUTONIC ROCKS

Unit 14: Leucocratic quartz monzonite

Unit 13: Biotite granodiorite

Intrusive Contact

Unit 12: Amphibolite dykes

Intrusive Contact

HERMON GROUP

Shovel Lake Formation

Unit 11: Biotite schist and phyllite, calc-silicate paragneiss

Unit 10: Felsic metavolcanics: layered, fragmental and massive, quartzofeldspathic gneisses; biotite gneiss; colour index 10

Unit 9: Intermediate metavolcanics: biotite-hornblende gneiss, hornblende-biotite gneiss, hornblende gneiss, generally fragmental, locally porphyritic; colour index: 10-35

Unit 8: Mafic metavolcanics: massive fine grained amphibolite, locally pillowed, coarse grained amphibolite, minor fragmental amphibolite

Table 1, cont'd.

Todd Lake Formation

Unit 7: Dolomitic marble, calc-silicate paragneiss, para-amphibolite, metachert; 7a) volcanic pebble metaconglomerate

Unit 6: Dolomitic marble

Otter Creek Formation

Unit 5: Calcareous marble

Unit 4: Rusty graphitic psammite and pelite

Unit 3: Quartzofeldspathic psammite, pelitic schist

Unit 2: Biotite-hornblende gneiss, hornblende-biotite gneiss, minor amphibolite

Unit 1: Massive felsic metavolcanics

Unit A: Marble, calc-silicate paragneiss, pegmatite

weather pink and display smooth rounded surfaces although, in some outcrops, ridging and furrowing indicative of layering occur. At one locality, wispy, fine (1 cm) layering defined by faint colour differences is interpreted as relict flow-banding.

Unit 2. Biotite-hornblende gneisses and hornblende-biotite gneisses constitute most of unit 2, which is characterized by a pale-orange weathered surface (due to the alteration of disseminated biotite flakes). The leucocratic phases of the gneisses consist of feldspar and subordinate quartz; black and green amphiboles, biotite, and chlorite, in varying proportions, constitute the mafic phases. Garnet, pyrite, calcite, and muscovite occur as accessory minerals. Garnet amphibolite and coarse brown-red weathering schists consisting of biotite and chlorite commonly occur as layers within the gneisses.

Compositional layering within the gneisses themselves is conspicuously absent on the outcrop scale - they display discontinuously ridged and pitted weathered surfaces or surfaces on which areas from 1 cm to 20 cm across stand out in relief. These surface textures reflect the fragmental nature of unit 2. This observation, the lack of compositional layering, and the overall homogeneity of the unit suggest that it is mainly of pyroclastic origin.

Unit 3 outcrops on the northwest and southeast limbs of the Clare River Synform (Figure 4) from Highway 41 to where they join, 2 km southeast of Otter Creek. Its exposure width averages 50 m over a strike length of 10 km. Massive, fine to medium grained quartzofeldspathic psammitic, consisting of quartz and feldspar with subordinate biotite and muscovite composes the majority of unit 3. Locally, calcite, sillimanite and/or garnet are present. It usually forms gray, smooth-weathering outcrops, although locally, sillimanite knots constitute a large portion of

the rock, which results in a rough textured surface. Pelitic layers make up about 10 percent of unit 3 and differ from the psammite only in having a higher proportion of muscovite.

Unit 4 differs from unit 3 in that it contains disseminated sulphides and graphite. It is characterized in the field by its rusty weathering and is occasionally stained dark orange or black. The psammite is fine grained and consists mainly of quartz and feldspar with subordinate muscovite, graphite, garnet, pyrite, and sillimanite. It possesses a marked closely-spaced fracture cleavage and some samples are friable.

The psammite is locally interbedded with rusty pelitic layers which are fissile and display penetrative schistosity. These layers are generally 1 to 3 cm thick and, locally, pelitic horizons consisting of a number of such beds occur. They consist mainly of muscovite, quartz, and feldspar, and commonly contain sillimanite knots and garnet porphyroblasts.

Unit 5 consists of coarse grained calcareous marble that is composed essentially of calcite, with traces of quartz, muscovite, and graphite. It displays a granoblastic texture, weathers white to light gray, and crumbles easily. Compositional layers, 1 to 2 cm thick, are locally preserved, and are defined by alternating gray and white bands or by quartz-rich layers that stand out in relief.

Unit 5 is greatly thickened in the core of the Clare River Synform. On the northwest limb, it averages 60 m in thickness, and can be traced for about 2 km, until it is lost in the low swampy ground surrounding Otter Creek. On the southeast limb, poor exposure coupled with the marked attenuation of all rock units make it impossible to separate unit 5 from the adjacent dolomitic marbles of unit 6.

At their contact, units 4 and 5 are intercalated in beds of 2 to 20 cm thickness, and in that region the greatest concentration of

minor folds and interference patterns in the entire structure are exposed (Plate 5, c and d).

Todd Lake Formation

Unit 6 is exposed in three separate areas and the recognition of this fact aided considerably in the solution of the overall geometry of the Clare River Structure.

The major exposure crops out on the limbs of the Clare River Synform, and extends from Highway 41 to the core, about 2 km east of Otter Creek. On the northwest limb, the outcrop width (about 500 m) is exaggerated by F2 parasitic folds. On the southeast limb, unit 6 is severely attenuated, but can be traced continuously from the core area northeast along the north shores of Bruno, Little Mitten, the Mitten Lakes back to Highway 41. This feature was not formerly recognized. The second outcrop area begins 5 km northeast of Hungerford and extends to Highway 41 at Kaladar. The third exposure area occurs 1 km east of Bogart. It was formerly mapped with the calcareous marble of unit 18 (Wilson, 1933; Burns, 1951) and the increase in exposure width was ascribed to flow during folding (Wilson, 1933).

Two main varieties of dolomitic marble are interbedded in unit 6. The first is a uniform fine to medium grained rock with a granoblastic texture, and consists essentially of dolomite and subordinate tremolite, with traces of quartz, feldspar, muscovite, and scapolite.

The second variety is characterized by knots and discontinuous lenses of calc-silicate which give the weathered surface of the outcrop a jagged aspect. These knots are composed of diopside and tremolite and are locally altered to talc and serpentine. They are the result of metamorphism and tectonism of competent siliceous-dolomite layers in the less

competent pure marbles (Plate 5a, 5b). The calc-silicate knots form a chaotic pattern in the vicinity of mafic dykes, which probably results from the reaction of irregular quartz veins with the dolomite during metamorphism.

In addition to the two varieties of dolomitic marble, thin lenses of calcareous marble occur locally in unit 6.

Unit 7. The contact between units 6 and 7 is arbitrarily drawn where uniform dolomitic marble outcrops (unit 6) give way to alternating outcrops of metasediments and marble (unit 7). Unit 7 has an exposure width of 180 to 275 m on the northwest limb of the Clare River Synform and extends from Highway 41 to where it abruptly pinches out about 6 km along strike. It is not encountered on the southwestern limb.

In addition to dolomitic marble, unit 7 includes polymictic metaconglomerate (7a), white weathering finely banded quartzite (metachert?), well layered para-amphibolite in which the amphibole occurs as dark gray radiating clusters in the cleavage planes, fine grained biotite phyllites with clots consisting of fine biotite (biotite eyes), hard dark-gray feldspathic psammite with varying amounts of hornblende and biotite, and calcareous epidote-bearing calc-silicate paragneisses. Individual horizons are thin (2 to 10 m) and the area is poorly exposed, making it impossible to trace them for any distance along strike.

Polymictic conglomerate (7a) is exposed at several different levels in the unit. Its most common pebble type (50 percent) is a fine grained pink-weathering leucocratic gneiss which is similar in appearance to the rhyolite of the Shovel Lake Formation. Milky-white vein quartz pebbles constitute another 30 percent of the pebble population and the remainder consist of marble, amphibolite, biotite schist, and calc-silicate paragneiss. The matrix weathers greenish-gray and consists of

epidote, diopside, biotite, carbonate, quartz, and feldspars.

The pebbles are highly deformed. A rough visual estimate puts their axial ratios at 1 to 5 to 10, with the long axis parallel to F2 fold axes. While no obviously angular pebbles were observed, sorting is rather poor; pebbles range in size from 1 by 4 to 5 by 20 cm measured on a surface perpendicular to foliation.

Shovel Lake Formation

One metasedimentary and three metavolcanic units constitute the Shovel Lake Formation. The metavolcanics are differentiated on the basis of their colour index as follows: felsic: colour index between 0 and 10; intermediate: colour index between 10 and 30; and mafic: colour index greater than 30.

A stratigraphic order for the metavolcanics could not be established; the three units outcrop as lateral equivalents over the entire exposure area of the Formation. The metasediments, on the other hand, overlie the metavolcanics.

Unit 8. The mafic metavolcanic rocks which comprise unit 8 are all metamorphosed to amphibolites. The largest body of unit 8 extends from the north end of Stoco Lake to where it tapers out 1 km northwest of Hungerford. Numerous other amphibolite bodies occur throughout the map area. They are discontinuous and lack a clearly defined stratigraphic position within the Formation. While it is probable that the complex pattern displayed by the amphibolite bodies is in part due to folding, there is reason to believe that it is in part a primary feature, as indicated by the discontinuity of the units of the Shovel Lake Formation as compared with the continuity of units of the metasedimentary formations.

The amphibolites weather dark gray or black and two types can be

distinguished on the basis of their field characteristics. The first displays smooth weathered surfaces that are locally ropey, and contains irregular epidote-rich clots and lenses. Pillows were observed in several outcrops of this member near the village of Sulphide, but unfortunately, only one pillow from which a top could be inferred was found (Plate 2a). The danger of using one facing determination to interpret major structures was pointed out by M. E. Wilson (1956, p. 53) as follows:

"The pillow like forms in metamorphosed volcanic rocks at Sulphide..... Only three bun-like forms possibly indicate tops; two to the north and one to the south were seen. These can scarcely be used with certainty to determine tops."

Other primary features observed in these amphibolites around Sulphide include relict laths of plagioclase, and amygdules.

The flows in this unit consist essentially of hornblende and plagioclase with minor quartz, epidote, sphene, and biotite. Locally scapolite is abundant. Grain size varies from fine to very coarse. In the coarse grained varieties, a clotty texture in which aggregates of essentially pure hornblende occur in matrix composed of plagioclase and hornblende is interpreted as a relict cumulate texture.

The second type of amphibolite, which constitutes a subordinate part of unit 8, is gradational with the hornblende gneiss of unit 9. The best exposures occur near the village of Sulphide. They display rough weathered surfaces characterized by discontinuous ridges and grooves, which are the result of differential weathering between fist-sized andesitic fragments and more mafic matrix.

Unit 9 comprises a variety of gneisses which have a colour index ranging from 10 to 30. They are generally fine grained and display mosaic textures in thin section. Plagioclase, subordinate quartz and, occasionally, microcline, make up the leucocratic portion of the gneisses;

biotite and hornblende constitute the mafic phases. Biotite predominates over hornblende in the more felsic gneisses and the reverse holds true for the mafic members. Apatite, epidote, muscovite, calcite, ilmenite, magnetite, and sphene are common accessory minerals.

Most of the gneisses of unit 9 display monotonous, alternating light and dark layers. Within layers, discontinuous lenses occur and these are interpreted as highly flattened fragments within pyroclastic deposits (Plate 2e). Outcrops in which layering is continuous and obvious fragments are lacking make up a small portion of unit 9. They may be uniform pyroclastics or epiclastic sediments: they cannot be distinguished in the field because most primary features have been destroyed by metamorphism. Chemical evidence (Chapter 7) suggests that epiclastic sediments do compose a part of unit 9.

Thick (greater than 2 m), uniform, hornblende gneisses, interpreted as meta-andesite flows, make up a small portion of unit 9. One such flow, characterized by relict aligned plagioclase laths, 1 to 4 cm long (Plate 2b), can be traced for over 1 km in the core area of the Clare River Synform east of Otter Creek.

Outcrops in which well preserved fragmental textures occur are abundant throughout unit 9. They display a variety of sizes and compositions, such as fist-sized mafic fragments in an intermediate matrix, large felsic fragments in an intermediate matrix (Plate 2d), and small intermediate fragments in a felsic matrix (Plate 2c).

Unit 10 comprises quartzofeldspathic gneisses, derived from felsic volcanics, with a colour index of less than 10. Mappable bodies of unit 10 are discontinuous and locally terminate abruptly. In addition, lenses of unit 10 too small to be represented on Figure 3 occur in outcrops of unit 9. Two varieties of felsic volcanics can be distinguished in the field.

The first is massive, forms rounded outcrops with smooth pink weathered surfaces and has a faint foliation. Locally, flow layering is apparent. The rock is fine grained, displays a mosaic texture, and is composed essentially of quartz, plagioclase, and microcline with traces of muscovite. Mafic minerals are absent. This variety of unit 10 is interpreted as metamorphosed rhyolite flows.

The second type is characterized by ridged and furrowed to nodular outcrop surfaces, reflecting the layered and fragmental nature of this member. It is slightly more mafic than the massive variety, generally containing a small percentage of biotite, and garnet is locally present. Primary features can be identified locally and include: layers which contain an abundance of fine equant pink feldspar crystals (crystal tuff?), discontinuous layers which are interpreted as highly flattened fragments in a matrix of similar composition, and streaky discontinuous dark layers in pink layered rocks (fiamme in ash flow tuffs?). These textures indicate the pyroclastic nature of this variety of unit 10.

Unit 11 comprises mappable bodies of metasediments within the Shovel Lake Formation. The main exposure begins at Shovel Lake and extends as a lens with a maximum thickness of about 300 m to Highway 41, a distance of 4.7 km. This exposure is high in the succession of the Shovel Lake Formation, and its shape suggests that it is the transposed core of an F1 fold. In addition, several thin lenses of unit 11 occur between the volcanics of the Shovel Lake Formation and the overlying Flinton Group in the area east of Hungerford.

As in the case of the metasediments of unit 7, unit 11 is heterogeneous, consisting of calc-silicate paragneisses, calcareous biotite schists and phyllites, and quartzofeldspathic psammites.

The calc-silicate gneisses are gradational with the gneisses of unit 9. They display continuous compositional layering and a pitted weathered surface due to recessive weathering of carbonate rich lenses. Microcline, quartz, and plagioclase comprise the leucocratic phases, actinolite, biotite, and epidote the mafic phases. Sphene, calcite, and opaques are common accessories. These calc-silicate gneisses are particularly common in exposures of unit 11 east of Hungerford.

The biotite schists and phyllites are fine grained, uniform, dark-gray weathering rocks consisting mainly of disseminated biotite, quartz, and feldspar, with accessory carbonate and epidote. This rock commonly contains eyes of biotite, which are lens-shaped areas composed of fine biotite flakes. Layers of biotite schist displaying these eyes are found in most exposures of unit 11.

The feldspathic psammites display continuous compositional layering which is in the order of 5 to 10 cm thick. Samples are hard, granular, dark gray aggregates of plagioclase and quartz with subordinate biotite and amphibole. The psammite is far less abundant than the other members mentioned above.

Metamorphosed Plutonic Rocks

Unit 12. The amphibolite dykes that comprise unit 12 are more widely distributed than indicated on Figure 3. They are common in the marbles of the Todd Lake Formation, but are too small to be represented on a geological map of the scale of Figure 3. In the Shovel Lake volcanics, the writer was not able to separate amphibolite dykes from flows as primary textures are destroyed and any bodies originally cross-cutting the volcanics could have been rotated into parallelism with the stratigraphy. It is important to note that amphibolite dykes are not observed to cut the

Flinton Group.

The following discussion is concerned with the large mafic dyke which cuts units 2 and 3, and is spatially associated with their contact. It is nowhere wider than 60 m and is generally about 30 m wide. Locally, this dyke alternates with unit 3 on a scale of metres. It is unclear whether this is a feature related to its emplacement, or the result of isoclinal folding. Although the dyke cuts units of the Otter Creek Formation, it is folded with them and is clearly pre-tectonic.

The dyke is generally a uniform medium-grained amphibolite which possesses a strong linear fabric and a less pronounced planar fabric. It consists mainly of hornblende and plagioclase with accessory chlorite, opaques, quartz, and apatite. Locally, relict igneous textures, in which blocky aggregates of hornblende (5 mm to 2 cm across) occur in a matrix of hornblende and plagioclase, are preserved. The hornblende aggregates appear to be pseudomorphic after pyroxene.

Zones of strain gradation commonly occur in unit 12. In these, the rock changes from an undeformed blocky aggregate of hornblende and plagioclase to a fine-grained streaky amphibolite over less than a metre. The dyke is also observed to be more highly deformed near its contacts.

Locally an ultramafic phase of unit 12 occurs. It is a dark green to black hornblendite, composed of ortho- and clino-amphiboles with accessory chlorite, spinel, opaques, and occasionally, garnet.

Unit 13. The Mellon Lake Pluton occurs along the southeastern boundary of the map area and is separated from the supracrustal rocks of the Clare River Structure by unit A and unit 14.

Within the map area the Mellon Lake Pluton is a medium to coarse grained migmatite of granodioritic composition consisting of about 10 percent pink quartz-microcline-plagioclase mobile phase and 90 percent light

gray biotite-plagioclase-quartz residual phase. The strong foliation of the rock is defined by the alignment of biotite flakes and to a lesser extent by quartz and plagioclase, paralleling the migmatitic layering. Locally many refolded folds of the mobile layers about the same axis are present in a single outcrop.

The Mellon Lake Pluton can be grouped with the biotite-diorite series as defined by Lumbers (1967a), believed to have formed 1250 ± 25 m.y. ago (Silver and Lumbers, 1966). There is no direct evidence of its relationship to the supracrustal rocks of the Clare River Structure, but on the basis of its correlation with the rocks of the biotite-diorite series, it both predates and postdates different units in the Clare River Structure.

Unit 14 is a uniform, pink weathering quartz monzonite composed essentially of quartz, plagioclase, and microcline, with accessory magnetite. Locally it contains a few percent of biotite. On smooth surfaces, it displays S1 foliation defined by tabular areas of quartz (1 cm by 1 cm by 2 mm) and to a lesser extent, by feldspar. These are believed to be highly flattened, recrystallized quartz and feldspar crystals which composed the undeformed quartz monzonite. Locally, trains of magnetite grains parallel this foliation.

In some exposures of unit 14 in the vicinity of Stoco Lake, a strong linear fabric is developed. Schwerdtner et. al. (1977) interpret this lineation as being coeval with the foliation described above, and they ascribe the variation between lineated and foliated gneisses to finite strain variation. The writer and others (R. L. Brown, J. M. Moore, pers. comm., 1976) examined several of the outcrops in question, and formed the opinion that this lineation is the result of the crenulation of a pre-existing planar fabric. This observation is consistent with the sequence of fabric elements throughout the Clare River Structure (Chapter 3).

Unit 14 structurally underlies the supracrustal rocks of the Clare River Structure: the contact is interpreted as a decollement (Chapter 3). The body of unit 14 on the northwest margin of the structure is known as the Addington Pluton. It is continuous with the lens of unit 14 exposed along the entire southeast margin of the structure.

There is no direct evidence to establish the age relationship between the Addington and the Mellon Lake Plutons. If it is accepted that the Mellon Lake Pluton is a member of the biotite-diorite series of Lumbers (1976b), and that the Addington Pluton belongs to the quartz monzonite group (Lumbers, 1976b), then the Addington Pluton is the younger. On a local scale, Psutka (1976) reports that dykes similar in composition to the Addington Pluton cut outcrops of the Northbrook Pluton. The Mellon Lake Pluton is similar in composition to the Northbrook Pluton, and if they are equivalent in age these data further support the contention that the Addington Pluton postdates the Mellon Lake Pluton.

The relative age of the Addington Pluton with respect to the supracrustal rocks of the Clare River Structure is not directly established. Amphibolite bodies are enclosed in the Addington Pluton near Stoco Lake, but it is not clear whether these are dykes, xenoliths, or infolded lenses. On the basis of stratigraphic correlations discussed in Chapter 4, the supracrustal rocks must in part postdate, and in part predate the Addington Pluton.

Flinton Group

Unit 15. 2.5 km northeast of Stoco Lake, a small lens-shaped area of granite clast conglomerate (unit 15) occurs at the contact of the Addington Pluton and the supracrustals of the Clare River Structure. Its maximum width is about 60 m, and it can be traced along strike for only

about 600 m. Along strike, unit 15 thins and the pebbles become so highly deformed that it is impossible to distinguish them from gneissic layers. The conglomerate was recorded only where distinct clasts could be made out.

The clasts, viewed on a two dimensional surface, vary in size from about 30 by 40 cm to 2.5 by 3.0 cm. A higher proportion of large pebbles occurs in exposures that are adjacent the Addington Pluton. The pebbles are angular to moderately rounded and poorly sorted. They are mainly white to pink-weathering and are similar in appearance to the Addington Pluton. Many of them contain magnetite. Less common are gray weathering biotite gneiss pebbles. The pebbles float in a gray to pink matrix that consists of biotite, plagioclase, quartz, and microcline. In outcrops immediately adjacent to the Addington Pluton, the matrix is more quartzofeldspathic than in the outcrops further away from it. The contact between unit 15 and the Shovel Lake volcanics is not exposed.

Beatty Formation

Unit 16 is a spectacular porphyroblastic pelitic schist that serves as a marker horizon throughout the entire map area. Its outcrop pattern clearly defines a type III (Ramsay, 1967 p. 530) interference pattern between F1 and F2 folds (Chapter 3). It begins south of Bruno Lake, follows along Otter Creek, then thickens in the core of the Clare River Synform just east of the crossroads of Otter Creek. From there, it can be traced northeast for about 9 km to the core of the Hungerford Antiform, then down the north limb where it gradually thins to only a few metres and finally lenses out south of Sulphide. Half a km east of the Hungerford Antiform a lens of pelitic schist representing another F1 fold core is exposed along a strike length of about 5 km (Figure 4).

Compositional layering is locally preserved in the pelitic schist

and individual layers range in thickness from 1 or 2 cm (Plate 4c) up to 30 to 40 cm. Layers vary both in texture and composition. Highly aluminous layers, which weather light gray and consist mainly of muscovite with subordinate quartz and kyanite, occur along with dark gray iron-rich layers consisting of quartz, plagioclase, biotite, staurolite, and garnet, with only traces of muscovite. All gradations between these extremes exist. Semi-pelite layers, made up essentially of quartz and plagioclase with subordinate garnet, biotite, and muscovite constitute a minor part of unit 16.

Layers containing porphyroblasts of magnetite occur adjacent to layers containing porphyroblasts of hematite, suggesting that the layers vary in oxidation state. Where this feature is observed, the light gray micaceous layers contain the hematite, while the darker layers rich in biotite, garnet, and staurolite contain the magnetite.

Proceeding northeast to southwest, changes in the metamorphic grade alter the appearance of the pelite. In the northeast, the micaceous layers contain large oligoclase porphyroblasts, locally abundant kyanite, and bits of fibrous sillimanite. The assemblage in the biotite-rich layers is staurolite, garnet, biotite, muscovite, quartz, and plagioclase. The staurolite overgrows the S1 foliation, and crystals are often several cm in length. Kyanite crystals, up to 5 cm long, have a preferred linear orientation.

Toward the southwest, highly micaceous layers become rare, and the rock takes on a more gneissic appearance. Large kyanite, oligoclase, and staurolite crystals disappear (although finer grained varieties of staurolite and plagioclase are present) while sillimanite becomes abundant, taking the form of sillimanite-quartz-plagioclase knots. The large oligoclase crystals break down by altering to fine micas and it is on these micas that sillimanite commonly nucleates. On the southeast limb of the

Clare River Structure, the assemblage observed in the field is sillimanite, biotite, muscovite, quartz, and plagioclase. The sillimanite occurs as fibrous mats.

Unit 17 comprises a fine grained rusty psammite which has a maximum exposure width of 30 m. On the north limb of the Hungerford Antiform, it is only 3 - 6 m wide. In the core of the Hungerford Antiform, considerable thickening has occurred. In the core of the Clare River Synform at Otter Creek, numerous unmappable lenses of rusty psammite are present in the marble, producing a far more intricate map pattern than is possible to represent on the scale of Figure 3.

The rock displays a rusty to orange-black weathered surface, and is fissile. Occasionally it is calcareous, and locally hard and massive. The essential mineralogy is muscovite, graphite, pyrite, quartz and feldspar.

Southwest of Sulphide, the underlying pelitic schist (unit 17) lenses out and, at one point, rusty lenses associated with pyroclastic rocks of the Shovel Lake Formation are directly overlain by rusty psammite which in turn is overlain by Flinton Group calcareous marble (unit 18) of the Bogart Formation.

Bogart Formation.

Unit 18 is the third of four distinct marble units of the Clare River Structure. It is exposed mainly in the core of the Clare River Synform near the old schoolhouse at Otter Creek. The intricate pattern it displays there is the result of inhomogeneous flow and the interference between F1 and F2 folds. From Otter Creek, it can be traced northeast to the core of the Hungerford Antiform, and then back southwest around the nose of the Clare River Synform 1 km west of Bogart. Its apparent thick-

ness is only 30 to 60 m on the fold limbs. South of Otter Creek, on the southeast limb of the Clare River Synform, it is lost in an area of pegmatite where thin slivers of calcareous marble, along with the other rock types of the Bogart Formation occur as individual lenses in large bodies of pegmatite (unit 23).

On older maps of the Clare River Structure (Wilson, 1933; Burns, 1951; and Pearson, 1958) unit 18 is shown to thicken considerably where it crosses the concession road about 1 km west of Bogart. At this point, one finds a few small outcrops of calcareous marble (unit 18), then 250 m west of these, several large outcrops of dolomitic marble with diopside and tremolite knots. At one point they are separated by an outcrop of unit 10, illustrating that they are not the same horizon. The dolomitic marble is interpreted as belonging to the Todd Lake Formation (unit 6).

Unit 18 displays a white or pale brown weathered surface and consists essentially of a medium grained granoblastic mosaic of calcite grains with accessory muscovite or phlogopite, graphite, and quartz. Compositional layering is expressed on clean outcrops as faint white and gray colour banding which varies from less than 1 cm to 5 or 10 cm in thickness. It is accentuated by interbedded layers of rusty paragneiss, which stand out in relief, and which form boudins in the less competent marble.

Unit 19 consists of rhythmically layered calc-silicate paragneiss. The best exposures occur in the core of the Clare River Synform east of Otter Creek where the apparent thickening is due to the fold axis going through the horizontal. It is also thickened in the core of the F1 Hungerford Antiform where complete transposition of compositional layering into the axial plane of F1 folds can be observed.

Outcrops of unit 19 are ridged and furrowed, reflecting its layered nature. The layers are 2 to 3 mm thick and alternate between a

biotite-rich and a biotite-poor composition. The biotite-rich layers display a mauve tint (because of alteration of the biotite) and the other layers are pale green, due to the abundance of diopside. Unit 19 is calcareous and tends to be friable. The essential mineralogy is quartz, diopside, phlogopite or biotite, microcline and calcite.

Unit 20. The characteristic features of unit 20 are its dark gray colour, rhythmic compositional layering, and the ubiquitous presence of discontinuous lenticular quartzofeldspathic lenses, which generally define the axial planes of minor folds. Locally, several generations of these lenses are intricately folded into confusing patterns.

Unit 20 is fine grained and has a colour index of between 20 and 30. The essential mafic constituents are hornblende and biotite, with a smaller amount of diopside; the felsic minerals are plagioclase, microcline, and subordinate quartz. Epidote and actinolite are present locally and calcite is ubiquitous. The texture comprises a granoblastic aggregate of microcline and plagioclase, with fine biotite flakes aligned parallel to the compositional layering. Locally, clots of biotite and hornblende, measuring 1 cm by 2 to 4 mm, are oriented at a high angle to the foliation.

Primary sedimentary layering is well preserved in outcrops southeast of Bogart. The layers are rhythmic, 2 to 10 cm thick, and display colour grading (Plate 3b, c). The colour grading in the layers is thought to reflect primary graded bedding, even though recrystallization has destroyed any hint of grain size gradation. It is reasoned that the darker portion of the layers (now rich in hornblende and biotite) were originally rich in clays, and thus indicate the upper portion of the beds. Southeast of Bogart, these layers face north.

Immediately west of Bogart, features which resemble flame structures (Plate 3a) give facing directions which conflict with those indicated

by the graded bedding. It is not apparent whether this is because the "flame structures" are tectonic in origin and therefore, unreliable as top indicators, or because minor structural complications result in facing reversals.

Northwest of Bogart, compositional layering is transposed into parallelism with the quartzofeldspathic segregations, resulting in an S0/S1 compositional surface in which primary features are no longer evident. These S0/S1 composite layers differ from S0 in being uniform in colour and continuous for only a few metres.

Unit 21 is poorly exposed; it underlies swampy ground which separates outcrops of units 20 and 22 over most of its exposure area. It is composed of a white to brown-orange weathering fine grained marble, with a granular texture, consisting mainly of dolomite with minor tremolite and muscovite. The tremolite is widely dispersed, occurring in small knots which have no preferred orientation. Compositional layering in the rock is defined by faint colour streaking, and by the more resistant layers standing out in relief. Planes of mica flakes also define compositional layering and locally, these flakes have been folded by F2 folds. The lower part of this unit comprises a rusty, locally calcareous paragneiss and schist. It is exposed in only three localities and is only a few metres thick.

Unit 22 is exposed mainly in the core of the Hungerford Antiform where it underlies cleared rolling fields about 1 km southwest of the village of Sulphide, and on the southeastern margin of the Clare River Structure, adjacent to Addington Pluton.

Its characteristic field features are a uniform gray weathered surface, rounded outcrops, and a lack of consistently expressed surface textures, such as ridging or colour banding. Foliation is expressed.

variably by widely spaced fractures, discontinuous ridging, and alignment of calcite pods. Samples of unit 22 display a light gray colour with a marked greenish tint due to diopside grains, and have a granular texture.

A brownish weathering rind about 1 cm thick is also characteristic.

Another distinctive field feature is the presence of lenses and pods of coarse pink calcite, which measure from 1 by 3 to 5 by 10 cm. The essential mineralogy is diopside, calcite, quartz, and feldspar with traces of biotite, hornblende, and epidote.

S0 is commonly preserved in unit 22. Southwest of Sulphide, the beds are in the order of 20 cm to 1 m in thickness and are non-rhythmic. The definition of these layers is emphasized by how they display cleavage (Plate 3d), some forming only a widely spaced fracture cleavage while others form ridges and grooves penetrative on a 1 cm scale. In exposures along the southeast margin of the Clare River Structure, compositional layers are much thinner (1-2 cm), alternate rhythmically, and are isoclinally folded (Plate 4a). The distinction between units 20 and 22 is not as clear in these outcrops as in the exposures around Sulphide.

Unit 23. Along the southeast contact of the Clare River Structure, the rocks of the Bogart Formation are intruded by pegmatite, to the extent that individual units can no longer be traced. Unit 23 was defined as calc-silicate gneiss and pegmatite; the former is mainly similar to unit 22 although slivers of marble and of units 19 and 20 are locally seen. At the contact with unit 14, coarse grained pink pegmatite constitutes about 90 percent of unit 23; it encloses thin lens of calc-silicate paragneiss which parallel the regional foliation. Unit 23 is truncated against unit 14 south of Bruno Lake.

Post-Flinton

Unit 24. Pegmatites are common throughout the area. They occur as pods, as irregular bodies at the contact with the Addington Pluton, and as discrete dykes cutting various rocks adjoining the Addington Pluton. Also included in unit 24 are the aplitic dykes that occur commonly in units 20 and 22 south of Otter Creek. Very coarse grained pegmatite pods occur in the pelitic schist outcrops about 2 km northwest of Otter Creek.

The pegmatites are composed of microcline, quartz, plagioclase, and muscovite; no exotic minerals were noted. Although some of the pegmatite bodies show a faint foliation, most appear to be undeformed. The disquieting observation of an isoclinally folded pegmatite vein with no discernible fabric, south of Bruno Lake, exhorts caution in interpreting the position of a rock in the deformational sequence based solely on fabric evidence.

Unit 25. Two flatlying outliers of Paleozoic sandstone occur in the northeastern section of the map area. One measures 750 by 130 m, the other is about 330 m in diameter. The rock consists of clean quartz-rich sandstone that locally contains quartz pebbles 2 - 3 cm in diameter. The samples were tested for radioactivity, and gave no response.

These basal sandstones are members of the Nepean or Potsdam sandstone of Lower Ordovician or Cambrian age.

PLATE 1

Aerial photograph (1:28512) of the southwestern margin of the Clare River Structure. The pattern supports the writer's contention that unit A is separate from the supracrustals of the Clare River Structure. The previous interpretation (Burns, 1951) is illustrated in Figure 11.

PLATE 2

Relict Textures in the Shovel Lake Volcanics

- a) Pillowed metabasalts at Sulphide. The pillow in the centre of the photograph faces northwest.
- b) Relict plagioclase phenocrysts in porphyritic meta-andesite. 1 km east of Otter Creek.
- c) Lapilli tuff. Flattened biotite-rich accidental fragments (meta-andesite) occur in a dacitic matrix 2.6 km south of Beatty Lake.
- d) Volcanic breccia. Flattened felsic fragments up to 50 cm long constitute about 70% of the rock. A more mafic biotite bearing matrix surrounds them. 1 km southeast of Hungerford.
- e) Alternating compositional layers in intermediate tuffs of the Shovel Lake Formation. Highly flattened fragments are visible in some of the layers. 0.3 km northeast of Hungerford.

PLATE 3

Primary Features Preserved in Metasediments of
the Flinton Group in the Clare River Structure

- a) Flame structures (?) in unit 20, horizontal surface, facing west.
0.5 km west of Bogart.
- b) Colour gradation in primary layering of unit 20, horizontal surface,
facing northeast. 0.6 km southeast of Bogart.
- c) Quartzofeldspathic segregations that dextrally offset primary
layering in unit 20. 0.6 km southeast of Bogart.
- d) Compositional layering defined by the pronounced development of
spaced cleavage (S2) in the central layer. Unit 22, 1.4 km southwest
of Hungerford.
- e) Granite pebble conglomerate (unit 15) exposed adjacent to the Addington
Pluton. 3.3 km west of Sulphide.

3. GEOMETRY OF THE CLARE RIVER STRUCTURE

Introduction

The supracrustal rocks of the Clare River Structure are isolated from adjacent metasedimentary belts by the Addington and Mellon Lake Plutons to the north and south respectively. On existing maps (e.g. Hewitt, 1964) these bodies are shown to merge near the southwest margin of the structure at Stoco Lake. In gross aspect, the geometry of the Structure is that of a synform within the underlying plutonic bodies.

In view of this fact, it is not surprising that the Clare River Structure was originally interpreted as a simple syncline and was named the Clare River Syncline (Wilson, 1933). Burns (1951) supported this interpretation, although he pointed out the existence of several complicating features, such as the pillow at Sulphide facing in the "wrong" direction, which he attributed to parasitic folding. Pearson (1958) suggested that cross-folding of the Clare River "Syncline" around the northwest trending axes resulted in doubly plunging lineations in the vicinity of Otter Creek, but it was not until 1969 that Venkitasubramanyam (1969), in compiling a regional model, suggested that the Clare River "Syncline" was in fact a second phase fold in a polydeformed terrain.

In detail, the supracrustal rocks of the Clare River Structure form a set of large scale refolded folds, illustrated by the macroscopic type III interference pattern traced out by several of the lithologic units (Figures 3 and 4). In models in which a passive flow fold mechanism is invoked, such a pattern is produced when the flow direction of the later folds is at a high angle to the axial plane of the earlier folds, and the angle between the respective fold axes is small (Ramsay, 1967, p. 521).

The structure of the Clare River supracrustal rocks can be geo-

metrically explained in terms of two phases of deformation. During the first phase large scale northeast trending, shallowly plunging, reclined to recumbent folds were formed. The majority of the supracrustal rocks are preserved on the overturned upper limb of one such fold, the Hungerford Antiform; its axial surface trace and a bit of its lower limb occur adjacent to the contact with the Addington Pluton; the majority of the lower limb is faulted out along the Addington contact.

The second deformational event produced upright tight to isoclinal northeast plunging folds of which the Clare River Synform¹ is the most important. Although not clearly defined by the stratigraphy, additional second phase closures must occur to the north of the Clare River Synform in order to explain the expanded exposure width of the Shovel Lake and Todd Lake Formations.

Mesoscopic Folds and Fabric Elements

Introduction

In this section a general description of the minor structures associated with the two main deformational phases, and also of later structures apparently not related to any features expressed on the map pattern will be given. A more detailed description of the interaction of the phases will be discussed under the headings of the individual subareas.

1. The axial surface trace of the Clare River Synform in part coincides with the axial surface trace of the structure traditionally called the Clare River Syncline (see Figure 7). When it was recognized that the Clare River "Syncline" was not a simple syncline, it was called the Clare River Synform (Moore and Thompson, 1972; Schwerdtner, 1977). In this thesis the term Clare River Structure replaces Clare River Syncline and Clare River Synform. The term Clare River Synform is retained to describe the major F2 fold in the Clare River Structure.

D1 Structures.

S1 is the dominant planar fabric throughout the map area, and with the exception of the southwestern corner, where primary layering is disposed at a high angle to S1, S1 is defined by parallel gneissic layering and mineral orientation in a composite S0/S1 surface. For convenience, this surface was mapped as S1 when S0 and S1 were parallel.

Indisputable first phase minor folds are rare, even though major first phase closures are plainly traced out on the map by the stratigraphy. Where they occur, F1 minor folds are generally long-limbed isoclinal folds from which it is impossible to determine vergence (Plate 4a, b). Many folds, particularly in the area west of Otter Creek, do not appear to fold a foliation, but by virtue of their position, symmetry and orientation, are interpreted as F2 structures.

L1 lineations are not well developed throughout most of the Clare River Structure. Only along its southwestern margin, and to a lesser extent in the core area of the Clare River Synform around Bogart, do L1 lineations occur. In this case, they are S0/S1 intersection lineations and parallel F1 fold axes. The long axis of stretched pebbles in the polymictic conglomerate (unit 7a) of the Todd Lake Formation may represent a first phase elongation direction, but is now parallel to the attitude of L2. It is possible that L1 is masked by L2 in the northeastern section of the map area.

During D1, a major decollement formed along the contact between the Addington Pluton and the Clare River supracrustal rocks. This is indicated by the fact that the Addington Pluton is not repeated within the Clare River Structure by the Hungerford Antiform (see Figure 6). In plan (Figure 3) the decollement can be seen along the southern contact of the Clare River Structure, where several supracrustal units are sheared out

against unit 14. This surface, along with the Hungerford Antiform, was folded to an upright orientation during D2.

D2 Structures

L2 is the predominant linear feature in the Clare River Structure. It is generally expressed as a strong colour streaking and ridging on S1 surfaces with biotite edges and hornblende prisms paralleling it. This lineation is particularly well exposed in outcrops on Highway 7 southwest of the town of Kaladar.

Where the Shovel Lake volcanics are folded in the core of the Clare River Synform, L2 is observed parallel to the fold axis, and is clearly seen to be the result of the folding of S1 into folds whose amplitude is the same or less than the thickness of compositional layering. On the limbs of F2 folds, its origin is obscure because it is very difficult to observe these small amplitude folds.

In micaceous layers, L2 is expressed as an edge lineation on S1, as micas originally parallel to that surface have been rotated out of it during D2. The evolution of the prominent L2 lineation is depicted in Figure 5.

S2 planar fabric is relatively weak throughout the map area. It is defined by mineral grains, mineral clots, crenulation cleavage, and a spaced cleavage which in some cases exhibits small offsets, all at a small angle to S1. While these features are all a result of D2, they were caused by rotation of pre-existing fabric elements rather than by mineral growth perpendicular to the principal compressive strain direction during D2. This accounts for their scatter (Figure 4). An exception to the above occurs in the marbles of the area, where calcite grains have recrystallized parallel to the axial planes of F2 folds.

Figure 5. Formation of the prominent L2 lineation. A) During D1, long limbed isoclinal folds, a strong foliation, S1, and a weak intersection lineation, L1, were developed. B) During D2, a compressive strain at a high angle to S1 resulted in the rotation of micas, and the formation of small amplitude F2 folds. These are respectively expressed as an edge lineation and a strong colour streaking on S1 planes.

F2 minor folds (Plate 4c, d; Plate 5) abound in the core zones of the major second folds, and their symmetry clearly marks the Clare River Synform, backing up the pattern traced out by the stratigraphy. They vary in style from open to isoclinal, depending on the rock unit in which they are observed. On the average, they can be described as close to tight. The majority plunge to the northeast, although fluctuations through the horizontal do occur locally, producing pseudo-type I interference patterns on the outcrop. Axial plane cleavage is usually weak.

Later Structures

Later structures include numerous kink bands and crenulations of the mica foliation that occur in the pelitic schist of the Beatty Formation (unit 16). In some cases these are parallel to the S2 axial surfaces, but generally are oriented in a more east-west direction, and are quite variable in orientation over the outcrops in which they occur. These may represent a late stage of D2, or they may be related to the open north-west trending warps of the entire structure (D3).

A lineation defined by chlorite slickensides is widespread on jointing surfaces parallel to the foliation in the Addington Pluton, north of the Clare River Structure. They are downdip, and clearly are late in the evolution of the area.

Detailed Description of Subareas

Introduction

The map area was divided into eight subareas (Figure 4), each chosen to illustrate a particular aspect of the entire structure. These subareas will be discussed in detail below.

Table 2. Structural history of the Clare River Structure

Phase of Deformation	Fabric Elements	Major Structures	Minor Fold Style	Metamorphic Grade
Later structures	-down dip slickensides on joint surfaces parallel to S1.	-open warps around NW trending axes.	-crenulations with NW axial surface trends in pelitic layers.	-post metamorphic
D2	-penetrative E to NE lineation, developed on S1 surfaces. -weak crenulation or fracture cleavage, NE trending, vertical to slightly dipping to the SE.	-map-scale folds; tight to isoclinal, NE trending, shallow to moderately plunging, axial surfaces vertical or slightly overturned NW.	-common, variable, generally close to tight, orientations and symmetry reflect that of the major structures; weak axial surface cleavage. F2 crenulations common in the pelitic units.	-lower amphibolite ?
D1	-penetrative mineral foliation, NE trending, vertical to steeply dipping to the SE.	-map-scale recumbent to reclined NE trending folds, major tectonic slides.	-rare; long limbed isoclines with attenuated limbs and highly thickened hinge zones.	-middle amphibolite

Subarea 1

The accurate representation of the outcrop pattern in subarea 1 is hindered by the fact that much of its southern portion is underlain by flood plain deposits of the Clare River. In spite of this, there is sufficient outcrop to illustrate that the supracrustal rocks of the Clare River Structure are truncated at a high angle by the Addington Pluton. This contact has been interpreted as a syn-D1 decollement surface, as illustrated in section F-F' of Figure 6. The area was subsequently folded during D2 to form the Clare River Synform, which is open, slightly inclined to the northwest, and in subarea 1, plunges moderately to the east.

The writer's interpretation in this area differs from those of previous workers (Wilson, 1933; Burns, 1951) in two main respects. First, rocks of unit A are thought to be isolated from the rest of the supracrustal rocks of the Clare River Structure. This change is important, as previous workers joined units A and 18 (Figure 7), obscuring the decollement surface which the writer's mapping indicates. Air photo interpretation (Plate 1) supports the writer's contention.

Secondly, previous workers have combined units 18 and 6 as one marble in this subarea, explaining the thickening in the hinge zone as a tectonic feature (Figure 7). Unit 18 is distinctive over several miles and differs markedly from the dolomitic marble of unit 6. In addition to this fact, outcrops of unit 10 and unit 17 crop out between the two marble units. It is in keeping with the structural style of the Clare River Structure to ascribe the presence of the dolomitic marble lens (unit 6) under discussion here to an S1 synformal closure. Minor structures are not present to either confirm or refute this interpretation.

Immediately east of Bogart, a body of calc-silicate paragneiss with quartzofeldspathic segregations (unit 20) occurs within unit 22. Its

LEGEND

A

Flinton Group

Shovel Lake Formation

Todd Lake Formation

Otter Creek Formation

All Marble

Addington Gneiss

Axial Surface Trace of the Clare River Synform

B

Metavolcanics and Paragneisses

Marble

Addington Gneiss

Axial Surface Trace of the Clare River Syncline

Figure 7. Maps of the Clare River Structure contrasting the interpretation of Burns (1951) and that of the writer. A is modified after Figure 3; B is modified after Ambrose and Burns (1956). Numbers locate some of the major differences in interpretation.

significance is unclear; it may be a rock unit identical to unit 20 in a higher stratigraphic position and therefore, in a strict sense, a different map unit, or the map pattern may be the result of a pre-D1 fault.

Attitudes of minor structures in subarea 1 are illustrated in Figure 4. The area is unique in that primary layering is commonly observed distinct from S1. F1 axial planes and hinge lines can be followed around the nose of the Clare River Synform. D1 minor structures in the calc-silicate units are complex, changing vergence and plunge frequently over small areas. Paucity of outcrop and lack of a marker horizon in this area prevents the fine details of the structure from being worked out.

As unit 20 is followed from southwest to northeast around the closure of the Clare River Synform, first phase folds are seen to be progressively tightened and a metamorphic layering distinct from S0 begins to form. In the southwest portion of the area, quartzofeldspathic segregations are observed to form along planes that offset S0 in a dextral sense (Plate 3c). As folds become tighter, S0 becomes obscure and the quartzofeldspathic segregations become continuous over several metres, forming a tectonic surface that is easily mistaken for S0. This feature is best observed on outcrops just north of Bogart where the Clare River forms two small lakes. On the south side, S0 is well preserved and is cut by quartzofeldspathic segregations that lie approximately in the axial planes of the F1 folds. North of the river however, S0 is no longer visible, having been supplanted completely by continuous alternating quartzofeldspathic and mafic bands of secondary origin.

D2 structures are more common in the northern half of this area. F2 fold axes and fold nullions plunge mainly to the east and southeast. While vergences of the minor folds mainly reflect the major structure, local symmetrical folds indicate complications that are not represented by

the stratigraphy. Mapping was not carried out in sufficient detail to illustrate these structures.

Towards the western end of subarea 1, a northwest trending foliation is observed to be cutting S1 at a small angle, and it contains the lineation mapped throughout the area as L2. In the writer's opinion, this feature reflects the regional warping of D2 structures around the southwestern end of the Elzevir Pluton (Figure 13).

Subarea 2

The cores of the first phase Hungerford Antiform and the second phase Clare River Synform are straddled by subarea 2, the third dimension of which is illustrated in the cross section E-E' of Figure 6. The expanded outcrop width of unit 22 west of Otter Creek is explained by its stratigraphic repetition in the cores of a series of F2 antiform-synform pairs. East of Bogart, the expanded exposure area of the calc-silicate paragneiss (unit 19) in the core of the Clare River Synform is due both to thickening by flow during folding and to inflections through the horizontal of F2 hinge lines. The intricate pattern displayed by the unit 18 southwest of Otter Creek is the result of interference between F1 and F2 folds. Mesoscopic interference patterns (Plate 5 e, f) are common in the area.

F2 minor folds (Plates 4 c and 4 d, 5) are present throughout the area, and they vary in style and orientation in response to several factors. These include rock type, intensity of D2 deformation at the point of observation (which can vary considerably on a single outcrop) and the initial orientation of the surfaces being folded. Because S1 fabric is often poorly developed in the sandy carbonate rocks of units 18, 19 and 22 in this subarea, some difficulty was encountered in associating folds

with D1 or D2. In the absence of evidence to the contrary (i.e. refolded folds) all folds were mapped as F2. F1 folds are localized in the extreme southwestern part of subarea 2, and these can be traced around the core of the Clare River Synform in subarea 1.

The pattern of L2 lineations in subarea 2 is distinctly bimodal. An examination of Figure 4 shows that the east-southeast trending lineations are spatially restricted to the area including and to the north of the F1 core, while in the southern portion of the area they plunge to the northeast. The best explanation for this feature is that it is the result of D2 fold axes and intersection lineations being formed on the divergent limbs of the F1 Hungerford Antiform. In addition, the predominance of F2 folds with S symmetry on both limbs of the F1 fold indicates that both limbs had the same sense with respect to the F2 axial surface. Figure 8 illustrates a model to explain these features.

The above interpretation is embodied in the cross sections (Figure 6) which show a major tight to close F1 fold (Hungerford Antiform) overturned to the northwest, the upper limb of which is refolded by F2 folds which are slightly overturned to the northwest. Because the lower limb of the Hungerford Antiform was oriented at a low angle to the F2 axial surface at the initiation of D2, parasitic F1 folds on this limb were tightened and rotated rather than folded during the second phase of deformation.

Subarea 3

In subarea 3, L2 plunges mainly to the east-southeast, as was the case in the northern section of subarea 2, and the spread in the poles of S1 is seen to be caused by D2 folding in the western part of the area.

Two folds of the contact between the Addington Pluton and the

Figure 8. Illustration of the pattern of L2 lineations formed on the divergent limbs of an F1 fold.

Shovel Lake volcanics occur in this subarea. The more westerly fold is seen in the field to be of second phase origin. many mesoscopic F2 folds are observed in its vicinity. The origin of the larger re-entrant to the east is unclear. No minor structures are associated with it, and it may be an F1 structure or it may be due to an early fault.

Although stratigraphy does not confirm it, there must be major F1 folds in this subarea. This is indicated by the fact that the D1 axial surface trace defined in subarea 1 must continue into subarea 3. It is possible that the large amphibolite body represents a major fold hinge; locally there is stratigraphic symmetry on either side of it, but the primary intertonguing of these volcanic rocks in adjacent areas indicates that it is difficult to make any structural assumptions based on their outcrop pattern.

Subarea 4

Subarea 4 is underlain by the F1 Hungerford Antiform. This fold is attenuated on the limbs and highly thickened in the hinge, in marked contrast to the more concentric form displayed by the F2 Clare River Synform. While the outcrop pattern shows no sign of the F1 fold having been refolded during D2, it was probably rotated and tightened at the time, as is illustrated in the cross sections C-C' and D-D' of Figure 6. The axial surface of the F1 fold is presently trending about 050° and is slightly overturned to the northwest.

While no unequivocal F1 minor structures were found, several symmetric tight to isoclinal folds in the core of the F1 antiform were mapped as first phase on the basis of their position, their symmetry with respect to the F1 structure, and their style. Their hinges parallel the prominent L2 lineation, and were probably rotated to this orientation during D2. S1

is defined by compositional layering and a weak biotite orientation.

S2 is a poorly developed spaced cleavage, while L2 is a pronounced colour streaking and weak biotite edge lineation. The strong colour-streaking lineation is formed because there are small offsets on the S2 cleavages which segment competent calc-silicate layers, while the biotite-rich layers were more ductile and able to flow. Often, the competent segments have been rotated slightly. The rocks break along S1 and these small offsets stand out as a pronounced colour streaking and ridging lineation.

The metavolcanic rocks on the north limb of the Hungerford Antiform in subarea 4 exhibit only an S1 planar fabric and an L2 lineation. They appear to be an uncomplicated sequence of southwest dipping rocks, but as this unit is traced both to the northeast and southwest, it becomes clear that there are a number of highly attenuated fold cores running through it.

Subarea 5.

In addition to the core of the F2 Clare River Synform, subarea 5 includes an antiform-synform pair defined by the contact between the Beatty and Shovel Lake Formations. Their rounded hinge zones suggest that they are D2 structures, but the expanded exposure width of unit 16 in that area is due to the interference of F1 and F2 folds (Figure 6, section D-D').

This juxtaposition of F1 and F2 hinge zones is indicated both by the presumed extension from subarea 3 of the F1 core defined by the rusty graphitic psammite (unit 17) of the Beatty Formation, and by the fact that F2 minor folds are seen to plunge in opposite directions without changing vergence, a feature most readily explained by assuming that these minor folds formed on the divergent limbs of an F1 structure. In fact the

calcareous marble of the Otter Creek Formation (unit 5) in the core of the Clare River Synform exhibits mesoscopic interference patterns showing the feature described above (Plate 5d).

Second phase minor folds are abundant in the pelitic schist (unit 16) and in the calcareous marble (unit 5) and these characteristically mirror the major structure, having rounded hinge zones and the appropriate vergence. Late kinks and crenulations in the pelitic schist (Plate 4d) and the late cleavage defined by aligned carbonate grains in the marble units may have formed during the waning stages of D2 or they may be related to the open warping of the entire structure about northwest trending axes.

In the dolomitic marble (unit 6), the transposition of competent siliceous layers has resulted in a rough surface texture on which floating knots and lenses of coarse diopside and tremolite stand out in relief (Plate 5a, b). In the hinge area of the Clare River Synform these layers define fragmented symmetrical F2 minor folds. On the limbs, vergence cannot be determined from the knots, but they define a strong L2 lineation.

In the Shovel Lake metavolcanics, minor folds are sparse, but L2 lineations are strong and consist of colour streaking, ridging, biotite edges, and aligned hornblende prisms on S1 surfaces. These lineations plunge down-dip in the core of the Clare River Synform, and it is clear from observations in this area that the colour streaking and ridging lineation that is ubiquitous in the Shovel Lake Formation is L2. No L1 lineations were folded around this core.

The Clare River Synform in subarea 5 is upright, and plunges moderately to the northeast. This is illustrated by the stereoplot of poles to S1, which can be seen to form a great circle which puts the plunge of the synform at about forty-five degrees to the northwest.

Subarea 6

The outcrop pattern in subarea 6 is straightforward, defining the F2 Clare River Synform. This structure is corroborated by second phase minor structures. Just west of Bruno Lake, a wedge of unit 2 lies out of sequence between units 3 and 4; this feature is thought to be the result of a pre-D2 fault. There is a marked variation in axial zone thickness exhibited by the different rock units (i.e. units 3 and 4), and this is further exaggerated because the fold is shallowly plunging.

Minor structures associated with the D2 deformation are abundant in the vicinity of the contact between units 4 and 5 where the calcareous marble and rusty graphitic psammite are interlayered. These folds stand out in relief because of the differential weathering of the layers, and are close to tight with slightly thickened, rounded hinge zones (Plate 5 c, d), unlike F1 hinges which are highly thickened in their axial zones. These folds vary in size but usually their amplitudes are less than a metre.

Compositional layering-cleavage intersections are common in the gneisses of unit 2, and these outline the Clare River Synform. The S2 cleavage is subparallel to S1, and is defined by a weak spaced-cleavage and by rotated mica flakes that cross S1. Micaceous layers were observed to exhibit crenulation cleavage in hinge zones of F2 folds. L2 is pronounced, and is defined both by a strong colour streaking on S1 planes, and by the orientation of mica flakes.

The attitude of the Clare River Synform in subarea 6 is made quite clear in the stereoplots of Figure 4. The fold is upright, tight, plunges shallowly to the northeast, and becomes nearly horizontal at Highway 41.

In the southwestern portion of subarea 6, rocks of the Beatty

Formation exhibit a southeast plunging mica edge lineation which can be followed for nearly 7 km to where L1 is folded around the hinge of the Clare River Synform in subarea 1. Unfortunately L2 crenulation and mica edge lineations locally plunge to the southwest too, making separation of the lineations on the basis of their orientation impossible. In the absence of evidence to the contrary, all southwest plunging lineations were mapped as L1.

Subarea 7

In addition to the Hungerford Antiform, there are three other closures defined by the stratigraphy in subarea 7. These are believed to be F1 folds for the following reasons: 1) their style is the same as that of the first phase Hungerford Antiform; 2) second phase minor structures do not change sense across them and, 3) their outcrop pattern indicates that they are doubly plunging, while L2 lineations exhibit a constant northeast plunge. Presently these folds are isoclinal and overturned to the northwest, with a mean axial surface trend of about 030° , dipping 70° to the southeast.

South of the pelitic schist outcrops, the metavolcanics of the Shovel Lake Formation have an expanded outcrop width, which is explained by large parasitic F2 folds on the northwest limb of the Clare River Synform. This interpretation, along with the postulated geometry of the F1 folds are shown in Figure 6, sections B-B'.

Minor structures are of little help in corroborating the above. In the pelitic schist, two isoclinal folds were observed with S2, defined by rotated sillimanite knots, crossing their limbs. These folds have an amplitude of several metres and are markedly extended in their hinge zones. Second phase minor structures are widespread. They vary from tight through

to open, are generally less than a metre in amplitude, and verge north-west, indicating that the second phase synform occurs to the southeast.

When micaceous layers are folded, crenulations parasitic to the minor folds are seen to develop, while in the more competent layers, sillimanite knots have rotated into parallelism with D2 (Plate 4c). A second set of crenulations trend to the east of F2 folds. When these two sets of structures were observed in a single outcrop, their distinction was clear, but when separate, there is no criterion other than orientation to separate them. It is probable that most of the crenulations are related to the second phase of deformation, although the possibility that they are related to the northwest trending warps of the entire structure must also be entertained.

Fabric elements in the Shovel Lake Formation are rather monotonous. S1 is defined by compositional layering which is usually discontinuous when traced over several metres and L2 is a pervasive lineation which is defined by biotite edges, colour streaking, and the alignment of amphiboles. This lineation parallels the F2 fold axes in the pelite and has a mean orientation of 40° towards 045° .

Subarea 8

Subarea 8 is separated from subarea 7 in order to illustrate the difference in orientation of the major fabric elements. The S1 surface is shallower also (30° towards 050°). An interesting feature is observed in the core of the F1 antiform, which splits into two smaller antiformal hinges separated by a synformal hinge of Shovel Lake Formation, as is illustrated in section A-A' of Figure 9.

Conclusions

Although polyphase deformation has been documented in areas adjacent to the Clare River Structure (Rivers, 1976; Thompson, 1972; Moore, 1967), and it has long since been postulated that the Clare River Structure is a second phase structure (Venkitesurbramanyan, 1969), the writer's detailed interpretation of the geometry of the Clare River Structure makes several important contributions to the understanding of the structural style and stratigraphy of the Kaladar-Dalhousie Trough.

The writer's mapping illustrates for the first time the folding of major first phase axial surface traces by second phase folds. Previous documentation for two phases of deformation in the Kaladar-Dalhousie Trough consisted only of minor structures.

Major thrust faulting associated with the early stages of deformation has been documented in the Adirondack Lowlands in New York State (Foote and Carl, 1977) and suggested in the Hastings Basin (Carmichael, 1968). In the Kaladar-Dalhousie Trough, the truncation of the supracrustal rocks in the Clare River Structure by the Addington Pluton (Figure 3) provides clear evidence of a major tectonic slide that was active during D1. The timing of the formation of this contact is discussed more fully in Chapter 8.

The fact that the rocks of the Clare River Structure form a fault contact with the underlying Addington Pluton indicates that the former is a nappe. The nappe is interpreted as having formed by shearing out of the limbs of reclined to recumbent F1 folds during D1. This is the first indication of such structures in the Kaladar-Dalhousie Trough, although they have been described in the Adirondack Lowlands (e.g. Brocum, 1971). Stratigraphic evidence, discussed in the following chapter, indicates that the majority of the supracrustal rocks exposed in

the Clare River Structure are overturned, further strengthening the argument for calling it a nappe.

PLATE 4

Minor Structures in the Clare River Structure

- a) F1 isoclinal folds in unit 23. 2 km southwest of Bruno Lake.
- b) F1 isoclinal folds in unit 19. 1.6 km south of Otter Creek.
- c) Quartz-plagioclase-sillimanite knots oriented at an angle to S1 due to rotation of the knots in a more ductile matrix during D2. Pelitic Schist (unit 16) 4.4 km northeast of Hungerford.
- d) Crenulation of S1 mica fabric by D2. Inclusions of S1 in the garnet (lower left) are visible. Same locality as c.

PLATE 5 \

Minor Structures in the Clare River Structure.

- a) Tight F2 fold in marbles of the Todd Lake Formation. 3 km southwest of Beatty Lake.
- b) Close up of a, showing boudinage of more competent calc-silicate layers and the reconstitution of compositional layering in the hinge zones of the ductile layers.
- c) F2 close to tight folds defined by competent layers in calcareous marble of unit 5 in the core of the Clare River Structure, 2.3 km southwest of Bruno Lake.
- d) Type III interference pattern. The variable plunge of the F2 fold axes results because the limbs of the refolded F1 fold originally had different dips. 2 km southwest of Bruno Lake.
- e) Type III interference pattern between F1 and F2 folds in marbles of unit 18, 0.5 km southeast of Otter Creek.
- f) Interference pattern between F1 and F2 folds in calc-silicates (unit 18) 0.4 km southeast of Otter Creek. The irregular shape is due to inhomogeneous flow between competent and incompetent layers.

4. STRATIGRAPHY AND CORRELATION

Introduction

The regional stratigraphy of the Hastings Basin (Figure 2) has been presented by Lumbers (1967 a, b), based on mapping by himself and by members of the Ontario Department of Mines. The Grenville Supergroup was divided into the predominantly volcanic Hermon Group and the predominantly sedimentary Mayo Group. These have been subdivided into a variety of formations as presented in Table 3 and Figure 9.

Lumbers was unable to confidently extend these units to the Kaladar-Dalhousie Trough and he therefore defined the Skootamatta formation (Lumbers, 1967a, p. 156) as:

"A prominent sequence of interbedded metasediments and metavolcanics characterized by abundant metaconglomerate that overlies and interfingers with the upper part of the Tudor Volcanics."

Since that time, a large amount of new data in the Kaladar-Dalhousie Trough have been obtained (Moore, 1967; Thompson, 1972; Moore and Thompson, 1972; Brown et al., 1975; Psutka, 1976). This work has led to the renewed assertion that there is a major unconformity within the Grenville Supergroup, and to the formal definition of the Flinton Group.

The Grenville - Hastings Controversy

One of the controversies surrounding the interpretation of the stratigraphy of both the Kaladar-Dalhousie Trough and the Hastings Basin is the presence or absence of a major unconformity separating two suites of supracrustal rocks. This controversy arose prior to 1900 with the work of Macfarlane, Vennor, and Selwyn¹. They described a suite of

1. Geological Survey of Canada, Reports of Progress, 1863-66, 1877-78.

Table 3. General lithology of formations of the Hermon and Mayo Groups. (after Lumbers, 1967b).

Group	Formation	Lithology
Mayo	Lasswade	marble, minor calcareous metasandstone and meta-siltstone, rare recrystallized chert
	Apsley	poorly sorted feldspathic metasandstone with upper and lower members of calcareous metasandstone and metasiltstone
	Dungannon	80% marble; remainder mainly calcareous metasandstone and metasiltstone, poorly sorted feldspathic metasandstone, well sorted quartz-rich metasandstone, and rare recrystallized chert; mafic meta-volcanic flows and iron-formation rare near base of formation overlying Tudor metavolcanics
Hermon	Burnt Lake	mainly rhyolite, trachytic, dacitic, and andesitic metavolcanics with minor metabasaltic flows; lensoid marble units and sandy siliceous metasediments found locally
	Turriff	75% pillowed basaltic and andesitic metavolcanic flows; remainder dacitic and felsic flows and pyroclastic rocks, and minor iron formation
	Vansickle	mainly poorly sorted feldspathic metasandstone locally with abundant marble, felsic and mafic metavolcanics, metaconglomerate, and well sorted quartz-rich metasandstone; arkosic metasediments commonly associated with felsic metavolcanics
	Oak Lake	mainly metamorphosed felsic pyroclastic rocks and arkose with some mafic and felsic metavolcanic flows; rare marble and metaconglomerate in upper part of formation
	Tudor	mainly metamorphosed, pillowed, basaltic and andesitic flows; minor dacitic and felsic metavolcanics and rare iron formation

Figure 9. Section showing the hypothetical relationship of the Mayo and Hermon Groups (after Lumbers, 1967b).

rocks consisting of conglomerate, argillite, schist, and well-bedded blue limestone that were much less metamorphosed than the type Grenville Group¹ described by Logan (1846) near the village of Grenville, Quebec. The former were termed the Hastings Group². Since that time, opposing views have emerged concerning their relationship. Some observers consider the Hastings Group to be a less metamorphosed part of the Grenville Group while others regard it as a younger group which unconformably overlies the rocks of the Grenville Group.

Based on detailed mapping of selected small areas over a wide area of the Kaladar-Dalhousie Trough, Miller and Knight (1914) concluded that the Hastings Group is indeed unconformable on the Grenville Group, the time interval being punctuated by the intrusion of the majority of the granitic rocks in the area. Other workers (Wilson, 1933; Smith, 1958 and 1969) postulated that the Hastings Group unconformably overlies the Grenville Group, but that the intrusive rocks of the area postdate it. Finally, Meen and Harding (1944), Burns (1951), and Ambrose and Burns (1956) concluded that the existence of the Hastings Group is open to question because conglomerates identified as "Hastings type" occur at different stratigraphic levels, and therefore do not mark a major unconformity.

The Flinton Group

On the basis of detailed mapping by Moore (1967), between Cloyne and Ompah (Figure 10), the concept of an unconformity between rocks correlated with the Hastings Group and rocks correlated with the Grenville

1. 2. Historically these were referred to as the Hastings Series and Grenville Series, but since they are rock-stratigraphic rather than time stratigraphic units, they will be referred to as the Hastings and Grenville Groups, in keeping with the code of stratigraphic nomenclature.

Figure 10. Distribution of the Flinton Group in a portion of the Kaladar-Dalhousie Trough. Compiled from MacDonald (1970), Rivers (1976), Moore and Thompson (1972), Thompson (1972), and this study. Numbers refer to sections in Figure 11.

Group was resurrected. Moore termed the former the schist-conglomerate suite and the latter the marble-paragneiss-metavolcanic suite. The following lines of evidence were cited by Moore (1967, p. 64) for the unconformity:

- 1) "Stratigraphy within the schist suite and adjacent conglomerate is consistent, but the unit contacts a variety of rocks; the suite is strikingly persistent whereas others are lenticular.
- 2) "Quartz pebble layers are ubiquitous at the base of the schist; the composition of the matrix varies with the adjacent rocks.
- 3) "In the area southwest of Marble Lake, truncation of the structures in the older rocks, by the schist, is evident locally on the mesoscopic scale, and generally on the map. Lithologic boundaries in the underlying rocks can be traced under the schist band.
- 4) "Northeast striking schistosity is parallel to the bedding in the schist suite and conglomerate, but generally transects layering in the underlying rocks southwest of Myer Cave, indicating two periods of folding. In much of the Northbrook-Ompah area, structures in the older and younger rocks are sub-parallel, rendering local detection of an unconformity difficult or impossible.
- 5) "Between Bishop Corners and Ardoash, sills and dykes of salic and intermediate meta-igneous rocks are common in the paragneiss, marble and metavolcanic rocks, but are absent from the conglomerate and the schist suite, indicating that all igneous activity in this area predates these units."

Thompson (1972) mapped the schist-conglomerate suite between Madoc and Bishop Corners and Moore and Thompson (1972) proposed formal stratigraphic nomenclature for the suite, terming it the Flinton Group after the informal name proposed by Burns (1951) for the metasediments in the Flinton area. It was felt that the name Flinton Group was more appropriate than the name Hastings Group for two reasons: first, rocks mapped as part of the Hastings Group occur below the unconformity described by Moore and Thompson, and secondly, the majority of the rocks mapped by Moore and Thompson occur outside of Hastings County. They also

defined the Grenville Supergroup to include both the Flinton Group, and the rocks historically included in the Grenville and Hastings Groups in the Central Metasedimentary Belt.

The Flinton Group has been traced over a strike length of over 100 km (Figure 10) by Moore and Thompson (1972) and broken down into formations described at type localities along the belt. Figure 11 shows columnar sections of the Flinton Group over most of the area mapped by Moore and Thompson. This work of Moore and Thompson (1972) is the first formal attempt at stratigraphy and sedimentology in the Grenville Province.

Stratigraphy of the Clare River Structure

Introduction

Based on the interpretation of the Clare River Structure as a simple syncline, Wilson (1933) grouped the Clare River supracrustal rocks with both the Hastings Group and the Grenville Group, their unconformable contact being marked by Hastings Type conglomerate (unit 7a). Harding (1944) on the other hand, who examined the lateral equivalent of this unit, mapped it as an intraformational conglomerate within the Grenville Group. Burns (1951) dispensed entirely with the Hastings-Grenville terminology in favour of a local informal stratigraphic nomenclature, because, according to his structural interpretation, "Hastings type" conglomerate occurred at different stratigraphic levels. The geometry of the Clare River Structure has been revised in this present study, and since formal stratigraphic nomenclature (Moore and Thompson, 1972) has superseded the informal units in the northwestern portion of Burns' (1951) map area, his stratigraphic succession has not been reproduced here.

According to the writer's interpretation, (see Chapter 3) the

LESSARD FORMATION

- 5c. Calcareous shale
- 5a. Calcareous feldspathic sandstone, conglomerate
- 5b. Plutonic pebble conglomerate

BISHOP CORNERS FORMATION

- 4c. Feldspathic sandstone and conglomerate
- 4b. Quartzite, quartzite conglomerate
- 4a. Shale

FERNLEIGH FORMATION

- 7. Black shale, calcareous shale

MYER CAVE FORMATION

- 6c. Dolostone conglomerate
- 6b. Black shale
- 6a. Dolostone, limestone

KALADAR METASEDIMENTS

- 8. Limestone
- 7. Calcareous shale
- 6. Rusty sandstone
- 5. Shale
- 4. Plutonic clast conglomerate

BOGART FORMATION

- 22. Calcareous sandstone
- 21. Dolostone, rusty siltstone
- 20. Calcareous graywacke
- 19. Calcareous shale
- 18. Limestone

BEATTY FORMATION

- 15. Plutonic clast conglomerate
- 17. Rusty siltstone
- 16. Shale

Figure 11. Columnar sections of the Flinton Group in a part of the Kaladar-Dalhousie Trough. Section locations are given in Figure 10. Data from Thompson, 1972; Moore and Thompson, 1972; Psutka, 1976; and this study. Rock names before metamorphism are inferred.

FLINTON-FERNLEIGH AREA

KALADAR
METASEDIMENTS correlative with the
Lessard Formation

metres

CLARE RIVER
STRUCTURE correlative with the
Lessard Formationcorrelative with the
Bishop Corners Formation

8

fault
contact

supracrustal rocks of the Clare River Structure lie in fault contact with the Addington Pluton (unit 14) and its equivalent, a narrow pink-weathering quartz monzonite body marking the southeastern boundary of the structure.

A narrow belt of calc-silicate paragneiss and marble, cut by white pegmatite (unit A) separates this lens of the Addington Pluton from the Mellon Lake Pluton on the southeast. Unit A was previously thought (Burns, 1951) to join up with the marbles of the Clare River Structure, but in the present interpretation, it is considered a separate, albeit thin, metasedimentary belt which separates the Mellon Lake and Addington Plutons (Plate 1).

The fault contact between the Clare River supracrustals and the Addington Pluton defines a second phase synform (Figure 6). Above this contact, Clare River supracrustals form a geometrically complex structure most readily described as a megascopic type III interference pattern. Fold axial planes and stratigraphic units are truncated against the Addington Pluton and its contact with the Clare River supracrustals can be classified as a decollement.

The supracrustal succession in the Clare River Structure was divided by the writer into five formations (Figure 12). From top to bottom they are: the Otter Creek Formation, the Todd Lake Formation, the Shovel Lake Formation, the Beatty Formation, and the Bogart Formation. The upper two formations are interpreted as belonging to the Flinton Group, and the lower three are part of the Hermon Group as defined by Lumbers (1967 b).

The majority of this succession is exposed on the overturned southeast limb of the first phase Hungerford Antiform and it can be observed in reverse sequence along the trace of the axial plane of the northeast plunging second phase Clare River Synform from Bogart to Highway 41. In the absence of evidence to the contrary, it is assumed that no tectonic breaks occur within the succession.

INFERRED PREMETAMORPHIC LITHOLOGY

Figure 12. Stratigraphic succession in the Clare River Structure. Thicknesses are based on the exposure widths of the various units, taken from Figure 3, and are uncorrected for strain.

One purpose of this section is to speculate on the environment of deposition of the sediments of the Clare River Structure. In order to do so, salient features are extracted from the descriptions of rock units (Chapter 2) and presented in the discussion that follows.

Hermon Group

Otter Creek Formation. The Otter Creek Formation consists of five units which record a change from active volcanism to the accumulation of limestone in a basin starved of clastic detritus. The bottom of the formation is not exposed: its lowest unit consists of massive felsic volcanic rocks (unit 1) which are locally flow-layered. Overlying these are uniform biotite-hornblende gneisses and schists (unit 2) which are locally fragmental.

The rest of the formation is of sedimentary origin. Unit 2 is overlain by quartzofeldspathic psammite with minor pelite (unit 3) followed by rusty quartzofeldspathic psammite (unit 4) and calcareous marble (unit 5). Units 4 and 5 are finely interbedded in the vicinity of their contact. The quartzofeldspathic psammite is probably derived from felsic volcanic edifices (unit 1); denudation of the volcanic terrain gradually lessened the supply of clastic debris ending in deposition of the carbonate of unit 5.

Todd Lake Formation. The Todd Lake Formation is composed of uniform dolomitic marble (unit 6) that is overlain by a heterogenous series of immature metasediments (unit 7). The marbles indicate a general lack of clastic detritus being deposited in the basin of accumulation. However, some clastic material did continue to find its way into the basin. This is indicated by the presence of scattered detrital feldspar grains in the marble.

Unit 7 consists of a variety of lithologies, including dolomitic marble, various calc-silicate gneisses and schists, chert, and most importantly, polymictic metaconglomerate (unit 7a). The pebbles in this conglomerate appear to be vein quartz, pink-weathering felsic volcanic fragments, and gray-weathering meta-andesite pebbles (biotite-hornblende gneisses) with subordinate amphibolite, calc-silicate, and marble pebbles. No granite pebbles were observed. The matrix of the conglomerate consists of calcite, diopside, epidote, biotite, microcline, plagioclase, and quartz.

The conglomerate does not occur as one mappable horizon, but as beds 1 - 10 m thick, at several levels within the formation. Pebble composition and size vary slightly from bed to bed. 20 km northeast of the map area, the lateral equivalent of unit 7a is a marble pebble metaconglomerate.

Pebble composition suggests that the conglomerate is in part derived from a volcanic terrain, and in part from a limestone terrain. This unit is interpreted as a beach facies that separates a volcanic island from a carbonate basin. The repetition of the conglomerate at different levels indicates fluctuating conditions, probably in relative sea level.

The writer envisions the environment of deposition of the preceding formations as a warm shallow sea dotted with volcanic islands. The West Indies or the Philippines are possibly modern sedimentological analogues.

Shovel Lake Formation. The Shovel Lake Formation consists of intermediate (50%), felsic (30%), and mafic (20%) metavolcanic rocks. Towards the top of the formation, sediments presumably derived from the volcanics are present as fine grained biotite schists and calc-silicate paragneisses. The details of the lithology of this formation are discussed more fully in Chapters 2 and 7.

The Shovel Lake Formation is the most widely distributed formation in the Clare River Structure. Its complex internal pattern is due both to folding and to the original interfingering of the volcanic members. The random variation in composition up section, the lack of lateral continuity of units, and the predominance of felsic and intermediate pyroclastics suggest that the Shovel Lake Formation was formed mainly as the result of explosive eruptions from volcanic centres rather than as flood lavas from fissure vents.

Flinton Group Correlatives

Rocks belonging to the Flinton Group are exposed mainly in the southwest portion of the Clare River Structure. The reasons for including these in this group will be discussed in the next section. In addition to the basal Beatty Formation and the overlying Bogart Formation, an isolated wedge of granite pebble metaconglomerate (unit 15) outcrops 3 km west of Sulphide at the contact between the Addington Pluton and the Clare River Structure.

The majority of the pebbles in this conglomerate resemble the Addington Pluton, and they decrease both in size and abundance away from the pluton. The pebbles are poorly sorted and some still retain a sub-angular shape even after being flattened during deformation. This suggests that at the time of deposition, the pebbles were angular. Adjacent to the Addington Pluton, the matrix is quartzofeldspathic, but as one moves away from the contact, the proportion of biotite and hornblende in the matrix increases.

It is probable that this unit formed as a colluvial deposit adjacent to the uplifted Addington Pluton. The Flinton Group post-dates the Addington Pluton (Thompson, 1972) and on this basis, the conglomerate.

is correlated with the Flinton Group.

There is no movement between this conglomerate and the Addington Pluton, and therefore the decollement which elsewhere separates the Clare River supracrustals from the Addington Pluton must be assumed to pass between the Shovel Lake Formation and the conglomerate in this area.

Beatty Formation. The basal unit of the Beatty Formation is a porphyroblastic pelitic schist (unit 16) which serves as a marker horizon, as it is easily distinguished and is widespread throughout the map area. It abruptly overlies amphibolite, intermediate volcanics, and felsic volcanics of the Shovel Lake Formation at various points along its contact, but it is gradational with the calc-silicate paragneisses of that formation, especially southeast of Hungerford. Locally, thin lenses of quartz pebble conglomerate (?), and cross-bedded quartzite (J. M. Moore, pers. comm., 1977) are found near the base of the pelite northeast of Hungerford.

A thin but persistent lens of rusty quartzofeldspathic psammite (unit 17) overlies the pelite. As the Beatty Formation is traced southwest from Hungerford along the north limb of the Hungerford Antiform, the pelitic schist pinches out and the rusty psammite directly overlies pyrite bearing lenses which occur near the top of the Shovel Lake Formation west of Sulphide. These lenses in the Shovel Lake Formation are thought to be exhalites associated with the volcanic activity and unit 17 may represent a weathering product of these lenses.

In most cases the pelite abruptly overlies the volcanics of the Shovel Lake Formation. The contact between them is clearly unconformable, as the pelite blankets volcanics of different composition. Locally a continuous unbroken sequence from meta-andesite through calc-silicate paragneiss to pelite occurs and this suggests that the contact between the

Shovel Lake and Beatty Formations is locally disconformable.

Bogart Formation. Two marble and three calc-silicate paragneiss members comprise the Bogart Formation. Each of the three calc-silicate units has a significant calcite content.

The lowest member of the Bogart Formation is a graphitic, muscovite-bearing calcareous marble (unit 18). It is in part intercalated with the rusty psammite of the Beatty Formation at their mutual contact. Unit 18 is overlain by calc-silicate paragneiss that is characterized by layers 1 - 5 cm thick that are alternately rich and lacking in biotite. This undisturbed fine alternation of layers suggests deposition below wave base in a basin supplied with both carbonate and clastic material.

These thinly layered calc-silicate paragneisses are overlain by more mafic metasediments (colour index 30%) in which plagioclase predominates over quartz as the leucocratic component. Primary layering is well preserved in this unit. Beds range from 1 - 10 cm in thickness and are graded in colour from light to dark, a feature interpreted as graded bedding (Chapter 2). Locally, current-ripple and convolute lamination, and rip up clasts (?) are preserved. These features suggest that this unit was deposited by turbidity currents. The high proportion of mafic constituents and the dominance of plagioclase over quartz suggests a mafic volcanic source area for unit 20, most likely to be the mafic volcanics of the underlying Hermon Group.

A poorly exposed unit consisting of buff-weathering dolomitic marble and rusty paragneiss (unit 21) separates unit 20 from thickly bedded (20 cm - 1 m) calc-silicate paragneisses which lack hornblende and biotite. Locally, poorly preserved structures resembling cross-stratification are observed in layers 10-30 cm thick. This feature, the thick layering, and lack of colour grading are suggestive of shallow water deposition.

There is a close lithological correspondence between the Bogart Formation and the Lessard Formation (Figure 11) defined by Thompson (1972) in the Flinton Area. Thompson interpreted the Lessard Formation as a shallow marine deposit, and it is probable that the Lessard and Bogart Formations are equivalent. Unit 20 is an exception since graded units are not recognized at Flinton.

Facing Direction of the Clare River Supracrustals

Southwest of Bogart, individual layers in several outcrops of unit 20 exhibit light to dark colour gradations and this colour change is interpreted as relict graded bedding. These beds are right way up and as they occur on the lower limb of the first phase Hungerford Antiform, it follows that the more widely exposed upper limb of that structure is overturned. One pillow observed by the writer in basalts at Sulphide indicated the opposite facing direction. M. E. Wilson (1956) found that pillows in the same area gave conflicting top indications and questioned the validity of their use. Features resembling flame structures occur in unit 20 at Bogart, and they also appear to give a facing sense that is contrary to that obtained from the graded beds. The outcrop upon which they occur is part of a series of large parasitic folds and because it is not clear on which fold limb they occur they cannot be used with confidence to determine tops.

It is clear that these sporadic and conflicting primary top indicators are not sufficient to establish the facing of the entire succession in the Clare River Structure. To firmly establish this facing it was necessary to correlate the Clare River succession with the rocks described by Moore and Thompson (1972) to the north. The Beatty and Bogart Formations are interpreted as belonging to the Flinton Group, and on the

basis of this correlation the majority of the stratigraphic sequence in the Clare River Structure must be inverted. This correlation is discussed in detail below.

Correlation with the Flinton Group

The correlation of the Bogart and Beatty Formations with the Flinton Group, described by Moore and Thompson (1972), underlies the stratigraphic and structural interpretation of the Clare River Structure.

This correlation is based on three lines of evidence:

- 1) the lithologic similarity of an individual marker horizon,
- 2) the correspondence of the stratigraphy above and below that marker horizon in both areas, and
- 3) the timing of an intrusive episode.

The most obvious lithologic correlation is that between the pelitic schist member of the Bishop Corners Formation (Moore and Thompson, 1972), and the pelitic schist member of the Beatty Formation. In the field, they are indistinguishable. Both consist of interlayered light gray (muscovite-rich) and dark gray to black (biotite-rich) varieties. Metamorphic index minerals such as staurolite, kyanite, sillimanite, and garnet are ubiquitous and outcrops in which these occur as large porphyroblasts are common in both the Bishop Corners and the Beatty Formations. Both members are thin, laterally continuous for tens of kilometres, and both contain lenses of quartzite and quartz pebble conglomerate.

The pelitic schist of the Beatty Formation overlies a variety of lithologies along its lower contact with the Shovel Lake Formation. It is overlain along its entire length by a thin rusty psammite layer followed by the marbles and calc-silicate paragneisses of the Bogart Formation.

72

Similarly the pelitic schist of the Bishop Corners Formation overlies a variety of lithologies, including mafic volcanics of the Tudor formation, intermediate metavolcanics, metatuffs or metagraywackes, and marbles. The correspondence between the overlying Bogart Formation in the Clare River Structure and the Lessard Formation, which overlies the Bishop Corners Formation in the Flinton Syncline, has already been noted.

Finally, Moore (1967) pointed out, that there was no igneous activity evident during the deposition of the Flinton Group in the Bishop Corners - Fernleigh Area. In contrast, the rocks underlying the Flinton Group are cut by dykes of intermediate and felsic compositions. This is also the case in the Clare River Structure: both the Todd Lake and Otter Creek Formations are cut by mafic and intermediate dykes while none were observed to cut outcrops of the Beatty and Bogart Formations.

Other Correlations

The stratigraphy of the rocks underlying the Flinton Group has not been well documented over a wide area. In the Cloyne area (Figure 10), Sethuraman and Moore (1973) have described a continuous section of meta-volcanic rocks with an apparent thickness of 7 km. Its lower portion consists of tholeiitic basalts which give way to a calc-alkalic suite comprising andesite flows and pyroclastics capped by rhyodacite pyroclastics. These volcanic rocks are correlative with the Hermon Group (as defined by Lumbers, (1976b) and the tholeiites that compose the lower portion of the section are part of the Tudor formation. These volcanics are overlain by marbles and metagraywackes. Morton (pers. comm., 1977) remapped much of the above area and he has suggested that erosional unconformities mark the transition from tholeiitic basalt to andesite, and the volcanic - marble contact.

The volcanics of the Shovel Lake Formation are chemically similar to the calc-alkalic suite at Cloyne (Chapter 7). The thick basalts that form the lower portion of the sequence at Cloyne are not represented in the Clare River Structure and conversely, at Cloyne, there are no units corresponding stratigraphically or lithologically to the Otter Creek and Todd Lake Formations. Other than the fact that the Flinton Group overlies both, and both contain thicknesses of calc-alkalic volcanics, there is little correspondence between the pre-Flinton sections in the Clare River and Cloyne areas.

Thompson (1972) found that calcareous schist and psammite directly underlie the Bishop Corners Formation in the Madoc and Actinolite areas. The upper member of the Shovel Lake Formation in the Clare River Structure is lithologically similar to and occupies the same stratigraphic position. The units are thought to be equivalent on the basis of this correspondence. The fact that these calcareous schists and psammites are associated with marbles and felsic volcanics at Madoc, a varied suite of volcanics in the Clare River Structure, and mafic volcanics at Actinolite indicates that their basin of accumulation was characterized by abrupt facies changes over short distances.

Lumbers' (1967 a, b) reconstruction of the stratigraphy of the Hastings Basin west of the Elzevir Batholith is shown in Figure 9. The volcanic pile in the immediate vicinity of the Elzevir Batholith is dominated by the Tudor basalts, but further to the west, these basalts become much thinner and the volcanic stratigraphy becomes more varied and is laterally discontinuous. Sedimentary sequences are common within the volcanics.

Immediately east and southeast of the Elzevir Batholith the Tudor basalts also dominate the pre-Flinton stratigraphy: only thin

lenses of more varied volcanics and sediments separate the Tudor formation from the Flinton Group. Further southeast, in the Clare River Structure, the Tudor basalts are not represented; the Hermon Group is present as two volcanic suites separated by a thick pile of marbles. Clearly, these units belong to the upper portion of the Hermon Group. In keeping with Lumbers' (1967 a, b) reconstruction, we can tentatively correlate the volcanic portion of the Otter Creek Formation with the Oak Lake Formation; the upper part of Otter Creek and the Todd Lake Formations with the Vansickle Formation; and the Turriff and Burnt Lake Formations with the Shovel Lake Formation.

Outside of the areas mapped by the writer and by Moore and Thompson, (Moore, 1967; Thompson, 1972), the significance of the Flinton Group in the regional stratigraphy has not yet been recognized. Rivers (1976), discovered Flinton Group lithologies at several structural levels in his map area. Intercalated with and overlying these, are rocks characterized as pre-Flinton by Moore and Thompson. Barring unrecognized structural complexities, Rivers' mapping suggests that the stratigraphy established by Moore and Thompson (1972) in the Flinton and Fernleigh synclines, and extended by the writer to the Clare River Structure may be more difficult to recognize in adjacent areas to the northeast.

Similarly, to the west of the Elzevir Batholith, Lumbers does not recognize a major unconformity comparable to that described by Moore and Thompson (1972) between the Flinton Group and underlying strata. He believes that conglomerates recognized as "Hastings type" are intraformational. It must be remembered that prior to the detailed mapping of Moore (1967) this was also thought to hold in the Kaladar-Dalhousie Trough.

More work is clearly necessary to establish and correlate stratigraphy in the Central Metasedimentary Belt of the Grenville Province.

In the Kaladar-Dalhousie Trough, the northeastern extension of the Clare River Structure, along with metasedimentary belts to the southeast await detailed mapping. In the Hastings Basin, detailed mapping of areas containing conglomerates similar to those contained in the Flinton Group should be undertaken to establish their structure and the stratigraphic position of these conglomerates. It is only by such detailed structural mapping that stratigraphy can be established and confidently correlated between adjacent areas in the Central Metasedimentary Belt of the Grenville Province.

5. STRUCTURAL CORRELATION

Introduction

Metamorphic and structural events corresponding to those that affected the Clare River Structure have been discussed by several writers (Rivers, 1976; Psutka, 1976; Mitchell, 1974; Thompson, 1972; Moore, 1967) who worked on the supracrustal belts to the northwest (Figure 2). In this section, these events will be described and correlated.

Regional Correlation of Structures

Pre-D1 Structures

Evidence of at least one phase of deformation predating D1 is preserved in exposures of the Tudor formation 3 km and 8 km southwest of Flinton (Figure 13; Appendix VI). This event is marked by a steeply dipping east-trending mineral foliation defined by the alignment of biotite, chlorite, and amphibole (Thompson, 1972; Appendix VI, this study). The isolated body of volcanics 8 km south of Flinton appears to be a synform in the Elzevir Batholith (Figure 13).

As the western boundary of the Flinton Syncline is approached, these early structures are rotated into parallelism with the northeast-trending D1 Flinton Syncline, complete transposition having taken place where the contact is reached. The open warps of the isolated volcanic body in the Elzevir Batholith are believed to be due to the D1 deformation of the Flinton Group (Figure 13).

There is ample evidence to show that this phase of deformation predates the deposition of the Flinton Group. At several localities, along the western margin of the Elzevir Batholith dykes of Elzevir

Figure 13. Correlation of major structures in a part of the Kaladar-Dalhousie Trough. Compiled from MacDonald (1970), Rivers (1976), Psutka (1976), Moore and Thompson (1972), Thompson (1972), and this study.

granodiorite truncate the pre-D1 foliation in the volcanics, and since the Flinton Group unconformably overlies the Elzevir Batholith (Thompson, 1972), the deformation under discussion must predate its deposition. Locally, breccias consisting of fragments of the Tudor volcanics, with disoriented foliation, occur in a matrix of Elzevir granodiorite, further illustrating this point. It is probable that this pre-D1 event is related to the emplacement of the Elzevir Batholith. This phase of deformation is discussed in more detail in Appendix VI.

D1 Structures

The writer's interpretation of the deformational history of the Clare River Structure is in general agreement with the interpretations of belts to the northwest proposed by Rivers (1976) and Thompson (1972). The first phase of deformation generated a pervasive penetrative foliation defined by mineral orientation and by transposed compositional layering. An L1 elongation lineation defined by the axes of stretched pebbles and the alignment of acicular minerals, described by both Rivers (1976) and Thompson (1972), is not well defined in the Clare River Structure. Although it is probable that L1 is masked by the strong L2 lineation in the northeastern portion of that structure, it is clearly absent in the core of the Clare River Synform east of Otter Creek (Figure 8). The absence of L1 is not incongruous with other areas; pebbles strained during D1 in the Flinton conglomerate north of the Northbrook Antiform are highly flattened but locally show little preferred elongation (Chappell, unpublished mapping, 1974).

The positions of major F1 axial surface traces are shown in Figure 13. In both the Clare River Structure and the Ompah area, the D1 deformation is thought to have produced major-reclined to recumbent,

northeast trending, shallowly plunging folds. These were subsequently refolded to upright orientations during D2. It is only in the southwestern portion of the Clare River Structure that F1 axial surface traces are seen to be bent around F2 folds. Rivers (1976) suggested that the absence of megascopic interference patterns between F1 and F2 in the Ompah area is due to the fact that nearly complete transposition of primary layering into S1 has occurred. This feature is also observed in the northeastern portion of the Clare River Structure.

The Flinton and Fernleigh Synclines are interpreted by Moore and Thompson (1972) as upright F1 structures which have not been noticeably affected during D2. This differs from the interpretations of both the writer and Rivers (1976), who view the present upright orientation of D1 structures as due to their being refolded during D2. The upright attitude of the Flinton Syncline may be the result of its proximity to the Elzevir Batholith.

The reason for the upright orientation of the Fernleigh Syncline is less clear. Rivers (1976) prefers to call it the Fernleigh Belt, seeing no evidence of it being a syncline where it extends into his map area. East of Cloyne, the Fernleigh Syncline can be documented both on the basis of stratigraphic symmetry and primary facing evidence (Moore and Thompson, 1972). Northeast of the bend which marks a change in trend from north-northwest to northeast (Figure 13), stratigraphic symmetry disappears and the belt becomes thinner, Moore and Thompson (1972) maintain that the syncline persists, attributing the lack of stratigraphic symmetry to major facies changes in the Flinton Group. Rivers (1976) contends that the Fernleigh Syncline dies out along its length to become a monocline which he terms the Fernleigh Belt. A third possibility is that the southern limb of the Fernleigh Syncline was sheared out during D1.

There is clearly scope for additional work in and around the Fernleigh Syncline. In addition to discovering its true configuration, work should be undertaken to understand the geometric relations between D1 and D2 structures with a view to explaining why rocks of the Flinton Group, which are continuous for tens of km along strike in the Fernleigh Syncline, have only locally been repeated across strike by either D1 or D2.

The Kaladar Metasediments lie immediately northwest of the Clare River Structure (Figures 2,10). Psutka (1976) interpreted the belt as comprising a series of F1 folds which form the overall geometry of a syncline within the Northbrook Pluton. On the southeast, the Kaladar metasediments are bounded by the Addington Pluton. Psutka interprets this contact as a syn-D1 to slightly post-D1 thrust fault which truncates both the Northbrook Pluton and the Kaladar Metasediments.

The realization that major low angle faulting accompanied D1 is a major step towards understanding the details of the geometry of the Kaladar-Dalhousie Trough. It was shown that the Clare River Structure is a major nappe lying on the Addington Pluton. The lower contact of the Addington Pluton is a shallowly-dipping fault contact with the Kaladar Metasediments, on the Northbrook Gneiss. This suggests that the Addington Pluton is also a nappe. This interpretation is embodied in Figure 14.

D2 Structures

The structural grain of the Kaladar-Dalhousie Trough is defined by large, northeast-trending, gently plunging F2 folds (Figure 13). All the major F2 antiforms are marked by the large granitic bodies of the area; the Clare River Synform is an important F2 synform; and several smaller F2 synforms are exposed north of Ompah. The axial surfaces of these major structures are vertical to slightly overturned to the northwest

Figure 14. Cross section across the Flinton Syncline, Kaladar Belt and Clare River Structure. Section location shown in Figure 13. 1) Elzevir Batholith, 2) Flinton Syncline, 3) Northbrook Pluton, 4) Kaladar Belt, 5) Addington Pluton, 6) Clare River Structure.

and, except for the Cross Lake Antiform, they are tight to isoclinal.

With the exception of the strong L2 lineation in the Clare River Structure, D2 fabric elements are subordinate to those generated during D1. Only a minor amount of mineral growth occurred during D2, resulting in a poorly defined planar fabric. F2 minor folds are widespread throughout the area (Figure 13).

More detailed mapping is necessary in the areas north and east of the Cross Lake Antiform to define the pattern of F2 folds. An important problem for future consideration is the relationship between the major F2 antiforms defined by the plutonic bodies and structures in the cover.

D3 Structures

The major open folds of F2 axial surfaces (Figure 13) around northwest trends, south of the Elzevir Pluton, have been attributed to a D3 deformational event (Thompson, 1972). He separated minor structures related to this phase from D2 minor structures mainly on the basis of their orientation; in only one case was interference between F2 and F3 structures recorded. In the Clare River Structure, some of the crenulations of S1 in the pelitic schist (unit 16) may be related to these structures.

Rivers (1976) records a D3 event characterized by northeast-trending, subhorizontal minor folds which possess shallowly dipping axial surfaces. These folds show no axial plane cleavage and do not affect the outcrop pattern. Thompson (1972) records a similar event, which is D4 in his deformational sequence. It is clear that Thompson's D4 and Rivers' D3 are the same event.

Polyphase Versus Continuous Deformation

There is little doubt that there are refolded folds in the Kaladar-Dalhousie Trough. The two major phases, D1 and D2, a sporadic D3 event, and a locally preserved pre-D1 event, have been discussed in the preceding section of this chapter. In the Central Metasedimentary Belt outside the Kaladar-Dalhousie Trough, refolding has been described by several workers (e.g. Appleyard and Stott, 1975; Appleyard, 1974; Divi and Fyson, 1973; Carmichael, 1968; Wynne-Edwards, 1967; Foote and Carl, 1977).

A question has arisen as to whether these folds are the result of polyphase deformation or a continuous progressive deformation. Polyphase deformation is the result of a series of separate pulsatory compressions during an orogenic cycle (Ramsay, 1967) whereas superimposed folding during one progressive deformation is the result of a change in the orientation of the layering with respect to the axes of infinitesimal strain ellipsoid. Wynne-Edwards suggested (1972) that these may be the result of the same stress system, continuous deformation occurring in the deeper levels, episodic deformation in the higher levels of an orogenic area.

When deformational phases are separated by time markers, polyphase deformation can readily be established. If regional unconformities or intrusive events occur between deformational events, there clearly has been a time break between the phases. More commonly used features to separate the phases of deformation are metamorphic culminations and fabric relationships. These are not necessarily indicative of prolonged time intervals; they may be the result of changes in rock rheology in response to conditional changes during a continuous compressional event. However, these criteria do indicate the sequence of metamorphic and deformational events, and have become accepted as criteria to establish polyphase deformation. Phases of deformation need not be separated by long time intervals.

The deformational events in the Kaladar-Dalhousie Trough can be classified as polyphase. The pre-D1 and D1 events are separated by an unconformity between the Flinton Group and the Hermon Group and by an intrusive episode.

D1 and D2 can be classified as polyphase in that they are clearly sequential. This is evidenced by the fact that S1 fabrics are folded by F2 folds. A line of evidence suggesting that the temporal gap between D1 and D2 was not significant is that only one metamorphic event is recorded (Rivers, 1976; Thompson, 1972; Houslow and Moore, 1967) in the Flinton Group, and that both D1 and D2 occurred during this metamorphic event.

Wynne-Edwards (1967) suggested that synchronous refolding was the mechanism responsible for the refolded folds which he mapped in the Westport area of the Frontenac Axis (Figure 2). Yet crenulated schistosity is a common feature in that area (J. M. Moore, R. L. Brown; pers. comm., 1978), as it is in the Kaladar-Dalhousie Trough and the Adirondack Lowlands. If the arguments for polyphase deformation are accepted in these adjacent areas, it seems reasonable to propose that the Frontenac Axis was also subjected to polyphase deformation.

Conclusions

The detailed studies of the writer, Rivers (1976), Thompson (1972), and Moore (1967) are broadly compatible. Minor problems in this synthesis, such as the attitude of the Flinton Syncline during D1 and the configuration of the Fernleigh Syncline remain to be solved. It is hoped that this synthesis will provide a framework for future work. One important aim of future studies should be to study the structural relationships between the F2 antiforms formed by the plutonic bodies and the cover. Another is to correlate structural events between the Kaladar-Dalhousie Trough and the Hastings Basin, west of the Elzevir Batholith.

6. TIMING OF METAMORPHISM AND DEFORMATION

Introduction

Depending on their relationship with a tectonic fabric, minerals may be classified as syn-, pre-, or post-tectonic (Rast, 1958; Zwart, 1960; Spry, 1968; Spry, 1969). Minerals such as micas, amphiboles, and tabular quartz grains, which exhibit a preferred orientation, are generally considered to be syn-tectonic. Syn-tectonic minerals may also occur as porphyroblasts which are characterized by curved inclusions of the external foliation, although in special cases, they may contain straight inclusion trails. Pre-tectonic minerals are wrapped by external foliation and they may be broken or exhibit strain extinction; post-tectonic minerals randomly overgrow the foliation, or the foliation may simply abut against them. Some hypothetical examples of these features are illustrated in Figure 15.

The special case of a syn-tectonic porphyroblast overgrowing straight inclusion trails will occur if the finite strain during the time of crystal growth was irrotational (Elliott, 1972) or if the rate of porphyroblast growth was substantially greater than the rate of formation of mineral fabric. Zwart (1960) described porphyroblasts which contained a plane internal fabric that was continuous with the external fabric but rotated with respect to it. To account for the straight inclusion trails, Zwart suggested that the rate of porphyroblast growth was high with respect to rate of deformation. When elongate and oriented porphyroblasts contain concordant, straight inclusion trails, or when they contain straight inclusions that are finer than, but continuous with the external foliation, they are probably syn-tectonic (Figure 15d). On the other hand, randomly oriented porphyroblasts overgrowing straight inclusion trails similar to the external foliation are post-tectonic (Figure 15e).

Figure 15. Hypothetical examples of syn-, pre-, and post-tectonic mineral growth. a) Pre-tectonic garnet grain containing a discordant internal foliation wrapped by a later external foliation. Pressure shadows form at both ends of the grain. b) Pre-tectonic quartz crystals exhibiting undulatory extinction and polygonization in response to strain. c) Syn-tectonic snowball garnet containing curved inclusions of the external foliation. d) Syn-tectonic kyanite crystals aligned parallel to the external foliation and containing fine inclusions of it. e) Post-tectonic idioblastic staurolite crystals overgrowing foliation. f) Post-tectonic crossed muscovite flakes growing across the foliation.

Mineral Textures in the Clare River Structure

Introduction

Two phases of deformation of the Clare River Structure can be documented in the field. In this section, these deformational phases will be related to mineral growth and fabric development, as observed in thin section. The pelitic schist member of the Beatty Formation (unit 16) is particularly well suited for this study as it is widespread in the map area, always exhibits a well defined foliation, and contains a variety of index minerals, often as porphyroblasts.

Forty-four thin sections of pelitic schist were examined in this study and the relationship of the mineral phases with respect to the principal foliation was determined in each section. This fabric data is tabulated in Appendix III.

Foliation (S1) Forming Minerals

The penetrative fabric mapped as S1 in the field is the dominant mineral fabric present in all thin sections. It is defined by flakes of muscovite and biotite, and by tabular grains of quartz, plagioclase and opaque minerals, all of which exhibit the same preferred orientation. In samples containing a high proportion of muscovite, the muscovite occurs as parallel interleaved layers containing tabular inclusions of opaque minerals and quartz. In psammite samples, muscovite and biotite occur in the form of individual flakes growing along the grain boundaries of tabular quartz, plagioclase and opaques. In the sillimanite zone, pods of fibrous sillimanite generally lie in S1. It is with respect to this fabric, which can be discerned throughout the map area, that mineral growth can be dated.

Relations Between Porphyroblasts and Foliation

Kyanite, plagioclase, staurolite, and garnet are present as porphyroblasts in the pelitic schist. The first three are restricted to the northwestern exposures of the pelite while garnet is common throughout.

These porphyroblasts can be spectacular. Idioblastic staurolite crystals up to 8 cm long and garnets 4 cm across, which display a variety of perfect crystal faces, can be found in pelite exposures in the core of the Hungerford Antiform.

An examination of Appendix III illustrates that the porphyroblasts nearly always contain straight inclusion trails which are continuous with the external foliation (Plates 6a,c,h). Usually the porphyroblasts are randomly oriented and contain inclusions which are the same size as the external foliation. Porphyroblasts displaying these features are post-tectonic.

In many sections, porphyroblasts display discordant inclusion trails (Figure 16b, c; Plate 6h). These discordant trails are the result of a rigid rotation of the porphyroblast subsequent to its growth; they do not represent an earlier foliation. The evidence of this is that these porphyroblasts usually have been rotated so that their long axis is oriented at an angle to the foliation (Plate 6h, Figure 16 c); the rotation is commonly observed in sections which show other post-D1 structural events; and finally, in all sections in which these discordant trails are present, individual porphyroblasts in which the inclusion trails are continuous with the external foliation can be found.

Also indicative of post-tectonic mineral growth are the crossed micas which are present to some degree in nearly all sections. These muscovite and biotite flakes, which are oriented at a high angle to the S1 foliation, exhibit a blockier habit than the same minerals growing parallel

Figure 16. Textural relationships of the pelitic schist. a) Post-tectonic kyanite crystals overgrowing straight inclusions of S1. Slide C75-94b, 0.6 km southwest of Shovel Lake. b) Same slide as (a). In this case, kyanite overgrows S1 and has been subsequently rotated to produce a discordant relationship with S1. c) Xenoblastic garnet overgrows S1 and was subsequently rotated during D2 to produce the discordant trails. Slide C75-68a, 1.6 km west of Todd Lake. d) Same slide as (c). In this case, the garnet contains inclusions of S1 that are continuous with the external S1. Both have been deformed by F2 folds. e) Syn-tectonic staurolite crystal is aligned in the S1 foliation plane and also contains inclusions of it. Slide C75-88b, 0.4 km northwest of Shovel Lake. f) Syn-tectonic garnets, containing inclusions of S1, which have been rotated differentially with respect to the external S1. Sillimanite parallels S1. Slide C75-258, 2.8 km northwest of Mitten Lake.

Note: All scale bars are 1 mm.

to the fabric. A typical length to width ratio for crossed micas is in the order of 3 to 1, versus 20 to 1 for those lying in the foliation. These micas indicate that temperatures were maintained at a level sufficient to crystallize muscovite and biotite after the first phase deformational event had ceased.

There is some evidence to show that porphyroblast growth also proceeded during D1. It is often observed that porphyroblasts contain inclusion trails that are finer than but continuous with the external fabric (Plate 6h). These fine inclusions are thought to represent an early stage in the development of S1. S1 wraps these porphyroblasts as it continues to develop (Figure 16 e). Staurolite, kyanite, and plagioclase grains are locally found aligned in the S1 plane, which suggests that they began to grow syn-tectonically. The most lucid evidence of syn-tectonic mineral growth was observed in section C75-270a (Appendix III). In this section, staurolite grains have overgrown an early stage of S1 and have been subsequently rotated so that their long axes lie in the plane of the S1 foliation. S1 wraps these staurolite porphyroblasts and in several cases, the external fabric is continuous with the discordant internal fabric.

Curved inclusion trails, the standard indicator of syn-tectonic mineral growth, are rare in the pelite in the Clare River Structure. The absence of curved trails in porphyroblasts thought to be syn-tectonic suggests that they grew rapidly with respect to the rate of strain. This is also supported by the absence of size variation of inclusions from the centre to the rim of syn-tectonic porphyroblasts, where the inclusions are noticeably finer than the external foliation.

As the pelitic schist is traced southwest towards the core of the Clare River Synform, prominent porphyroblasts of kyanite, staurolite, and plagioclase disappear as pods of fibrous sillimanite become common. In thin section, small fragments of staurolite lying in the foliation plane are

commonly observed in the sillimanite zone. The fact that they are localized and are often in optical continuity suggest that they are relicts of large poikiloblasts which were replaced by their inclusions as metamorphism and deformation progressed.

In the core area of the Clare River Synform, and on the southern limb of that structure (Figure 3), the cores of garnet porphyroblasts commonly enclose very fine randomly oriented inclusions of quartz and opaque minerals (Plate 6e). This suggests that in this area, metamorphism was initiated prior to deformation.

These garnets possess idioblastic rims which are free of inclusions. Such inclusion zoning is common in garnets in the sillimanite zone and was observed in one case in staurolite porphyroblasts. In some cases, the garnet cores contain randomly oriented inclusions, in others, fine inclusions are aligned and represent an early stage of S1 fabric development. The rims may be free of inclusions, or less frequently, they may overgrow the external fabric (Plate 6a). This inclusion zonation suggests two discrete phases of garnet growth. Rivers (1976) observed this feature in similar rocks 60 km to the northeast and suggested that it was a result of different garnet forming reactions.

D2 Textures

The second phase of deformation in the Clare River Structure occurred after virtually all mineral growth had ceased. The only exception occurs in marble units where aligned grains of calcite or dolomite lie in the axial plane of F2 folds. In other units, D2 is evidenced in the field by the rotation of minerals into its axial direction, by kinks of existing fabric, and by crenulation cleavage.

F2 folds of the S1 foliation are common in thin sections of the

pelite. These folds are small in amplitude and vary in style from gentle warps to tight chevron folds. In the noses of such folds, muscovite and biotite generally exhibit strain extinction (Plate 6b), although occasionally, they are annealed. Quartz and plagioclase are commonly strained around F2 folds. 2 km northwest of Hungerford, sillimanite needles are broken and pulled apart around the nose of one F2 fold (Plate 6d). Muscovite and biotite often lie parallel to the axial surfaces of the F2 folds, but are assumed to have been rotated to that orientation during D2. Porphyroblasts have also been rotated during D2.

S2 planar fabric was observed to be expressed as crenulation cleavage (Plate 6b), or as a series of discontinuous spaced fractures (Plate 6g) that are marked by dark residue. These observations are in keeping with the field observation that S2 is expressed as a crenulation cleavage or a spaced fracture cleavage.

Grade of Metamorphism

The portion of the Clare River Structure mapped by the writer lies entirely within the medium grade of metamorphism, as defined by Winkler (1974), or the amphibolite facies (Turner, 1968). Mafic metavolcanic rocks consist essentially of plagioclase and hornblende; siliceous carbonates are characterized by tremolite, diopside, and dolomite; and the pelitic rocks contain almandine, staurolite, kyanite and sillimanite. The latter are typical of a part of the classical Barrovian facies-series (Miyashiro, 1961). The petrography of all thin sections examined is presented in Appendix I.

Three mineral zones can be recognized in the field: a staurolite-kyanite zone, a sillimanite-staurolite zone and a sillimanite zone. The positions of these zones and their correlation with the pelitic zones mapped by Rivers (1976), and Moore and Thompson (1972) is presented in Figure 17.

Figure 17. Mineral zones in the Kaladar-Dalhousie Trough. Metamorphic boundaries are conjectural where broken. After Moore and Thompson (1972), Rivers (1976), and this work.

Timing of Metamorphism and Deformation

It is apparent from the preceeding discussion that one metamorphic event affected the Clare River Structure and that it was broadly coeval with the first phase of deformation. The evidence suggests that the second phase took place when the temperatures were below that of the metamorphic peak.

In detail, both the onset and the peak of metamorphism were diachronous if the S1 foliation is used as a time marker. In the staurolite-kyanite zone, the index minerals post-date S1 for the most part and the metamorphic peak can be considered to have occurred slightly after D1. In the sillimanite zone on the other hand, staurolite has clearly been deformed by S1, and sillimanite tends to parallel the foliation, suggesting that the peak of metamorphism was syn-tectonic.

Incipient metamorphic textures are partially preserved as fine inclusions in the cores of porphyroblasts. In the staurolite-kyanite zone, these early textures are aligned wherever observed. Conversely, in the core of the Clare River Structure, randomly oriented inclusions within garnets are common (Plate 6e), the metamorphism having been initiated prior to deformation.

Correlation with Adjacent Areas

Metamorphism of the supracrustal belts north of the Clare River Structure has been studied by Thompson (1972), Sethuraman (1970), Psutka (1976), and Rivers (1976). They all cite evidence of a single major metamorphic event, although locally, Thompson has found vestiges of a pre-Flinton metamorphism in volcanic rocks west of Flinton, probably related to the intrusion of the Elzevir Batholith. There is little doubt that the metamorphic event which affected the Clare River Structure is the same as

that which affected the rocks to the north.

Detailed fabric studies led both Thompson (1972) and Rivers (1976) to conclude that the metamorphic culmination occurred between the D1 and D2 structural events. Although porphyroblasts grew to some extent during D1, most index minerals overprint the S1 fabric. A similar time relationship between D1 and the metamorphic event has been documented by the writer in the Clare River Structure in the staurolite-kyanite zone, although most sillimanite in the sillimanite zone appears to be syn-tectonic.

In the Clare River Structure, D2 occurred at a relatively low temperature. No mineral growth is evidently related to D2: muscovite and biotite display undulatory extinction around the hinge zone of F2 folds, and quartz is commonly strained in samples affected by D2. These findings are at variance with those described by Thompson (1972) and Rivers (1976) who suggest that temperatures remained high during D2. Although no penetrative mineral fabric defining S2 was found by either Thompson or Rivers, their observations of micas annealing in the hinge zones of F2 folds, of pegmatites cutting across F2 folds, and of chlorite and biotite lying in S2 led them to this suggestion. It seems that temperatures during D2 were higher to the north of the Clare River Structure.

PLATE 6

Metamorphic Textures of the Pelitic Schist

- a) Garnet with inclusion free core and subidioblastic rim overgrowing tabular opaques which define S1. Plane light, slide C75-112a, 0.4 km east of Beatty Lake.
- b) F2 folds of S1 foliation. Hinge zones display strain extinction. Crossed polars, slide C75-051d, 1.2 km southwest of Shovel Lake.
- c) Euhedral staurolite crystals overgrowing straight inclusions of S1 that are continuous with the external S1. The staurolite crystals have been slightly rotated during D2. Plane light, slide C75-055, 1.7 km south of Shovel Lake.
- d) F2 folds of sillimanite needles that grew during D1. The needles appear to be broken in the hinge zones. Plane light, slide C75-256b, 2.7 km west of Bruno Lake.
- e) Subidioblastic garnet crystals. Cores contain slightly aligned inclusions, rims are inclusion free. External S1 wraps the garnets. Crossed polars, slide C75-157a, 0.5 km east of Otter Creek.
- f) Spaced S2 cleavage, defined by fine dark east-west trending lines cutting penetrative west-northwest trending fabric, defined by mica flakes. Crossed polars, slide C75-157a, 0.5 km east of Otter Creek.
- g) Subidioblastic staurolite contains fine inclusions of quartz and opaques which parallel S1. Sillimanite, which also parallels S1, nucleates on the faces of the staurolite crystal. Plane light, slide C75-73a1, 2.4 km northwest of Bruno Lake.
- h) Garnet crystals containing straight inclusion trails of S1 that is finer than the external S1. The garnets have been subsequently rotated to produce discordant inclusion trails. Crossed polars, slide JM-16a, 2.8 km northwest of Hungerford.

Note: all scale bars are in mm.

7. GEOLOGY AND GEOCHEMISTRY OF THE SHOVEL LAKE FORMATION

Introduction

The presence of metamorphosed volcanic rocks in the Clare River Structure has formerly been documented (Wilson, 1933; Burns, 1951). These are particularly apparent in the vicinity of the village of Sulphide where primary volcanic structures and textures (Plate 2) are clearly preserved. The purpose of this chapter is to describe the stratigraphic position of these volcanics in the light of the revised structural interpretation, to discuss their chemistry, and to compare them with the volcanic rocks in the Bishop Corners-Donaldson area described by Sethuraman and Moore (1973).

The writer has recognized two stratigraphically distinct meta-volcanic units within the Clare River Structure: the Shovel Lake Formation and the two lower members of the Otter Creek Formation. They are separated by the metasediments of the upper Otter Creek Formation and the Todd Lake Formation respectively (Figure 3, 12). The metavolcanics of the Otter Creek Formation consist of massive felsic metavolcanics (unit 1) which are overlain by hornblende-biotite gneisses and biotite-hornblende gneisses (unit 2). No chemical data are currently available for them and they shall not be discussed further.

Within the confines of the area underlain by the Shovel Lake Formation (Figure 3), Burns (1951) mapped two metavolcanic units which he considered to be stratigraphically distinct. The following excerpt is taken from Ambrose and Burns (1956, p. 45):

"The rocks within the syncline (Clare River Structure) consist of a lower series of pillow lavas and meta-volcanics, exposed around the nose and along the north-west flank to Hungerford. These are overlain by an alternating succession of thick biotite-amphibole gneisses, some garnet sillimanite gneiss, and thin but remarkably persistent layers of white to grey, coarsely

crystalline limestone which serve as marker beds. This alternating series is overlain by a second thick layer of metavolcanics..."

The distinction of these two volcanic units hinges on the interpretation of the structure as a syncline. The present writer has shown that the structure is in fact a megascopic type III interference pattern (Chapter 3), and that the two volcanic units described by Burns are equivalent.

Distinction of Ortho- and Paragneisses

Field Criteria

It is proposed that much of the area which was formerly mapped as paragneiss is also of volcanic origin. The alternating succession of garnet-sillimanite gneiss and biotite-amphibole gneiss noted above is the result of isoclinal folding rather than stratigraphic intercalation. The garnet-sillimanite gneiss (pelitic schist, unit 16) forms the base of the Beatty Formation of the Flinton Group while the biotite-amphibole gneiss (unit 9) maps as an integral part of the Shovel Lake Formation, along with the amphibolites and hornblende gneisses previously recognized as metavolcanics, and the felsic metavolcanics that were also previously mapped as paragneisses.

It is generally difficult to decide unequivocally the protolith of the biotite-hornblende and hornblende-biotite gneisses of unit 9 on the basis of a single outcrop in the field. Primary features are preserved only locally, and the gneissic foliation is easily mistaken for primary bedding. It was only by mapping this unit over a large area that the writer was able to confidently identify it as being mainly of volcanic origin. The features described in the next paragraphs are indicative of this origin.

An examination of the geological map (Figure 3) shows that the Shovel Lake Formation is internally complicated, consisting of a variety of units which cannot be traced any great distance. In the nose of the Clare River Synform, where the Formation is well exposed in open fields, individual units could be walked for a maximum of only a few hundreds of metres before termination. This lack of continuity of members stands in marked contrast to the continuity over tens of kilometres of the proven metasediments of the overlying Flinton Group, and is a point of argument in favour of the metavolcanic origin of the gneisses of the Shovel Lake Formation. Stauffer et al. (1975) note the same feature in the Amisk Group of Saskatchewan.

In a metasedimentary formation one would expect to find compositional and textural variation within individual outcrops. The rocks of unit 9 are monotonously similar from outcrop to outcrop, consisting of thin, discontinuous gneissic layers which vary slightly in colour index and biotite-hornblende ratio from one to the next. Unit 7, a metasedimentary unit derived in part from a volcanic terrain, is characterized by its heterogeneity, consisting of volcanic pebble metaconglomerate, chert, marble, and a variety of calc-silicate gneisses that are exposed over a width that is small in comparison to that of the volcanically derived gneisses of the Shovel Lake Formation.

The pronounced compositional layering that characterizes unit 9 and some of unit 10 can easily be mistaken for modified sedimentary layering, but careful examination of this layering reveals that it is generally discontinuous (Plate 2e). While it may be argued that this discontinuity is the result of boudinage during deformation, the absence of this feature in sedimentary units above, below, and within the Shovel Lake Formation refutes this argument. It is the view of the writer that

these discontinuous layers are in fact highly flattened fragments of a pyroclastic deposit, an hypothesis that accounts for the discontinuous layering, the general monotony of the succession, and the common intercalation of amphibolites within it.

Finally, the rocks in the northeastern part of the map area are the lateral equivalents of rock units in the nose of the Clare River Synform that are demonstrably flows and pyroclastics on the basis of primary textures and structures (Plate 2).

Chemical Criteria

It is suggested by the above arguments that the biotite-hornblende and hornblende-biotite gneisses of the Shovel Lake Formation are part of a volcanic succession. This assumption was further tested using chemical criteria.

A variety of statistical and chemical techniques exist for determining the protolith of gneisses in the andesite-dacite-arkose compositional range (e.g. Blackburn, 1974; Shaw, 1972; Davoine, 1969). Five of these were attempted for the gneisses of the Shovel Lake Formation (Appendix IV). Although the different techniques often yield conflicting results for the same sample, an examination of Table 10 (Appendix IV) suggests an overall metavolcanic protolith, in agreement with the field data. Yet, an examination of the individual analyses of the Shovel Lake Formation (Tables 4, 5) shows that some of them differ substantially from average analyses of tholeiitic and calc-alkalic volcanic suites (Tables 6, 7). In particular, in the andesite-dacite compositional range, several samples of the Shovel Lake Formation (Table 4) are substantially higher in TiO_2 and lower in CaO than those listed in Table 6. The tests listed in Appendix IV do not satisfactorily discriminate individual analyses of the

Table 4. Major element analyses of rocks of the Shovel Lake Formation

SAMPLE	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	H ₂ O	CO ₂	TiO ₂	P ₂ O ₅	S	MnO	TOTAL
C74-188	72.73	13.71	0.94	2.95	0.93	0.23	5.28	2.69			0.43	0.06	0.01	0.04	100.00
C74-193	44.58	16.62	2.97	9.28	11.46	11.41	1.99	0.51			0.87	0.09	0.02	0.21	100.00
C75-067	70.16	16.18	0.68	2.15	0.78	1.66	6.96	0.81			0.36	0.17	0.01	0.07	100.00
C75-080c	73.13	14.05	0.56	1.78	0.34	1.87	5.07	2.73			0.37	0.04	0.03	0.04	100.00
C75-100	73.20	13.98	0.69	2.18	0.21	1.40	5.93	1.97			0.30	0.09	0.01	0.03	100.00
C75-149	53.24	14.12	2.46	7.69	4.83	9.55	3.32	1.67			2.06	0.87	0.05	0.17	100.00
C75-151	55.25	16.17	2.07	6.47	4.59	7.92	3.90	1.28			1.56	0.67	0.02	0.12	100.00
C75-174a	49.19	14.54	2.54	7.95	12.02	10.08	2.35	0.15			0.87	0.12	0.03	0.17	100.00
C75-200	51.69	14.13	2.92	9.11	7.67	9.43	2.45	0.46			1.60	0.37	0.02	0.17	100.00
C75-222	50.61	13.96	2.56	8.01	10.55	9.90	2.77	0.12			1.08	0.27	0.02	0.16	100.00
C75-350	44.06	14.32	3.94	12.19	10.16	10.72	2.13	0.33			1.59	0.37	0.04	0.18	100.00
C76-043	50.21	12.91	2.88	9.00	8.87	10.87	2.63	0.93			1.36	0.13	0.03	0.20	100.00
JM76-3e	74.86	14.27	0.41	1.36	0.29	0.18	6.48	1.70			0.33	0.03	0.12	0.02	100.00
C75-028	60.20	15.00	4.50	6.80	1.90	3.41	4.10	1.66	0.77	0.20	1.31	0.55	0.00	0.16	100.6
C75-035a	72.60	11.60	1.20	4.40	1.41	3.73	4.00	0.27	0.30	0.40	0.53	0.11	0.07	0.09	100.9
C75-047a	61.40	16.10	4.00	3.60	2.57	4.55	3.80	1.77	0.80	0.10	1.04	0.40	0.02	0.13	100.8
C75-056b	74.10	12.40	1.80	2.20	1.50	1.65	3.70	1.35	0.20	0.00	0.43	0.09	0.02	0.07	100.5
C75-061	57.60	14.00	8.10	6.90	1.67	3.31	3.60	1.93	0.60	0.40	1.82	0.75	0.05	0.16	101.0
C75-080a	67.20	14.60	2.90	2.70	0.93	2.38	5.90	1.19	0.30	0.20	0.68	0.21	0.03	0.06	99.5
C75-080b	58.50	18.60	2.40	4.60	3.29	3.07	4.90	2.76	0.70	0.00	1.13	0.35	0.01	0.14	100.8
C75-086a	60.10	14.60	5.10	2.20	4.49	3.76	3.50	3.51	0.80	0.00	0.87	0.21	0.02	0.12	99.7
C75-098	53.40	17.30	3.80	6.10	4.28	7.05	4.00	1.68	0.70	0.60	1.18	0.18	0.03	0.17	100.9
C75-205	64.40	14.90	3.60	2.40	1.49	3.31	4.60	1.78	0.70	1.20	1.06	0.35	0.01	0.08	100.2
Ch-1	46.56	14.75	2.35	9.85	9.44	10.14	2.29	0.30	L.O.I.	2.69	1.14	0.38		0.09	100.0
Ch-2	49.79	17.44	4.65	5.97	4.56	7.88	3.91	1.63		2.33	1.38	0.34		0.09	100.0
P-27	48.19	15.77	3.35	8.65	6.00	11.20	2.34	0.33	1.87	0.76	1.57			0.12	100.31
P-67	48.70	17.10	3.39	7.12	5.10	7.20	4.27	1.13	1.70	0.61	1.12			0.11	98.62
P-351	49.73	14.74	2.37	9.82	6.62	9.00	3.10	0.75	1.98	0.03	1.39			0.15	99.74
P-78	51.78	14.19	2.51	7.36	4.61	12.10	2.34	0.30	1.27	1.45	1.25			0.13	99.23

Sources: First Group: X-ray fluorescence analyses, Dr. G. R. Holland, Science Laboratories, Durham, England.

Corrected to 100% on a volatile free basis, total iron recalculated to give FeO/Fe₂O₃ of about three to one.

Second Group: X-ray fluorescence analyses, Geological Survey of Canada, courtesy of Dr. J. H. Bourne.

Third Group: Pearson, 1958.

Note: L.O.I.: lost on ignition.

Table 5. Trace element analyses of rocks of the Shovel Lake Formation

SAMPLE	Ba	Nb	Zr	Y	Sr	Rb	Zn	Cu	Ni	Cr	V
C74-188	424	20	529	86	39	43	58	4	14	9	
C74-193	91	1	41	21	198	18	113	5	216	143	
C75-067	107	14	554	44	161	21	69	0	9	3	
C75-080c	974	22	565	74	146	37	11	5	12	15	
C75-100	396	20	545	81	76	31	11	3	12	10	
C75-149	241	14	342	52	303	25	116	137	51	62	
C75-151	472	8	270	38	908	22	126	4	59	79	
C75-174a	84	4	68	21	264	4	96	94	249	504	
C75-200	286	6	163	48	340	7	128	67	47	136	
C75-222	51	5	108	25	433	3	113	8	223	525	
C75-350	84	4	141	41	163	5	142	6	258	620	
C76-043	73	2	82	29	202	22	330	72	71	452	
JM76-3e	462	23	603	79	99	25	19	3	11	15	
C75-028	190		70		140	20	80		80	205	
C75-035a	30		380		70	0	20		80	205	
C75-047a	270		130		430	30	120		130	342	
C75-056b	270		220		100	30	90		80	274	
C75-061	360		270		210	50	90		110	205	
C75-080a	230		250		230	20	30		190	205	
C75-080b	530		200		210	50	100		250	205	
C75-086a	460		130		190	80	60		240	205	
C75-098	240		0		310	40	50		290	547	
C75-205	290		180		200	50	60		200	205	
P-27			420		200	168		21	165	100	590
P-67			370		200	n.d.		40	39	21	580
P-351			470		163	n.d.		86	140	110	860
P-78			430		335	n.d.		59	29	40	410

Source: as in Table 4.

Table 6. Major element abundances in volcanic rocks of island arcs (from Jakes and White, 1972).

	Island arc tholeiites				Calc-alkaline rocks					Shoshonitic		
	Basalt Tholeiitic andesite dacite				High-Al Low-Si Low-K Andesite High-K Dacite Sho-					Latite Abyssal tholeiite		
					basalt	andesite	Low-K andesite	Andesite	High-K andesite		Shonite	
SiO ₂	51.57	57.40	79.20	50.59	54.54	59.05	59.64	58.52	66.80	53.74	59.27	49.34
TiO ₂	.80	1.25	.23	1.05	1.13	.69	.76	.76	.23	1.05	.56	1.49
Al ₂ O ₃	15.91	15.60	11.10	16.29	16.26	17.07	17.38	16.20	18.24	15.84	15.90	17.04
Fe ₂ O ₃	2.74	3.48	.52	3.66	2.31	3.90	2.54	2.93	1.25	3.25	2.22	1.99
FeO	7.04	5.01	.90	5.08	5.40	2.57	2.72	3.28	1.02	4.85	3.19	6.82
MnO	.17	-	-	.17	.12	.15	.09	.09	.06	.11	.10	.17
MgO	6.73	3.38	.36	8.96	6.97	3.25	3.95	4.14	1.50	6.36	5.45	7.19
CaO	11.74	6.14	2.06	9.50	7.50	7.09	5.92	5.59	3.17	7.90	5.90	11.72
Na ₂ O	2.41	4.20	3.40	2.89	3.64	3.80	4.40	3.64	4.97	2.38	2.67	2.73
K ₂ O	.44	.43	1.58	1.07	1.49	1.27	2.04	2.67	1.92	2.57	2.68	.16
P ₂ O ₅	.11	.44	-	.21	.23	.20	.28	.25	.08	.54	.41	.16
H ₂ O	.45	-	-	.81	1.31	.64	1.08	1.47	.26	1.09	1.44	-

Table 7. Trace element abundances in volcanic rocks of island arcs (from Jakes and White, 1972).

	Island arc tholeiites			Calc-alkaline association			Shoshonitic association		
	Basalt	Andesite	Dacite	Basalt	Andesite	Dacite	Basalt	Andesite	Dacite
SiO ₂	52%	58%	63%	52%	58%	63%	52%	58%	63%
Rb	5.0*	6.0	15	10	30	45	75	100	120
Ba	75	100	175	115	270	520	1000	850	900
Sr	200	220	90	330	385	460	700	850	850
Y	-	-	23	20	21	20	-	-	-
Ni	30	20	1	25	18	5	20	-	-
V	270	175	19	255	175	68	200	-	-
Cr	50	15	4	40	25	13	30	-	-
Zr	70	70	125	100	110	100	50	150	200

* Data in p.p.m.

Shovel Lake Formation.

In order to recognize analyses of metasediments (or altered volcanics) and eliminate them from the various chemical plots presented in the next section, individual analyses from the Shovel Lake Formation (Tables 4, 5) were compared with analyses of volcanics from other orogenic areas (Tables 6, 7).

In the range of silica from fifty-six to sixty-five percent, analyses which showed a TiO_2 value greater than one percent coupled with a CaO value of less than three percent were eliminated. On this basis, samples C75-028, C75-061, C75-080b, and C75-205 (Table 4) were eliminated. It should also be noted that the former two are inordinately high in total iron (11.3 and 15 percent respectively) even for volcanic rocks of tholeiitic trend (Table 6). One other analysis is eliminated, C75-086a. Its high MgO and K_2O as compared with CaO and Na_2O , and its high ferric/ferrous ratio, suggest a sedimentary rather than volcanic protolith.

The high TiO_2 values found in the excluded analyses suggest that the samples were taken from volcanically derived epiclastic rocks in which mechanical sorting resulted in the enrichment of the heavy mineral fraction and concomitantly TiO_2 . These data illustrate the need for extreme caution in sampling metamorphosed volcanic rocks for chemical analyses. Flows are to be preferred over meta-tuffs for sampling purposes because recrystallization destroys textural criteria necessary to distinguish tuffs from epiclastics of similar composition.

Metavolcanic Rocks of the Shovel Lake Formation

Field Characteristics

The rest of this chapter will be taken up with the classification of those rocks of the Shovel Lake Formation identified as volcanic. Field

characteristics which bear on this classification will be discussed in this section. For a more complete description of the field characteristics of each unit, the reader is referred to Chapter 2.

In the field, the volcanic rocks of the Shovel Lake Formation were separated on the basis of colour index; amphibolites having a colour index greater than 30, intermediate rocks between 10 and 30, and felsic rocks having a colour index of less than 10. Mixed outcrops were mapped to correspond with their major constituent. It was found that these field divisions of mafic, intermediate, and felsic correspond to basalt, andesite and rhyolite respectively, with dacites overlapping the andesite-rhyolite boundary.

Relict textures and structures are rare. The basaltic members have been metamorphosed to amphibolites consisting of a granoblastic polygonal aggregate of feldspar with hornblende prisms that tend to be aligned parallel to the foliation. Typically they consist of 50 to 60 percent hornblende, 30 to 40 percent plagioclase with accessory quartz, biotite, sphene, opaques, carbonate, and apatite.

Rocks in the intermediate ranges are composed of a mosaic of plagioclase, quartz, and microcline with biotite (10-20%) in slight excess over hornblende (5-15%) as the mafic phases. Sphene, opaques, hematite, epidote, and chlorite are minor constituents. The felsic members are recrystallized aggregates of plagioclase, quartz, and microcline in decreasing abundance with accessory biotite, muscovite, chlorite and opaques.

A rough visual estimate from Figure 3 illustrates that about half of the Shovel Lake Formation is andesite whereas basalt and rhyolite compose about 20 and 30 percent respectively. In terms of areal distribution, basalts are slightly more common in the Sulphide area while

rhyolites are more widely distributed in the northeast. It was found that in the andesite range, pyroclastics are far more abundant than flows; the converse holds true in the basaltic range while the proportion of fragmental and massive rocks in the rhyolitic range are approximately equal.

Owing to the lack of primary facing indicators and to the complexity of the outcrop pattern, which is the result both of primary interfingering of stratigraphic units and two phases of folding, no attempt can reasonably be made at proposing a stratigraphic succession. Something has been gained, however from the careful mapping of the volcanic horizons.

Figure 3 shows that basalt, andesite, and rhyolite are exposed at both the lower and upper contacts of the Shovel Lake Formation. This suggests to the writer that the evolution of these units was contemporaneous rather than sequential, as in the classical upward progression from basalt to andesite to dacite to rhyolite. The writer is cognizant of the fact that these sequential evolutions are usually demonstrated over wider areas and thicker successions, and that the Shovel Lake Formation may represent a portion of such a sequence.

The Shovel Lake Formation is bounded above and below by the Flinton Group and the Todd Lake Formation respectively. This contrasts sharply with the Bishop Corners area to the north where the base of the volcanic succession is not exposed, but available field evidence (Brown et al., 1975) suggests that those volcanics were deposited on a basement comprising mafic and ultramafic plutonic rocks.

Morton (pers. comm, 1977) has suggested that there is an erosional interval in that section, marked by the deposition of calcareous graywackes and marbles and also a change in volcanic regime from uniform subaqueous tholeiitic basalt flows to a more varied calc-alkaline suite ranging from basalts through andesite flows and pyroclastics to dacites.

The diversity of the volcanic rocks in the Shovel Lake Formation and the fact that they are underlain by metasediments suggest that they correlate with the calc-alkaline portion of the suite exposed in the Bishop Corners area.

Little can be said about the stratigraphic thickness of the volcanic rocks because of the lack of data to decipher the precise fold geometry, and the amount of strain within them. A minimum thickness of 1 km can be postulated on the basis of the map pattern illustrated in Figure 3. A more reasonable estimate, taking into account the marked flattening of fragments, is 3 or 4 km.

Geochemical Classification

Because a study of the volcanic chemistry was not initially an integral part of the thesis plan, the sampling distribution and density throughout the area (Figure 18) are less than ideal. The analyses, listed in Tables 4 and 5, are derived from different sources. The first group of 13 analyses were performed by X-ray fluorescence at Science Laboratories, Durham, England under the supervision of Dr. G. H. Holland. They are normalized to a hundred percent on a volatile-free basis and total Fe is recalculated to give an $\text{FeO}/\text{Fe}_2\text{O}_3$ ratio of about three to one. The second group of 10 analyses were kindly provided by Dr. J. H. Bourne of the Geological Survey of Canada. They were performed at the Geological Survey analytical laboratories by X-ray fluorescence and emission spectroscopy, and rapid chemical techniques. The analyses labeled P are from Pearson (1958) and those labeled CH are from Church (1954) corrected to a hundred percent by Pearson.

The metamorphic grade throughout the Clare River area is middle to upper amphibolite facies. The pelites contain kyanite-staurolite in the

northeast and sillimanite in the southwest. At these grades it is commonly assumed that some constituents, especially alkalis are mobile. In spite of this, the assumption will be made that this mobility has not markedly affected the original chemistry. In support of this contention, a comparison between analyses of other volcanic suites (Tables 6 and 7) and the Shovel Lake Formation (Tables 4 and 5) shows that they have not been greatly affected in either major or trace elements.

Common volcanic rocks are broadly classified into subalkaline and alkaline series (Irvine and Baragar, 1971; Miyashiro, 1974) and one of the most widely used methods of discrimination between the two is the alkali-silica plot, shown in Figure 19 for the metavolcanics of the Shovel Lake Formation. It is clear that the rocks fall predominantly into the subalkaline field with only four of the analyses falling into the alkaline field as defined by Irvine and Baragar (Samples C74-193, C75-350, CH-2, and P-67). The trace element compositions of all these samples do not resemble those of alkaline rocks though. In particular, Ba and Sr are significantly lower than those of alkaline basalts (cf. Carmichael et al., 1974, p. 494; Table 8, shoshonitic basalt). It is probable that these samples have been selectively affected by secondary processes.

Subalkaline suites can be classified as tholeiitic or calc-alkaline. Tholeiitic suites, dominated by basalts, are found in ocean basins and continental flood basalt terrains such as the Columbia River Plateau. The calc-alkaline suite is generated at active continental margins and basalts associated with this suite are high in alumina. Both are found in island arcs (Jakes and White, 1972) with tholeiitic suites being more characteristic of young island arcs whereas the calc-alkaline suite characterizes more mature island arcs (Miyashiro, 1974).

The two suites display distinctive differentiation trends.

Figure 19. Alkali-silica plot for metavolcanics of the Shovel Lake Formation. Field divisions after MacDonald, 1968, and Irvine and Baragar, 1971. Closed squares represent metabasalts.

Bowen (1928) recognized the differentiation trend in which SiO_2 , Na_2O , and K_2O increase, while total iron, MgO , and CaO decrease, resulting in the basalt-andesite-dacite-rhyolite association characteristic of orogenic belts. This differentiation has come to be known as the calc-alkaline trend. Fenner (1929) suggested that the residual liquid in a differentiating basaltic magma would be enriched rather than depleted in iron and this trend was documented by Wager and Deer (1939) in their study of the Skaergaard Intrusion of tholeiitic affinity.

Calc-alkaline and tholeiitic suites are commonly distinguished by their trends on an AFM ternary plot. The former show a trend directly away from the FM side of the triangle while the latter initially trend parallel to this axis towards the iron apex. Although the entire suite of the Shovel Lake Formation displays a distinctly calc-alkaline trend in Figure 20, the more mafic members of the suite show a parallel trend to the FM axis and chemically they are tholeiitic basalts.

Irvine and Baragar (1971) obtained good separation of the more mafic members of Cenozoic tholeiitic and calc-alkaline suites by plotting Al_2O_3 against normative plagioclase composition. The underlying assumption here is that the basalts associated with the calc-alkaline suite are high in alumina. Jakes and White (1972) also cite this criterion as characteristic of calc-alkaline basalts. Most of the Shovel Lake basalts plot as tholeiites (Figure 21).

Miyashiro (1973, 1974) has proposed plots to distinguish between tholeiitic and calc-alkalic volcanic suites using the iron-magnesium ratio as an indicator of fractional crystallization. Tholeiites display a lower rate of silica enrichment and a greater increase in iron with increasing fractional crystallization. Based on trends of known belts, Miyashiro has divided calc-alkaline and tholeiitic fields on plots of SiO_2 against

Figure 20. AFM plot for metavolcanics of the Shovel Lake Formation. A: $\text{Na}_2\text{O} + \text{K}_2\text{O}$; F: total iron; M: MgO . Field division after Irvine and Baragar, 1971. Closed squares represent metabasalts.

Figure 21. Plot of Al_2O_3 versus normative plagioclase composition for the metavolcanics of the Shovel Lake Formation. Field division after Irvine and Baragar, 1971. Closed squares represent metabasalts.

FeO^*/MgO and FeO against FeO^*/MgO where FeO^* is total iron expressed as FeO . These lines are shown on Figures 22 and 23 along with data from the Shovel Lake Formation. Miyashiro (1974) points out that the field separations shown in Figure 22 and 23 are useful only in the range of FeO^*/MgO between 2.0 and 5.0. At values below 2.0, calc-alkalic and tholeiitic trend lines intersect, and above 5.0, the trend lines can show considerable curvature.

The slope of the trend for the Shovel Lake volcanics, ignoring these points beyond FeO^*/MgO greater than 5.0, is clearly greater than the slope of the dividing line (Figure 5), and the suite is calc-alkaline in its differentiation trend. Although a number of analyses fall into the field which Miyashiro characterized as tholeiitic, a tholeiitic differentiation trend is not established. The limited data from the Shovel Lake volcanics suggest a continuous differentiation trend, in marked contrast to the sharp break in composition and trend displayed on a similar plot for the suite in the Bishop Corners-Cloyne area (Brown et al., 1975, Figure 3).

The dividing line between the fields illustrated in Figure 23 was derived by Miyashiro (1974) by distinguishing analyses characterized as tholeiitic or calc-alkaline on a SiO_2 vs. FeO^*/MgO plot and drawing the best separation between them. This line of separation is not entirely adequate for the Shovel Lake analyses. Three of the five samples characterized as calc-alkaline in Figure 22 fall close to but on the tholeiitic side of the dividing line in Figure 23.

The variation of titanium with iron-magnesium ratios was also used by Miyashiro (1974) to distinguish calc-alkaline and tholeiitic trends, the former being characterized by decreasing trends and the latter by initial increase then decrease respectively. Figure 24 illustrates the data from the Shovel Lake Formation, which displays a convex downward,

Figure 22. Plot of SiO_2 versus FeO^*/MgO for the metavolcanics of the Shovel Lake Formation, where FeO^* is total iron expressed as FeO . Field division after Miyashiro, 1974. Closed squares represent metabasalts.

Figure 23. FeO^* (total iron expressed as FeO) versus FeO^*/MgO for the metavolcanics of the Shovel Lake Formation. Field division after Miyashiro, 1974. Closed squares represent metabasalts.

Figure 24. Plot of TiO_2 versus FeO^*/MgO for the metavolcanics of the Shovel Lake Formation, where FeO^* is total iron expressed as FeO . Closed squares represent metabasalts.

decreasing trend typical of calc-alkaline suites.

To this point, only the classification of the volcanic rocks of the Shovel Lake Formation has been discussed. In summary, they comprise a subalkaline suite which shows tholeiitic affinities in the mafic members, coupled with an apparently continuous differentiation trend which is characterized as calc-alkaline.

It is widely recognized that subalkaline basalts occur in a variety of present day tectonic environments, most noticeably as ocean floor tholeiites, continental flood basalts, and the foundation of island arcs. A variety of discriminatory diagrams based on major and trace elements have been used to separate these various fields. The discriminant function method of Pearce (1976) was chosen to compare the Shovel Lake basalts to basalts of known tectonic regime.

The reason that this method was chosen is that it utilizes major elements, which are uniformly available for all the rocks analyzed. Trace elements on the other hand were considered less suitable because data were not uniformly available for all the rocks analyzed (Table 5). Of the basalt analyses, five were screened out as outlined in Pearce (1976). In samples C74-193, C74-174a, C75-222, and C75-350, the sum of CaO and MgO is greater than twenty percent, while sample Ch2 shows an $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio in excess of that outlined by Pearce for acceptance in his data.

Figure 25 shows that the basalts from the Shovel Lake Formation fall into the calc-alkaline basalt - low-potassium tholeiite field while Figure 26 which is designed to discriminate between these two types indicates that the basalts more closely resemble modern low-potassium tholeiites, which are associated in the current tectonic framework with the early stages of an island arc on oceanic crust.

Figure 25. Major element discriminant function plot, designed by Pearce, 1976, for the amphibolites of the Shovel Lake Formation. Fields are as follows. WPB: within plate basalt; SHO: shoshonite; CAB: calc-alkaline basalt; LKT: low potassium tholeiite; OFB: ocean floor basalt.

Figure 26. Major element discriminant function plot, designed by Pearce, 1976, for the amphibolites of the Shovel Lake Formation. Fields are as follows. OFB: ocean floor basalt; LKT: low potassium tholeiite; CAB: calc-alkaline basalt; SHO: shoshonite.

Conclusions

Two stratigraphically distinct metavolcanic formations occur in the Clare River Structure. The lower one comprises the lowermost two units of the Otter Creek Formation and does not appear to have an equivalent in the rocks of the Bishop Corners-Cloyne succession to the north as described by Sethuraman and Moore (1973).

The Shovel Lake Formation consists of a series of basalts, andesites, dacites, rhyolites, and minor epiclastic sediments which are bounded above by the Flinton Group and below by marbles of the Todd Lake Formation. On the basis of their stratigraphic position, they correlate with the calc-alkaline suite of volcanic rocks which make up the upper half of the Bishop Corners - Cloyne succession.

Chemically, the suite shows a calc-alkaline trend which appears to be continuous from tholeiitic basalts through to rhyolites. Unfortunately, the volcanic stratigraphy within the Shovel Lake Formation could not be worked out in sufficient detail to correlate stratigraphic position with chemistry. It is clear that basalts and rhyolites appear as lateral equivalents at both the top and bottom of the Shovel Lake Formation, but there are insufficient data to tell whether the chemistry of either of these changed from top to bottom of the succession.

In terms of modern equivalents, the basalts resemble low-potassium tholeiites characteristic of the early stages of formation of island arcs and the progression along the calc-alkaline differentiation trend is suggestive of the maturation of the arc system. Plate tectonic scenarios involving subduction and ocean closure to generate the Bishop Corners volcanic suite have been suggested by several authors (Brown et al., 1975; Moore, 1977; Condie and Moore, 1977; Sethuraman and Moore, 1973). The writer's data are consistent with these models.

8. TECTONIC EVOLUTION OF THE KALADAR-DALHOUSIE TROUGH

Introduction

In this final section, the writer will discuss models of the depositional environment and of the tectonic evolution of the Kaladar-Dalhousie Trough and adjacent areas. The stratigraphic succession, the style and timing of deformation, and the present geographic distribution of the rock associations are important in this discussion. It is clear that the detailed field mapping of the sort presented by Thompson (1972) and in this study must be extended and correlated over a wider area in order to better understand the tectonic evolution of the Kaladar-Dalhousie Trough, the Central Metasedimentary Belt, and ultimately, of the Grenville Province.

Paleogeographic Models

The majority of the supracrustal rocks in the Grenville Province are confined to the Central Metasedimentary Belt (Figure 1) and they have been termed the Grenville Supergroup (Moore and Thompson, 1972; Wynne-Edwards, 1972). The Grenville Supergroup is normally characterized as consisting of pelite, marble, and quartzite. Wynne-Edwards (1972) interprets this association as being typically miogeosynclinal. Infolded remnants of the Grenville Supergroup are found scattered throughout the Grenville Province, and on this basis, Wynne-Edwards (1972) suggests that most, if not all of the Grenville Province, was covered with these platformal sediments and as an extension, that the entire Grenville Province was ensialic at the time of their deposition.

The Hastings Basin and the Kaladar-Dalhousie Trough contain the thickest successions of the Grenville Supergroup, and these are not typical

of the miogeosynclinal rocks exposed elsewhere in the Grenville Province. In the Hastings Basin, Lumbers (1976b) has estimated the thickness of the supracrustal succession to be 8200 m, the majority of which comprises metavolcanic rocks of the Hermon Group (Figure 9). In the Kaladar-Dalhousie Trough, Sethuraman and Moore (1973) have documented a 7000 m succession of tholeiitic and calc-alkalic metavolcanics which belong to the Hermon Group. Lumbers found no evidence of basement to the Hermon Group. Recently, Brown et al. (1975) have found field evidence of basement to the Hermon Group which consists of mafic and ultramafic plutonic rocks. This evidence is more fully discussed in Appendix VI. There is no evidence of sialic basement to the Grenville Supergroup in the Hastings Basin or the Kaladar-Dalhousie Trough.

Lumbers (1967a) suggested that the Hastings Basin evolved as an island arc. This has been corroborated by studies of the volcanic chemistry (Sethuraman and Moore, 1973; Brown et al., 1975; Condie and Moore, 1977). In their lower portion, the volcanics are low potassium tholeiites (Sethuraman and Moore, 1973) that were extruded on oceanic crust. Higher in the section, the volcanics display calc-alkalic trends and are intercalated with marbles, graywackes and volcanic pebble conglomerates. The latter association probably evolved in a warm shallow sea dotted with volcanic islands.

Southeast of the Kaladar-Dalhousie Trough, in the Frontenac Axis and the Adirondack Lowlands (Figure 2), mafic volcanic rocks have not been recognized. The Grenville Supergroup in this area consists of marble, quartzite, and aluminous paragneiss (metamorphosed shale); its thickness has been estimated at 4600 m (Wynne-Edwards, 1967; Engel and Engel, 1953) and cross-bedded quartzites up to 150 m thick have been recorded (Wynne-Edwards, 1967). This association is typical of a platform-type sedimentary accumulation.

The boundary between these contrasting associations coincides with the boundary between the Kaladar-Dalhousie Trough and the Frontenac Axis (Figure 2). No detailed stratigraphic correlations have yet been made between these two areas, but two lines of evidence suggest that the pre-Flinton succession in the Hastings Basin and Kaladar-Dalhousie Trough is correlative with the platform-type sediments of the Frontenac Axis. First, with the exception of the Flinton Group in the Kaladar-Dalhousie Trough, the supracrustal rocks of each of the areas are intruded by granitic and granodioritic plutons. Secondly, in looking for a likely source of the quartzite pebbles of the Flinton conglomerate, Thompson (1972) suggested that they were derived from the quartzites of the Frontenac Axis.

In any attempt to reconstruct actualistic models of the Central Metasedimentary belt, the contrasting environments discussed above must be taken into account. The boundary between the two areas runs roughly north-south (J. M. Moore, pers. comm. 1977). If it corresponds to a shelf-slope boundary, then the distribution of land masses at the time of the deposition of the pre-Flinton Grenville Supergroup was such that a major land mass, from which the detritus for the quartzites of the Frontenac Axis was derived, lay to the east of the exposed Grenville Province, in what is now the Appalachian Orogen. Radiometric evidence (Krogh and Davis, 1969, 1971) suggests that another older land mass lay northwest of Bancroft during the evolution of the Hastings Basin and Kaladar-Dalhousie Trough.

Plate Tectonic Models

Recently, attempts have been made to explain the Grenville Province in the framework of plate tectonics (Baer et al., 1974). Two approaches have been taken: the first attempts to account for the evolution of the entire Grenville Province in terms of a model (Dewey and

Burke, 1973; Irving et al., 1974; Wynne-Edwards, 1976); the second attempts to explain geological history of specific areas in the Grenville Province using actualistic models (Brown et al., 1975; Baer, 1976).

Dewey and Burke (1973) interpret the Grenville Province as a basement reactivation zone, the result of a continental collision. They view the Tibetan Plateau as a modern analogue exposed at a higher level.

Irving et al. (1974) draw on paleomagnetic evidence to suggest that the Grenville Province moved from the position of North Africa to its present position, suturing to the rest of the Canadian Shield (Laurentia) at about 1000 m.y. Wynne-Edwards (1976) holds that the Grenville Province evolved as an ensialic orogenic belt. In the "millipede model", he envisions continental crust becoming thinned and heated as it passes over mantle hot spots, and then deforming as a result of the compression generated by its own momentum.

These models are important in that they provide a framework in which to interpret more detailed studies of the geology of the Grenville Province. There is an inherent danger in using them though, namely that details which do not agree with the model may be overlooked.

The second approach, that of interpreting a sequence of events in a specific area and applying actualistic models, has been applied mainly in the Central Metasedimentary Belt of the Grenville Province.

Baer (1976) notes that the Central Metasedimentary Belt is roughly bounded by two lineaments; the northeast trending McArthur's Mills (also called Bancroft-Renfrew) lineament on the northwest, and the north trending Chibougamau-Gatineau lineament on the east. He interprets these as the boundaries of an aulacogen in which the Grenville Supergroup was deposited. This model does not account for the volcanic chemistry, nor the tectonic style observed in the area.

Lumbers (1967a) suggested that the Hastings Basin and the Kaladar-Dalhousie Trough evolved as an island arc. Detailed studies of the geochemistry of the Tudor volcanics (Sethuraman and Moore, 1973; Condie, 1975; Condie and Moore, 1977) indicate that they are the products of island-arc volcanism, corroborating the interpretation of Lumbers (1967a). An actualistic model involving island arc generation, followed by continental collision (Brown et al. 1975), explains both the geochemistry of the volcanic rocks and the polyphase deformation which affected the area. Further work is necessary to establish the position of the suture, the extent of the island arc suite, and the nature of its boundaries with non-volcanic areas.

The Role of Diapirism and Emplacement of the Plutonic Bodies

"Diapir" and "diapirism" are terms which have sometimes been misused in the literature. Classically, the terms diapir and diapiric structure were applied to features formed when mobile rocks are considered to have been injected into weak zones in overlying rocks (A.G.I. Glossary of Geology). The term diapirism is used to describe the process. Diapirism does not imply a specific mechanism. The American Geological Institute Glossary of Geology states that a diapir may be formed either by geostatic load (gravity) or tectonic stresses (lateral compression). Since the model work of Ramberg (1963), diapirs and diapirism are often implicitly attributed to a gravity mechanism. In the following discussion, the terms diapir and diapirism are used in the classical context.

Appleyard and Stott (1975) unduly generalize when they suggest that all the trondhjemite plutons of Lumbers' (1967b) biotite-diorite series are diapirs. There is ample evidence to show that the Elzevir Batholith was emplaced in fluid state. Dykes of the Elzevir Batholith intrude the

Tudor volcanic rocks and intrusive breccias are common along its margin. Similar features are observed on the southern margin of the Weslemkoon Pluton. Only if all igneous intrusions are taken to have formed diapirically (Stephansson, 1975) can Appleyard and Stott's (1975) suggestion that the trondhjemites of the biotite-diorite series are diapirs, be accepted.

Schwerdtner (oral comm., 1974) suggested that the Clare River Structure is an interdiapiric depression. This model is based on the strain patterns displayed by the Addington Pluton, which are deduced by the relative strength of L versus S fabrics. The Addington Pluton in the core of the Clare River Structure shows a strong linear fabric while on the limbs, it displays a planar fabric. Schwerdtner interpreted this as being due to inhomogeneous strain during a continuous episode of diapiric intrusion. The writer and others (R. L. Brown, J. M. Moore) examined these fabrics in the field, and concluded that the lineation in the Addington Pluton was the result of crenulation of the pre-existing S1 planar fabric that occurs on the limbs. The crenulation of S1 to produce a strong L2 is a common feature in the Clare River Structure.

The granite bodies which define the major F2 antiforms in the Kaladar-Dalhousie Trough were clearly emplaced prior to the D1 event which affected the Flinton Group. Two lines of evidence underlie this fact. First, unconformable relationships between these plutonic bodies and the Flinton Group have been documented in several areas (Thompson, 1972; Psutka, 1976), and they must, therefore, predate the deposition of the Flinton Group. Secondly, the plutonic bodies contain D1 fabric elements which are deformed by F2 folds.

It has been suggested (Schwerdtner, oral comm., 1974) that gravity driven upwelling of the plutonic bodies is responsible for the

structural pattern in the Kaladar-Dalhousie Trough. Clearly, this cannot apply to the initial emplacement of the plutons, as they predate the majority of the deformation in the Kaladar-Dalhousie Trough. If a gravitational mechanism is to be proposed for this deformation it must occur during a later remobilization of the granitic bodies. With the exception of Schwerdtner et al. (1977), there are no detailed studies of strain patterns in the plutons and adjacent supracrustal rocks to evaluate this possibility. As discussed previously, the writer's interpretation of lineation patterns in the Addington Pluton differs from that of Schwerdtner et al. (1977). The fact that the plutons of the Kaladar-Dalhousie Trough are strongly elongated in a northeasterly direction, and not circular or oval as is the case of classical examples of gravity emplaced structures, suggests that gravitational upwelling was not responsible for the deformation in the Kaladar-Dalhousie Trough.

With the exception of the Addington Pluton, the major plutonic bodies of the Kaladar-Dalhousie Trough have not been incorporated into the overlying supracrustal rocks by the long-limbed reclined to recumbent folds which characterize the D1, although they do contain D1 fabric elements. This feature is tentatively ascribed to the differing responses of the layered supracrustal rocks and the uniform underlying granitic rocks to the effects of the first phase of deformation. Clearly, more detailed studies of the effects of deformation on the plutonic rocks of the Kaladar-Dalhousie Trough are warranted.

The Addington Pluton differs from the other plutons of the area in that it appears to sit as a wedge in the supracrustal rocks (Figure 14). Its initial emplacement must predate the deposition of the Flinton Group; it contains D1 structures, and is locally overlain by granite clast conglomerate which has been strained during D1. It is thought to have been

emplaced into the supracrustals as a fault slice during the early stages of D1.

A Model of Tectonic Evolution

The oldest rocks recorded in the Central Metasedimentary Belt are rhyolites of the Tudor formation, the lowest member of the Hermon Group, dated at 1310[±]15 m.y. (Silver and Lumbers, 1966). Locally exposed field relations (Brown et al., 1975) and volcanic chemistry (Condie, 1975; Condie and Moore, 1977; Brown et al., 1975; Sethuraman and Moore, 1973) suggest that the Tudor formation was deposited on simatic crust and represents the commencement of the formation of an island arc. Actualistically, this can be ascribed to the initiation of an ensimatic subduction zone. The Tudor formation which consists mainly of tholeiitic basalts, gives way to a more varied suite of calc-alkalic volcanic rocks (Sethuraman and Moore, 1973; Brown et al., 1975; Condie and Moore, 1977) as the island arc matures. The emplacement of granodiorite bodies at depth is probably coupled with the intrusion of the calc-alkalic suite.

The earliest diastrophic event in the Kaladar-Dalhousie Trough, recorded by an east-trending foliation preserved in metavolcanics west of Flinton (Figure 13) is believed to be related to the emplacement of the Elzevir Batholith.

The Elzevir Batholith shows features associated with mesozonal emplacement. It possesses a contact aureole that is preserved locally and foliation in the Tudor volcanics is concordant with its margin. Subsequent to the formation of this foliation, it was disrupted by dykes of Elzevir and locally, the volcanics are brecciated.

The Elzevir Batholith is a member of the biotite-diorite series, dated at 1250[±]25 m.y. (Silver and Lumbers, 1966). Other members of this

series include the Northbrook, Cross Lake, Mellon Lake, Weslemkoon, and Hinchinbrook Plutons. The crystallization date of the Hinchinbrook Pluton is 1277 m.y. (Wallach, 1974). These plutons were intruded into mafic volcanic rocks and associated sediments of the Hermon Group, but unfortunately, with the exception of the Elzevir Batholith, they were remobilized during deformation of the Flinton Group, destroying most evidence of their intrusion.

The plutons described above represent the first evidence of sialic basement in the Central Metasedimentary Belt. Although it has been suggested that they are the result of the remobilization of a pre-existing older basement (Wynne-Edwards, 1967), the writer is unaware of any field or isotopic evidence of the sialic basement to the Tudor formation, the oldest rocks in the Kaladar-Dalhousie Trough and the Hastings Basin.

Following the emplacement of the large grandiorite plutons into the rocks of the Hermon Group, the entire area was uplifted, unroofing of the plutons occurred, and resulted in the deposition of the Flinton Group.

After deposition of the Flinton Group, the entire area was buried, metamorphosed, and subjected to two major phases of deformation. The style observed in the Kaladar-Dalhousie Trough - reclined to recumbent D1 structures followed by upright D2 structures - is also recorded in the Hastings Basin (Carmichael, 1968; Divi and Fyson, 1973) and the Adirondack Lowlands (Foose and Carl, 1977). Fyson (1971) has pointed out that this style is characteristic of many paleozoic orogenic belts.

In the Kaladar-Dalhousie Trough, the fact that there is only one major metamorphic culmination (Hounslow and Moore, 1967; Sethuraman and Moore, 1973; Thompson, 1972; Rivers, 1976, and this study) suggests that the two phases of deformation occurred during one episode of burial during which stress conditions changed.

During the first phase of deformation, a pervasive S1 foliation was developed and deformed pebbles in metaconglomerates suggest flattening by a factor of three to four times. The peak of metamorphism roughly coincides with the end of D1. Pegmatites associated with the metamorphism were emplaced at 1050 ± 20 m.y. (Silver and Lumbers, 1966). The metamorphic grade is thought to have remained high during the second phase of deformation (Thompson, 1972; Rivers, 1976), which resulted in the upright F2 folds that give the region its northeast-trending structural grain. Map scale warps around northwest-trending axial surfaces and mesoscopic warping about subhorizontal axes occurred. No metamorphism was associated with these events.

It is believed that the metamorphism and deformation of the Flinton Group took place during the closing of the ocean in which the Tudor volcanics were extruded. More work is necessary to define the position of the suture and the nature of the boundaries between the areas where volcanics of the Hermon Group are exposed (Hastings Basin, Kaladar-Dalhousie Trough) and adjacent areas. The tectonic evolution of Kaladar-Dalhousie Trough is summarized in Table 8.

Table 8. Tectonic evolution of the Kaladar-Dalhousie Trough

Age	Event
500±m.y.	-deposition of basal Nepéan sandstone
	-uplift and prolonged erosion
950±m.y.	-final cooling of the Precambrian complex
1050±20 m.y.	-intrusion of pegmatites
	-D2: reorientation of shallowly-dipping D1 structures by upright tight to isoclinal F2 folds
	-metamorphism spanning upper greenschist to upper amphibolite facies
	-D1: generation of major reclined to recumbent F1 folds, formation of pervasive S1 foliation, formation of nappes and decollement structures
	-deposition of the Flinton Group in a shallow marine environment
	-uplift and erosion of the basement complex
1250±25 m.y. ¹	-intrusion of plutons of the biotite diorite series.
1277±m.y. ²	-Greenschist grade metamorphism and structural features related to this intrusive episode are preserved locally.
	-extrusion of calc-alkalic volcanic rocks; deposition of clastic sediments and marbles in a shallow marine environment; development of intraformational unconformities
1310±15 m.y.	-extrusion of low-K tholeiitic basalts on oceanic crust

¹ Silver and Lumbers, 1966² Wallach, 1974

9. SUMMARY OF CONCLUSIONS

The Clare River Structure is one of several supracrustal belts which occupy synformal depressions between flanking plutons in the Kaladar-Dalhousie Trough (Figure 2). By detailed mapping of that portion of the Clare River Structure southwest of Highway 41 (Figure 3), the writer was able to work out its structural geometry and stratigraphic succession.

The supracrustal rocks of the Clare River Structure can be divided into five formations. They are as follows:

Otter Creek Formation: massive felsic metavolcanics, biotite-hornblende gneisses and schists, quartzofeldspathic psammite and pelite, rusty psammite, and calcareous marble.

Todd Lake Formation: dolomitic marble; volcanic pebble metaconglomerate, metachert, para-amphibolite, calc-silicate paragneiss.

Shovel Lake Formation: mafic, intermediate, and felsic metavolcanics, calc-silicate paragneisses.

Beatty Formation: pelitic schist, rusty psammite.

Bogart Formation: calcareous marble, dolomitic marble, and various calc-silicate paragneisses.

The Beatty and Bogart Formations are correlative with the Flinton Group, as defined by Moore and Thompson (1972) and the Otter Creek, Todd Lake, and Shovel Lake Formations are part of the Hermon Group, as defined by Lumbers (1967b). The lower three formations are thought to have accumulated in a shallow marine environment characterized by active volcanism and carbonate deposition. The Beatty and Bogart Formations were also deposited in a shallow marine environment, but appear to postdate active volcanism.

The outcrop pattern displayed by the rock units of the Clare River Structure is that of a megascopic type III interference pattern, which is accounted for by two major phases of deformation. D1 produced reclined to recumbent northeast-trending folds which were refolded by northeast-trending tight to isoclinal, upright F2 folds. The Hungerford Antiform and the Clare River Synform are major F1 and F2 folds respectively; together they constitute the Clare River Structure. D3 warps the entire structure around northwest-trending axes.

D1 produced a penetrative foliation (S1) defined by both mineral orientation and compositional layering. Long limbed F1 isoclinal folds were observed in only a few locations. L1 is not observed in most of the Clare River Structure. D2 resulted in a penetrative lineation, L2, locally abundant F2 minor folds which vary considerably in style, and a weak S2 crenulation cleavage. Later minor structures occur locally.

The supracrustal rocks of the Clare River Structure lie in fault contact on the Addington Pluton, as indicated by the marked attenuation of rock units against it, on the southeast margin of the structure. This fault was formed during D1 and folded into a synform during D2. It can be classified as a decollement surface. Because the Clare River supracrustals formed a low angle fault contact with the underlying Addington Pluton during D1, the Clare River Structure can be classified as a nappe. Stratigraphic correlation with adjacent areas, which suggests that most of the structural succession is inverted, further supports this interpretation.

Diastrophic events recorded in the Clare River Structure can be correlated with those described in adjacent areas. On a regional scale, D1 generated large scale, northeast-trending recumbent structures which were refolded into upright orientations during D2. D3 resulted in warps around northwest-trending axes, which are visible on a regional scale.

This sequence of events occurred during one continuous phase of metamorphism, documented by the writer and others (Thompson, 1972; Rivers, 1976; Hounslo and Moore, 1967; and Sethuraman, 1970). In the Clare River Structure, the grade of metamorphism is amphibolite facies. Three zones, kyanite-staurolite, sillimanite-staurolite, and sillimanite, can be delineated (Figure 17). Porphyroblast growth took place mainly after D1 in the kyanite-staurolite zone, but sillimanite grows in the S1 foliation plane in the other two zones. The metamorphic peak was syn- to post D1. Metamorphic grade during D2 was lower; micas are strained around the hinges of F2 folds.

The metavolcanic rocks of the Shovel Lake Formation can be classified as subalkaline. The amphibolites show tholeiitic affinities and the intermediate and felsic rocks display a poorly defined calc-alkaline differentiation trend. More data are needed to substantiate this trend. The basalts of the formation chemically resemble low potassium tholeiites, which are characteristic of the early stages of island arc evolution.

The structural style, stratigraphy, and volcanic chemistry observed in the Clare River Structure and throughout the Kaladar-Dalhousie Trough are not only consistent with, but taken as a whole, strongly support the plate tectonic model for the evolution of the Central Metasedimentary Belt of the Grenville Province proposed by Brown et al. (1975).

REFERENCES

- Ambrose, J.W. and Burns, C.A. 1956. Structures in the Clare River Syncline: A demonstration of granitization. In: The Grenville Problem (J.E. Thompson ed.), Royal Society of Canada Special Publication Number 1, pp. 42-52.
- American Geological Institute Glossary of Geology. (M. Gary, R. McAfee Jr., C.L. Wolfe eds.). Washington, D.C. 1972. 805 pp.
- Appleyard, E.C. 1974. Basement/cover relationships within the Grenville Province of eastern Ontario. Canadian Journal Of Earth Sciences, V. 11, pp. 369-379.
- Appleyard, E.C. and Stott, G.M. 1975. Grenville gneisses in the Madawaska Highlands, eastern Ontario. In: Guidebook, Part A, Precambrian Geology (P.G. Telford ed.): Geological Association of Canada, Waterloo, pp. 26-62.
- Baer, A.J. 1976. The Grenville Province in Helikian times: a possible model of evolution. Philosophical Transactions of the Royal Society of London, V. 280, pp. 499-515.
- Baer, A.J., Emslie, R.F. and Tanner, J.G. 1974. Grenville geology and plate tectonics. Geoscience Canada, V. 1, no. 3, pp.54-61.
- Blackburn, W.H. 1974. Origin of the cordierite-bearing gneisses of the Ganonoque-Westport area, Ontario. Annales de la Société Géologique de Belgique, V. 97, pp.297-309.
- Bowen, N.L. 1928. The Evolution of Igneous Rocks. Princeton University Press, Princeton, New Jersey. 334 pp.
- Brocoum, S.J. 1971. Structural and metamorphic history of the major Precambrian gneiss belt of the Hailesboro-West Fowler-Balmat area, Adirondack Lowlands, New York. Unpublished Ph.D. thesis, Columbia University, New York.
- Brown, R.L., Chappell, J.F., Moore, J.M., Jr., and Thompson, P.H. 1975. An ensimatic island arc and ocean closure in the Grenville Province of southeastern Ontario. Geoscience Canada, V. 2, no. 3, pp. 141-144.
- Burns, C.A. 1951. The Clare River Area, Southeastern Ontario. Unpublished M.Sc. thesis, Queens University, Kingston.
- Carmichael, D.M. 1968. Structure and progressive metamorphism in the Whetstone Lake area, Ontario, with emphasis on the mechanisms of prograde metamorphic reactions. Unpublished Ph.D. thesis, University of California, Berkeley.
- Carmichael, I.S.E., Turner, F.J., and Verhoogen, J. 1974. Igneous Petrology. McGraw-Hill, New York. 739 pp.
- Cermignani, C. 1971. Geochemical study of the marbles in the Tweed area. Unpublished M.Sc. thesis, University of Toronto, Toronto.

Church, J.F. 1954. Study of amphibolites from Sulphide, Ontario. Unpublished M.Sc. thesis, University of New Brunswick.

Cloos, E. 1946. Lineation: a critical review and annotated bibliography. Geological Society of America, Memoir 18, 122 p.

Condie, K.C. 1975. Geochemistry of Proterozoic volcanic rocks from the Grenville Province, eastern Ontario. Abstracts with programs. Geological Association of Canada Annual Meeting, Waterloo, Ontario. p. 739.

Condie, K.C. and Moore, J.M. 1977. Geochemistry of Proterozoic volcanic rocks from the Grenville Province, eastern Ontario. In: Volcanic regimes in Canada (W.R.A. Baragar, L.C. Coleman, and J.M. Hall, eds.), Geological Association of Canada, Special Paper Number 16, pp. 149-168.

Currie, A.L. 1972. Analysis of structures in a part of the Clare River Synform, Ontario. Unpublished M.Sc. thesis, University of Toronto, Toronto.

Davoine, P. 1969. La distinction géochimique ortho-para des leptynites. Société Française de Mineralogie et de Cristallographie, Bulletin. v. 92, pp. 59-75.

Dewey, J.F. and Burke, K.C.A. 1973. Tibetan, Variscan, and Precambrian basement reactivation: products of continental collision. Journal of Geology, v. 81, pp. 683-692.

Divi, R.D. and Fyson, W.K. 1973. Folds and strain in Grenville metamorphic rocks, Bancroft, Ontario, Canada. Geological Society of America Bulletin, v. 84, pp. 1607-1618.

Dixon, J.M. 1975. Finite strain and progressive deformation in models of diapiric structures. Tectonophysics, v. 28, pp. 89-124.

Elliott, D. 1972. Deformation paths in structural geology. Geology Society of America Bulletin, v. 83, pp. 2621-2638.

Engel, A.E.J. and Engel, C.E. 1953. Grenville series in the northwest Adirondack Mountains, New York. Geological Society of America Bulletin, v. 64, pp. 1013-1097.

Fenner, C.N. 1929. The crystallization of basalts. American Journal of Science, 5th series, v. 18, pp. 225-253.

Fitzpatrick, M.M. 1950. A gravitational study of the Clare River Syncline area, Ontario. Transactions of the Royal Society of Canada, v. 44, section 4, pp. 21-34.

Foose, M.P. and Carl, J.D. 1977. Setting of alaskite bodies in the northwestern Adirondacks, New York. Geology, v. 5, pp. 77-80.

Fyson, W.K. 1971. Fold attitudes in metamorphic rocks. American Journal of Science, v. 270, pp. 373-382.

Harding, W.D. 1944. Geology of Kaladar and Kennebec Townships. Ontario Department of Mines, Annual Report v. 51, 1942, pt. 4, pp. 51-74.

- Hewitt, D.F. 1956. The Grenville region of Ontario. In: The Grenville problem (J.E. Thompson ed.). Royal Society of Canada Special Publication Number 1, pp. 22-41.
- Hewitt, D.F. 1964. Geological notes for maps nos. 2053 and 2054, Madoc-Gananoque area. Ontario Department of Mines Geological Circular No. 12, 33 pp.
- Hounsflow, A.W. and Moore, J.M. 1967. Chemical petrology of Grenville schists near Fernleigh, Ontario. *Journal of Petrology*, v. 8, pp. 1-28.
- Irvine, T.N. and Baragar, W.R.A. 1971. A guide to the chemical classification of the common volcanic rocks. *Canadian Journal of Earth Sciences*, v. 8, pp. 523-548.
- Irving, E., Emslie, R.F., and Ueno, H. 1974. Upper Proterozoic paleomagnetic poles from Laurentia and the history of the Grenville Structural Province. *Journal of Geophysical Research*, v. 79, pp. 5491-5502.
- Jakes, P. and White, A.J.R. 1972. Major and trace element abundances in volcanic rocks of orogenic areas. *Geological Society of America Bulletin*, v. 83, pp. 29-40.
- Kamilli, D.C. 1974. Progressive metamorphism of the Cross Lake granodiorite, Ontario, Canada. *Geological Society of America Bulletin*, v. 85, pp. 1755-1762.
- Krogh, T.E. and Davis, G.L. 1969. Old isotope ages in the northwestern Grenville Province, Ontario. In: Age relations in high grade metamorphic terrains. (H.R. Wynne-Edwards ed.) *Geological Association of Canada Special Paper Number 5*, pp. 189-192.
- Krogh, T.E. and Davis, G.L. 1971. The Grenville Front interpreted as an ancient plate boundary. *Carnegie Institute of Washington Yearbook* 70, pp. 239-240.
- Krogh, T.E. and Hurley, P.M. 1968. Strontium isotope variation and whole-rock isochron studies, Grenville Province of Ontario. *Journal of Geophysical Research*, v. 73, pp. 7107-7125.
- Logan, W.E. 1846. Geological Survey of Canada, Reports of Progress, 1845-1846. pp. 40-51.
- Lumbers, S.B. 1967a. Stratigraphy, plutonic activity, and metamorphism of the Ottawa River remnant in the Bancroft-Madoc area of the Grenville Province of southeastern Ontario. Unpublished Ph.D. thesis, Princeton University, Princeton, New Jersey.
- Lumbers, S.B. 1967b. Geology and mineral deposits of the Bancroft-Madoc area. In: Guidebook - Geology of parts of eastern Ontario and western Quebec (S.E. Jenness ed.). *Geological Association of Canada*, Kingston, pp. 13-29.
- MacDonald, G.A. 1968. Composition and origin of Hawaiian lavas. *Geological Society of America, Memoir* 116, pp. 477-522.

- MacDonald, G. 1970. Geology of the Kingston area; a compilation. Carleton University Geological Paper, 70-2.
- Meen, V.B. 1944. Geology of the Grimsthorpe-Barrie area. Ontario Department of Mines, Annual Report, v. 51, 1942, pt. 4, pp. 1-50.
- Miller, W.G. and Knight, C.W. 1914. The pre-Cambrian geology of southeastern Ontario. Ontario Bureau of Mines Annual Report, v. 22, 1913, pt. 2. 151 pp.
- Mitchell, W.J. 1974. Stratigraphy and structure of the Kaladar metasediments. Unpublished B.Sc. thesis, Carleton University, Ottawa.
- Miyashiro, A. 1961. Evolution of metamorphic belts. Journal of Petrology, v. 2, pp. 277-311.
- Miyashiro, A. 1973. The Troodos Ophiolitic Complex was probably formed in an island arc. Earth and Planetary Science Letters, v. 19, pp. 218-224.
- Miyashiro, A. 1974. Volcanic rock series in island arcs and active continental margins. American Journal of Science, v. 274, pp. 321-355.
- Moore, J.M., Jr. 1967. Grenville structure, stratigraphy, and metamorphism, southeastern Ontario; Northbrook-Ompah area. In: Guidebook - Geology of parts of eastern Ontario and western Quebec (S.E. Jenness ed.). Geological Association of Canada, Kingston, pp. 59-71.
- Moore, J.M. and Morton, R.L. 1978. Geology, Clarendon Lake area, Ontario. (in editorial hands).
- Moore, J.M., Jr. and Thompson, P.H. 1972. The Flinton Group, Grenville Province, eastern Ontario, Canada. In: Twenty-Fourth International Geological Congress, Section 1, Montreal, pp. 221-229.
- Park, R.G. 1969. Structural correlation in metamorphic belts. Tectonophysics, v. 7, pp. 323-338.
- Pearce, J.A. 1976. Statistical analysis of major element patterns in basalts. Journal of Petrology, v. 17, pp. 15-43.
- Pearson, G.R. 1958. Granitic Gneisses around the Clare River Syncline, Hastings and Lennox and Addington Counties, Ontario. Unpublished Ph.D. thesis, Queens University.
- Psutka, J.F. 1976. Provenance of the plutonic pebbles in the Kaladar metaconglomerate and geology of the associated metasediments. Unpublished B.Sc. thesis, Carleton University, Ottawa.
- Ramberg, H. 1963. Experimental study of gravity tectonics by means of centrifuged models. Bulletin of the Geological Institutions of the University of Uppsala, v. XLII, pp. 1-97.
- Ramsay, J.G. 1967. Folding and Fracturing in Rocks. McGraw-Hill, New York, N.Y., 568 pp.

- Rast, N. 1958. Metamorphic history of the Schichallion Complex. Transactions of the Royal Society of Edinburgh, v. 63, pp. 413.
- Rivers, C.J.T. 1976. Structures and textures of metamorphic rocks, Ompah area, Grenville Province, Ontario. Unpublished Ph.D. thesis, University of Ottawa, Ottawa, Ontario.
- Schwerdtner, W.M., Bennett, P.J., and Jones, T.W. 1977. Application of L-S fabric scheme to structural mapping and paleostrain analysis. Canadian Journal of Earth Sciences, v. 14, pp. 1021-1032.
- Sethuraman, K. 1970. Petrology of Grenville metavolcanic rocks in the Bishops Corners-Donaldson Area, Ontario. Unpublished Ph.D. thesis, Carleton University, Ottawa.
- Sethuraman, K. and Moore, J.M., Jr. 1973. Petrology of metavolcanic rocks in the Bishop Corners - Donaldson Area, Grenville Province, Ontario. Canadian Journal of Earth Sciences, v. 10, pp. 589-614.
- Shaw, D.M. 1972. The origin of the Apsley Gneiss, Ontario. Canadian Journal of Earth Sciences, v. 9, pp. 18-35.
- Silver, L.T. and Lumbers, S.B. 1966. Geochronological studies in the Bancroft - Madoc area of the Grenville Province, Ontario, Canada. (abstract). Geological Society of America, Special Paper 87, p. 156.
- Smith, B.L. 1958. Geology of the Clarendon - Dalhousie Area, Ontario. Department of Mines, Annual Report, v. 65, 1956, pt. 7, pp. 1-46.
- Smith, B.L. 1969. The Ompah Syncline in relation to the Grenville stratigraphy of eastern Ontario. In: Age relationships in high-grade metamorphic terrains (H.R. Wynne-Edwards ed.). Geological Association of Canada Special Paper Number 5, pp. 225-228.
- Stauffer, M.R., Mukherjee, A.C., and Koo, J. 1975. The Amisk Group: An Aphebian (?) Island Arc Deposit. Canadian Journal of Earth Sciences, v. 12, pp. 2021-2035.
- Spry, A. 1963. The chronological analysis of crystallization and deformation of some Tasmanian Precambrian rocks. Geological Society of Australia Journal, v. 10, pp. 193-208.
- Spry, A. 1969. Metamorphic Textures. Pergamon Press, Oxford. 350 pp.
- Stephansson, O. 1975. Polydiapirism of granitic rocks in the Svecofennian of central Sweden. Precambrian Research, v. 2, pp. 189-214.
- Talbot, C.J. 1974. Fold nappes as asymmetric mantled gneiss domes and ensialic orogeny. Tectonophysics, v. 24, pp. 259-276.
- Thompson, P.H. 1972. Stratigraphy, Structure and metamorphism of the Flinton Group in the Bishop Corners - Madoc Area, Grenville Province, Eastern Ontario. Unpublished Ph.D. thesis, Carleton University, Ottawa.
- Turner, F.J. 1968. Metamorphic Petrology. McGraw-Hill, New York. 403 pp.

Venkitasubramanian, C.S. 1969. Large-scale superimposed folds in Precambrian rocks of the Actinolite - Kaladar area, southern Ontario. Unpublished Ph.D. thesis, Queen's University.

Wager, L.R. and Deer, W.A. 1939. The petrology of the Skaergaard intrusion, Kangerdlugssuaq, East Greenland. Meddel. om Grønland, v. 105, no. 4, 352 pp.

Wallach, J.L. 1974. Origin of the Hinchinbrooke Gneiss and its age relationship to the Grenville Group rocks of southeastern Ontario. Abstract. Geological Association of Canada Annual Meeting, St. John's, Nfld. p. 96.

Wilson, M.E. 1933. The Clare River Syncline. Transactions of the Royal Society of Canada, v. 27, section 4, pp. 7-11.

Wilson, M.E. 1940. Map 559A, Madoc Area. Geological Survey of Canada, Ottawa.

Wilson, M.E. 1956. Structures in the Clare River Syncline: A demonstration of granitization. Discussion. In: The Grenville problem (J.E. Thompson ed.), Royal Society of Canada Special Publication Number 1, p. 53.

Winkler, H.G.F. 1974. Petrogenesis of Metamorphic Rocks. Third Edition. Springer-Verlag, New York. 320 pp.

Wolff, J.M. 1978. Kaladar Area, Frontenac and Lennox and Addington Counties, Southern Ontario. Ontario Geological Survey Preliminary Map P. 1563, Geological Series. Scale 1:15840. Geology 1977.

Wynne-Edwards, H.R. 1967. Westport map area, Ontario, with special emphasis on the Precambrian rocks. Geological Survey of Canada, Memoir 346.

Wynne-Edwards, H.R. 1972. The Grenville Province. In: Variations in tectonic styles in Canada (R.A. Price and R.J.W. Douglas eds.). Geological Association of Canada Special Paper Number 11, pp. 263-334.

Wynne-Edwards, H.R. 1976. Proterozoic ensialic orogenesis: the millipede model of ductile plate tectonics. American Journal of Science, v. 276, pp. 927-953.

Zwart, H.J. 1960. Relations between folding and metamorphism in the Central Pyrenees, and their chronological succession. Geologie en Mijnbouw, v. 39, pp. 163-180.

APPENDIX I: Mineralogy of Thin Sections Examined

List of mineral abbreviations:

ac	actinolite	mi	microcline
ap	apatite	mu	muscovite
bi	biotite	oam	orthoamphibole
cc	calcite	op	opaques
ch	chlorite	opx	orthopyroxene
cl	clay minerals	pc	plagioclase
cm	cummingtonite	ph	phlogopite
cz	clinozoisite	qz	quartz
di	diopside	sc	scapolite
do	dolomite	si	sillimanite
ep	epidote	sp	sphene
fh	ferrohastingsite	st	staurolite
ga	garnet	to	tourmaline
gd	gedrite	tr	tremolite
hb	hornblende	zo	zoisite
hc	hercynite	zr	zircon
ky	kyanite		

Mineralogy of Thin Sections

Unit 2

C75-241c pc hb qz bi op
 C75-251 pc ch qz bi ga op cc

Unit 3

C75-299 qz pc mu bi si ga ap to op st
 C75-333 qz pc bi ga op ap ch

Unit 4

CR-20 qz pc bi st ky si ga op zr cl
 C75-225a qz mi pc bi mu op
 C75-329a qz pc bi ga mu op

Unit 6

C75-59a do mu mi pc qz cc ch
 C75-72b do cc ph tr di
 C75-76 tr cc mu qz
 C75-176b di tr

Unit 8

C74-193* hb pc op ch
 C74-200 hb pc qz ga op
 C75-174* hb pc bi op sp ap cc
 C75-200* pc hb qz op bi ap
 C75-222* hb pc bi ch ap ep op
 C75-350* hb pc op ap cc bi ch

* Denotes specimen submitted for whole rock analysis.

C76-43* hb pc sp ch op

Unit 9

C75-30 pc hb cm qz op ap cc bi ch
 C75-33b pc fh qz ep sp ap op
 C75-33c pc fh bi qz sp op ap ep
 C75-35a* qz fh pc mi ch op ap cc
 C75-43 pc qz bi op ep ap
 C75-47a* pc bi qz mi fh ep sp ap op ch cc to
 C75-49 pc qz bi mu ch cc op ep ap sp
 C75-56b* qz pc bi mu op ap ep
 C75-80b* pc bi qz op ap ep sp
 C75-098* pc hb qz bi sc ep sp ch cc op ap
 C75-132b pc bi hb qz ch op cc ap ep
 C75-149a* pc hb bi qz cc sp ep ch ap sc op
 C75-151* pc hb bi ch cc op ep zo
 C75-205* pc mi qz bi mu ch sp op cc
 C75-290 pc qz bi op sp ap ep

Unit 10

C74-188* qz pc mi bi op ch mu
 C75-67* pc qz bi mu ch ap op cc ga
 C75-80a* pc qz bi mu op sp ap
 C75-80c* pc qz mi mu ch cc ep
 C75-100* pc qz mi mu ch op
 C75-264 pc qz mu bi cc op
 JM-3e* pc qz mi mu ch op

Unit 11

C75-28* pc qz bi fh op ga ap ch
 C75-61* qz bi op pc mu ap cc to
 C75-86a* mi qz bi hb pc ep sp op ap to
 C75-94a mi qz ac ep pc bi op sp ap
 C75-110 qz mi mu pc bi op ch
 C75-113 mi bi qz pc cc op ap

Unit 12

C75-291b cm gd op ch
 C75-291d ac cm ch gd ga op
 C75-297a ac oam ga hc cc op ch
 C75-297b ac cl hc ch oam
 C75-329b hb pc qz ap op cl
 C75-332 pc fh qz sp cc zo ep ap

Unit 16

C75-50a qz mu pc cl to op
 C75-51a qz mu pc ky bi ap op si to
 C75-51d mu to op bi
 C75-53a qz mu bi pc to op ch
 C75-55 qz mu bi st ga pc op to ky
 C75-62a qz mu st bi pc to ap op

C75-64b pc qz bi to op ap
 C75-68a qz mu bi ga cl ap op
 C75-69a qz mu ga st op
 C75-69b qz mu st ga bi op
 C75-70b qz mu ch st ga bi to op pc(altered)
 C75-73a qz bi mu st pc ch si op to ap
 C75-75b qz ky pc mu bi op cl to
 C75-78a qz bi mu si cl pc to op ap
 C75-85a qz mu st ga pc bi to sp op ch
 C75-88b qz mu bi st pc op to ap
 C75-93b qz mu bi cl si pc ky st ap to op
 C75-94d qz mu bi ky st pc op to ap
 C75-97b qz pc bi st si ch ga
 C75-97d pc qz bi mu st ga ap op
 C75-103a pc ky mu bi qz op ch to ap
 C75-103b mu qz bi pc cl op to ch
 C75-103d qz mu pc op ch bi to ap sp
 C75-106 qz mu ky st bi cl to op
 C75-107a qz pc mu bi st ga ky op
 C75-111a qz mu bi ky pc op cl to ap
 C75-111c qz mu ph ky cl pc to op
 C75-112 qz mu bi st ga pc to op ap
 C75-156a bi qz pc mu si cl ga op to ap
 C75-157a qz pc mu bi ga ch op
 C75-168a qz pc mu bi si to ap op
 C75-168b qz bi pc ga st ch ap op
 C75-180a qz pc bi ga st si to op ap
 C75-185a qz mu pc bi cl si ga ap op
 C75-197c qz bi pc si ga st mu op
 C75-209 qz mu bi si pc op cl ap
 C75-233 qz mu bi si cl ch ep op to ap
 C75-256a qz mu bi pc si cl to op ap
 C75-256b qz bi mu ga si cl op
 C75-258b qz mu bi si ga cl op
 C75-267 qz mu pc bi si cl to op
 C75-270a qz mu bi si cl st ga to op
 C75-278b qz bi pc ga si mu to ap op
 C75-287a qz bi pc ga si cl ch st ap op
 JM-16a qz mu bi ga op cz ap ch ep

Unit 19

C75-130 mi qz di ph cc tr sp ap
 C75-268 mi qz di ph tr pc cc tp op

Unit 20

C75-121b pc mi di ep cc sp qz ap op

Unit 22

C75-139 mi qz ac ep di cc ap sp
 C75-353 pc bi ac di cc sp

APPENDIX II. Construction of Cross Sections (Figure 9).

The cross sections illustrated in Figure 9 are extremely important in understanding the geometry of the Clare River Structure, and an explanation of their construction will be presented here. Three items of information were used in their construction: 1) the depth to basement, 2) the outcrop width of the units, and 3) the nature and orientation of minor structures observed at the surface.

The first step was to construct a vertical profile along each section line of the contact between the Addington Gneiss (basement) and the Clare River supracrustals. These profiles are based on a gravity study of the Structure by Fitzpatrick (1970). Along two section lines, one parallel to the axis of the Clare River Structure and one perpendicular to it, he calculated three depth to basement profiles, based on different density contrasts between basement and supracrustal rocks. Using Fitzpatrick's middle profiles, the writer extrapolated the depth to basement (Addington Gneiss) for each vertical section illustrated in Figure 9.

The next step was to plot the dips and outcrop widths of the various units on the sections. These were extrapolated below the surface in the light of major structures which were demonstrated by the plan view (Figures 3 and 8), and the minor folds and interference patterns present locally. The complicated patterns displayed in the cross sections are the result of space problems imposed by the profiles of the contact between the Addington Gneiss and the supracrustals.

Two important features emerge from the construction of these cross sections. The first is the nature of the contact between the Addington gneiss and the supracrustals. Because it is not repeated by the Fl Hungerford Antiform, it must have been a plane of dislocation during D1. In addition, supracrustal rocks can be seen to be sheared out along it.

Secondly, in order to explain the outcrop width of the Todd Lake and Shovel Lake Formations, a series of F2 folds must be present in the northeastern portion of the map area. Due to a lack of marker horizons and minor folds in this area, these folds were not outlined by surface mapping.

APPENDIX III. Mineral Textures Observed in Thin Sections of the Pelitic Schist (unit 16). \

In this Appendix, the textural features observed in each thin section are described and related to S₁, the principal foliation observed in all sections. The descriptions are tabulated in order of increasing metamorphic grade, which increases from northeast to southwest (Figure 17). All station locations are shown in Figure 28.

Abbreviations not listed in Appendix 1

foln	foliation
S _e	foliation outside a porphyroblast
S _i	foliation enclosed within a porphyroblast
xl	crystal
xls	crystals

SECTION	PRE-D1	SYN-D1	POST-D1
C75-111a	-pc xls contain discordant inclusion trails defined by fine op	-S1 foln defined by mu, bi, tabular qz -ky generally lies in S1	-ky contains straight inclusion trails of S1 -crossed bi overgrows S1 -some ky xls have been rigidly rotated -pc and ky contain straight inclusion trails of S1
C75-111c		-S1 foln defined by mu, bi, tabular qz	-crossed mu overgrows foln -mu kinked by F2 folds, ky rigidly rotated so that S_1 is not always continuous with S_e
C75-112a		-S1 foln defined by mu, bi, and tabular qz. Ga and st occasionally contain curved S_1 that is continuous with S1 -st lies parallel to S1	-crossed bi overgrows S1 fabric -S1 abuts against ga and st porphyroblasts
C75-106		-S1 foln defined by mu and tabular qz. Pc and ragged ky contain straight inclusions of S1, but S1 wraps around them. Ky and pc appear to be syn-D1	-crossed mu and bi are common -micas are bent and exhibit undulatory extinction
C75-107a		-S1 foln defined by individual xls of mu and bi	-subidioblastic ga, xenoblastic st overgrow straight inclusions of S1 -crossed mu and bi are common
C75-88b		-S1 defined by mu and bi	-st overgrows S1 and contains straight inclusion trails of S1 (in one xl, the trail is slightly curved) but S1 subsequently wraps the st -crossed mu and bi occur
C75-103a		-S1 foln defined by stacked layers of mu, and by larger individual flakes of mu and bi	-ky tends to be parallel to S1, and overgrows straight inclusions of S1. Pc overgrows straight inclusions of S1 -crossed mu and bi present, but they are overgrown by mu parallel to S1 -mu is kinked and broken by D2

SECTION	PRE-D1	SYN-D1	POST-D1
C75-103d		-S1 foln defined by tabular qz, op, and layers of mu	-pc porphyroblasts overgrow straight inclusions of S1 -S1 is folded into F2 chevron folds -pc folded by F2 -qz develops undulatory extinction during D2
C75-103b		-S1 foln defined by mu layers, bi, fibrous clay minerals, and aligned pc poikiloblasts	-pc overgrows straight inclusions of S1, but is aligned in S1 -fibrous clays are forming after pc -S1 is folded by F2. Qz shows deformation bands, mu and bi are strained in F2 hinge zones
C75-94d		-S1 foln defined by individual flakes of mu, bi, and aggregates of clay minerals after pc	-ky and st porphyroblasts contain straight inclusion trails of S1 -crossed mu and bi are common -si nucleates on fibrous clays
C75-51a		-randomly oriented ky overgrows fine grained inclusions (early S1) continuous with coarser external S1 foln -S1 defined by individual flakes of mu and fine tabular op	-crossed bi grows at right angles to S1 -one patch of si nucleates parallel to the crossed bi
C75-50a		-S1 defined by stacked layers of mu, tabular op and qz, fibrous clays -poikiloblastic pc contains S-shaped inclusion trails of S1	-crossed mu common -S1 is folded by F2 folds. Mu shows strain extinction in the hinge zones
C75-53a		-S1 foln defined by mu, tabular qz and op	-S1 foln bent by F2 folds. Mu shows strain extinction in the hinge zones
C75-51d		-S1 foln defined by solid layers of mu flakes enclosing tabular op	-S1 folded by tight chevron-style F2 folds. Mu flakes are strained in the hinge zones
C75-55		-S1 foln is defined by solid matted mu and subordinate op in one portion of section, and by tabular qz, mu, and bi in the other portion	-ga and idioblastic st porphyroblasts contain straight inclusions of S1 (in one ga, trail is slightly curved) -S1 is bent by F2 folds and st and ga have been rigidly rotated during D2
C75-64b		S1 foln defined by aligned bi flakes	

SECTION	PRE-D1	SYN-D1	POST-D1
C75-62a		-S1 defined by mu, bi, and tabular op	-subidioblastic st contains a few tabular op that are continuous with S1. S1 mu and bi abut against st -crossed mu and bi are common
C75-68a		-S1 is defined by mu, bi, tabular qz and op	-ga overgrows straight inclusion trails of S1. External S1 is continuous with S1 in some ga, discordant in others -crossed mu common -S1 foln is bent by F2 folds -rotation of some ga during D2 to produce discordant trails
C75-69a		-S1 defined by layers of mu enclosing tabular op and qz in one portion of slide and by larger individual mu flakes, tabular qz, and op in the other portion of the slide	-st, ga porphyroblasts are randomly oriented and contain inclusions of S1 continuous with external S1 -crossed mu and bi are common -S1 is kinked by D2 and qz exhibits deformation bands
C75-75b		-ky and pc enclose an early stage of S1 defined by trails of op and subordinate qz -external S1 defined by coarser long slender mu flakes	-crossed mu, bi, and op are aligned
C75-85a		-S1 defined by layers of mu, individual mu flakes, tabular qz and op	-st, ga, pc overgrow S1, and contain straight concordant inclusions of it -poikiloblasts of ga, pc, are breaking up into pieces parallel to S1 -crossed bi has cleavage at a high angle to S1 but morphologically lie in S1 -S1 and ga are deformed by F2 folds
C75-73a1		-S1 foln defined by bi, tabular op. Some si parallels S1	-st overgrows straight inclusion trails of S1 and qz contains tabular op parallel to S1 -crossed bi is common and some si nucleates parallel to the bi -subidioblastic st contains inclusion trails parallel to the crossed bi

SECTION	PRE-D1	SYN-D1	POST-D1
C75-73a2		<ul style="list-style-type: none"> -st contains straight inclusions of S1 -S1 is defined by si that nucleates on the faces of st -st xls impinge on one another due to flattening related to D1 	<ul style="list-style-type: none"> -si fibres often oriented at an angle to S1 -occasional crossed mu xls present
C75-97b		<ul style="list-style-type: none"> -ga and st contain straight inclusion trails of S1 which are much finer than external S1 -external S1 is defined by aligned bi flakes 	<ul style="list-style-type: none"> -rims of ga are clear of inclusions, per a second growth -crossed bi common -ga xls rigidly rotated, resulting in discordant inclusion trails
C75-97d		<ul style="list-style-type: none"> -S1 foln defined by flakes of mu and tabular op 	<ul style="list-style-type: none"> -poikiloblastic st and ga contain straight inclusion trails of S1 that are continuous with external fabric -crossed mu and bi grow at forty-five degrees to S1
C75-78a		<ul style="list-style-type: none"> -S1 foln defined by parallel mu flakes and by si flakes and fibrolites patches 	<ul style="list-style-type: none"> -pc overgrows qz that is aligned and continuous with S1 -pc is highly altered
C75-278b		<ul style="list-style-type: none"> -idioblastic porphyroblasts of ga overgrow very fine inclusions representing an early stage of S1 -S1 foln defined by bi flakes which wrap ga 	<ul style="list-style-type: none"> -mu, bi grow across foln -abundant crossed bi overgrows fabric
C75-70b		<ul style="list-style-type: none"> -recrystallized qz preserves within it fine tabular opaques which represent an early stage of S1 -S1 foln defined by layers of mu flakes, tabular op and qz 	<ul style="list-style-type: none"> -mu bent by F2 folds, and shows undulatory extinction in their hinge zones -qz displays undulatory extinction -some mu rotated into the axial planes of the F2 folds
C75-258b		<ul style="list-style-type: none"> -ga overgrows early S1 foln defined by tabular qz and op. Some ga shows inclusion free rims -S1 foln defined by mu, bi, tabular qz, si. Locally, S1 is continuous with inclusion trails in ga -external S1 wraps ga locally 	<ul style="list-style-type: none"> -ga porphyroblasts have been rotated, resulting in discordant inclusion trails -some si fibrolite patches appear to have rotated -crossed mu and bi common -mu and bi is bent around the hinges of D2 folds

SECTION	PRE-D1	SYN-D1	POST-D1
C75-256a		-S1 defined by layers of flakes of mu and bi, and si as needles and patches of fibrolite	-crossed mu and bi occur locally -S1 openly folded by F2 folds -qz shows undulatory extinction
C75-256b		-cores of ga xls display fine straight inclusion trails of qz and op believed to be an early stage of S1 -S1 foln defined by mu and bi flakes, si needles and patches, and tabular qz	-early ga overgrown by inclusion free rims and one exceptional rim overgrows external S1 foln -crossed mu and bi are common -S1 foln is openly folded by F2 folds. Mu and bi is bent around their hinges, and crossed mu and bi lie roughly in the axial planes of these folds
JM-16a		-ga contains straight aligned inclusions of tabular qz and op that are finer grained than but continuous with the external S1 -S1 defined by stacked flakes of mu, bi, tabular op and qz	-ga has been rigidly rotated, in some cases producing discordant inclusion trails -S1 is bent by F2 folds. Undulatory extinction of micas occurs in the noses of these folds
C75-270a		-st which is breaking down, contains inclusions that are usually discordant with the external foln, but locally, the external foln is continuous with the inclusion trails, suggesting that it is an early stage of S1 -S1 foln defined by mu, bi, si, op, and pieces of what was once a large poikiloblastic st xl -continued generation of S1 breaks apart the large relict xls of st, resulting in remnants parallel to S1, and also in rotation of st with straight inclusions -bi appears to pseudomorph st -S1 foln wraps around st in most cases	

SECTION	PRE-DI	SYN-DI	POST-DI
C75-267		-S1 defined by stacked layers of mu, plates of bi. S1 nucleates parallel to S1 on clays that are alteration products of pc	-pc poikiloblastically encloses S1, but its alteration products give rise to si that is parallel to S1. D1 must continue after growth of pc -crossed mu and bi overgrow fabric -S1 is folded by F2 folds, and mu is strained in their hinges. Some rotation of mu and bi into the axial planes of these folds occurred -mu, bi, minor si grow across the foln -small idioblastic ga xls are clear of inclusions and are probably later than those containing inclusions
C75-197c		-ga contains straight to slightly curved inclusions of S1 that are continuous with but finer than the external S1 -S1 foln defined by bi flakes, si needles	
C75-180a	-ga contains a dusting of random inclusions	-pc contains straight inclusions of S1 that are continuous with but slightly finer than external S1 -S1 foln defined by aligned pc poikiloblasts, bi, op and tabular qz -S1 wraps early ga -ga overgrows an early phase of S1 -S1 foln defined by stacked layers of mu, si, bi, clay mineral patches, and qz	-st poikiloblastically contains straight inclusions of S1. In one case, st overgrows S1 that has wrapped around an early ga -crossed mu and bi overgrow S1
C75-185a			
C75-168b	-xenoblastic st, idioblastic ga contain fine random inclusions	-S1 foln defined by bi, ch, op, qz -S1 wraps around st and ga	-qz displays pronounced undulatory extinction
C75-168a		-S1 foln defined by mu, bi, si needles	-some pods of fibrous si grow across foln -crossed mu and bi are common -qz displays undulatory extinction
C75-156a		-no planar fabric observed	-si, mu, bi, qz, pc, clay minerals randomly oriented in slide

SECTION	PRE-D1	SYN-D1	POST-D1
C75-157a	-some ga enclose randomly oriented inclusions	-some ga contain five weakly aligned inclusions that are mainly discordant but occasionally continuous with external S1 -S1 foln defined by mu, bi, and lenticular qz -S1 poorly defined by aligned bi, some mu	-S2 defined by spaced shear planes that are filled with altered material. Mu and bi are rotated towards S2. -qz is highly strained and lenticular as a result of D2
C75-209			-coarse mu, si, equant op, qz have recrystallized with random orientations -qz is highly strained and displays serrated grain boundaries
C75-233		-S1 poorly defined by aligned mu, bi, si	-D2 results in folds of S1 fabric, strain extinction of micas in hinge zones, strain extinction in qz and pc, and strain induced polygonization of qz
C75-287a	-ga cores contain fine disoriented inclusions -garms are inclusion free	-S1 defined by stacked layers of bi, si, clay mineral patches -S1 foln wraps around subidioblastic ga.	-crossed bi are common -qz and pc are highly strained, locally qz displays a mortar texture

APPENDIX IV. Chemical Tests to Distinguish between Ortho- and Paragneisses.

A variety of chemical tests have been proposed to distinguish between volcanically derived orthogneisses and paragneisses whose compositions fall into the intermediate to felsic range (e.g. Blackburn, 1974; Shaw, 1971; Davoine, 1969). The five tests quoted below are extracted from Blackburn (1974).

1. If the MgO content is greater than CaO while K₂O is greater than Na₂O, the rock is deemed sedimentary.
2. If normative corundum is greater than five percent, the rock is sedimentary.
3. Davoine (1969) used analyses of 33 rhyolites, 33 dacites, and 32 arkoses to establish the criteria below:
 - a) paragneisses average less than 2.5 percent CaO and less than 7 percent (Na₂O plus K₂O).
 - b) orthogneisses average greater than 2.5 percent CaO.
4. Shaw (1972) outlined the fields of 278 igneous rocks and 506 sedimentary rocks on an albite-orthoclase-quartz diagram. The gneisses of the Shovel Lake Formation are plotted on this diagram in Figure 27. Normative mineralogy is shown in Table 9.
5. Shaw (1972) proposed the following discriminant function to distinguish between ortho- and paragneisses:

$$DF3 = 10.44 - 0.21 SiO_2 - 0.32 Fe_2O_3 \text{ (total Fe)} - 0.98 MgO \\ + 0.55 CaO + 1.46 Na_2O + 0.54 K_2O$$
 Positive values indicate probable igneous parentage; negative values are sedimentary.

Dr. J.H. Bourne of the Geological Survey of Canada kindly supplemented the writer's chemical data by providing 10 chemical analyses of gneisses of the Shovel Lake Formation in order to test the discriminatory methods cited above. The results are presented in Table 9.

These methods do not satisfactorily discriminate between gneisses of igneous and sedimentary parentage in the Shovel Lake Formation, as discussed in Chapter 7.

Figure 27. Normative quartz-albite-orthoclase plot for the intermediate and felsic gneisses of the Shovel Lake Formation. Igneous field (solid line), sedimentary field (dashed line) and vacant fields (stippled) taken from Shaw (1972).

Table 9. Normative mineralogy of rocks of the Shovel Lake Formation.

Sample	Q	C	Or	Ab	An	Lc	Ne	Di	He	En	Fs	Fo	Fa	Mt	Il	Cr	Ap	Po
C74-188	28.20	1.84	15.91	44.68	0.75	0.00	0.00	0.00	0.00	2.32	3.96	0.00	0.00	1.36	0.82	0.00	0.14	0.03
C74-193	0.00	0.00	3.02	10.71	34.93	0.00	3.32	11.76	5.37	0.00	0.00	16.19	9.35	3.44	1.65	0.00	0.21	0.06
C75-067	21.03	1.24	4.79	58.87	7.51	0.00	0.00	0.00	0.00	1.94	2.88	0.00	0.00	0.97	0.68	0.00	0.39	0.03
C75-080c	28.13	0.00	16.15	42.90	5.71	0.00	0.00	0.40	0.88	0.66	1.68	0.00	0.00	0.81	0.70	0.00	0.09	0.08
C75-100	26.90	0.00	11.65	50.18	5.71	0.00	0.00	0.10	0.47	0.48	2.70	0.00	0.00	1.00	0.57	0.00	0.21	0.03
C75-149	3.29	0.00	9.88	28.09	18.69	0.00	0.00	11.50	7.34	6.69	4.90	0.00	0.00	3.57	3.91	0.00	2.02	0.14
C75-151	4.83	0.00	7.57	32.99	22.83	0.00	0.00	6.21	3.66	8.55	5.79	0.00	0.00	3.00	2.96	0.00	1.55	0.06
C75-174a	0.00	0.00	0.89	19.89	28.69	0.00	0.00	12.33	4.19	11.05	4.31	9.23	3.97	3.44	1.65	0.00	0.28	0.08
C75-200	3.58	0.00	2.72	20.73	26.20	0.00	0.00	9.59	5.21	14.65	9.14	0.00	0.00	4.23	3.04	0.00	0.86	0.06
C75-222	0.00	0.00	0.71	23.44	25.31	0.00	0.00	13.01	4.76	14.17	5.95	4.26	1.97	3.71	2.05	0.00	0.63	0.06
C75-350	0.00	0.00	1.95	15.92	28.56	0.00	1.15	11.06	7.16	0.00	0.00	14.15	11.57	4.48	3.02	0.00	0.86	0.11
C76-043	0.00	0.00	5.50	22.25	20.67	0.00	0.00	17.76	8.55	5.96	3.29	5.53	3.37	4.15	2.58	0.00	0.30	0.08
JM76-3e	29.40	1.52	10.06	54.83	0.70	0.00	0.00	0.00	0.00	0.72	1.16	0.00	0.00	0.60	0.63	0.00	0.07	0.33
C75-028	16.45	1.58	9.87	34.88	13.40	0.00	0.00	0.00	0.00	4.76	11.14	0.00	0.00	4.10	2.50	0.04	1.28	0.00
C75-035a	36.80	0.00	1.60	33.84	12.90	0.00	0.00	1.64	2.48	2.75	4.76	0.00	0.00	1.74	1.01	0.04	0.26	0.19
C75-047a	17.76	0.62	10.55	32.38	20.10	0.00	0.00	0.00	0.00	6.45	5.39	0.00	0.00	3.71	1.99	0.07	0.93	0.06
C75-056b	41.31	2.08	8.04	31.51	7.65	0.00	0.00	0.00	0.00	3.76	1.88	0.00	0.00	2.63	0.82	0.06	0.21	0.06
C75-061	15.10	1.78	13.48	30.63	11.59	0.00	0.00	0.00	0.00	4.18	14.99	0.00	0.00	4.84	3.48	0.04	1.75	0.14
C75-080a	21.36	0.00	7.13	50.56	9.97	0.00	0.00	0.23	0.28	2.24	3.11	0.00	0.00	3.20	1.31	0.05	0.49	0.08
C75-080b	6.74	2.81	16.36	41.54	12.97	0.00	0.00	0.00	0.00	8.21	4.85	0.00	0.00	3.49	2.15	0.04	0.81	0.03
C75-086a	10.67	0.00	21.13	30.14	14.00	0.00	0.00	2.05	0.85	10.43	4.96	0.00	0.00	3.50	1.68	0.05	0.50	0.06
C75-098	0.65	0.00	10.02	34.73	24.50	0.00	0.00	4.53	3.36	8.65	7.35	0.00	0.00	3.92	2.26	0.12	0.42	0.08
C75-205	21.87	0.23	10.75	39.75	14.44	0.00	0.00	0.00	0.00	3.79	2.43	0.00	0.00	3.97	2.06	0.05	0.83	0.03
Ch-1	0.00	0.00	1.82	19.91	29.89	0.00	0.00	10.22	5.47	7.35	4.51	8.47	5.73	3.50	2.23	0.00	0.91	0.00
Ch-2	0.00	0.00	9.89	33.94	25.88	0.00	0.00	5.71	4.12	1.26	1.04	5.43	4.95	4.28	2.69	0.00	0.81	0.00
P-27	0.36	0.00	2.00	20.31	32.37	0.00	0.00	12.11	8.09	9.71	7.44	0.00	0.00	4.57	3.06	0.00	0.00	0.00
P-67	0.00	0.00	7.02	36.03	25.38	0.00	1.05	5.80	4.20	0.00	0.00	7.47	6.83	3.99	2.24	0.00	0.00	0.00
P-78	7.15	0.00	1.84	20.50	28.30	0.00	0.00	16.10	10.54	4.90	3.71	0.00	0.00	3.77	2.46	0.00	0.00	0.00
P-351	0.00	0.00	4.54	26.85	24.67	0.00	0.00	9.94	7.39	7.52	6.42	3.33	3.13	3.52	2.70	0.00	0.00	0.00

Table 10. Results of tests to distinguish ortho- and paragneisses.

SAMPLE	TEST 1	TEST 2	TEST 3	TEST 4	TEST 5
C74-188	I or S	I	S	I or S	I
C75-067	I	I	S	n.d.	I
C75-080c	I	I	S	I or S	I
C75-100	I	I	S	I or S	I
C75-151	I	I	I	n.d.	I
JM76-3e	I or S	I	S	I or S	I
C75-028	I	I	I	I or S	I
C75-035a	I	I	I	I or S	I
C75-047a	I	I	I	I or S	I
C75-056b	I	I	S	S	S
C75-061	I	I	I	I or S	I
C75-080a	I	I	S	n.d.	I
C75-080b	I or S	I	I	I or S	I
C75-086a	S	I	I	I or S	I
C75-098	I	I	I	n.d.	I
C75-205	I	I	I	I or S	I

I: igneous S: sedimentary n.d.: not discriminated

Appendix V. List of 1974 Stations and Corresponding Samples.

<u>STATION</u>	<u>SAMPLE</u>
C74-216	C74-180
C74-217	C74-181
C74-218	C74-182
C74-223	C74-183
C74-227	C74-184
C74-229	C74-185
C74-230	C74-186, C74-187
C74-231	C74-188, C74-189
C74-232	C74-190, C74-191
C74-233	C74-192, C74-193
C74-234	C74-194, C74-195
C74-235	C74-196
C74-236	C74-197, C74-198
C74-240	C74-199, C74-200, C74-201
C74-241	C74-202
C74-242	C74-203
C74-243	C74-204, C74-205, C74-206
C74-244	C74-207
C74-245	C74-208, C74-209
C74-247	C74-210

APPENDIX VI. GEOLOGY AND GEOCHEMISTRY OF THE KALADAR COMPLEX

Introduction

A number of gabbro and diorite plutons are exposed in the Ontario portion of the Central Metasedimentary Belt of the Grenville Province. In the Hastings Basin these intrusions are generally oval in shape, cover an area in the order of 20 to 30 square kilometres and are emplaced into meta-volcanics and metasediments of the Grenville Supergroup (Figure 29).

They have been studied by Lumbers (1967a) who mapped four of them (the Tudor Gabbro, the Lingham Lake, Thanet and Jocko Lake Complexes) in detail during the course of his regional compilation of the geology of a large portion of the Ontario Section of the Central Metasedimentary Belt. He divided the mafic intrusions into two groups, the late gabbro and diorite group associated with the nepheline syenite suite of intrusions, and an older diorite group, included in the biotite diorite series (Figure 30) dated at 1250[±]25 m.y. (Silver and Lumbers, 1966). The Kaladar Complex belongs to the earlier diorite group.

The intrusions of this group share several common features including trondhjemitic differentiation trends, average modal compositions ranging from diorite to granodiorite and a close spatial association both with the other groups in the biotite diorite series and the Hermon Group metavolcanics (Lumbers, 1967a). Gabbro and diorite with a hypidiomorphic-granular texture make up the bulk of these intrusions, minor pyroxenite is ubiquitous and peridotite occurs in some of them. Generally complete uranization of the pyroxenes has occurred in the gabbros and diorites. The tonalite portions of these intrusions contain greater than 10 percent quartz with biotite and primary hornblende.

The timing of these intrusions with respect to major diastrophic

Figure 29. Location and distribution of mafic intrusions of the early diorite group in a portion of the Grenville Province of southeastern Ontario. Geology modified after MacDonald, 1970. 1) Kaladar Complex, 2) Elzevir Batholith, 3) Skootomatta Diorite, 4) Lingham Lake Complex, 5) Tudor Gabbro, 6) Mount Moriah Syenite, 7) Thane Complex. Study area enclosed by dashed lines.

Figure 30. Plutonic events in a part of the Ontario segment of the Central Metasedimentary Belt. After Lumbers, 1967b.

and metamorphic events is still unclear. Lumbers (1967a) and Smith (1958) both argued on structural grounds that the emplacement of the plutons of the diorite group must postdate regional deformation; Lumbers points out that contact aureoles occur around all the diorite group plutons that he has mapped. Yet well established age relationships east of the Elzevir Batholith (a member of the biotite diorite series) show that polyphase deformation and regional metamorphism spanning the amphibolite facies occurred after the emplacement of the biotite diorite series (Brown et al., 1975; Thompson, 1972). This problem will be taken up later in this discussion.

The Kaladar Complex

Miller and Knight (1914) first described the Kaladar Complex as a body of "gneissic syenite" occupying several square miles south and southwest of Flinton and recorded that it was intruded by dykes of the "fundamental granite gneiss" (Elzevir granodiorite). In later mapping (Harding, 1944; Lumbers, 1967a) it was not distinguished from the Elzevir granodiorite until Thompson (1972) mapped a portion of it as hornblende diorite in the course of his mapping of the Flinton Group. The writer's interest was stimulated by the discovery of fragments of the Kaladar Complex in fragmental units of the Tudor metavolcanics, an observation which suggested that the Kaladar Complex contains the oldest rocks exposed in the Ontario section of the Central Metasedimentary Belt. This discovery prompted detailed mapping of the complex, the results of which form the basis of this study.

Figure 31 illustrates the details of the geology of the Kaladar Complex and its surroundings. It is a crudely oval-shaped body trending northeast and occupying an area of about 21 square kilometres; it is

bounded to the north and south by metabasalts of the Tudor formation (unit 1), to the west by the Elzevir Batholith (unit 8) and to the east by the metaconglomerates of the Flinton Group (unit 9).

The dominant rock types of the Kaladar Complex are metadiorite (unit 4) and metagabbro (unit 3), their gradational boundaries were arbitrarily established in the field at a colour index of 40. Within these units a complete range of compositions exists, from rocks transitional between diorite and quartz diorite right through to hornblendite. Anorthositic gabbros which now consist of plagioclase and actinolite are distinguished as unit 4a. Rocks of units 3 and 4 tend to be equigranular and coarse-grained with a hypidiomorphic-granular texture (Plate 7a) and locally they exhibit a primary alignment of plagioclase laths. The blocky outlines of the fibrous aggregates of amphibole suggest that they are uralitized pyroxenes. Highly deformed equivalents of this unit occur as sill-like bodies in the Tudor basalts of the south (Figure 31).

Two distinct rock types of ultramafic composition (unit 2) occur as isolated lenses within the body of the complex. Closely associated talc schists and serpentinites, the origin of which is unclear, make up the first type (2a). They may represent altered peridotites of the complex; alternatively, they may be related to a similar ultramafic body that was tectonically emplaced into the metasediments of the Flinton Group (Thompson, 1972). The second type of ultramafic bodies exposed in the area are composed of chlorite-actinolite schist (unit 2b) which appear to be altered equivalents of hornblendite.

The second major unit (unit 6) of the complex is a northeast-trending lenticular body that includes quartz diorite, granodiorite, and quartz monzonite. It occurs in the east central portion of the intrusion (Figure 31). These rocks exhibit a duality of texture in which the matrix

is a granoblastic aggregate of sodic plagioclase, quartz, microcline, and biotite which enclose relict phenocrysts of plagioclase and hornblende. The latter appear to have been rotated during deformation. The hornblende exhibits varying degrees of replacement by biotite in the form of fine aggregates of flakes that retain the orientation of the hornblende. The leucocratic body is zoned, showing an increase in quartz, a decrease in mafics, and an increase in biotite relative to hornblende towards its centre.

A small body of andesite porphyry (unit 5) occurs in the south-central portion of the complex and numerous dykes of the unit occur proximally to it. These dykes cut the hornblende diorite (unit 4), but are cut in turn by dykes related to the leucocratic body described above. Unit 5 is very distinctive, containing phenocrysts of plagioclase up to 2 cm across in a fine matrix of hornblende and plagioclase. The proportion of phenocrysts varies from sparsely scattered to up to 50 percent of the rock. Some of the amphibolite dykes cutting the complex are probably phenocryst poor varieties of this unit, others may be related to the Tudor metabasalts.

Leucocratic dykes are found throughout the complex. Although it is often impossible to distinguish individual dykes in the field, it is clear that some are related to the leucocratic body of the complex (unit 6) while others are related to the Elzevir Batholith (unit 8). The large dyke (unit 7) in the south-central portion of the complex differs from the leucocratic body of unit 6 only in having sharp contacts with the host diorites: it generally shares the plagioclase phenocrysts and the texture where biotite occurs as clots of fine flakes, although locally neither of these features are evident. Granodiorite dykes that are similar in appearance to the Elzevir Batholith cut the Tudor volcanics on the northwest.

contact and they also appear to cut the Kaladar Complex along its western contact with the Elzevir Batholith. These dykes in which biotite is evenly distributed as fine flakes exhibit a granoblastic texture of quartz and feldspar. With a few exceptions, they are concordant with their host rocks.

The Tudor metabasalts (unit 1), bounding the complex to the north and south, have been metamorphosed to fine-grained amphibolites. Their massive, uniform and fine grained nature suggests that they were originally flows, and pillow structures, relict pyroxene phenocrysts, and flow breccias locally are preserved. Thinly laminated biotite rich layers are a minor constituent in the volcanics and probably represent interflow sediments. These basalts are equivalent to the lower portion of the pile described by Sethuraman and Moore (1973) 20 km to the north.

Only a small portion of the Elzevir Batholith was examined during the course of this study, and in the immediate vicinity of the Kaladar Complex, it is a pale pink-weathering medium-grained granodiorite with a granoblastic texture and a foliation defined by the alignment of biotite flakes. Away from the complex, the granodiorite changes to a grey-weathering quartz diorite consisting of quartz, plagioclase and biotite.

Field Relations

That the Kaladar Complex belongs to the early diorite group (Lumbers, 1967a) is indicated both by its close spatial relationship with the Tudor volcanics and with the Elzevir batholith, a member of the trondhjemite and soda granite group (Figure 30). There are conflicting age relationships between units of the Kaladar Complex and the Tudor meta-volcanics, suggesting that there exists a close temporal relationship between the two. Fragments of coarse grained gabbro are found in volcanic units on both the northern and southern contacts between the Kaladar

Complex and the Tudor metabasalts (Plate 7b). In addition, hornblendite fragments occur in basalt flows near the contacts with the complex. These fragments are mineralogically and texturally similar to rocks of unit 3 of the complex, indicating that it is at least in part older than the Tudor metavolcanics. This relationship is further evidenced by the occurrence of amphibolite dykes, some of which are related to the Tudor volcanics, that are observed to cut the diorite.

That the Kaladar Complex in part postdates the extrusion of the Tudor metabasalts is suggested by the presence of sill-like bodies of deformed metagabbro which intrude them (Figure 31) and also by the presence of numerous amphibolite xenoliths which occur in all phases of the complex (Plate 7c). The most probable explanation of these conflicting age relationships is that the Kaladar Complex and the Tudor basalts are contemporaneous and different manifestations of the same intrusive cycle.

Within the Kaladar Complex itself, the leucocratic phase (unit 6) postdates the diorite (unit 4), as indicated by the fact that the former intrudes the latter. It would also appear that the more mafic phases of the complex are the oldest. This is illustrated by the fact that angular gabbroic fragments similar to unit 5 are commonly enclosed within the diorite of unit 4 (Plate 7d). Another interesting feature seen within the complex is a breccia in which highly chloritized diorite fragments are enclosed by fresh diorite of very similar composition (Plate 7e) pointing to the episodic nature of the igneous activity.

A period of deformation separates the intrusion of the Elzevir Batholith and the Kaladar Complex, illustrated by the fact that dykes of Elzevir granodiorite cut across a foliation in the Tudor metavolcanics (Plate 7f). Also pointing to this fact is the occurrence of an intrusive breccia (Plate 7g) in which marginal aplitic phases of the Elzevir Batholith

3 3

OF/DE

enclose angular fragments of foliated Tudor metabasalt.

One puzzling feature in the light of the above data is the presence of a dyke of diorite (Thompson, 1972) cutting a foliation in the Tudor metovolcanics, illustrated in Plate 7h. It is tentatively suggested that this dyke is related to the Elzevir and has been contaminated by its host rocks resulting in its dioritic composition; and is not related to the emplacement of the Kaladar Complex.

The unconformable nature of the Flinton Group with respect to the basement complex has been well documented (Thompson, 1972; Moore and Thompson, 1972). Effects of the Flinton deformation, although not so pronounced as in the Flinton Group are present within the Kaladar Complex, where an early east-west trend can be seen to be rotated into and locally crenulated by the Flinton trends (Figure 31). At least two phases of deformation, separated by a major unconformity have affected the Kaladar Complex.

Geochemistry of the Kaladar Complex

Major and trace element analyses for rocks of the Kaladar Complex and surrounding units are presented in Tables 11 and 12. The rock units which correspond to each analyzed sample are listed in Table 13 and all samples are located in Figure 31.

In order to characterize the suite, AFM and alkali-silica diagrams are presented in Figures 32 and 33 respectively. In the alkali-silica diagram, all samples can be seen to fall into the subalkaline field. The Al₂O₃ plot shows a well defined calc-alkaline differentiation trend, with the exception of only four points. Samples 111 and 226 are metamorphosed basalts of the Tudor formation and fall within the tholeiitic trend defined by Sethuraman and Moore (1973) for equivalent metabasalts to the north.

Table 11. Major element analyses of rocks of the Kaladar Complex

SAMPLE	SiO ₂	Al ₂ O ₃	Fe ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	H ₂ O	CO ₂	TiO ₂	P ₂ O ₅	S	MnO	O excess
C74-111	46.07	11.09	3.58	11.51	9.10	9.47	2.10	0.33	2.44	0.55	2.17	0.22	0.49	0.23	0.25
C74-137	51.71	2.97	1.66	5.41	20.97	12.38	0.35	0.02	2.64	1.20	0.48	0.04	0.02	0.16	0.01
C74-171	46.36	14.16	2.17	7.08	12.89	9.90	1.33	0.90	2.83	0.76	1.39	0.04	0.05	0.16	0.02
C74-172	43.48	7.84	2.60	8.49	18.41	11.74	0.77	0.24	2.88	1.49	1.76	0.05	0.17	0.17	0.09
C74-178	55.36	14.22	1.84	5.90	7.59	6.35	3.00	1.74	1.72	0.80	1.03	0.34	0.02	0.11	0.01
C74-226	46.99	11.88	3.62	11.67	6.77	11.59	2.28	0.24	1.88	1.22	1.38	0.22	0.04	0.24	0.02
C74-236	47.94	0.79	1.46	5.19	31.93	1.25	0.02	0.01	8.64	2.52	0.05	0.02	0.21	0.08	0.12
C74-238	51.39	20.15	0.66	2.14	6.98	11.59	3.12	0.78	2.04	0.94	0.13	0.04	0.02	0.04	0.01
GR-136	67.58	15.37	0.88	2.77	2.18	3.84	4.84	1.67	-	-	0.57	0.23	0.02	0.06	0.01
C74-145	70.68	15.81	0.41	1.28	1.36	2.39	5.85	1.74	-	-	0.30	0.12	0.04	0.03	0.02
C74-217	66.67	15.56	0.90	2.84	2.12	4.00	4.17	2.89	-	-	0.50	0.18	0.00	0.07	0.00
C74-231	52.55	13.49	1.98	6.18	12.00	9.21	2.31	1.16	-	-	0.78	0.21	0.01	0.14	0.00
C74-235	71.74	14.96	0.52	1.67	1.63	2.20	5.34	1.42	-	-	0.35	0.13	0.00	0.03	0.00
C74-237	44.55	2.38	2.28	7.13	43.21	0.04	0.08	0.01	-	-	0.05	0.00	0.21	0.17	0.10
C74-242	72.87	14.54	0.55	1.73	0.67	1.83	4.85	2.53	-	-	0.28	0.11	0.00	0.04	0.00
C74-243	51.47	21.27	0.85	2.69	8.46	11.22	2.55	1.13	-	-	0.13	0.14	0.05	0.06	0.02
C75-008	48.95	7.08	2.16	6.74	25.01	9.44	0.26	0.00	-	-	0.10	0.14	0.01	0.12	0.00
C75-012	56.11	15.66	1.80	5.63	6.50	6.61	3.48	2.53	-	-	1.19	0.38	0.01	0.01	0.00
C75-017	63.46	16.26	1.10	3.45	4.21	2.96	4.37	3.11	-	-	0.74	0.25	0.00	0.00	0.00
C75-022	66.58	15.65	0.81	2.53	2.52	3.61	4.28	3.27	-	-	0.49	0.20	0.00	0.06	0.00
C75-028	51.91	15.72	1.89	5.90	9.39	9.43	2.94	1.29	-	-	1.04	0.36	0.01	0.13	0.00
C75-029	50.55	8.60	2.17	6.76	18.00	12.56	0.75	0.14	-	-	0.17	0.11	0.03	0.18	0.01

Source: X-ray fluorescence analyses, Dr. G. R. Holland, Science Laboratories, Durham, England.

Total iron recalculated to give an FeO/Fe₂O₃ ratio of about three to one. First group recalculated to 100%; second group recalculated to 100% on a volatile free basis.

Table 12. Trace element analyses of rocks of the Kaladar Complex

SAMPLE	Ba	Nb	Zr	Y	Sr	Rb	Zn	Cu	Co	Ni	Cr
C74-111	85	4	144	54	125	6	130	91	72	61	149
C74-137	43	4	16	12	33	1	89	9	66	473	1543
C74-171	341	8	27	16	451	20	84	10	65	228	546
C74-172	99	10	56	29	86	4	107	44	66	331	942
C74-178	364	21	57	29	356	64	102	5	46	104	313
C74-226	72	4	39	29	193	5	116	21	64	27	37
C74-236	21	3	2	1	43	1	70	410	102	2185	3127
C74-238	110	3	8	2	965	22	26	4	48	261	296
Gr-136	269	11	150	13	505	40	61	4		31	34
C74-145	212	7	142	5	312	39	30	17		13	15
C74-217	342	8	169	14	394	86	54	3		27	35
C74-231	276	6	83	13	385	33	57	47		135	431
C74-235	261	8	145	6	525	34	23	27		26	23
C74-237	37	4	1	0	7	3	83	4		2339	3248
C74-242	296	15	214	5	386	59	58	4		6	16
C74-243	169	2	12	0	945	32	24	8		207	460
C75-008	45	3	14	11	25	0	86	4		912	1676
C75-012	407	21	233	24	395	74	73	27		101	193
C75-017	215	15	191	16	322	94	51	17		47	78
C75-022	297	10	169	15	356	97	42	57		43	61
C75-028	277	12	54	29	495	33	87	58		148	301
C75-029	93	0	13	9	137	5	100	0		474	644

Source: as in Table 11.

Table 13. Rock units corresponding to the chemical analyses listed in Tables 11 and 12.

C74-111:	fine grained amphibolite of the Tudor formation (unit 1).
C74-137:	actinoliteite inclusion in metabasalt of the Tudor formation.
C74-171:	Metagabbro (unit 3) of the Kaladar Complex.
C74-172:	Hornblendite (unit 3) of the Kaladar Complex.
C74-178:	Hornblende diorite (unit 4) of the Kaladar Complex.
C74-226:	Fine grained amphibolite of the Tudor formation (unit 1).
C74-236:	Metamorphosed ultramafic (unit 2a, talc schist) of the Kaladar Complex.
C74-238:	Anorthositic gabbro (unit 4a) of the Kaladar Complex.
GR-136:	Elzevir granodiorite, 8 km south of the Kaladar Complex.
C74-145:	Elzevir granodiorite (unit 8).
C74-217:	Biotite diorite (unit 6) of the Kaladar Complex.
C74-231:	Hornblende diorite (unit 4) of the Kaladar Complex.
C74-235:	Biotite diorite (unit 6) of the Kaladar Complex.
C74-237:	Metamorphosed ultramafic (unit 2a, serpentinite) of the Kaladar Complex.
C74-242:	Elzevir granodiorite (unit 8).
C74-243:	Anorthositic gabbro (unit 4a) of the Kaladar Complex.
C75-008:	Chlorite-actinolite schist (unit 2b) of the Kaladar Complex.
C75-012:	Biotite-hornblende diorite (unit 4) of the Kaladar Complex.
C75-017:	Biotite-quartz diorite (unit 6) of the Kaladar Complex.
C75-022:	Biotite-quartz diorite (unit 6) of the Kaladar Complex.
C75-028:	Metagabbro (unit 3) of the Kaladar Complex.
C75-029:	Chlorite-actinolite schist (unit 2b) of the Kaladar Complex.

Figure 33. AFM plot for the rocks of the Kaladar Complex. A: $\text{Na}_2\text{O} + \text{K}_2\text{O}$; F: total iron; M: MgO .

Figure 33. Alkali-silica plot for the rocks of the Kaladar Complex. Field division after MacDonald, 1968.

Samples 238 and 243 fall below the well defined calc-alkaline trend and both belong to the anorthositic gabbro of unit 4a. Transitional contacts between this unit and the gabbros of unit 3 are visible in the field. These field and chemical data point to the possibility of two differentiation trends, the first a gabbro-anorthosite trend, and the second, more widely represented calc-alkaline gabbro-diorite-tonalite-granodiorite trend. Rocks belonging to the early trend are characterized by coarse grain size, subidiomorphic plagioclase laths with interstitial unaltered pyroxene and a lack of quartz and biotite. It is interesting to note that on the southern contact, it is rocks of this suite that occur in fragmental volcanic units. It is proposed that the gabbro-anorthosite trend is the result of a physical process of gravity fractionation whereas the calc-alkaline trend is the result of chemical differentiation.

That the gabbro-anorthosite suite formed earlier than the calc-alkaline rests on two lines of evidence: 1) that quartz-normative diorite samples contain fragments of gabbro (Plate 7d) and 2) that the volcanics (unit 1) contain fragments of the gabbro-anorthosite trend, but phases of the calc-alkaline trend contain amphibolite fragments.

That the two suites are contemporaneous is indicated by the transitional contacts between the hornblende-gabbro-anorthositic gabbro on one hand and the gabbro-diorite-quartz diorite-granodiorite on the other.

That the two types of ultramafic rocks described earlier represent a fundamental difference in initial rock type rather than a variation in metamorphic grade is supported by a comparison of analyses C74-236 and C74-237 (talc schist and serpentinite respectively) with C74-137, C75-008, and C75-029 (chlorite-actinolite schists). The former are significantly higher in Cr, Ni and MgO, and lower in CaO (Tables 11 and 12). The

chlorite-actinolite schists are chemically similar to the hornblendite portions of the metagabbro (cf. C74-172), supporting the contention that they are sheared equivalents of the latter.

In Figures 32 and 33, it will be observed that samples of the Elzevir Batholith (Gr-136, C74-145, C74-242) fall along the trends defined by the samples of the Kaladar Complex. Although the Elzevir Batholith resembles the leucocratic members of the Kaladar Complex chemically, field and textural relationships described earlier suggest that they are not closely related.

Timing and Effects of Processes

Because the age relationships between the Kaladar Complex and the surrounding rocks are well established, as are the metamorphic and tectonic processes which affected these rocks (Thompson, 1972; Moore and Thompson, 1972; Brown et al., 1975), an excellent opportunity is afforded to study these affects, in addition to the deuteric processes which appear to have played a role in the evolution of the Kaladar Complex. Textures and structures permit some elaboration on the established sequence of events.

The pyroxenes of the metagabbro and metadiorite of the Kaladar Complex are completely altered to amphiboles, which occur either as single large crystals of blue-green hornblende overgrowing relict pyroxenes that are often outlined by opaque dustings, or as aggregates of pale green fibrous amphiboles that are often partly recrystallized to small idioblastic grains. The actinolite schists of unit 2 (Figure 31) are composed entirely of fibrous amphiboles.

Card and Pattison (1975) in discussing the Nipissing Diabase suggested the following reaction for the above process:

(1)

This reaction would explain the opaque dustings (now hematite) that are observed to be outlining relict pyroxenes within large amphibole crystals.

Relict plagioclase crystals within the metagabbro and metadiorite are clouded with alteration minerals that are too fine to be identified. While some recrystallization to a granoblastic mosaic of untwinned plagioclase has occurred along the margins of these large crystals, mortar textures and bent or kinked twins are more common. This suggests that the Kaladar Complex remained relatively unreactive during the deformational episodes that effected the adjacent rocks.

That the solutions responsible for the uralitization process were of deuteric rather than metamorphic origin is supported by the fact that the diorite was not strained during the uralitization process. The close association of metamorphism and deformation in the Tudor and Flinton rocks are documented, and had the diorite been wetted with the uralitizing solutions during these tectonic events, it would have developed a penetrative fabric. On the contrary, the effects of deformation can be seen only sporadically in shear zones in the diorite and gabbro of the complex, where chlorite-actinolite schists indicate a lower temperature regime than the hornblende-plagioclase assemblages formed deuterically.

If the argument that the majority of the alteration of the mafic phases of the complex is deuteric is accepted, why they were largely unaltered during subsequent metamorphic and tectonic events then must be examined. The interaction between the Kaladar Complex and the fluid phase is the key to the problem. The effects of the fluid on rocks (as a medium for metamorphic reactions, a lubricant to facilitate strain, and as an agent of heat transfer) and the source of the fluid phase (metamorphic reactions, meteoric water) must be taken into account.

The gabbro and diorite were largely unreactive during the first metamorphism of the adjacent volcanic rocks and were not strained. The Flinton deformation, a much more important regional event, affected the diorite to some extent, causing a penetrative foliation near its border and sporadic shear zones away from it.

Temperatures within the diorite did not reach the levels necessary to reverse the reactions responsible for the uranization process, and it can be shown that the metamorphic grade increases towards the Flinton contact (see following discussion). It is probable that the impermeability of heat conducting fluids is responsible to some extent for this feature.

After the deuteritic alteration of the Kaladar Complex, the area underwent a phase of deformation and metamorphism characterized in the Tudor metabasalts by an east-west trending subvertical foliation, near vertical mineral lineation and the mineral assemblages hornblende-plagioclase. Although the mafic rocks of the Kaladar Complex are unaffected by this episode, both satellite gabbroic intrusions in the Tudor basalt (Figure 31) and the leucocratic member of the complex possess the east-west trending foliation.

After this pre-Flinton episode, which is brought to light only by detailed mapping; uplift and erosion led to the deposition of the Flinton Group (Thompson, 1972). The Flinton Group was then subjected to two major phases of deformation, coupled with a metamorphic event spanning the upper greenschist and amphibolite facies. The effects of this metamorphism and deformation on the rocks of the Kaladar Complex are penetrative near the Flinton contact, but rather sporadic away from it.

The east-west foliation in the Tudor basalts is crenulated around north-east trending axes parallel to the trend of the Flinton Group, and

these are interpreted to be related to the Flinton deformation. Within the Kaladar Complex, sporadically distributed northeast-trending shear zones are the result of this deformation. In these zones, the hornblende-plagioclase assemblage and hypidiomorphic granular texture are completely destroyed, resulting in a actinolite-chlorite schist. The alteration along these zones, probably triggered by the introduction of water along cracks in the rock, can be explained by reactions of the type:

These zones have sharp contact (over a few cm) and are always under 30 m wide.

Although the Kaladar Complex occurs in the staurolite-kyanite zone of regional metamorphism as defined by metapelites in the adjacent Flinton Group (Thompson, 1972), mineral zonation occurs within the Kaladar Complex which indicates a decrease in metamorphic grade away from the Flinton contact. Although a comprehensive study of these assemblages is beyond the scope of this presentation, they will be discussed briefly.

Small ultramafic bodies in the western part of the complex on the Skootomata River show assemblages talc-serpentine-magnesite-quartz, whereas adjacent the Flinton Group, ultramafic rocks consist of anthophyllite schists. In the shear zones that parallel the Flinton trend, best exposed in the northwestern region of the complex, chlorite-actinolite (and locally garnet) are indicative of lower grade of metamorphism than staurolite-kyanite at the Flinton contact. It is interesting to speculate that this gradient is in part due to the impermeability of the Kaladar Complex to heat conducting fluids released during the Flinton metamorphism.

Regional Implications

The regional implications of the age relationships between the

Kaladar Complex and its surroundings are very important. As shown previously, the Kaladar Complex belongs to the early diorite group of Lumbers (1967a); it is the oldest unit yet described, in part predating the Tudor basalts. Although one could argue that the Kaladar Complex is unrelated to the early diorite group, the correspondence of its attributes to those characteristic of that group, suggests otherwise.

Bearing in mind the fact that without detailed mapping of the Kaladar Complex, its age relationships would not have been discovered and it would have been mapped as a late to post-tectonic intrusion; the arguments for the post-tectonic emplacement of the diorite group must be examined.

Two arguments have been made by Lumbers (1967a) for the age of these bodies. Firstly, he points out that fold axes tend to be deflected around these intrusions and postulates that this is evidence of the forceful intrusion and shouldering aside of country rock. Another equally plausible explanation for this feature is that these intrusions acted as rigid bodies during the deformation, and the deflection of the fold axes is analogous to the deflection of a metamorphic fabric around a pre-tectonic rigid porphyroblast.

Lumbers (1967a) second line of evidence for the plutons being late in the cycle of events is that they possess contact aureoles. The volcanic rocks immediately adjacent to the southern contact of the Kaladar Complex also appear hornfelsic in the field, but in view of the age relations discussed above, this must be a relict feature. Perhaps the hornfelses described by Lumbers (1967a) are akin to these. In the light of this present study, interpretations of all the bodies of the diorite group should be reexamined.

Conclusions

The Kaladar Complex is the oldest rock unit that has been discovered in the Grenville Province of southeastern Ontario. Based on field evidence, it at least in part predates the Tudor volcanics, previously considered by Lumbers to be the oldest units in the same area. This study also emphasizes the care that must be taken in assigning a relative age to a plutonic body based on the absence the effects of deformation and metamorphism.

This study provides detailed evidence of a mafic plutonic basement to the Tudor volcanics, previously considered to be the oldest rocks in the area. The earliest sialic rocks exposed in this area postdate both the Tudor volcanics and the mafic plutons. There is no evidence of sialic basement.

Lumbers (1967a) was the first to suggest that the Tudor volcanics were erupted in an island arc environment. The chemistry of the Tudor volcanics (Sethuraman and Moore, 1973; Brown et al., 1975; and Condie and Moore, 1977) further supports an ensimatic island arc origin. This study further supports an ensimatic origin for the Tudor volcanics.

REFERENCES

- Brown, R. L., Chappell, J. F., Moore, J. M., Jr, and Thompson, P.H. 1975. An ensimatic island arc and ocean closure in the Grenville Province of southeastern Ontario, Canada. *Geoscience Canada*, 2, pp. 141-154.
- Card, K. D., and Pattison, E. F. 1973. Nipissing Diabase of the Southern Province, Ontario. In *Huronian stratigraphy and sedimentation* (G. M. Young, ed.). Geological Association of Canada Paper Number 12, pp. 7-30.
- Condie, K. C. and Moore, J. M. 1977. Geochemistry of Proterozoic volcanic rocks from the Grenville Province, eastern Ontario. In: *Volcanic regimes in Canada* (W. R. A. Baragar, L. C. Coleman, and J. M. Hall, eds.), Geological Association of Canada, Special Paper Number 16, pp. 149-168.
- Harding, W. D. 1944. Geology of Kaladar and Kennebec Townships. Ontario Department of Mines, Annual Report 51, Part 4, 1942, p.p. 51-74.
- Irvine, T. N. and Baragar, W. R. A. 1971. A guide to the chemical classification of the common volcanic rocks. *Canadian Journal of Earth Sciences*, 8, pp. 523-548.
- Lumbers, S. B. 1967a. Stratigraphy, plutonic activity, and metamorphism of the Ottawa River remnant in the Bancroft - Madoc area of the Grenville Province of southeastern Ontario. Unpublished Ph.D. thesis, Princeton University, Princeton, New Jersey.
- Lumbers, S. B. 1967b. Geology and mineral deposits of the Bancroft - Madoc area. In *Guidebook to geology of parts of eastern Ontario and western Quebec* (S. E. Jenness, ed.). Geological Association of Canada, Kingston, Ontario, pp. 13-30.
- MacDonald, G. 1970. Geology of Kingston area: a compilation. Carleton University Geological Paper 70-2. Ottawa, Canada.
- MacDonald, G. A. 1968. Composition and origin of Hawaiian lavas. *Geological Society of America. Memoir* 116. pp. 477-522.
- Miller, W. G. and Knight, C. W. 1914. The Precambrian geology of southeastern Ontario. *Ontario Bureau of Mines*, 22, part 2, 151 p.
- Moore, J. M., Jr. and Thompson, P. H. 1972. The Flinton Group. Grenville Province, eastern Ontario, Canada. *Proceedings of the twenty-fourth International Geological Congress, Section I*, pp. 221-229.
- Sethuraman, K. and Moore, J. M., Jr. 1973. Petrology of metavolcanic rocks in the Bishop Corners - Donaldson area, Grenville Province, Eastern Ontario. *Canadian Journal of Earth Sciences*, 10, pp. 589-614.
- Silver, L. T. and Lumbers, S. B. 1966. Geochronological studies in the Bancroft - Madoc area of the Grenville Province, Ontario, Canada. (abstract). *Geological Society of America, Special Paper* 87, p. 156.

Smith, B. L. 1958. Geology of the Clarendon-Dalhousie area. Ontario Department of Mines Annual Report, 65, Part 7, pp. 1-46.

Thompson, P. H. 1972. Stratigraphy, structure and metamorphism of the Flinton Group in the Bishop Corners - Madoc area, Grenville Province, Eastern Ontario. Unpublished Ph.D. thesis, Carleton University, Ottawa.

Wynne-Edwards, H. R. 1972. The Grenville Province. In: Variations in tectonic styles in Canada (eds. R. A. Price and R. J. W. Douglas). Geological Association of Canada Special Paper 11, pp. 264-334.

PLATE 7

Field Photographs of the Kaladar Complex

- a) Subhedral plagioclase laths showing primary alignment in the metadiorite of the Kaladar Complex. The interstitial mafic mineral is amphibole, replacing pyroxene.
- b) Metagabbro and other accidental fragments in a fragmental volcanic unit near the southern contact between the Kaladar Complex and the Tudor metavolcanics. Note the similarity between the fragments and the diorite pictured in (a).
- c) Amphibolite fragments in the leucocratic member of the Kaladar Complex. The mafic minerals in the host consist of clots of fine flakes of biotite replacing hornblende.
- d) Angular fragments of metagabbro enclosed in the metadiorite of the Kaladar Complex.
- e) Metadiorite (light areas) enclosing orange-brown weathering fragments that are similar in composition, but are more highly altered. These fragments preserve relict hydrothermal effects, and the brecciation illustrates episodic igneous activity.
- f) Dyke of Elzevir granodiorite cutting foliation (parallel to the hammer handle) in the Tudor metavolcanics.
- g) Pink aplitic border phases of the Elzevir Batholith brecciating the Tudor metavolcanics.
- h) Diorite dyke cutting foliation in the Tudor metavolcanics.

SYMBOLS

Analyzed sample (1974,1975)

Geological Contact(defined, approximate, transitional)

Outcrop

Pre-Flinton foliation, lineation

Stream

Building

Road

FIGURE 31. GÉOLOGY OF THE KALADAR COMPLEX

LEGEND

Flinton Group Metased

9

UNCONFORM

Elzevir Batholith: mainly
quartz diorite

8

INTRUSIVE COM

KALADAR COMPLEX

7

Granodiorite and quartz

6

Leucocratic Unit, colour
quartz diorite, granodior

5

Andesite porphyry with

4

Metadiorite, colour index
4a) Anorthositic gabbro

3

Metagabbro, colour index
deformed coarse-grained
intruding Tudor metavol

2

Ultramafic Rocks: 2a) Ta
schist, serpentinite; 2b)

TUDOR METAVOLCANI

1

Amphibolite: mainly fir
locally pillowed, locally

Compiled from mapping by Chappell (1974, 75), T
and 1975 Carleton field camp.

LEGEND

Flinton Group Metasediments

UNCONFORMITY

Elzevir Batholith: mainly granodiorite and quartz diorite

INTRUSIVE CONTACT

KALADAR COMPLEX

Granodiorite and quartz diorite dykes

Leucocratic Unit, colour index < 20 . Includes quartz diorite, granodiorite and quartz monzonite

Andesite porphyry with plagioclase phenocrysts

Metadiorite, colour index between 20 and 40.

4a) Anorthositic gabbro.

Metagabbro, colour index > 40 3a) Highly deformed coarse-grained amphibolite intruding Tudor metavolcanics.

Ultramafic Rocks: 2a) Talc schist, anthophyllite schist, serpentinite; 2b) chlorite-actinolite schist

TUDOR METAVOLCANICS

Amphibolite: mainly fine grained, uniform; locally pillowed, locally fragmental

Compiled from mapping by Chappell (1974, 75), Thompson (1972), and 1975 Carleton field camp.

IV

V

483

824 CR12

CR13

CR14

CR15

CR16

FIGURE 28

CATIONS IN THE CLARE RIVER STRUCTURE

1974

1975

1976

1977

V Subarea boundary and number

— Selected geologic contact

TATION NUMBERS CORRESPOND FOR 1975, 76, & 77

SAMPLES GIVEN IN APPENDIX V

IENTED SPECIMENS, 1974

LES- ANALYZED BY PEARSON (1958)

D SAMPLE

KM

1

2

0

1

MILES

140F14

Plan view of contact between the Addin
Gneiss and the Clare River Supracrustal

LEGEND

SUPRACRUSTAL SUCCESSION OF THE CLARE RIVER STRUCTURE

FLINTON GROUP

BOGART FORMATION (18-23)

- 22 Uniform calc-silicate paragneiss
- 23 Dolomitic marble, rusty paragneiss
- 24 Calc-silicate paragneiss with quartzofeldspathic stringers
- 25 Rhythmically layered calc-silicate paragneiss
- 26 Calcareous marble

BEATTY FORMATION (16-17)

- 16 Rusty graphitic psammite
- 17 Pelitic schist

UNCONFORMITY

HERMON GROUP

8 SHOVEL LAKE FORMATION

TODD LAKE FORMATION (6-7)

- 7 Dolomitic marble, calc-silicate paragneiss, para-amphibolite metachert, volcanic pebble metaconglomerate
- 6 Dolomitic marble

1 OTTER CREEK FORMATION

KILOMETRES

MILES

* See text under subarea 1

G'

E

F

80F

FIGURE 6: SERIAL SECTIONS THRO

ROUGH THE CLARE RIVER STRUC

100F

RUCTURE

110F

HORIZONTAL SCALE = VERTICAL SCALE

D1

C

D2

D1

15

43

70

31

30F

D2

4

D1

D2

7

44 40

25

35

42

30

50F

D2

D1

8

KALADAR

STOCO LAKE

CLARE RIVER

D1

130F

Subarea boundary

Bedding (S0), tops

S1 foliation (inclined)

L1 lineation

F1 fold axis, vergence

S2 foliation (inclined)

L2 lineation

F2 fold axis, vergence

Axial surface trace

Axial surface trace of

Axial surface trace of

Axial surface trace of

Axial surface trace of

Selected geological

LEGEND

boundary, number	— 2
, tops unknown	— T
n (inclined, vertical)	— + +
	— >
s, vergence	— < >
n (inclined, vertical)	— + +
n	— +
s, vergence	— < >
e trace of overturned F1 synform (defined, assumed)	— U —
e trace of overturned F1 antiform (defined, assumed)	— ∩ —
e trace of F2 synform	— T —
e trace of overturned F2 synform (defined, assumed)	— U —
e trace of overturned F2 antiform (defined, assumed)	— ∩ —
ological contacts	— —

140F

D1

2

D2

D1

FIGURE 4 : STRUC

5

D2

STRUCTURAL GEOLOGY OF THE C

D1

D2

CLARE RIVER STRUCTURE

190 F

S1 foliation (incl
L1 lineation
F1 fold axis, ve
S2 foliation (incl
L2' lineation . .
F2 fold axis, ve
Axial surface trac
Axial surface trac
Axial surface trac
Axial surface trac
Axial surface trac
Selected geologic
Stereonet diagram

Lake (permanent)
Stream
Road
Cart track

Notes: 1) Defined
2) For clar

defined folds based on stratigraphy

for clarity, some intervening F1 synforms in subareas 4, 7, and 8, are omitted.

BOGART

LEGEND

70f

14

80°
48°

90F

QUEBEC

76°

72°

48°

8

100F

OTTER CREEK

BRUNO LAKE

13 OF

- Geological boundary
- Rock outcrop
- Fault
- Limit of geological
- Road
- Cart track
- Stream
- Lake (permanent)
- Geological boundary
- Geology by J. F. C.

PALEOZOIC

25 Basal sandstone

24 Pegmatite

FLINTON GROUP

BOGART FORMATION (18-23)

22 Uniform calc-silicate paragneiss

21 Dolomitic marble, rusty paragneiss

20 Calc-silicate paragneiss with quartz stringers

19 Rhythmically layered calc-silicate

18 Calcareous marble

BEATTY FORMATION (16-17)

17 Rusty graphitic psammite

16 Pelitic schist

140A

METAMORPHOSED PL

14 Leucocratic quartz monzonite

13 Biotite granodiorite

12 Amphibolite dykes

SYMBOLS

ary (defined, approximate, assumed)

al mapping

, seasonal)

ary (transitional)

map 1974-1977

HELIKIAN

HERMON GROUP

SHOVEL LAKE FORMATION (8-11)

11 Biotite schist and phyllite, calc-

8 9 10 Felsic metavolcanics layered massive quartzofeldspathic gneiss color index < 10

9 Intermediate metavolcanics, gneiss, hornblende biotite gneiss.

LEGEND

UNCONFORMITY

UPERGROUP AND ASSOCIATED PLUTONIC ROCKS

INTRUSIVE CONTACT

3-23)

paragneiss

y paragneiss

ss with quartzofeldspathic

la-silicate paragneiss

-17)

ite

23

Calc-silicate paragneiss pegmatite

15

Granitic pebble metaconglomerate

UNCONFORMITY

ED PLUTONIC ROCKS

nzonite

ADDINGTON PLUTON

MELLON LAKE PLUTON

INTRUSIVE CONTACT

INTRUSIVE CONTACT

ON (8-11)

ite, calc-silicate paragneiss

s layered, fragmental and
athic gneisses; biotite gneiss

anics biotite hornblende
gneiss hornblende gneiss

150F

LEGEND

UNCONFORMITY

UP AND ASSOCIATED PLUTONIC
ROCKS

INTRUSIVE CONTACT

athic } 23 Calc-silicate paragneiss, pegmatite

15 Granitic pebble metaconglomerate

UNCONFORMITY

ROCKS

ADDINGTON PLUTON

MELLON LAKE PLUTON

INTRUSIVE CONTACT

INTRUSIVE CONTACT

gneiss

and
gneiss

gneiss

INDEX MAP

GEOLOGY OF THE CLARE RIVER STRUCTURE SOUTHEASTERN ONTARIO

- Geological bo
- Rock outcro
- Fault
- Limit of geolo
- Road
- Cart track
- Stream
- Lake (perma
- Geological b

Geology by J

E

~~Base map en~~
topographic

S

SYMBOLS

geological boundary (defined, approximate, assumed)

rock outcrop

road

unit of geological mapping

road

road track

stream

lake (permanent, seasonal)

geological boundary (transitional)

compiled by J F Chappell, 1974-1977

This map enlarged from 1:50,000 federal
topographic maps 31C/11 and 31C/6

HELIKIAN

17 Rusty graphitic ps

16 Pelitic schist

METAMORPHIC

14 Leucocratic quartz

13 Biotite granodiorite

12 Amphibolite dykes

HERMON GROUP

SHOVEL LAKE FORMATION

11 Biotite schist and

8 9 10 Felsic, metavolcanic
massive quartzofeldspathic
color index < 10

9. Intermediate mafic gneiss, hornblende gneiss, generally fragmental
index 10-35

8. Mafic metavolcanic, locally pillowed, coarse fragmental amphibolite

TODD LAKE FORMATION

7 Dolomitic marble, biotite, metachert;

6 Dolomitic marble

OTTER CREEK FORMATION

5 Calcareous marble

4 Rusty graphitic ps

3 Quartzofeldspathic

2 Biotite-hornblende, minor amphibolite

1 Massive felsic m

A Marble, calc-siliceous

graphitic psammite

schist

15

Granitic pebble metaconglomerate

UNCONFORMITY

MORPHOSED PLUTONIC ROCKS

atic quartz monzonite

ADDINGTON PLUTON

granodiorite

MELLON LAKE PLUTON

INTRUSIVE CONTACT

bolite dykes

INTRUSIVE CONTACT

N GROUP

LAKE FORMATION (8-11)

schist and phyllite, calc-silicate paragneiss

fic metavolcanics: layered, fragmental and
e quartzofeldspathic gneisses; biotite gneiss;
index < 10

mediate metavolcanics: biotite-hornblende
, hornblende-biotite gneiss, hornblende gneiss,
illy fragmental, locally porphyritic; color
10-35

c metavolcanics: massive fine grained amphibolite,
pillowed, coarse grained amphibolite, minor
ental amphibolite

LAKE FORMATION (6-7)

tic marble, calc-silicate paragneiss, para-amphi-
, metachert; 7a, volcanic pebble metaconglomerate

itic marble

CREEK FORMATION (1-5)

reous marble

graphitic psammite and pelite

zofeldspathic psammite, pelitic schist

e-hornblende gneiss, hornblende-biotite gneiss,
amphibolite

ve felsic metavolcanics

le, calc-silicate paragneiss, pegmatite

CONFORMITY
ROCKS

ADDINGTON PLUTON

MELLON LAKE PLUTON

SIVE CONTACT

SIVE CONTACT

ss

ss;

polite,

phi-
nerate

ss,

END

06 03 80

FIN