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STRATIGRAPHY, STRUCTURE AND METAMORPHISM
OF THE FLINTON GROUP IN THE BISHOP CORNERS-
MADOC AREA, GRENVILLE PROVINCE, EASTERN ONTARIO

by

PETER H. THOMPSON, BSc.

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

Department of Geology
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Ottawa, Ontario
June, 1972

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ABSTRACT

The Flinton Group, part of the Grenville Supergroup, comprises metamorphosed carbonate and clastic sedimentary rocks. In the study area, it is exposed in narrow isoclinal synclines for 40 km along an easterly-to northeasterly-trending belt. These rocks unconformably overlie plutons, metavolcanic rocks, marble, and metagraywacke. Deposition of Flinton rocks is estimated to have occurred in the period between 1075 and 1100 m.y. ago. A transition from pelitic schist and marble at Bishop Corners and at Madoc to quartzite metaconglomerate and quartzofeldspathic psammite between these localities is interpreted to be a facies change between a relatively low-energy, shallow-water depositional environment and a higher energy, fluvial-deltaic-beach environment.

The structural geometry of the Flinton Group can be explained, for the most part, by three phases of deformation. The regional foliation (S_1) is parallel to the axial surfaces of first phase isoclinal folds characterized by continuity for tens of km parallel to the strike of the axial surface and by fold axes with variable plunge. These structures were refolded during the second phase of deformation about axial surfaces (S_2) with a northeast trend. Both S_1 and S_2 are warped by a third phase of folding with northwest-trending axial

surfaces. A gently-dipping crenulation in micaceous rocks and evidence of cataclasis in the Flinton Group are also present. Metamorphic grade was increasing during the first phase of deformation (D_1), attained a maximum just before or during D_2 , and began to decrease before D_3 .

Mineral zoning is clearly defined in calcareous and pelitic rocks of the Flinton Group. Textural and chemical criteria indicate the rocks attained a state close to chemical equilibrium during metamorphism. The application of linear programming to chemical analyses of all the minerals in six samples from the Flinton Group demonstrates that the mineral zoning is the result of a gradient in physical conditions at the time of metamorphism. Bulk composition was not an important variable with respect to the development of mineral zoning. Metamorphic grade increases from that typical of the lowermost amphibolite facies at Bishop Corners and at Madoc to upper middle amphibolite facies in the central part of the study area.

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Dr. T.M. Gordon of the Geological Survey of Canada made possible the application of linear programming in this study. Dr. Richard L. Brown assisted in removing the shroud of mystery, at least in part, from the structural geology in the study area. Thermodynamics became understandable and useful to the writer under the tutelage of Dr. G.B. Skippen. Professor J.A. Donaldson and Dr. Brian Jones contributed helpful criticism in the areas of sedimentology and stratigraphy. Dr. David H. Watkinson and Paul Mainwaring introduced the writer to microprobe analysis.

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I INTRODUCTION

General Statement

A study of the metamorphism of a sequence of sedimentary rocks cannot be separated from an examination of their sedimentological characteristics and structural geometry. Furthermore, the structural geometry cannot be completely understood without knowledge of the stratigraphy. Within the Grenville Province of eastern Ontario, the Flinton Group, part of the Grenville Supergroup (Moore and Thompson, in press), comprising metamorphosed carbonate and clastic sedimentary rocks, provides an excellent opportunity for such a study. Primary sedimentary structures such as cross-bedding are preserved; relatively simple structural geometry exists; rock units are continuous for tens of kilometers from areas of lowest metamorphic grade to the area of highest grade; and mineral zoning is clearly defined in calcareous and pelitic rocks.

Purpose of Study

The objectives of this study are:

- 1) To describe the stratigraphy and structural geometry of the Flinton Group and the mineralogical vari-

ation in these rocks in the Bishop Corners-Madoc map area (Figure 1);

2) To interpret the time and environment of deposition and infer the provenance area of these rocks;

3) From the structural geometry and petrographic evidence interpret the structural history and the timing of deformation with respect to regional metamorphism;

4) To test the validity of the assumption of equilibrium during metamorphism, and to determine if variation in bulk composition is an important variable in the development of the mineral zoning.

Methods of Study

Detailed mapping on a scale of 1:15,340 (one inch = one-quarter mile) was done over an area of approximately 110 km². Outcrop distribution and rock unit contacts were recorded in the field on aerial photographs and the data transferred to geologic maps of the same scale (Figures 2 and 3). Petrography was concentrated on the rocks of the Flinton Group, but thin sections of other rock units in the area were examined as well. Mineral assemblages from each thin section are included as Appendix 3. The chemical composition of all the minerals in six rocks, two each from calcareous psammite, carbonate schist, and pelitic schist units of the Flinton Group were determined by electron microprobe. An IBM linear

Figure 1: Location map and distribution of the Flinton Group (solid pattern, and of rocks tentatively correlated with the Flinton Group (striped pattern). Geology mainly from G. MacDonald (1970).

programming package was used in the assessment of bulk composition as a variable in the development of mineral zoning in these rocks.

Location and Access

The Bishop Corners-Madoc map area is located 200 km southwest of Ottawa, Ontario near the southeastern border of the Grenville Province (Figure 1). The area includes parts of Anglesea, Kaladar, Elzevir and Madoc Townships. Madoc is a major population centre in the region (1400 souls). Bishop Corners refers to the junction of a gravel road with Highway 41 (Figure 2, UTM 279608).¹ Access is good, with two major highways, 41 and 7, passing through the area and a network of gravel-surfaced secondary roads in the intervening areas.

General Geology

Metamorphosed rocks of the Elzevir and Northbrook plutons and mafic metavolcanic rocks intruded by the plutons are the predominant rock types in the Bishop Corners-Madoc map area. Metamorphosed graywacke, pyroclastics and dolostone are associated with the volcanic rocks. The Finton Group unconformably overlies the plutonic rocks. Regional metamorphism, ranging from con-

¹ The Universal Transverse Mercator grid is superimposed on the geologic maps of the study area to facilitate the designation of specific localities.

ditions typical of the lowermost amphibolite facies to the upper middle amphibolite facies, and three major phases of deformation have affected all the rocks in the area. Detailed lithologic descriptions of the units in the Flinton Group are included in Chapter II. The following section briefly describes rock units, defined on the geologic maps (Figures 2 and 3), which are not part of the Flinton Group.

Lithology

It should be noted that many of the rock units observed in the Flinton area (Figure 2) are not present in the Madoc area (Figure 3). To avoid the occurrence of gaps in the numbering of rock units in the Madoc area a separate legend was introduced. In this section, rock units present in the Flinton area will be described first. Unit numbers are the same as those on the geologic maps (Figures 2 and 3).

Flinton and Black River map areas

Unit 1. Gray-green to dark green-weathering, fine-grained, massive amphibolite is considered to be the oldest rock in the area. Locally, pillows and fragmental textures are preserved. Mineral composition is variable, but hornblende and plagioclase compose most of the rock in all cases. Hornblende grains are generally randomly

oriented in the plane of the schistosity (S_1). Contacts with other rock units are sharp. The homogeneity and lateral extent of these rocks suggest they were originally mafic volcanic flows.

Unit 2. These rocks weather pale green, or commonly a distinctive splotchy white and green. Chlorite and plagioclase are more prominent and the rocks are more heterogeneous than those in Unit 1. Relict fragmental textures are common. Some rocks definitely are metamorphosed agglomerates. Where Unit 2 occurs near the contact of the Flinton Group, the rock commonly exhibits a gneissic character, with layers (5 - 10 mm thick) of massive hornblende interspersed with plagioclase-rich layers. The heterogeneity of Unit 2 suggests the rocks are probably derived from pyroclastics and graywacke-type sediments associated with the volcanic flows of Unit 1.

Unit 2A (hornblende biotite quartzofeldspathic schist) is distinguished on the basis of its mineralogy and appearance in outcrop. The gray-weathered surface is characterized by discontinuous ridges 5 - 15 cm wide. Biotite is the predominant mafic mineral. Plagioclase and quartz compose the bulk of the rock. Locally well-defined layers (2 - 10 cm thick), continuous for tens of meters, are made up entirely of aggregates of hornblende. The layered aspect of this rock and the high plagioclase content compared to associated amphibolite indicate the

schist is probably derived from sedimentary or pyroclastic rocks.

Unit 3. Metavolcanic rocks east of Highway 41 near Bishop Corners (Figure 2) contain more feldspar, quartz, and biotite and less hornblende than Unit 1. The rock is distinctly gray to grayish-beige on the weathered surface. Fragmental and amygdaloidal textures are preserved. Pillow structures are rare; relict porphyritic textures occur locally. Unit 3 appears to be derived from an association of massive and pyroclastic rocks of andesitic composition.

Unit 4. Although salic metavolcanic rocks are widely distributed to the northeast (Sethuraman, 1970), in the Flinton map area (Figure 2) they are not common. Adjacent to the basal contact of the Flinton Group southeast of Bishop Corners a striking purplish-white, magnetite-rich rock occurs. Ellipsoidal-shaped features suggestive of spherulites are present. Much of the rock exhibits a fragmental texture. Thickness of the unit is less than 20 m and the lateral extent 200 - 300 m. Rare occurrences of pink-weathering, aphanitic, discontinuous layers were observed associated with Units 1 and 2. Most of these are too small to be represented on the geologic map.

Unit 5. Gray- to beige-weathering marble is common at Bishop Corners, in the southern part of Figure 2, and

east of Actinolite (Figure 3). Calcite marble (5a) is rare but where present it is a medium-grained white- or gray-weathering rock. Aphanitic or fine-grained dolomite marble (5b), locally containing tremolite, diopside or epidote, composes most of the marble outcrops in the study area. Compositional layering is well-defined in some cases but is generally absent. Boudined layers of quartzite are common in the southern part of the Flinton area (Figure 2) and eastern Black River area (Figure 3). Massive aggregates of tremolite and diopside are associated with the quartzite.

Unit 6. Unit 6 is composed of: black pyritic phyllite (6b) in the Bishop Corners area, and calcareous schist, calcareous psammite (6a) and volcanic pebble metaconglomerate (6c) in the Black River area. The black-weathering phyllite is characterized by thin bedding or laminations (5 - 30 mm thick), a fissile habit, and high sulphide content. The calcareous schist and psammite and associated volcanic pebble metaconglomerate weather gray-green or brown and generally contain more epidote and hornblende than the calcareous rocks of the Flinton Group in the area. Cross-bedding and structural geometry indicate Units 6a and 6c underlie the Flinton rocks. The volcanic pebble conglomerate is poorly sorted with respect to size and composition of clasts. A wide assortment of metavolcanic pebbles and cobbles is present ranging from mafic to

salic in composition. Some clasts exhibit a striking relict porphyritic texture. Pink calcite marble clasts also occur in the rock. The matrix is composed of biotite, white mica, quartz, carbonate, Fe-Ti oxide and plagioclase. Units 6a and 6c are thought to be sedimentary rocks, probably with some pyroclastic material, which are associated with the massive amphibolites in the Black River area.

Unit 7. Southwest of Flinton (Figure 2) a large body (3 - 4 km²) of greenish-gray-weathering hornblende diorite is bounded by metavolcanic rocks on the north and south, Flinton Group on the east, and granodioritic rocks to the west. Fine- to medium-grained, the rock is homogeneous. Two foliations defined by mineral lenticles are observed near the eastern contact. Plagioclase and amphibole compose most of the rock. In places an hypidiomorphic granular texture is preserved. Along the eastern margin of the body a coarse-grained variety of the diorite (7a) is present, characterized by randomly-oriented euhedral plagioclase crystals and interstitial plumose amphibole. The rock has a distinctive white and green weathered surface. Contacts between the diorite and other rock units are not exposed. However, small fine-grained granodioritic dikes are seen to cut the diorite.

Unit 8. The most abundant rock unit in the study area is fine- to medium-grained metamorphosed biotite granodiorite (8a) of the Elzevir and Northbrook plutons. Plagioclase, quartz and biotite are the main constituents, with smaller amounts of microcline and epidote. Apatite, zircon, tourmaline and allanite are common accessories. The rock is generally homogeneous, even-grained and gray-weathering. Near contacts with mafic metavolcanic rocks, however, inclusions of amphibolite occur. A prominent foliation defined by mineral lentils is present in the Northbrook pluton west of the main belt of Flinton rocks (Figure 2). Two foliations, somewhat less prominent, occur in the granodiorites west of the Flinton rocks in the Elzevir pluton. Westward toward the core of this body the foliation disappears and hypidiomorphic granular texture predominates. In the Northbrook pluton (Figure 2) patches of pink-weathering biotite quartz monzonite (8b) occur. Contacts between 8a and 8b are sharp in the northeastern part of the area but more diffuse to the southwest. Contacts between plutonic rocks (8a, 8b) and the Flinton Group are sharp where they can be observed. The mineralogical similarities and close spatial relationships of the metamorphosed Elzevir and Northbrook granodioritic bodies indicate these rocks are probably part of the same large plutonic mass.

Unit 9. A small proportion of the map areas in Figures 2 and 3 is underlain by pink-weathering metamorphosed quartz monzonite (9a) of the Addington pluton. The fine- to medium-grained rock is characterized by tablets of quartz (1 - 2 cm long) which with biotite lenticles define a well-developed foliation. East of Actinolite (Figure 3) along the western contact of the Addington body, a dioritic variety is present.

Units 10 - 15. (Flinton Group - See Chapter II.)

Unit 16. Two fairly large bodies (1 - 2 km²) and several smaller masses (0.1 - 0.2 km²) of metamorphosed ultramafic rocks have been mapped in the study area. Buff- or red-brown-weathering, these rocks are composed mainly of serpentine and anthophyllite with smaller amounts of tremolite and talc. Small amounts of Fe-rich dolomitic marble and amphibolite are associated with the ultramafic rocks. As a result of the radial growth of anthophyllite in the thicker parts of the ultramafic body, in the central southern part of Figure 2, the rock is only weakly foliated or massive. Nearer contacts, a foliation is better developed. This body of ultramafic rocks transects formation contacts in the Flinton Group at a low angle (10° - 15°). Preliminary work on the body which occurs in the area of overlap between Figures 2 and 3 indicates the ultramafic rocks are deformed by the second phase of deformation in the region and are probably assoc-

iated with the mafic metavolcanic rocks (J.M. Moore, personal communication). From this information it would appear that the ultramafic rocks which cut the Flinton Group must have been emplaced tectonically. The presence of mylonite and intensely-strained quartz in the Flinton Group adjacent to the ultramafic body supports this contention.

Unit 17. Pink-weathering feldspar pegmatite occurs closely associated with diopside-bearing calc-silicate rocks of the Flinton Group (Figure 2) and with the contact between the Addington pluton and pre-Flinton meta-sedimentary rocks in the southern part of Figure 2. The remainder of this discussion is limited to the pegmatite associated with Flinton rocks. Pegmatite is restricted to the metamorphic mineral zone characterized by diopside in the calcareous rocks of the Flinton Group (see Chapter IV). Microcline and sodic plagioclase are the major mineral constituents of these rocks. Minor amounts of muscovite, biotite, magnetite and quartz are present. Associated with concentrations of biotite in these rocks are red euhedral garnets (1 - 10 mm). Grain size of the feldspar ranges from 2 to 200 mm most commonly, but crystals as large as 0.5 - 1 m long were observed. In thin section, the rock exhibits a granoblastic texture. Contacts with the Flinton Group are sharp and generally steeply-dipping.

Most of the pegmatite occurs within the limits of the calc-silicate rocks of the Flinton Group (Figure 2). Rarely does pegmatite cut across quartzite metaconglomerate. The presence of garnet, the spatial relationship to the calcareous metasedimentary rocks, and the lack of exotic minerals suggest the pegmatite is derived by partial melting of Flinton rocks.

Madoc area

Unit 1. Metavolcanic rocks in the Madoc area are similar to those described from Bishop Corners. The gray- or green-gray-weathering fine-grained rocks commonly exhibit relict amygdaloidal or fragmental textures. Several well-preserved flow contacts are preserved. Hornblende is present in minor amounts, biotite being the principal mafic mineral. These rocks are derived from a sequence of flows and pyroclastic rocks of andesitic composition.

Unit 2. Fine-grained, buff-weathering dolomitic marble, mainly massive, locally exhibiting a fragmental texture, is a prominent rock unit in the southern portion of the Madoc area (Figure 3). Gray-weathering, medium-grained calcitic marble with compositional layering is less common.

Unit 3. Clastic metasedimentary rocks not included in the

Flinton Group near Madoc are quite heterogeneous. Although locally thin bedding or laminations are present, generally the rock is a rather poorly-sorted breccia with a calcareous quartzofeldspathic matrix. Fragments are commonly salic metavolcanic material or carbonate.

Biotite, white mica, quartz, plagioclase, Fe-Ti oxides, and dolomite and/or calcite are the most common matrix minerals in these rocks.

Unit 4. Unit 4 is composed of metamorphosed rhyolitic dikes, salic metavolcanic rocks (felsite breccia), porphyritic plutonic rocks and one occurrence of meta-ignimbrite. Actinolite sheaves in dolomitic marble immediately adjacent to the contact of a rhyolitic dike indicate igneous activity post-dated the deposition of the marble. Nowhere in the area do these dikes cut Flinton rocks. All the rock types included in Unit 4 are composed mainly of potassic feldspar, quartz and smaller amounts of biotite. The rocks are commonly brecciated, but the brecciation appears to have occurred before the regional metamorphism. The fine-grained, gray-white weathering plug of metamorphosed porphyritic intrusive rock west of Madoc may be related to the intrusion of the dikes. The relationship of the salic breccias located north of the syncline containing the Flinton to the salic rocks near the town of Madoc is not known. Associated with these breccias at one locality

(Figure 3, UTM 027333), is an outcrop of gray-weathering very fine-grained felsite which contains beige-coloured inclusions very similar in form to the fiamme structures (flattened pumice fragments) which occur in ignimbrites (welded tuffs). It seems likely that the salic rocks in the Madoc area are related to one period of igneous activity which clearly post-dates the deposition of the dolomite marble unit.

Units 5 - 8 (Flinton Group, see Chapter 2).

II STRATIGRAPHY AND SEDIMENTOLOGY

Introduction

Geologists have worked sporadically for more than one hundred years on the complex sequence of metamorphosed sedimentary, volcanic, and plutonic rocks of the Bishop Corners-Madoc area, but because the stratigraphy is complicated by regional metamorphism and polyphase deformation, stratigraphic order has become apparent only recently. Murray (1853) started a prolonged controversy when he suggested that the conglomerates of the Madoc region should be included in a separate unit, the Hastings Series, rather than in the Grenville Series as defined by Logan (1846). Both authors considered these rocks to be Precambrian. In 1906 an International Committee concluded that the Hastings rocks were part of the Grenville Series. The name Hastings was preserved to indicate differences in metamorphic grade and lithology between the rocks of the Madoc region and the marble-quartzite-paragneiss association at the type locality of the Grenville Series. The first regional mapping in the study area (Miller and Knight, 1914) revealed an unconformity between Hastings conglomerate and plutons which intrude metavolcanic, and associated metasedimentary, rocks. The volcanics, thought to be the oldest rocks in the area, were correlated with the Keewatin Series of

northwestern Ontario, and the metasedimentary rocks with the Timiskaming Series. Later workers did not reach the same conclusions. Meen and Harding (1944) restored the Hastings rocks to the Grenville Series and observed that plutonic rocks intrude all the metasedimentary rocks. Burns (1951) and Ambrose and Burns (1956) concluded, on the basis of detailed mapping, that conglomerate occurs at least three times in the stratigraphic succession, and that grouping of the rocks into either the Hastings Series or the Grenville Series on the basis of lithology was therefore misleading. Furthermore, the plutonic rocks were interpreted to be the product of regional metasomatism which occurred after the deposition of the sedimentary rocks. Some clarification of this confusing assortment of hypotheses has resulted from detailed mapping carried out by J.M. Moore (1967), K. Sethuraman (1970) Sethuraman and Moore, (in press), and the writer.

Moore (1967) mapped metasedimentary rocks east of Bishop Corners (Figure 1) and established an angular unconformity between these rocks and the underlying metavolcanics. Sethuraman (Sethuraman and Moore, in press) described in detail the stratigraphy and petrology of the metavolcanic rocks in the same area. The author extended Moore's detailed mapping southwest to Madoc (Figures 2 and 3). The metasedimentary rocks were shown to overlies unconformably the Elzevir and Northbrook plutons, which intrude the older, mafic portion of the

metavolcanic rocks studied by Sethuraman. Rocks above the unconformity have been named the Flinton Group, part of the Grenville Supergroup, and the underlying metavolcanics, metagraywacke and marble have been correlated with Lumbers' (1967) Hermon Group (Moore and Thompson, in press).

The purpose of this chapter is to describe the stratigraphy of the Flinton Group in the Bishop Corners-Madoc region, to interpret the environment of deposition, and to infer the provenance of these rocks, thereby demonstrating the validity of such an approach in the regionally metamorphosed, complexly deformed rocks in the Grenville Province.

Stratigraphic Units.

Stratigraphic units mapped in the area are defined on the basis of lithologic constancy and position within the metasedimentary sequence. Lithologic units composed of a single rock type are grouped into formations on the basis of mineralogy, grain size, bulk composition, and inferred pre-metamorphic lithology. Rock units within a formation are typically less than 200 m thick and continuous for 10 to 15 km. The formations described in subsequent sections of the paper either pinch out or grade laterally and/or vertically into adjacent rock units. They will be discussed in the order in which they occur in the map area from northeast

to southwest. Mineral assemblages in thin sections of rocks from each unit are tabulated in Appendix 3.

The Bishop Corners and Myer Cave Formations, and the Lessard Formation in the low grade part of the study area, have been defined by Moore and Thompson (in press). In this Chapter, the Madoc and Stewart Formations are defined and the portion of the Lessard Formation in the high grade area is described.

Bishop Corners Formation

The Bishop Corners Formation, the basal unit of the Flinton Group in the Fernleigh, Bishop Corners, Flinton, and Black River synclines (Figure 1), includes pelitic schist, quartzite, quartzite metaconglomerate and micaceous quartzofeldspathic psammite. The formation has been mapped along strike for more than 100 km (Moore and Thompson, in press). Along the 40 km of the formation which occur in the study area (Figures 2 and 3), the thickness does not exceed 200 m. Although contacts are poorly exposed, the basal contact is sharp where it can be observed; whereas the upper contact with the Lessard Formation is commonly gradational within a zone 5 to 15 m thick. Within the Bishop Corners Formation, the predominant lithology grades from pelitic schist in the northeast to quartzite metaconglomerate and psammite in the southwest. Bishop Corners rocks are continuous from the area of lowest metamorphic grade to that of the high-

est grade. An increase in grain size and in the proportion of potassium feldspar and epidote, and a decrease in the proportion of mica in the matrix of the metaconglomerate are the only observable effects of metamorphism on the coarse clastic units of the formation. Relict clastic textures are not common but have been observed in quartzite pebbles, tourmaline and sphene. Amphibole occurs in the metaconglomerate in the high grade area, but only in the transition zone between the Bishop Corners Formation and overlying calcareous rocks. Pelitic schist, more sensitive to metamorphism, displays a variety of mineral assemblages (see Chapter 4).

Pelitic Schist (10).¹ The lowermost aluminous unit of the Bishop Corners Formation consists of a gray-weathering muscovite-rich schist, black-weathering schist lacking muscovite, and a rock transitional between these two. At Bishop Corners the basal schist is gray-weathering and composed essentially of muscovite, quartz, biotite, Fe-Ti oxides, garnet, chloritoid and chlorite, or chloritoid and staurolite. 20 - 100 m thick, the unit is continuous for 6 - 7 km in the study area. Tourmaline sometimes occurs as relict detrital grains surrounded by metamorphic overgrowths. Hornblende is found in the basal portion of the schist, coexisting with muscovite and chloritoid at several localities. Poikiloblasts of

¹ numbers refer to map units on Figure 2.

biotite, garnet, chloritoid, staurolite, and hornblende, not exceeding 10 mm, lie in a matrix of quartz, muscovite, and oxide with an average grain size less than 0.5 mm. The rock generally appears fresh; textural evidence of disequilibrium is not present. Quartzite pebbles (1 cm) occur near the basal contact of the gray schist, increasing in size and abundance in outcrops southwest and southeast of Bishop Corners.

Three km southwest of Bishop Corners, the aluminous schist is predominantly the black-weathering variety lacking muscovite. The black schist is much more feldspathic than the gray schist. The unit occurs at the contact between the Flinton Group and mafic metavolcanic rocks for 40 km, and does not exceed 20 m in thickness at any point. The essential minerals present are biotite, plagioclase (An_{20-25}), chlorite, garnet, oxide, and quartz. Tourmaline and apatite are accessories. Poikiloblastic hornblende and garnet, 5 to 20 mm parallel to the longest dimension, overgrow the fine-grained lepidoblastic biotite and equigranular feldspathic zones in the rock. The black schist commonly comprises alternating mafic and quartzofeldspathic laminae, 3 to 6 mm thick. Near the contact with metavolcanics, hornblende is an essential mineral, and there are discontinuous layers (probably relict bedding) of garnet-biotite amphibolite within the garnet-biotite schist. Staurolite and less commonly kyanite are present in the

hornblende-free portions of the schist. At several localities, the plagioclase in the black schist is completely altered to fine-grained aggregates of white mica which preserve the granoblastic texture of the unaltered feldspar. Tourmaline is commonly more abundant in these rocks than in the rest of the unit.

Near the contact with overlying quartzite and micaceous quartzite, the black schist generally contains more muscovite, quartz, and magnesian chlorite than is typical of the rest of the unit. The only known occurrences of the assemblage kyanite-chlorite in the study area are in this rock. Muscovite-bearing black schist represents a mineralogical composition transitional between the gray pelitic schist and the black pelitic schist.

Quartzite (11a). The distinctive gray- or white-weathering crossbedded quartzite in the Bishop Corners Formation is best exposed 2 km southeast of Bishop Corners (Figure 2, UTM 300605) and northeast of Flinton (Figure 2, UTM 232526). Quartz composes approximately 90 percent of the rock. Hematite, mica and feldspar, with accessory epidote and tourmaline, make up the other 10 percent. Thickness ranges from 10-30 m. The unit occurs as layers or lenses continuous, parallel to strike, for 2-10 km. Irregular pegmatitic bodies of quartz in the quartzite southeast of Bishop Corners (UTM 298603) contain kyanite. Generally, the quartzite lies between the basal pelitic schist and overlying

quartzite metaconglomerate. The contact with pelite is sharp, though locally the quartzite becomes a mica-ceous quartzite near the base of the unit. The upper contact is either gradational, with a gradual transition from quartzite to pebbly quartzite to meta-conglomerate, or the two rocks are interbedded with the proportion of quartzite beds decreasing upward. Well-preserved cross-bedding is a distinctive and useful feature of this unit (Plate 1a, Chapter 3). The cross-beds are the trough type. Depth ranges from 6-50 cm and length from 0.5-5 m parallel to compositional layering.

Quartzite Metaconglomerate (11). The quartzite-pebble metaconglomerate unit in the Bishop Corners Formation is what was referred to as the "Hastings Conglomerate" by earlier workers. Resistant to weathering and repeated by isoclinal folding, this unit appears to be more extensive than is actually the case. Along strike, the metaconglomerate is continuous from a point 2 km southeast of Bishop Corners to the southern closure of the Flinton syncline northeast of Actinolite (Figure 1). The rock is also present west of Actinolite (Figure 3). Maximum thickness is probably less than 200 m. Superficially, the rock is a homogeneous lithologic unit; in detail, the matrix mineralogy, proportion of matrix to pebbles, and pebble composition are variable. All gradations between pebbly schist, pebbly quartzite, sandy metaconglomerate and metaconglomerate occur. Bedding is

evident in places and lateral transitions are common. Cross bedding and scour-and-fill features provide top criteria. In areas of low strain the shape of pebbles suggests they were well-rounded or sub-rounded and ellipsoidal before deformation and metamorphism. Size is extremely variable (1-25 cm, long dimension) but most pebbles are in the range of 5-15 cm. The conglomerate is generally well-sorted with respect to the size and composition of pebbles. Most of the pebbles (50-67%) are gray or white quartzite. Bedding, cross bedding, and small faults are preserved in the pebbles despite recrystallization and deformation. In several cases, relict detrital textures were observed. The presence of minor amounts of feldspar, tourmaline and epidote indicate that the original material was more likely quartzite than chert. Massive quartz pebbles ("vein quartz") are the next most common type, composing up to 25 percent of the total pebble population. Black quartzite pebbles compose less than 10 percent of the clasts in the conglomerate. Granitic clasts are rare. Very fine-grained pink quartzofeldspathic pebbles (rhyolite?) and fine-grained dolomitic marble pebbles occur at some localities; for example, the outcrops in the middle of the roadcut on Highway 41 north of Northbrook. Although locally the matrix is quartzofeldspathic, it is mainly composed of muscovite, biotite and hematite, with lesser amounts of quartz, feldspar, carbonate, and epidote. Apatite and tourmaline are invari-

ably present as accessories. Over the entire conglomeratic unit, the proportion of matrix averages about 40 percent; however, the upper part of the unit adjacent to the biotite-carbonate schist of the Lessard Formation contains less than 20 percent matrix. Pebble composition in this conglomerate is almost entirely white quartzite, much more homogeneous than is typical of the unit as a whole.

Pebbles throughout the unit have been deformed during one or more of the phases of deformation that affected these rocks. The intensity of strain in pebbles of all compositions varies on a regional scale along the strike of the metaconglomerate. Within a hand specimen, the intensity of strain can be seen to relate to pebble composition; for example, quartzite pebbles are more severely deformed than vein-quartz pebbles. A study which attempted to measure the finite strain in these pebbles is included as Appendix 1.

Micaceous quartzofeldspathic psammite (12). East of Flinton the quartzite-quartzite conglomerate unit grades upward into a micaceous quartzofeldspathic psammite and conglomeratic psammite. The gradational transitions at the base of the unit, and at the top with the overlying calcareous rocks of the Lessard Formation, are sufficiently sharp (within a zone 0.5-5 m thick) for the psammite to be mapped as a separate unit. The relatively high proportion of feldspar distinguishes the rock from

those already described. For 15 km in the middle portion of the main Flinton belt, the quartzofeldspathic psammite occupies the crest of the central anticline and separates the quartzite metaconglomerate from the Lessard Formation in the adjacent syncline to the west (Figure 2). The thickness ranges from 20 to 100 m. Quartz, plagioclase, mica, and potassic feldspar are the essential minerals in the rock. Carbonate and oxides are also present and tourmaline and apatite are accessories. Characteristically, the rock weathers to form a series of discontinuous ridges and grooves 5 to 20 cm wide and several tens of meters long which probably reflect differences in feldspathic content in adjacent layers in the rock. Where the psammite is conglomeratic, the pebbles are the same composition as those in the quartzite metaconglomerate. Generally, the matrix in the conglomeratic quartzofeldspathic psammite weathers relative to the pebbles to a much greater degree than does the matrix in the quartzite conglomerate, allowing the two rocks to be distinguished.

Myer Cave Formation

Dolomitic marble, graphite- and sulphide-bearing schist, pelite and dolostone metaconglomerate overlying the Bishop Corners Formation in the Fernleigh syncline have been included in the Myer Cave Formation (Moore and Thompson, in press). Three outcrops of the

marble member of the formation (15) exposed near Bishop Corners are the only known occurrences of the Myer Cave Formation in the study area. The rock is a fine-grained, buff-weathering, massive dolomitic marble.

The Lessard Formation.

The Lessard Formation, overlying the Bishop Corners Formation in the Bishop Corners, Flinton, and Black River synclines, is composed of calcareous, micaeous quartzofeldspathic psammite, biotite-carbonate schist, epidote-bearing and epidote-free calc-silicate paragneiss (high grade metamorphic equivalents of the psammite and the schist respectively), and plutonic pebble metaconglomerate. The formation name is taken from the Lessard family farms (Figure 2, UTM 271518). There is a gradational change from schist and psammite to paragneiss along the strike of the isoclinal folds containing Flinton rocks. The formation is distinguished on the basis of high carbonate content and the relatively restricted occurrence of metaconglomerate. Along strike, the formation is exposed for 7.5 km in the Bishop Corners syncline and 15 km in the Flinton structure (Figure 2). The thickness ranges from 50 m to a maximum of 120 m. Compositional layering, (2-20 cm thick) outlined mainly by variable carbonate content, is present. Cross-bedding is rare in the psammite. Relict sedimentary textures were not observed. Locally the psammite,

schist, and their metamorphic equivalents are conglomeratic, containing quartzite pebbles and, less commonly, granitic pebbles. The relationship between the calcareous psammite and the biotite-carbonate schist is not clear. At one locality, the schist overlies the psammite, but generally the two rocks occupy the cores of adjacent synclines, with the intervening anticline composed of Bishop Corners Formation. This relationship also holds in the high grade area where the paragneiss units are separated by an anticline. A lateral interfingering or gradational relationship is likely. Although the stratigraphic position relative to the Bishop Corners Formation is the same and the carbonate content of both the Lessard and Myer Cave Formations is high, a lateral relationship between the two cannot be proven due to poor exposure in the critical northeastern portion of the area. Southwest, in the direction of increasing metamorphic grade, there is a gradational change from calcareous psammite and calcite biotite schist to epidote-bearing and epidote-free calc-silicate paragneiss. Stratigraphic relationships remain the same. The microcline isograd (Figure 14, Chapter 4) was chosen as the boundary between the schist and psammite, and the paragneisses.

Biotite-carbonate Schist (14). The biotite-carbonate schist occupies the core of the easternmost of the two synclines in the major Flinton structure for a dis-

tance of nine km, and a small doubly-plunging syncline near the western contact of the main Flinton belt (Figure 2). The thickness of the schist does not exceed 75 m. There is compositional variation within the unit with carbonate content increasing upward. The relatively siliceous basal zone commonly contains quartzite pebbles and granitic pebbles occur much more frequently than in the underlying Bishop Corners Formation. Biotite, ferroan calcite, ferroan dolomite, plagioclase, and quartz compose most of the rock. Accessory amounts of apatite, tourmaline, zircon and pyrite are present. Average grain size is generally about 0.5-1 mm. The large number of tiny zircons present, high carbonate and low opaque mineral content distinguish the schist from other micaceous rocks in the area. Actinolitic amphibole and microcline occur in the calcareous portion of the unit, increasing in abundance with increasing metamorphic grade. The distinctive brown weathered surface of the rock commonly reveals many small-scale (amplitude of several cm) isoclinal first phase folds.

Calcareous Quartzofeldspathic Psammite (13). Characteristically, differential weathering has produced a grooved and pitted, gray-brown surface on outcrops of calcareous psammite. The rock has a sandy texture with an average grain size less than 2 mm. Discontinuous compositional layering in the rock probably represents bedding. The psammite is exposed along strike for 15 km in

the westernmost of the two synclines in the major Flinton structure (Figure 2) with thickness ranging from 50 to 120 m. The core of the Bishop Corners syncline is also composed of calcareous psammite and the Black River syncline contains this rock as well. Quartzite pebbles and rarely granitic pebbles are present where the psammite is conglomeratic. Essential minerals include quartz, plagioclase, microcline, muscovite, biotite, ferroan calcite, and ferroan dolomite. Where dolomite is absent, epidote and actinolitic amphibole are present in small amounts; both become more abundant with increasing metamorphic grade. Accessory tourmaline, apatite, zircon, and Fe-Ti oxides are common.

Epidote-free Calc-silicate Paragneiss (14a). The high grade equivalent of the biotite-carbonate schist is a brown-green weathering, fine-grained paragneiss composed of diopside, actinolite, quartz, microcline, calcite, and sphene. The common accessory minerals are zircon, apatite, and pyrrhotite. Characteristically, discontinuous alternation of diopside-amphibole layers and quartz-feldspar-calcite layers is present. Amphibole is randomly oriented. Whereas the eastern contact of the unit is not exposed, the sharp and conformable western contact was observed at several localities. The basal siliceous zone present in the biotite-carbonate schist at lower grades is also associated with the paragneiss. This zone differs mineralogically from the

calc-silicate rock in the occurrence of epidote, abundant biotite, and hornblende rather than actinolite. Just south of the diopside isograd, a thin layer of tremolite-calcite marble separates the basal zone from diopside paragneiss. A small body of phlogopite-calcite marble exposed near the southeastern contact of the serpentinite (UTM 236418) is included in the epidote-free calc-silicate paragneiss unit of the Lessard Formation. The decreasing thickness of the paragneiss toward the southwest is more likely the result of structural geometry than a stratigraphic thinning.

Epidote-bearing Calc-silicate Paragneiss (13a). In the zone of highest metamorphic grade the stratigraphic equivalent of the calcareous psammite is a fine-grained paragneiss composed essentially of quartz, microcline, plagioclase, epidote, amphibole, diopside, and calcite. Sphene, apatite and oxide are also present. Occurrence of scapolite is sporadic; where present it is intergrown with plagioclase. Differential weathering and discontinuous compositional layering result in a pink and green grooved outcrop surface. Amphibole is randomly oriented in the rock. At some localities, the discontinuous layering is suggestive of a rock containing intensely deformed pebbles. The presence of pebbles was confirmed, however, only in the southwestern parts of the unit. The epidote-bearing paragneiss is exposed over a much larger area than the epidote-free paragneiss. The

epidote-bearing variety is continuous for 8.5 km from the microcline isograd (Figure 14 and Figure 2, UTM 233 441) south to the closure of the Flinton syncline. Thickness is difficult to estimate due to complex structural geometry but probably is less than 75 m.

Plutonic-pebble Metaconglomerate (13b). 1.6 km northeast of the southwest closure of the Flinton syncline, epidote-bearing paragneiss grades into plutonic-pebble metaconglomerate. Because of the similarity between discontinuous quartzofeldspathic layers in the paragneiss and the more intensely deformed clasts in the conglomerate, conglomerate was mapped only where deformed clasts could be clearly distinguished. A useful characteristic in this regard is the textural similarity between pebbles and the adjacent Northbrook granodiorite and Addington quartz monzonite plutons. Quartzofeldspathic zones in the paragneiss bear no resemblance to these rocks. Together with aplitic material, pebbles of granodiorite and quartz monzonite compose most of the pebble population. Vein-quartz pebbles are present but quartzite is rare. Near the contact with the Northbrook pluton, cobble-size clasts are composed entirely of biotite granodiorite. The conglomerate weathers dark green with pale gray or pink clasts. The matrix of the rock is fine-grained and essential minerals are hornblende, diopside, epidote, microcline, plagioclase, quartz, and calcite. The metaconglomerate

exposed north of Kaladar (Figure 1) (Walton, et al, 1964) and conglomerate in the Clare River synform (McKinstry, 1971) have similar matrix mineralogy and pebble composition.

There is a gradation, both laterally and upward, in the succession between the metaconglomerate and epidote-bearing calc-silicate paragneiss. The similar mineralogy of the epidote-bearing paragneiss and the matrix of the metaconglomerate suggest the possibility that the latter is a pebbly facies of the paragneiss unit. Stratigraphically, the plutonic-pebble metaconglomerate overlies directly both the Bishop Corners Formation and the Northbrook pluton. The plutonic-pebble metaconglomerate is included in the Lessard Formation because of the calc-silicate mineralogy of the conglomerate matrix and the fact that plutonic pebbles are more common in the calcareous rocks of the Lessard Formation than in the basal Bishop Corners Formation.

Associated with the metaconglomerate 4.5 km northeast of the locality described above is a rock composed of interlayered epidote paragneiss and biotite-quartz-feldspar. The epidote-bearing layers are similar in composition to the matrix of the metaconglomerate and the leucocratic layers are almost identical to the adjacent Northbrook granodiorite. The layered aspect of this rock and the association with conglomerate was the basis for including this rock in the metaconglomerate

unit.

The apparent thickening of the Lessard Formation near the southern closure of the Flinton syncline (Figure 2) results in part from the shallow dips of the layering in Flinton rocks caused by refolding of the syncline over the second phase Northbrook antiform.

Black River Area.

Quartzite metaconglomerate, pelitic schist, quartzite and psammite in the Black River syncline west of Actinolite (Figures 1 and 3) are very similar to the Flinton rocks described above. Cross-bedding indicates that these relatively homogeneous rock units overlie a wide variety of rocks including amphibolite, marble and metapyroclastic rocks. Wherever bedding can be discerned in the underlying rocks, there appears to be a conformable relationship between the two groups of rocks, or at most, a small angular discordance. Two rock types associated with metavolcanics and marble in the Black River area, a calcareous epidote-hornblende psammite or schist and a polymictic metaconglomerate with an amphibolitic matrix, are not observed elsewhere in the study area. The metaconglomerate occurs north and south of the main Black River structure. Clasts are dominantly marble and metavolcanic material. The calcareous schist and psammite separate the overlying quartzite metaconglomerate and pelite from the metavolcanic rocks in most

of the area. Within the metavolcanic sequence, an extensive unit of pink-weathering, very fine-grained metarhyolite occurs south of the Black River syncline. The origin of a pelitic schist closely associated with the metarhyolite is problematical. The schist is exposed south of the Black River syncline (Figure 3), separated from the basal quartzite metaconglomerate in the syncline by a strip of ground 5 to 150 m wide in which marble, calcareous phyllite, amphibolite and marble metaconglomerate are exposed. Mineralogical similarities between the schist and pelitic rocks in the Bishop Corners Formation, and the fact the schist is a homogeneous unit in contact with a variety of rock-types, form the basis for the decision to include the pelitic schist in the Flinton Group. However, the possibility the rock is the metamorphosed equivalent of a tuff associated with the nearby metarhyolite cannot be dismissed.

The Black River "syncline" is actually a complex structure composed of two or more tight isoclinal synclines and intervening anticlines. Reversals of way-up as defined by cross-bedding outline the major structures. Coexisting staurolite, biotite and garnet in the pelitic schist and rarity of amphibole in the calcareous psammite indicate the metamorphic grade to be the same as that near Flinton, intermediate between the highest and lowest grades in the region.

As was the case with the rocks in the Flinton

and Bishop Corners synclines, the younger metasedimentary rocks in the Black River structure can be grouped into a non-calcareous basal unit comprising pelitic schist (10)², quartzite (11a), quartzite metaconglomerate (11), micaceous quartzofeldspathic psammite (12), and an upper unit of calcareous quartzofeldspathic psammite (13). Although these rocks are similar to corresponding units in the Bishop Corners-Flinton area (Figure 2), there are small differences. The micaceous quartzofeldspathic psammite in the Black River area contains more carbonate, and where the psammite is conglomeratic, it contains more rhyolitic pebbles than does the quartzofeldspathic psammite in the Flinton area. Furthermore, in the basal unit quartzite is limited to discontinuous lenses in the quartzite-pebble metaconglomerate. The matrix of the conglomerate is somewhat less micaceous, and the average pebble size in the metaconglomerate is approximately 3 cm, about one-third that in the corresponding unit near Flinton (Figure 2).

Despite these differences, however, the rocks in the Black River syncline are essentially the same as those in the northeastern half of the Flinton map area (Figure 2). The same sequence, a non-calcareous basal unit overlain by a calcareous unit, occurs above the unconformity in both areas. On the basis of these similarities, the rocks of the Flinton Group present in the

² Numbers refer to map units in the Black River area (Figure 3).

Black River and Flinton map areas are considered to be parts of the same mappable units, the basal Bishop Corners Formation and overlying, calcareous Lessard Formation.

The basal unit of the Bishop Corners Formation (Black River area) is composed of quartzite metaconglomerate with minor quartzite and pelitic schist. The thickness of the unit, though somewhat variable, does not exceed 40 m. The lower contact is sharp, but the contact with the overlying quartzofeldspathic psammite is gradational over several m. The matrix of the conglomerate is made up of quartz, plagioclase, microcline, biotite, muscovite, and hematite. Quartzite pebbles, less intensely deformed than those in the Flinton area, are by far the most common pebble composition, with vein quartz the next most important. Quartzite, locally cross-bedded (trough type), occurs as discontinuous lenses in the metaconglomerate. Gray- or red-weathering fine-grained pelitic schist occupies the nose of a tight isoclinal first phase fold parasitic to the main Black River structure to the north (Figure 3). As a result, only the sharp lower contact of the unit is exposed. Biotite, staurolite, garnet, oxide, quartz, plagioclase and muscovite are the essential constituents of the schist. Poikiloblasts of staurolite, garnet and plagioclase locally attain maximum sizes several cm across.

The gray-weathering micaceous quartzofeldspa-

thic psammite overlying the metaconglomerate unit in the Black River syncline appears to be the predominant rock type in the structure. Repetition by isoclinal folding, and gradational contacts with adjacent rock units, obscure the true thickness of the unit, but it is probably less than 100 m. The psammite is locally conglomeratic with quartzite, "vein quartz" and meta-rhyolite the most common pebble lithologies. Microcline, plagioclase, quartz, mica, and minor amounts of carbonate are the principle minerals in the rock. The carbonate content is variable; all gradations from non-calcareous to calcareous psammite occur. Cross-bedding and minor first phase folds, common in the psammite, delineate the major structures of the complex syncline. Cross-beds are the trough type and are generally less than 10-20 cm deep and 1-2 m long (parallel to compositional layering).

The uppermost unit in the Black River sequence is composed of a calcareous quartzofeldspathic psammite sufficiently micaceous to appear schistose in outcrop. This rock unit has characteristics similar to those of both the calcareous psammite and the carbonate-biotite schist in the Lessard Formation in the Flinton area. Calcite, ankerite (locally), quartz, microcline, plagioclase, muscovite, biotite and epidote are the essential minerals in the rock. Tourmaline and apatite are common accessories. The calcareous psammite is a massive unit;

bedding was not observed. Restricted to a single narrow syncline near the northern edge of the Black River syncline (Figure 3), the distribution of the psammite is extremely limited in comparison to that of the calcareous psammite and its metamorphic equivalent in the Flinton syncline (Figure 2). Thickness is 5-20 m.

Madoc area

Madoc is located near the northern boundary of the Paleozoic cover which extends over much of southeastern Ontario (Figure 1). Outliers of flat-lying Paleozoic rocks are common in the region. Metamorphic rocks in the vicinity of Madoc (Figure 3) resemble those exposed northeast of Bishop Corners. A sequence of carbonate metaconglomerate, pelitic schist and graphitic calcite marble is underlain by metavolcanics, metagraywacke, and calcite and dolomite marble. The metavolcanic rocks are mainly andesitic, but rhyolitic material, both massive and with pyroclastic structures, is common. The marble has been metamorphosed along the contacts of rhyolitic dikes which predate the regional metamorphism. A small body of metamorphosed ignimbrite occurs north of Madoc (see Chapter 1). Once again, the basal unit of the younger metasedimentary rocks in the area is homogeneous and in contact with a variety of rocks - metaandesite, meta-rhyolite, dolomitic marble and metagraywacke - suggesting an unconformable relationship.

The major structure is the first phase Madoc syncline, outlined by the carbonate metaconglomerate and pelite units northwest of Madoc. The syncline is a more open style of fold (apical angle = 35°) than the first-phase structures at Bishop Corners (apical angle = $10-15^{\circ}$). The rocks near Madoc were not as intensely deformed as those already described in the central and northeastern parts of the map area; consequently, formations are thicker, and probably closer to the thickness of the original sedimentary unit. Metamorphic grade is comparable to that at Bishop Corners, with chloritoid and staurolite coexisting in the pelitic rocks. The metasedimentary rocks from the Madoc area assigned to the Finton Group have been divided into two formations: the basal Madoc Formation, and the overlying Stewart Formation.

Madoc Formation

The Madoc Formation, overlying metavolcanic rocks and marble north of Madoc (Figure 3), is composed of carbonate pebble metaconglomerate and fine-grained pelitic schist. The formation is exposed along strike for 2 km on the southeast limb and 2 km on the northwest limb of the Madoc syncline. Formation thickness is constant on each limb, but greater (300 m) on the northeast than on the southeast limb (190 m). The carbonate metaconglomerate pinches out toward the east on the south-

east limb. The lower contact of the Madoc Formation is sharp. The upper contact is not exposed; however, the pelitic rocks observed near (25 m) the inferred contact do not indicate a gradational transition into the marble units of the overlying formation. An extensive covered area obscures any extension of the Madoc Formation east of the Madoc syncline.

The basal unit of the Madoc Formation is a beige-weathering metaconglomerate (5a)³ composed of dolomite marble, quartzite and quartz clasts in a fine-grained dolomitic matrix containing some calcite. Minor amounts of muscovite appear in the matrix. Dolomite marble clasts, composing more than 80 percent of the pebble population, are the most intensely deformed (Z:Y:X = 3:2:1). Fragments of gray phyllite, mostly disc-shaped, appear in the conglomerate at several localities on the northeast limb of the syncline. The rock is poorly sorted, with clast size ranging from 2-3 mm to 5-10 cm parallel to the longest dimension. At one point where the contact with underlying meta-andesitic rock is exposed, the matrix at the base of the conglomerate contains amphibole, chlorite, and small lithic fragments similar in composition to the adjacent metavolcanic rocks.

Interbedded in the conglomerate is a gray-weathering, slightly calcareous phyllite (5b) 25 m

³ Numbers refer to map units Figure 3, Madoc area.

thick. Biotite porphyroblasts, oblique to a poorly developed muscovite schistosity, tend to weather out of this rock, resulting in a characteristic pitted appearance on the outcrop surface. Contacts between this unit and carbonate metaconglomerate are sharp.

The gray, fine-grained pelitic schist (6) overlying the metaconglomerate unit resembles the muscovite-bearing pelitic unit in the Bishop Corners Formation. Thickness ranges from 100 to 300 m. High Fe-Ti oxide content and coexisting chloritoid and staurolite are features common to both rocks. Lenses of fine-grained quartzite metaconglomerate, similar to those in Bishop Corners pelite, occur in the schist of Madoc (Figure 3, UTM 025330). Muscovite, quartz, oxide, and biotite or chlorite are the predominant minerals. Chloritoid schist is relatively restricted, occurring as discrete layers in the sequence.

Stewart Formation

Named for the Stewart family farm north of Madoc (Figure 3, UTM 024326), the Stewart Formation is composed of dolomite marble, graphitic calcite marble, and micaceous calcite marble. The dolomitic variety (7), a very fine-grained rock which locally is a recrystallized breccia, forms the basal unit of the formation. Distribution of the unit is not known because of poor exposure. The unit is at least 200 m thick. The rock is

sufficiently resistant to weathering to form a prominent ridge of outcrop parallel to the northwestern limb of the Madoc syncline.

Upward in the sequence, the next rock type is a gray-weathering calcareous marble (8), characterized by well-defined bedding on the scale of several cm and by abundant pyrite and graphite. Thickness could not be accurately established due to poor exposure, but is a minimum of 400 m. Black prismatic grains of scapolite occur in certain layers, commonly associated with dark gray or black calcite.

Although exposure is poor northeast of Madoc, the unit inferred to be youngest in the metamorphic rocks of the area is a beige- to gray-weathering micaceous marble characterized by thick bedding (30 cm to several m), and multicoloured carbonate breccia layers. An excellent example of the breccia and associated soft-sediment deformation features is exposed in a roadcut on the new Highway 7 bypass, north of Madoc (Figure 3, UTM 035316).

Type Sections

No single stratigraphic section in the study area contains all the rock units described above. A group of type localities, one for each rock unit, are defined which, if taken together, define a type section for each formation. The Myer Cave Formation is omitted because only three outcrops occur in the study area. The

unit is described more completely by Moore and Thompson (in press). Table 1 presents the UTM coordinates for the type localities. These points all occur within Zone 18 of the Universal Transverse Mercator grid system.

Table 1. Type localities for rock units of the Flinton Group.

Rock Units	UTM Coordinates of Type Locality, Figures 2 and 3
BISHOP CORNERS FORMATION	
- gray pelitic schist	275600
- black pelitic schist	260509
- quartzite	299602
- quartzite metaconglomerate	299602
- micaceous quartzofeldspathic psammite	299602
LESSARD FORMATION	
- biotite carbonate schist	273509
- calcareous quartzofeldspathic psammite	269509
- epidote-free calc-silicate paragneiss	241443
- epidote-bearing calc-silicate para- gneiss	234439
- plutonic pebble metaconglomerate	214371
MADOC FORMATION	
- carbonate metaconglomerate	023331
- calcareous phyllite	015318
- pelitic schist	025330
STEWART FORMATION	
- dolomitic marble	024326
- calcareous marble	026324
- micaceous marble	035316

Correlation

On the basis of detailed mapping in the area on which this thesis is based, and in the region northeast of Bishop Corners, (Moore, 1967; Sethuraman, 1970), Moore and Thompson (in press) defined the Flinton Group, part of the Grenville Supergroup. The Flinton Group includes clastic metasedimentary rocks and marble which overlie an angular unconformity. The unconformity provides a datum which makes possible correlation of an heterogeneous sequence of metasedimentary rocks with a complex structural and metamorphic history across a distance of more than 100 km. Stratigraphic units of the Flinton Group described in the preceding section have been correlated on the basis of lithologic characteristics and position within the stratigraphic sequence relative to the unconformity.

Cross-bedded quartzite and psammite in the Bishop Corners Formation, and the presence of pebbles in Flinton conglomerates similar in composition to underlying rocks indicate that the Flinton Group was deposited on top of the unconformity. Northeast of Bishop Corners, detailed mapping (Moore, 1967) revealed an angular unconformity between a sequence of metavolcanics, meta-graywacke and marble, and overlying metasedimentary rocks. Contacts in the rocks beneath the unconformity can be matched across the belt of overlying rocks. South of Flinton (Figure 2) the basal conglomerate of the Flinton

Group transects contacts between metamorphosed plutonic rocks and metavolcanic rocks at a high angle (30° - 60°). Further southwest, near the closure of the Flinton syncline (Figure 1) the fact the Flinton Group is in contact (Figure 2) with granodiorite, marble, amphibolite and metagraywacke suggests that an unconformity is present. North of Madoc, poorly defined bedding in the metavolcanic rocks trends oblique to the contact of the carbonate conglomerate at the base of the overlying meta-sedimentary rocks (Figure 3). Also, the basal unit once again is in contact with a variety of rock types.

A schematic representation of the space and time relationships between the formations of the Flinton Group in the study area is presented in Figure 4. Pre-metamorphic rock types have been inferred in the accompanying table of formations. For the rock units in areas of low grade, however, and for rock types not significantly affected by metamorphism (such as non-calcareous psammite and quartzite metaconglomerate), the inference is minimal.

Sedimentology

Lithologic variation takes place in the Flinton Group laterally and vertically in the sequence, and across the strike of the major structures in which these rocks are exposed. Figure 5 illustrates the regional variation along strike in the Flinton Group. Two major associations of rocks are evident: the pelitic schist-

Figure 5: Schematic representation of the variations laterally and vertically in the Flinton Group (solid pattern) in the study area. Thickness of the plutonic pebble conglomerate has been increased for purposes of illustration.

marble (shale-limestone) association at Madoc and Bishop Corners, and the metaconglomerate-psammite (conglomerate-sandstone) association predominant in the intervening area. Cross-bedding is commonly observed in the coarse-grained association, and intercalation of sandstone and conglomerate within rock units is common. In the northeast section of the map area, there is a gradational transition between these two associations of rocks. Near Bishop Corners in the formation of the same name, quartzite pebbles are limited to thin discontinuous lenses in the pelitic schist, or, in some localities, to a layer 1 or 2 m thick at the base of the pelitic unit. Southwest and southeast of Bishop Corners, the proportion of pebbles and micaceous quartzite in the formation increases, as does the average pebble size (from less than 1 cm to 15 or 20 cm parallel to the longest dimension). Where quartzite metaconglomerate and psammite compose most of the Bishop Corners Formation, pelitic material is present as matrix in the conglomerate and as a basal schist less than 25 m thick. A gradational change westward from the coarse-grained association to the fine-grained rocks at Madoc is not as well-documented, because of poor exposure. In the Black River area (section 4, Figure 4), however, the average size of pebbles in Flinton conglomerate is half the size of pebbles in the conglomeratic units exposed near the southern closure of the Flinton syncline and farther to

the northeast.

The gradational transition between the sandstone-conglomerate association and the shale-limestone association in the Flinton Group is interpreted as a facies change between coarse clastic sediments deposited in a high energy, fluvial-deltaic or beach environment, and fine-grained sediments deposited in a relatively low energy, shallow water environment.

There are significant changes in the Flinton Group upward in the sequence as well as laterally. Generally, the calcareous formations at the top of the sequence comprise finer-grained rock types than those at the base. Northeast of Bishop Corners, graphitic and pyritic schists, rocks with relatively low ferric-ferrous ratios, overlie marble and hematite-bearing pelitic schist (relatively high ferric-ferrous ratio). A similar relationship occurs at Madoc where the pyritic and graphitic marbles of the Stewart Formation overlie the oxide-rich pelitic schist in the Madoc Formation. The biotite-carbonate schist of the Lessard Formation and the laterally equivalent high-grade paragneiss contain a much lower proportion of opaque minerals than the underlying conglomeratic and psammitic rocks. Where present, the opaque mineral is a sulphide. The trend toward smaller grain size and more reduced rocks upward in the Flinton Group suggests the upper part of the sequence was deposited in deeper water than that in which the basal part was deposited. The high ferric-ferrous ratios

typical of the Bishop Corners Formation probably indicate a subaerial depositional environment, whereas rocks higher in the Flinton sequence are subaqueous deposits.

Corresponding with the variation along the strike of the Flinton Group in the map area (Figure 5) is a change in the rocks below the unconformity. The shale-limestone association at Bishop Corners and Madoc is underlain by andesitic and rhyolitic metavolcanic rocks, marble and metagraywacke (Figures 2 and 3). In the intervening area, mafic metavolcanics, dioritic and granodioritic plutons underlie the conglomerate-sandstone association. On a smaller scale, there is a correspondence between pebble composition in the lower part of Flinton conglomerates and rocks beneath the unconformity, particularly along the southeast contact of the Flinton Group. At the contact near the southern closure of the Flinton syncline, metaconglomerate contains pebbles of granodiorite and quartz monzonite similar to the Addington and Northbrook plutons located to the southeast. The feldspathic psammites in the Bishop Corners and Lesard Formations are spatially related to the large plutons (Figures 2 and 5). The carbonate metaconglomerate exposed east of Bishop Corners is closely associated with a large body of pre-Flinton dolomitic marble (Figure 7, Moore and Thompson, in press). Rhyolitic pebbles in quartzite metaconglomerate south of Bishop Corners

and in the Black River area are similar to metarhyolite in the underlying metavolcanic rocks. In these examples, the probable source rock for clasts in metaconglomerate units is either adjacent to the southeast contact of the Flinton Group or within a few km to the southeast. Near Bishop Corners, increase in pebble size indicates that the source area of the most common pebble type, quartzite, is also southeast across the sequence. In the Black River area (Figure 3), the basal quartzite-quartzite conglomerate unit thickens to the southeast. However, near Flinton (Figure 2) pebble size increases toward the northwest across the synclines containing the quartzite conglomerates and, along the length of the Flinton syncline, there is a general tendency for the basal Bishop Corners Formation to become thinner toward the southeast. At several localities (for example, Figure 2, UTM 230440-250440) the formation thins toward the Northbrook pluton (southeast) and toward the Elzevir pluton (northwest). These large masses of plutonic rock apparently controlled to some extent the deposition of the Flinton Group in the study area. Southwesterly and/or northeasterly transport of sediment may have prevailed in the area between the two plutons (Figures 1 and 2).

Although much of the material composing the Flinton sedimentary rocks was probably derived locally, the nearest mappable quartzite which could be the prov-

enance area for the quartzite pebbles occurs 60 km southeast of the study area. The distance is not unreasonable for a fluvial transport system considering the lack of vegetation at the time of deposition more than 1000 million years ago. Periodic flooding similar to that occurring in present-day environments was very likely an important process during deposition of the quartzite conglomerate in the Flinton Group. However, the quartzite layers commonly present in the pre-Flinton marbles in the study area were probably the source of some of the quartzite pebbles in the Flinton Group. The marble may have been more extensive prior to the unroofing of the plutonic rocks.

The sedimentological characteristics of the Flinton Group can be summarized as:

- 1) a major angular unconformity at the base of the sequence;
- 2) two lithologic facies: a conglomerate-sandstone association and a shale-limestone-dolomite association;
- 3) continuity of relatively thin (10-100 m) rock units for tens of kilometers;
- 4) a relatively highly oxidized basal formation (red beds);
- 5) decrease in grain size and ferric-ferrous ratio upward in the sequence;
- 6) compositional maturity;

- 7) both a local and a distant (60 km) provenance area.

Time of Deposition

Some of the pebbles found in the Flinton Group are texturally and mineralogically similar to the adjacent Northbrook-Elzevir granodiorite and Addington quartz monzonite plutons. Lumbers (1964, Figure 7) included the granodiorite in a "sodic granite group" and the quartz monzonite in a "potassic granite group". Silver and Lumbers (1965) concluded the apparent ages of the two associations were 1250 ± 25 m.y. and 1125 ± 25 m.y., respectively. The ages were obtained from a preliminary study of zircons from various plutonic bodies in a large area near the southwestern border of the Grenville Province which includes the present study area. Pegmatites similar to those which cut the Flinton Group in the study area were dated at 1050 ± 20 m.y. by the same authors. Krogh and Hurley (1968) quote an apparent age of 1035 ± 60 m.y. as the minimum time of formation of the Addington pluton north of Kaladar (Figure 1). The age may represent a major rehomogenization of the Rb-Sr isotopes in the rock related to regional metamorphism. If the plutonic pebbles in the Flinton Group are derived from the Elzevir-Northbrook and Addington plutons, as they appear to be, and if the apparent ages obtained by Silver and Lumbers (1965) apply to these rocks,

the time of deposition of the Flinton Group would postdate the intrusion of the Elzevir-Northbrook and Addington plutons (1250 ± 25 m.y. and 1125 ± 25 m.y., respectively) and predate the regional metamorphism and emplacement of pegmatites (1050 ± 20 m.y.). A maximum time of deposition for the Flinton Group is indicated by isotopic data obtained from zircons in quartzite metaconglomerate in the Bishop Corners Formation (Black River area) by Boutcher, Davis, and Moorhouse (1965). The apparent ages from $\text{Pb}^{206}/\text{U}^{238}$, $\text{Pb}^{207}/\text{U}^{235}$, and $\text{Pb}^{207}/\text{Pb}^{206}$ ratios are 1060 ± 32 m.y., 1090 ± 33 m.y., and 1140 ± 34 m.y. respectively (average 1096 ± 33 m.y.). These numbers support the contention that deposition postdates the intrusion of the Addington quartz monzonite. In fact, the zircons may have been derived from that body. The unconformity at the base of the Flinton Group probably developed during a relatively long period of erosion (25 m.y.?). The presence of kyanite indicates these rocks were deeply buried (>15 km) at the time of metamorphism. Deposition of the Flinton Group very likely occurred between 1075 and 1100 m.y. ago.

Conclusions

In the Bishop Corners-Madoc region, metamorphosed sedimentary rocks of the Flinton Group comprise a shale-limestone-dolostone facies and a sandstone-conglomerate facies. The transition between them is interpreted

ted as a change from a low energy shallow water depositional environment to a relatively high energy fluvial-deltaic or beach environment. Except for quartzite pebbles which probably had, for the most part, a more distant source, the provenance of the Flinton Group was local. Sediment transport directions were from the southeast and possibly from the northeast or southwest. The plutonic bodies created local variations in these trends. The Flinton Group is believed to be a transgressive sequence deposited on a relatively deeply eroded erosion surface between 1075 and 1100 m.y. ago.

This work has demonstrated that the basic principles of stratigraphy and sedimentology can be applied to Grenville rocks despite regional metamorphism and several periods of deformation, an approach that is essential to reconstructing the complex geological history of these rocks.

III STRUCTURAL GEOMETRY

Introduction

In the Bishop Corners-Madoc area (Figure 1) the structural geometry of the metasedimentary rocks of the Flinton Group can for the most part be explained by three major phases of deformation. In addition, a gently-dipping crenulation, locally developed in micaceous rocks, is present. The latest event is cataclasis which affected rocks near the contact of a body of metamorphosed ultramafic rock which cuts the Flinton Group.

The study area overlaps the area east of Madoc mapped by Bain (1960) and the northwest corner of the area covered by G. Venkitasubramanyam (1969) in a structural study of the Clare River synform. Bain interpreted the outcrop pattern as the result of extensive faulting. Venkitasubramanyam demonstrated that there were two phases of deformation in the region and that the Clare River synform and Northbrook antiform were second phase structures. J.M. Moore (personal communication) and K. Sethuraman (1970), working east of Bishop Corners, recognized three phases of deformation, the second corresponding to Venkitasubramanyam's first phase. The purpose of this chapter is to describe the structural geometry of the Flinton Group in the study area, and from the geo-

metry interpret the structural history of these rocks and the timing of deformation with respect to regional metamorphism.

Structural History

The structural geometry presented in this chapter is based on data collected mainly from rocks of the Flinton Group in the Flinton, Black River, and Madoc map areas (Figure 1). Somewhat more sparsely distributed data were collected from pre-Flinton rocks. Six deformational events (Table 2) can be recognized in the field. A pre-Flinton foliation is moderately well-developed in the Elzevir pluton and metavolcanic rocks west of the main belt of metasedimentary rocks south of the town of Flinton (Figure 1). Structures associated with D_1 and D_2 , the first and second phases of deformation affecting the Flinton Group, can be observed on scales ranging from microscopic to the regional scale of Figure 1 everywhere but in the Madoc area. At this locality, although D_1 structures are prominent, evidence for D_2 is absent. D_3 folds occur on a regional scale as well, but small-scale structures are common only in the Black River area. D_3 features are sparsely developed near Madoc, and in the Flinton area are restricted to localities where the trend of S_1 changes greatly. Post- D_3 structures have a more limited distribution. The

Table 2 - Structural History in the Bishop Corners-Madoc area

Deforma- tion Event	Observed Elements	Fold Geometry apical angle	strike and dip/plunge of axial surface/hinge-lines	Fold Type ¹	Elonga- tion Lineation	Metamorphic Grade
myloniti- zation	planar fabric		025° - 035°, steep SE			
D ₄	S ₄ axial surf. L ₄ hinge-lines L ₄ of crenula- tion	30° 120°	variable, shallow SE or NW	variable, flexural slip?		post date metamorphic recrystalliza- tion
D ₃	S ₃ axial surf., L ₃ cleavage L ₃ fold hinge- lines	40°- 60°	100° - 120°, steep SE or NW	70°-90°, NW or SE flexural slip -flexural flow		greenschist? lower amphibolite?
D ₂	S ₂ cleavage, axial surf., L ₂ fold hinge- lines S ₂ /S ₁ intersec- tion lineation	20°- 60°	030° - 045° steep SE or NW	10°-90° NW and SW flexural flow		middle amphibolite
D ₁	S ₁ cleavage, fol- iation, axial surfaces L ₁ elongation lineation, fold hinge lines S ₁ /S ₀ intersec- tion of S ₁ with layering	10°- 35°	000° - 085° steep, direc- tion variable	0 - 90° direction variable flexural flow variable, NE to SW		greenschist to middle amphib- olite
pre-Flin- ton folia- tion	S _x mineral len- ticles cleavages?		050°-085° steep NW			greenschist?

¹after Donath and Parker, (1964)

crenulation (D_4) is present in schistose rocks everywhere along the belt. In the southwestern third of the Flinton map area (Figure 2) intensely deformed quartz grains exhibiting undulose extinction and deformation bands occur in the metasedimentary rocks. Mylonite in epidote-bearing paragneiss near the western contact of the lenticular ultramafic body may have developed at the time of emplacement of the serpentinite. The preservation of the strained quartz textures indicates that these features developed late in the structural history after metamorphic recrystallization ceased.

Structural Geometry

Pre-Flinton Foliation. In the metamorphosed Elzevir pluton and metavolcanic rocks west of the Flinton syncline the older rocks contain a foliation (S_x) which is cut by the dominant foliation in the region (S_1). The older foliation swings around slightly to a more easterly trend ($070^\circ - 085^\circ$) and disappears altogether toward the central part of the pluton (Figure 6). Approaching the western contact of the Flinton Group, S_x is increasingly difficult to measure as S_1 becomes better developed. At the contact of the pluton with Flinton rocks S_x is completely transposed by S_1 . At one outcrop (Figure 2, UTM 232483) of fragmental metavolcanic rocks, a dike similar in composition to the Elzevir diorite (unit 7, Figure 2) cuts a foliation defined by deformed

fragments. The Flinton Group unconformably overlies the diorite. S_x was probably developed before deposition of the Flinton Group during the intrusion of the plutonic rocks into the volcanic rocks.

D_1 Structures. The Flinton Group is exposed in a series of tight to isoclinal synclines and intervening anticlines throughout the study area (Figures 6 and 7). Detailed mapping of the stratigraphy of the Flinton Group was the most important means of identifying these folds. Cross-bedding (Plate 1a) provided reliable top determinations at several localities. These structures can be traced along strike for 5 - 20 kilometers, but the distance between the axial surface traces of adjacent synclines is generally less than 250 meters. To reduce the large number of lines on the trend maps, axial surface traces of anticlines were omitted. Only bedding (S_0) can be traced around the closures of these structures and the dominant foliation (S_1) in the region is parallel to the axial surfaces, indicating that these are first phase (D_1) folds. The attitude of the hinge-lines of the map-scale folds can be inferred from S_0/S_1 intersections and from minor first phase folds; both the direction and angle of plunge of major fold hinge-lines are variable. For example, the Bishop Corners syncline (Figure 6) plunges southwest at Bishop Corners and northeast near Flinton. East of Flinton, in the main belt of meta-sedimentary rocks, there is a doubly-plunging syncline

Plate 1

- a) Cross-bedding in quartzite (Bishop Corners Formation).
- b) S_1 defined by deformed pebbles in quartzite metaconglomerate (Bishop Corners Formation).
- c) D_1 minor fold in calcareous quartzofeldspathic psammitic (Lessard Formation).

a

b

c

PLATE 1

a

b

c

PLATE 1

with a core of biotite carbonate schist (Figure 6). First phase minor folds generally deform only compositional layering. At some localities, however, in the core zone of major D_1 folds, a small proportion of the pebbles in metaconglomerate containing very little matrix is folded. The axial surfaces of these folds are subparallel to the S_1 fabric defined by the unfolded pebbles in the rest of the rock. Angles of plunge of D_1 minor folds range from zero to ninety degrees. Generally, the folds have smaller apical angles and plunge more steeply ($60^\circ - 70^\circ$) in schistose rocks than in quartzofeldspathic units ($20^\circ - 45^\circ$). Apical angles of $10 - 20$ degrees are common in the schist, apical angles of $30 - 60$ degrees in the psammite (Plate 1c). On a larger scale, the major first folds in the Flinton and Black River areas (Figures 6 and 7) have smaller apical angles ($10^\circ - 15^\circ$) than the more open Madoc syncline (35°).

Structural domains were defined to illustrate the variation in orientation of structural elements from area to area along the trend of the Flinton Group. The dominant foliation (S_1) in all the rocks in the area is parallel to the axial traces of major D_1 folds. S_1 is defined by micas, amphibole, and deformed fragments in the metavolcanic rocks; mineral lenticles in the plutonic rocks; and mica or deformed pebbles or both in meta-sedimentary rocks (Plate 1b). The attitude of S_1 , generally steeply-dipping, is reasonably constant within

each of the five structural domains defined in Figures 6 and 7. The strike of S_1 varies from 085° to 175° within the map area. The stereonet diagrams on the D_1 trend maps illustrate the distribution of poles to S_1 , S_1/S_0 intersections and hinge-lines (L_1) of D_1 minor folds in each domain. Note that the linear elements on the diagrams are rather variable within the plane of S_1 in Domains I - III but more tightly clustered in Domains IV and V. When these diagrams are combined to produce one stereoplot for the entire area, a greater degree of variation for both poles to S_1 and the linear elements is apparent (Figure 8). The effect of D_2 on S_1 is most evident in the hinge-zone of the Northbrook antiform. Although the relationship is not clearly defined it is possible to fit a small circle to the points representing S_0/S_1 intersections and D_1 hinge-lines. The axis of the circle plunges gently ($5 - 10^\circ$) toward 110° . A better fit can be obtained about a steeply-dipping plane (great circle) with a strike of about 028° ; that is, the average azimuth for S_1 in the map area. Even though there is a possibility of relating some of the variability in attitude of L_1 to a later phase of folding trending north-northwest, the stereoplot data are not convincing and variation may well be due to heterogeneous strain during D_1 . The variable trend of S_1 is attributed mainly to D_2 and D_3 , but the Elzevir pluton was probably a rigid mass during D_1 , causing S_1 to develop a variable

Figure 8: Poles to S_1 (•) and L_1 (•) (D_1 minor fold hinge-lines and S_0/S_1 intersections) for all 5 domains. Poles to S_1 inside area outlined by the fine line occur in the hinge-zone of the Northbrook antiform.

attitude at that time in some parts of the area.

Figure 9 illustrates the variation in orientation of the elongation lineation (L_1) in the study area. The lineation is defined by long axes of deformed pebbles, hornblende, and epidote which are parallel to S_1 , the regional foliation. At several localities the hornblende lineation is folded by D_2 minor folds, indicating that this structure was developed during D_1 . On Figure 9 only the epidote lineations and one gently-plunging hornblende lineation are from Domain III. The influence of D_2 is so great in this domain, particularly near the closure of the Northbrook antiform, that D_1 mineral lineations are difficult to distinguish. Pebbles in the metaconglomerate are intensely flattened in Domain III to a pancake shape and do not define a preferred orientation. The elongation direction is parallel to the hinge-lines of first phase folds in Domain IV (Black River area), but over the rest of the study area, the elongation direction commonly plunges more steeply than the folds. From a comparison of Figures 8 and 9, it is clear the orientation of the D_1 elongation lineation is less variable along the belt than the orientation of S_1/S_0 intersections and hinge-lines of minor folds. This relationship together with the variability of hinge-lines in different rock types indicates most of the spread of D_1 linear structures (Figure 8) results from heterogeneous strain during D_1 .

Figure 9: D_1 elongation lineations in all 5 domains;
pebbles, •; hornblende, x; epidote, ▲

Within Domains II and III, westward from the Flinton rocks toward the core of the Elzevir pluton, the trend of S_1 becomes more easterly (050°) and the intensity of development of foliation decreases in the plutonic rocks. The more easterly orientation may represent the attitude of S_1 before the second and third phases of deformation, preserved within the relatively rigid mass of the Elzevir plutonic rocks.

D_2 Structures. From Figure 1 it is evident that large-scale structures such as the Clare River synform and Northbrook antiform, with northeast trending axial surfaces, define the northeast-southwest structural grain of this part of the Grenville Province. As a result of the work of Venkitasubramanyam (1969) on the Clare River structure and the writer's observations in the vicinity of the closure of the Northbrook antiform, D_1 folds are known to be refolded by these structures. The D_2 - D_3 trend maps (Figures 10 and 11) illustrate the evidence for a second phase of deformation in the Bishop Corners-Madoc area. Though difficult to detect in places, a second cleavage (S_2) is present in micaceous rocks such as pelitic schist and biotite amphibolite. Locally, the intersection of S_1 and S_2 produces a mica edge lineation on S_1 . Moderately open (apical angle = 30° - 60°) small-scale asymmetrical folds (Plate 2a) deforming S_1 are common in certain parts of the area. The axial surfaces

Plate 2

- a) D_2 minor folds deforming S_1 where it is defined by flattened pebbles.
- b) S_1/S_2 intersection lineation in the hinge-zone of the Northbrook antiform.
- c) Crenulation of S_1 in the hinge-zone of the Northbrook antiform. S_1 is defined by mineral lenticles in granodiorite of the Northbrook pluton.

PLATE 2

a

b

c

PLATE 2

a

b

c

dip steeply northwest or southeast, striking 035° - 045° . Stereonet diagrams on Figures 10 and 11 indicate the orientation of poles to S_2 and of D_2 minor fold hinge-lines (L_2) in Domains I - IV. D_2 structures were not observed in Domain V, the Madoc area. The fact that S_1 is very close to the general trend of S_2 in this area is the most likely explanation for the absence of evidence for D_2 .

Figure 12 is a stereonet diagram which includes the D_2 data for the entire map area. The spread of poles to S_2 on this diagram is less than that for poles to S_1 (Figure 8). The D_2 hinge-lines fit a great circle with a trend of 038° , the average azimuth of S_2 for the area. The spread of data is probably due to variable amounts of strain during D_2 and a range of attitudes for S_1 and S_0 before D_2 .

Where the strike of S_1 is appropriate, the development of minor folds which deform S_1 is the most convincing evidence for D_2 in the Flinton and Black River areas. The plunge of these folds and their form viewed down the plunge (S, Z, or M) depends on the angle between the strike of S_1 and the strike of S_2 , and the amount and direction of dip of the earlier schistosity. This dependence is demonstrated by the variable attitude of D_2 hinge-lines in Figure 12 and on the trend maps (Figures 10 and 11). At the closure of the southwest-plunging Northbrook antiform (Figure 10) there is a change from

Figure 12: Poles to S_2 ($\cdot$) and L_2 ($\bullet$) (D_2 minor fold hinge-lines) for all 5 domains.

S- to M- to Z-shaped folds across the closure. The cluster of Z folds northwest of the closure is related to the interference pattern caused by superposition of D_1 and D_2 folds in this area. All these folds plunge to the southwest, as does a distinctive (Plate 2b) lineation produced by the intersection of S_1 and S_2 . The crenulation of S_1 in the hinge-zone of the antiform is extremely well-developed (Plate 2c). Another D_2 fold closure is indicated by a change of vergence in the western part of the Black River area (Domain IV) but the intersection lineation and crenulation are not present. The position of the axial surface representing the closure (Figure 11, UTM 120340) is based on the location and vergence of minor D_2 folds in the area. The lack of outcrop to the southeast of the inferred closure precludes accurate positioning of the axial surface trace.

Northeast of the closure of the Northbrook antiform, D_2 minor folds are S-shaped (Figure 10). Just north and northwest of the closure the minor folds are Z-shaped. In between these two areas there does not appear to be a major D_2 closure. The trend of S_1 , however, changes from an orientation more northerly (030°) than S_2 (045°) to a trend more southerly (065°). This change is sufficient to account for the change in vergence of the D_2 minor folds. It is probable the variation in the trend of S_1 reflects a pre- D_2 orientation related to the adjacent Elzevir batholith.

D₃ Structures. Major open folds or warps deform the axial surface traces of the Clare River synform and Northbrook antiform (Figure 1). The trace of the axial surface of these open folds trends approximately north-northwest. In the Bishop Corners - Madoc area, mesoscopic evidence for a third phase of deformation (D₃) is more sparsely distributed than that for D₁ and D₂. Both a weakly developed cleavage (S₃) (in pelitic schist) and third phase minor folds occur in the Black River area (Domain IV, Figure 11). D₃ minor folds deform S₁ and are distinguished on the basis of the trend of their axial surface (090° - 110°) from D₂ minor folds (035° - 045°). Also, S₂ is deformed by D₃.

Axial surface traces of minor folds in the marble at Madoc have a similar orientation to those in the Black River area. In Domains I, II, and III, evidence for D₃ is limited to areas where there is a major change in the trend of S₁; that is, at either end of the Flinton area (Figure 10).

Evidence for the superposition of D₃ on D₁ structures is present at Madoc in the form of interference patterns (Plate 3a) in the marble in the core of the D₁ syncline. In salic metavolcanic rocks (Unit 4, Figure 9) southeast of the main D₁ structure in the Black River area, a D₂ Z-shaped fold is deformed by a D₃ S-shaped fold.

Plate 3

- a) Interference patterns in marble (Stewart Formation) near Madoc (note mechanical pencil for scale).
- b) Mylonitized epidote-bearing calc-silicate paragneiss (Lessard Formation). Scale bar represents 0.5 mm (nicols uncrossed).
- c) Mylonitized epidote-bearing calc-silicate paragneiss (Lessard Formation). Scale bar represents 0.5 mm (nicols crossed).
- d) Internal fabric preserved in a staurolite porphyroblast is oblique to S_1 defined by mica and Fe-Ti oxide in the matrix of the rock. Scale bar represents 0.2 mm.
- e) Garnet porphyroblast preserves the fabric of the matrix composed of chlorite, muscovite, opaques, and margarite in pelitic schist from Zone 1 (see Chapter 4). Scale bar represents 0.5 mm.
- f) Garnet porphyroblast preserves outline and small opaque inclusions of a biotite porphyroblast in pelitic schist from Zone 1. Scale bar represents 0.5 mm.

PLATE 3

a

d

b

e

c

f

PLATE 3

a

d

b

e

c

f

The data for the third phase of deformation in all five structural domains is presented in Figure 13. The generalized attitude of S_3 taken from the stereonet data corresponds to the axial surface trace of the large scale warps in the trend of S_2 and S_3 .

The outcrop pattern of the Flinton Group and associated rocks can be explained by a structural geometry based on these three phases of deformation - D_1 , D_2 , and D_3 .

D_4 Structures. A distinctive gently-dipping crenulation or crinkle which occurs in micaceous rocks of the Flinton Group is thought to have developed after the third phase of deformation. Metamorphic grade was high enough during D_3 for micas to grow parallel to S_3 in the Black River area. However, at the time of D_4 , although some of the mica plates bent in the hinge zones of the crenulation recrystallized, most of the mica and chloritoid porphyroblasts in the hinge zones exhibit undulatory extinction. The crenulation is present almost everywhere schistose rocks occur in the map area. At several localities (e.g. UTM 234406, Figure 2), small-scale folds (amplitude 1-2 cm) with axial surfaces dipping 5 - 35 degrees southeast occur. These structures are attributed to D_4 as well.

Mylonitization. A north-northeast trending mylonite zone is found in the epidote-bearing paragneiss (Figure 2, UTM 229409) near the western contact of a

Figure 13: Poles to S_3 ($\cdot$) and L_3 ($\bullet$) (D_3 minor fold hinge-lines) for all 5 domains.

body of metamorphosed ultramafic rocks (Unit 16) which cuts the Flinton Group. The mylonitized paragneiss (Plate 3b,c) contains severely strained quartz and microcline, broken diopside grains, and thin anastomosing seams of very fine-grained material which suggest intense granulation has taken place. Data collected independently of this study from an ultramafic body (Figure 3, UTM 330760) located 6 km southwest of the one discussed here indicate that these rocks are associated with pre-Flinton metavolcanic rocks of the Tudor Formation (J.M. Moore, personal communication). The mylonite may have developed during the tectonic emplacement of the serpentine anthophyllite body into its present position. In any case, the mylonite must have been formed at a late stage in the thermal history of these rocks because intensely strained aggregates of quartz and feldspar in the rock have not recrystallized. Strained quartz is common in Flinton rocks in the southern part of the Flinton area, but only one occurrence of mylonitized rock has been found.

Timing of Metamorphism and Deformation

Petrographic work by J.M. Moore (personal communication), A.W. Hounslow (1967), K. Sethuraman (1970), and the writer (Chapter 4) has revealed only one period of regional metamorphism affecting the meta-sedimentary rocks of the Flinton Group. In the Bishop

Corners-Madoc area, the minerals representing the highest metamorphic grade in an individual thin section (for example, kyanite, staurolite, chloritoid, garnet, hornblende and epidote) always overgrow a pre-existing foliation defined by micas, tabular opaque minerals, and the long axes of quartz grains. In some samples, S_1 is deformed around garnets and hornblende, and also included within these minerals. Most commonly, however, porphyroblastic growth occurred after formation of S_1 was complete.

Garnet and chloritoid in pelitic schist from the Flinton area and staurolite from the Black River area in some cases contain an internal fabric which is at a high angle ($60^\circ - 90^\circ$) to S_1 in the surrounding matrix (Plate 3d). The relict fabric may have developed during a period of deformation postdating deposition of the Flinton Group but occurring before D_1 , and for which no other evidence is preserved. Although a more detailed petrofabric study in Flinton rocks might reveal more evidence to support this possibility, at this stage a more acceptable explanation is that the relict fabric represents an early stage in the development of S_1 .

Biotite and hornblende defining S_1 in granodiorite and amphibolite respectively, were intensely folded in the hinge zone of the Northbrook antiform during D_2 . Even though complete recrystallization of these minerals to polygonal arcs occurred, no new material

grew parallel to S_2 . The same is true of D_2 minor folds present in the epidote-bearing and epidote-free calc-silicate paragneisses. Pelitic rocks from the Flinton and Black River areas contain a cleavage, S_2 , defined by the small proportion of mica and chlorite in the rock which have grown during D_2 . No porphyroblastic growth is related to the orientation of S_2 in these rocks. In addition, it is not clear whether or not porphyroblasts present in the rock before D_2 continued to grow during this phase of deformation. In rocks where garnet has overgrown everything else in the thin section (Plate 3e,f) there is a possibility S_2 was included but no conclusive evidence was found. Although new mineral growth was not extensive during D_2 , one line of evidence suggests that temperatures were probably high at the time in at least the central part of the area. Pegmatites considered to be the product of partial melting of rocks of the Flinton Group (Chapter 1) cut axial surfaces of D_2 minor folds in these rocks. Contact effects are not observed in country rock adjacent to pegmatite. Isograds drawn on a regional map (Figure 14, Chapter 4) are not folded about D_2 axial traces; in fact, they seem to cross the northeast trend of S_2 at Madoc and Bishop Corners. In the area between Flinton and Actinolite the isograds strike approximately parallel to S_2 . The timing of metamorphism with respect to D_2 is not clearly established. Metamorphic grade attained a maximum after S_1 had devel-

oped and was declining at the time of D_3 (see next paragraph). Because new mineral growth was not extensive during or after D_2 , the maximum is interpreted to have occurred shortly before D_2 . The metamorphic grade remained high during this phase of deformation. Time could be a factor in the lack of new mineral growth during D_2 ; possibly the rocks were not hot long enough for new material to crystallize.

Metamorphic grade was high enough during D_3 for some micas to grow parallel to S_3 in pelitic schist in the Black River area, but porphyroblasts have not overgrown this trend. Micas folded by D_3 have recrystallized, but the limited distribution of evidence for D_3 suggests metamorphic grade was lower than that during D_2 . The fact chloritoid porphyroblasts and most mica folded in the hinge zones of the D_4 crenulation exhibit undulatory extinction indicates D_4 occurred near the time when metamorphic recrystallization had essentially ceased.

Textural evidence suggests metamorphic grade in the Bishop Corners-Madoc area was increasing during D_1 , attained a maximum shortly before D_2 , remained high for some time, and, before D_3 , began to decrease. From the geochronological data presented in Chapter 2, it is apparent that three major phases of deformation (D_1 , D_2 , D_3), including the development of a northeasterly trend typical of the Grenville Province, and the accompanying regional metamorphism, occurred between the time of

deposition of the Flinton Group, (1075 - 1100 m.y. ago), and 950 - 1000 m.y. ago, the time at which the system became closed with respect to radiogenic argon - a relatively short period near the end of the history of the Grenville province. The question as to whether the period of metamorphism and deformation outlined here represents the entire Grenville "orogeny" or only the last stage in a more complex orogenic history remains to be answered.

IV METAMORPHIC PETROLOGY

Introduction

Regional metamorphism in the Bishop Corners-Madoc area of the Grenville province (Figure 1) resulted in mineral zonation in metasedimentary rocks of the Flinton Group. Textural evidence indicates that mineral assemblages from each zone crystallized under conditions closely approaching chemical equilibrium. Detailed microprobe analyses of all the minerals in six samples support, for the most part, the contention that equilibrium was attained; however, in some samples plagioclase, garnet, or amphibole are heterogeneous. Compositional criteria provide a more precise evaluation of the degree to which chemical equilibrium was attained in the rocks.

In the study area, two major sedimentary facies changes occur in the Flinton Group (Chapter 2). The bulk composition of individual rock units probably changes significantly along strike as well. Changes in bulk composition must therefore be considered as important variables, in addition to pressure and temperature in the development of mineral zonation in these rocks.

Conventionally, the importance of compositional variation in determining mineralogical variation is evaluated by graphical methods in the form of ACF, AFM,

or AKFM diagrams. These methods generally require that a rock, a multicomponent chemical system, be expressed in terms of a simplified system that can be represented by a geometric form. For example, mineral assemblages in pelitic schists are commonly depicted within the AKFM tetrahedron, a three-dimensional compositional space defined by four constituents: Al_2O_3 - K_2O - FeO - MgO . Constituents such as MnO , TiO_2 , Fe_2O_3 , CaO , and Na_2O are omitted. A compositional space in which most rock compositions could be rigorously defined would have many more than three dimensions. Once the number of dimensions exceeds three, evaluation of the importance of compositional variation is no longer a graphical problem, but a mathematical one requiring linear algebra to obtain a solution.

H.J. Greenwood (1967) demonstrated the application of linear programming to the solution of problems involving metamorphic rocks and multidimensional composition spaces. Linear programming, a mathematical method based on linear algebra, allows consideration of all the constituents of a complex mineral assemblage, thereby avoiding most of the simplifications required by the graphical method.

The aim of this chapter is to describe the mineralogical zonation in rocks of the Flinton Group in the study area, to test the validity of the assumption of equilibrium in these rocks, and to show how linear

programming can be used to demonstrate that primary compositional variation is not an important variable in the development of the mineral zonation.

Previous Work

Earlier workers in the Bishop Corners-Madoc area and vicinity (Figure 1), including Meen (1944), Harding (1944), Ambrose and Burns (1956), Burns (1951) and Bain (1960) mentioned the occurrence of metamorphic minerals such as staurolite and sillimanite but detailed studies of mineral zoning with respect to metamorphic grade were not done. Lumbers (1967) defined three metamorphic zones; low, intermediate and high, over a large area including the present study area. The boundary between the low and intermediate zones on Lumbers' Figure 1 (1967) is shown as parallel to the strike of the Flinton Group in the Bishop Corners-Madoc area. Except for the northeast end, the study area lies south-east of this boundary in the intermediate zone. B.L. Smith (1958) noted an increase in metamorphic grade northeastward from Myer Cave (Figure 1). Moore (1967) elaborated on these observations and Hounslow and Moore (1967) studied the chemical aspects and mineral relations in pelitic schist of the Flinton Group in the same area. Metavolcanic rocks east and northeast of Bishop Corners were described by K. Sethuraman (1970) and Sethuraman and

and Moore (in press). Sethuraman (1970) concluded that conditions of metamorphism in the area were between the low pressure Abukuma facies series and the intermediate pressure Barrovian facies series. The presence of sillimanite in metasedimentary rocks north of Kaladar (Figure 1) and in the Clare River synform (Burns, 1951) indicates that metamorphic grade increases southeastward from the study area. In the study area it will be shown that metamorphic grade ranges from that typical of the lowermost amphibolite facies (garnet zone) at Bishop Corners and Madoc to upper middle amphibolite facies (diopside zone) between these localities. Mineral zone boundaries are not parallel to the strike of the Flinton rocks in the Bishop Corners-Madoc area as shown by Lumbers (1967) but cut across the structural trend.

Mineral Zoning

Mineral zoning is most clearly defined, in the study area, in pelitic schist of the Bishop Corners Formation and in the aluminous, iron-bearing calcareous rocks of the Lessard Formation. The regional distribution of mineral zone boundaries in these rocks is illustrated in Figure 14. Location of the sillimanite zone boundary and the diopside zone boundary in metavolcanic rocks east of Bishop Corners is based on data from J.M. Moore (personal communication) and K. Sethuraman (1970). Quartzite, quartzite metaconglomerate, and quartzofeldspathic psammite in

Figure 14: Regional distribution of isograds in the Madoc-Ompah area, Grenville Province, eastern Ontario. Isograds 5 and 6 are taken from Sethuraman, (1972).

the Flinton Group do not exhibit distinctive mineral zoning. In the diopside zone, however, the grain-size of these rocks is larger, and they contain a greater proportion of epidote and microcline and less mica than they do elsewhere in the study area. Chlorite is not present between a point 2 km south of Flinton and Actinolite (Figure 14). Muscovite and quartz co-exist throughout the area. As mineral zoning in metamorphosed volcanic and plutonic rocks was not studied, only a few general observations are given. Chlorite is not present in metavolcanic rocks in the diopside zone. Actinolitic amphibole in the Elzevir diorite (Unit 7, Figure 2) and epidote in the biotite granodiorite (unit 8) are the products of metamorphism. The degree of recrystallization decreases toward the core of the Elzevir batholith.

Petrographic work by J.M. Moore (personal communication), Hounslow and Moore (1967), Sethuraman (1970) and the writer has revealed that mineral zoning in the Flinton Group is the result of one period of regional metamorphism. The internal fabric of some garnet, chloritoid and staurolite porphyroblasts in pelitic schist (see Chapter 3) suggests the possibility of a pre-D₁ period of deformation and metamorphism, but there are no relict porphyroblasts or other mineralogical evidence to support an earlier metamorphism affecting the Flinton Group. White mica alteration of plagioclase (this study) and chlorite in the metavolcanic rocks (Sethuraman, 1970)

are attributed to retrograde metamorphism and weathering. At some localities in the study area plagioclase is completely replaced by fine-grained aggregates of white mica which preserve the original metamorphic texture. Tourmaline is prominent in such rocks in at least two cases.

The following parts of this section will describe the mineral zoning in the biotite-carbonate schist, calcareous psammite, and pelitic schist in the study area.

Biotite-carbonate schist. A general description of the biotite-carbonate schist and its metamorphic equivalent, the epidote-free calc-silicate paragneiss, is presented in Chapter 2, along with the stratigraphic relationships between these rocks and the rest of the Flinton Group. Part of the Lessard Formation, the schist-paragneiss unit occurs in the uppermost portion of the Flinton Group for a strike length of 10 km (Figure 2). Opaque minerals are not common and when present are usually pyrite or pyrrhotite, indicating that these rocks have a relatively low ferric-ferrous ratio compared to the Fe-Ti oxide-bearing rocks which make up most of the Flinton Group in the study area. The lack of muscovite and epidote further distinguish the carbonate schist from calcareous quartzofeldspathic psammite, the other unit in the Lessard Formation. The basal portion of the schist-paragneiss unit contains more quartz and feldspar and less carbonate than the upper portion. The mineral assemblages used to outline mineral zoning occur in the upper

part of the unit.

Mineral zoning is illustrated in Figure 15. The thinness of the unit (50 - 75 m) makes the choice of a trend for the zone boundaries somewhat arbitrary. Data from the calcareous psammite aid in the location of these lines. Mineral zones can be adequately distinguished on the basis of individual minerals. Zone index minerals are discussed after the zoning in the psammite and schist has been described.

The appearance of the rock does not change greatly within Zones I and II or within Zones III and IV, but at the boundary marked by the appearance of microcline (Figure 15), there is a distinct transition over 200 - 300 m from biotite-carbonate schist to microcline-actinolite paragneiss. All along the unit, calcite and biotite are present, though somewhat different in composition and modal percent from place to place. Variation in bulk composition in the schist and paragneiss results in a wide range in the modal composition of mineral assemblages in each zone. The proportion of plagioclase, however, is relatively constant throughout.

Although reaction relationships between minerals are generally not evident in the textures of these rocks, the spatial relationships of some minerals give an indication as to which are involved in the appearance or disappearance of zone index minerals. In Zone I, ferroan-dolomite (see discussion in mineral chemistry

Figure 15: Mineral zoning in the biotite-carbonate schist - epidote-free paragneiss unit (Lessard Formation). Complete mineral assemblages are indicated by the symbols (all rocks contain quartz and plagioclase). Mineral symbols: BI - biotite, CC - calcite, FD - ferroan dolomite, AC - actinolite, MI - microcline, SP - sphene, DI - diopside.

section) and ferroan-calcite are closely associated in carbonate-rich lenses and layers in the rock. In Zone II, particularly in the northern part, actinolite is commonly associated with calcite in similar carbonate lenses. Grain size of actinolite and biotite increases toward the south in Zones II and III (0.1 - 0.3 to 0.4 - 1.5 mm). The marked increase in microcline defining the northern boundary of Zone III corresponds to a sharp reduction in the amount of biotite in the rock. Microcline in some samples contains inclusions of biotite. Sphene also becomes a prominent mineral at this point. Inclusions of actinolite and calcite are found in diopside in Zone IV.

Complete chemical analyses of all the minerals (except accessories) in samples from Zone I (Rock 1) and Zone IV (Rock 2) are presented in Tables 3 and 4 in the mineral chemistry section of this chapter. Petrographic descriptions and modal analyses of these rocks are given in Appendix 2. Mineral assemblages in the samples not described in detail in this chapter are listed in Appendix 3.

Calcareous quartzofeldspathic psammite. The calcareous quartzofeldspathic psammite and its metamorphic equivalent, epidote-bearing calc-silicate paragneiss (both in the Lessard Formation) occur in the cores of synclines in the Bishop Corners, Flinton, and Black River structures (Figures 1 - 3). The total exposure along strike is 23

km. Stratigraphy and lithologic characteristics of this unit are discussed in Chapter 2. The high carbonate content (15 - 25%) of the psammite distinguishes it from the quartzofeldspathic psammite in the underlying Bishop Corners Formation. The calcareous psammite contains 1 - 3 percent Fe-Ti oxides (ilmenite, hematite, and magnetite), resulting in a higher oxidation ratio than that of the carbonate schist. Mineral assemblages are more complex in the oxidized rocks. All the minerals which occur in the schist (except the sulphides) are present in psammite with the addition of Fe-Ti oxides, white mica and epidote. Composition is variable within the unit. Compositional layering (2 - 30 cm) is well-defined. The carbonate-rich portions yield the distinctive mineralogical variation along strike.

Figure 16 illustrates the mineral zoning observed in the calcareous psammite-epidote-bearing paragneiss unit in the northeastern half of the study area. In order to show the details of the mineral variation, only the northeast side of the zoning pattern illustrated in Figure 14 is shown. Sample density and the continuity of the psammite-paragneiss unit is much better in this part of the study area. The mineral zoning for the psammite is quite similar to that for the carbonate schist (Figure 15). The same number of zones is present; Fe-dolomite and diopside are the index minerals for Zones I and IV, respectively. The zones differ in that epidote

Figure 16: Mineral zoning in the calcareous quartzofeldspathic psammite - epidote-bearing calc-silicate paragneiss unit (Lessard Formation). Complete mineral assemblages are indicated by the symbols (all rocks contain quartz and plagioclase). Mineral symbols: PT - phengite, OX - Fe-Ti oxide (hematite-ilmenite), EP - epidote, HN - hornblende, QZ - quartz, PL - plagioclase. All other symbols same as Figure 15.

rather than actinolite is present in Zone II in the rock with the higher oxidation ratio. Amphibole does not appear until Zone III in the psammite. As in the case of the schist-paragneiss unit, the psammite becomes a paragneiss near the boundary between Zones II and III. That is, the sandy or granular texture of the rock disappears with an increase in grain size and reduction in the proportion of mica and carbonate. The transition is gradual over several hundred meters.

In Zone I (Figure 16), phengite occurs in two habits: fine grained lepidoblastic aggregates associated with biotite and oriented parallel to S_1 , and larger porphyroblastic plates which are oblique to S_1 . Fe is 0.2% lower in the fine-grained phengite but this may result from the instrumental difficulties encountered in analyzing very fine-grained aggregates. The other elements determined were the same in both micas within the limits of uncertainty associated with these analyses. The amount of fine-grained phengite in the psammite decreases southward in Zone II, and it is absent from Zones III and IV. The size of the larger grains increases toward the south. For example, in Zone I, matrix phengite laths are 0.1 - 0.3 mm long, the oblique micas are 0.3 - 0.6 mm. In the southern part of Zone II, porphyroblasts of white mica (phengite?) oblique to S_1 defined by biotite are mostly 1 - 2 mm long. The large grains disappear at the southern boundary of Zone II.

Biotite is less abundant in the calcareous psammite (5 - 10%) than in the carbonate schist (25 - 45%). In the psammite the proportion of biotite is about the same throughout Zones I and II, but decreases greatly with the appearance of actinolite in the rock. In the epidote-paragneiss in the diopside zone there is only a trace of biotite, or none at all.

Epidote in the calcareous psammite in the northern part of Zone II is in the form of small (< .5 mm) grains and spongy poikiloblasts, some of which contain metamict cores (relict detrital allanite?). There is a close association between phengite, biotite, and epidote in this zone. Grain size and modal percent of the mineral increase toward the south. Epidote is an essential constituent of the paragneiss equivalent to the psammite throughout the diopside zone.

Microcline is present in small amounts in Zones I and II, increasing toward the south. Microcline becomes an essential mineral in Zones III and IV. Whereas the microcline in the paragneiss is clearly related to disappearance of micas at the northern boundary of Zone III, in Zones I and II the mineral may be partly of detrital origin.

Sphene is also present in all four zones. The small size, well-rounded habit, and golden yellow-brown colour of the sphene in Zones I and II distinguishes it from the larger, sometimes poikiloblastic grains of sphene

in the paragneiss. An increase in sphene in the psammite-paragneiss unit corresponds to a decrease in biotite, muscovite and oxide content of the rock. There is a noticeable reduction in the amount of Fe-Ti oxide in the rock along strike from Zone I to IV.

The distribution of paragneiss bearing amphibole, but without diopside, is more restricted in the psammite-paragneiss unit than it is in the schist-paragneiss unit. Another significant difference between the two units is the presence of an almost colourless amphibole rimming some of the grains of amphibole in the epidote-bearing paragneiss.

Plagioclase, commonly zoned and partly altered to white mica, is relatively constant in proportion throughout.

Detailed petrographic descriptions of Rock 3 (Zone I) and Rock 4 (Zone IV, Figure 16) are included in Appendix 2. Complete chemical analyses of all but the accessory minerals in these rocks are presented in Tables 5 and 6 in the mineral chemistry section of this chapter.

Zone boundaries defined in the biotite-carbonate schist (Figure 15) are included on Figure 16. The boundaries between Zones I and II, representing the disappearance of Fe-dolomite in the schist and psammite, do not coincide. Fe-dolomite is stable in the carbonate schist beyond the point where it disappears from the psammite. Furthermore, whereas actinolite appears south of the boundary in the schist, epidote appears south of

the boundary in the calcareous psammite. To take into account the effect of bulk composition on the disappearance of Fe-dolomite, the Fe-dolomite zone is subdivided into Zones I and IA (Figure 16) and Zone II is designated the epidote-actinolite zone.

Bulk composition does not affect the position of the boundaries between Zones II and III. At this point in both rock units, a substantial decrease in the amount of mica and a corresponding increase in the microcline and amphibole is observed. Zone III represents a second actinolite zone in the carbonate schist. To avoid confusion, Zone III is designated the microcline zone, even though small amounts of microcline (1 - 5%) occur in the calcareous psammite in Zones I and II.

Zone IV is characterized by the presence of diopside. The boundary marking the first appearance of diopside in the epidote-bearing paragneiss (calcareous psammite) coincides with that in the epidote-free paragneiss (biotite-carbonate schist). Another unique characteristic of Zone IV is the widespread occurrence of pegmatite (Figure 16).

Pelitic schist. The pelitic schist, part of the Bishop Corners Formation, occurs at the base of the Flinton Group, adjacent to rocks beneath the unconformity.

The unit, generally less than 50 m thick, is sparsely exposed in the study area because the basal contact of the Flinton Group is commonly covered by a swamp. In Chapter 2 three types of pelitic schist were described: gray schist containing muscovite (20 - 40%), black schist without muscovite, and a rock transitional between these two. The mineral zoning described here is limited to the muscovite-bearing schists. Some systematic variation in mineral assemblages is present in the muscovite-free schist but low sample density precludes definition of mineral zones. Gray pelitic schist in the Fernleigh syncline near Bishop Corners (Figure 3) and the Madoc area contains 30 to 40 percent muscovite. South of the Bishop Corners area, and near Flinton (Figure 2) the modal percent of muscovite in the pelitic schist is 5 to 15%. These rocks contain considerably larger amounts of plagioclase (20%) than the muscovite-rich schists where plagioclase is not common. Both rock types contain 2 - 10% Fe-Ti oxides.

Several unusual features are found in the pelitic schist discussed here. Mineral assemblages containing four ferro-magnesian minerals (e.g. chloritoid-biotite-chlorite-garnet) are common. The calcium mica, margarite, is present at several localities both at Bishop Corners and near Madoc. South of Bishop Corners, chloritoid coexists with hornblende in muscovite-rich pelitic schist. The mineralogical similarities between the pelitic rocks at Bishop Corners and those at Madoc

are striking.

The absence of pelitic rocks in the central part of the study area and between the Black River and Madoc areas is the reason for the small number of mineral zones shown in Figures 17 and 18. Zones I and II are distinguished on the basis of the presence or absence of chloritoid. The mineral assemblages in Zone I near Bishop Corners (Figure 17) can be further subdivided into those containing chlorite ($\text{Fe:Mg} > 1.0$) and those containing staurolite. The dotted line is drawn between the occurrences of chloritoid-chlorite-biotite-garnet and chloritoid-staurolite-biotite-garnet. The subdivision is not used throughout the study area as a zone boundary because the pelitic rocks at Madoc (Figure 18) cannot be separated in this way.

Chlorite in Zone I is length slow with brown to purplish-brown anomalous interference colours. In Zone II, the mineral is length fast with gray interference colours. The chlorite in Zone II is a magnesium-rich chlorite by the criteria of Albee (1962), consistent with mineral assemblages in other areas where chlorite is coexisting with staurolite and/or kyanite. Near Bishop Corners, chloritoid is intimately associated with staurolite. The textural relationship between these minerals suggests two possibilities: staurolite growing inside or adjacent to chloritoid, or nucleation of chloritoid on staurolite. Further to the northeast there is

Figure 17: Distribution of pelitic rocks (Bishop Corners Formation) in the northern half of the Flinton map area (Figure 2) and mineral zoning in muscovite-bearing pelitic rocks. All rocks contain quartz and muscovite. Complete mineral assemblages are not indicated by the symbols in this figure.

Figure 18: Distribution of pelitic rocks in the Madoc area (Madoc Formation) and in the Black River area (Bishop Corners Formation) and mineral assemblages in muscovite-bearing pelites. All rocks contain quartz, muscovite, and Fe-Ti oxides. Mineral symbols: CT - chloritoid, GA - garnet, ST - staurolite, CH - chlorite, Ky - kyanite, BI - biotite.

specific evidence of the decomposition of chloritoid, in the form of white mica pseudomorphs after chloritoid (J.M. Moore, personal communication). Garnets in the pelitic schist commonly include chloritoid grains and exhibit distinctive overgrowth textures (Plate 3e,f, Chapter 3).

Detailed petrographic descriptions and modal analyses of two rocks, one from each zone (Rocks 5 and 6, Figure 17), are presented in Appendix 2. Chemical analyses of all the minerals (except accessories) in each rock are included in Tables 7 and 8 in the mineral chemistry section of this chapter.

Before the mineral zoning in the pelitic schist, calcareous psammite, and carbonate schist can be discussed in terms of changes in metamorphic grade and before the term isograd can be substituted for zone boundary, two questions must be answered: Was a state of chemical equilibrium attained at the time the mineral assemblages in these rocks were formed? Is the mineral zoning the result of changes in physical conditions or of variable bulk composition?

Mineral Chemistry

To evaluate the extent of chemical equilibrium in metamorphic rocks, the compositions of coexisting minerals must be known. To apply linear programming to the problem of variable bulk composition, the compositions of

all the minerals in a pair of rocks and the uncertainties associated with these compositions need to be measured. Two samples from each of the biotite-carbonate schist, calcareous psammite, and pelitic schist units were chosen to include as much mineralogical variation between the two sample localities as possible (sample localities on Figures 15, 16, and 17). Thin sections of rocks from these localities were viewed and an unweathered sample containing all the minerals common to the particular mineral zone was selected. Rocks 1 and 3 are from Zone I (Fe-dolomite) in the carbonate schist and calcareous psammite, respectively. Rocks 2 and 4 are from the paragneiss equivalents of these rocks in the diopside zone (IV). One pelitic schist (Rock 5) was taken from Zone I (chloritoid present) and the other (Rock 6) is from Zone II (chloritoid absent). Petrographic descriptions and modal analyses of these rocks are presented in Appendix 2.

Chemical analyses presented in this chapter were determined using a Cambridge Microscan 5 electron microprobe. The instrument has two spectrometers and a take-off angle of 75° . Silicate standards were used when analyzing for Al, Si, Na, K, Ca, and Mg, pure element standards for Fe, Mn and Ti, and millerite for S. Silicates, oxides, and sulphides were analyzed with an operating voltage of 15 kilovolts and specimen current of $45-55 \times 10^{-9}$ amps. To avoid decomposition of calcite

by the electron beam, 13 kilovolts and specimen current of 35×10^{-9} amps were used for carbonate analyses. Six thin sections, one from each sample, were studied. Elements were analyzed in pairs and not all 10 listed above were done in every mineral. 20 to 30 analyses, with a ten-second counting period, were randomly placed on a single relatively large grain as a test for homogeneity in each mineral. If zoning was not present 10 randomly-placed ten-second counting periods made up a typical analysis for each pair of elements in each grain. The homogeneity test was done for each pair of elements. For zoned garnets, compositions for the core and rim of each grain were obtained. Homogeneity of the rim was checked with at least 10 ten-second counts at different spots on the rim. Another ten stops were made in the core zone. Zoned plagioclase was analyzed by a similar method with the object of determining the extremes of compositional variation in each grain. The number of grains of each mineral analyzed ranged from 5 to 20, with the larger number required for fine-grained minerals. To measure the composition of aggregates of very fine grains (short dimension less than 2 - 5 microns) of white mica in the matrix of pelitic schist, 40 ten-second counting periods at 20 spots scattered around the thin section were used. With this number of spots it was possible to distinguish irregular compositions which could be attributed to hidden quartz or feldspar grains.

EMPADR VII (Rucklidge and Gasparri, 1969), a probe data processing program, applies corrections (Sweatman and Long, 1969) to the raw data and prints out the compositions of minerals analyzed in the form of mean weight percent of the elements, atomic proportions of the oxides, and oxide weight percentages. Associated with each mean oxide weight percent value is a variance which is derived from the variance of the sets of counts for each element. The variance is expressed as a percentage of the total for the sample.

Chemical analyses of all the major minerals in Rocks 1 to 6 and the uncertainty associated with each oxide weight percentage are presented in Tables 3 to 8.

Discussion of Chemical Analyses.

Microprobe analytical procedures do not differentiate between ferric and ferrous iron. The iron content of a mineral is calculated as though the iron were entirely ferrous. For the samples (3, 4, and 6) containing minerals whose ferric iron content could be approximated or calculated, ferric iron was included as a constituent. The amount of ferric iron in magnetite and ilmenite was calculated from the probe data on the basis of stoichiometry. The iron in epidote (Rock 4) and in the feldspars in Rocks 3, 4, and 6 was assumed to be entirely ferric. This was not done for the iron in plagioclase and microcline in Rocks 1 and 2 because nothing was

Table 3 - Mineral Analyses - Rock 1

	BIOTITE (6)*	PLAGIOCLASE (5)	FE-DOLOMITE (4)	CALCITE (4)	RUTILE (11)
SiO ₂	37.70 ± .30	61.63 ± .37	-	-	-
TiO ₂	1.94 ± .21	-	-	-	96.28 ± .70
Al ₂ O ₃	17.25 ± .22	22.67 ± .31	-	-	-
FeO**	12.70 ± .28	0.54 ± .15	7.54 ± .39	0.52 ± .14	1.17 ± .20
MnO	-	-	0.38 ± .13	0.48 ± .14	-
MgO	15.98 ± .22	-	16.72 ± .31	0.34 ± .10	-
CaO	0.00	4.61 ± .30	29.46 ± .33	54.86 ± .46	-
Na ₂ O	0.11 ± .07	8.67 ± .33	-	-	-
K ₂ O	10.03 ± .30	0.19 ± .12	-	-	-
CO ₂ ***	-	-	46.22 ± 1.16	44.02 ± .84	-
Total	95.71 ± .72	98.31 ± .68	100.32 ± 1.31	100.22 ± .97	97.45 ± .73
Number of ions based on x oxygens					
x	22	32	4	4	2
Si	5.54	11.11			
Al	2.46	4.84			
Al	.53				
Ti	.22				
Fe ³⁺	1.56	0.08	0.40	0.03	0.99
Fe ²⁺	3.50		0.02	0.03	0.01
Mn			1.58	0.03	
Mg	.03	0.89	2.00	3.91	
Ca	1.88	3.03			
Na		0.04			
K					
%An		23			

* numbers in brackets refer to number of grains analyzed

** FeO - Total Fe as FeO

*** CO₂ from fixed stoichiometry

- not determined

Table 4 - Mineral Analyses - Rock 2

	DIOPSIDE (11)*	CALCITE (10)	PLAGIOCLASE (8)	MICROCLINE (8)	ACTINOLITE (10)	SPHENE (13)	BIOTITE (5)
SiO ₂	53.35 ± .24	-	62.20 ± .31	64.98 ± .31	53.54 ± .42	30.65 ± .24	38.72 ± .28
TiO ₂	0.03 ± .05	-	-	-	0.13 ± .07	36.94 ± .70	1.95 ± .15
Al ₂ O ₃	0.62 ± .23	-	23.27 ± .25	18.05 ± .27	3.20 ± .37	2.50 ± .25	16.92 ± .24
FeO**	7.04 ± .24	0.31 ± .09	0.03 ± .04	0.02 ± .02	10.10 ± .25	0.27 ± .07	13.72 ± .27
MnO	0.22 ± .09	0.17 ± .07	-	-	0.20 ± .07	-	-
MgO	14.23 ± .22	0.24 ± .09	-	-	18.46 ± .29	0.02 ± .05	15.54 ± .24
CaO	24.89 ± .34	55.52 ± .55	5.63 ± .21	0.00	12.64 ± .27	28.18 ± .29	0.00
Na ₂ O	0.23 ± .09	-	8.49 ± .21	0.65 ± .16	0.45 ± .13	-	0.05 ± .01
K ₂ O	-	-	0.23 ± .12	16.26 ± .25	-	-	9.28 ± .30
CO ₂ ***	-	44.12 ± .57	-	-	-	-	-
Total	100.59 ± .59	100.36 ± .97	99.85 ± .51	99.87 ± .51	98.47 ± .75	97.56 ± .84	96.19 ± .61

x	Number of ions based on x oxygens					
	6	4	32	32	23	22
Si	1.97	-	11.07	12.03	7.55	5.65
Al	.03	-	4.87	3.93	.45	2.35
Al	.001	-	-	-	.08	.56
Ti	.22	-	-	-	.01	.21
Fe ³⁺	.01	.018	.004	.003	1.19	1.67
Fe ²⁺	.79	.012	-	-	.02	5.82
Mn	.99	.024	-	-	3.88	3.38
Ca	.02	3.958	1.08	.23	1.91	.01
Na	-	-	2.92	3.84	.10	1.73
K	-	-	.05	-	-	-
4An	-	-	-	-	-	-

* numbers in brackets refer to number of grains analyzed

** FeO - Total Fe as FeO

*** CO₂ from fixed stoichiometry

- not determined

Table 5 - Mineral Analyses - Rock 3

	PHENGITE (7)	PLAGIOCLASE A*(6)	PLAGIOCLASE S(3)	PLAGIOCLASE C(6)	CALCITE (5)	FE-DOLONITE(2)	BIOTITE (7)	FE-TI OXIDE
SiO ₂	45.81 ± .31	65.97 ± .41	66.46 ± .23	65.47 ± .46	-	-	37.36 ± .21	-
TiO ₂	1.24 ± .12	-	-	nd	-	-	1.53 ± .12	9.41 ± .37
Al ₂ O ₃	30.41 ± .33	20.93 ± .33	20.85 ± .26	21.01 ± .36	-	-	17.18 ± .26	-
Fe ₂ O ₃ **	-	0.21 ± .14	0.22 ± .08	0.19 ± .15	-	-	-	47.60 ± .50
FeO***	4.37 ± .22	-	-	-	1.00 ± .16	5.14 ± .26	12.78 ± .25	38.19 ± .50
MnO	-	-	-	-	0.42 ± .16	0.58 ± .12	-	-
MgO	1.80 ± .16	-	-	-	1.65 ± .20	17.09 ± .29	15.10 ± .22	-
CaO	0.02 ± .05	1.97 ± .29	1.35 ± .08	2.26 ± .09	52.82 ± .44	29.99 ± .37	0.10 ± .07	-
Na ₂ O	0.33 ± .09	13.62 ± .25	10.87 ± .26	10.36 ± .24	-	-	0.15 ± .05	-
K ₂ O	10.73 ± .22	0.07 ± .07	0.06 ± .05	0.07 ± .07	-	-	9.72 ± .19	-
CO ₂ ****	-	-	-	-	44.12 ± .53	45.68 ± .54	-	-
Total	94.54 ± .67	99.77 ± .55	100.14 ± .45	99.36 ± .66	100.01 ± 1.05	98.48 ± 1.14	93.92 ± .53	95.20 ± .64

Number of ions based on x oxygens

	22	32	32	32	4	4	22
x	6.28	11.63	11.66	11.59	5.59	5.59	5.59
Si	1.72	4.35	4.31	4.38	2.41	2.41	2.41
Al	3.19	-	-	-	.62	.62	.62
Al	.11	-	-	-	.17	.17	.17
Ti	.50	.03	.03	.03	1.60	1.60	1.60
Fe ³⁺	.37	.37	.25	.43	.28	.28	.28
Fe ²⁺	.003	3.63	3.70	3.55	.02	.04	.04
Mn	.09	.02	.01	.02	1.64	1.64	1.64
Ca	1.88	-	-	-	2.06	2.06	2.06
Na	-	-	-	-	4.02	4.02	4.02
K	-	-	-	-	1.92	1.92	1.92
%An	-	9	6	11	-	-	-

* A - "average"; S - sodic; C - calcic

** Total Fe as Fe₂O₃ in Plagioclase; Fe₂O₃ is calculated on basis of stoichiometry for Fe-Ti oxides.

*** Total Fe as FeO - not determined

**** CO₂ from fixed stoichiometry

Table 6 - Mineral Analyses - Rock 4

	MICROCLINE (5)	PLAGIOCLASE A (5)	PLAGIOCLASE C (5)	PLAGIOCLASE S (5)	EPIDOTE (5)
SiO ₂	63.75 ± .44	62.00 ± .65	60.77 ± .33	64.13 ± .45	38.04 ± .22
TiO ₂	-	-	-	-	0.14 ± .07
Al ₂ O ₃	19.16 ± .38	23.72 ± .52	24.69 ± .29	22.28 ± .31	23.59 ± .32
Fe ₂ O ₃ *	0.05 ± .05	0.04 ± .05	0.04 ± .04	0.05 ± .04	12.83 ± .29
FeO**	-	-	-	-	-
MnO	-	-	-	-	0.07 ± .05
MgO	-	-	-	-	0.00
CaO	0.00	4.78 ± .51	5.79 ± .26	3.28 ± .24	23.80 ± .27
Na ₂ O	0.06 ± .15	8.80 ± .40	8.19 ± .23	8.77 ± .24	-
K ₂ O	15.57 ± .29	0.14 ± .10	0.17 ± .08	0.12 ± .07	-
CO ₂ ***	-	-	-	-	-
Total	98.59 ± .67	99.48 ± 1.03	99.65 ± .57	99.63 ± .65	98.47 ± .56

Number of ions based on x oxygens

x	32	32	32	13
Si	11.90	11.04	10.83	3.13
Al	4.22	4.98	5.19	2.29
Al				.01
Ti	.01	.01	.01	.79
Fe ³⁺				3.09
Fe ²⁺				.01
Mn				.01
Ca	.02	.91	1.11	2.10
Na	3.71	3.04	2.83	2.11
K		.03	.04	
%An		23	28	16

* Total Fe as Fe₂O₃ in feldspars and epidote

** Total Fe as FeO

*** CO₂ from fixed stoichiometry

- not determined

Table 6 (con'd) - Mineral Analyses - Rock 4

	CALCITE (2)	DIOPSIDE (5)	HORNBLende (5)	ACTINOLITE (2)	SPHENE (5)
SiO ₂	-	52.29 ± .30	47.17 ± .42	52.85 ± .51	30.45 ± .24
TiO ₂	-	0.06 ± .05	0.21 ± .07	0.14 ± .09	36.99 ± .47
Al ₂ O ₃	-	1.36 ± .22	7.71 ± .38	3.12 ± .44	1.85 ± .29
Fe ₂ O ₃ *	-	-	-	-	-
FeO**	0.30 ± .09	8.58 ± .28	13.02 ± .30	9.89 ± .28	1.23 ± .16
MnO	0.25 ± .09	0.34 ± .09	0.27 ± .09	0.24 ± .09	-
MgO	0.26 ± .10	13.66 ± .27	15.01 ± .28	18.66 ± .21	0.01 ± .00
CaO	55.18 ± .56	24.37 ± .27	12.41 ± .24	12.24 ± .26	29.23 ± .31
Na ₂ O	-	0.57 ± .12	1.17 ± .18	0.91 ± .50	-
K ₂ O	-	-	-	-	-
CO ₂ ***	43.96 ± .58	-	-	-	-
Total	99.97 ± 1.01	101.23 ± .62	96.97 ± .77	98.05 ± .30	99.76 ± .69
Number of ions based on x oxygens					
x	4	6	23	23	20
Si		1.94	6.95	7.52	4.00
Al		.06	1.05	.48	
Al			.29	.04	.29
Ti		.002	.02	.01	3.66
Fe ³⁺					.14
Fe ²⁺	.02	.27	1.60	1.18	4.09
Mn	.02	.01	.03	.03	
Mg	.76	.76	3.29	3.96	.002
Ca	3.94	.97	1.96	1.87	4.12
Na		.04	.33	.25	
K					
%An					

* Total Fe as Fe₂O₃ in feldspars and epidote

** Total Fe as FeO

*** CO₂ from fixed stoichiometry - not determined

Table 7 - Mineral Analyses - Rock 5

	MUSCOVITE(20)	CHLORITE(8)	GARNET A(5)	GARNET C(5)	GARNET R(5)	CHLORITOID(5)	BIOTITE(5)	ILMENITE(10)
SiO ₂	44.70 ± .54	23.00 ± .24	36.99 ± .36	36.99 ± .42	36.99 ± .29	25.02 ± .24	34.15 ± .31	-
TiO ₂	0.17 ± .09	0.06 ± .04	-	-	-	0.01 ± .00	1.56 ± .19	58.51 ± .71
Al ₂ O ₃	37.02 ± .43	21.78 ± .36	20.89 ± .31	20.82 ± .32	20.95 ± .30	38.94 ± .40	17.63 ± .37	-
Fe ₂ O ₃	-	-	-	-	-	-	-	-
FeO*	2.49 ± .19	30.73 ± .36	35.78 ± .37	32.06 ± .36	39.49 ± .38	26.24 ± .36	22.71 ± .37	41.92 ± .43
MnO	-	0.03 ± .03	3.68 ± .19	6.79 ± .25	0.57 ± .12	0.06 ± .04	-	-
MgO	0.65 ± .11	10.76 ± .22	1.17 ± .12	.73 ± .09	1.60 ± .16	2.02 ± .16	8.02 ± .20	-
CaO	0.00	0.01 ± .01	3.10 ± .20	4.56 ± .23	1.63 ± .17	0.00	0.01 ± .04	-
Na ₂ O	0.82 ± .39	0.09 ± .07	-	-	-	-	0.22 ± .09	-
K ₂ O	9.55 ± .31	-	-	-	-	0.00	8.92 ± .28	-
Total	95.40 ± .88	86.46 ± .61	101.42 ± .61	101.60 ± .61	101.23 ± .62	92.29 ± .61	93.22 ± .73	100.43 ± .83

Number of ions based on x oxygens

x	22	28	24	24	24	24	12	22	6
Si	5.96	5.09	5.97	5.97	5.97	5.97	2.04	5.42	
Al	2.04	2.91	.03	.03	.03	.03	3.84	2.58	8.00
Al	3.78	2.78	3.94	3.92	3.96			.72	
Ti	.02	.01					.001	.19	
Fe ⁺⁺	.28	5.69	4.82	4.32	5.33				5.82
Fe ⁺⁺	4.21	12.10	.50	.93	.08	1.80	2.04	3.01	2.15
Mn		.01	.29	.18	.39	.004	.24	1.90	1.71
Mg	.13	3.55	.53	.79	.28			.002	
Ca		.01						.07	
Na	.21							1.81	
K	1.62	.04							

* Total Fe as FeO
- not determined

Table 8 - Mineral Analyses - Rock 6

	MUSCOVITE (10)	BIOTITE (5)	PLAGIOCLASE A*(5)	CHLORITE (8)	STAUROLITE (5)	KYANITE (5)
SiO ₂	45.15 ± .23	36.41 ± .24	56.30 ± .82	25.20 ± .26	28.01 ± .23	37.78 ± .23
TiO ₂	0.35 ± .09	1.46 ± .12	-	0.05 ± .05	0.45 ± .10	0.13 ± .09
Al ₂ O ₃	36.37 ± .34	19.23 ± .33	27.97 ± .72	22.87 ± .29	53.79 ± .46	61.18 ± .39
Fe ₂ O ₃ **	-	-	0.06 ± .04	-	-	-
FeO***	2.20 ± .14	13.05 ± .33	-	16.00 ± .25	12.17 ± .22	0.77 ± .16
MnO	0.02 ± .02	0.11 ± .05	-	0.27 ± .07	0.76 ± .11	-
MgO	0.73 ± .12	14.53 ± .22	-	20.49 ± .26	2.28 ± .12	-
CaO	-	-	9.61 ± .45	-	-	-
Na ₂ O	0.90 ± .11	0.25 ± .07	5.97 ± .41	-	-	0.08 ± .05
K ₂ O	9.79 ± .22	9.27 ± .20	0.05 ± .02	-	-	0.06 ± .05
Total	95.51 ± .52	94.31 ± .64	99.96 ± 1.25	84.88 ± .54	97.46 ± .59	100.00 ± .49

	Number of ions based on x oxygens					
x	22	22	32	28	46	20
Si	6.01	5.42	8.00	5.23	7.75	4.10
Al	1.99	2.58	10.11	2.77	17.56	7.81
Al	3.72	.79	5.92	2.82	.10	.01
Ti	.04	.16	-	.01	-	-
Fe ³⁺	.24	1.62	.01	2.77	2.82	.07
Fe ²⁺	.002	.01	-	.05	.18	-
Mn	.15	3.22	-	6.33	.94	.02
Mg	.23	.07	1.85	-	-	.01
Ca	1.66	1.76	2.08	-	-	-
Na	-	-	.01	-	-	-
K	-	-	-	-	-	-
%An	-	-	-	-	-	-

* A - "average" plagioclase

- not determined

**

Total Fe as Fe₂O₃ in plagioclase; Fe₂O₃ in oxides based on stoichiometry

*** Total Fe as FeO in silicates; fixed stoichiometry for oxides

Table 8 (con't.) - Mineral Analyses - Rock 6

	GARNET A* (9)	GARNET C (3)	GARNET R (3)	OX1 (5)	OX2 (2)	OX3 (15)
SiO ₂	37.38 ± .25	37.36 ± .26	37.41 ± .24	-	-	-
TiO ₂	-	-	-	9.41 ± .37	0.04 ± .01	51.80 ± .50
Al ₂ O ₃	21.37 ± .30	21.47 ± .28	21.26 ± .31	-	-	-
Fe ₂ O ₃ **	-	-	-	47.60 ± .50	68.90 ± .50	-
FeO***	22.57 ± .34	22.43 ± .38	22.71 ± .31	38.19 ± .50	30.80 ± .50	44.10 ± .50
MnO	12.65 ± .23	13.02 ± .22	12.30 ± .25			
MgO	3.10 ± .14	3.14 ± .15	3.07 ± .13			
CaO	2.89 ± .17	2.87 ± .17	2.92 ± .17			
Na ₂ O	-	-	-			
K ₂ O	-	-	-			
Total	99.96 ± .61	100.29 ± .60	99.67 ± .58	95.20 ± .37	99.74 ± .71	95.90 ± .71

Number of ions based on x oxygens

x	24	24	24
Si	5.99	5.97	6.00
Al	.01	.03	6.00
Al	4.02	4.01	4.02
Ti			
Fe ³⁺	3.02	3.00	3.05
Fe ²⁺	1.72	1.76	1.67
Mn	.74	.75	.73
Mg	.50	.49	.50
Ca			
Na			
K			

* A - "average" garnet; C - garnet core; R - garnet rim
 ** Total Fe as Fe₂O₃ in silicates; fixed stoichiometry for oxides
 *** Total Fe as FeO in silicates; fixed stoichiometry for oxides
 - not determined

known about the ferric-ferrous ratio in the other minerals in these samples.

Although analysis of most of the minerals in the samples used for this study was straightforward, Fe-Ti oxides and chemically zoned minerals required special treatment. The Fe-Ti oxides were the most difficult to deal with. Submicroscopic lamellae of one mineral in another, for example, of hematite in ilmenite, could not be clearly distinguished, with the result that variable compositions intermediate to those of the host and the inclusions were obtained. The small size of the lath-shaped opaque minerals added to the difficulties. For Rock 5, the composition of the Fe-Ti oxide present was consistent from grain to grain in most of the grains. The mineral has the optical properties of ilmenite. Grains of Fe-Ti oxide in Rock 3 were also consistent but the composition was not appropriate for a single mineral. An artificial composition was calculated from stoichiometry using the magnetite-ulvospinel series as a basis.¹ The average compositions for ilmenite in Rock 5 and the Fe-Ti mineral grains in Rock 3 are included in Tables 5 and 7. These results were used to obtain an average composition for the oxide mineral in each rock. The chemical analyses of oxide minerals in Rock 6 generally totaled less than 85% and therefore were not suitable for this

¹ further work has demonstrated the oxide grains to be composed of hematite and ilmenite. This information would not seriously affect the linear programming results.

study. However, the oxides could be grouped into three types; anhedral grains with a composition approaching that of the Fe-Ti oxide in sample 3; lath-shaped grains similar in composition to the ilmenite in sample 5; and euhedral grains which did not correspond to any of the compositions measured in this study. The rhombic outline of the mineral suggests it to be magnetite. To make a preliminary run with linear programming with the pair of pelitic schists (5 and 6), the compositions for ilmenite in Rock 5, Fe-Ti oxide in Rock 3, and a magnetite from the literature representing the euhedral oxide grain were substituted for the three opaque minerals present in Rock 6 (Table 8).²

Zoned plagioclase grains were analyzed at 15 to 20 different spots located randomly over the grain. The information from each grain was first processed as a unit in order that a mean composition could be obtained. The mean composition is not the bulk composition because no attempt was made to determine the relative amounts of the various compositions present in each grain. The original probe data for each grain were then separated into three groups, one representing the most sodic compositions, the second representing the most calcic compositions, and the third group including the compositions

² Further work has shown the anhedral grains to be composed of ilmenite and hematite; the lath-shaped (plates) grains are ilmenite and the euhedral rhombic grains are magnetite.

between these extremes. On the second processing run each grain was represented by the calcic group and the sodic group. The intermediate compositions were omitted. From this information, the composition of the most sodic and most calcic portions of the plagioclase grains in the rock were obtained. The extreme sodic, calcic, and "average" compositions are designated PLAGIOCLASE S, PLAGIOCLASE C, AND PLAGIOCLASE A respectively, on Tables 5 and 6. Knowledge of the extreme plagioclase compositions in two rocks is useful when an attempt is made to find an hypothetical bulk composition common to both. Plagioclase in Rocks 1 and 3, the biotite carbonate schist (An_{23}) and epidote-free calc-silicate paragneiss (An_{27}) is not zoned. The zoned plagioclase in Rock 6 (chloritoid absent) was not treated as outlined above because plagioclase is not present in the other member of the pair of pelitic schist samples (Rock 5). A comparison of extreme compositions was not required. The "average" composition of plagioclase in Rock 6 is An_{47} . Plagioclase in Rock 4 (epidote-bearing paragneiss) exhibited normal zoning (calcic core, sodic rim); Rock 6 (pelitic schist) exhibited reverse zoning (sodic core, calcic rim); and in Rock 3 zoning was not as well-defined, compositional variation being patchy. PLAGIOCLASE A, S, and C in Rock 3 and PLAGIOCLASE S in Rock 4 have compositions in the peristerite range ($An_3 - An_{22}$). The extreme sodic and calcic compositions

determined in Rock 3, and most sodic in Rock 4 are also in the peristerite range. If the plagioclases in Rocks 3 and 4 are intergrowths of albite and oligoclase, the width of the individual lamellae must be less than the diameter of the microprobe electron beam.

Analysis of garnet compositions was done somewhat differently because the chemical zoning was more regular than that in plagioclase. Furthermore, the plentiful very fine-grained inclusions in the garnets in sample 5 (pelitic schist) made random spots less desirable. The ten-second counting periods were restricted to localities near the rim and in a small portion of the core. Data from garnets in this sample together with a preliminary study carried out on the garnets illustrated in Plate 3f (Chapter 3) indicate that chemical zoning is continuous across the well-preserved overgrowth fabrics. During the processing of the data the compositions for rim and core were kept separate. An "average" composition similar to that described for plagioclase was calculated from the rim and core compositions. These compositions are designated GARNET A (average), GARNET C (core), and GARNET R (rim) in Tables 7 and 8.

White mica in the samples of pelitic schist is very close to the boundary used by Deer, Howie, and Zussman (1962) to distinguish phengite from muscovite ($\text{Si} : \text{Al} > 3:1$). In Rock 5 the silicon-aluminum ratio in white mica is 2.98:1.02, and in white mica in Rock 6 the

ratio is 3.01:.99. Both minerals are phengitic muscovites. White mica in Rock 3 is phengite, Si:Al = 3.14:0.86.

The Fe-Mg-Ca carbonate in Rock 3 (Mg:Fe = 5.85:1.00) is ferrodolomite according to the definition of Deer, Howie, and Zussman (1966, p. 494, Mg:Fe > 4:1). The analogous mineral in Rock 1 has a Mg:Fe ratio of 3.95:1.00, just barely making it ankerite. The Mg:Fe ratio used to distinguish ferrodolomite from ankerite was arbitrarily set at 4:1, in order that the commonly occurring Fe-Mg-Ca carbonates with Mg:Fe = 1:1 would be included in the ankerite group. The labelling of mineral zones in the study area is simplified if the Fe-Mg-Ca carbonate in Rock 1 is designated Fe-dolomite. Considering the arbitrary nature of the definition stated above, and the likelihood of a complete solid solution between dolomite and ankerite (Deer, Howie, and Zussman, 1962), such a designation is reasonable.

The transition from colourless to blue-green amphibole within the same grain in the epidote-bearing calc-silicate paragneiss (4) corresponds to an increase in Fe, Ti, Mn, Al and Na, and a decrease in Si and Mg. The amount of Ca remains almost constant. These changes occur across a distance of 5 - 10 microns in at least one grain. The extreme compositions are designated ACTINOLITE (colourless) and HORNBLENDE (blue-green) in Table 6. Mineral names were applied on the basis of

compositional criteria from Deer, Howie, and Zussman
(1962).

Discussion of Uncertainty

The uncertainty quoted for each constituent in Tables 3 to 8 was arrived at from the following expression:

$$U_i = \frac{1}{2} \sqrt{\sum_{j=1}^n v_i^j}$$

U_i - uncertainty of constituent i
 n - number of grains analyzed
 v_i^j - variance of constituent i in grain j

The variance for each constituent in each grain is based on the variance of the set of counts obtained at the time of the analysis for a particular element. The variance results from spot to spot variation in the composition of a mineral grain and instrumental factors such as drift during the analysis of the grain. The uncertainty calculated this way does not reflect the variation from grain to grain in a particular sample.

Theory

Equilibrium

Equilibrium is a state of minimum energy for a system of fixed composition under a particular set of intensive parameters. A system has attained equilibrium when the values for all the intensive parameters are the same in all parts of the system.

For the purpose of this study a system is defined as a thin section of rock. The system is composed of phases, entities which can be separated on the basis of physical properties. A fluid phase is assumed to have coexisted with the mineral phases at the time of metamorphism. Phases consist of chemical species. The chemical species can be dealt with as constituents in the form of oxides or as components (defined below).

If the effects of electromagnetic and gravita-

tional fields and surface area are neglected, the change in free energy of a system can be related to the intensive and extensive parameters of the system by the fundamental thermodynamic equation:

$$dG = TdS + PdV + \sum_i \mu_i dn_i,$$

where G = Gibbs' free energy

T = temperature

P = pressure

μ_i = chemical potential of component i

S = entropy

V = volume

n_i = mole number of component i

The definition of the term "component" used in this study is as follows:

"The actual components of a system are the independently variable chemical species which contribute significantly to the free energy of the system." (G.B. Skippen, lecture notes, 1972).

From the preceding equation the intensive parameters defining the state of the system are temperature, pressure, and the chemical potential of each component. During regional metamorphism the temperature can be safely assumed to be constant throughout a system the size of a polished thin section (0.05 X 20 X 40 mm). The pressure on the solid phases in the system $P_{\text{lithostatic}}$ may be different from the fluid pressure, P_{fluid} . For a rock

undergoing deformation, even with $P_f = P_l$, the pressure in one dimension differs from that in a second dimension. For systems the size of those used in this study, the assumption that the magnitude of each of these pressures is equal in all parts of the system is valid.

Chemical potential (μ) was defined by J.W. Gibbs to account for the proportion of the free energy of a system which can be attributed to compositional variation. The chemical potential of component i is the change in free energy of the system with a change in the number of moles of component i , the temperature, pressure and mole numbers of other components being constant.

Although the temperature, pressure, and composition of the fluid phase in the rock at the time of metamorphism cannot be measured, the compositions of the minerals (solid phases) are known. Chemical potential is related to composition in an ideal solid solution by the following equation:

$$\mu_i^\alpha = \mu_i^\circ + RT \ln X_i^\alpha$$

μ_i^α = chemical potential of component i in solution (mineral) α

μ_i° = chemical potential of pure component i

R = gas constant

T = temperature, $^\circ K$

X_i^α = mole fraction component i in mineral α

For chemical equilibrium in a system, the intensive parameter μ_i and hence X_i is constant within each phase. The composition of each grain of a particular phase is the same in all parts of the system. Natural minerals are far from ideal. Crystallographic imperfections, submicroscopic inclusions and departures from pure end member compositions are common in most minerals. Despite these complications, the extent to which the composition of a mineral varies within a thin section should be a practical measure of the validity of the assumption that chemical equilibrium was attained in a metamorphic rock.

On the basis of textural criteria, all six of the rocks used for detailed work in this study are equilibrium assemblages. Reaction textures and pseudomorphs are not present.

Element distribution relationships between pairs of coexisting minerals have been used as criteria for chemical equilibrium (Kretz, 1959; Hounslo and Moore, 1967; Albee, 1968). The approach is most useful where the mineral pairs occur in a group of rocks which have different bulk compositions. Chemical data are limited to six samples in this study and no more than two of the six have a common mineral pair appropriate for calculating distribution coefficients.

The distribution relationships between different grains of a pair of minerals in a single rock are use-

ful. If chemical equilibrium was attained at the time of recrystallization, the points representing each pair of grains should cluster tightly on a distribution diagram if the composition of the sample volume is homogeneous. Figure 19 illustrates that this is true for all six rocks.

Detailed chemical analyses of the minerals in these samples demonstrate that the state of equilibrium indicated by textural and distribution coefficient criteria does not perfectly conform to the definition of chemical equilibrium. Garnet, plagioclase, and actinolite in several of the rocks exhibit chemical zoning. Zoned garnets are common in metamorphic rocks. Mineral assemblages containing garnets have been used in many areas to define metamorphic zones and isograds. The fact that the garnets were found to be zoned did not change the systematic relationships between mineral assemblages in these metamorphic terrains. Zoned garnets represent a deviation from chemical equilibrium, but not a serious one. One commonly used rationalization is the assumption that only the rim of the garnet is in equilibrium with the rest of the mineral assemblage. Another is to treat the garnet as an "extra" phase in a mineral assemblage because it contains significant amounts of an "extra" component such as MnO. Regardless of how the problem is dealt with, zoned garnets are common in rocks which in all other respects, both textural and chemical, conform

$$x_{Fe} (AC, DI) = \frac{Fe}{Fe + Mg + Mn}$$

$$x_{Fe} (BI, PT, GA, FD) = \frac{Fe}{Fe + Mg + Ca + Mn}$$

Figure 19: Distribution relationships in the six samples analyzed. Mineral symbols: AM - amphibole. All other symbols are as in Figures 15-18.

to the model of chemical equilibrium. In the course of this study, chemical zoning was found to occur in minerals other than garnet. Plagioclase in two of the aluminous iron-bearing calcareous rocks and in one of the pelitic samples, and actinolite in one of the calcareous rocks, are compositionally heterogeneous. In each case, the other minerals in these rocks are, within narrow limits, completely homogeneous. Deviations from perfect chemical equilibrium such as these are not surprising in systems as complex as natural rocks. The composition of even a single mineral grain is never precisely the same at every point for a variety of reasons. Trace amounts of impurities, crystallographic imperfections, instrumental factors and departure from equilibrium may all contribute to some degree to the variation. More detailed microprobe analyses of complete mineral assemblages will probably reveal more instances of zoned minerals. For the purpose of this study, however, the fact that the distribution coefficients for mineral pairs from each of the six samples plot in tight clusters on distribution diagrams is taken as a criterion that these rocks attained a state sufficiently close to chemical equilibrium that Duhem's theorem can be applied.

Pages 131 - 133 (including Table 10) omitted.

The hypothetical bulk composition

The problem dealt with in this study can be stated as follows: given two rocks, each composed of a different mineral assemblage, is the difference in mineralogy the result of differences in bulk composition or in physico-chemical conditions at the time of crystallization? Paraphrasing Duhem's Theorem (Chapter 13, Prigogine and Defay, 1954): given a system with a fixed composition under a particular set of intensive variables, there is only one set of phases, each with a fixed composition, which can coexist in the system at equilibrium. If the composition of two rocks is the same, any differences in mineralogy must be attributed to different physical conditions at the time of crystallization.

The position of a triangle defined by three minerals on an AFM projection is fixed for some set of values for temperature, pressure and fugacity of water (Thompson, 1957). If bulk composition is constant, changes in the topology of the projection can only occur

if the intensive parameters change. For example, a rock (A) composed of chlorite, garnet, and biotite (plus quartz and muscovite) is represented on an AFM projection in Figure 20a, and a rock (B) composed of staurolite, garnet, and biotite (quartz, muscovite) is represented in Figure 20b. Both rocks are equilibrium assemblages. The diagrams represent two different topologies. If the bulk composition of the rocks is the same, the rocks must have formed under different sets of intensive parameters. However, two rocks are not likely to have exactly the same composition.

An important characteristic of the triangles (Figure 20 a,b) is that a range of bulk compositions can be represented by the three minerals defining each triangle. By hypothetically varying the modal proportions of the minerals in a rock, one can change the bulk composition while the topology remains unmodified. That is, as long as all three minerals are present and of constant composition, the topology of the diagram can be seen not to change, and the physical conditions to be constant. For rocks A and B, if the modal compositions are hypothetically varied a set of bulk compositions is found which can be expressed as an arbitrary combination of either staurolite-garnet-biotite or chlorite-garnet-biotite. These hypothetical compositions are indicated by the shaded area in Figure 20c. Any one of these compositions is a closed system that can be represented by

Figure 20: Graphical representation of the hypothetical bulk composition problem. Mineral symbols same as in Figure 18.

two different mineral assemblages (two AFM topologies). This can only occur if the physical conditions at the time of formation were different for rock A than for rock B. The example illustrates a significant point. The bulk compositions of two rocks can be quite different (x and y, Figure 20c) yet the principles of the AFM projection with respect to physical conditions still apply.

Figures 20d and 20e present a case for which there are not hypothetical bulk compositions that can be represented by arbitrary combinations of minerals in the two rocks. If the chlorite in both rocks was identical, the point representing that mineral composition would be common to both rocks. However, that composition cannot be represented by a combination of all the three minerals in each rock. In this case, nothing can be said about the relative physical conditions at the time of formation of these rocks.

Duhem's theorem is a generalized statement of the principle demonstrated by the preceding examples. If there is an hypothetical bulk composition that can be expressed as an arbitrary combination of one set of minerals in rock A or a different set of minerals in rock B, and both rocks are equilibrium assemblages, the rocks must have formed under different sets of intensive parameters. The problem of finding the hypothetical bulk composition can be represented in a simplified way geometrically, but

a rigorous solution to the problem in real rocks can only be found mathematically.

Mathematical method

Linear Algebra

If there is an hypothetical bulk composition that can be expressed as some combination of the minerals in each of two rocks, A and B, the statement,

$$A - B = 0 \quad (1)$$

should be true. That is, a balanced chemical reaction can be written between all the minerals in rock A and those in rock B. Uncertainties in the analyzed compositions of the phases in each rock and departures from chemical equilibrium make the occurrence of a perfect balance between two natural rocks extremely unlikely. A more realistic statement would be that

$$-U \leq (A - B) \leq U \quad (2)$$

where U represents the uncertainties associated with rocks A and B. $A - B = 0$ is included in the range defined by expression (2).

If rock A contains m minerals and rock B contains n different minerals, and both rocks are composed of i components, then expression (2) can be expanded to take the form of Greenwood's inequalities (20) (Greenwood, 1967, p.474).

$$\left[\sum_{j=1}^m \nu_j \delta A_{ij} + \sum_{k=1}^n \nu_k \delta B_{ik} \right] \leq \left[\sum_{j=1}^m \nu_j A_{ij} - \sum_{k=1}^n \nu_k B_{ik} \right] \quad (3)$$

$$\left[\sum_{j=1}^m \nu_j A_{ij} - \sum_{k=1}^n \nu_k B_{ik} \right] \leq \left[\sum_{j=1}^m \nu_j \delta A_{ij} + \sum_{k=1}^n \nu_k \delta B_{ik} \right]$$

where, ν_j - stoichiometric coefficient for mineral j
 ν_k - stoichiometric coefficient for mineral k
 A_{ij} - amount of component i in mineral j in rock A
 B_{ik} - amount of component i in mineral k in rock B
 δA_{ij} - uncertainty in the amount of component i in mineral j in rock A
 δB_{ik} - uncertainty in the amount of component i in mineral k in rock B

and,

$$\nu_j \geq 0 \quad \nu_k \geq 0$$

$$\sum_{j=1}^m \nu_j = 1 \quad \sum_{k=1}^n \nu_k = 1$$

$j = 1, m$
 $k = 1, n$

The term between the inequality signs corresponds to A - B in expression (2) and the terms outside the inequality signs correspond to the ν terms.

In the approach used by Greenwood (1967) the U-terms in expression (3) were treated as constants rather than related to the variable A_{ij} 's and B_{ik} 's. To accomplish this the stoichiometric coefficients and uncertainty terms (δA_{ij} δB_{ik}) were replaced by the average uncertainty for component i in all the minerals. The result was that the same uncertainty was applied to the amount of component i in each mineral j regardless of

how large or small the amount was. A simple yet brilliant modification of Greenwood's inequalities (20, 21) (expression (3) this study) by T.M. Gordon (personal communication) permits the uncertainty for each component in each mineral to be used to obtain a solution.

In reference to expression (3):

by definition, for all values of i , j , and k ,

$$\nu_j \geq 0 \quad \nu_k \geq 0$$

$$\delta_{B_{ik}}, \delta_{A_{ij}} \geq 0$$

that is, inside the brackets, the uncertainty terms are positive. Because addition of equal numbers to both sides of an inequality does not change the sense of the inequality, expression (3) can be rearranged and terms collected.

$$\left[\left(\sum_{j=1}^m \nu_j A_{ij} + \sum_{j=1}^m \nu_j \delta A_{ij} \right) - \left(\sum_{k=1}^n \nu_k B_{ik} - \sum_{k=1}^n \nu_k \delta B_{ik} \right) \right] \geq 0 \quad (4)$$

$$\left[\left(\sum_{j=1}^m \nu_j A_{ij} - \sum_{j=1}^m \nu_j \delta A_{ij} \right) - \sum_{k=1}^n \nu_k B_{ik} + \sum_{k=1}^n \nu_k \delta B_{ik} \right] \leq 0 \quad (5)$$

or:

$$\sum_{j=1}^m \nu_j (A_{ij} + \delta A_{ij}) - \sum_{k=1}^n \nu_k (B_{ik} - \delta B_{ik}) \geq 0 \quad (6)$$

$$\sum_{j=1}^m \nu_j (A_{ij} - \delta A_{ij}) - \sum_{k=1}^n \nu_k (B_{ik} + \delta B_{ik}) \leq 0 \quad (7)$$

$$\text{or: } (\text{Rock A} + \text{uncertainty}) - (\text{Rock B} - \text{uncertainty}) \geq 0 \quad (8)$$

$$(\text{Rock A} - \text{uncertainty}) - (\text{Rock B} + \text{uncertainty}) \leq 0 \quad (9)$$

To illustrate expressions (6) and (7) an hypothetical example using two special rocks (A is composed of calcite and quartz, B is composed of wollastonite and an interstitial fluid phase) is presented below. The summations in the algebraic expressions have been expanded to form a matrix where the minerals are the columns and the constituents the rows. The following hypothetical chemical analyses and associated uncertainties were used:

	Calcite	Quartz	Wollastonite	Fluid
CaO	$54.40 \pm .50$	0.00	$51.75 \pm .25$	0.00
SiO ₂	0.00	$100.00 \pm .25$	$48.25 \pm .50$	0.00
CO ₂	$45.25 \pm .65$	0.00	0.00	$100.00 \pm .25$

In the composition matrix the letters P and M designate the instances where the uncertainty was added and subtracted respectively, from the associated oxide weight percent.

	Calcite	Quartz	Wollastonite	Fluid
CaO	$\nu_{CC} 55.00P + \nu_{QZ} 0.00P - \nu_{WO} 51.50M - \nu_{F1} 0.00M \geq 0$			
SiO ₂	$\nu_{CC} 0.00P + \nu_{QZ} 100.25P - \nu_{WO} 47.75M - \nu_{F1} 0.00M \geq 0$			
CO ₂	$\nu_{CC} 45.90P + \nu_{QZ} 0.00P - \nu_{WO} 0.00M - \nu_{F1} 100.35M \geq 0$			
CaO	$\nu_{CC} 54.00M + \nu_{QZ} 0.00M - \nu_{WO} 52.00P - \nu_{F1} 0.00P \leq 0$			
SiO ₂	$\nu_{CC} 0.00M + \nu_{QZ} 99.75M - \nu_{WO} 48.75P - \nu_{F1} 0.00P \leq 0$			
CO ₂	$\nu_{CC} 44.60M + \nu_{QZ} 0.00M - \nu_{WO} 0.00P - \nu_{F1} 99.65P \leq 0$			

This composition matrix together with,

$$v_{cc} + v_{qz} = 1; v_{wo} + v_{F1} = 1$$

$$\left. \begin{array}{l} v_j \geq 0 \quad v_k \geq 0 \\ \text{and } \sum B_{ik} \sum A_{ij} \geq 0 \end{array} \right\} \text{ for all values of } i, j \text{ and } k$$

are the constraints within which a solution must be found if a feasible one exists.

Linear programming is a mathematical method for solving systems of inequalities and equalities such as this.

Linear Programming

For a system of inequalities there is either an infinite number of solutions or there are none. There is no unique set of solutions for such a system as there would be if the minerals in rock A balanced perfectly those in rock B. The solution set to a system of inequalities in 2-space is illustrated in Figure 21.

- 1) $x \geq 0$
- 2) $y \leq -x + 8$
- 3) $y \leq 1/3x + 4$
- 4) $y \geq 0$
- 5) $y \geq 4/3x - 8$

Figure 21: Solution set in 2-space.

The shaded area includes all the solutions possible within the constraints listed to the right of the graph. Linear programming finds a subset of the set of solutions which satisfy the constraints by mapping out the extreme points of the shaded area. The procedure involves maximizing or minimizing an objective function such as the dashed line, $y = \frac{7}{11}x + 7$ on Figure 21. The value, $\frac{43}{11}$, obtained when the objective function intersects extreme point (5,3) is the maximum for that function given this particular set of solutions. The details of the linear programming method are found in Spivey (1963), Hadley (1962) and Gass (1958).

Linear programming is ideally suited to finding an hypothetical bulk composition common to a pair of metamorphic rocks because a virtually unlimited number of constituents and minerals can be considered simultaneously. Greenwood (1967) discussed how mobile or excess components can be incorporated into the system. For example, H_2O , CO_2 , and SiO_2 (for quartz-bearing rocks) are treated as minerals which are present in both rocks in amounts sufficient to balance any reaction between these rocks.

In the following section an attempt is made to construct a geometrical analogy of the application of linear programming to problems in metamorphic rocks.

To describe geometrically the problem stated above, the concept of a composition space and the repres-

entation of minerals as vectors in such a space need to be discussed. Each dimension in a composition space represents one constituent or component. For example, in a two-dimensional orthogonal composition space, the ordinate defines the proportion of one constituent and the abscissa the proportion of the second constituent. Minerals composed of these constituents plot as points in the 2-space.

Figure 22: Minerals as vectors in 2-space.

Those minerals composed entirely of CaO and SiO_2 in Figure 22 plot on the line representing unity (100 per cent); other minerals plot between this line and the origin of the space. Each point in the space corresponds to an ordered pair of numbers indicating its position in the composition space. The ordered pair of numbers associated with a mineral indicates the proportions of the constituents in the mineral. Each ordered

pair of numbers also designates a vector in the composition space, with its initial point at the origin. The concept of a mineral as a vector (Greenwood, 1967)

Figure 23 a) Mineral vector as the sum of two component vectors.
b) Mineral vector with associated uncertainty.

makes possible the graphical and algebraic representation of a mass balance between several minerals. For example, $\frac{1}{2}\text{lime} + \frac{1}{2}\text{quartz} = \text{wollastonite}$ is the same as $\frac{1}{2}(100,0) + \frac{1}{2}(0,100) = (50,50)$. The latter is illustrated in Figure 23a. The usefulness of this property of mineral vectors is particularly evident when dealing with the multidimensional composition space necessary to describe a natural rock. Figure 23b illustrates the common occurrence of a mineral which does not total 100 percent and in addition has uncertainties attached to the proportions of each component present in the mineral. In the examples used here, for purposes of illustration, all minerals will be assumed to total 100 percent without any uncertainty.

A three-dimensional composition space is analogous to the two-dimensional space. The 100 percent

line becomes a plane and each vector (mineral) is designated by an ordered triple of numbers.

Figure 24 a) Mineral vector in 3-space.

The vectors $(0,0,100)$, $(0,100,0)$, $(100,0,0)$ (Figure 24a) represent three minerals composed entirely of one of the three constituents defining the composition space. The 100 percent plane is similar to the staurolite-garnet-biotite triangle discussed previously, in that any point on the plane represents a linear combination of the three minerals defining the plane. In other words, any vector intersecting the plane can be expressed in terms of some combination of the vectors $(0,0,100)$, $(100,0,0)$ and $(0,100,0)$; for example,

$$\frac{1}{2}(100,0,0) + \frac{1}{2}(0,100,0) + \frac{1}{2}(0,0,100) = (25,25,50)$$

The vector $(25,25,50)$ can be thought of as the bulk composition of a rock composed of 25% X, 25% Y, and 50% Z.

Although the geometrical representation of vectors, 100 percent planes and composition spaces is limit-

ed to three dimensions, the algebraic representation of these elements can be carried into as many dimensions as required. In a n -dimensional composition space a vector would be designated by an ordered n -tuple of numbers (50,25,30,60..... n) and the 100 percent element would be a hyperplane with $n-1$ dimensions. Vector addition, scalar multiplication and other operations applicable to vectors in 2- or 3-space function in the same way in n -space. For rocks, a composition space with 11 or 12 dimensions would be appropriate, but for the purpose of a geometric analogy of the linear programming method, three dimensions is sufficient.

Figure 24b Rock A and Rock B on the 100 percent plane in 3-space.

In Figure 24b vectors a, b, c, d, e, and f represent six minerals which lie on the 100 percent plane in a three-dimensional orthogonal composition space. Minerals a, b, c compose rock A and d, e, f, rock B. The bulk composition of these rocks is not known. The triangular subspace of the 100 percent plane bounded by vectors a, b, and c represents all linear combinations of positive amounts of those vectors with the restriction that the resultant vector intersects the 100 percent plane. Similarly, the area d-e-f is all linear combinations of positive amounts of the vectors d, e, and f which sum to 100 percent. Once again, any vector intersecting the 100 percent plane within the boundaries of either of these subspaces can be represented by a linear combination of the three minerals defining the subspace. Any vector which occurs in the portion of overlap between the two triangles a-b-c and d-e-f can be represented as a linear combination of either a, b, and c, or d, e, and f. In this example, a graphical approach revealed a set of vectors (bulk compositions) which can be expressed as a combination of either the minerals in rock A or the minerals in rock B. From Duhem's theorem, if a-b-c and d-e-f are equilibrium assemblages they must have crystallized under different sets of intensive parameters.

Within the area of overlap on the 100 percent plane in Figure 24b there are an infinite number of sol-

utions. All that can be said is that there is an intersection between that part of the composition space defined by the minerals in rock A and the part defined by the minerals in rock B. Linear programming provides a method of determining the extreme points of a solution set similar to that in Figure 24b in a multidimensional composition space.

In order to illustrate graphically the method by which linear programming maps out the extreme points of a solution set in multidimensional space it is necessary to make a step which does not occur in the actual mathematical method. Even though the analogy is no longer a rigorous one, it is a useful approximation which aids in understanding the linear programming method.

Figure 25 presents the 100 percent plane from Figure 24b superimposed arbitrarily on an orthogonal grid which does not have compositional coordinates. There are no uncertainties in the compositions of each mineral.

Figure 25: The 100 percent plane from Figure 24b superimposed on an arbitrary grid.

The purpose of this step is to simplify the graphical description of the inequalities defining the solution set and illustrate the maximization of the objective function. Note the similarities between this figure and the linear programming solution in 2-space. Normally the objective function maximized when dealing with metamorphic rocks is the stoichiometric coefficients of the minerals in the reaction equating the two rocks under study. In this case, since there is a group of hypothetical bulk compositions which can be expressed as either rock A or rock B, the expression

$$\nu_a a + \nu_b b + \nu_c c = \nu_d d + \nu_e e + \nu_f f$$

can be solved for some set of stoichiometric coefficients. By maximizing ν_j for each j in turn, the extreme points of the solution set can be found. ν_j can be represented on the graph in Figure 25 as a linear function parallel to the side of the triangle opposite the apex j . For example, ν_a can be represented by the function $y = \frac{x}{24} + \frac{19}{4}$. By sweeping across the solution set the linear function finds the extreme point (4,5) on the arbitrary orthogonal grid which maximizes the function ν_a . This point corresponds to the solution

$$.30a + .25b + .45c = 1.00d + 0.00e + 0.00f$$

By maximizing each stoichiometric coefficient in turn, all the extreme points of the solution set are determined, thereby defining the limits of the range of bulk compo-

sitions which could be expressed as either a-b-c or d-e-f under a particular set of physical conditions.

For real rocks, a precise balance is not likely to be obtained, nor is the composition space likely to have only three dimensions. The linear programming method used in this study must map out the extreme points of a solution set in a multidimensional composition space. Furthermore, the boundaries of the solution set are not as precisely defined as those in Figure 25 because there is some degree of uncertainty attached to the compositions of minerals in rocks.

Application of Linear Programming

In the Bishop Corners-Madoc area three rock units of the Flinton Group: biotite-carbonate schist, calcareous psammite, and pelitic schist, exhibit mineralogical variation along strike (Figures 15, 16, 17, and 18). From each unit a pair of samples was collected so as to include as much mineralogical variation between the two sample localities as possible. Rocks 1 and 2 occur in Zones I and IV, respectively, in the biotite carbonate schist and its metamorphic equivalent (Figure 15). Rocks 3 and 4 are from Zones I and IV in the calcareous psammite and its metamorphic equivalent (Figure 16). Rocks 5 and 6 were chosen as representative of the muscovite-bearing pelitic schist in Zones I and II (Figure 17).

It follows from Duhem's theorem that if there is an hypothetical bulk composition that can be expressed as an arbitrary combination of either the minerals in rock A or of those in rock B and both rocks are equilibrium assemblages, any mineralogical differences must be attributed to differences in physico-chemical conditions at the time the rocks were formed. Furthermore, if there is such an hypothetical composition, a balanced chemical reaction can be written between the minerals in rock A and those in rock B. The balance is not a perfect one for real rocks because of uncertainties in the compositions of the minerals and deviations from chemical equilibrium. The "reaction" is balanced within the limits of these uncertainties. Using linear programming, an attempt has been made to find hypothetical bulk compositions for each of the pairs, 1-2, 3-4 and 5-6. All six samples have been demonstrated to approximate equilibrium assemblages. If hypothetical compositions are found for each pair, differences in mineralogy can be attributed to different physico-chemical conditions at the time of recrystallization, and the mineral zoning between the sample localities must be the result of these differences rather than variable bulk composition.

Rocks A and B can be expanded into composition matrices with oxide constituents as rows and the minerals in A and B as columns similar to the example given in the theory section of this chapter, except for the fact that

the six rocks used in this study are composed of 9 - 10 oxide constituents and each pair includes 14 - 16 minerals. For example, even without expanding each mineral into its constituents, inequality (8) for the pair of rocks (1 and 2) from the carbonate schist unit would be the following lengthy expression:

$$\begin{aligned} & \nu_{QZ1} QZ1 + \nu_{BI1} BI1 + \nu_{PA1} PA1 + \nu_{FD1} FD1 + \nu_{CC1} CC1 + \\ & \nu_{RU1} RU1 - \nu_{QZ2} QZ2 - \nu_{DI2} DI2 - \nu_{CC2} CC2 - \nu_{PA2} PA2 - \\ & \nu_{MI2} MI2 - \nu_{AC2} AC2 - \nu_{SP2} SP2 - \nu_{BI2} BI2 \geq 0 \end{aligned}$$

For this example, the uncertainties are included in the symbol identifying each mineral.

Two inequalities of the form of (8) and (9) together with the limitations that the stoichiometric coefficients and uncertainties be numbers greater than or equal to zero, and that the stoichiometric coefficients for the minerals in each rock sum to unity, are the constraints which must be satisfied simultaneously for a set of feasible solutions to be obtained. For systems of inequalities, there are no unique solutions; there is either no solution (infeasible) or there is an infinite number of feasible solutions within the limits of the solution set defined by the inequalities. In this study, the set of feasible solutions is a set of hypothetical bulk compositions which can be represented as either the minerals in Rock A or the minerals in rock B. The solu-

tion set can be thought of as an intersection in composition space between the subspace defined by the minerals in rock A and the subspace defined by the minerals in rock B. Linear programming provides a subset of the feasible set by maximizing some objective function. The optimal solutions obtained this way are equivalent to the extreme points of a solution set in two-space (Figure 21). For the problem of the hypothetical bulk composition in metamorphic rocks, the stoichiometric coefficient of each mineral in rocks A and B is maximized in turn. In each case, an hypothetical chemical reaction relating the minerals of rock A to those in rock B is obtained in which one of the stoichiometric coefficients is a maximum for that particular set of feasible solutions. The "reaction" is balanced within the limits of the uncertainties in the chemical data.

Although the significance of linear programming solutions will be discussed in more detail later in this section, the introduction of the term ΔX_k (Greenwood's ΔX_k ; 1967, p.479) at this point is pertinent. ΔX_k represents the uncertainty associated with the amount of mineral k which occurs in any of the solutions in the solution set. The term is calculated from the weighted averages of the uncertainties associated with each of the oxide constituents of mineral k.

$$\Delta v_k = \sum_i^j \frac{U_i \cdot P_i}{a_k} \quad (10)$$

(after Greenwood, 1967)

1...j = constituents in mineral k

U_i = uncertainty of constituent i in mineral k

P_i = weight percent constituent i in mineral k

$$a_k = \sqrt{\sum_i P_i^2}$$

If the stoichiometric coefficients (v_k) for mineral k in the solution set are less than Δv_k , the coefficients are indistinguishable from zero. If most of the v_k 's for the minerals in the solution set are indistinguishable from zero, the intersection can not be considered significant.

Chemical data for the three pairs of rocks examined in this study were transformed into composition matrices of the type described above, and a linear programming package³ was used to perform the operations necessary to test for solutions to the problem. A set of feasible solutions was found for each of the three pairs of rocks. In all three cases H_2O , CO_2 and quartz (SiO_2) were treated as "excess minerals". That is, all three species were assumed to have been present in both rocks at the time of metamorphism in amounts sufficient to balance any requirements for hypothetical chemical reactions (mass balances) between the minerals of Rock 1

³ IBM Mathematical Programming System Extended (Systems Dimensions Ltd.).

Table 11 - Optimal solution set of stoichiometric coefficients for Rocks 1 and 2

Objective Function	QZ1	BI1	PA1	FD1	CC1	RU1	QZ2	DI2	CC2	PA2	MI2	AC2	SP2	BI2
Δ BI1	.0723	.9128*	-	-	.0149	-	-	-	-	-	.1006	.0665	-	.8329
Δ PA1	.0119	.1913	.7888*	-	.0079	-	-	-	-	.7727	.0231	.0100	-	.1910
Δ FD1	.0174	.2911	.6611	.0304*	-	-	-	-	.0073	.6344	.0189	.0216	.0015	.3163
Δ CC1	.1097	.5866	.0021	-	.3016*	-	-	-	.2652	-	.1227	.1440	.0119	.4562
Δ RU1	.0495	.4549	.4497	-	.0323	.0137*	-	-	-	.4273	.0557	.0406	.0421	.4343
Δ DI2	.0576	.4819	.3412	-	.1194	-	-	.0639*	.0867	.3287	.0604	-	.0060	.4524
Δ CC2	.1035	.5939	.0016	-	.3010	-	-	-	.2721*	-	.1186	.1372	.0002	.4719
Δ PA2	.0119	.1913	.7888	-	.0079	-	-	-	-	.7728*	.0263	.0100	-	.1910
Δ MI2	.1414	.8092	.0017	-	.0476	-	-	-	-	-	.1597*	.1784	.0156	.6463
Δ AC2	.1414	.0920	.0017	-	.0476	-	-	-	-	-	.1597	.1784*	.0156	.6463
Δ SP2	.0588	.6499	.2475	-	.0306	.0132	-	-	-	.2261	.0784	.0386	.0433*	.6137
Δ BI2	.0723	.9128	-	-	.0149	-	-	-	-	-	.1066	.0665	-	.8329*
Δ Δk	.0000	.0066	.0052	.0052	.0047	.0070	.0000	.0043	.0031	.0042	.0041	.0059	.0075	.0053

QZ - quartz
 BI - biotite
 PA - plagioclase (average)
 FD - Fe-dolomite
 CC - calcite

DI - diopside
 MI - microcline
 AC - actinolite
 SP - sphene
 RU - rutile

The number associated with each mineral symbol refers to the rock in which the mineral occurs.

* stoichiometric coefficient maximized

and those of Rock 2. The stoichiometric coefficients for H_2O and CO_2 were not included in the constraint $\sum_{j=1}^n v_j = \sum_{k=1}^m v_k = 1.0$, for all values of j and k , for either Rock 1 or Rock 2. All three species were assumed to be 100 percent pure.

Biotite carbonate schist (Rocks 1 and 2). The initial attempt to obtain an intersection between Rocks 1 and 2 was not successful. Analyses for MnO had only been done for the carbonates in these rocks. The MnO in calcite in Rock 2 was not sufficient to balance that present in calcite and ferroan dolomite in Rock 1. A feasible solution set could only be obtained if the constituent MnO was removed from the composition matrix. In some cases (see solution for Rocks 5 and 6), the linear programming package obtains a feasible solution by omitting the phase or phases containing the constituent which cannot be balanced. The minerals containing MnO in Rock 1 could not be omitted from the solution set for the pair 1-2 because the CaO in the calc-silicate minerals and calcite in Rock 2 could not be balanced. Actinolite and diopside were analyzed for MnO and with this small addition of MnO to Rock 2, a set of feasible solutions was obtained. The subset of optimal solutions resulting from maximizing the stoichiometric coefficients of each mineral in the two rocks is presented in Table 11.

Each row of the table is the hypothetical chemical reaction (mass balance) obtained when the objective function, the stoichiometric coefficient of one of the minerals, is maximized. In order for this set of "reactions"

or solutions to be written, there must be a set of hypothetical bulk compositions that can be expressed as some combination of either the minerals in Rock 1 or those in Rock 2. The solution for maximum $\sum_{BI1}$ is an interesting example:

The balance is not perfect because of the uncertainties in chemical data. The similarity (except for BI2) between this "reaction" and Carmichael's (1970) reaction (5), biotite + calcite + quartz = Ca-amphibole + K-feldspar + $CO_2 + H_2O$, is coincidental. Although any one of the hypothetical reactions may have occurred during the metamorphic history of Rock 2, the fact that these "reactions" are part of the solution set does not mean they actually did occur.

Most of the stoichiometric coefficients in the solution set, and in particular all maximum $\sum_k$'s, are much larger (3 - 150X) than the $\Delta \sum_k$ term. From Table 11 it is evident that minerals which are common to both rocks tend to have the largest stoichiometric coefficients. If the hypothetical bulk composition is to be common to both rocks, the solution will tend to include large amounts of minerals common to both rocks even if the minerals differ somewhat in composition. This tendency occurs because the portion of the composition space included in the common mineral is common to both rocks even if no other portion of the composition space is.

This problem, with respect to the significance of intersections, will be dealt with later in this section.

For the purpose of this study the main point to be made from Table 11 is that there is a set of hypothetical bulk compositions which can be expressed as an arbitrary combination of either the minerals in the biotite-carbonate schist or the minerals in the epidote-free calc-silicate paragneiss. Therefore, the contrast in mineral assemblages, and thus the mineral zoning which occurs between the two sample localities, is the result of a gradient in physico-chemical conditions at the time of recrystallization.

Calcareous quartzofeldspathic psammite (Rocks 3 and 4).

Rock 3, calcareous psammite, and Rock 4, epidote-bearing calc-silicate paragneiss, contain zoned plagioclase. The actinolite in Rock 4 is heterogeneous as well. For the first attempt to find a solution set for this pair of rocks "average" plagioclase compositions were included in the composition matrix. For the second attempt, the most sodic plagioclase composition in Rock 3 and the most calcic plagioclase in Rock 4 were used. In both instances, the colourless and blue-green amphiboles (Rock 4) were included as two distinct minerals. The attempt to find a solution with two amphiboles is thought to be a more severe test for a solution set than if an average amphibole composition was used.

The optimal solution set for the first run with

average plagioclase compositions in both rocks is presented in Table 12. Although the minerals common to both rocks, plagioclase and calcite, have the largest stoichiometric coefficients, most of the others are also significantly larger than the Δv_k terms. Certainly the maximized coefficients for each mineral are all larger (3 - 150 times as large) than Δv_k . The coefficients for SP2 are all relatively small. Apparently very little sphene is required to balance the titanium present in biotite and Fe-Ti oxide in Rock 1.

The second attempt using plagioclases with extreme compositions was also successful. Table 13 illustrates the changes which occurred in the v_k 's as a result of changing the plagioclase compositions. The solution sets for both runs are summarized by listing only the maximum stoichiometric coefficients for each phase. None of the coefficients were reduced to values approaching the Δv_k 's. However, every one was reduced, some more than others. v_{CC1} and v_{CC2} changed the least amount, .001 and .002 respectively. The largest reductions occurred in the maximum coefficients for AC4 and HN4. On the basis that the closer the maximized coefficients approach zero, the less significant is the intersection (see below), the use of extreme plagioclase compositions reduced the significance of the intersection between Rocks 3 and 4, but the intersection is still

Table 12 - Optimal solution set of stoichiometric coefficients
for Rocks 3 and 4 (using "average" plagioclases).

Objective Function	QZ3	PT3	PA3	CC3	FD3	BI3	OX3	QZ4	MI4	PA4	EP4	CC4	DI4	HN4	AC4	SP4
↓MU3	.1218	.3434*	.4214	.0288	.1830	.0234	-	-	.2211	.4440	.0821	-	-	.2407	-	.0121
↓PA3	.0403	.1977	.7051*	.0513	.0380	.0079	-	-	.1243	.7854	-	-	-	.0857	-	.0047
↓CC3	.0405	.0007	.0039	.9954*	-	-	-	-	.0046	-	.0000	.9166	.0829	-	-	-
↓FD3	.1235	.1472	.3230	.2561	.2685*	-	.0053	-	.0928	.3253	.0181	.2941	-	-	.2664	.0033
↓BI3	.0599	.1134	.2710	.5178	-	.0975*	.0003	-	.1296	.2852	-	.4324	-	.1473	-	.0055
↓OX3	.0991	.1802	.0224	.6685	.1051	-	.0238*	-	.1194	-	.0848	.6015	-	.1858	-	.0085
↓MI4	.1218	.3434	.4214	.0288	.1830	-	.0234	-	.2211*	.4440	.0821	-	-	.2407	-	.0121
↓PA4	.0403	.1977	.7051	.0513	.0380	.0079	-	-	.1243	.7854*	-	-	-	.0857	-	.0047
↓EP4	.0991	.1802	.0224	.6685	.1051	-	.0238	-	.1194	-	.0848*	.6015	-	.1858	-	.0085
↓CC4	.0261	.0027	.0082	.9891	-	-	-	-	.0018	-	.0000	.9377*	-	-	.0605	-
↓DI4	.1660	.2152	.4709	.0666	.2392	-	.0081	-	.1352	.5098	.0278	-	.3209*	-	-	.0063
↓HN4	.1132	.3194	.4609	.0149	.1891	-	.0156	-	.2046	.4880	.0542	-	-	.2435*	-	.0097
↓AC4	.1235	.1472	.3230	.2561	.2685	-	.0053	-	.0928	.3253	.0181	.2941	-	-	.2664*	.0033
↓SP4	.1175	.3379	.4339	.0318	.1746	-	.0218	-	.2172	.4595	.0763	-	-	.2312	-	.0158*
ΔJK	.0000	.0050	.0046	.0048	.0050	.0045	.0075	.0000	.0058	.0087	.0045	.0056	.0047	.0064	.0066	.0061

PT - phengite
PS - sodic plagioclase
EP - epidote
PC - calcic plagioclase
OX - Fe-Ti oxide
HN - hornblende

All other symbols same as in Table 11

Table 13: Comparison of the maximized $\mathcal{U}_K$'s from the optimal solution sets for Rocks 3 and 4 using average plagioclase compositions (PA3, PA4) with those obtained for 3 and 4 using extreme plagioclase compositions (PS3, PC4).

$\mathcal{U}$ maximized	$\mathcal{U}_K(\text{PA3, PA4})$	$\mathcal{U}_K(\text{PS3, PC4})$	$\Delta \mathcal{U}_K$
PT3	.3434	.3378	.0050
PA3	.7051	-	.0046
PS3	-	.4834	
CC3	.9954	.9953	.0048
FD3	.2685	.1928	.0050
BI3	.0975	.0858	.0045
OX3	.0238	.0236	.0075
MI4	.2211	.2148	.0058
PA4	.7954		.0087
PC4	-	.5907	
EP4	.0848	.0841	.0045
CC4	.9377	.9375	.0056
DI4	.3209	.2411	.0047
HN4	.2435	.1844	.0064
AC4	.2664	.2201	.0066
SP4	.0158	.0108	.0061

a valid one.

For Rocks 3 and 4, the calcareous psammite and epidote-bearing paragneiss, there is a set of hypothetical bulk compositions that can be expressed as arbitrary combinations of either the minerals in Rock 3 or those in Rock 4. Therefore, the mineral zoning in the psammite-paragneiss unit between the sample localities is the result of different physico-chemical conditions at the time of metamorphism.

Calcareous psammite(3) and epidote-bearing paragneiss (2). A major difference between the schist-paragneiss unit (Rocks 1 and 2) and the psammite-paragneiss unit (Rocks 3 and 4) is the presence of Fe - bearing minerals such as epidote and magnetite in the latter. The iron content of epidote and a proportion of that in magnetite can be assumed to be ferric iron. The amount of ferric iron present in the minerals in the schist-paragneiss unit cannot be assumed with the same degree of certainty. In this respect at least, the two rock units are compositionally different. Furthermore, the presence of the pairs phengite-biotite and hornblende-epidote in the psammite-paragneiss unit shows that these rocks are more aluminous than the schist-paragneiss unit which contains only one mica and an aluminum-poor amphibole, actinolite. The greater proportion of iron in Fe-dolomite in the schist and the rela-

tively high colour index, 25 - 40 compared to 2 - 5 for the psammite, indicate the schist probably has a higher total iron content than the psammite. An attempt was made to find an intersection between the calcareous psammite (Rock 3) using an average plagioclase composition and the epidote-free calc-silicate paragneiss (Rock 2). To the surprise of all concerned, a set of feasible solutions was found.

The optimal solution set of stoichiometric coefficients is included as Table 14. The set of solutions for the pair 3-2 differs considerably from those for pairs 1-2 and 3-4. The coefficients generally are much lower. The maximized coefficients for PA3 and OX3 are zero, and the coefficient for SP2 is less than $\Delta\psi_{\text{sphene}}$. With respect to minerals common to both rocks, for biotite and calcite the coefficients are high, but they are very low or zero for both plagioclases. Clearly the plagioclases in Rocks 3 and 2 are different minerals. The ψ_k 's for CC3, FD3, BI3, DI2, CC2, MI2, AC2, and BI2 are consistently larger (5 - 20X) than the $\Delta\psi_k$ terms for these minerals. For example, maximizing ψ_{DI2} and multiplying each coefficient by 100:

$$2.55\text{CC3} + 4.65\text{FD3} + 82.80\text{BI2} \approx 10.00\text{DI2} + 9.60\text{CC2} + 1.21\text{PA2} + 4.33\text{MI2} + 71.86\text{BI2} (\pm \text{QZ}, \text{H}_2\text{O}, \text{CO}_2)$$

This hypothetical mass balance and others like it in the solution set demonstrate that a mass balance can be found

Table 14 - Optimal solution set of stoichiometric coefficients for Rocks 3 and 2 (using PA3).

Objective Function	PT3	PA3	CC3	FD3	BI3	OX3	DI2	CC2	PA2	MI2	AC2	SP2	BI2
✓ PT3	.0289*	-	.0565	-	.9146	-	-	.0493	.0150	.0680	.0106	-	.8572
✓ PA3	-	-*	-	-	-	-	-	-	-	-	-	-	-
✓ CC3	-	-	.5252*	-	.4748	-	.0934	.4484	.0049	.0509	-	-	.4024
✓ FD3	-	-	.0155	.0570*	.9275	-	.0976	-	.0116	.0434	-	-	.8474
✓ BI3	-	-	.0004	.0047	.9949*	-	-	-	-	.0619	.0270	-	.9111
✓ OX3	-	-	-	-	-	*	-	-	-	-	-	-	-
✓ DI2	-	-	.1255	.0465	.8280	-	.1000*	.0960	.0121	.0433	-	-	.7486
✓ CC2	-	-	.5252	-	.4748	-	.0934	.4484*	.0049	.0509	-	-	.4024
✓ PA2	.0179	-	.0068	-	.9754	-	.0180	-	.0162*	.0600	-	-	.9058
✓ MI2	.0118	-	.0044	-	.9838	-	-	.0007	-	.0732*	.0171	-	.9091
✓ AC2	-	-	.2805	.0212	.6983	-	-	.2544	.0078	.0507	.0801*	-	.6070
✓ SP2	-	-	.4393	-	.5607	-	-	.3920	.0052	.0529	.0746	.0019*	.4734
✓ BI2	-	-	.0004	.0047	.9949	-	-	-	-	.0619	.0270	-	.9110*
Δ OK	.0050	.0046	.0048	.0050	.0045	.0075	.0043	.0031	.0042	.0041	.0059	.0075	.0053

* stoichiometric coefficient maximized
mineral symbols same as for Tables 11 and 12

between a part of Rock 3 and a part of Rock 2. Despite the imposed restriction limiting Fe^{3+} to Rock 3 and other compositional differences, an intersection was found. Although an intersection which does not make use of all the minerals in both rocks and for which many of the coefficients are zero, is less significant than one which uses all the minerals, the intersection is still a valid one. The optimal solution set for Rocks 3 and 2 means that the differences in the two mineral assemblages must be, at least in part, the result of different physical conditions at the time of metamorphism.

Pelitic schist (Rocks 5 and 6). By substituting composition data from oxide minerals in other rocks for the Fe-Ti oxides in Rock 6, and using the average composition of plagioclase and garnet in 6 and garnet in 5, an attempt was made to find an intersection with Rock 5. As in the case of Rocks 3 and 2, one sample contains minerals composed of constituents not present in the other. Plagioclase (PA6), Fe-Ti oxide (O16) and magnetite (O26), do not occur in Rock 5. Ilmenite, chlorite, biotite, muscovite, and quartz are prominent in both rocks. The differences in mineral assemblage result in large imbalances in the constituents CaO and Fe_2O_3 . The fact that a mineral is not present in both members of a pair of rocks is not necessarily important but if the components of that mineral are not included in some other min-

eral in the second rock, the absence prohibits a solution involving all the minerals of both rocks. The pair of rocks was subjected to the linear programming method and a feasible set of solutions was obtained (Table 15). This time the maximized stoichiometric coefficients for PA6, Ol6, and O26 are zero and those for GA6 and GA5 are only one-third the Δv_k terms for these minerals. The reasonable solutions involving MU5, CH5, CT5, OX5, BI6, KY6, and OX6 indicate that, as is the case of the intersection for the pair 3-2, only parts of Rocks 5 and 6 are involved. Because CaO and Fe_2O_3 could not be balanced with mineral compositions in Rock 5, PA6, Ol6, and O26 were omitted from the solution set. Even though some of the mineralogical differences between Rocks 5 and 6 are the result of bulk composition, the fact that an intersection was found demonstrates that the physical conditions at the time the rocks recrystallized were different. It is interesting to note that the minerals involved in the solution set are the ones chosen on the basis of petrographic study to define mineral zoning in the pelitic rocks.

Significance of Intersections

The solution sets presented in this chapter indicate that various types or degrees of intersections are possible. The first attempt with Rocks 1 and 2 was infeasible because MnO could not be balanced. With ad-

Table 15 - Optimal solution set - Rocks 5 and 6
PA6, GA6, GA5 - stoichiometric coefficients

Obj. Funct.	MU5	CH5	GA5	CT5	BI5	OX5	MU6	BI6	PA6	CH6	ST6	KY6	GA6	OI6	O26	O36
↓ MU5	.7137*	-	-	.0658	-	.2205	.6568	-	-	-	-	.0874	-	-	-	.2548
↓ CH5	.1387	.1952*	-	.0494	-	.6168	-	.1506	-	-	-	.1397	-	-	-	.7094
↓ GA5	.0528	.0374	.0022*	.1703	-	.7373	-	.0499	-	-	-	.1352	-	-	-	.8129
↓ CT5	.12409	-	-	.1977*	-	.6782	.0948	.0196	-	-	.0272	.1109	-	-	-	.7476
↓ BI5	.0017	.0452	-	.0988	.1151*	.7393	-	.1057	-	-	-	.0772	-	-	-	.8166
↓ OX5	.0316	.1255	-	.0953	-	.7475*	-	.02976	-	.0519	-	.0931	-	-	-	.8249
↓ MU6	.7073	-	-	.0689	-	.2239	.6887*	-	-	-	-	.0528	-	-	-	.25849
↓ BI6	.1485	.1888	-	.0571	-	.6055	-	.1613*	-	-	-	.1422	-	-	-	.6959
↓ PA6	-	-	-	-	-	-	-	-	*	-	-	-	-	-	-	-
↓ CH6	.0316	.1255	-	.0953	-	.7475	-	.0298	-	.05192*	-	.0931	-	-	-	.8249
↓ ST6	.0414	.0241	-	.1959	-	.7386	-	.0389	-	-	.0273*	.1192	-	-	-	.8144
↓ KY6	.1219	.1184	-	.1299	-	.62981	-	.1171	-	-	-	.1578*	-	-	-	.7248
↓ GA6	.0530	.0377	-	.1710	-	.7368	-	.05009	-	-	-	.1356	.0020*	-	-	.8123
↓ OI6	-	-	-	-	-	-	-	-	-	-	-	-	-	*	-	-
↓ O26	-	-	-	-	-	-	-	-	-	-	-	-	-	-	*	-
↓ O36	.0316	.1255	-	.0953	-	.7475	-	.02976	-	.0519	-	.0931	-	-	-	.8249
ΔΔK	.0075	.0059	.0055	.0059	.0065	.0083	.0043	.0053	.0114	.0053	.0055	.0045	.0054	.0075	.0066	.0070

* stoichiometric coefficient maximized

ditional data, 0.22% and 0.25% MnO in actinolite and diopside respectively, a set of optimal solutions was obtained. A change from "average" plagioclase compositions in Rocks 3 and 4 to a sodic composition in one rock and a calcic composition in the other reduced most of the maximized stoichiometric coefficients in the solution set, some as much as 30 percent. The intersections between the pairs 3-2 and 5-6 involved only parts of each rock. An optimal solution is possible which involves only the minerals common to both rocks with the other stoichiometric coefficients indistinguishable from zero. The need for a measure of the significance of an intersection is apparent.

In a system of inequalities, for example an optimal solution set, a coefficient may have a value greater than zero but because it is small enough to be within the uncertainties in the system it is indistinguishable from zero. Greenwood (1967) defined the term Δv_k , the uncertainty associated with the amount of mineral k used in any hypothetical mass balance, to deal with this problem. A stoichiometric coefficient with a value less than Δv_k is indistinguishable from zero. If the v_k 's for a particular solution set are equal to or less than the Δv_k 's for each mineral, the solution set is not a significant one.

To account for uncertainties in mineral analyses, Greenwood (1967) used a constant error vector for

each oxide constituent in all the minerals in the rocks he worked with. With this approach it was possible to change the error vector by a fixed amount and note the effects on the stoichiometric coefficients. Greenwood concluded that if the proportional change in the coefficients was much less than that introduced in the error vector, the intersection was significant. In this study, the uncertainties for each constituent in each mineral are known and used throughout the mathematical analysis. Although these uncertainties could be reduced systematically, each composition matrix for each pair of rocks would have required major alterations. The procedure was not considered practical for the purpose of this study.

The magnitude of the uncertainties in chemical data is important to the linear programming method. There is the possibility that an intersection between two rocks is not a true intersection but only an intersection within the range of the uncertainties associated with each rock. The smaller the uncertainties, the better is the possibility that the intersection is a valid one. If reducing the uncertainties in the problem causes the set of solutions to become infeasible, the feasible solution set is not significant. In this study, uncertainties were kept as small as possible by keeping the uncertainties in composition within one grain of a mineral (U) separated from those caused by grain to grain variation.

Only the uncertainty U was attached to the chemical data used in the linear programming analysis.

The uncertainties used in this study are sufficiently small that an intersection can be considered significant providing the stoichiometric coefficients for at least one mineral, not present in both rocks, is greater than the Δv_k terms for those criteria. From a comparison of the magnitudes of the coefficients in the optimal solution sets it is evident that the intersections between samples from the same rock unit are more significant than the intersection between two rocks from different rock units. The minerals composing the solution set for the pelitic rocks are the ones chosen to define the mineral zoning in these rocks on the basis of petrographic study.

Metamorphic Grade

Estimates of metamorphic grade are based on the mineral zoning observed in rocks which occur in regionally metamorphosed terrains and contact aureoles. The most important variables in the development of mineral zoning are temperature, pressure, and bulk composition. Metamorphic grade can only be defined in terms of changes in physical conditions (temperature and pressure) at the time of metamorphism. The distinction of mineral zoning induced by variation in bulk composition from that caused by a gradient in physical conditions is a funda-

mental petrologic problem.

Mineral zoning in the Bishop Corners-Madoc area occurs in Fe-bearing, aluminous, potassic, calcareous rocks of the Lessard Formation and in pelitic schist of the Bishop Corners Formation (Figures 15 - 18). These rocks contain more minerals and constituents than can be dealt with realistically by the conventional graphical approach. On the basis of significant intersections found by the linear programming method the mineral zoning is attributed to a gradient in physical conditions at the time of metamorphism. Bulk composition was not an important variable with respect to the development of mineral zoning in the pelitic schist or in the calcareous rocks of the Lessard Formation.

A gradient in physical conditions during metamorphism is indicated by the calcareous rocks and the pelitic rocks in the study area, but which way is "up"? In which direction does metamorphic grade increase? The answer to this question is dependent to some degree on the assumptions made about the fluid phase in the rock during metamorphism.

For the purpose of applying linear programming to metamorphic rocks in the study area, several assumptions were made about the enigmatic fluid phase. A fluid phase is assumed to be present in the rock in excess at the time of metamorphism. H_2O and CO_2 are considered to be the major constituents of the fluid. The

composition of the fluid phase is assumed to be buffered by the solid phases in the rock.

The metasedimentary rocks considered here occupy narrow isoclinal synclines (0.25 - 1.5 km perpendicular to the strike of the axial surface) which are surrounded by masses of relatively anhydrous, relatively impermeable mafic metavolcanic and granodioritic plutonic rocks. Mineralogical changes requiring the evolution of large amounts of water have not occurred in the quartzofeldspathic rocks adjacent to the calcareous or pelitic rocks in the areas where these rocks exhibit the most striking mineralogical variation. Accordingly, H_2O and CO_2 were derived mainly from the calcareous and pelitic rocks themselves. The existence of a large scale regional reservoir of fluid that controls the composition of the fluid phase is not likely in this setting. This evidence, though qualitative, supports the contention that the fluid phase composition was buffered by the mineral assemblages in each rock unit. Therefore, regional variation in composition of the fluid phase is not considered responsible for the observed mineral zoning.

An increase in metamorphic grade in the context of the statements made at the beginning of this section infers an increase in temperature. In terms of the range of conditions attributed to all metamorphic processes, pressure is an important variable, but on the scale of

the study area temperature is the most significant parameter. From experimental and field data on the stabilities of minerals and mineral assemblages, the direction in which temperature increased during metamorphism can be inferred from the mineral assemblages in the study area. The assemblage staurolite-biotite is known to be stable at higher temperatures than chloritoid-chlorite (Albee, 1965; Winkler, 1970). Pelitic rocks in Zone I in the study area (Figure 17) containing chloritoid and Fe-chlorite are interpreted to have formed at lower temperatures than those in Zone II which contain staurolite, biotite, and kyanite. In rocks of appropriate composition, diopside is stable at higher temperatures than the assemblage dolomite-quartz (Turner, 1968; Winkler, 1970). The addition of small amounts of Fe and Mn to the rock may alter the compositions of these minerals to Fe-bearing diopside and Fe-dolomite-quartz respectively, but the relative stabilities of these minerals in regard to temperature would not change. In calcareous rocks of the Flinton Group, Zones I and IA are characterized by Fe-dolomite and Zone IV by diopside, indicating a temperature (metamorphic grade) increase from Zone I to Zone IV. The fact that pegmatite interpreted as the product of partial melting of the Flinton rocks (see Chapter 1) is restricted to the diopside zone also suggests high temperatures in that area.

The increase in metamorphic grade is marked by

the appearance and disappearance of certain minerals in the pelitic and calcareous rocks (Figures 14 - 18). Boundaries between mineral zones are located on the basis of the presence or absence of these index minerals. The boundaries are isograds in the original sense of the word according to Tilley (1925); that is, lines on a map joining points of the first appearance of a particular mineral. This definition of an isograd is extended to include lines joining points of disappearance of index minerals. The isograds defined in the study area are listed below.

Pelitic rocks	Zone I (chloritoid present)
	Upper limit of chloritoid isograd
	Zone II (chloritoid absent)
Calcareous rocks:	Zone I (Fe-dolomite)
	Upper limit of Fe-dolomite isograd
	Zone II (actinolite-epidote)
	Microcline isograd
	Zone III (microcline)
	Diopside isograd
	Zone IV (diopside)

It is interesting to note that the mineral zonation observed in the calcareous rocks in the study area occurs above the upper limit of chloritoid in the pelitic rocks, approximately equivalent to the staurolite zone in pelites. The isograds defined in aluminous, Fe-bearing,

potassic rocks of the Lessard Formation are a useful addition to the isograds generally observed in pelitic and siliceous carbonate rocks. In metamorphic terrains such as the study area where pelitic schist and siliceous marble are rare or restricted in distribution, the impure calcareous rocks provide a means of evaluating variation in metamorphic grade.

In this chapter, the distribution of mineral zones in the study area has been described and the fact that the zones are the result of a gradient in the physical conditions of metamorphism has been demonstrated. A separate problem, the chemical reactions which produced the zoning, can only be dealt with briefly here. Based on a comparison of the mineral assemblages in each zone, hypothetical reactions are proposed which may have occurred in the rocks during metamorphism. The fact that these reactions are idealized representations of what actually happened needs to be emphasized. The processes of change in metamorphism are not likely to involve chemical species that can be identified as particular minerals.

Calcareous Rocks. The effect of bulk composition on the sequence of isograds is demonstrated by the fact that the disappearance of Fe-dolomite at the boundary between Zones I and II (Figures 15 and 17) corresponds to the appearance of epidote in the calcareous psammite and the appearance of actinolite in biotite carbonate schist.

Hypothetical reactions representing the upper limit of the Fe-dolomite isograd are presented below.

In the biotite-carbonate schist:

In the calcareous psammite:

The boundary between Zones II and III (microcline isograd) is the point at which the mica content of both psammite and schist is greatly reduced and the amount of microcline sharply increases. The microcline isograd is the second actinolite isograd in the biotite-carbonate schist. Carmichael's (1969) reaction (5) appears to have taken place.

The same reaction appears to have occurred in the calcareous psammite. In addition, muscovite disappears from the psammite at the microcline isograd.

This reaction may account for the reduction of the modal proportion of muscovite which occurs southward within

Zone II.

The diopside isograd separating Zones III and IV can be represented in both rock units by the reaction

Pelitic Rocks. On Figure 17 the chloritoid upper limit isograd crosses the strike of the pelitic rocks both northeast and southwest of the sample locality for Rock 5 near Bishop Corners. The upper limit of chloritoid can be represented by several different reactions, for example;

At Bishop Corners the line separating chloritoid-bearing rocks containing chlorite from those containing staurolite may represent the first appearance of staurolite in the area. The reaction which took place may have been:

Albee (1965) constructed a petrogenetic grid for the Fe-Mg silicates of pelitic schists containing quartz and muscovite for two sections in $P_{\text{solid}} - P_{\text{H}_2\text{O}} - T$ space. Garnet, chloritoid, staurolite, kyanite, chlorite, and biotite were considered to occur within the system $\text{SiO}_2 - \text{Al}_2\text{O}_3 - \text{FeO} - \text{MgO} - \text{K}_2\text{O} - \text{H}_2\text{O}$. For the

purpose of graphical representation garnet, staurolite, and chloritoid were assumed to have the composition of the pure iron end member of the respective Fe-Mg solid solution. Chlorite and cordierite were assumed to be pure Mg end members. Biotite with a Mg:Fe ratio of 1:1 was used.

Pelitic schist in the study area commonly contains four ferromagnesian minerals with quartz, muscovite, and Fe-Ti oxides. It is interesting to note that if the assemblages are equilibrium assemblages, (from the data for Rock 6, this would appear to be a good possibility), and the assemblages are treated as though they conformed to Albee's (1965) ideal pelitic system, then each four-phase assemblage represents a univariant curve on Albee's petrogenetic grid. In fact, Rock 6 (staurolite-chlorite-biotite-kyanite-garnet) would represent an invariant point. Because of the high Mn content of garnet in Rock 6, this is probably not the case; however, for the sake of discussion, the Mn content will be ignored. Figure 26 illustrates the superposition of Albee's petrogenetic grid onto the study area. The distribution of reaction curves south of the invariant point is identical to that for Albee's Figure 6 (p.530, 1965). North of the invariant point the reaction curves appear to represent a case where the invariant point (G, Cd) (Albee, Figure 5B) is superimposed on the point (K, Cd) in the same figure. That is, mineral assemblages in the

Figure 26: Superposition of Albee's petrogenetic grid onto the northern half of the Flinton area. All mineral assemblages contain quartz, muscovite, and Fe-Ti oxides. Only those three-mineral assemblages which aid in the positioning of reaction curves are included. Mineral symbols: CD - cordierite, all others as in Figure 18.

study area represent a section through $P_{\text{solid}} - P_{\text{H}_2\text{O}} - T$ space intermediate between Albee's examples for the condition that $P_{\text{H}_2\text{O}}$ is less than P_{solid} (Albee, Figures 5 and 6, 1965). The changes in the topology of the quadrilateral (a modified AKFM projection) across several of the curves in Figure 26 require more than one reaction (e.g., curves 1 and 2). These degenerate reactions are attributed to the idealized compositions used to derive the petrogenetic grid.

The common occurrence of four-phase assemblages such as chloritoid-staurolite-biotite-garnet, together with the occurrence of the rare Ca-mica, margarite, and rocks containing the combination hornblende-chloritoid-muscovite suggest the physical conditions of metamorphism are rather closely defined near Bishop Corners. The first appearance of staurolite and the upper limit of chloritoid are two of the criteria used by Winkler (1970) to define the boundary between the low-stage and intermediate-stage divisions of metamorphism. The third criterion, breakdown of chlorite in the presence of quartz, is not valid in the study area. Mg-chlorite commonly coexists with quartz, staurolite and kyanite beyond the upper limit of chloritoid. The mineral assemblages at Bishop Corners conform to those typical of the lowermost amphibolite facies in Turner's (1968) classification.

Diopside-bearing calcareous rocks are commonly associated with sillimanite-bearing pelitic rocks

(Sethuraman, 1970; Burns, 1951; J.M. Moore, personal communication) northeast and southeast of the study area. The presence of pegmatite in the calcareous rocks of the Flinton Group in the diopside zone indicates high temperatures at the time of metamorphism. However, muscovite and quartz coexist in quartzofeldspathic metasedimentary rocks in the diopside zone. The mineralogical evidence indicates conditions of metamorphism in the zone of highest grade in the study area were typical of those associated with the upper middle amphibolite facies.

Conclusions

A state was attained during metamorphism which was sufficiently close to chemical equilibrium that Duhem's theorem can be applied to the rocks used in this study. The linear programming method demonstrates that mineral zoning in calcareous and pelitic rocks of the Flinton Group is the result of a gradient in the physical conditions of metamorphism and that bulk composition is not an important variable in respect to the zoning. Metamorphic grade increased toward the southwest from Bishop Corners and eastward from Madoc during the metamorphism of the Flinton Group in the study area.

V CONCLUSIONS

Metamorphosed carbonate and clastic sedimentary rocks of the Flinton Group, part of the Grenville Supergroup, can be traced in a northeasterly-trending belt for more than 100 km (Figure 1). Detailed mapping in the Bishop Corners-Madoc area has shown that these rocks overlie unconformably plutonic rocks which in turn postdate metavolcanic rocks, marble and metagraywacke. Geochronological evidence drawn from other authors indicates deposition of the Flinton Group in the study area occurred between 1050 ± 25 m.y. and 1125 ± 25 m.y. ago. The unconformity at the base of the Flinton Group and the depth at which the rocks were metamorphosed suggests the period in which deposition occurred can be narrowed to between 1075 and 1100 m.y. ago.

The Flinton Group is divided into five formations in the Bishop Corners-Madoc area:

Bishop Corners Formation - pelitic schist, quartzite, quartzite metaconglomerate, micaceous quartzofeldspathic psammite.

Lessard Formation - calcareous quartzofeldspathic psammite, biotite-carbonate schist, epidote-bearing and epidote-free calc-silicate paragneiss, plutonic pebble metacon-

glomerate.

Myer Cave Formation - dolomite marble.

Madoc Formation - carbonate pebble metaconglomerate,
pelitic schist.

Stewart Formation - dolomite marble, graphitic
calcite marble, micaceous calcite marble.

Formations are generally less than 200 m thick and continuous for 10 - 20 km parallel to strike. Cross-bedding, preserved in quartzite layers, provides reliable top determinations. Correlation of rock units in the study area is based on lithologic characteristics and stratigraphic position with respect to the unconformity at the base of the Flinton Group.

Two major lithologic associations are present in the study area (Figure 1). At Bishop Corners and Madoc the major rock types, pelitic schist and marble, are derived from fine-grained sedimentary rocks. In the central part of the area, metamorphosed quartzofeldspathic psammite and conglomerate predominate. The transition between the two associations is interpreted as a facies change between a relatively low energy shallow marine depositional environment and a higher energy fluvial-deltaic-beach environment.

Vertically in the Flinton Group there is a general decrease in grain size and reduction in the proportion of ferric iron oxide minerals present. The relative-

ly high ferric-ferrous ratio typical of basal portions of the Bishop Corners Formation suggests a subaerial depositional environment, whereas rock units higher in the sequence are subaqueous deposits. Flinton rocks were derived from both a local and a distant provenance area. Transport directions were controlled to a large extent by the plutonic rocks.

Three major phases of deformation explain, for the most part, the structural geometry of the Flinton Group. The bulk of the Flinton rocks are exposed in the Flinton and Bishop Corners synclines. The Black River and Madoc synclines also contain the Flinton Group. The Bishop Corners and Madoc structures are simple synclines, whereas the other two consist of two major synclines and an intervening anticline. The major structures, products of the first phase of deformation (D_1), are characterized by continuity along strike for tens of kilometers and by fold axes with variable plunge. The axial surfaces parallel the regional foliation. Assymetrical minor folds, with steep axial surfaces striking northeast, deform the first foliation on a mesoscopic scale (D_2), and major first-phase isoclinal folds are re-folded to form the Northbrook antiform and Clare River synform. Evidence for a third phase (D_3) is relatively limited. Minor folds which deform second structures about northwest-trending, steeply-dipping axial surfaces can be correlated with map-scale open folds deforming

the Northbrook antiform and Clare River synform. A post-metamorphic crenulation, sporadically developed in micaceous rocks in the area, is thought to represent a fourth phase of deformation. Strained quartz grains and mylonite indicate that a late stage of brittle deformation affected the Flinton Group, and may be associated with the tectonic emplacement of a serpentine-anthophyllite body which cuts Flinton rocks. Axial ratios of deformed pebbles suggest that the thickness of lithologic units of the Flinton Group was substantially reduced during deformation. Metamorphic grade in the area was increasing during D_1 , attained a maximum just before or during D_2 , and began to decrease before D_3 .

Regional metamorphism has resulted in well-defined mineral zoning in pelitic and calcareous rocks of the Flinton Group in the study area. Textural and chemical evidence indicates the rocks recrystallized under conditions closely approaching chemical equilibrium.

Pelitic rocks in the study area are characterized by mineral assemblages containing four Fe-Mg silicates, for example, chloritoid-biotite-chlorite-garnet. These assemblages conform reasonably well to the petrogenetic grid proposed by Albee for the Fe-Mg silicates in pelitic rocks.

The presence or absence of chloritoid was found to be the most useful basis for distinguishing mineral zones in the pelitic rocks.

Mineral zoning is better developed in the aluminous, Fe-bearing, potassic, calcareous rocks where four mineral zones are defined - ferroan dolomite, actinolite-epidote, microcline, and diopside.

A problem considered in this study is to determine whether the mineral zoning in the Flinton Group is the result of variable bulk composition or a gradient in physical conditions at the time of metamorphism. The pelitic and calcareous rocks in the study area contain too many minerals and constituents for the problem to be dealt with by conventional graphical methods. A linear programming technique was applied which takes into account all the constituents and minerals in the rocks. The method requires complete chemical analyses of all the minerals in a pair of samples and a measure of the uncertainty associated with the amount of each constituent in each mineral. The application of linear programming to six samples from the Flinton Group demonstrated that the mineral zoning is the result of a gradient in physical conditions during the metamorphism of these rocks. Bulk composition is not an important variable in respect to the development of the mineral zones mapped.

The boundaries between these zones are thus isograds in the original sense of the word in that they

represent the first appearance or disappearance of zone index minerals. Furthermore, the isograds illustrate the gradient in physical conditions which existed during regional metamorphism.

Metamorphic grade increases from that typical of the lowermost amphibolite facies at Bishop Corners and at Madoc to upper middle amphibolite facies between these localities. All the variation in mineralogy of the calcareous rocks occurs within Zone II (chloritoid absent) in the pelitic rocks. This zone approximately corresponds to the staurolite zone in the pelites. The isograds defined in calcareous rocks containing Al, Fe and K are a useful addition to those generally observed in pelitic and siliceous carbonate rocks. In metamorphic terrains such as the study area where pelitic rocks and pure siliceous carbonate rocks are limited in distribution, the "impure" carbonate rocks provide an alternative means of evaluating variation in metamorphic grade.

Contributions

1) This work has demonstrated that the basic principles of stratigraphy and sedimentology can be applied to rocks of the Grenville Province despite regional metamorphism and several periods of deformation. A stratigraphic succession for the rocks in the study area was determined and the presence of an unconformity at the base of the Finton Group confirmed. Sedimentary facies

were shown to be present in the Flinton Group and the depositional environment and provenance of these rocks was interpreted.

2) Although polyphase deformation has been described in the Grenville Province of eastern Ontario by previous workers, the structural history of an area in the context of the stratigraphy and metamorphism of the rocks had not been done before the present study. Furthermore, this study has shown that three phases of deformation, including the northeast trend typical of the Grenville Province, and the accompanying regional metamorphism, occurred between the time of deposition of the Flinton Group (approximately 1075 - 1100 m.y.) and 950 - 1000 m.y., the time at which the rocks became closed systems with respect to radiogenic argon - a relatively short period near the end of the tectonically active history of the Grenville Province.

3) Linear programming has for the first time been applied to a group of metamorphic rocks in order to determine if mineral zonation is the result of a gradient in physical conditions at the time of metamorphism or of variable bulk composition.

4) Isograds which have not generally been recognized in the past have been defined in calcareous rocks containing Al, Fe, and K. These boundaries are useful in that they provide a means of evaluating variation in metamorphic grade in terrains where siliceous carbonate or pelitic rocks are rare or absent and in a range of conditions to which pelites and pure siliceous carbonate rocks are relatively insensitive.

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Appendix 1

Finite Strain in a Quartzite Pebble
Metaconglomerate Near Flinton, Ontario.

Introduction

Some measure of the total (finite) strain in deformed rocks is essential to a complete understanding of their structural history. The shape of a pebble in a deformed conglomerate is the result of the superposition of a strain state, represented by the strain ellipsoid, upon the original shape of the pebble. The relative influence of these two factors depends on the orientation of the pebble with respect to that of the strain ellipsoid, the configuration and intensity of strain, and the relationship between the pebble and surrounding matrix. If the pre-deformation shape and orientation are known, a minimum measure of the total strain can be obtained from axial ratios of deformed pebbles.

Ramsay (1967) has demonstrated on a theoretical basis that variation in original orientation and shape may explain a significant amount of the regional variation observed in strained pebbles. Hossack (1968), using Ramsay's approach, pointed out that the error in location of points on a strain plane diagram could be as much as ± 50 percent where the intensity of strain was low; however, the error associated with original shape and orientation factors decreased with increasing strain. Hobbs and Talbot (1967) concluded the error introduced by ignoring the state of pebbles before deformation was so large that axial ratio data from deformed pebbles were useless for strain analysis.

From studies of undeformed sedimentary rocks (Griffiths, 1967; Potter and Pettijohn, 1963), pebbles are known to generally have an ellipsoidal shape and a preferred orientation. Using this information, Elliot (1970) derived a "shape factor grid", with which a quantitative estimate of the original shape and orientation of deformed pebbles can be determined from the shape and location on the grid of the cluster of pebble axial ratios.

This appendix presents results of a preliminary study of strain in a quartzite-pebble metaconglomerate of the Grenville Province near Flinton, Ontario. Axial ratio data from four samples at one locality are analyzed graphically in an attempt to evaluate the reliability of various techniques in the literature, and the feasibility of regional strain analysis.

Geological Setting

In the Madoc-Flinton-Fernleigh area of eastern Ontario the Flinton Group (Moore and Thompson, in press) comprises clastic metasedimentary rocks and marble exposed in narrow isoclinal synclines for more than 100 km along strike (Figure 27). These rocks unconformably overlie granodioritic plutons which intrude a sequence of metavolcanic rocks, metagraywacke and marble. The present understanding of the stratigraphy of the Flinton Group is based on detailed mapping carried out by J.M.

Moore (1967), K. Sethuraman (Sethuraman and Moore, in press), and the writer. The quartzite metaconglomerate used in this study is part of the Bishop Corners Formation, the basal unit of the Flinton Group at the sample locality. The conglomerate is present along the entire length of the Flinton syncline (Figure 27).

In pelitic schist near the sample locality, kyanite, biotite, and staurolite coexist with muscovite and quartz, indicating metamorphic grade in the vicinity was equivalent to that of the middle amphibolite facies. Textural evidence suggests metamorphic grade in the Bishop Corners-Madoc area was increasing during D_2 , remained high for some time, and, before D_3 , began to decrease.

The complex structural history of the Flinton Group includes at least four phases of deformation. Major isoclinal folds, products of the first phase of deformation, are characterized by continuity along strike for tens of kilometers and by hinge lines with variable plunge. Their axial surfaces are parallel to the regional foliation. Asymmetrical minor folds, with steep axial surfaces trending northeast, deform the first foliation on the scale of an outcrop, and regional first-phase isoclinal folds are folded to form structures such as the Northbrook antiform and Clare River synform (Figure 27). A second cleavage associated with these structures is developed in micaceous rocks. Evidence for a

third phase is relatively limited. Minor folds which deform second structures about northwest-trending, steeply dipping axial surfaces can be correlated with regional open folds deforming the Northbrook antiform and Clare River synform. A gently-dipping crenulation, sporadically developed in micaceous rocks in the area, which in part postdates metamorphic recrystallization, represents a fourth phase of deformation. Strained quartz grains and mylonite indicate that, at one locality at least, a late stage of cataclastic deformation affected the Flinton Group.

Description of Samples

Four large samples of quartzite metaconglomerate were collected from an outcrop 3.5 km northeast of Flinton (Figure 27). The samples, weighing 20 to 30 kilograms each, were selected from frost-loosened blocks lying on the surface of the outcrop. The strongly foliated metaconglomerate (Plate 1b, Chapter 3) comprises sixty to seventy percent deformed pebbles in a fine grained schistose matrix of muscovite, biotite, quartz, plagioclase, hematite and minor amounts of epidote, tourmaline and carbonate. In order of decreasing abundance, most of the pebbles are white, gray, or brown quartzite. White-weathering, medium-grained "vein quartz" is the other prominent pebble type. To test variation of strain in pebbles of different composition, five

Figure 27: Distribution of the Flinton Group in southeastern Ontario and sample locality (●) for the metaconglomerate used in this study.

pebble types were distinguished by eye in the samples: 1) "vein quartz" (27.2), 2) white quartzite (22.2), 3) gray-brown quartzite (30.3), 4) red-brown quartzite (12.0), and 5) black quartzite (8.3). The numbers in parentheses refer to modal percent of each pebble type present in sample D. Types 3 and 4 were clearly differentiated in only one of the samples; in the other three, these pebbles were treated as a single group (Type 3). Pebble size ranges from 0.5 to 25 cm parallel to the longest dimension, but most are between 5 and 15 cm long. All the pebbles in the metaconglomerate are tri-axial ellipsoids. The longest, intermediate, and shortest axes of the pebble ellipsoids have been designated Z, Y, and X respectively. The convention $Z > Y > X$, though contrary to that used by Flinn (1956) and Ramsay (1967), is consistent with the system used by Hossack (1968) and by crystallographers.

Preparation of Samples and Calculation of Axial Ratios

The samples (A, B, C, D) were cut into blocks bounded by three perpendicular planes (Figure 28); the ZY plane parallel to the foliation in the sample, the ZX plane perpendicular to the foliation and parallel to the direction of maximum elongation (Z), and the YX plane perpendicular to both the foliation and the Z direction. Approximately twenty sections were cut in each sample. The sections were varnished to improve distinction of

Figure 28: Method for obtaining principal sections from the samples.

pebbles and matrix.

Elliptical outlines of pebbles on each section were traced onto acetate film (Figure 29) to record the distribution of pebbles and allow accurate measurement of the principal axes. The angle (α) between the long axis of each pebble and a reference line was also recorded. In the ZX and YX sections, the trace of the foliation was the reference line. In the ZY section, the dominant trend of the long axes was chosen as a reference line.

The axial ratio and α -angle for each pebble ellipse on a particular principal section are required for Elliot's (1970) shape factor grid. At each sample locality, Flinn (1956) and Hossack (1968) averaged the axial ratios of all the pebble ellipses in each of two principal sections: Z/Y and Y/X. Each sample was represented by a point on their deformation plots. In order that variation in strain with pebble composition would not be obscured, for this study the average Z/X, Z/Y, and Y/X ratios of each of the five pebble types were calculated for each sample, resulting in several points per sample on the deformation plots. 800 axial ratios were obtained, approximately 60 for each of the Z/X, Z/Y, and Y/X sections in each sample (Table 16). The number of axial ratios measured for each pebble type varies with its abundance. Only sample D contains all five pebble types.

Figure 29: Examples of principal sections used in this study. Numbers refer to pebble composition.

Table 16 - Axial ratios from samples A, B, C, and D

Sample	Z/Y	Number	Stand. Dev.	log _e Z/Y	Y/X	Number	Stand. Dev.	log _e Y/X	Z/X	Number	Stand. Dev.	log _e Z/X	Z/Y·Y/X	log _e Z/Y + log _e Y/X	$\bar{\Sigma}_s$	$\bar{v}$
A 1	2.2	9	0.7	0.79	1.9	26	0.5	0.64	2.3	14	1.2	0.83	4.2	1.43	1.00	-0.17
2	3.2	56	1.2	1.16	2.5	39	0.9	0.92	7.5	28	2.8	2.02	8.0	2.08	3.61	-0.12
3	4.0	8	1.1	1.39	2.3	13	1.1	0.83	7.3	11	2.0	2.00	9.2	2.22	3.79	-0.28
B 1	1.8	8	0.6	0.59	1.8	37	0.5	0.59	3.3	22	1.5	1.19	3.2	1.18	1.22	0.00
2	3.5	24	1.2	1.25	2.9	45	1.2	1.06	9.6	40	4.4	2.26	10.0	2.32	4.51	-0.08
3	4.2	13	2.0	1.44	3.8	13	0.9	1.34	7.2	23	2.6	1.97	16.0	2.77	4.47	-0.05
5	4.8	8	3.2	1.57	3.3	5	0.5	1.19	17.6	6	9.1	2.87	16.0	2.76	7.00	-0.13
C 1	1.3	4	0.1	0.26	2.2	12	1.5	0.79	-	-	-	-	2.9	1.05	-	-
2	3.7	13	1.6	1.31	2.4	17	1.0	0.88	6.3	10	2.5	1.84	8.9	2.18	3.39	-0.26
3	3.8	9	1.7	1.34	2.6	10	0.5	0.96	-	-	-	-	9.9	2.29	-	-
D 1	2.8	16	1.8	1.03	2.1	32	0.7	0.74	3.5	21	1.4	1.25	5.9	1.77	1.84	-0.23
2	5.0	10	2.2	1.61	2.5	39	0.9	0.92	6.3	12	3.5	1.84	13.0	2.53	3.93	-0.37
3	4.5	18	2.1	1.50	2.1	39	0.8	0.74	7.3	21	4.8	2.00	9.5	2.25	3.90	-0.38
4	5.2	4	1.4	1.65	3.2	17	1.5	1.16	6.1	12	1.6	1.81	17.0	2.81	4.23	-0.26
5	6.1	8	2.2	1.81	4.2	5	0.8	1.44	7.9	5	5.8	2.07	26.0	3.24	5.54	-0.18
B.A.	1.4	126	0.2	0.334	1.4	126	0.3	0.336	1.96	126	-	0.672	1.96	0.670	0.391	0.00
R.H.	1.3	203	.13	0.262	1.4	203	0.2	0.336	1.82	203	-	0.598	1.82	0.598	0.313	0.11

The average axial ratios for each pebble type are listed in Table 16. As a check on the accuracy of the cutting of the principal sections, the relation $Z/X = (Z/Y)(Y/X)$ can be used. Of the thirteen average Z/X ratios measured in samples A, B, C, and D, only four were close to the product of the average Z/Y and Y/X ratios (Table 16). Data for pebble types 1 and 3 were not available on the Z/X section for sample C. Part of the error can be associated with imprecise cutting of the principal sections. Great care must be taken to insure accurate location of the saw cuts. The fact that the four internally consistent ratios are not restricted to one sample suggests that inaccurate cutting is not the only source of error. Possibly there is variation in orientation of pebbles of different composition in the same rock. Sample B includes three of the four internally consistent values.

Plotting Methods

Flinn deformation plot. To compare the deformation of pebbles from different parts of the Funzie conglomerate, Flinn (1956) plotted the average Z/Y ¹ ratio for the pebbles at each sample locality on a graph with cartesian coordinates (Figure 30). Z, Y, and X are the maximum, intermediate and minimum axes of the pebble ellip-

¹ The convention used here, $Z > Y > X$, differs from that used by Flinn (1956), $X > Y > Z$.

Figure 30: Deformation plot after Flinn (1956).

$$d = \sqrt[3]{ZYX}$$

I - VI: zones of deformation

BA - quartzite pebbles from a modern beach deposit.

RH - quartzite pebbles from undeformed conglomerate.

soids. Three-dimensional strain was measured by comparing Z, Y, and X to "d", the diameter of a sphere with the same volume as the pebble ellipsoid ($d = \sqrt[3]{ZYX}$). Using this relationship, curves representing strain in each of the three principal directions were plotted on the graph (Figure 30). Strain increases radially outward from the origin of the graph. If the assumption that the pebbles were spherical before deformation is valid, the position with respect to these curves of a point representing the average axial ratios for a group of pebbles indicates the amount the pebbles have been strained. Flinn (1956) emphasized that by comparing deformed pebbles to a spherical reference state he was not suggesting undeformed pebbles are spherical - comparison to a sphere of equal volume is merely convenient. In fact, the radial distance between the position of a sphere on the deformation plot ($Z/Y = Y/X = 1.0$) and the position of points representing pebbles from an undeformed beach deposit (B.A.) and an undeformed conglomerate (R.H.) is not large (Figure 30).

The line representing pebbles for which the Y axes did not change throughout deformation divides the graph into two fields; the constrictional strain field where $Y < d$ and the field of flattening where $Y > d$. Most deformed pebbles are triaxial ellipsoids which plot in one or the other of these fields. Perfect prolate (cigar-shaped) or oblate (pancake-shaped) ellipsoids are

rare. Prolate ellipsoids plot on the line $Y/X = 1.0$ with $Z/Y > 1$ and oblate ellipsoids on the line $Z/Y = 1.0$ with $Y/X > 1$. The sections of the deformation plot where either $Z/Y < 1.0$ or $Y/X > 1.0$ are not defined because if either Z/Y or Y/X was less than one, the relationship $Z > Y > X$ would not hold.

When Flinn plotted the Funzie conglomerate pebbles on the deformation diagram, he discovered a regional variation in the intensity and in the symmetry of strain. The points representing the average shape of pebbles in a particular region could be separated geographically on the plot (Figure 30). Where the rock was exposed in the north (Zone VI), the strain was less intense than that in conglomerate from the south (Zones I and II). The symmetry of strain changed from flattening in Zones I and II to constrictional in Zones III to VI. Another important point was that the radial distance from a deformed pebble in Zone I and to one of the undeformed pebbles is three times the distance between the average undeformed pebble and a reference sphere. Evidently, the more intense the strain, the less important is the original shape of the pebble.

Axial ratio data from four samples of Flinton quartzite metaconglomerate were plotted on a Flinn deformation plot (Figure 31). Each point on the diagram represents the average Z/Y and Y/X ratios for a particular pebble type in one of the four samples. Two points

Figure 31: Data from samples A, B, C, D of Flinton conglomerate plotted on a Flinn diagram. Numbers associated with data points refer to pebble composition: 1 - vein quartz, 2 - white quartzite, 3 - gray-brown quartzite, 4 - red-brown quartzite, 5 - black quartzite. Dotted lines enclose pebbles of similar composition. BA and RH represent average axial ratios of undeformed pebbles (see Figure 30).

representing the undeformed pebble populations on Flinn's diagram are included. All but two of the points plot in the field of constrictional deformation. The amount of strain in each sample varies with pebble composition, ranging from less than 100 percent ($Z = 2d$, $X = 0.5d$) for Type 2 to 500 percent ($Z = 5d$, $X = 0.2d$) for Type 5. If the pebbles all had the same initial shape and similar strain histories, a minimum measure of the finite strain in the rock would be indicated by the most highly strained pebbles. Data are more complete for Type 2 and Type 3 pebbles than for Types 4 and 5. Type 1 (vein quartz) is compositionally distinct from the others. For these reasons, the intermediate group on the deformation plot (Figure 31) was chosen as the safest measure of the minimum total strain at the sample locality.

Quartzite pebbles in the Flinton conglomerate are more intensely strained than those in the Funzie conglomerate. Note that the radial distance between the location of a spherical pebble on the plot and undeformed pebbles is about one-seventh the distance between the undeformed pebbles and the most strained pebbles in the intermediate group. A spherical reference state is a better approximation for the rocks used in this study than for the Funzie conglomerate.

The Natural Strain Deformation Plot. Ramsay (1967)

developed a deformation plot based on natural or logarithmic strain rather than simple axial ratios. The use of natural (logarithmic) strain reduces the scatter of a group of data and results in relatively less complex deformation paths (Ramsay, 1967). It is possible to convert axial ratios to natural strain units using the following relationships. If e_1 , e_2 , and e_3 represent elongation parallel to the Z, Y, and X axes of a pebble respectively; ϵ_1 , ϵ_2 , and ϵ_3 represent natural strain parallel to Z, Y, and X, and $\log_e x$ is the natural logarithm of x , then:

- (1) $Z = 1 + e_1$; $Y = 1 + e_2$
- (2) $\epsilon_1 = \log_e (1 + e_1)$; $\epsilon_2 = \log_e (1 + e_2)$
- (3) $\epsilon_1 = \log_e Z$; $\epsilon_2 = \log_e Y$
- (4) $(\epsilon_1 - \epsilon_2) = \log_e (Z/Y)$

Similarly, $(\epsilon_2 - \epsilon_3) = \log_e (Y/X)$, $(\epsilon_1 - \epsilon_3) = \log_e (Z/X)$

The natural logarithm of Z/Y is plotted along the ordinate of the deformation plot and the natural logarithm of Y/X along the abscissa (Figure 32). Whereas on the axial ratio plot the simplest possible deformation path is a curve, on the natural strain diagram, the simplest path is a straight line with the form $\log_e (Z/Y) = K \log_e (Y/X)$ where K is the slope of the line. The line with K equal to one divides the constrictional field of the diagram ($K > 1$) from the flattening field ($K < 1$).

Figure 32: Data from the Flinton metaconglomerate plotted on the natural strain deformation diagram developed by Ramsay (1967). Dotted lines enclose pebbles of similar composition. Symbols are the same as those in Figure 31.

Figure 32: Data from the Flinton metaconglomerate plotted on the natural strain deformation diagram developed by Ramsay (1967). Dotted lines enclose pebbles of similar composition. Symbols are the same as those in Figure 31.

The average axial ratios used for the Flinn diagram were recalculated to natural strain units (Table 16) and plotted on a natural strain deformation plot (Figure 32). The scatter of the data is reduced. Once again, the data can be grouped with respect to pebble composition. Type 1 shows the least strain; Types 4 and 5, the maximum. The representative value of strain for the sample was derived from the intermediate group composed of the most common pebble compositions. Except for C_1 and B_1 , all the points occur in the constrictional strain field.

The Strain Plane Deformation Plot. The parameters used by Hossack (1968) to describe the magnitude and symmetry of strain in deformed pebbles differ from those used by Flinn and Ramsay. These terms, $\bar{\epsilon}_s$ and ψ (based on natural strain units), are the coordinates of a polar graph resulting from the projection of a three-dimensional orthogonal strain state onto a plane. A detailed description of the polar graph (strain plane) is given by Hossack. The terms can be calculated from axial ratio data using the following equations:

$$\bar{\epsilon}_s = \frac{\sqrt{3}}{2} \bar{\gamma}_o$$

where $\bar{\gamma}_o$ is the Natural Octahedral Unit Shear (Nadai, p.71, 1956)

$$\bar{\gamma}_0 = \frac{2}{3} [(\xi_1 - \xi_2)^2 + (\xi_2 - \xi_3)^2 + (\xi_3 - \xi_1)^2]^{\frac{1}{2}}$$

Substituting natural logarithms of the pebble axial ratios for the natural strain terms:

$$\bar{\gamma}_s = 0.867 [(\log_e Z/Y)^2 + (\log_e Y/X)^2 + (\log_e Z/X)^2]^{\frac{1}{2}}$$

$$\text{Lode's Parameter, } \gamma = \frac{(\log_e Y/X - \log_e Z/Y)}{(\log_e Z/X)} \quad (\text{Gay, 1969})$$

Lode's parameter is analogous to the k values used on the natural strain plot as a measure of strain symmetry. The radius on the strain plane equivalent to plane strain, $\gamma = 0$, separates the constrictional field where Lode's parameter is negative, from the flattening field where Lode's parameter is positive (Figure 33). Simple extension or ideal constrictional deformation of a sphere results in a cigar-shaped prolate ellipsoid ($\gamma = +1$), whereas simple compression or ideal flattening produces a pancake-shaped oblate ellipsoid ($\gamma = -1$).

Hossack (1968) calculated the amount and symmetry of strain from pebble axial ratios at 81 localities near Bygdin and Olefjedd, southern Norway. He contoured the values of $\bar{\gamma}_s$ and γ at these localities to illustrate the regional variation of these parameters. In this work Hossack assumed the original shape of the pebbles was spherical and ignored the predeformation orientation. An effort was made, however, to place limits

on the error associated with these assumptions. He concluded the error was as high as ± 50 percent in areas of "low" strain ($\bar{\epsilon}_s = 1.1$), and ± 25 percent in areas of "high" strain ($\bar{\epsilon}_s = 2.2$).

Data from deformed pebbles measured in this study were converted to the parameters used by Hossack (Table 16) and plotted on a strain plane diagram (Figure 33). The distribution of points is much narrower perpendicular to the direction of increasing strain, and the pebble types form more distinct groups than on previous plots. The magnitude of $\bar{\epsilon}_s$ ranges from about 1.0 to a maximum of 7.0 with the intermediate group between 3.5 and 4.5. Because sample C was only 12-15 cm thick in the Z direction, the only complete pebble ellipses present on the ZX principal section were Type 2. For this reason points C_1 and C_3 do not appear on this plot. The radial distance between the reference sphere at the origin of the plot and the undeformed pebbles is only one-tenth the distance between the undeformed pebbles and the upper part of the intermediate group of deformed pebbles. Here a spherical reference state is a reasonable approximation. B_1 plots on the line representing plane strain ($\gamma = 0$). The rest of the data are in the field of constrictional deformation. The quartzite pebbles in this study are considerably more strained than those used by Hossack. The error estimated by Hossack to be associated with the original shape

Figure 33: Data from the Flinton metaconglomerate plotted on the strain plane diagram used by Hossack (1968). Symbols as in Figure 31.

and orientation factors would be significantly reduced for the Flinton metaconglomerate.

From the work of Flinn and Hossack, it is apparent that comparing deformed pebbles to a reference sphere is a workable approximation. Where possible, however, an effort should be made to determine the original shape and orientation of the pebbles.

The Shape Factor Grid. A quantitative estimate of the initial shape and orientation of pebbles in a deformed rock can be determined using the shape factor grid developed by Elliot (1970). The method eliminates the need for the assumption made by Flinn (1956) and Hossack (1968) that the initial pebble shape was a sphere, and Ramsay's (1967) assumption of an initially random distribution of ellipses. The shape and orientation of the distribution of data from undeformed sedimentary rocks was determined by plotting on the grid data from undeformed pebble populations. Matrix algebra was used to "deform" these distributions. Elliot (1970) observed that certain characteristics of the undeformed distribution were not altered by the strain. For example, the heart or delta-shaped distribution typical of an unimodal population of ellipsoids with a preferred orientation moves to the right across the grid and rotates but the basic form does not change. The initial undeformed pebble distribution can be determined from the distribution of deformed pebbles without knowledge of the form

and orientation of the strain ellipsoid (Elliot, 1970, p. 2221).

The shape factor grid is a polar graph. The radius vector, ϵ , is a measure of the magnitude of strain, the polar angle is twice the angle (α) between the long axis of the pebble ellipse and a reference line. The parameter ϵ is based on the natural logarithm of the axial ratio. The following equations taken from Elliot (1970) outline the derivation.

$$\begin{array}{ll} \text{If } A_1 = e^{\epsilon_1}, A_2 = e^{\epsilon_2} & A_1 = \text{long half} \\ \text{the area of the ellipse (omitting} & \text{axis of} \\ \text{constants) is} & \text{ellipse} \\ A_1 \cdot A_2 = e^{\epsilon_1 + \epsilon_2} & A_2 = \text{short half} \\ \text{If } 2\epsilon_A = \epsilon_1 + \epsilon_2 & \text{axis of} \\ & \text{ellipse} \end{array}$$

$$\begin{array}{l} \text{Define } \epsilon \text{ so that } \epsilon_1 = \epsilon_A + \epsilon \\ \text{and } \epsilon_2 = \epsilon_A - \epsilon \end{array}$$

$$\text{The axial ratio of the ellipse is: } \frac{A_1}{A_2} = e^{\epsilon_1 - \epsilon_2}$$

$$\text{Axial ratio} - \frac{\text{long axis}}{\text{short axis}} = e^{2\epsilon}$$

$$\epsilon = \frac{1}{2} \ln \left(\frac{\text{long axis}}{\text{short axis}} \right)$$

The deformed pebble or final ellipsoid is the result of the superposition of the strain ellipsoid upon an initial ellipsoidal shape. Elliot subscripts the value for each of these ellipsoids ϵ_f , ϵ_s , and ϵ_i ,

respectively. ϵ_f or ϵ_i can be the radius vector of the polar graph.

If a significant distribution becomes apparent when data from deformed pebbles are plotted on the shape factor grid, the form and orientation of the strain ellipsoid, and the initial shape and orientation of the pebbles can be determined. From this, the magnitude of the total strain is evident. The symmetry of strain, a major variable in regional strain analysis, is not represented on the shape factor grid, but strain symmetry can be evaluated if a grid is constructed for each of the three principal sections.

The most likely distribution of undeformed pebbles (Elliot, 1970) has the form of a triangle with the base at the origin of the graph (Figure 34). The area immediately adjacent to the origin represents spherical pebbles. Spherical pebbles are rare in natural undeformed pebble populations and as a result, the polar graph is generally free of data points in the vicinity of the origin. The lack of data in this area causes the triangle to become heart-shaped, with the "notch" at the origin. The notch or gap in the distribution is referred to as the initial circle point (Elliot, 1970). Because the problems of initial shape and orientation do not exist for spherical pebbles the initial circle point in a deformed distribution yields directly the axial ratio and orientation of the strain ellipsoid

Figure 34: Shape factor grid with a heart-shaped initial distribution to the left of the origin. The general form of the distribution is preserved after a strain of $\epsilon_s = 0.60$ (after Elliot, p.2224, 1970). The notch in the heart corresponds to the initial circle point.

with respect to the reference line in each principal section. The total strain ϵ_f measured along the radius vector to the initial circle point is the strain applied to all the pebbles in the distribution.

The form of the distribution illustrates the result of adding this strain to a group of non-spherical pebbles with a variety of orientations.

Using the ϵ_f and α values for each pebble ellipse on the principal sections in samples B and D, a point representing each ellipse was plotted on the shape factor grid. A polar graph was constructed for the Z/X, Z/Y, and Y/X sections and the collection of points outlined using Melcher's method (Flinn, 1958). Data from both samples were combined because the wide variation in axial ratios in these rocks necessitates a large number of points (>100) to define a useful distribution. Elliot (1970) mentions that the correct diameter of circle used to contour the collection of points on the grid is the one resulting in a distribution which includes most of the data. A circle containing 5 percent of the plot area was used here.

Elliot (1970, p. 2230) pointed out the need to restrict strain analysis to pebbles of a particular composition. The deformation plots in Figures 31, 32, and 33 from this study illustrate the importance of considering pebble composition when measuring finite strain. The relationship clearly defined on the plots where

Figure 35: Distribution obtained from Z/X sections in samples B and D. Pebble Type 1 (thin line) can be distinguished from Types 2, 3, and 4 (heavy line). There are not enough data to separate Type 5 (designated by the number 5 on the plot).

average axial ratios were used is not as obvious on the shape factor grid. In Figure 35, Type 2 (vein quartz) pebbles can be separated from quartzite pebbles but there were not enough points representing Type 5 (black quartzite) for these pebbles to be distinguished as a group from the other quartzite pebbles.

On the shape factor grids in Figures 36 - 38, data from Types 1 and 5 were not included. The remaining points correspond to the intermediate group of pebbles present on the Flinn, natural strain, and strain plane deformation diagrams. Unfortunately, due to the inexact form of the Z/Y and Y/X distributions, the selection of an initial circle point was somewhat arbitrary. The shape factor grid for the Z/X section contains the best-developed distribution of the three (Figure 36). This section includes the longest and shortest axes of the pebble ellipsoids and therefore exhibits the greatest degree of strain, $\epsilon_f = 0.60$. The heart-shaped outline of the distribution indicates the initial shape of the pebbles in the samples was ellipsoidal and they defined a unimodal preferred orientation. There are fewer data (55 points) from the Y/X sections of samples B and D than from the Z/Y and Z/X sections. The initial circle point of this narrow heart-shaped distribution (Figure 37) indicates a value of approximately 0.25 for the magnitude of strain. The poorly developed form of the distribution and low density of points leaves this value

Figure 30: Data for the Z/X section from samples B and D of the Flinton conglomerate plotted on the shape factor grid (Elliot, 1970). A large black dot indicates the position of the initial circle point on each grid.

Figure 37: Data for the Y/X section from samples B and D of the Flinton conglomerate, initial circle point as on Figure 36.

open to question. The distribution for the Z/Y section (Figure 38) is based on 120 points. Once again, however, the form of the distribution is difficult to evaluate. A broad, heart-shaped distribution indicating a minimum finite strain (ϵ_f) of 0.45 was interpreted from this diagram.

If the relation $Z/X = (Z/Y)(Y/X)$ is converted to natural strain units, the equation becomes $\log_e(Z/Y) = \log_e(Z/Y) + \log_e(Y/X)$. That is, the amount of strain in the Z/X section should equal the sum of the strains in the other two sections. In this case, the relationship holds rather well; the strain in the ZX section ($\epsilon_f = 0.60$) is quite close to being equal to the sum of the strain in the Z/Y and Y/X sections ($0.25 + 0.48 = 0.73$). The fact that there is a reasonably close correspondence is an internal check of the quality of the data.

Discussion

Strain and structural geometry. In the Flinton area most of the strain evident in deformed pebbles probably occurred during the first phase of deformation (D_1). The intermediate and longest axes of the pebble ellipsoids, Y and Z, lie in S_1 , the regional foliation. The shortest axis X is perpendicular to this surface. The long axes of most pebbles define a well-developed preferred orientation which generally plunges down the dip of S_1 . S_1 and flattened pebbles contained in S_1 were

Figure 38: Data for the Z/Y section from samples B and D of the Flinton conglomerate, initial circle point as on Figure 36.

deformed into moderately-plunging asymmetrical minor folds during the second phase of deformation. The pebbles were strained during the second phase of deformation (D_2) but the amount of strain which occurred then is not known. Regardless of when the strain occurred, an estimate of the total strain is useful and further fabric work on the pebbles may allow determination of the timing of deformation more precisely.

Comparison of plotting methods. The Flinn deformation plot is the simplest to use and in areas of moderate to high strain indicates the regional variation of finite strain if measurements are restricted to a single pebble composition. The scatter of data is much greater on this type of plot than on one based on natural strain units. Although the natural strain deformation diagram (Ramsay, 1967) does not indicate directly the absolute magnitude of strain, the method is useful for measuring strain in a relative sense within a particular region. In this study, the Hossack strain plane plot was found to be the best method of the three based on average axial ratios. Data scatter was less on this plot than on either the Flinn or natural strain plots. The separation of pebbles with respect to composition was very good, and the use of a spherical reference state was a better approximation on the strain plane plot than on the other two diagrams.

The advantage the shape factor grid has

over other graphical methods is that the initial shape and orientation of deformed pebbles can be estimated. Furthermore, axial ratios are not averaged - all the data collected are plotted on the polar graph. Distinction of the effects of different pebble compositions is not as clear on the shape factor grid as on the Hossack plot, nor is symmetry of strain directly evident. For measuring the amount of strain, however, Elliot's grid is superior to the others in that it separates finite strain from the original ellipsoidal shape and orientation of the pebbles.

Pebble/matrix relationship and volume change. The validity of the strain analysis carried out here depends on two assumptions: the strain recorded was a homogeneous strain, and the volume of the pebbles did not change during deformation. For a rock such as the quartzite metaconglomerate used in this study, in which the matrix is markedly different in composition from the pebbles, the model of homogeneous strain is inaccurate. The ductility contrast between the pebble and matrix results in a rotational component of strain, the magnitude of which is not evident in the outcrop. Gay (1969) has suggested a method of dealing with this problem. However, the relative amount of matrix and pebbles present is important. If the proportion of matrix in the rock is sufficiently low that most of the pebbles are in contact, as is the case with the samples used here (Plate

1b, Chapter 3), the assumption of homogeneous strain is probably reasonable.

If the volume of the individual pebbles was not constant during deformation, the values for finite strain obtained in this study would not be correct. However, the quartzite pebbles undergo little mineralogical transformation during metamorphism, nor is there evidence in the Flinton conglomerate for transfer of material in the form of pitted pebbles or quartz "tails" (see Figure 5-48, Ramsay, 1967).

Symmetry of strain. Once an approximation of the original shape of the pebbles is obtained from the shape factor grid, the symmetry of strain is best evaluated from the plots based on average axial ratios. The majority of the data for Flinton pebbles on the Flinn (Figure 30), natural strain (Figure 31), and strain plane (Figure 32) deformation diagrams is located in the constrictional field. The distribution of points is immediately adjacent to the line representing plane strain (Y axis constant during strain) which separates the two major strain symmetries. From this it is evident that at the sample locality strain in the metaconglomerate was much greater parallel to the Z and X axes of the pebbles than parallel to Y.

Magnitude of strain. In this study, four methods have been used to analyze finite strain. All four methods indicate maximum strain occurred parallel to Z and

X with much less strain parallel to Y, but in each case the strain is measured by a different parameter. Using point B₂, one of the internally consistent sets of ratios, as a reference point in the intermediate group of data on the Flinn, Ramsay, and Hossack plots, and the initial circle point on the shape factor grid, the results for the Z/X section are summarized in Table 17:

Table 17 - Finite strain parameters

Plot	Finite Strain parameters	Value
Flynn	a/d^*	$X = 0.29d$ $Z = 3.6d$
Natural Strain	$\epsilon_1 - \epsilon_3$	2.3
Strain Plane	$\overline{\epsilon}_s$	4.5
Shape Factor Grid	ϵ_f	0.60

*a = Z, Y, or X, where Z, Y, and X represent the longest, intermediate and shortest axes of the pebble ellipsoids.

$d = \sqrt[3]{ZYX}$ the diameter of a sphere with volume equal to that of the pebble ellipsoid.

Deformation plots based on average axial ratios, such as the strain plane diagram, are useful for evaluating the variation of finite strain in a relative sense in a particular region. Elliot (p.2228, 1970) has shown, however, that average axial ratios result in an overestimate of the absolute value of finite strain. This fact is demonstrated in Table 17. Assuming pebbles were initially spherical results in a further overestimate of total strain. The ϵ_f value from the shape factor grid (0.60) is a measure of the strain which would occur in a pebble which was chosen from the central part of the distribution (Figure 36) where the pebbles were ellipsoids before deformation, $\epsilon_f = 0.90$. Similarly, the minimum measure of strain indicated on the Flinn, Ramsay, and Hossack plots is the result of the superposition of a strain state onto an ellipsoidal pebble. The assumption of a spherical reference state and the use of averaged axial ratio data combine to produce a large overestimate of the absolute finite strain in a deformed conglomerate. The finite strain obtained from the shape factor grid for the Z/X sections of samples B and D, $\epsilon_f = 0.60$, was chosen as the most accurate estimate of finite strain in the Flinton conglomerate at the sample locality.

The magnitude of strain arrived at in this study applies only to deformation of the pebbles. Extrapolation to a larger scale can only be done in a

general way if at all. There is no doubt, however, that the strain measured in the pebbles represents a minimum for the rock as a whole. The amount of strain which occurred in the matrix of the conglomerate is not known. If finite strain in the Z/X section of the conglomerate was at least 0.60, the rock unit may have been shortened by an appropriate amount perpendicular to the foliation; that is, parallel to the X axis of the pebble ellipsoids. The line of symmetry on the distribution in Figure 35 represents the orientation of bedding in the principal section, Z/X, after strain. The angle between the reference direction (foliation) and bedding is 11 degrees. In the vicinity of the sample locality, bedding is sub-parallel to the regional foliation (S_1) except near first phase fold hinges. If the strain measured in this study occurred when bedding was sub-parallel to the foliation (the Z/Y plane of the pebble ellipsoids) the thickness of the conglomerate would be substantially less than it was before deformation. The value $\epsilon_f = 0.60$ for a hypothetical spherical pebble amounts to a shortening of approximately 350 percent parallel to the X axis of the pebble and hence nearly perpendicular to the present orientation of bedding. This is equivalent to decreasing the thickness of the conglomerate unit to less than one-third its original value. Less competent units would have thinned to a much greater degree. In a study of the stratigraphy

of intensely deformed rock units such as those of the Flinton Group, finite strain is an important consideration.

Conclusions

Graphical analysis is a valid and useful method for evaluating finite strain in deformed pebbles. For a comprehensive study of strain the shape factor grid should be used in conjunction with a Flinn deformation plot or Hossack strain plane diagram. The plots based on average axial ratios are useful for measuring variation of strain with pebble composition and strain symmetry. The shape factor grid yields an estimate of the initial pebble shape and orientation, and the magnitude of finite strain. A minimum value of 350 percent was obtained for the constrictional finite strain in the pebbles of quartzite metaconglomerate near Flinton, Ontario. This is equivalent to decreasing the metaconglomerate to less than one-third its pre-deformation thickness.

Appendix 2

Petrographic Descriptions and Modal
Analyses of the Analyzed Rocks (1-6).

Petrographic Descriptions of the Analyzed Rocks

Biotite-carbonate schist (Rock 1). Rock 1 (sample number 160569) is a fine-grained schistose rock composed of thin (0.5-1 mm) discontinuous laminae of biotite, plagioclase, and minor carbonate interlayered with laminae of quartz, minor plagioclase, and two carbonates, ferroan calcite and Fe-dolomite. The layering is tightly folded (apical angles - $10-20^{\circ}$) and most of the mica is oriented parallel to the axial surfaces (S_1) of the folds. A modal analysis is presented in Table 18. Biotite (0.05 - 0.20 mm) forms unaltered plate-shaped grains. The grains are pleochroic, brown to beige, and in some cases contain tiny zircon inclusions which produce pleochroic halos in the mineral. Untwinned plagioclase (0.01 - 0.20 mm) tends to be interstitial to biotite in biotite-rich zones. Small amounts of white mica are generally present and in places the plagioclase is completely altered to white mica. With quartz and carbonate, plagioclase forms a granoblastic mosaic. Anhedral quartz grains ranging from 0.01 to 0.7 mm in diameter (mostly between 0.1 and 0.3 mm) are free of inclusions. The carbonates were distinguished by a staining technique (Evamy, 1963). Both calcite and Fe-dolomite occur in anhedral grains (0.1 - 0.4 mm diameter). Some grains contain inclusions of quartz, feldspar or biotite. Pleochroic green to colourless, euhedral tourmaline grains

Table 18: Modal Analyses of Calcareous Rocks

	Rock 1	Rock 2	Rock 3	Rock 4
Quartz	19.9	24.6	29.1	20.7
Biotite	44.0	0.1	9.1	-
Plagioclase	17.7	16.6	18.3	17.1
Microcline	-	11.2	-	15.9
Calcite	6.8	16.9	11.5	6.9
Fe-dolomite	11.1	-	9.9	-
Hornblende	-	-	-	2.8
Actinolite	-	8.7	-	0.5
Diopside	-	20.2	-	9.6
Sphene	-	1.4	0.5	0.8
Epidote	-	-	-	17.3
Muscovite	-	-	16.8	-
Ti-Magnetite	-	-	4.3	-
Rutile	.1	-	-	-
Tourmaline	0.2	trace	0.1	trace
Pyrite	0.1	-	-	-
Pyrrhotite	-	trace	-	-
Apatite	0.1	0.3	0.4	0.2
Zircon	trace	-	trace	trace
Arsenopyrite	trace	-	-	-

(0.03 mm long) are common. Rutile, apatite, and pyrite occur in amounts less than 0.1 modal percent. Trace amounts of arsenopyrite and zircon are also present.

Epidote-free calc-silicate paragneiss (Rock 2).

Rock 2 (sample number 182444-1B) is fine-to medium-grained with irregular interlayering (1 - 5 mm) of laminae composed mainly of quartz, microcline, plagioclase, and calcite with laminae composed mainly of calc-silicates, quartz, and feldspar. Calcite, quartz, and feldspar in the quartzofeldspathic layers are generally more coarse-grained (1 - 5 mm) than they are in the calc-silicate-rich zone (0.2 - 0.5 mm). The long dimensions of the larger quartz and microcline grains, and those of a large proportion of the calc-silicate grains are parallel or subparallel to compositional layering. Overall, the rock is extremely fresh-looking, with a striking granoblastic mosaic texture with straight-line grain boundaries present in the quartzofeldspathic portions. A modal analysis is presented in Table 18. Alteration of plagioclase to white mica is much less prominent than in Rock 1. Twinning is absent. Grain size ranges from 0.3 to 0.7 mm. Microcline occurs as unaltered anhedral grains (0.5 - 1.5 mm). Grid twinning is very well-developed. Calcite occurs as irregular anhedral grains (.5 - 3 mm) and as aggregates of smaller grains. Quartz and feldspar are common inclusions. Colourless diopside (1 - 2 mm) ranges from subhedral

grains free of inclusions to spongy poikiloblasts which include quartz, plagioclase, microcline, actinolite, and calcite in varying amounts. Some grains are slightly altered along the margin to an amorphous brownish material (chlorite?). Actinolite exhibits very faint pleochroism (pale green to very pale green). Quartz and carbonate are commonly present as inclusions but generally the amphibole contains fewer inclusions than does diopside. Grain size of actinolite ranges from 0.4 - 1.5 mm. Cross sections are commonly euhedral. Sections parallel to cleavage are subhedral. Sphene is present as well-rounded to subhedral grains (0.1 - 0.3 mm). Some grains are slightly pleochroic, brown to paler brown. A few grains of biotite (0.1 mm) occur in the rock. Pleochroic brown to pale beige, the mineral is similar to that in Rock 1. Small amounts of anhedral apatite, tourmaline, pyrrhotite and zircon are also present.

Calcareous quartzofeldspathic psammite (Rock 3).

Rock 3 (sample number 130469) is very fine-grained to fine-grained (0.1 - 0.5 mm) with discontinuous compositional layering (2 - 4 mm) resulting from transposition of bedding during D_1 . Mica-rich zones with plagioclase and quartz are interlayered with quartz-carbonate zones. The rock is unaltered and lacks reaction textures or pseudomorphs. Most of the mica in the rock defines a preferred orientation (S_1). However, a large proportion

(20% - 30%) of the mica is at various angles to S_1 . Most mica grains are lath-shaped (cross-sections of plates). A modal analysis of Rock 3 is presented in Table 18. Quartz occurs as equant anhedral grains 0.1 - 0.4 mm in diameter. Plagioclase (0.1 - 0.2 mm) is present partly as equant anhedral grains and partly as irregularly-shaped material interstitial to quartz and mica. Patchy extinction indicates compositional variation. If the strain were the cause of the irregular extinction, quartz would also show undulatory extinction. Quartz does not. Many of the grains contain cores partly altered to white mica. Euhedral plates of muscovite (0.1 - 0.4 mm, long dimension) cut across all other minerals. Only the larger grains contain inclusions (quartz). Muscovite is also present in the form of aggregates, of finer grain size than the plates. Generally these grains are oriented parallel to S_1 . Pleochroic biotite (0.1 - 5 mm), dark olive brown to pale beige, occurs as anhedral to subhedral grains. This mineral has a cloudy appearance but no clear evidence of alteration. Tiny (.001 mm) zircons produce pleochroic halos. Anhedral grains of ferrodolomite occur all over the thin section, associated with calcite in some cases, with biotite and muscovite in others. Inclusions, usually of calcite, are rare. Ferroan calcite makes up most of the carbonate-rich zones in the rock. Anhedral grains, some equant, some with amoeboid shapes,

range from 0.2 to 1.0 mm. Small grains (0.05 - 0.10 mm) of Fe-Ti oxide are scattered randomly through the rock, generally as discrete grains, rarely as inclusions. Tourmaline, sphene, apatite, and zircon are common accessories. The well-rounded habit and small size (0.01 - 0.02 mm) of the sphene grains suggest they may be relict detrital grains.

Epidote-bearing calc-silicate paragneiss (Rock 4).

The fine-grained granoblastic texture, even grain-size and lack of a preferred orientation distinguish Rock 4 (sample number 16770) texturally from the calcareous psammite. D₂ minor folds are outlined by thin layers (0.5 - 1.0 mm) of very fine-grained epidote. Although irregular grain boundaries are common, the overall impression is that the rock is approaching quite closely textural equilibrium. Zoned plagioclase and the presence of two amphiboles suggests the possibility that chemical equilibrium was not attained in the rock. Table 18 includes a modal analysis. Small amounts (5%) of alteration are present in plagioclase. Along the grain boundaries of diopside and some of the epidote, alteration to Fe oxide and possibly chlorite has occurred. This alteration is attributed to weathering because similar features were observed at a more advanced stage in a deeply weathered specimen of the epidote-bearing paragneiss. Quartz (0.3 - 0.5 mm) occurs as

equant anhedral grains exhibiting weak undulatory extinction. Grains of microcline (0.1 - 0.5 mm) have an equant anhedral habit and are generally totally grid-twinned. Plagioclase (0.1 - 0.6 mm, long dimension) commonly has very irregular grain boundaries and is in places twinned (albite type). White mica alteration occurs along fracture cleavage and twin planes and typically is concentrated in the central portion of the grain. Calcite (0.1 - 1.0 mm, long dimension) is present in a variety of forms ranging from equant grains with straight boundaries to those with cusped boundaries. Where calcite occurs with diopside, amphibole, and epidote, it is embayed by these minerals and included within them. Faintly pleochroic, pale yellow to colourless, epidote is present as discrete grains and as fine-grained aggregates. Grain size varies from 0.1 to 1.0 mm. It is subhedral or anhedral, and exhibits a variety of anomalous high order interference colours. Pale-green non-pleochroic diopside is present as grains free of inclusions and as spongy poikiloblasts containing inclusions of quartz, microcline, amphibole and epidote. The alteration on the grain boundaries penetrates the mineral in the form of finger-shaped channelways 1 - 2 microns long. Most of the amphibole in the rock is pleochroic pale green to pale blue-green and has a maximum extinction angle of 20° . The mineral is not altered. Subhedral to euhedral grains (0.2 - 1.0 mm, long dimension) exhibit first order yellow interference colours. Parts

of some of the grains, both on the rim and in the core, are composed of colourless amphibole. The extinction angle and birefringence do not change but there are chemical differences between the two amphiboles (see Table 6, Chapter 4). The contact between the two is sharp in some cases and gradational over several microns in others. Small (0.05 - 0.2 mm) anhedral grains of sphene, slightly pleochroic brown to paler brown, anhedral grains of apatite, and blue-green tourmaline occur in the rock. Two or three grains of Fe-Ti oxide are present as inclusions in epidote but otherwise the rock is devoid of opaque minerals.

Pelitic schist (Rock 5). Rock 5 (sample number 110969) is a fine-grained porphyroblastic chlorite-muscovite schist. A very pronounced S_1 defined by micas, ilmenite plates and the long dimensions of quartz grains is present. Some chlorite and most biotite are oriented at various angles to S_1 . Compositional layering is not present. S_1 is crenulated (D_4). The rock is unaltered and there are no features indicative of textural disequilibrium. Table 19 includes a modal analysis. Elliptical quartz grains (0.05 - 0.3 mm, long dimension) are evenly distributed in the rock. Undulatory extinction is present in some grains. Muscovite occurs as very fine-grained (short dimension, 2 - 5 microns) lepidoblastic aggregates oriented parallel to S_1 . Pleochroic chlorite (pale green to green) is both

Table 19: Modal Analyses of
Pelitic Rocks.

	Rock 5	Rock 6
Quartz	30.0	22.2
Muscovite	36.1	15.8
Biotite	3.0	11.5
Chlorite	23.0	7.2
Plagioclase	-	26.5
Garnet	2.6	.4
Chloritoid	3.3	-
Staurolite	-	3.0
Kyanite	-	7.9
Ilmenite	1.7	2.6
Ti-Magnetite	-	2.5
Pyrite	trace	-
Tourmaline	trace	.1
Apatite	-	-
Zircon	trace	trace

parallel and oblique to S_1 . The mineral exhibits anomalous purple to grayish-purple interference colours and positive elongation. Grain size ranges from 0.1 - 0.3 mm, long dimension. Commonly in the form of rounded grains (0.1 - 0.3 mm) oriented with the C axis parallel to S_1 , biotite is also pleochroic (beige to medium brown). The large grains exhibit deformation bands. Anhedral garnets (1 - 2 mm dia.) contain inclusions of quartz and ilmenite. In the cores of some of the larger garnets there is preserved a much finer-grained fabric than that present in the matrix outside the garnets. Chloritoid occurs as large (4 - 6 mm long) porphyroblasts containing quartz and ilmenite inclusions. Blue-gray to gray pleochroic grains oblique to S_1 with a subhedral to euhedral habit are typical of this rock. Chloritoid overgrows the S_1 presently defined in the matrix of the rock, but in the larger porphyroblasts there is preserved an earlier finer-grained fabric which is at an angle of 50° to the S_1 in the matrix. The degree of flattening in the older fabric in the chloritoid suggests it postdates the older fabric preserved in the garnets. Ilmenite is present in the form of small blebs and plates (0.01 - 0.1 mm, long dimension). Pyrite (0.2 - 0.3 mm) is also present. Tourmaline and zircon are common accessories.

Pelitic schist (Rock 6). Compositional layering (10 - 20 mm) is prominent in this sample (sample number

2370). Contacts are sharp in some cases, gradational in others. Layers composed almost entirely of lepidoblastic muscovite or granoblastic plagioclase are the most distinctive. Porphyroblasts of staurolite, kyanite, and garnet are not restricted to any particular layer. Micas and Fe-Ti oxide laths define S_1 approximately parallel to compositional layering. An unusual feature of this rock is the coexistence within a field 2 mm in diameter of staurolite, kyanite, biotite, chlorite, and garnet, with quartz muscovite and Fe-Ti oxide. The proportion of muscovite is lower and that of biotite higher than in Rock 5 (compare modal analyses in Table 19). Except for white mica in some plagioclase grains, the rock shows no indications of alteration.

Quartz grains or aggregates of grains 2 - 4 mm in diameter are present in this rock. S_1 is deformed around them. Grain boundaries within the granules are sutured and undulatory extinction is prominent. Quartz in the rest of the rock is finer-grained and less intensely strained. Muscovite occurs as fine-grained aggregates oriented parallel to S_1 in layers composed largely of this mineral. Where it is less abundant the preferred orientation is not as well developed. Biotite is pleochroic, medium brown to pale beige, and generally oriented at various angles to S_1 . Grain size ranges from 0.5 to 1.0 mm and habit is euhedral to subhedral. Length fast and exhibiting gray interference colours,

the chlorite in Rock 6 is a magnesian variety. The mineral occurs in small (0.1 - 0.3 mm) grains. The pleochroic colour scheme is pale green to colourless. Plagioclase is abundant. Where this mineral composes most of the rock, a polygonal mosaic texture has developed. Alteration to white mica is less prominent in most of the grains in this rock than in plagioclase in rocks already described. Zoning is much more clearly defined, however. Commonly the outer one-sixth of a grain has a distinct composition (more calcic) from that of the rest of the grain. In other grains, the compositional change is continuous from the centre of the grain to the core. Grain size is variable (0.2 - 0.8 mm). Inclusions of quartz, white mica, and opaque minerals are common. Albite twinning occurs in many grains. Grain boundaries in some portions of the plagioclase-rich layers are extremely irregular, suggesting the possibility that polygonization as a result of strain has occurred. Anhedral to euhedral porphyroblasts of staurolite (0.1 - 0.5 mm, long dimension) exhibit a strong yellow to colourless pleochroism. Quartz and Fe-Ti oxide are the only inclusions found in this mineral. Kyanite porphyroblasts, subhedral to euhedral, range in size from 0.5 to 1.0 mm (long dimension). Garnets (1 - 2 mm) are not common in Rock 6. Where present, they are colourless, slightly elongate parallel to compositional layering, and contain fewer inclusions than are found in

garnets from Rock 6. Opaque minerals have three habits in this sample: plates 0.05 - 0.2 mm long; irregularly shaped anhedral grains, 0.2 - 0.8 mm; and euhedral grains with rhombic outlines. Plates are most common in micaeous layers and anhedral grains are scattered all over the thin section. The euhedral grains are not common. Probe data indicates the plates are probably ilmenite, and the blebs are an intergrowth of at least two Fe-Ti oxides, and the rhombic grains may be magnetite. Blue-green tourmaline and apatite are present in small amounts in the rock.

Appendix 3

Mineral assemblages from thin sections examined for this study. (Samples outside the areas of Figures 2 and 3 are omitted.)

Index to Mineral Name Abbreviations

MINERAL	CODE
ACTINOLITE	AC
ALLANITE	AL
AMPHIBOLE	AM
ANKERITE(Fe-DOLO.)	AK (FD)
ANTHOPHYLLITE	AN
APATITE	AP
BIOTITE	BI
CALCITE	CC
CARBONATE	CB
CHLORITE	CH
CHLORITOID	CT
CLINOZOISITE	CL
DIOPSIDE	DI
DOLOMITE	DO
EPIDOTE	EP
GARNET	GA
HEMATITE	HE
HORNBLENDE	HN
KYANITE	KY
MAGNETITE	MT
MICROCLINE	MI
MUSCOVITE	MU
OLIVINE	OL
OPAQUES	OP
PHLOGOPITE	PH
PLAGIOCLASE	PL
PSEUDOMORPH	Ps
PYRITE	PY
PYROXENE (DIOPSIDE)	PX
PYRRHOTITE	PO
QUARTZ	QZ
SCAPOLITE	SC
SERPENTINE	SE
SILLIMANITE	SL
SPHENE	SP
STAUROLITE	ST
TALC	TA
TOURMALINE	TO
TREMOLITE	TR
ZIRCON	ZI

I FLINTON GROUP

Bishop Corners Formation

1) Gray pelitic schist

Sample Number	Mineral Assemblage
0304I69	BI-CH-PL-QZ-HN-MU-CT-SP-OP-TO
050769	MU-BI-QZ-HN-GA-TO-PL-OP
050969	CT-ST-BI-MU-QZ-PL-OP
0510A69	QZ-MU-PL-BI-BA-CT-OP-TO-CH
091269	BI-MU-QZ-CH-OP-AP-SP-TO-ZI
100869	MU-QZ-BI-GA-OP-CH-TO-AP
100969	CT-PL-QZ-BI-ST-MU-OP-AP-ZI
1009I69	MU-QZ-BI-GA-CT-OP-ST-TO
110969	MU-QZ-CH-OP-GA-CT-BI-PL-ZI-SP
1205A69	QZ-MU-CT-BI-GA-CH-OP-PL-TO-AP
1206A69	GA-QZ-MU-CH-HN-CT-PL-PY-TO
1206AI69	QZ-MU-HN-EP-OP-BI-PL-TO-CH
P270	QZ-MU-PL-BI-OP-GA-TO-AP
P370	MU-QZ-OP-BI-Ps-GA-TO
P3A70	MU-QZ-BI-GA-Ps-ST-OP-TO
2370	QZ-MU-PL-ST-KY-BI-GA-CH-OP-TO-AP
2570	MU-QZ-CH-BI-EP-PL-GA-TO-AP-OP-SP
4870	QZ-MU-BI-PL-OP-GA-CH-AP-TO-EP
7870	MU-QZ-HN-OP-PL-BI-CB-TO
81A70	MU-OP-QZ-ST-KY-CH-TO-Ps
8870	QZ-EP-CH-MU-BI-OP-PL-AP
8970	MU-QZ-CH-EP-GA-OP
89A70	QZ-MU-CH-GA-OP-TO
89B70	MU-QZ-OP-CT-HN-GA-CH-BI-TO
90C70	MU-QZ-CH-GA-OP-PL-EP-TO-AP
9570	MU-QZ-CT-OP-BI-CH-GA-HN
95B70	QZ-MU-CT-CH-OP-SP-TO
95C70	GA-MU-QZ-CH-BI-OP-AP-CT-TO-ZI
98A70	MU-QZ-CT-GA-CH-OP-RI-TO-ZI
100B70	MU-PL-MI-OP-CB-QZ-CH
100C70	QZ-MU-CT-BI-GA-OP-CH-TO
101A70	QZ-MU-CT-OP-Ps-EP-TO
101D7E	QZ-MU-CH-BI-CT-GA-OP
111B7E	QZ-MU-BI-CH-EP-CB-OP-AP-TO
11670	MU-QZ-BI-OP-GA-Ps-TO
116A70	MU-QZ-BI-GA-CH-OP
116C70	MU-QZ-BI-CT-ST-CH-OP-BI
116D70	MU-QZ-BI-CT-GA-ST-Ps-OP-TO-ZI
14270	QZ-BI-PL-MU-ST-KY-CH-OP-AP-TO-EP
142A70	QZ-PL-UM-CH-OP-BI-GA-AP-EP-TO
142AI70	CH-BI-PL-QZ-EP
142B70	QZ-BI-CH-PL-KY-MU-OP-AP-EP
14470	QZ-MU-BI-ST-PL-CH-TO-AP-EP-OP
15470	PL-BI-GA-CH-QZ-ST-TO-AP-OP-PY-KY-ZI
24370	QZ-CH-OP-PL-BI-GA-TO-AP-ZI

Sample Number

Mineral Assemblage

244A70	CH-GA-QZ-OP-PL-BI-AP
244B7E	QZ-BI-CH-ST-MU-PL-KY-OP-ZI-TO
25271-1	BI-PL-QZ-CH-GA-ST-AP-TO-EP-ZI-OP
25271-2	PL-ST-OP-QZ-TO-BI-AP
252A71	BI-MU-PL-KY-QZ-CH-TO-OP-AP-PL-ST-ZI
183311-3	QZ-MU-ST-OP-BI-PL-CH-TO
1833-11-3A	MU-QZ-ST-PL-BI-OP-CH-TO
1833-11-3B	QZ-ST-PL-MI-MU-BI-OP-CH-TO-AP-ZI
1833-11-3C	MI-QZ-BI-EP-MU-OP-PL-AP-TO-SP
183312-3	PL-QZ-BI-ST-GA-MT-OP-TO-AP
183412-1	QZ-MU-BI-ST-GA-OP-TO
183412-1I	QZ-ST-MU-BI-OP-TO-AP-ZI
183413-2	QZ-MU-ST-BI-MT-PL-OP-TO-AP
183413-2A	QZ-BI-ST-OP-MU-PL-TO
183413-2B	QZ-GA-BI-ST-PL-OP-MU-TO-AP
183413-3	QZ-MU-GA-ST-BI-OP-TO-AP
183413-3A	QZ-MU-BI-GA-TO-ST

2) Black pelitic schist

0703A69	QZ-PL-HN-BI-GA-CH-CB-TO-OP-ZI-AP
0704A69	QZ-PL-BI-GA-ST-CH-OP-TO-AP
090369	PL-CH-QZ-BI-CB-OP
1307B69	BI-CH-GA-ST-PL-QZ-AP-TO-MT-OP
150269	PL-BI-OP-QZ-TO-HN-CH
180569	QZ-CH-HN-GA-BI-PL-OP-TO-AP-ZI
270	BI-MU-QZ-PL-EP-TO-OP
470	QZ-BI-CH-PL-GA-OP-EP
10B70	CH-OP-QZ-PL-BI-HN-GA-EP
1170	UM-QZ-CH-BI-GA-TO-EP-AP-CB-OP
24A70	GA-BI-QZ-CH
3970	PL-CH-OP-BI-HN-GA-EP-TO-AP
48C70	CH-QZ-BI-MU-OP-GA-TO-AP
5770	CH-QZ-BI-MU-OP-EP-TO-AP
63A70	BI-PL-QZ-GA-OP-TO-AP-CH
65B70	BI-PL-QZ-OP-HN-GA-CH-AP
73B70	BI-HN-QZ-MU-OP-GA-EP-TO
83I70	MU-CH-EP-PL-GA-OP
9770	HN-BI-MU-OP-QZ-PL-GA-CH-EP
99A70	QZ-BI-GA-OP-TO-CH-AP
99B70	CH-QZ-HN-OP-PL-TO-BI
11270	PL-QZ-BI-CT-OP-GA-TO-EP-CB-AP-OP
11570	PL-QZ-CT-BI-HN-GA-OP-CB-TO
12370	BI-QZ-PL-EP-OP
123B70	BI-PL-AM-EP-QZ
142I70	PL-BI-QZ-EP-CH-ZI-AP-TO-OP
142C70	HN-BI-PL-MU-QZ-OP-EP-AP-TO
14570	PL-BI-OP-QZ-GA-AP
145A70	PL-BI-QZ-GA-OP-AP-ZI

Sample Number	Mineral Assemblage
157A70	QZ-BI-OP-GA-AP-SP-TO
159A70	QZ-MU-MT-TO-EP-CH
17070	QZ-BI-PL-MU-OP-GA-AP-TO-ZI
21670	QZ-MU-BI-OP-EP-AP
22170	QZ-BI-PL-GA-SP-EP-AP-ZI-TO
22370	QZ-PL-MI-BI-MU-EP-OX-AP-TO
227A70	QZ-MI-EP-PL-BI-SP-MU
229B70	QZ-BI-EP-CC-AM
23370	QZ-MU-OP-AP-SP
182344-4	QZ-MU-BI-EP-OP-TO
182444-1	QZ-PL-AM-BI-CB-OP-TO-SP-ZI
144B70	QZ-MU-EP-OP-CB-TO
170A70	CC-BI-QZ-PL-MI-GA-MU-CH-SP-TO-AP
223C70	QZ-PL-HN-EP-BI-CH-SP-AP

Lessard Formation

- 1) Calcareous quartzofeldspathic psammite - epidote-bearing calc-silicate paragneiss.

130469	QZ-MU-PL-CB-BI-OP-AP-TO-ZI-FD
130569	QZ-BI-PL-MU-CH-CB-SP-TO-OP-ZI
1305A69	QZ-CB-CC-MI-PL-MU-BI-OP-TO-SP-FD
1305A69	QZ-PL-CB-UM-BI-OP-TO-AP-SP
170569	QZ-MI-PL-BI-EP-CB-OP-SP-TO-AP
200269	QZ-MI-CC-BI-PL-EP-MU-SP-AP-TO
210369	DI-EP-CC-SC-QZ-PL-AC-HN-MI-SP-TO-AP-OP
420369	QZ-MU-MI-PL-SP-CB-DI-CL-EP-AM-CC-SP-SC-OP
4203A69	DI-EP-MI-SC-PL-QZ-AM-BI-SP-CC-AP-ZI
4301B69	PL-HN-DI-SC-EP-SP-QZ-CB-AP
4570	QZ-CC-MU-BI-EP-TO-AP-SP
52A70	QZ-CB-MU-EP-OP-TO-CH
58B70	QZ-MU-BI-CB-PL-OP-TO
63C70	QZ-CB-MU-BI-PL-OP-TO-AP-EP
129D70	QZ-CC-PL-MU-BI-EP-TO-SP-OP-AP
13570	QZ-MU-CC-BI-EP-MI-TO-OP-SP
13770	QZ-CB-MU-MI-EP-BI-PL-OP-TO-SP
151D70	QZ-PL-BI-MU-CB-EP-OP-AP-TO-SP
16170	QZ-PL-BI-MU-EP-OP-TO-AP
16770	MI-PL-QZ-EP-CC-DI-AC-SP-AP-OP
17170	QZ-MI-BI-EP-AC-SP-AP-CH
185A70	BI-CC-MU-QZ-PL-OP
18670	QZ-CC-PL-UM-BI-EP-OP-TO
189C70	QZ-CB-BI-MU-EP-OP-TO
189B70	CC-QZ-MU-BI-MI-EP-PL-OP-TO-AP
196A70	
208B70	QZ-MI-EP-PL-DI-AM-SC-SP
21070	QZ-MI-CB-DI-EP-AM-OP-SP-AP
21270	QZ-MI-EP-CC-MU-BI-SP-AP-OP

21370 QZ-EP-MI-BI-AM-SP-OP
 21470 DI-EP-MI-QZ-AM-SP-CC-PL-OP-AP-TO
 216B170 QZ-MI-EP-DI-MU-BI-AM-SP-AP
 225A70 QZ-MI-EP-AC-CC-SP-TO-BI-AP
 22770 QZ-MI-PL-AM-EP-SP-SC-AP-BI-TO
 228B70 PL-EP-DI-MI-AM-QZ-SP-AP-CH-OP
 22970 QZ-EP-CC-MI-AM-DI-SP-OP-AP
 25171 MI-PL-QZ-AM-CB-EP-SP-BI-AP
 251A71 QZ-MI-CC-AM-EP-SP-OP-AP-TO
 251670 QZ-EP-MI-BI-CB-PL-MU-SP-AP-TO
 251C70 QZ-MI-PL-EP-CB-BI-MU-OP-SP-AP-TO
 261A71 QZ-MI-EP-PL-AM-CB-SP-OP-BI
 182344-2 QZ-EP-AM-BI-PL-CB-SP-CH-SC
 18360200-2 QZ-EP-DI-PL-AM-SP-OP
 183620-5 MI-PL-AM-EP-DI-QZ-BI-SP-AP
 183821-1 QZ-MI-EP-AM-CB-SP-OP-AP
 183322-2A QZ-MI-EP-AM-DI-SP-CH-CB-OP-AP-ZI

2) Biotite carbonate schist - epidote-free calc-silicate paragneiss.

160469 QZ-CB-BI-PL-MI-MU-TO-SP-FD
 160569 BI-QZ-CC-PL-TO-OP-FD
 200669 QZ-MI-CC-BI-AC-OP-AP-ZI
 P470 QZ-BI-CC-PL-MU-TO-AP-ZI-OP-FD
 12170 BI-CC-QZ-AM-PL-EP-OP-SP-ZI-TO
 13170 BI-CB-QZ-PL-OP-CH-AP-ZI-EP
 13670 BI-QZ-MI-CC-PL-CH-AP-SP-ZI
 14770 QZ-CB-BI-AC-PL-CH-AP-ZI
 15370 QZ-CC-BI-AC-PL-MI-AP-SP-ZI-CH
 15870 AC-QZ-MI-PL-CB-CC-BI-SP-AP-ZI-OP
 158I70 QZ-PL-BI-CB-AC-SP-ZI-PY-AP
 16570 QZ-BI-CC-MI-PL-SP-TO-AP
 170670 QZ-CC-MI-DI-AC-TR-SP-OP-BI-AP
 17370 AM-MI-BI-QZ-SP-CB-CH-AP-ZI
 17570 AM-MI-CB-QZ-PL-SP-BI-OP-AP
 18070 CB-BI-QZ-PL-AM-AP-TO
 211A70 QZ-PL-BI-AM-MI-CB-SP-EP-AP-TO-OP-ZI
 211B70 MI-DI-CB-AM-QZ-SP-AP-BI-OP
 214A70 AM-MI-QZ-DI-BI-SP-EP-OP-ZI-AP
 24171 QZ-PL-BI-CB-AM-TO-AP-ZI
 253A71 BI-QZ-PL-CB-AM-PY-CH-AP-ZI
 25471 QZ-BI-PL-CB-AM-CH-AP-ZI-OP
 27671 QZ-CC-AM-PL-BI-MI-SP-AP-ZI-OP
 276A72 CC-QZ-SE-PL-SP-CH-ZI
 27772 DI-MI-AC-PL-SP-QZ-CC-ZI
 277A71 DI-MI-AC-PL-SP-CC-BI-AP-OP-ZI-AN
 277A172 AM-MI-QZ-PL-BI-CC-SP-CH-EP-ZI-AP
 182750 QZ-CC-BI-PL-SP-CH-ZI-OP
 182444-1B CC-QZ-MI-PL-DI-AM-SP-OP-BI-TO-ZI
 182444-1C PL-MI-AM-QZ-SP-BI-CB-AP-ZI
 182444-3 MI-DI-QZ-PL-AM-CB-SP-BI-OP-AP
 182444-4 BI-QZ-PL-AM-CC-AP-OP-ZI

Sample Number	Mineral Assemblage
18270	PB-BI-AM-GA-QZ-OP-AP-CH
252C71	PL-BI-MU-QZ-AP-EP-TO-OP-ZI
FBS70 (10770)	QZ-BI-PL-HN-MU-OP-AP-SP-TO-CB
185327-1	BI-HN-PL-GA-QZ-CH-OP

3) Quartzite, quartzite metaconglomerate, and mica-ceous quartzofeldspathic psammite.

PTCGA,B (150169)	QZ-MU-CH-OP-TO-CB-BI-ZI
010469	QZ-MU-BI-PL-EP-CH-CB-OP-TO-AP
0104A69	QZ-MU-BI-CB-EP-PL-CH-OP
070869	QZ-MU-BI-CH-OP
1201A69	QZ-MU-BI-CH-EP-OP-CB-PL
1201B69	QZ-BI-MU-PL-MI-EP-CH-OP-AP-TO
130769	QZ-MU-BI-CH-CB-PL-TO-AP-OP-SP
1307I69	MU-BI-CH-EP-CB-QZ-PL-TO-SP
170469	QZ-BI-MU-MI-PL-CB-AP-OP-TO-SP
1706A69	QZ-PL-MI-BI-MU-CH-TO-AP
1903AI69	QZ-MU-BI-MI-PL-CB-OP-TO-AP
200569	BI-MU-EP-CB-TO-QZ-CH-TR
3706C69	QZ-CH-BI-MU-TO-CB-SP-AP
3706B69	MU-QZ-BI-EP-OP-CH-TO-SP
420269	QZ-BI-EP-MI-SP-CB-ZI-CH-AP
4204A69	QZ-MI-EP-AC-BI-CH-SP-CB-TO-AP
4301A69	QZ-BI-EP-MI-CB-MU-OP
430269	QZ-EP-AM-MI-SP-CH-BI-TO-MI
4302A69	QZ-BI-EP-MU-CH-MI-SP-CB-PL
4302B69	QZ-BI-EP-CH-MI-PL-MU-SP-ZI
4303A69	MU-HN-EP-SP-QZ
370	QZ-MU-CB-EP-PL-TO-AP-OP
10A70	QZ-MU-BI-PL-EP-CH-OP-CB-TO-AP
13A70	QZ-MU-BI-ST-GA-CT-TO-SP-OP-EP-KY
2170	QZ-MU-BI-CB-CH-EP-TO-AP-OP
30A70	QZ-PL-SI-CH-EP-CB-AP-SP
3170	QZ-MU-EP-AP-TO-OP-CB
4470	QZ-BI-CH-MU-CB-TO-AP-OP
45A70	QZ-MU-EP-CH-BI-PL-CB-SP-TO-AP-OP
46C70	QZ-MU-OP-SP-TO
46D70	QZ-MU-EP-CH-BI-CB-OP-TO-AP
47B70	QZ-MU-CH-EP-CH-TO-OP
47D70	QZ-BI-EP-MU-PL-CB-OP-TO-AP
45470	QZ-MU-CH-BI-CB-OP-TO
60A70	QZ-PL-BI-UM-OP-CB-CH
100A70	QZ-MU-CH-PL-EP-CB-OP-BI-TO-AP
12770	QZ-BI-CB-MU-EP-CH-OP-TO-AP
13270	QZ-PL-BI-CB-TO-AP-OP-CH-EP
142D70	QZ-BI-MU-GA-EP-CH
129C70	QZ-MI-PL-BI-EP-MU-OP-CB-AP-TO-ZI
136A70	QZ-PL-BI-MU-SP-EP-CB-TO-AP
14870	QZ-BI-CH-CB-OP
151A70	GA-QZ-OP-MU-BI-EP-PL-TO-AP

182445-1	BI-PL-CC-QZ-AP-SP-TO-ZI-OP-CH
182445-5	BI-CC-PL-MI-QZ-AM-SP-OP-CH-ZI
182445-6	BI-QZ-CC-PL-AM-SP-AP-ZI-OP

Madoc Formation

1) Carbonate metaconglomerate

300769	CC-DO-QZ-MU-TA-OP
310469	BI-OP-QZ-MU-CB-ZI-TO
3104A69	DO-CC-QZ-MU-TA

2) Pelitic schist

2501E69	MU-QZ-CH-BI-OP-GA
2501A69	QZ-MU-CH-CB-OP
270569	QZ-MU-OP-GA-Ps-CH-TO
2705I69	QZ-MU-OP-CH-PL-GA-BI-CB-CC
2705AI69	BI-GA-QZ-TO-MU-CH-OP-PL-MI
2705B69	QZ-MU-CH-OP-MI-PL-CT-GA-CB-CC-TO
2705C69	QZ-MU-PL-MI-CH-OP-CT-BI-GA
2705CI69	QZ-CH-MU-PL-MI-OP-CT-GA-CH-CB-CC-ST
2705D69	QZ-MU-CH-PL-MI-TO-OP-GA-BI-KY-CB
2705EI69	QZ-MU-CH-OP-TO-CT
2705E69	QZ-PL-MI-MU-CH-OP-CT-CB-TO
2902A69	QZ-MU-TO-PL-OP-AP
300269	QZ-CH-MU-OP-SP
3002I69	QZ-CH-MU
3002A69	ST-QZ-OP-MU-TO
3002C69	ST-Ps-OP-MU-QZ-TO-GA-PL-MI
3002CII	CT-ST-OP-QZ-MU-CH-Ps-TO
3002CIII	CT-OP-CH-QZ-MU-Ps
300469	GA-BI-ST-CT-PL-MU-QZ-OP-CH
3004A69	QZ-MU-OP-TO
3004D69	CT-ST-BI-MU-QZ-OP-TO-ZI
3004E69	CH-BI-QZ-PL-PY-CB-OP-AP
3004F69	QZ-MU-PL-OP-BI
300669	QZ-MU-OP-PL-Ps-ST
310169	MU-QZ-OP-Ps-TO
3101AI69	GA-QZ-MU-CH-OP-TO
3101C69	MU-OP-QZ-PL-CH-Ps-KY-ST-TO-ZI
310269	OP-QZ-MU-BI-TO
3102A69	CT-ST-GA-BI-OP-QZ-MU
310369	ST-OP-QZ-MU-CH-Ps

Stewart Formation

2703A69	CC-DO-QZ-PL
2711A69	DO-CC-QZ-PL-MU-MI-OP-SP
2711F69	MU-OP-BI-QZ-PL-CB-Ps-TO-ZI
280169	CC-QZ
2801A69	HN-SC-CC-DO-QZ-OP-MU
2902B69	PL-CC-EP-MU-QZ-TO-OP-SP-CH
3001I69	CC-PY-PL-SP-MU

II HERMON GROUP

- 1) Metapyroclastic and metasedimentary rocks associated with the metavolcanic rocks.

0102AI69	QZ-BI-CH-EP-CB-PL-PY-TO
0102AII69	
030569	PL-BI-AK-CH-QZ-MU-OP-TO-AP
030669	MU-DO-CC-EP-QZ-AP-PY
0306A69	MU-PL-EP-DO-QZ-CC-OP-TO
100769	PL-BI-QZ-MU-CH-CB-OP-PY-AP-Ps
1006I69	QZ-MU-BI-CH-OP-CB-SP-TO
1201C69	QZ-PL-BI-CH-MU-EP-TO-AP-OP
1803B69	CB-QZ-PL-MI-BI-CH-MU-OP-EP
220269	QZ-PL-CB-BI-OP-PY-MU-TO-ZI
2203A69	BI-MU-QZ-PL-CB-OP-TO-ZI
220569	PL-MI-CC-MU-BI-SP-AP-PY
2206A69	QZ-BI-CB-PL-PY-AP
230169	PL-BI-CC-SP-OP
2602B69	MU-QZ-BI-CB-TO-AP-SP-PL-OP
2605C69	MU-QZ-PL-BI-OP-TO-AP
2701B69	BI-MU-QZ-MI-PL-TO-CB-SP-AP-OP
2701C69	BI-MU-QZ-PL-CB-TO-OP
2704A69	CB-MU-QZ-PL-BI-SP-OP
270669	BI-QZ-PL-CB-TO
290169	CC-QZ-MU-EP-OP
2908B70	BI-MU-QZ-PL-OP-CB-TO
23A70	BI-MU-QZ-CH-EP-CB
46A70	MU-QZ-EP-BI-PL-PY-OP
4770	QZ-MU-PL-BI-EP-CB-OP
73C70	QZ-MU-TO-BI-EP-CH-OP-AP
8270	PL-MU-EP-OP-QZ-CB-BI
85C70	MU-QZ-PL-CH-OP-BI-EP
86A70	PL-OP-MU-CH-QZ-BI-EP-AP
10570	EP-OP-PL-CH-BI-QZ-EP-MU
105A70	PL-MU-QZ-OP-CH-BI
124A70	MU-QZ-BI-PL-OP-CB-TO-ZI-AP
128B70	AM-BI-CH-MU-SP-EP-TO-QZ
189D70	QZ-AM-CB-BI-OP-EP-CH
19170	MI-PL-QZ-CH

191A70	HN-PL-CB-BI-QZ-GA-OP-AP
216C70	PL-HN-QZ-SP-EP-AP
183620-1	PL-AM-BI-OP-EP-AP-SP-MI
183821-3	PL-MI-EP-AM-QZ-BI-SC-SP-CB-ZI-TO-AP

2) Metavolcanic rocks

03011169	PL-HN-BI-CB-CH-QZ-OP-MU-EP-ZI-TO
0302A69	PL-BI-QZ-CB-CH-HN-MU-EP-OP-TO
040469	BI-CH-HN-EP-QZ-PL-AP-OP
140169	AM-CH-PL-QZ-CB-ST-EP
180369	HN-CH-BI-QZ-PL-CB-OP
5170	PL-CH-AM-CC-QZ-OP
2007A69	HN-PL-QZ-MU
200869	AM-BI-PL-QZ-OP-SP-EP-AP-CB
2008C69	AM-PL-CH-QZ-TO
2302A69	BI-PL-QZ-AP-CB-OP
2602A69	MI-BI-QZ-PL-MU-CB-AP-OP-TO
270969	MI-BI-QZ-MU-TO-CB
271069	BI-CC-QZ-PL
2803B69	PL-MU-CB-QZ-TO
3104B69	AM-CH-BI-EP-OP-PL
320369	QZ-MI-MU-BI-CB
3202A69	MI-MU-QZ-OP-PL
320269	BI-CH-EP-MU-OP-QZ-PL-CB-TO
3204A69	BI-CH-EP-QZ-PL-OP-SP-ZI
420569	HN-QZ-PL-AP-OP
430569	PL-HN-BI-QZ-OP-AP
123A70	HN-QZ-TO-OP
229A70	PL-AM-BI-OP-CB
24971	HN-PL-EP-OP-AP-QZ-SP
252B70	AM-BI-TO-EP-OP-AP-QZ
183721-2	HN-PL-OP-QZ-AP-ZI

3) Serpentinite and Marble

031169	QZ-DO-CC-MU-TR-CH-SP
210169	SE-PY-PO-PX
2102A69	AM-CH-SE-OP-AP
2102B69	DO-CC
2301A69	CC-OP-MI-QZ-MU
2503A69	QZ-TR-AC-CC-PH-OP-PY
17770	AC-TA-CH-SE-PL
19670	CC-QZ-MU-EP-PH-AP-TO-DO
22870	AM-TA-SE-OL-OP
216D70	AM-CH-OP
25371	AN-CH-TR-PX-OP-PL
183620-2A	CC-DI-EP-QZ-PL-SP-CH-AM-AP
183822-2	PX-AM-CC-CH-BI-OP
182445-6A	AN-TA-CB-CH-OP

III PLUTONIC ROCKS AND PEGMATITES

190169	QZ-BI-EP-HN-SP-PL-MI-AP
2103A69	MI-QZ-PL-GA-MU-MT-OP
420169	PL-QZ-BI-EP-SP-AP-OP-CH-TO
430469	MI-QZ-BI-EP-MU-AP-TO-OP
4305A69	QZ-PL-BI-MU-EP-AP-CB-CH
430669	PL-MI-QZ-BI-MU-OP-EP-AP-ZI-SP
12870	PL-AM-BI-QZ-CH-SP-EP-AP-CB-OP
131A70	PL-QZ-BI-MU-AP-CH-CB
149170	PL-AM-QZ-BI
15070	BI-PL-HN-QZ-OP-SP-AP-CH-TO-CB
150170	BI-PL-QZ-HN-SP-CB-TO
15170	PL-QZ-BI-MI-EP-AM-TO-SP-AP-CB-ZI
159170	PL-BI-QZ-EP-SP-CB-TO-AP-AC
165A70	PL-QZ-BI-EP-CH-AP-MU-SP-AL-CB-ZI
20770	MI-QZ-BI-EP-OP
21570	PL-BI-QZ-EP-SP-MU-CH-AP-CB-ZI-OP
215A70	PL-QZ-BI-EP-SP-MU-AP-CH-ZI
22670	PL-QZ-BI-SP-EP-CH-CB-AP
22770	PL-MI-QZ-BI-EP-SP-AP-CH
23070	MI-PL-QZ-BI-GA-AP-CH-ZI
183621-1	PL-MI-BI-QZ-AM-CH-SP-AP-ZI
183817-1	PL-QZ-BI-SP-EP-AP-TO

Appendix 4

UTM coordinates of fieldwork stations
which occur within the boundaries of
Figures 2 and 3.

FIELD WORK 1969

TRAVERSE NUMBER	STATION	UTM CO- ORDINATES
--------------------	---------	----------------------

(1)	1	280600
	2	283593
	3	284587
	4	283572

(2)	1	276602
	2	275602
	3	273601
	4	271601
	5	268603
	6	262601
	7	259600
	8	259599
	9	261511
	10	261511

(3)	1	281607
	2	282608
	3	284608
	4	271601
	5	268603
	6	262601
	7	259600
	8	259599
	9	261511
	10	289611
	11	293612
	12	291612
	13	293613
	14	292613
	15	292615
	16	292616
	17	293615
	18	294614
	19	294613

(4)	1	291611
	2	290611
	3	289611
	4	288611
	5	287610
	6	286610

(5)	1	285611
	2	285610
	3	283609
	4	282608
	5	282608
	6	281608

FIELD WORK 1969

TRAVERSE NUMBER	STATION	UTM CO- ORDINATES
--------------------	---------	----------------------

(5)	7	283611
	8	284610
	9	282610
	10	282609
	11	282609
	12	283609
	13	283609
	14	283609
	15	283609
16	284610	

(6)	1	284611
	2	284611
	3	285617
	4	286616
	5	285613

(7)	1	285587
	2	291593
	3	296597
	4	297598
	5	299599
	6	300601
	7	303604
	8	305606
	9	307607

(8)	1	297598
	2	299600
	3	300603
	4	300604
	5	300604

(9)	1	285611
	2	286611
	3	288612
	4	289612
	5	289612
	6	290612
	7	291614
	8	291614
	9	289614
	10	291616
	11	291617
	12	290617
	13	286615

FIELD WORK 1969

TRAVERSE NUMBER	STATION	UTM CO- ORDINATES
(10)	1	285612
	2	286612
	3	286612
	4	286612
	5	286613
	6	287614
	7	288614
	8	288614
	9	289614
	10	289615
(11)	1	286610
	2	286609
	3	285609
	4	280606
	5	280606
	6	280604
	7	280603
	8	280602
	9	280602
(12)	1	279599
	2	280600
	3	282601
	4	282602
	5	279602
	6	279602
(13)	1	259520
	2	260525
	3	262528
	4	262529
	5	264533
	6	265534
	7	267539
(14)	1	278557
	2	273554
	3	271545
	4	270543
(15)	1	271533
	2	272532
	3	273532
	4	272530
	5	273529
	6	269518
	7	273527

FIELD WORK 1969

TRAVERSE NUMBER	STATION	UTM CO- ORDINATES
(15)	8	273522
	9	280522
(16)	1	265512
	2	266513
	3	267514
	4	268515
	5	268516
	6	269518
	7	271519
	8	273522
	9	267513
(17)	1	274512
	2	279511
	3	270515
	4	270513
	5	268509
	6	270507
(18)	1	284593
	2	283593
	3	283592
	4	284595
	5	285596
(19)	1	265482
	2	264486
	3	262490
	4	262492
	5	261494
	6	261495
	7	260496
	8	258506
(20)	1	240455
	2	240455
	3	241454
	4	242452
	5	243452
	6	244452
	7	244450
	8	238453
(21)	1	231407
	2	233408
	3	234408
	4	235408

FIELD WORK 1969			FIELD WORK 1969		
TRAVERSE NUMBER	STATION	UTM CO- ORDINATES	TRAVERSE NUMBER	STATION	UTM CO- ORDINATES
(21)	5	228406	(29)	3	082329
(22)	1	038287		4	093332
	2	038285		5	014337
	3	039287		6	011339
	4	037287		7	015333
	5	039290	(30)	1	026323
	6	036282		2	026323
(24)	1	016296		3	024328
	2	019298		4	023329
	3	118298		5	022327
	4	014300		6	021327
	5	017302		7	021325
(25)	1	014306		8	024319
	2	012303	(31)	1	017315
	3	011306		2	018317
(26)	1	010306		3	017317
	2	009305		4	016318
	3	010302		5	016322
	4	009306	(32)	1	028317
	5	018307		2	018317
(27)	1	024306		3	015317
	2	023307		4	014314
	3	021307	(37)	1	118343
	4	024308		2	121344
	5	025311		3	125345
	6	026310		4	133347
	7	027309		5	143350
	8	028307		6	151352
	9	029309	(42)	1	207374
	10	028312		2	208373
	11	035316		3	209370
(28)	1	027314		4	211368
	2	020298		5	214369
	3	021297			
	4	031301			
	5	033303			
	6	028303			
	7	029304			
	8	025305			
	9	022303			
(29)	1	067323			
	2	070323			

FIELD WORK 1970

STATION	UTM CO-ORDINATES
1	271527
2	270524
3	270524
4	269520
5	267517
6	269518
7	274531
8	273533
9	269534
10	267534
11	266532
12	265530
13	265537
14	263534
15	279525
16	278527
17	274527
18	275529
19	282530
20	288528
21	284569
22	281565
23	277562
24	278565
25	279568
26	279570
27	281575
28	283572
29	286571
30	289573
31	290573
32	287561
33	287575
34	289577
35	290579
36	289580
37	287581
38	283584
39	291593
40	289567
41	291572
42	292572
43	293574
44	292575
45	294577
46	297602
47	295608
48	295607
49	302605
50	269558
51	271564
52	273566

FIELD WORK 1970

STATION	UTM CO-ORDINATES
53	273571
54	272570
55	270553
56	269549
57	268544
58	267539
59	266544
60	267546
61	268549
62	268552
63	261526
64	263525
65	262521
66	262518
67	260519
68	261521
69	262531
70	263533
71	258515
72	260514
73	261511
74	313566
75	311571
76	311579
77	306586
78	281609
79	283610
80	282609
81	280606
82	280604
83	281598
84	285602
85	285599
86	285598
87	275600
88	275602
89	275600
90	273594
91	276594
92	275585
93	268575
94	275590
95	276590
96	276592
97	283587
98	277594
99	281588
100	279612
101	277608
102	283605
103	284607
104	284608

FIELD WORK 1970

STATION	UTM CO-ORDINATES
105	286607
106	285606
107	259509
108	265510
109	270509
110	282507
111	285596
112	295597
113	289600
114	288592
115	299615
116	299617
117	290540
118	285547
119	283544
120	278543
121	274542
122	271534
123	261516
124	262521
125	261525
126	259498
127	257496
128	255492
129	261495
130	263496
131	268497
132	261491
133	259488
134	258485
135	259482
136	260481
137	260487
138	296627
139	299629
140	300633
141	252488
142	255491
143	253492
144	255493
145	258495
146	257492
147	264490
148	262491
149	257488
150	252483
151	248472
152	251469
153	252469
154	250475
155	240460

FIELD WORK 1970

STATION	UTM CO-ORDINATES
156	244460
157	247462
158	248462
159	245469
160	250467
161	246466
162	253479
163	253479
164	255478
165	257478
166	258477
167	234438
168	237438
169	238439
170	241438
171	237447
172	239447
173	241446
174	244449
175	244450
176	218488
177	220484
178	225486
179	272509
180	270505
181	266501
182	264504
183	111337
184	113337
185	115336
186	116335
187	115332
188	117340
189	123342
190	118339
191	119336
192	026314
193	025316
194	130341
195	133337
196	130336
197	022333
198	023332
199	025333
200	025332
201	018333
202	018337
203	018326
204	028325
205	014317
206	027329

FIELD WORK 1970

STATION	UTM CO-ORDINATES
207	233429
208	234427
209	236429
210	237430
211	238430
212	233442
213	237440
214	241440
215	231439
216	229409
217	233413
218	234416
219	234409
220	234410
221	234411
222	236414
223	226402
224	227404
225	229403
226	224403
227	223395
228	226396
229	229398
230	234422
231	235422
232	235427
233	233427

FIELD WORK 1971

STATION	UTM CO-ORDINATES
240	274509
241	267509
242	266508
243	265507
246	269527
247	268519
248	225472
249	232451
250	231446
251	232441
252	253471
253	249473
254	268499
255	305586
256	298582
257	295585
261	232418
262	235418
263	301479

FIELD WORK 1971

STATION	UTM CO-ORDINATES
275	244457
276	242445
277	243448
1839-1	181380
2038-1	208383
2137-7	212371
2137-8	214371
2137-9	215372
2344-1	244445
2344-2	245444
2344-3	246444
2344-4	247444
2345-1	237452
2345-2	239451
2444-1	240452
2444-2	241452
2444-3	241453
2444-4	242450
2445-1	245454
2445-2	244455
2445-3	243454
2445-4	243450
2445-5	243453
3311-1	114332
3311-2	113239
3311-3	118234
3312-1	127338
3312-2	125337
3312-3	120335
3312-4	124339
3412-1	129339
3412-2	130340
3412-3	125346
3412-4	125344
3412-5	127345
3413-1	130342
3413-2	129340
3413-3	135342
3413-4	131344
3413-5	136345
3413-6	133346
3414-5	141345
3618-1	179363
3618-2	184363
3618-3	185366
3619-1	190366
3619-2	190366
3619-3	185366
3620-1	206370
3620-2	200360
3620-3	208370

FIELD WORK 1971

STATION	UTM CO-ORDINATES
3620-4	201368
3620-4a	209367
3620-5	202364
3621-1	213369
3720-1	207375
3720-2	208373
3721-1	214370
3721-2	215375
3721-3	214372
3721-4	212374
3721-5	211375
3721-6	212377
3717-1	179378
3817-1	175387
3821-1	218384
3821-2	216383
3821-3	218379
3821-4	214380
3821-5	219388
3821-6	220387
3822-1	219387
3822-2	221387
3922-1	223394
5327-1	273520
5327-2	272529
5327-3	273532

LEGEND

GRENVILLE SUPERGROUP AND ASSOCIATED PLUTONIC ROCKS (HELIKIAN)

18 Metamorphosed ultramafic rocks

17 Pegmatite

INTRUSIVE CONTACT

FLINTON GROUP (10-14)

LESSARD FORMATION (13, 14)

14a. epidote-free biotite-carbonate schist; 14b. epidote-free calc-silicate paragneiss

MYER CAVE FORMATION

15 Dolomitic marble

13 Calcareous quartzofeldspathic psammite and metaconglomerate; 13a. epidote-bearing calc-silicate paragneiss; 13b. plutonic pebble metaconglomerate

BISHOP CORNERS FORMATION

12 Quartzofeldspathic psammite and metaconglomerate

11 Quartzite metaconglomerate and quartzite; 11a. quartzite

10a Pelitic schist

UNCONFORMITY

METAMORPHOSED PLUTONIC ROCKS

9a. biotite quartz monzonite

ADDINGTON PLUTON

8a. biotite granodiorite and quartz diorite; 8b. biotite quartz monzonite

NORTHBROOK AND ELZEVIR PLUTONS

7a. coarse grained hornblende diorite; 7b. coarse grained contact phase

INTRUSIVE CONTACT

HERMON GROUP (1-6)

5a. calcitic marble; 5b. dolomitic marble; 5c. quartzite

6a. calcareous schist; 6b. pyritic phyllite

TUDOR FORMATION (1-4)

4 Sodic metavolcanics; locally fragmental

3 Fine-grained massive gray-weathering metavolcanics; locally amygdaloidal

2 Chlorite-amphibole schist, hornblende-biotite schist, gneissic amphibolite, locally fragmental; 2a. hornblende-biotite quartzofeldspathic schist

1 Massive fine-grained amphibolite; locally pillowed, fragmental

Geological boundary; defined, approximate, assumed

Area of outcrop

Road, highway

Geology by Peter H. Thompson, 1969, 70, 71.

Base-map from Forest Resources Inventory Maps, Ontario Department of Lands and Forests.

Numbers around edge of the map refer to the one thousand metre Universal Transverse Mercator Grid, Zone 18.

BISHOP CORNERS

1 2 3 4 5 6 7 8 9 10 11 12 13 14 15 16 17 18 19 20 21 22 23 24 25 26 27 28 29 30 31 32 33 34 35 36 37 38 39 40 41 42 43 44 45 46 47 48 49 50 51 52 53 54 55 56 57 58 59 60 61 62 63

2 of .

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SKOOTAMATTA

RIVER

SKOOTAMATTA

FLINTON

CREEK

A detailed map of the Skootamatta area, showing a large body of water with numerous small islands and a complex network of channels. The map is labeled with 'SKOOTAMATTA' in the upper right. It features a grid of latitude and longitude coordinates along the edges. The map includes various labels such as '8a', '11', '13a', '13b', '16', '17', '2a', '5a', '5b', '7', '9a', and '10'. There are also several small crosses marked on the map. The map is oriented with North at the top.

1 of

GEOLOGY OF THE MADOC AREA

MADOC AREA

LEGEND

9 Undifferentiated Paleozoic

UNCONFORMITY

GRENVILLE SUPERGROUP
(HELIKIAN)

FLINTON GROUP (5-8)

STEWART FORMATION (7-8)

8 Graphitic calcite marble; 8a, marble breccia and micaceous calcite marble

6 Dolomitic marble

MADOC FORMATION (5-6)

5b Pelitic schist, phyllite

5a, dolomitic marble metaconglomerate;
5b, calcareous phyllite

UNCONFORMITY

HERMON GROUP (1-4)

4a, metamorphosed potassic feldspar porphyry
4b, felsite breccia, meta-ignimbrite

MOI

FIGURE 3

GEOLOGY OF THE MADOC AND BLACK RIVER AREAS SOUTHEASTERN ONTARIO

BLACK RIVER AREA

FIGURE 3

GEOLOGY OF THE MADOC AND BLACK RIVER AREAS SOUTHEASTERN ONTARIO

2 of .

AREA

BLACK RIVER AREA

K RIVER AREAS

RIVER AREA

LEGEND

GRENVILLE SUPERGROUP AND ASSOCIATED PLUTONIC ROCKS (HELIKIAN)

Metamorphosed ultramafic rocks

37 Pegmatite

INTRUSIVE CONTACT

FLINTON GROUP

LESSARD FORMATION (13)

Biotite-carbonate schist; 14a, epidote-free calc-silicate paragneiss

Dolomitic marble

Calcareous quartzofeldspathic psammite metaconglomerate; 13a, epidote-bearing calc-silicate paragneiss; 13b, plutonic pebble metaconglomerate

MADOC

LEGEND

Undifferentiated Paleozoic

UNCONFORMITY

GRENVILLE SUPERGROUP (HELIKIAN)

FLINTON GROUP (5-8)

STEWART FORMATION (7-8)

Graphitic calcite marble; 8a, marble breccia and micaceous calcite marble

Dolomitic marble

MADOC FORMATION (5-6)

Pelitic schist, phyllite

5a, dolomitic marble metaconglomerate;
5b, calcareous phyllite

UNCONFORMITY

HERMON GROUP (1-4)

Salic metavolcanics: felsite breccia, meta-ignimbrite;
4a, metamorphosed potassic feldspar porphyry

Dolomitic marble

Calcareous metagraywacke

TUDOR FORMATION

Fine-grained gray- or green-weathering metavolcanics,
commonly amygdaloidal, fragmental

Geological boundary: (defined, approximate, assumed)

Area of outcrop

Road, highway

Geology by Peter H. Thompson, 1969, 70, 71

Base-maps from Forest Resources Inventory Maps, Ontario
Department of Lands and Forests

Numbers around edge of maps refer to the one thousand
metre Universal Transverse Mercator Grid; Zone 18

30f

GRENVILLE SUPERGROUP AND ACTINOLITE

Metamorphosed ultramafic rocks

INTRUSIVE

FLINTON GROUP

LESSARD FORMATION (13)

Biotite-carbonate schist; 14a, epidote-calc-silicate paragneiss

Calcareous quartzofeldspathic psammite metaconglomerate; 13a, epidote-bearing silicate paragneiss; 13b, plutonic pebbly metaconglomerate

BISHOP CORNERS FORMATION (10-12)

Quartzofeldspathic psammite and metaconglomerate

Quartzite metaconglomerate and quartzite; 11a, quartzite

Pelitic schist

UNCONFORMABLE

METAMORPHOSED PLUTONIC ROCKS

9a, biotite quartz monzonite; 9b, diorite

8a, biotite granodiorite and quartz diorite

Hornblende diorite; 7a, coarse grained gabbro

INTRUSIVE

HERMON GROUP

Marble: 5a, calcitic; 5b, dolomitic

TUDOR FORMATION (1-4)

Salic metavolcanics

Fine-grained massive gray-weathering locally amygdaloidal

Chlorite-amphibole schist, hornblende-schist, gneissic amphibolite, locally fragmental; 2a, hornblende-biotite quartzofeldspathic

Massive fine-grained amphibolite locally pillowed, fragmental

NOTE: Units 3, 6b, 7, 13a, 13b, 14, 14a, 15, and 17 are not shown on this map.

2 of 1

DOMAIN II

56
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LEGEND

POST FLINTON ROCKS

Metamorphosed ultramafic rocks

Pegmatite

Structural Domain boundary

Contact; contact in pre-Flinton rocks

Pre-Flinton foliation

Axial surface trace of D_1 synclines; overturned

Trend of S_1 in pre-Flinton rocks

Bedding (S_0) top known; inclined, vertical, dip not known

Bedding (S_0) top not known; inclined vertical

Plunge of D_1 minor folds (F_1) and S_0/S_1 intersections: inclined, horizontal, vertical

Stereonet diagrams, poles to S_1

F_1 hinge line and S_0/S_1 intersection

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47
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43
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LEGEND

POST FLINTON ROCKS

Metamorphosed ultramafic rocks

Pegmatite

Structural Domain boundary

Contact; contact in pre-Flinton rocks

Pre-Flinton foliation

Axial surface trace of D_1 synclines; overturned

Trend of S_1 in pre-Flinton rocks

Bedding (S_0) top known; inclined, vertical, dip not known

Bedding (S_0) top not known; inclined, vertical

Plunge of D_1 minor folds (F_1) and S_0/S_1 intersections:

inclined, horizontal, vertical

Stereonet diagrams, poles to S_1

F_1 hinge line and S_0/S_1 intersection

40 of 4

FIGURE 6
D₁ STRUCTURAL TREND MAP, FLINTON AREA
SOUTHEASTERN ONTARIO

1 of

MADOC AREA

LEGEND

- Contact, contact in pre-Flinton rocks
- Pre-Flinton foliation
- Axial surface trace of D_1 synclines; overturned
- Trend of S_1 in pre-Flinton rocks
- Bedding (S_0) top known: inclined, vertical, dip not known
- Bedding (S_0) top not known: inclined, vertical
- Plunge of D_1 minor folds (F_1) and S_0/S_1 intersections: inclined, horizontal, vertical
- Stereonet diagrams, poles to S_1
- F_1 hinge line and S_0/S_1 intersection

Geology by Peter H. Thompson, 1969, 70, 71

FIGURE 7

D_1 STRUCTURAL TREND MAP FOR MADOC AND BLACK RIVER SOUTHEASTERN ONTARIO

LEGEND

- pre-Flinton rocks
- Flinton rocks
- D₁ synclines; overturned
- D₂ synclines; overturned
- D₃ synclines; overturned
- D₄ synclines; overturned
- D₅ synclines; overturned
- D₆ synclines; overturned
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- D₉₈ synclines; overturned
- D₉₉ synclines; overturned
- D₁₀₀ synclines; overturned

Geology by Peter H. Thompson, 1969, 70, 71

FIGURE 7
STRUCTURAL TREND MAP FOR THE
BLACK RIVER AREA
EASTERN ONTARIO

1 of

MADOC AREA

DOMAIN V

MADOC

LEGEND

- Contact, contact in pre-Flinton rocks
- Pre-Flinton foliation
- Axial surface trace of D_1 synclines; overturned
- Trend of S_1 in pre-Flinton rocks
- Bedding (S_0) top known: inclined, vertical,
- dip not known
- Bedding (S_0) top not known: inclined, vertical
- Plunge of D_1 minor folds (F_1) and S_0/S_1 intersections: inclined, horizontal, vertical
- Stereonet diagrams, poles to S_1
- F_1 hinge line and S_0/S_1 intersection

Geology by Peter H. Thompson, 1969, 70, 71

FIGURE 7

D_1 STRUCTURAL TREND MAP FOR
MADOC AND BLACK RIVER AREA
SOUTHEASTERN ONTARIO

LEGEND

- contact in pre-Flinton rocks
- Flinton foliation
- surface trace of D_1 synclines; overturned
- S_1 in pre-Flinton rocks
- (S_0) top known: inclined, vertical
- known
- (S_0) top not known: inclined, vertical
- D_1 minor folds (F_1) and S_0/S_1
- directions: inclined, horizontal, vertical
- great circle diagrams, poles to S_1
- line and S_0/S_1 intersection

Geology by Peter H. Thompson, 1969, 70, 71

FIGURE 7

STRUCTURAL TREND MAP FOR THE MOC AND BLACK RIVER AREAS SOUTHEASTERN ONTARIO

MADOC AREA

LEGEND

- Contact, contact in pre-Flinton rocks
- Pre-Flinton foliation
- Axial surface trace of D_1 synclines; overturned
- Trend of S_1 in pre-Flinton rocks
- Bedding (S_0) top known: inclined, vertical, dip not known
- Bedding (S_0) top not known: inclined, vertical
- Plunge of D_1 minor folds (F_1) and S_0/S_1 intersections: inclined, horizontal, vertical
- Stereonet diagrams, poles to S_1
- F_1 hinge line and S_0/S_1 intersection

Geology by Peter H. Thompson, 1969, 70, 71

FIGURE 7
 **D_1 STRUCTURAL TREND MAP FOR THE
 MADOC AND BLACK RIVER AREAS
 SOUTHEASTERN ONTARIO**

BLACK RIVER AREA

ACTINOLITE

34 18

16 33

15

14

data from NE part
of Black River area

30f

BLACK RIVER AREA

DOMAIN IV

ACTINOLITE

BLACK RIVER AREA

ACTINOLITE

1 of

2 of .

DOMAIN II

D2

- Axial surface trace of D₁ synclines
- Trend of S₁ in pre-Flinton rocks
- Axial surface trace of D₂ folds: antiform
- Trend of S₂
- Plunge and vergence of D₂ minor folds (F₂)
- Trend of S₃
- Plunge and vergence of D₃ minor folds
- Stereonet diagrams:
- pole to S₁
- F₂ hinge line
- pole to S₃
- F₃ hinge line

DOMAIN II

D2

2 of .

55

54

21

53

52

51

50

20

49

48

47

55

54

53

52

51

50

49

48

DOMAIN III

D2

48

47

46

30f

45

19

44

43

42

41

40

18

39

LEGEND

- Structural domain boundary
- Axial surface trace of D_1 synclines
- Trend of S_1 in pre-Flinton rocks
- Axial surface trace of D_2 folds: antiform
- Trend of S_2
- Plunge and vergence of D_2 minor folds (F_2)
- Trend of S_3
- Plunge and vergence of D_3 minor folds
- Stereonet diagrams:
- pole to S_1
- F_2 hinge line
- pole to S_3
- F_3 hinge line

DOMAIN III

D2

D3

45

44

43

42

41

40

39

28

38

37

- Axial surface trace of D_1 synclines.
 Trend of S_1 in pre-Flinton rocks
 Axial surface trace of D_2 folds: antiform.
 Trend of S_2
 Plunge and vergence of D_2 minor folds (F_2)
 Trend of S_3
 Plunge and vergence of D_3 minor folds
 Stereonet diagrams:
 pole to S_1
 F_2 hinge line.
 pole to S_3
 F_3 hinge line.

1 of

MADOC AREA

LEGEND

- Axial surface trace of D₁ synclines
- Trend of S₁ in pre-Flinton rocks
- Axial surface trace of major D₂ folds
- Trend of S₂
- Plunge and vergence of D₂ minor folds (F₂)
- Trend of S₃
- Plunge and vergence of D₃ minor folds
- Stereonet diagrams:
 - pole to S₂
 - F₂ hinge line
 - pole to S₃
 - F₃ hinge line

FIGURE 11

D₂ AND D₃ STRUCTURAL MAP OF THE MADOC AND BLACK MOUNTAIN AREAS, SOUTHEASTERN SCOTLAND

0 KILOMETRES
0 MILES

LEGEND

- Axial surface trace of D₁ synclines
- Trend of S₁ in pre-Flinton rocks
- Axial surface trace of major D₂ folds
- Trend of S₂
- Plunge and vergence of D₂ minor folds (F₂)
- Trend of S₃
- Plunge and vergence of D₃ minor folds
- Stereonet diagrams:
- pole to S₂
- F₂ hinge line
- pole to S₃
- F₃ hinge line

FIGURE 11

D₂ AND D₃ STRUCTURAL TREND MAP DOUGLASS AND BLACK RIVER AREAS SOUTHEASTERN ONTARIO

2 of.

18

17

39

39

16

15

38

38

14

37

37

BLACK RIVER AREA

36

35

ACTINOLITE

34

18

17

16

33

14

15

D2

D3

12

36

DOMA

11

31

MADOC

30

35

31

30

01

02

03

04

34

33

11

12

D3

30f

12

BLACK RIVER AREA

36

DOMAIN IV

ACTINOLITE

3

17

33
16

12

13

14

15

D2

BLACK RIVER

DOMAIN IV

MADOC

D3

D2

4 of 4

BLACK RIVER AREA