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**PETROLOGY OF THE FLINTON CREEK ULTRAMAFIC ROCKS,
CENTRAL METASEDIMENTARY BELT,
GRENVILLE PROVINCE**

by

Frederick Dean Ford, B.Sc. (Hons)

**A thesis submitted to
the Faculty of Graduate Studies and Research
in partial fulfilment of
the requirements for the degree of
Master of Science**

**Department of Earth Sciences
Carleton University
Ottawa-Carleton Geoscience Centre
Ottawa, Ontario
November, 1989**

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Frederick Dean Ford, B.Sc. (Hons)
in partial fulfilment of the requirements
for the degree of Master of Science

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Carleton University
December, 1989

Abstract

Metamorphosed ultramafic rocks are exposed in elongate lenses on the eastern margin of the Flinton synclinorium, in the Central Metasedimentary Belt of the Grenville Province. The ultramafic rocks are composed of two, essentially bi-mineralic, lithologies: enstatite-magnesite rock, and anthophyllite-carbonate rock. The first rock type consists of coarse-grained (5x3cm), euhedral enstatite crystals (En_{90}), intergrown with coarse-grained (2x2cm), euhedral magnesite crystals. Other minerals present are anthophyllite ($X_{Mg}=0.88$), chromium-magnetite, dolomite, forsterite (For_{90}), talc ($X_{Mg}=0.97$) and tremolite. The second rock type is composed of spherically radiating (1cm in diameter), medium-grained, subhedral anthophyllite crystals intergrown with fine-grained, anhedral magnesite and dolomite. Minor minerals are chromium magnetite and talc. Enstatite-magnesite rocks occur only in the interior of the larger ultramafic lenses and are always surrounded by a mantle of anthophyllite-carbonate rock.

Ultramafic lenses are surrounded by variably deformed mafic schists and gneisses. Both ultramafic and mafic rocks are believed to be tectonically emplaced into the Flinton Group sediments. The development of mono-mineralic reaction zones between the mafic and ultramafic rocks suggests that both have been metamorphosed *in situ*. The following mineral zones are encountered while progressing outwards from ultramafic rock: tremolite zone, biotite zone, chlorite zone. Surrounding pelites have been metamorphosed in the staurolite-kyanite zone and sillimanite zone (for higher grade assemblages).

Mineral assemblages in the two ultramafic rock types may be related by the reaction $8 \text{ enstatite} + \text{CO}_2 + \text{H}_2\text{O} = \text{anthophyllite} + \text{magnesite}$. The anthophyllite + magnesite assemblage results from increased water activity along the margins of the ultramafic lenses. Mineral assemblages present in both rock types require the presence of fluids comparatively rich in CO_2 .

The growth of late-stage forsterite in the enstatite-magnesite rock may be an indication that the area experienced isothermal decompression during uplift.

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Dan Marshall added greatly to my limited knowledge of personal computers to the point where I can now freely engage in a digit-head conversation without finding my brow extensively furrowed. Dan also provided the eccentric program FUGCOEFF, used to verify the results of my shaky thermodynamic calculations.

Peter Jones provided valuable instruction on the operation of the Carleton University microprobe as well as advice on standard selection, practical analytical capabilities of the instrument and precision and accuracy of the E.D.S. method.

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Information about current research and other local geological activities was provided by members of the Resident Geologist's Office, Tweed. In particular, I would like to thank Peter LeBaron for providing access to analytical data on the talc bodies surrounding the Elzevir Batholith.

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Chapter One - Introduction

1.1 General Statement

The area surrounding Madoc, Ontario, referred to as the Hastings metamorphic low (Hewitt, 1956; Carmichael *et al.* 1978), represents an area of greenschist facies metamorphism in the Grenville province. Moving away from the Madoc area, there is a continuous transition from greenschist, through amphibolite and even to granulite facies metamorphism (Carmichael *et al.*, 1978). To date, pelitic rocks have been the most thoroughly studied in the area but a smaller amount of work exists on the metamorphism of the volcanic rocks (Sethuraman and Moore, 1973), carbonate rocks (Hutcheon and Moore, 1973) and granitoids (Kamilli, 1974) which surround the Hastings low.

Metamorphosed ultramafic rocks occur throughout the Central Metasedimentary Belt (Wynne-Edwards, 1972), usually associated with gabbroic intrusives. Very little work has been devoted to the study of the metamorphism of these rocks.

Trommsdorff and Evans (1974) established isograds and mineral zones in the metaperidotites of the Lepontine Alps. The pattern of isograds they obtained was consistent with the distribution of isograds already established in rocks of other compositions. A similar study has not yet been attempted in the Grenville province; however, this thesis could be considered a first step. In this study, the metamorphic assemblage of a Grenville province ultramafic body is described. This information is used, along with the mineral assemblages of rocks with other bulk compositions to better understand the regional metamorphism which occurred in this part of the Grenville province at least 1050 million years ago.

1.2 Statement of Purpose

The objectives of this study are:

1) To describe the minerals and mineral textures of the Flinton Creek ultramafic rocks and determine if changing pressures and temperatures of regional metamorphism have produced any variation in minerals and mineral assemblages.

2) To interpret geochemical data and attempt to determine the origin of the ultramafic body.

3) To interpret structural data to establish the method and timing of ultramafic emplacement.

4) To use thermodynamic data and mineral compositions obtained by microprobe analyses to develop a petrogenetic grid for magnesite-bearing ultramafic rocks which may then be used, along with textural evidence, to establish the pressure and temperature conditions of metamorphism and the composition of the accompanying fluid.

1.3 Location, Access and Physiography

The study area is located in southeastern Ontario, 200 km southwest of Ottawa; see Fig. 2.1. The ultramafic body straddles the boundary between Elzevir Township, Hastings County, and Kaladar Township in the County of Lennox and Addington. The study area is 2 km south of the town of Flinton and 8 km west of Kaladar.

The Flinton Creek ultramafic rocks occur within walking distance of the Upper Flinton Road which may be reached directly from Ontario highway 7, or from highway 41 through the town of Flinton.

The ultramafic rocks are generally found in low-lying, swampy areas and in stream gullies. Despite this unfavourable setting, the rocks are fairly well exposed in elongate, steep-sided outcrops which rise 3-4 metres above the surrounding ground surface. Still, several large swampy areas are totally devoid of outcrop. The exposure is excellent along a power line which cuts the body near its widest point (5.5 km north

of highway 7 on the Upper Flinton Road), and in a quarry (9.5 km north of highway 7, 850 m east of the Upper Flinton Road), where the ultramafic rock was mined for decorative stone.

1.4 Recent Work

Table 1.1 summarizes both unpublished (thesis) and published work completed near the study area over the last twenty years. Note that in recent years, a large proportion of the work has been carried out by Carleton University students under the direction of Dr. John Moore and Dr. Richard Brown.

1.5 Symbols and Units

The symbols used in this thesis are those common in geological literature. Units are standard S.I. units. Symbols and corresponding units are listed in Table 1.2. Mineral formulas and abbreviations are listed in Table 5.1.

Table 1.1

Theses and Published Papers Concerning the Precambrian geology of south-eastern Ontario

Author(s)	Year	Subject of Study
Lumbers	1967a	Large scale study of regional stratigraphy, petrology of plutonic intrusions, timing of pluton emplacement, and the regional metamorphism of meta-volcanic rocks.
Hounslow and Moore	1967	Prograde metamorphism and geochemistry of pelitic rocks on the eastern margin of the Hastings metamorphic low.
Carmichael	1969	Mechanism of prograde metamorphic reactions in pelitic schists of the Whetstone Lake area, on the western margin of the Hastings metamorphic low.
Carmichael	1970	Definition and mapping of isograds based on discontinuous reactions in calcareous and pelitic rocks, plus the relationship between isograds from the two rock types.
Thompson	1972	Geology of the Flinton Group sediments including stratigraphy, structure and metamorphism of pelitic and calcareous rocks.
Hutcheon and Moore	1973	Mapping the tremolite isograd and calcite geothermometry in the Marble Lake area.
Sethuraman and Moore	1973	Prograde metamorphism and minor geochemistry of volcanic rocks on the eastern margin of the Hastings metamorphic low.
Thompson	1973	Definition of reaction isograds and mineral zones in impure calcareous rocks of the Lessard formation.
Psutka	1976	Geology of the Kaladar Metasediments and provenance of plutonic pebbles from the Kaladar metaconglomerate.
Chappell	1978	Structure and metamorphism in the Clare River synclinorium. Mapping, petrology and geochemistry of the Kaladar complex is in appendix 1.
Connelly	1987	Method and timing of Elzevir batholith emplacement.
Harnois	1988	Possible origin of the Ore Chimney formation and its relation to Gold deposition.

Table 1.2**Symbols and units**

a_i	Activity of species "i"
(aq)	Subscript designating an aqueous species
C_p	Heat capacity at constant pressure
Δ	Difference, final state minus initial state
f_i	Fugacity of species "i"
γ_i	Fugacity coefficient of species "i"
G	Gibbs function, or Gibbs free energy
H	Enthalpy, or heat function
K	Equilibrium constant
k_i	Power series heat capacity coefficient
μ_i, μ_i^0	Chemical potential of species "i", chemical potential of pure "i"
P	Pressure; kilobars or bars (1kbar=100MPa)
R	Gas constant (8.314 J/(K · mole), 1.9872 cal/(K · mole))
S	Entropy
T	Temperature; Kelvin scale
T_r, P_r	Reference temperature and pressure; 298.15 K, 0.1 MPa (25°C, 1 bar)
V_s	Molar volume of a solid phase
X, X_i	Mole fraction, mole fraction of species "i"

Chapter Two - Regional Geology

2.1 Introduction

The study area is located in the Central Metasedimentary Belt of the Grenville Province; the youngest orogenic belt in the Canadian Shield (Davidson, 1986). The Central Metasedimentary Belt (Wynne-Edwards, 1972) has been divided by Davidson (1986) into three subsections: the Bancroft terrain, the Elzevir terrain and the Frontenac terrane. The Elzevir Terrain, which encompasses the study area, is roughly equivalent to segment IVb, the Hastings Basin, of Wynne Edwards (1972). The terrain is characterized by a succession of metavolcanic rocks, marbles, and fine to medium-grained clastic metasediments; the Grenville Supergroup (Wynne-Edwards, 1972). Metamorphic "depressions" located within the terrain contain the few occurrences of greenschist to lowermost amphibolite facies metamorphic rocks in the Grenville Province (Davidson, 1986). Primary structures are often preserved in this area allowing stratigraphic principles to be applied during field mapping.

2.2 Stratigraphy

The stratigraphy first defined by Lumbers (1967a) for the Madoc-Bancroft region, west of the Elzevir batholith (see Fig. 2.1), may be applied to Grenville Supergroup rocks surrounding the study area, (boxed in Fig. 2.1). The stratigraphy of the overlying Flinton Group was defined by Moore and Thompson (1972, 1980). Figure 2.2 details the stratigraphic column in the area of interest.

2.2.1 Hermon Group

Predominantly mafic metavolcanic rocks (Unit 1, Fig. 2.4), have been correlated by Moore and Thompson (1980) with the Tudor Formation, the lowest member of the Hermon Group as defined by Lumbers (1967a). The metavolcanics of this formation are typically dark, fine to medium-grained, quartz-poor rocks, believed to be derived primarily from basic volcanic rocks (Sethuraman and Moore, 1973).

Figure 2.1 General geology of part of the eastern Elzevir Terrain, Central Metasedimentary Belt, Grenville Province. The study area occurs on the eastern margin of the Flinton Synclinorium, within the rectangular box shown on the main map. The detailed geology of this area is shown in Figure 2.4. Note the location and geologic setting of several places, units and structures mentioned in the text. In particular, note the apparent unconformity which exists between the Flinton Group (black) and the underlying volcanics and sediments in the Bishop Corners (Bi) - Myers Cave (My) area. Here, contacts between the older, underlying units can be traced across the Fernleigh Syncline. Place names are abbreviated as follows: Ac = Actinolite; Bi = Bishop Corners; Cl = Cloyne; Fe = Fernleigh; Fl = Flinton; Fo = Flower Station; Ka = Kaladar; Ma = Madoc; My = Myer Cave; Om = Ompah; Tw = Tweed. Figure is modified from Moore and Thompson (1980).

Figure 2.2 Stratigraphic column defined by Lumbers (1967a) and Moore and Thompson (1980) for Precambrian rocks in the vicinity of the study area. Note the presence of the major unconformity between plutonic rocks which intrude the Grenville Supergroup and the Flinton Group sediments.

2.2.2 Mayo Group

The Hermon Group is overlain, possibly with minor unconformity, by carbonate sedimentary rocks (Unit 2, Fig. 2.4) correlated by Moore and Thompson (1980) with the Mayo Group of Lumbers (1967a).

2.2.3 Plutonic Rocks

Lumbers (1967a) subdivided the plutonic rocks of the Bancroft-Madoc region into seven major groups, summarised in Figure 2.3. The relative age of each group was assigned on the basis of field relationships and isotopic data (Silver and Lumbers, 1966). Lumbers (1967a) demonstrated that the three oldest groups, the albite granite and syenite group, the diorite group, and the trondhjemite and sodic granite group, were spatially and temporally related. These three groups form the biotite diorite series (Lumbers, 1967a), which is characterized by trondhjemite-sodic granite differentiation trends and a spatial association with rocks of the Tudor Formation. Three of the seven plutonic rock groups defined occur within the study area.

The trondhjemite and sodic granite group is composed of trondhjemite with subordinate granodiorite and late phases of sodic granite and granitic pegmatite. This group is represented locally by the Elzevir and Northbrook batholiths.

A portion of the Kaladar Complex was mapped as hornblende diorite by Thompson (1972). The pluton is believed to be a member of the diorite group. As well, the Kaladar Complex has several characteristics of a biotite diorite series pluton including:

- 1) A spatial association with the Elzevir granodiorite, another member of the biotite diorite series,
- 2) A spatial association with Hermon Group metavolcanics, (Lumbers, 1967a).

The exact stratigraphic position of the complex remains controversial due to conflicting field evidence. Chappell (1978) noted fragments of hornblende and coarse-grained gabbro, similar to the Kaladar Complex, in fragmental volcanic rocks of the Tudor

Figure 2.3 Summary of the seven major plutonic groups found in the Elzevir Terrain (shown in black) as defined by Lumbers (1967a). Relative age of each group was assigned on the basis of field evidence, Lumbers (1967a), and isotopic data (Silver and Lumbers, 1966). Diagram from Lumbers (1967b).

Decreasing Age —————→

□ □ □ sedimentation and volcanism (oldest known lavas about 1310 m.y. old)

□ □ □ mafic sills and dykes

■ ■ ■ albite granite and syenite group

■ ■ ■ diorite group

■ ■ ■ trondhjemite and sodic granite group

biotite diorite series
(about 1250 m.y.)

□ □ □ mafic sills and dykes

■ ■ ■ late gabbro and diorite group

■ ■ ■ nepheline syenite group

■ ■ ■ potassic syenite
and monzonite group

■ ■ ■ quartz monzonite group

□ □ □ late pegmatite
(about 1050)

Figure 2.4 Geology of the Flinton Synclinorium and surrounding area, after Thompson (1972). Note that the Flinton Creek ultramafic rocks cross-cut lithologic units belonging to the Flinton Group. Ultramafic rocks are found in the Northbrook Batholith, along the Northbrook Batholith - Flinton Synclinorium contact, and within the Flinton Synclinorium surrounded by Flinton Group sediments.

GEOLOGY OF THE FLINTON AREA SOUTHEASTERN ONTARIO

After: Thompson (1972)

Formation. This suggests that the Kaladar Complex contains the oldest rock exposed in the area. However, the opposite age relationship is suggested by amphibolite xenoliths which are found throughout the body and gabbroic sills that appear to intrude the surrounding volcanics. The last two field relationships support isotopic data which suggest the biotite-diorite series was intruded shortly after the cessation of Tudor Formation volcanism.

The Addington granite is a member of the Quartz Monzonite Group (Lumbers, 1967a), the third plutonic rock group in the area.

2.2.4 The Unconformity

The presence of a major angular unconformity above the plutonic rocks in the area has been suggested by several authors, (Moore 1967; Moore and Thompson 1972, 1980; and Thompson 1972). The clastic sediments of the Flinton Group (Moore and Thompson 1972, 1980), are thought to lie above this unconformity on the older rocks.

Connelly (1985, 1986) argues that the contact at the base of the Flinton Group could be tectonic and is not necessarily a stratigraphic unconformity. Specifically, Connelly (1985) suggests that the proposed unconformity is a folded thrust plane.

2.2.5 Ore Chimney Formation

The Ore Chimney Formation (Moore and Thompson, 1980), is a thin (<10m), biotite schist which overlies the unconformity separating Grenville Supergroup mafic to intermediate volcanics from the Flinton Group sediments .

Moore and Thompson (1980), proposed that the Ore Chimney Formation was a metamorphosed regolith which formed from metavolcanic rocks on the pre-Flinton erosion surface. Work by Harnois and Moore (1988) further supported this conclusion. However, Springer (1982) suggested that the schist represents a sheared and metasomatised layer of the underlying mafic rocks.

2.2.6 Flinton Group

The Flinton Group outcrops in a series of narrow fold belts extending from near Madoc, north-east through Bishop Corners and Fernleigh to at least Flower Station (Moore and Thompson, 1980), (see Fig. 2.1). Also shown in this figure are metasediments, found separated from the continuous belts, that are believed to be correlative with the Flinton Group (Chappell, 1978; Psutka, 1976).

Based on mineralogy and inferred pre-metamorphic lithology, Thompson (1972) and Moore and Thompson (1972, 1980) grouped lithologic units into several formations, summarized in Table 2.1 below.

Table 2.1

Formations in the Flinton Synclinorium

Moore and Thompson (1972, 1980), Thompson (1972, 1973)

Formation	Members of the Formation	Equivalent High Grade Members
Lessard Formation	biotite-carbonate schist	epidote-free calc-silicate paragneiss
	calcareous muscovite quartzofeldspathic psammite	epidote-bearing calc-silicate paragneiss
	plutonic pebble conglomerate	
Bishops Corners Formation	pelitic schist	
	quartzite	
	quartzite metaconglomerate	
	micaceous quartzofeldspathic psammite	

2.2.7 Post-Flinton Rocks

2.2.7.1 Pegmatites

Simple pegmatites, composed primarily of microcline and sodic plagioclase with minor biotite, muscovite and quartz, cut the Bishop Corners and Lessard Formations near the south end of the Flinton synclinorium. A few large pegmatite bodies have been shown on Figure 2.4 but most are small dykes, 1-3m thick. The pegmatites are found in the area of highest metamorphic grade, leading Moore and Thompson (1980) to suggest they originate by partial melting during metamorphism.

2.2.7.2 Flinton Creek Ultramafic Rocks

The Flinton Creek ultramafic rocks (new name) are herein defined as a group of metamorphosed rocks with approximately ultramafic bulk composition (Wyllie, 1967), which are relatively rich in magnesium silicate minerals (enstatite, forsterite, anthophyllite) and occur in and around the Flinton Synclinorium. Rocks which meet this criterion are generally located adjacent to Flinton Creek (see Fig. 2.4), hence the name. The Flinton Creek ultramafic rocks are surrounded by a variety of different host-rocks (see Fig. 2.4), since the ultramafic rocks are found:

- 1) in the Northbrook Batholith,
- 2) along the Northbrook Batholith - Flinton synclinorium contact,
- 3) within the Flinton synclinorium, surrounded by Flinton Group sediments.

The Flinton Creek ultramafic rocks and an associated suite of metabasites comprise the Flinton Creek Mafic - Ultramafic Body.

There is conflicting evidence concerning the exact stratigraphic position of this unit.

Similar ultramafic pods have been described by Chappell (1978), in the Kaladar

Complex. If the Flinton Creek rocks are related to this occurrence, then they must be older than the Flinton Group. The same age relationship also holds if the Flinton Creek rocks are related to talc bodies, which could have had ultramafic protoliths, that have been found interbedded with the pre-Flinton Tudor mafic volcanic rock along the southern (Hunter, 1973) and western (LeBaron *et al.* 1987) sides of the Elzevir batholith.

Thompson (1972), on the other hand, demonstrates that ultramafic rocks of the Flinton Creek unit cross-cut lithologic contacts in the Flinton Group as indicated on Fig. 2.4. This relationship was confirmed during the course of the present study. Pegmatites which post-date the Flinton Group (Thompson, 1972) were never observed intruding the ultramafic rocks, even though the pegmatites and ultramafic rocks occur in close proximity. This is consistent with the suggestion that the Flinton Creek ultramafic rocks are younger than the Flinton Group (although the ultramafic was also never clearly seen cross-cutting the pegmatite). Results presented in later sections will be used to resolve this apparent contradiction.

2.3 Regional Structure

Previous work by several authors, (Chappell, 1978; Rivers, 1976; Thompson, 1972; Moore, 1967) has been completed on the structure of the area surrounding the Flinton Creek body.

Provided that the contact at the base of the Flinton Group is not a tectonic one, then at least one period of pre-Flinton group deformation is necessary to explain the structural unconformity observed in the Bishop Corners - Myer Cave area, see Fig. 2.1 (Moore, 1967). At this location, contacts and layering within the north to north-west striking, steeply dipping rocks of the Tudor and Hermon formations are truncated by steeply dipping, north-east striking sediments of the Flinton Group. The existence of pre-Flinton Group deformation is further supported by the presence of angular foliated and lineated metavolcanic and gabbroic inclusions, found within undeformed Elzevir granodiorite (Moore and Thompson, 1980).

Moore and Thompson (1980) concluded that three deformational periods followed the deposition of the Flinton group. This is in general agreement with other authors who have completed more detailed structural studies in the area (Chappell, 1978; Rivers 1976; Thompson, 1972).

The first post-Flinton Group deformation, (D_1) (Thompson, 1972), produced the narrow isoclinal folds, containing the Flinton Group, which can be traced from 5 to 20 kilometres along strike, see Fig. 2. The regional planar fabric (S_1) is parallel to the axial surfaces of these first folds.

Second phase (D_2) deformational structures are difficult to see at map scale. Field evidence for D_2 in the Flinton area includes the refolding of D_1 -flattened conglomerate pebbles (S_1) about northeast trending axial surfaces.

The third phase of deformation is best seen on the map scale, (Fig. 2.1, in the Myers Cave area), where D_3 open folds with north-west trending axial surfaces may be seen warping the axial surface of the major isoclinal folds.

A final, late-stage, gently-dipping, crenulation (D_4) has been developed in micaceous rocks throughout the study area on the D_1 axial planar schistosity.

2.4 Regional Metamorphism of Pelitic Rocks

Surrounding the Hastings Metamorphic Low

Pelitic isograds have been mapped in detail on the eastern and western margins of the Hastings metamorphic low. Moving in an east to west direction, Carmichael (1970) located four successively higher-grade isograds within Grenville Supergroup pelitic rocks in the Whetstone Lake area, on the western margin of the Hastings low. The following mineral reactions were interpreted from these isograds:

- 1) chlorite + muscovite + garnet = staurolite + biotite + quartz (+ water);
- 2) chlorite + muscovite + staurolite + quartz = kyanite + biotite (+ water);
- 3) kyanite = sillimanite;
- 4) staurolite + muscovite + quartz = sillimanite + garnet + biotite (+ water).

Carmichael (1970) noted that only 250m up-grade from isograd 3 (the first appearance of sillimanite in kyanite-bearing rocks), sillimanite was formed as a product of a staurolite-consuming reaction in kyanite-free rocks (isograd 4). Thus in the Whetstone lake area, staurolite breakdown occurred in the sillimanite zone, close to the kyanite = sillimanite isograd.

On the eastern margin of the Hastings metamorphic low, pelitic rocks occur in the narrow Fernleigh syncline which runs from Bishop Corners northeast to Flower Station, (see Fig. 2.1). The prograde metamorphism of the lowermost, dominantly pelitic member of the Flinton Group in this syncline, the Bishop Corners Formation, has been documented by several authors, (Hounslow and Moore 1967, Carmichael, Moore and Skippen 1978, and Rivers 1976).

The lowest grade pelitic rocks in the area are found in the vicinity of Bishop Corners. Here porphyroblastic chloritoid co-exists with quartz + muscovite + biotite with or without garnet, (Carmichael *et al.*, 1978). Chlorite and garnet assemblages are not observed reacting to form staurolite, (Carmichael 1970, isograd I), possibly due to increased garnet stability caused by high manganese content, (Hounslow and Moore, 1967).

Moving eastward (up-grade) chloritoid is pseudomorphed by either coarse white mica, or fine aggregates of staurolite + quartz. Chloritoid disappears completely approximately half-way between Bishop Corners and Myers Cave, (see Fig. 2.5).

Near this last location, at Marble Lake, Boyce (1986) reported the following mineral assemblage for several samples: staurolite + muscovite + chlorite + kyanite + biotite. This is the invariant assemblage of Carmichael's (1970) isograd II which persists up-grade until Fernleigh, (Carmichael, Moore and Skippen, 1978) where the pair chlorite + staurolite no longer appear to be stable.

Co-existing kyanite and andalusite have been reported by Boyce (1986) at one location in the Fernleigh belt near Marble Lake. Both aluminosilicate polymorphs appear to be in stable co-existence, (D.M. Carmichael, pers. comm. 1989) and textural

Figure 2.5 Regional distribution of pelitic isograds in the Eastern Elzevir Terrain, Central Metasedimentary Belt, Grenville Province. Note that the Flinton Creek Ultramafic rocks occur in both the kyanite-staurolite, and sillimanite zones. However, most of the ultramafic rocks are found on or above the kyanite=sillimanite isograd. Isograd mineral abbreviations are listed on Figure 2.6. Regional geology is from Moore and Thompson (1980).

evidence could not be used to prove that one polymorph was forming at the expense of the other, i.e. retrograde andalusite forming after kyanite.

Isograd III occurs near Ardoch where sillimanite first appears as fibrolite knots with biotite and garnet, (Carmichael, Moore and Skippen, 1978). Staurolite persists beyond this point as irregular, microscopic crystals for approximately 2km. Thus, the situation at Ardoch is similar to that of the Whetstone lake area, with staurolite breakdown occurring shortly after entering the sillimanite zone.

Sillimanite growth at the expense of muscovite in the Fernleigh belt beyond Ardoch has been documented by Rivers (1976). Muscovite breakdown is believed by Rivers (1976) to occur according to the reaction:

The four-phase reaction assemblage which exactly locates the isograd was only found at one locality north-east of Clyde Forks, approximately 75 km north-east of the Flinton Creek ultramafics.

South-east of the study area, the high pressure assemblage sillimanite - orthopyroxene has been reported in the granulite grade pelitic gneisses of the Frontenac Arch, (Carmichael, Helmstaedt and Thomas, 1987). Cordierite reaction rims may separate the orthopyroxene from the aluminosilicate. This high-pressure assemblage is found only in a small area. More commonly, pelitic gneisses contain garnet - cordierite assemblages with or without orthopyroxene.

2.5 Regional Pressure-Temperature Gradient of the Eastern Elzevir Terrain

When the accumulated field evidence reviewed in the previous section is plotted on a pressure-temperature (P-T) grid for pelitic rocks, a pressure-temperature gradient for the Eastern Elzevir terrain may be established.

An important constraint in locating the gradient is the observation (Carmichael, 1970, Carmichael *et al.*, 1978) that staurolite breakdown in the Elzevir terrain occurs shortly after crossing into the sillimanite stability field. This situation corresponds to the point labelled Ardoc on Fig. 2.6.

Figure 2.6 Regional pressure-temperature gradient for the Eastern Elzevir Terrain. Approximate P-T conditions of mineral assemblages used to derive the gradient are shown as stars and labelled by location. The shaded area shows the pressure-temperature range derived by Harnois (1988) for mineral assemblages at the Orc Chimney mine using sphalerite geobarometry and garnet-biotite geothermometry.

The occurrence of kyanite and andalusite in textural equilibrium at Marble Lake (Boyce, 1986), suggests that the rocks equilibrated close to the kyanite=andalusite equilibrium curve. The point labelled Marble Lake on Fig. 2.6 lies close to this curve in the kyanite stability field and assumes the metastable persistence of andalusite at Marble Lake. The formation of andalusite from kyanite during late-stage decompression may also explain the presence of this assemblage.

Sphalerite geobarometry on three samples from the Ore Chimney mine, near Marble Lake, yielded pressure estimates ranging from 3.5 to 5.9 kbars (Harnois, 1988), (see shaded area, Fig. 2.6). These results suggest equilibrium pressures higher than required for the co-existence of kyanite and andalusite; however, the pressure range obtained from the data of Harnois (1988) is large. More analyses are necessary to better define an equilibrium pressure using this technique. Harnois (1988) used garnet-biotite thermometry as calibrated by Thompson (1976) to estimate equilibration temperatures for three samples at the same location. Garnet-rim compositions gave equilibrium temperatures which ranged from 475 - 500 C, (see shaded area, Fig. 2.6).

The presence of cordierite reaction rims between orthopyroxene and sillimanite (Carmichael *et al.*, 1987), may be used to locate the upper portion of the P-T gradient. The reaction texture suggests that the higher pressure assemblage, sillimanite - orthopyroxene became unstable relative to lower pressure cordierite - orthopyroxene before metamorphic quenching. For this to occur, temperature must have declined significantly through the invariant point labelled Frontenac Axis on Fig. 2.6. This point is the upper limit of the proposed geothermal gradient.

Chapter Three - The Flinton Creek Mafic - Ultramafic Body

3.1 Introduction

Reconnaissance mapping (at 1:25,000) and sampling of the Flinton Creek meta-ultramafic body commenced in the winter of 1987 and continued intermittently until the fall of 1988. The purpose of this program was to study the minerals in the ultramafic rocks and to see if the mineral assemblage varied with increasing metamorphic grade. Further detailed mapping (1:500) and sampling were completed in smaller, better exposed areas to describe the structural relationship of the mafic-ultramafic body with surrounding rocks and to determine the time and method of ultramafic rock emplacement.

3.2 Mode of Occurrence

Thompson (1972), represents the ultramafic rocks on the eastern margin of the Flinton synclorium as a single, continuous, tadpole-shaped unit (see Fig. 2.4). However, the very nature of the outcrop pattern in this area suggests that the ultramafic rock does not constitute a single, continuous body.

Rocks of the Flinton Creek body are exposed in discontinuous, steep-sided, lens-like outcrops. These trend sub-parallel to the structural fabric in surrounding units and vary in size from less than 1m x 0.25m to over 50m x 10m. The gullies which separate the ultramafic outcrops are deep and narrow, yet, for the most part, lack streams, suggesting that less competent material surrounding the ultramafic exposures has been eroded. These observations, and others detailed below, lead to the conclusion that the individual ultramafic outcrops are actually lenses or boudins, detached from one another and surrounded by other types of rock. Thus, even though the ultramafic rocks comprise about 90 percent of the outcrop in the Flinton Creek body, other rock-types could be more abundant since the area has only about thirty percent exposed bedrock.

3.3 Lithology

The Flinton Creek body is composed of the following three rock suites:

- 1) ultramafic rocks (enstatite-magnesite and anthophyllite-carbonate assemblages)
- 2) mafic rocks
- 3) metasomatic reaction rocks

3.3.1 Enstatite-Magnesite Rock

This rock is composed primarily of coarse-grained, euhedral enstatite and coarse to medium-grained magnesite (see Plate 3.1). Similar carbonate - orthopyroxenites from the Troms region of northern Norway were first discovered by Pettersen (1883), who applied the name *sagvandite* to the new rock type. *Sagvandite*-type rocks have also been described from Aaheim, western Norway (Kendel, 1970), and the Swiss Alps (Trommsdorff and Evans, 1970). More recent work on the Troms locality has been completed by Schreyer, Ohnmacht and Mannchen (1972) and Ohnmacht (1974).

The rock weathers a dun-brown colour, while the fresh surface is a greasy-green. The enstatite-magnesite rock has been found throughout the Flinton Creek body, except in the two most northerly ultramafic exposures, (see Fig. 2.4). Individual enstatite crystals are blocky, with either approximately rectangular or square cross-sections. Rectangular sections vary in size from 1cm×2cm to 5cm×15cm. The grain-size within a given sample tends to be fairly homogeneous (uniformly coarse-grained or very coarse-grained). Enstatite appears to have grown with no preferred orientation, and in carbonate-poor rocks, crystals grow together in a well-developed interlocking texture.

The amount of carbonate present in the rock varies from less than 10% to approximately 60%. X-ray diffraction and microprobe analysis have identified most of the carbonate present as magnesite, with lesser amounts of dolomite. No calcite was observed. Carbonate (magnesite) present in the enstatite-magnesite rock appears to be

Plate 3.1a Field photograph showing the typical weathered surface of an enstatite-magnesite rock outcrop. Coarse-grained, euhedral enstatite crystals weather high on the outcrop surface. The pyroxenes are intergrown with buff-coloured, coarse-grained, euhedral-subhedral magnesites. A Canadian 25 cent piece (diameter of 2.4cm) is shown for scale. Locality 101588, Morrisett Farm (see Appendix III).

Plate 3.1b Field photograph of partially weathered anthophyllite-carbonate rock. Anthophyllite needles grow in spherically radiating bundles or rosettes, approximately 1cm in diameter. These are surrounded by medium-grained, buff-weathering carbonate grains. Green markings on scale card are 1cm long. Locality 103188, Hearthstone Anthophyllite Quarry (see Appendix III).

both in equilibrium with, and replacing enstatite. Equilibrium magnesite occurs as coarse-grained, euhedral crystals intergrown with enstatite. Magnesite grains vary in size from 1cm x 1cm to 6cm x 6cm. In general, enstatite and magnesite usually have similar grain-size (very coarse-grained magnesite is usually found in equilibrium with very coarse-grained enstatite, rather than very coarse-grained magnesite and medium-grained enstatite). The medium-grained, anhedral replacement variety of magnesite is found in veins and irregular patches which clearly cross-cut and replace enstatite.

Amphiboles are non-uniformly distributed through the enstatite-magnesite rock, usually comprising less than ten percent of the rock's volume. They occur either in parallel or radiating aggregates, often closely associated with carbonate. Textural evidence, such as cross-cutting amphibole veinlets and amphibole overgrowths of enstatite, demonstrates that at least some amphibole growth post-dates pyroxene formation.

Throughout the body are several locations where the enstatite rock appears to have been affected either by low temperature hydrothermal alteration or severe weathering. When this is the case, the rock is reduced to a uniform mass of chlorite, talc, serpentine, carbonate and opaques, (in order of decreasing abundance). However, the weathered rocks still retain several characteristics which allow them to be mapped as enstatite-magnesite rocks. These features include:

- 1) The presence of an irregular, rough weathered surface on the outcrop, which is covered by ovoid, randomly oriented blocks, that were formerly enstatite crystals.
- 2) Preferential replacement of enstatite by chlorite and serpentine producing, on broken surfaces and slabs, dark-green ovoid areas which contrast against a lighter-coloured talc - carbonate background.

3.3.2 Anthophyllite-Carbonate Rocks

As the name suggests, this rock is composed of fine-grained, anhedral carbonate (magnesite and/or dolomite), and medium-grained subhedral anthophyllite. The weathered surface varies in colour from light grey to light brown, depending on the amount of anthophyllite and talc (grey) or carbonate (brown) present. The fine-grained anthophyllite needles grow in spherical radiating aggregates (rosettes), with an average radius of about 1cm. The presence and abundance of these rosettes makes the rock quite distinctive and easily recognisable in the field. Although a spaced cleavage is developed in the rock adjacent to lithologic contacts, anthophyllite appears to have grown without any preferential orientation. This is evident since anthophyllite needles grow equally well in all directions, forming perfectly spherical rosettes.

3.3.3 Relationship Between the Two Ultramafic Rock Types

During the course of this study, it became apparent that the two ultramafic rock types are spatially related both to each other, and to the proximity of the ultramafic - host rock boundary. Several observations led to this conclusion including:

1) The discovery that enstatite-magnesite rock is restricted to the larger (thicker) ultramafic outcrops.

2) The discovery that enstatite-magnesite rock always occurs in the core of the outcrop, or ultramafic slice and is always surrounded on all sides by an envelop of anthophyllite-carbonate rock.

(Confirmation of these first two relationships came during detailed mapping of the Hearthstone Anthophyllite Quarry (see Fig. 3.1); where several of the larger ultramafic slices were found to be cored by enstatite-magnesite rock.)

3) The discovery of zoned carbonate veins cutting through the enstatite-magnesite rock. The veins had a magnesite core surrounded by a zone of magnesite and anthophyllite.

These three observations lead to several hypotheses concerning the formation of the two ultramafic rock types:

Figure 3.1 Geology of the Hearthstone Anthophyllite Quarry (locality 103188). Note that the enstatite-magnesite rock is completely surrounded by anthophyllite-carbonate rock. This relationship also applies in the third dimension not shown by this map since a cap of anthophyllite-carbonate rock was discovered in one of the quarry walls above the larger enstatite-magnesite rock occurrence. Also note that the steeply-dipping schistosity present in the rocks is usually sub-parallel to the ultramafic / host-rock contact and that mafic schists and gneisses are separated from the ultramafic rocks by a sequence of monominerallic reaction zones.

Geology of the Hearthstone Anthophyllite Quarry

1) Fluid composition was the primary factor controlling the formation of either rock type. The existence of zoned veins and anthophyllite-carbonate rock mantles around enstatite rock suggests that fluid infiltration resulted in the production of anthophyllite-carbonate rock at the expense of enstatite rock.

2) Anthophyllite-carbonate rock formed as a result of retrograde metamorphism from the enstatite-magnesite rock. The core of enstatite rock survived because water could not reach it in time to catalyze the retrograde reactions.

3) Some combination of 1 and 2 above.

These three ideas will be tested later in chapters 4 and 5.

3.3.4 Metabasites

Metabasic rocks occur in small lens-like bodies on the margins of the larger ultramafic outcrops, except in the Hearthstone Anthophyllite Quarry where they are well exposed and surround the ultramafic rocks. The metabasites have variable composition, grain-size and degree of deformation. Weathered surfaces are either black or dull green. Both of these varieties have black fresh surfaces.

Metabasites are typical amphibolites composed of variable amounts of amphibole (60 to 100 percent) and plagioclase (0 to 30 percent). Other minor minerals observed in hand specimen are: quartz, biotite, chlorite, epidote, sphene, chalcopyrite, pyrite and pyrrhotite.

Amphibolite grain-size varies from very coarse-grained to fine-grained. As would be expected, this variation is partially related to deformation. Less deformed rocks consisting of nearly 100 percent amphibole have equant, granular textures while those with higher proportions of plagioclase are compositionally zoned, (with felsic and mafic domains). Grain-size in these rocks varies from medium- to coarse-grained. More-strongly deformed rocks are very fine-grained to medium-grained with well developed foliations and mineral segregations.

While the fine-grained metabasites closely resemble the surrounding Tudor

Formation metavolcanics (Lumbers, 1967a; Sethuraman and Moore, 1973), the coarser-grained, less deformed, rocks have textures which suggest that they were derived from mafic plutonic rocks. Pods of metabasite with a gabbroic appearance were found in the Hearthstone Anthophyllite Quarry surrounded by more strongly deformed, fine-grained amphibolites. This suggests that the fine-grained, volcanic looking metabasites might be produced by stress induced, grain-size reduction of the coarse-grained material during deformation.

3.3.5 Metasomatic Rocks

Frequently, the contact between an ultramafic body and the adjacent country rock is marked by monominerallic zones which represent the products of reaction between the dissimilar rock types. Several authors have studied these classic examples of metasomatism including Read (1934), Phillips and Hess (1936), and Sanford (1982). It is generally accepted that mass transport along chemical potential gradients produces these zones. At the Hearthstone Anthophyllite Quarry, reaction zones were found between the anthophyllite-carbonate rock and the surrounding mafic gneisses. The total thickness of all the reaction zones combined rarely exceeded half a metre. Despite their size, these reaction zones are important because their presence indicates that the Flinton Creek body was metamorphosed, on at least one occasion, in-situ.

A typical progression outwards from the ultramafic rocks into the surrounding mafic rocks reveals the following mineral zones:

- 1) Monominerallic tremolite zone. This zone has an average thickness of approximately 30 mm. Tremolite needles grow in radiating aggregates which nucleate on the outer margin of the zone and grow, and expand, inwards. The inner boundary with the anthophyllite-carbonate rock is gradational, whereas the outer boundary is quite sharp.

- 2) Monominerallic biotite zone. The thickness of this zone ranges from 1 to 50 mm. The outer contact with the chlorite zone appears to be gradational.

3) Chlorite zone. This zone is composed primarily of chlorite; however, up to five percent biotite may also be found. There is a sharp contact between the chlorite zone and the surrounding mafic host rocks.

The last zone is not always present. If this is the case, then the tremolite zone passes sharply into the biotite zone which passes gradationally into a biotite - colourless, non-pleochroic amphibole (actinolite) zone. This outermost zone is in sharp contact with the host rocks.

The initial boundary between the mafic and ultramafic rocks was probably at or near the sharp contact between the biotite zone and the tremolite zone. Evidence for this includes the presence of chromium magnetite on the ultramafic side of the contact while Ti-bearing phases (ilmenite) occur on the other side.

Exact mass transfer values cannot be calculated without knowing the modal mineralogy of the various zones, and the individual mineral compositions obtained by microprobe analyses. However, Sanford (1982) has proposed the following mass-transfer estimate: silicon, iron, and carbon dioxide are introduced into the ultramafic body while water and magnesium pass outwards. These gradients are consistent with the observed petrography of the reactions zones surrounding the Flinton Creek Ultramafic rocks.

3.4 Structural Geology of the Flinton Creek Body

As mentioned previously, Flinton Creek meta-igneous rocks cross-cut formation contacts in the Flinton Group at a low angle (10-15°), see Fig. 2.4. Thompson (1972) cited the presence of mylonite in the surrounding sediments as proof that the ultramafic rocks had been tectonically emplaced into their current setting. The results of this study provide further evidence in support of this hypothesis.

The excellent exposure at the Hearthstone Anthophyllite quarry, (see Fig. 3.1), provided the best opportunity to study the contact relationships between the ultramafic and neighbouring lithologies. At this location, ultramafic rocks occur in discontinuous,

elongate lenses which vary in length from 1m to 50m and in width from 0.5m to 10m. The zone containing the ultramafic rocks is approximately 25m wide. It is bordered to the east by the Northbrook Batholith and to the west by the Lessard Formation sediments. Within this zone, the ultramafic bodies are surrounded by mafic schists and gneisses. This observation is consistent with the idea presented earlier that the original ultramafic body has been tectonically pulled apart and surrounded by other rock-types.

Contacts between the Flinton Creek ultramafic rocks and the surrounding mafic rocks have a steep, but variable, dip (75-85°). The contact is always concordant with the schistose / gneissic foliation developed in the host rocks. An amphibole mineral lineation plunges down the dip of the foliation.

The degree of deformation in the surrounding mafic gneisses varies widely. The least-deformed rocks are medium to coarse-grained gabbros which occur in pods up to 1m in size, (see Plate 3.2a). These are surrounded by fine-grained, (most grains < 0.5mm), amphibolites which lack visible mineral segregation or foliation in hand specimen, although a weak amphibole foliation is visible in thin section. The degree of deformation in the mafic rocks increases as the contact with the ultramafic rocks is approached. The rock develops extremely thin (<1mm) mafic segregations composed mainly of amphibole. These segregations form thin planar structures, along which the rock cleaves. The mafic bands are separated from each other by plagioclase-rich segregations. The extreme lateral continuity of both mafic and felsic layers suggests that the rock is a straight gneiss (Hanmer, 1988). If the mafic and felsic layers are discontinuous, the rock develops a flaser-gneissic texture with small, (commonly < 1mm), epidote or plagioclase-rich "eyes" (see Plate 3.2b).

Microscopic examination reveals that all the mafic host rocks have been recrystallized. Triple junctions are common between plagioclase grains which display zoned yet uniform extinction. Amphiboles in the mafic layers have been recrystallized as they do not appear to be bent around the epidote augen.

Plate 3.2a Field photograph of coarse-grained gabbroic pods, similar in appearance to some of the Kaladar Complex gabbros, surrounded by more strongly deformed mafic gneisses. The gneisses appear to have been derived from the less-deformed material. Locality 103188, Hearthstone Anthophyllite Quarry (see Appendix III).

Plate 3.2b Photomicrograph of strongly deformed, mafic gneiss. Mafic segregations are composed predominantly of medium-grained, pleochroic amphibole. Felsic layers are composed of medium-grained plagioclase and fine-grained white mica. Note the presence of a large, dark epidote augen on the right-hand side of the photograph. The rock appears to have been recrystallised since plagioclase exhibits only weak undulatory extinction and amphiboles surrounding the large epidote augen appear to be neither bent nor broken. Sample 110188-01

Two reaction zone members immediately adjacent to the ultramafic boudins, the biotite zone and the chlorite zone, have an extremely strong schistosity. A late-stage crenulation cleavage, dipping 15-20°, is developed on the schistose surface. Sigmoidal biotite crystals were discovered in the biotite reaction zones surrounding lenses of amphibolite. Unfortunately, these were only found in loose blocks on the quarry floor. Since the original source material could not be located, the direction of movement implied by the biotite crystals could not be determined.

In contrast to the enclosing mafic rocks, the ultramafic rocks show little evidence of the deformation. As mentioned earlier, a spaced cleavage, defined by preferential growth of anthophyllite in the plane of the foliation, is developed near the host rock contact. The intensity of this foliation decreases rapidly over a distance of several metres moving into the ultramafic rock and therefore most of the anthophyllite-carbonate rock and all of the enstatite-magnesite rock are unfoliated. Enstatite crystals grow equally well in all directions and anthophyllite grows in spherical rosettes. A small number of these rosettes overgrow the marginal foliation in the anthophyllite-carbonate rock, providing evidence that mineral growth outlasted deformation. During field mapping, this spaced cleavage was usually noted on the margins of the ultramafic outcrops. This is yet another indication that ultramafic rocks have been pulled apart and boudinaged throughout the body.

3.4.1 Structural Geology - Discussion

Two possibilities concerning the emplacement of the Flinton Creek body are to be considered. The first involves intrusion of igneous material into the Flinton Group sediments followed by in-situ deformation and boudinage. The second proposes that mafic and ultramafic slices were emplaced into the Flinton Group sediments as part of a tectonic melange.

Evidence supporting the first hypothesis includes:

- 1) The concordant nature of the ultramafic lenses with the foliation in the surrounding mafic rocks.
- 2) The apparent increase in strain at the host rock - ultramafic contact.
- 3) The similar attitude of stretching lineations in both mafic gneisses of the Flinton Creek body and the Flinton Group conglomerates.

Since the area has been affected by several periods of deformation, the concordant contact of the ultramafic rocks with their host rocks does not rule out the tectonic emplacement proposal, since the ultramafic contact and surrounding structural fabric could have been transposed during any of the deformational periods. Furthermore, even though individual ultramafic lenses appear concordant, when taken as a group, they strike across an overturned syncline formed during D_1 (Thompson, 1972), (see Fig. 2.4, where the ultramafic rocks pass into the Flinton synclinorium). This means that the ultramafic rocks were emplaced after the first phase of deformation, which is further indicated by the lack of folding in the ultramafic rocks.

The relationship between the lineation and foliation in the mafic rocks of the Flinton Creek body is similar to that found in the nearby conglomerates of the Flinton Group where L_1 stretching lineations, defined by pebble elongation, tend to be steep and down the dip of the D_1 axial planar foliation. Since the Flinton Creek body was emplaced post- D_1 , it seems reasonable to suggest that the two lineations are unrelated. The exact time of formation of this fabric is uncertain; however, it is proposed that it is related to tectonic movement.

To produce the ultramafic lenses in the Flinton Creek body by boudinage requires extension in both the intermediate and maximum strain directions. The lack of other similar boudinaged structures in the Flinton synclinorium argues against the formation of ultramafic lenses by this process. Stretched pebbles in Flinton Group conglomerates show that the chief elongation direction was down-dip rather than along strike. Cross-bedding in the steeply dipping quartzites has not been transposed, which

would have been expected if a significant amount of strain had been strike-parallel. In conclusion, it is felt that the degree to which the ultramafic lenses have been pulled apart could not have been achieved without producing evidence for similar strain in the surrounding lithologies.

The strain gradient observed at the mafic - ultramafic contact results from a competence difference between the two dissimilar rock types. It cannot be used to distinguish between either of the two proposals since the local increase in strain could have developed during either late D_2 or D_3 deformation. Post-peak second-phase and later deformation can also account for the concordant nature of the ultramafic/host rock contact and for the development of the anthophyllite spaced cleavage in the border of the ultramafic lens.

The presence of gabbroic pods in finer-grained rocks of essentially the same composition was described earlier (see Plate 3.1). The existence of such features suggests that the mafic rocks have been inhomogeneously deformed. This type of deformation, coupled with the development of mafic straight gneisses, and the extreme variation in the size of the ultramafic slices strongly favours the second, tectonic emplacement, proposal.

It is possible to bracket the ultramafic rock tectonic emplacement time using field relationships. The lower age limit can be determined by noting that the Flinton Creek Ultramafic rocks crosscut D_1 (Thompson, 1972) structures in the area and must therefore have been emplaced after the first period of deformation. The development of the monominerallic reaction zones and recrystallization of the mafic gneisses provides an upper age limit since both require that the body be emplaced prior to peak regional metamorphism which occurred during D_2 . Therefore, the most likely time of emplacement for the Flinton Creek body would be at the beginning of D_2 , prior to peak metamorphic conditions.

The observation that late pegmatites never cross-cut the ultramafic rocks is difficult to reconcile with the existence of the metasomatic reaction zones around the

ultramafic lenses. The first observation suggests that the ultramafic rocks were emplaced after peak-metamorphism, since the pegmatites are interpreted to be the products of partial melting during the metamorphic heating, while the second observation argues for ultramafic emplacement prior to the regional metamorphic event. Two possible solutions to this enigma are suggested:

1) The ultramafic rocks behaved in a more ductile fashion than the surrounding lithologies, thereby avoiding pegmatite intrusion.

2) While the pegmatites are never observed intruding the ultramafic rocks, the opposite case of tectonically emplaced ultramafic cross-cutting the pegmatite is also not seen. Simple chance may be responsible for the negative observations in each case.

3.5 Geochemistry of the Flinton Creek Body

3.5.1 Introduction

Geochemical sampling of the Flinton Creek Ultramafic rocks was undertaken for two purposes.

- 1) To determine if lithologic and mineralogic changes observed in the ultramafic rocks (enstatite-magnesite rock vs. anthophyllite - carbonate rock) were related to changes in bulk chemical composition, rather than changes in metamorphic conditions.
- 2) To better understand and constrain the origin of the ultramafic rocks, for example, do the ultramafic rocks have an igneous or sedimentary protolith and if the protolith is igneous, were the rocks originally part of a layered intrusion, an ophiolite, or were they once komatiites?

Twenty-four samples were collected for whole-rock chemical analysis. The sample population may be broken down as follows, (see appendix 1 for exact descriptions):

- 1) Four enstatite-magnesite rocks, one altered to chlorite + serpentine.
- 2) Nine anthophyllite-carbonate rocks, four partially talcified
- 3) Six amphibolites
- 4) One ultramafic rock from the Actinolite body on the southern margin of the Elzevir batholith.
- 5) Four gabbros from the Kaladar Complex.

3.5.2 Bulk Composition of the Flinton Creek Ultramafic Rocks

Figure 3.2 is a three component diagram on which are projected the molar percentages of CaO , SiO_2 , and MgO determined from samples of the Flinton Creek rocks. These three components were chosen because they are the chief non-volatile constituents of the meta-ultramafic rocks. Also shown for comparison is the composition field covered by similar carbonate - orthopyroxenites from the Tromms region, (Ohnmach, 1974). Several observations can be made from this diagram:

- 1) The Flinton Creek ultramafic rocks are more calcic than their Norwegian counterparts, but span approximately the same Si:Mg ratio.
- 2) The anthophyllite-carbonate rocks are slightly more calcic than the enstatite-magnesite rocks; however, this statement should be treated with caution since the sample population is small and analytical error has not been taken into account.
- 3) The Flinton Creek rocks cluster in two chemically different groups, one being richer in lime and slightly poorer in silica than the other (although it is possible that the gap separating the two groups could be eliminated, or enlarged, if more samples were analyzed).

The calcium-rich group contains four anthophyllite-carbonate rocks (samples 1031-C1; 1104-C4, -C8, -C9), and one enstatite-magnesite rock (1015-C2). An examination of the sample descriptions, see Appendix A, reveals that all the samples in this group, with the exception of 1031-C2, were highly altered, (talcified or chloritized). In addition, all the calcium-rich samples come from within two metres of

Figure 3.2 Ternary diagram showing molar percent CaO-SiO₂-MgO for thirteen meta-ultramafic samples from the Flinton Creek Ultramafic rocks, one from the Actinolite body and two from the Kaladar Complex (Chappell, 1978). The compositional range covered by carbonate-orthopyroxenites from the Tromms region, Norway is shown by the stippled pattern (Ohnmach, 1974). Note that the Flinton Creek rocks appear to be more calcic than the Norwegian examples. Also note the broad calcium enrichment trend defined by Flinton Creek Ultramafic samples located close to the ultramafic-host rock contact.

the ultramafic - host rock contact and none have XRF totals greater than 83 weight percent. Thin sections of these rocks show that anthophyllite has been replaced by talc and enstatite has been pseudomorphed by fine-grained chlorite, serpentine and talc. The presence of these hydrous alteration products, combined with the large quantities of carbonate minerals present, explains the low analytical totals. It seems likely that the bulk chemistry of these rocks has been modified to a greater extent than the chemistry of the other ultramafic rocks; therefore these samples will not be utilized in later discussions concerning the origin of the Flinton Creek rocks.

The presence of the tremolite zone in the metasomatic reaction zone sequence has already been described. The formation of this monomineralic zone suggests, at the very least, that calcium was mobile during metamorphism. Since the Ca-rich samples are associated with the margin of ultramafic lenses, it is proposed that a similar metasomatic fluid carried calcium inwards from the surrounding host rocks, enriching the ultramafics. It is assumed that dolomite was formed as a result of these metasomatic fluids, since no other calcium-bearing phases have been described in the rocks from this group.

The slight calcium-enrichment of the anthophyllite-carbonate rocks, relative to the enstatite-magnesite rocks, is likely due to the same metasomatic process which created the Ca-rich group of ultramafics, see Fig 3.2. Calcium enrichment caused by contact metasomatism would be expected to affect the anthophyllite-carbonate rock to a greater degree since it surrounds the enstatite-magnesite rock.

Overall, the effect of this metasomatic fluid on the bulk composition of the ultramafic rocks at a distance greater than several metres from the ultramafic/host-rock contact appears to have been limited. The SiO_2 and MgO contents of both the enstatite-magnesite rock and the anthophyllite-carbonate rock are similar and since neither enstatite nor anthophyllite contains calcium, a difference in the amount of this component cannot explain why the anthophyllite-carbonate rock (which is rich in anthophyllite) appears to replace the enstatite-magnesite rock (which is rich in enstatite) as the ultramafic - host rock contact is approached.

3.5.3 Origin of the Flinton Creek Ultramafic Rocks

Different genetic theories have been proposed for enstatite-magnesite rocks. Barth (1926) favoured an igneous origin for carbonate - orthopyroxenites as the result of *in situ* crystallization of an ultramafic magma which had assimilated carbonates. Other authors, (Reitan and Geul, 1958; Randall, 1960), have suggested a sedimentary - metamorphic origin for these rocks. Reitan and Geul (1958) proposed that high-grade metamorphic differentiation of interbedded dolomites and pelites resulted in carbonate - orthopyroxenite formation. Randall (1960) interpreted the rocks as products of intensive metasomatism of limestone. If the Flinton Creek ultramafic rocks are the metamorphic and metasomatic products of igneous protoliths, then several other possible origins exist including:

- 1) Cumulates from a large diabase sill or stratiform intrusion.
- 2) Komatiitic magma.
- 3) Fragments of tectonically emplaced oceanic crust, (ophiolites) or upper mantle (alpine peridotites).

Field relationships prove that the enstatite-magnesite rocks cannot be considered as primary, unaltered, *in situ* crystallization products of an intruded magma. The presence of boudinaged slices, concordant contacts and contact metasomatic reaction zones, suggest that the ultramafic rocks have been at least partially metamorphosed and deformed with the surrounding host rocks.

Goles (1967) suggested that the average ultramafic rock contained approximately 2400 ppm chromium and 1500 ppm nickel. A summary of the whole-rock geochemical data for the Flinton Creek ultramafic rocks, presented in Table 3.1a., shows that the both the average chromium, (4500 ppm.) and nickel, (1800 ppm) concentrations exceed Goles' estimates. (In detail, all but four of the thirteen samples analyzed, contain greater than the suggested abundances of either Cr or Ni.) To reach such elemental concentrations in the sedimentary environment is viewed as unlikely, especially since Table 3.1a also shows that the rocks contain very little manganese or titanium, and no

TABLE 3.1a

Extreme and mean values from 13
enstatite-magnesite and
anthophyllite-carbonate rocks

	AVERAGE	MAXIMUM	MINIMUM
SiO ₂	40.35	51.79	25.80
Al ₂ O ₃	1.66	4.86	0.57
Fe ₂ O ₃ *	8.55	11.54	6.47
MgO	28.82	33.50	24.45
CaO	6.37	14.69	0.43
Na ₂ O	0.15	0.39	0.00
K ₂ O	0.02	0.10	0.01
TiO ₂	0.04	0.09	0.01
P ₂ O ₅	0.00	0.01	0.00
MnO	0.15	0.29	0.10
S	0.07	0.12	0.05
Ba	59	83	38
Cr	4544	7617	1575
Zr	<10	<10	<10
Sr	61	143	<10
Rb	1	13	<10
Y	<10	<10	<10
Nb	<10	<10	<10
Ni	1796	2302	1202
V	42	76	11
TOTAL	87.55	99.11	76.98

* total iron as Fe₂O₃

TABLE 3.1b

Extreme and mean values from 9
CO₂-poor ultramafic rocks
with XRF totals > 90 percent

	AVERAGE	MAXIMUM	MINIMUM
	46.97	51.79	39.12
	2.33	4.86	1.31
	10.00	11.54	6.51
	30.36	33.50	26.48
	2.40	4.47	0.43
	0.11	0.29	0.00
	0.01	0.02	0.01
	0.06	0.09	0.04
	0.01	0.01	0.00
	0.13	0.15	0.10
	0.07	0.12	0.05
	63	83	38
	4889	7617	2747
	<10	<10	<10
	30	98	<10
	<10	<10	<10
	<10	<10	<10
	<10	<10	<10
	1912	2302	1487
	53	76	28
	93.45	99.11	89.41

detectable phosphorus, elements which are to be expected in sedimentary rocks. For this reason, a sedimentary origin for the enstatite-magnesite rocks, as suggested by previous authors, is considered unlikely.

Based on high chromium and nickel concentrations in both the enstatite-magnesite rocks and the anthophyllite-carbonate rocks, it is proposed that prior to metamorphism and metasomatism, the Flinton Creek rocks belonged to the ultramafic family (the rocks were originally peridotites and/or pyroxenites). Similar conclusions were reached concerning the protolith of the carbonate - orthopyroxenites from the Tromms region in Norway, (Schreyer, et al., 1972, and Ohnmacht, 1974).

Table 3.1b shows the mean value of seven whole-rock analyses with totals greater than ninety percent. This limit excludes those rocks which contain significant amounts of hydrous alteration minerals and carbonate; therefore, it is felt that the rocks in this category come closest to the original pre-metamorphic composition.

Even with this restriction, the ultramafic rocks still display a considerable variation in SiO_2 (39 to 52 weight percent). Therefore, although the rocks are ultramafic, some of them cannot be considered ultrabasic since members of the ultrabasic rock suite, as defined by Wyllie (1967), contain less than 45 weight percent SiO_2 .

The average element concentrations for carbonate-poor ultramafic rocks given in Table 3.1b are shown again in Table 3.2. Shown for comparison are the average chemical compositions of a pyroxenite and a peridotite as determined by Le Maitre (1976). The SiO_2 content of the "average" Flinton Creek ultramafic rock is closer to that of a pyroxenite; however, the peridotite SiO_2 value is still approximately within one standard deviation, as given by Le Maitre (1976), of the Flinton Creek value. The MgO and CaO content of the Flinton Creek rock is far closer to that of the average peridotite and for this reason, it is concluded that the majority of the Flinton Creek ultramafic rocks are metaperidotites.

Le Baron et al. (1987) have suggested that the talc bodies discovered within

TABLE 3.2
Comparison of average Flinton
Creek Meta-Ultramafic analysis
with other common rock types

	Average	Maximum	Minimum	Average Pyroxenite*	Average Peridotite*	Average Harzburgite*	Average Dunite*
SiO ₂	46.97	51.79	39.12	46.27	42.26	39.93	38.29
Al ₂ O ₃	2.33	4.86	1.31	7.16	4.23	2.35	1.82
Fe ₂ O ₃ **	10.00	11.54	6.51	12.25	10.92	12.67	14.01
MgO	30.36	33.50	26.48	16.04	31.24	33.18	37.94
CaO	2.40	4.47	0.43	14.08	5.05	2.90	1.01
Na ₂ O	0.11	0.29	0.00	0.92	0.49	0.31	0.20
K ₂ O	0.01	0.02	0.01	0.64	0.34	0.14	0.08
TiO ₂	0.06	0.09	0.04	1.47	0.63	0.26	0.09
P ₂ O ₅	0.01	0.01	0.00	0.38	0.10	0.13	0.20
MnO	0.13	0.15	0.10	0.16	0.41	0.15	0.71
TOTAL	93.45	99.11	89.41	99.37	95.67	92.02	94.35
fe***	0.229	0.237	0.181	0.407	0.239	0.256	0.249

* Data from LeMaitre (1976)

** total iron as Fe₂O₃

*** fe = FeO/(FeO+MgO)

Tudor Formation volcanic rocks on the western margin of the Elzevir batholith are metamorphosed ultramafic komatiites. When the analyses from the Flinton Creek ultramafic rocks are plotted on a weight percent Al_2O_3 - $\text{FeO}+\text{Fe}_2\text{O}_3+\text{TiO}_2$ - MgO diagram, (Jensen Cation Plot), they also plot in the ultramafic komatiite field as defined by Jensen (1976). Table 3.3 summarizes primary element abundances (Table 3.3a: Viljoen and Viljoen, 1969; Bickle *et al.*, 1975; Nesbitt and Sun, 1976) and ratios (Table 3.3b: Ludden and Gelinas, 1982) for several komatiites from world-wide locations. Comparison of this data with that for the carbonate-poor ultramafic rocks in this study show several differences:

- 1) The Flinton Creek ultramafics are titanium-poor, averaging 270 ppm, relative to the other komatiites which have between 1000 and 2500 ppm titanium.
- 2) The Al_2O_3 : TiO_2 ratio for the Flinton Creek rocks is approximately 30 whereas the same ratio lies in the 15 to 20 range for recognised komatiites.
- 3) While the CaO : Al_2O_3 ratio typically lies in the 0.5 to 1.2 range for most komatiites, the ultramafic rocks from this study have higher ratios, (1.40) similar to that found in the Barberton Komatiites.
- 4) Analyses show ultramafic rocks in this study have no detectable zirconium or yttrium.

The differences in element ratios and abundances noted above suggest that the Flinton Creek Ultramafic rocks were not komatiites prior to metamorphism. Metasomatic processes could account for these differences; however, if the ultramafic rocks were selectively depleted in zirconium, yttrium and titanium and enriched in calcium. This would explain the low abundance of these elements and the higher Al_2O_3 : TiO_2 and CaO : Al_2O_3 ratios. Calcium metasomatism has already been documented in rocks near the margins of the ultramafic lenses; however, these rocks were not included in this comparison.

Figure 3.3 shows weight percent NiO plotted against weight percent Cr_2O_3 for analyses from the Flinton Creek Ultramafic rocks. A chromium enrichment trend is

TABLE 3.3a

Comparison of other known
komatiites with analyses from
the Flinton Creek Meta-
Ultramafic Rocks

	Flinton Creek	Barberton (1)	Rhodesia (2)	Ontario (3)
SiO ₂	46.97	48.26	49.06	45.58
Al ₂ O ₃	2.33	3.39	6.88	8.90
Fe ₂ O ₃ *	10.00	13.27	12.07	12.54
MgO	30.36	24.88	24.56	22.39
CaO	2.40	8.51	7.20	9.46
Na ₂ O	0.11	0.49	0.25	0.64
K ₂ O	0.01	0.05	0.17	0.09
TiO ₂	0.06	0.63	0.38	0.40
P ₂ O ₅	0.01	n.d.	0.08	0.02
MnO	0.13	0.20	0.20	0.23
S	0.07	n.d.	n.d.	n.d.
Ni	1912	707	n.d.	894
Cr	4889	2532	n.d.	3333
TOTAL	93.45	100.00	100.85	100.67
Fe**	0.229	0.324	0.307	0.335

* Total iron as Fe₂O₃

** Fe = FeO/(FeO+MgO)

n.d. not determined

TABLE 3.3b

Comparison of primary element ratios
for known komatiites and the
Flinton Creek Ultramafic rocks

	Ti (ppm)	Al ₂ O ₃ /TiO ₂	CaO/Al ₂ O ₃
<i>Flinton Creek Meta-Ultramafics</i>			
Average	390	29.01	1.38
Maximum	540	34.67	2.98
Minimum	300	21.83	0.26
<i>Abitibi, Canada (4)</i>			
Lamotte	2100	16-19	0.8-1.3
Munro	2400	19-22	0.9-1.25
<i>Western Australia (4)</i>			
Yilgarn	1000	20	1.0
<i>South Africa (4)</i>			
Barberton	1750	11	1.4-1.8
Rhodesia	--	19-22	1.0-1.25

(1) Average of samples 87J and 88J; Viljoen and Viljoen (1969)

(2) Average of samples NG213 and NG214; Bickle, Martin and Nisbet (1975)

(3) Average of samples 8, 9 and 10; Nesbitt and Sun (1976)

(4) Ludden and Gelinas (1982), pg. 337.

Figure 3.3 NiO (wt.%) versus Cr₂O₃ (wt.%) diagram after Malpas (1977); Malpas and Stevens (1979). The line drawn on the diagram separates NiO and Cr₂O₃ abundances for alpine ultramafic peridotites from ultramafic portions of tholeiitic stratiform intrusions. Note that only two of thirteen samples plot in the field defined by alpine peridotites, suggesting that the Flinton Creek ultramafic rocks may be part of a layered intrusive complex. As well, note the existence of a chromium-enrichment trend at approximately constant nickel. This trend is believed to be due to primary igneous processes, rather than being metasomatic in origin.

evident with approximately constant nickel. The line drawn on the diagram separates NiO and Cr_2O_3 abundances for alpine ultramafic peridotites from ultramafic portions of tholeiitic stratiform intrusions, represented by the Muskox Intrusion (after Malpas, 1977; Malpas and Stevens, 1979, pg. 34). Only two of the thirteen analyses fall in the alpine peridotite region. This, combined with the chromium enrichment trend suggests that the Flinton Creek rocks are ultramafic cumulates from either a stratiform-type intrusion or the ultramafic cumulate portions of an ophiolite. Field evidence supporting this conclusion was found in two locations where what may be relict igneous layering was discovered, (see Plate 3.3). It should be noted that the apparent chromium enrichment trend could be the result of chromium depletion of the Cr-rich ultramafic rocks at approximately constant nickel. This was considered and tested due to the already documented presence of metasomatic fluids. Both previously identified ultramafic rock groups, samples with high XRF totals (minor or no Ca-metasomatism) and samples with low XRF totals (Ca-metasomatism), plot over the entire length of the chromium trend. If one group had clustered around lower chromium abundances than the other, then an argument could have been made for a metasomatic origin of the apparent trend. However both groups are uniformly distributed and it is assumed that the chromium enrichment is due to primary igneous processes.

Analyses from the Kaladar Complex (Chappell, 1978; this study), and the Flinton Creek Body, including meta-ultramafics and metabasites, are plotted on a weight percent $\text{CaO} - \text{Al}_2\text{O}_3 - \text{MgO}$ diagram in Fig. 3.4a and an AFM diagram in Fig. 3.4b. Samples from the Kaladar Complex were analysed because Chappell (1978) reported the presence of ultramafic rocks in the Kaladar Complex similar to those found in the Flinton Creek Body. Therefore gabbro samples were collected to test for a possible genetic link between the two.

Figure 3.4a shows the fields for metamorphic peridotites, ultramafic cumulates, and mafic cumulates from ophiolite sequences (Coleman, 1977, pg.45). The five samples

Plate 3.3 Field photograph showing possible relict igneous layering in anthophyllite-carbonate rock. Cleavage in the rock is oriented at 90 degrees to the inferred layering, see cleavage symbol. Ruler shown for scale is 16cm long. Locality i02788 (see Appendix III).

Figure 3.4a CaO-Al₂O₃-MgO(wt.%) diagram showing the compositional fields covered by ophiolite sequence metamorphic peridotites, ultramafic cumulates and mafic cumulates (Coleman, 1977). Those samples not subjected to calcium metasomatism plot in the field defined by ultramafic cumulates. Note that the gabbros from the Kaladar complex have a general trend towards Al₂O₃ enrichment different from the trend displayed by ophiolite-related gabbroic rocks.

Figure 3.4b AFM diagram after Malpas (1977) and Malpas and Stevens (1979). Compositional fields covered by Bay of Islands gabbros, pyroxenites and dunites are shown along with several well known chemical differentiation trends. Note that most of the Flinton Creek ultramafic samples plot in, or near the field defined by Bay of Islands dunites. Gabbroic rocks from the Kaladar Complex show a well-defined calc-alkaline trend. While some of the metabasites associated with the Flinton Creek Ultramafic Rocks plot on this trend, other samples plot on the tholeiitic trend. This suggests that the metabasites associated with the Flinton Creek Ultramafic Rocks might have two different origins. Those samples which plot on the calc-alkaline trend may have gabbroic protoliths while those plotting on the tholeiitic trend may be volcanic.

previously identified as having been subjected to calcium-metasomatism plot outside the field of ultramafic cumulates, trending towards the CaO corner of the diagram, confirming this earlier conclusion. Results obtained earlier from the $\text{NiO} - \text{Cr}_2\text{O}_3$ diagram are confirmed in this diagram since the remaining ultramafic samples plot in the ultramafic cumulate field, with none plotting in the field for alpine peridotites (ultramafic tectonites).

Gabbroic samples from the Kaladar Complex show no apparent chemical link with the Flinton Creek Ultramafic rocks. The gabbros have a general trend towards Al_2O_3 enrichment different from the trend displayed by ophiolite-related gabbroic rocks. Few of the Kaladar Complex samples plot in the field covered by ophiolitic mafic cumulates since the Kaladar Complex rocks generally have higher $\text{Al}_2\text{O}_3:\text{CaO}$ ratios.

Metabasites analyzed from the Flinton Creek Body cannot be differentiated from Kaladar Complex gabbros since they plot on the same chemical trend.

The AFM diagram in Fig 3.4b (Malpas and Stevens, 1979, pg. 32), shows similar results to the previous diagram. As noted by Chappell (1978), the gabbroic rocks of the Kaladar Complex display a well defined calc-alkaline trend. This clearly falls off the trend defined by ophiolite related gabbros. The suggestion that the Flinton Creek Ultramafic rocks are metaperidotites is confirmed by the fact that most of the samples plot in the compositional field covered by Bay of Islands dunites rather than pyroxenites.

It is interesting to note that while some of the Flinton Creek metabasite samples plot near the calc-alkaline trend of the Kaladar Complex rocks, others plot on the Hawaiian tholeiitic trend. This suggests that the various metabasic rocks might be derived from two different protoliths; one being gabbroic, the other volcanic. This conclusion is supported by the observation that both coarse-grained (gabbroic) and medium-grained (volcanic) metabasites are associated with the ultramafic rocks. However, Sethuraman and Moore (1973) noted the presence of both calc-alkaline and

tholeiitic flows in the Tudor Formation; therefore both the calc-alkaline and tholeiitic metabasic varieties could have a volcanic origin.

Both of the ultramafic samples analyzed by Chappell (1978) from the Kaladar Complex have similar chemistry to the ultramafic rocks from the Flinton Creek body; however no obvious chemical link exists between these ultramafic rocks and the Kaladar Complex gabbros.

3.5.4 Summary - Geochemistry of the Flinton Creek Ultramafic Rocks

1) Ultramafic rocks near the host-rock/ultramafic contact have been enriched in calcium relative to rocks from the interior of ultramafic lenses.

2) In $\text{CaO-SiO}_2\text{-MgO}$ composition space, no systematic differences were noted between the bulk compositions of either the enstatite-magnesite rock or the anthophyllite-carbonate rock.

3) Chromium and nickel abundances confirm that the Flinton Creek ultramafic rocks had an igneous, rather than sedimentary protolith.

4) Comparison of un-metasomatised(?) whole-rock analyses with "average" ultramafic rock compositions suggests that the meta-ultramafic rocks were originally peridotites.

5) Comparison of the same data above (4) with whole-rock data from known komatiites suggests that the Flinton Creek rocks are not metamorphosed komatiites.

6) Low concentrations of nickel relative to chromium suggest that the Flinton Creek rocks are ultramafic cumulates either from a stratiform-type of intrusion or the cumulate portion of an ophiolite, rather than alpine-type peridotites.

7) The $\text{CaO} - \text{Al}_2\text{O}_3 - \text{MgO}$ diagram confirms the previous suggestion that the Flinton Creek Ultramafic rocks are cumulate rather than mantle derived. Calcium metasomatism of border-zone rocks may also be seen on this diagram.

8) The AFM diagram suggests that two different types of metabasic rocks, calc-alkaline and tholeiitic, are present in the Flinton Creek Body. The calc-alkaline rocks

may be metagabbroic or metavolcanic whereas the tholeiitic rocks are probably only metavolcanics. Flinton Creek Ultramafic rocks plot in the same field as Bay of Islands dunites rather than pyroxenites.

9) Major-element chemistry does not support a genetic link between the Flinton Creek Ultramafic rocks and the gabbros of the Kaladar Complex.

10) Points 1 to 8 above suggest that the Flinton Creek Ultramafic rocks are metamorphosed portions of either a stratiform-type of intrusion or the cumulate portion of an ophiolite.

3.6 Metamorphism in the Flinton Synclinorium

3.6.1 Introduction

In the Flinton syncline, the basal unit of the Bishop Corners Formation is the same pelitic schist that was studied by Hounslow and Moore (1967) in the Fernleigh syncline. Thompson (1972) mapped mineral occurrences within this unit between Madoc and Northbrook. Unfortunately, either due to a facies change in the Flinton Group sediments or to poor exposure of the basal unit, pelitic rocks within the Flinton synclinorium become scarce four kilometres south of the town of Flinton, near the northern-most ultramafic exposures. Pelitic rocks re-appear again in the Black River synclinorium 10km east of the southern-most ultramafic outcrop. This lack of exposure makes direct comparison between ultramafic and pelitic assemblages impossible. There are however, pelitic occurrences outside the Flinton syncline which, with minor extrapolation, allow comparison of the metamorphic assemblages in these two lithologies. A more direct comparison of metamorphic mineralogy is possible between ultramafic rocks and the impure calcareous rocks of the Lessard Formation, (Thompson, 1972, 1973).

3.6.2 "Impure" Calcareous Rocks

Thompson (1973) documented the prograde metamorphism of "impure" calcareous rocks belonging to the Lessard Formation.

This formation consists of two units, a biotite-carbonate schist (epidote-free calc-silicate paragneiss at higher metamorphic grade) and a calcareous psammite (higher grade equivalent = epidote-bearing calc-silicate paragneiss) (Thompson, 1972). Changes in the mineralogy of both units are consistent with a southward increase in metamorphic temperature and pressure within the Flinton synclinorium.

To account for the different mineral zoning patterns that result from bulk compositional differences between the schist and psammite members of the Lessard Formation, Thompson (1973) defined three general mineral zones based on mineral pairs that occur in the zonation in both units, see Fig. 3.5. These zones are:

- I) biotite-calcite zone;
- II) Ca-amphibole-microcline zone;
- III) diopside-microcline zone.

Zone I is further subdivided on the basis of the presence or absence and mode of occurrence of ferroan dolomite. In Zone Ia, Fe-dolomite occurs with phengite in the oxidized psammite and without phengite in the reduced schist. In zone Ib, epidote replaces Fe-dolomite in the presence of phengite and iron oxide. In zone Ic, ferroan dolomite is replaced by actinolite.

Comparing Fig. 2.4 and Fig. 3.5, it can be seen that with few exceptions, the Flinton Creek ultramafic rocks occur in Thompson's (1973) zone III, above the diopside isograd. Ultramafic rocks from the Hearthstone Anthophyllite quarry are found in zone II, while the two northern-most ultramafic occurrences are in zone Ic.

Figure 3.5 Generalized mineral zones in the Lessard Formation, from Thompson (1973). Zone I, biotite-calcite; Zone II, Ca-amphibole-microcline; Zone III, diopside-microcline. Zone I is subdivided due to differences in bulk composition of the schist and psammite units in the Lessard Formation (see Thompson (1973) pg. 70, for details). Comparison with Figure 2.4, which shows the general location of the ultramafic rocks within the Flinton Synclinorium, shows that all but two of the Flinton Creek Ultramafic Rock occurrences are found in mineral zone III. The two exceptional locations are the Hearthstone Anthophyllite Quarry, which occurs in zone II, and the two northern-most ultramafic occurrences, which are in zone Ic.

3.6.3 Pelites

The most northerly ultramafic occurrence in the Flinton Syncline is approximately coincident with the southernmost outcrop of pelitic rocks. At this location, Thompson (1972) reports the following mineral assemblage:

muscovite + biotite + garnet + chlorite + staurolite + kyanite.

Therefore the lowest grade ultramafic rocks have been metamorphosed to at least the kyanite - staurolite zone in pelitic rocks.

The Kaladar Metasediments occur in a narrow northeasterly trending belt, west of the Flinton Creek body (see Fig. 2.1). The most abundant lithology within the belt is a metamorphosed plutonic pebble conglomerate. Marble, calc-silicate, psammite and pelitic schist form the other major rock units (Psutka, 1976). Like the Flinton Group sediments, the Kaladar sediments lie unconformably on older plutonic rocks, in this case, the Northbrook Batholith (Psutka, 1976). Psutka (1976) noted the following mineralogy for the pelitic unit:

muscovite + biotite + staurolite + garnet + sillimanite

Staurolite is frequently found as armoured inclusions in garnet. Thus this locality is within the sillimanite zone, apparently near the upper limit of staurolite stability, (see Fig. 3.6).

The Black River syncline is found west of the town of Actinolite, approximately nine kilometres east of the study area. Thompson (1972) identified the following metamorphic minerals in pelitic schists from this location:

muscovite + biotite + garnet + staurolite

No aluminosilicate minerals were identified at this location. Pelites in the Black River syncline have not been metamorphosed above the temperature of staurolite breakdown.

3.6.4 Inferred Conditions of Ultramafic Metamorphism

Figure 3.6 summarizes the metamorphic conditions indicated by pelitic rocks surrounding the Flinton syncline. Staurolite plus kyanite appears to be stable north and west of the study area while sillimanite and staurolite breakdown occur just to the east. Thus it seems reasonable to assume the Flinton Creek Ultramafic rocks have been metamorphosed at pressure-temperature conditions between these two extremes. An upper and lower estimate of ultramafic metamorphic conditions can be obtained from the intersection of the pressure - temperature gradient derived for the Eastern Elzevir terrain with the upper and lower temperature pelitic isograds which surround the ultramafic rocks (see Fig. 3.6). The upper limit of metamorphism is believed to lie near 600C (875K) and 5.6 kbar (560 MPa), while the lower metamorphic limit is approximately 530C (800K) and 4.7 kbar (470 MPa), see Fig. 3.6.

Comparison of mineral assemblages in the Flinton Creek Ultramafic rocks with isograds found in surrounding rock types yields results in agreement with those reported by Trommsdorff and Evans (1974), for anthophyllite-carbonate and enstatite-magnesite assemblages in the Lepontine Alps. These authors report that enstatite - magnesite assemblages become stable just inside the diopside-calcite-quartz zone and persist into the sillimanite zone. Enstatite-magnesite assemblages from the Flinton Creek rocks are generally found above the diopside-microcline zone (Thompson, 1973) in the Lessard Formation (Moore and Thompson, 1980), and are believed to be in the sillimanite zone for pelitic rocks.

Figure 3.6 Inferred pressure-temperature conditions for ultramafic metamorphism based on surrounding pelitic mineral assemblages. A lower limit of ultramafic metamorphism is provided by the mineralogy of the southern-most (highest-grade), pelitic occurrence in the Flinton Synclinorium. At this location, Thompson (1972), reports co-existing muscovite + biotite + garnet + chlorite + staurolite + kyanite. The P-T conditions required for this univariant assemblage are shown by the left-hand stippled line. Psutka (1976), reported co-existing muscovite + biotite + staurolite + garnet + sillimanite in the Kaladar Metasediments east of the Flinton Creek Ultramafic Rocks. This univariant assemblage, indicated by the right-hand stippled line, provided an upper limit for ultramafic metamorphic conditions. The Flinton Creek Ultramafic Rocks must have been metamorphosed at conditions between these two limits, presumably close to the inferred P-T gradient for the Eastern Elzevir Terrain. Therefore, the phase relations between minerals observed in the Flinton Creek Ultramafic Rocks were investigated at three points, shown as open circles, along this gradient.

3.6.5 Prograde Metamorphism of the Flinton Creek Ultramafic Rocks

Despite the pressure-temperature gradient which exists in the area as suggested by changing pelitic and calcareous mineral assemblages, no progressive mineral changes were noted in the Flinton Creek ultramafic rocks. Enstatite-magnesite rock is found in the core of most ultramafic lenses from the Hearthstone Anthophyllite Quarry south. (The absence of this rock type from the two northern-most ultramafic occurrences is attributed to poor exposure.) Anthophyllite-carbonate rock is found over the entire strike length of the body, surrounding the enstatite-magnesite rock.

Chapter Four - Petrography of the Flinton Creek Ultramafic Rocks

4.1 Introduction

Approximately two-hundred thin sections were examined to better characterise the mineralogy of the ultramafic rocks and to determine the textural relationships between constituent minerals. Textures in both ultramafic rock types are complex, and are only partially understood at this time. The reader is also reminded that enstatite-magnesite rock was not found over the entire length of the body, only occurring as far north as the Hearthstone Anthophyllite Quarry.

4.2 Enstatite-Magnesite Rock

If the amount of carbonate in the enstatite-magnesite rock is small, the rock exhibits an interlocking texture comprised of mutually interfering, coarse-grained, subhedral enstatite crystals, see Plate 4.1a. They are colourless in plane-polarised light and have parallel extinction. Microprobe traverses across the large enstatite grains have revealed no compositional zonation (see Table AII.1, Sample 101788-1B). Enstatite crystals from different localities throughout the body have the same magnesian composition, approximately En_{90} (see Table AII.1). The lack of chemical zonation within individual crystals combined with the uniformity of pyroxene composition throughout the body suggests that the enstatite crystals have achieved chemical equilibrium, implying that the enstatite crystals have a metamorphic, rather than igneous, origin.

Even though the enstatite crystals are compositionally similar, there is textural evidence which suggests multiple periods of enstatite growth. This evidence includes:

- 1) The presence of well-developed kink-bands in some enstatite grains while others in the same thin-section appear to be undeformed.

- 2) Unusual, and contorted grain-boundaries between pyroxene crystals which suggest that an earlier generation of enstatite crystals have been partially resorbed prior to the final period of enstatite growth, see Plate 4.1b.

Plate 4.1a Photomicrograph of carbonate-poor enstatite-magnesite rock exhibiting a typical interlocking texture. At least seven mutually interfering, coarse-grained enstatite crystals can be found in the photograph. Field of view is 5mm long. Sample 101788-1B

Plate 4.1b Photomicrograph of two complexly intergrown enstatite crystals found in an enstatite-magnesite rock. Textures such as this suggest multiple periods of enstatite growth. Field of view is 5mm long. Sample 060489-2B.

Magnesite is the most abundant carbonate mineral present in the Flinton Creek carbonate-orthopyroxenites. Based on random microprobe analyses, it appears that most enstatite-magnesite rocks have a dolomite/magnesite ratio of less than approximately 0.1. Magnesite in these rocks is observed to be both in textural equilibrium with, and replacing, enstatite.

In carbonate-poor enstatite-magnesite rocks (those with <10% carbonate), magnesite occurs as large, 1-5cm single crystals in textural equilibrium with the surrounding coarse-grained enstatite crystals, see Plate 4.2a. Carbonate-rich varieties of the same rock-type have a matrix of fine-grained, anhedral magnesite "blebs" in which large, porphyroblastic enstatite crystals appear to be growing. These large enstatite crystals contain inclusions of the finer-grained magnesite, see Plate 4.2b.

Magnesite is also formed as a result of CO₂ metasomatism and replacement of enstatite. Magnesite grains formed in this process are similar to those found in carbonate-rich enstatite-magnesite rocks, occurring as irregular, fine to medium-grained, anhedral carbonate "blebs". This type of magnesite is found in carbonate-rich replacement patches and veins where the pre-existing enstatite has been almost entirely destroyed, see Plate 4.3a. This process may be described by the following metasomatic reaction:

"Blebby" magnesite is frequently associated with anthophyllite. Both minerals appear to have grown simultaneously since few anthophyllite needles cross-cut carbonate grains. Enclaves of the two minerals may be found within large single enstatite grains, and on enstatite grain boundaries where the two appear to be growing into the larger crystal (see Plate 4.3b). A reaction open only to volatile species, in particular H₂O and CO₂, may be written to describe the replacement of enstatite by anthophyllite and magnesite:

Plate 4.6 Photomicrograph of several optically continuous enstatite atolls (purple, labelled E) which have been surrounded by coarse-grained, type I olivine crystals. This texture suggests that type I olivine grew later than orthopyroxene. Olivine is intergrown with anthophyllite. Field of view is 5mm long. Sample 110488-C2B.

Plate 4.2a Photomicrograph of carbonate-poor, enstatite-magnesite rock showing apparent textural equilibrium between coarse-grained enstatite and magnesite crystals. Field of view is 5mm long. Sample 101788-1A.

Plate 4.2b Photomicrograph of carbonate-rich enstatite-magnesite rock with large, subhedral porphyroblastic enstatite crystals growing in a medium-grained matrix of anhedral magnesite. Field of view is 5mm long. Sample 011788-1.

Plate 4.3a Photomicrograph of relict enstatite crystal (grey), which has been almost completely replaced by anhedral, or "blebby" magnesite. Field of view is 5mm long. Sample 111587-2.

Plate 4.3b Photomicrograph showing replacement of enstatite (grey), by an anthophyllite - magnesite assemblage. Field of view is 5mm long. Sample 011788-1.

This reaction results in a 16.9 percent volume increase. Anthophyllite-magnesite rock produced by this reaction should be composed of approximately 90 percent anthophyllite, due to the large molar volume of anthophyllite relative to magnesite ($264.4 \text{ cm}^3/\text{mol}$ versus $28.018 \text{ cm}^3/\text{mol}$, Helgeson *et al.*, 1978). Few anthophyllite-carbonate rocks are this anthophyllite rich; however, many contain approximately 70 percent orthoamphibole suggesting that this equilibrium, coupled with later magnesite addition, is a reasonable explanation for anthophyllite-carbonate rock formation.

Reaction 4.2 above assumes chemical equilibrium existed between the solid phases with only the introduction of water and carbon dioxide. However, anthophyllite and magnesite could have been formed at different times in a stepwise sequence of reactions. First with the replacement of enstatite by magnesite according to reaction 4.1:

Followed by the growth of anthophyllite in the magnesium-carbonate:

Reaction 4.2 may be produced from 4.1 and 4.3 by addition with the elimination of $\text{SiO}_2(\text{aq})$. Thus reaction 4.2 assumes limited silicon mobility.

Table A2.2 shows that both textural varieties of magnesite have the same composition, (101788-1B is the coarse-grained type, 011788-01 has the "blebby" variety). Iron, calcium and manganese are the only elements which substitute for magnesium in the crystal structure; these elements show limited variation in magnesite from the enstatite-magnesite rock. Total ferrous iron varies between 3.5 and 4.7 weight percent. Lime and manganese oxide are constant at 0.4 and 0.2 weight percent respectively.

The Flinton Creek carbonate-orthopyroxenites frequently contain dolomite as well as magnesite. Dolomite occurs in small quantities, about 1 percent of the total rock volume, as fine to medium-grained, anhedral "blebs". It is always found co-existing in textural equilibrium with magnesite, often in a symplectic or wormy intergrowth (see Plate 4.4a). Unlike magnesite which sometimes forms euhedral

crystals, dolomite grains are always anhedral. Dolomite may be found in textural equilibrium with most silicate minerals, including enstatite and anthophyllite; however, in several thin sections, it appears to be a replacement product of tremolite. Table A2.3 shows several analyses of dolomite. Iron is the only element substituting for magnesium.

Talc is found both as a primary metamorphic mineral, and as a late-stage alteration product. Primary metamorphic talc occurs as medium to fine-grained, subhedral laths within enstatite crystals and along enstatite grain boundaries. The minerals are separated by sharp, well-defined margins and appear to be in textural equilibrium. Late talc is generally coarser-grained than its equilibrium counterpart and is commonly found as an irregular, patchy overgrowth replacing enstatite (see Plate 4.4b). On the assumption that all elements but Si are conserved, talc replacement of enstatite may be constrained by the following equilibrium:

The talc alteration appears to develop inwards from fractures and veins, suggesting that mass transfer through the fluid phase is an important part of the replacement process. Late-stage, vein-filling talc also occurs with chlorite. Talc in equilibrium with enstatite appears to approach its pure magnesian end-member composition (see Table A2.4), with only minor substitution of iron and smaller amounts of nickel for magnesium.

Chlorite is the main aluminum-bearing phase in the enstatite-magnesite rock. It occurs as medium to fine-grained, subhedral laths, commonly found within the large enstatite grains. The inclusions are usually oriented parallel to crystallographic planes in the enstatite. The results for microprobe analyses of several chlorite grains in equilibrium with enstatite are given in Table A2.5. Following the nomenclature of Hey (1954), chlorites from the enstatite-magnesite rock are members of the clinochlore group. Comparison of chlorite compositions with those of other minerals shows that chlorite contains the second highest concentration of chromium next to chromium-

Plate 4.4a Photomicrograph of a complex dolomite - magnesite intergrowth. Dolomite occurs as "wormy", anhedral patches (lighter, labelled D) in coarse-grained magnesite (darker, labelled M). This photograph also shows another example of the apparent textural equilibrium between magnesite and enstatite. Field of view is 5mm long. Sample 111587-1C.

Plate 4.4b Photomicrograph of an enstatite crystal (purple), which has been partially replaced by coarse-grained talc (light pink). Relict enstatite grains occur within the talc laths. Field of view is 5mm long. Sample 110488-C5.

magnetite. As well, chlorite in equilibrium with enstatite contains iron and minor amounts of nickel. The presence of chlorite in the enstatite has almost certainly influenced its composition. This is demonstrated by microprobe analyses of enstatite which show that it is almost devoid of aluminum, see Table A2.1.

Chlorite is also the most abundant alteration product in severely weathered enstatite-magnesite rocks, forming a fine-grained, mesh-textured mat with serpentine and talc.

Olivine has been found at all but one enstatite-magnesite rock occurrence in the Flinton Creek body. It is never an abundant phase, commonly comprising less than one percent of the total rock. The mineral is not uniformly distributed throughout the rock and it may be concentrated in ten-centimetre sized areas to the point where it locally comprises up to thirty percent of the rock volume.

In hand sample and thin section, olivine appears to have two distinct textural habits, see Plate 4.5. Olivine I occurs coarse-grained (4-8mm) crystals, found either singly or in groups, partially surrounded by enstatite (see Plate 4.5a). Olivine II forms 1-2mm grains which occur in mosaic textured anastomosing olivine veinlets, commonly 1-4mm wide (see Plate 4.5b). This second textural variety cross-cuts the surrounding enstatite-magnesite assemblage and was thus formed later; however, the relative timing of olivine I formation is not as clear.

On the macroscopic scale, the coarse-grained olivine has a texture which suggests that it is relict, having been surrounded and replaced by enstatite. Microscopic examination reveals the opposite. The olivine-rich areas, each composed of 1-4 large crystals, are replacing the bordering enstatite. Evidence for this includes:

- 1) The pristine nature of the olivine, even though it is surrounded by enstatite which has been extensively replaced by magnesite.
- 2) The presence of enstatite "atolls" surrounded by olivine, see Plate 4.6.
- 3) Embayments in enstatite grain boundaries filled with olivine.
- 4) The absence in olivine grains of undulose or strained extinction present in surrounding enstatite crystals.

Plate 4.5a Photograph of a thin section which contains type I (coarse-grained) olivine. One to three olivine crystals may be found in each of the large (1cm diameter), dark, sub-circular areas (labelled Ol I). Enstatite crystals (colourless mineral) surround the olivine grains. The texture shown suggests that the forsterite grains have been engulfed by enstatite. Note that the olivine-rich patches are connected to one another by tiny fractures. Oversized thin section is 7cm long. Sample 060489-2B.

Plate 4.5b Photograph of a thin section containing type II ("blebby" or vein related) olivine. Irregular veinlets of fine-grained olivine (dark green, labelled Ol II) crosscut the surrounding enstatite + magnesite assemblage (colourless). Oversized thin section is 7cm long. Sample 050888-2A.

Plate 4.6 Photomicrograph of several optically continuous enstatite atolls (purple, labelled E) which have been surrounded by coarse-grained, type I olivine crystals. This texture suggests that type I olivine grew later than orthopyroxene. Olivine is intergrown with anthophyllite. Field of view is 5mm long. Sample 110488-C2B.

Thus it appears that both olivine varieties formed later than the surrounding enstatite-magnesite assemblage.

In addition to similar formational timing, olivine I and olivine II both appear to be fluid / vein related since the large olivine I patches are connected to one-another by narrow, (<0.01mm wide), veinlets filled with serpentine and chlorite.

Both textural varieties have similar compositions, (Fe_{90}) see Table A2.6. Sample 060489-2B is an example of olivine I (pseudo-relict), while 050888-2A is an example of olivine II ("blebby"). Nickel, as well as iron substitute for magnesium.

The mineral changes leading to late-stage olivine formation were discovered during examination of the monominerallic olivine veinlets. At the microscopic scale, these veins are not in sharp contact with the surrounding enstatite-magnesite assemblage. Each veinlet has a core of irregular, "blebby" olivine which passes transitionally into the surrounding enstatite-magnesite rock, see Plate 4.7a. In this transition zone, olivine, which has a texture similar to that found in the core zone, is observed to pseudomorph the pre-existing magnesite alteration "blebs" in enstatite, see Plate 4.7b. The observed presence of enstatite, magnesite and forsterite and the assumed presence of the CO_2 bearing fluid suggests that the following equilibrium constrained the process of olivine formation:

It should be stressed that enstatite and magnesite co-exist in textural equilibrium elsewhere in the rock and forsterite formation appears to be controlled by vein proximity (fluid availability?).

Three different amphibole group minerals, tremolite, anthophyllite and magnesio-cummingtonite, are found in the enstatite-magnesite rock. The three cross-cut and overgrow each other producing complex textural relationships which are not fully understood.

Plate 4.7a Photomicrograph of a type II olivine veinlet. Three different mineral zones can be identified: Zone 1 contains enstatite which has been partially replaced by magnesite. Zone 2 contains anhedral "blebs" of forsterite which appear to be pseudomorphing similar textured magnesite. Zone 3 is contains only olivine. Note the apparent random distribution of opaque minerals throughout the photograph. Field of view is 5mm long. Sample 050888-2A.

Plate 4.7b Close-up photomicrograph of textures found in zone 2 on the margins of type II olivine veinlets (see Plate 4.7a above). Enstatite atolls (E) appear to be partially surrounded by olivine, suggesting late-stage forsterite growth. Note the similarity in grain-size and texture between olivine and the magnesite which had previously replaced enstatite grains. Field of view is 2.5mm long. Sample 050888-2A

Tremolite is the second of two Ca-bearing phases found in the enstatite-magnesite rock. It occurs as medium-grained, subhedral crystals which are commonly found in parallel aggregates at enstatite grain boundaries and junctions. Tremolite needles are generally wider than anthophyllite needles, permitting the former to be distinguished by its more equant shape. Textural evidence suggests calcium-amphibole was out of equilibrium with magnesite and anthophyllite when the final mineral assemblage was "frozen in". Several observations in different thin sections support this conclusion including:

- 1) Fine-grained anthophyllite needles are always observed overgrowing coarser-grained tremolites. The anthophyllite appears to nucleate preferentially on exposed ends of (001) tremolite needles and progresses inwards.
- 2) Tremolite grains in contact with magnesite are pitted and corroded, see Plate 4.8a.
- 3) Tremolite was observed to be reacting to a dolomite plus anthophyllite assemblage, see Plate 4.8b.

A solid-phase reaction can be written to explain this last observed breakdown:

Tremolite analyses are shown in Table A2.7. Substitution of iron for magnesium represents the only major deviation from the ideal mineral formula.

Magnesio-cummingtonite has been tentatively identified in the enstatite-magnesite rock. It occurs as large (5mm.) crystals, exhibiting inclined-, irregular or wavy extinction (see Plate 4.9a). Magnesio-cummingtonite crystals tend to be wider and therefore more equant than the other two amphibole varieties. The mineral is biaxial negative with a maximum extinction angle of 15 degrees. First order interference colours are most common but low second-order colours (blue) may also occur in thin sections that are 30 microns thick. Cummingtonite interference colours are always lower than anthophyllite. Magnesio-cummingtonite appears to be

Plate 4.8a Photomicrograph of apparent textural disequilibrium between tremolite (Tr) and magnesite (M). Tremolite needles in contact with magnesite have irregular, corroded grain boundaries. Tremolite crystals appear to be in textural equilibrium with surrounding enstatite grains (near extinction). Field of view is 2.5mm long. Sample 111587-1C.

Plate 4.8b Photomicrograph of relict tremolite (light-green mineral, labelled Tr) which has been partially replaced by anthophyllite (A) and dolomite (D). Field of view is 1.5mm long. Sample 110488-C2B.

Plate 4.9a Photomicrograph of typical magnesio-cummingtonite from the Flinton Creek Ultramafic Rocks. Note that the mineral has lower first order interference colours; inclined, irregular extinction; and an overall felted appearance. Field of view is 2.5mm long. Sample 050888-1A.

Plate 4.9b Photomicrograph of anthophyllite (A) overgrowing talc (Tc) and enstatite (E). Field of view is 5mm long. Sample 032188-3B.

microscopically intergrown with a second type of amphibole. Both have inclined extinction, the first with an extinction angle of approximately 10 degrees, the second has a larger angle of 20 degrees. No difference in birefringence was noted between the two. As expected, microprobe analyses yield unusual results (see Table A2.8), which indicate that one of the intergrown amphiboles is calcic (tremolite?) while the other is magnesian (magnesio-cummingtonite?).

Anthophyllite may be found as single acicular crystals and in parallel to radiating aggregates. Extremely fine needles also combine to form fibrous rosettes. The anthophyllite crystals are colourless, with parallel extinction. Interference colours may be as high as mid-second order, (yellow in standard-thickness thin section). With the exception of phases formed due to late-stage retrograde metamorphism anthophyllite appears to be the last silicate formed in the paragenetic sequence. Anthophyllite needles may be found overgrowing medium-grained talc laths inside large enstatite crystals, see Plate 4.9b. The talc which is being replaced appears to be the equilibrium, rather than the replacement variety. This overgrowth may be described by the following solid phase reaction:

Anthophyllite needles may be found replacing both enstatite and forsterite. These replacement processes are described by the following reactions.

Table AII.9 shows that iron substitution in the anthophyllite crystal structure causes the only major deviation from the ideal end-member composition.

Opaque minerals comprise up to five percent of either ultramafic rock type. Reflected light microscopy reveals that in most cases, the opaque mineral is chromium-magnetite; however minor amounts of pyrite and pyrrhotite also occur. Table A11.10 shows the results of microprobe analyses of chromium-magnetite in enstatite-magnesite rock. The mineral is found scattered uniformly throughout both ultramafic rock types as fine to medium-grained, anhedral to subhedral crystals. In the enstatite-magnesite rock, the opaque grains appear to have been overgrown by enstatite.

4.3 Anthophyllite-Carbonate Rock

Magnesite and dolomite are more abundant in the anthophyllite-carbonate rock than the enstatite-magnesite rock. In total, carbonate minerals comprise roughly thirty to fifty percent of the total volume of this rock. Magnesite is generally the most abundant carbonate mineral; however, the dolomite/magnesite ratio is strongly dependent on the proximity of the ultramafic / host-rock contact. Although the total amount of carbonate remains constant, as this boundary is approached, the amount of dolomite in the rock increases so that about one metre away from the contact, dolomite comprises up to ninety percent of the carbonate present. Dolomite is the only calcium-bearing phase present in the anthophyllite-carbonate rock.

Texturally, magnesite is found as coarse to medium-grained, anhedral "blebs". When the percentage of dolomite present in the rock is small, it occurs as fine to medium-grained, anhedral "blebs" within the coarser-grained magnesite and between anthophyllite needles. However dolomite grain-size appears to increase with the amount of dolomite present. Meta-ultramafic rocks near the host-rock contact are anthophyllite - dolomite rocks consisting of medium to coarse-grained, anhedral dolomite in textural equilibrium with anthophyllite. Dolomite and magnesite found in the anthophyllite-carbonate rock generally contain more inclusions than in the enstatite-magnesite rock and thus appear "dirtier".

Anthophyllite is the most common silicate in the anthophyllite-carbonate rock. As in the enstatite rock, it is found as medium-grained, subhedral needles which occur in parallel or radiating aggregates (rosettes), see Plate 4.10a. Since the anthophyllite needles do not frequently cross-cut carbonate grains, it is assumed they both formed simultaneously. No compositional differences could be detected between anthophyllite from enstatite-magnesite rocks (110488-C2B) or anthophyllite from anthophyllite-carbonate rocks, (103188-C1) (see Table AII.9).

Magnesio-cummingtonite is found in minor amounts (<5%), in the anthophyllite-carbonate rock, see Plate 4.10b. It occurs as fine needles which grow together forming large laths up to 1cm across. As in the enstatite-magnesite rock, a second amphibole appears to be intergrown with the magnesio-cummingtonite giving the crystals wavy, irregular extinction and a felted appearance. Coarse-grained, anthophyllite needles are also found overgrowing the cummingtonite. The mineral has also been partially replaced by magnesite.

Chlorite is less abundant in this rock type than in the enstatite-magnesite rock. It occurs throughout the anthophyllite-carbonate rock as randomly oriented subhedral laths.

Talc appears to be more abundant in this rock type than in the enstatite-magnesite rock, composing from five to fifteen percent of the total volume. It has two textural habits. Talc may be found as medium-grained, subhedral laths displaying excellent textural equilibrium with anthophyllite and magnesite. It also occurs as a fine-grained alteration product which pseudomorphs anthophyllite. Comparison of the composition of talc from anthophyllite-carbonate rocks with talc from enstatite-magnesite rocks (see Table AII.4), shows that talc from the former rock-type contains more aluminum than talc from the latter.

An unknown, highly pleochroic mineral, with pleochroic colours ranging from light yellow-green to dark emerald green, has been identified only in the anthophyllite-carbonate rock. The mineral has a well developed basal cleavage and is tentatively

Plate 4.10a Photomicrograph of typical anthophyllite - carbonate rock. Subhedral anthophyllite needles occur in radiating aggregates which are surrounded by a matrix of medium-grained, anhedral magnesite or dolomite. Field of view is 5mm long. Sample 011788-1.

Plate 4.10b Photomicrograph of anthophyllite - carbonate rock. Thin section contains magnesio-cummingtonite (MC) and talc (Tc) in addition to those minerals mentioned above in Plate 4.10a (anthophyllite, dolomite and magnesite). Field of view is 5mm long. Sample 10318-C1.

identified as chromium - phlogopite, although no satisfactory microprobe analyses could be obtained. The unknown mineral appears to be in textural equilibrium with talc.

4.4 Discussion - Petrography of the Flinton Creek Ultramafic Rocks

A paragenetic sequence for the Flinton Creek Ultramafic rocks was constructed using the petrography discussed in this chapter (see Fig. 4.1). The sequence depicted is a first order approximation, due to the complex textures present in the ultramafic rocks which make it difficult to determine the relative timing of mineral growth. Figure 4.1 shows that:

- 1) The anthophyllite-carbonate rock developed from the enstatite-magnesite rock.
- 2) The assemblage talc and enstatite became unstable relative to anthophyllite over time.
- 3) Dolomite-bearing assemblages became predominant relative to tremolite bearing assemblages.
- 4) Forsterite formation post-dates magnesite replacement of enstatite and is therefore late.
- 5) Anthophyllite growth post-dates tremolite and magnesio-cummingtonite formation.

In the previous section, it was suggested that the following equilibrium constrained observation 1) above:

Likewise, the following equilibrium was suggested for observation 2):

In both cases it is assumed that anthophyllite formation was due to an equilibrium process. Evidence in support of this assumption includes repeated observation of anthophyllite overgrowing talc and enstatite and the ubiquitous association of anthophyllite and magnesite. As well, microprobe analyses revealed that minerals have uniform compositions throughout the ultramafic body. This suggests that chemical equilibrium was maintained during metamorphism. Both the anthophyllite-magnesite association and the overgrowth of talc + enstatite by anthophyllite could be the result of kinetic factors however, if anthophyllite nucleates preferentially in carbonate and talc. Anthophyllite replacement of enstatite (Equilibrium 4.8) and forsterite (Equilibrium 4.9) has already been documented; therefore the direct replacement of talc by anthophyllite cannot be ruled out, and an equilibrium process involving both talc and enstatite (Equilibrium 4.7) might not be necessary for anthophyllite formation. However, the fact the anthophyllite needles rarely cross-cut magnesite grains combined with the observation that anthophyllite and magnesite generally occur in a 1:1 ratio suggests that anthophyllite grew in equilibrium with magnesite (as per equilibrium 4.2).

Ohnmacht (1974) and Schreyer *et al.* (1972), both report the occasional presence of corroded olivine in Norwegian carbonate - orthopyroxenites. Using this textural evidence, these authors concluded that enstatite is replacing relict olivine. Ohnmacht (1974) suggested that this replacement occurs according to the reaction:

The same author found only a few examples of olivine replacement by orthopyroxene and magnesite, the reverse of reaction 4.2. Whether the enstatite-magnesite member of the Flinton Creek Ultramafic rocks was formed by similar reactions is unknown due to a paucity of mineral assemblages which pre-date the enstatite-magnesite assemblage.

The timing of forsterite formation is therefore a major difference between the Flinton Creek Ultramafic rocks and the Norwegian carbonate-orthopyroxenites. Textural evidence, such as uniform olivine extinction (compared with enstatite which generally has undulose extinction) and lack of magnesite alteration, suggests that both

olivine I and II formed later than enstatite. Forsterite in the Flinton Creek rocks thus appears to be the result of late-stage metamorphic processes, rather than being relict igneous or early metamorphic in origin as suggested for the Norwegian occurrences (Ohnmacht, 1974, Schreyer *et al.*, 1972). A further textural distinction in the Flinton Creek rocks is the fact that olivine appears to be vein related, rather than occurring in isolated grains as in the Norwegian Sagvandites. Therefore, any explanation for forsterite formation must take into account the fact that enstatite + magnesite appears to be stable elsewhere in the rock except in close proximity to the olivine veinlets.

The observed mineral zonation surrounding the olivine II (fine-grained, anhedral or "blebby" olivine) veinlets (olivine - olivine+enstatite+magnesite - enstatite+magnesite) combined with the apparent vein (fluid) association suggests the following equilibrium constrained olivine II formation:

Note that reaction 4.5 is the reverse of a reaction Ohnmacht (1974) proposed for carbonate-orthopyroxenite formation from peridotite. It is assumed that olivine I (coarse-grained, anhedral olivine) was formed under similar conditions. The exact cause of the textural difference between olivine I and II is unknown. It is postulated that olivine grain-size and texture is related to the grain-size of the original magnesite which it partially replaced. The observation that magnesite itself has two textural habits; a coarse-grained "equilibrium" variety and a medium-grained "blebby" replacement type, supports this theory. Thin sections which contain type I coarse-grained olivine tend to also contain equilibrium-type magnesite, while enstatite in thin sections with type II "blebby" olivine has usually been heavily replaced by magnesite.

Rice, Evans and Trommsdorff (1974) reported the presence of magnesio-cummingtonite in an olivine-tremolite-cummingtonite-chlorite assemblage (with or without enstatite or magnesite) from the Lepontine Alps, Ticino, Switzerland. In these

rocks, cummingtonite occurs as prisms or needles extending homoaxially from tremolite crystals (Rice *et al.*, 1974). The same authors also report the frequent intergrowth of anthophyllite and cummingtonite.

Unlike the Alpine occurrences, magnesio-cummingtonite from the Flinton Creek ultramafic rocks is found as an intergrowth, rather than overgrowth, of tremolite. This conclusion is based on the observation of patchy or wavy extinction of magnesio-cummingtonite grains, produced by the intergrowth of two different mineral lamellae with inclined extinction, combined with microprobe analyses (see Table AII.8). This intergrown texture suggests that either tremolite and magnesio-cummingtonite were stable simultaneously, or that tremolite was epitaxially overgrown by magnesio-cummingtonite.

Rice *et al.* (1974) suggested that the co-existence of anthophyllite and magnesio-cummingtonite in parallel lamellae represented arrested stages in the inversion of one amphibole to the other during cooling. Specifically, Rice *et al.* (1974) postulated that the orthoamphibole (anthophyllite) was replacing the clinoamphibole (cummingtonite). Petrographic observations from the Flinton Creek Ultramafic rocks partially support these authors. Anthophyllite needles were observed overgrowing magnesio-cummingtonite grains, implying that the orthoamphibole became stable later than the clinoamphibole; however, whether this happened during cooling or heating of the rocks is unknown. Furthermore, it is not known whether the intergrowth of the anthophyllite and magnesio-cummingtonite in the Flinton Creek rocks is due to exsolution, as suggested by Rice *et al.* (1974), or due to the epitaxial overgrowth of anthophyllite on magnesio-cummingtonite.

Several minerals, including talc, chlorite and chromium magnetite, occur less frequently in the ultramafic rocks than those minerals already discussed.

Microprobe analyses suggest that the aluminum content of talc is related to the rock type it is found in. Table AII.4 lists the average analyses for talc from enstatite-magnesite rock (0.15 wt. % Al_2O_3) and anthophyllite-carbonate rock (1.13 wt. % Al_2O_3).

Thus talc from the anthophyllite-carbonate rock appears to be more aluminous than talc from the enstatite-magnesite rock. Anthophyllite-carbonate rocks generally contain a smaller amount of chlorite than enstatite-magnesite rocks and it therefore seems reasonable to suggest that talc is more aluminous in the anthophyllite-carbonate rock in response to this reduced chlorite abundance.

Chlorite occurs intergrown with enstatite in the enstatite-magnesite rock. Chlorite laths appear to be crystallographically oriented suggesting that either both minerals grew simultaneously, with chlorite nucleating on specific crystallographic sites in the enstatite, or that chlorite grew later in the already formed enstatite crystals. However this second possibility must be considered unlikely.

Although the exact formation process is unknown, the presence of chlorite appears to have influenced the enstatite composition. Table AII.1 shows that enstatite crystals from the Flinton Creek Ultramafic rocks are aluminum-poor, with a maximum of 0.24 wt. % Al_2O_3 (sample 032188-2B). Enstatite co-existence with chlorite is interpreted to be responsible for this Al-poor composition.

Chromium magnetite grains are found scattered throughout both the enstatite-magnesite rock and the anthophyllite-carbonate rock. It is often possible to find several medium to coarse-grained, anhedral opaque grains within a single coarse-grained enstatite crystal. This suggests that the large enstatite crystals engulfed the pre-existing opaque grains. Since chromium-magnetite is a common product of serpentinization (Coleman and Keith, 1971), the opaque grains could indicate that the Flinton Creek Ultramafic rocks passed through a serpentinite stage prior to metamorphism. This is supported by the fact that chromium magnetite pre-dates the existing metamorphic mineral assemblage and is scattered randomly throughout both rock types.

In conclusion, metamorphism of ultramafic rock is believed to have resulted in the formation of the minerals which compose the enstatite-magnesite rock and the anthophyllite-carbonate rock. Textural evidence documented in the previous two

sections suggests that several mineral assemblages became unstable during metamorphism, either due to changing pressure - temperature conditions or changing fluid composition. Microprobe analyses show that the mineral assemblages produced during the metamorphism of the ultramafic rocks are equilibrium assemblages.

Chapter Five

Phase Relations in a Portion of the System CaO-MgO-SiO₂-H₂O-CO₂

5.1 Introduction

Various authors have studied phase relations for portions of the system CaO-MgO-SiO₂-H₂O-CO₂. Unfortunately, none has dealt with mineralogy specific to the Flinton Creek ultramafic rocks.

The basic thermodynamic equations used to derive curve shapes on T-X sections was presented by Greenwood (1962). Using thermodynamic data from earlier work on three component subsystems, and experimental data obtained for the four component reaction, magnesite + talc = forsterite + H₂O + CO₂, Greenwood (1967) constructed a T-X section for the system MgO-SiO₂-H₂O-CO₂ at 2000 bars (see Fig. 5.1). Several features of this diagram are worth noting. Serpentine stability is restricted to low X_(CO₂) values, and the assemblages anthophyllite + magnesite and enstatite + magnesite are restricted to high X_(CO₂) values.

Johannes (1969) reported the results of an extensive experimental investigation of the system MgO-SiO₂-H₂O-CO₂. Three reactions of particular interest to this study were experimentally determined at a pressure of 2000 bars and temperatures up to 660C. The reactions investigated occur only in fluids rich in CO₂ and are as follows:

Using the experimental results combined with chemography, Johannes (1969) was able to construct a T-X_(CO₂) diagram at P=2000 bars (see Fig. 5.2). The experimentally determined results closely match the calculated results of Greenwood (1967); enstatite+magnesite and anthophyllite+magnesite assemblages both require high concentrations of CO₂ in the fluid phase, with X_(CO₂) generally greater than 0.83.

Figure 5.1a Calculated equilibrium curves for the CO₂-rich side of the 2000 bar T-X section. Shaded area indicates the T-X region where enstatite + magnesite bearing assemblages are stable. Stippled area shows the limit of anthophyllite + magnesite bearing assemblages. Diagram taken from Greenwood (1967).

Figure 5.1b Calculated equilibrium curves for the H₂O-rich side of the 2000 bar T-X section. Shaded area indicates the limit of serpentine stability. Diagram taken from Greenwood (1967).

Figure 5.2a Phase relationships for the CO_2 -rich side of the 2000 bar T-X_{CO_2} diagram. Curves 2, 4, 16, and 17 were experimentally determined by Johannes (1969) (experimental conditions are shown as numbered black squares). The position of the remaining curves was estimated. Horizontally ruled field shows the stability region of the assemblage enstatite + magnesite; perpendicularly ruled field shows the stability region of the assemblage anthophyllite + magnesite. Diagram taken from Johannes (1969), pg. 1097.

Figure 5.2b T-X_{CO_2} diagram for phases in the system $\text{MgO-SiO}_2\text{-H}_2\text{O-CO}_2$ at $P_{\text{fluid}}=2000$ bars, from Johannes (1969, pg. 1099). The solid-line equilibrium curves were experimentally determined (Johannes, 1969) while the dashed curves were derived from intersection point relationships of the equilibrium curves. Various magnesite-bearing assemblages have been listed. Abbreviations: A, anthophyllite; B, brucite; E, enstatite; F, forsterite; M, magnesite; P, periclase; Q, quartz; S, serpentine; T, talc.

5.2b

5.2a

Evans and Trommsdorff (1974) completed a field and thermodynamic study of phase relations in calcium-poor, CO₂-metasomatised ultramafic rocks from Val d'Efra, in the Central Alps. Although the rocks studied by Evans and Trommsdorff (1974) were poor in calcium, the results of their paper are of interest to this study since they dealt with the stability of talc + enstatite versus anthophyllite.

5.2 Choice of Components

The Flinton Creek ultramafic rocks may contain up to 10 weight percent iron; however, ferrous iron and manganese commonly substitute freely for magnesium in silicate crystal sites. Thus for the purpose of this study, only magnesium will be treated as a component. Iron and manganese will be treated as dilutants of the magnesium components.

Magnesian-chlorite is the primary aluminous mineral in the rocks. As it is always present, and does not appear to be involved in any reactions with higher or lower-grade aluminous minerals, aluminum has been removed from the list of components with an understanding that all assemblages contain chlorite.

5.3 Choice of Phases

Seven mineral phases observed in the Flinton Creek ultramafic rocks were selected for study. These are listed in Table 5.1 below. Note that several common meta-ultramafic minerals have not been observed in the Flinton Creek rocks including: quartz, diopside and serpentine. An enlarged study of phase relations, including the unobserved phases, was beyond the scope of this thesis.

Table 5.1
Mineral Formulas and Abbreviations

Mineral	Abbreviations	Formula
Anthophyllite	A, Anth	$\text{Mg}_7\text{Si}_8\text{O}_{22}(\text{OH})_2$
Dolomite	D, Dol	$\text{CaMg}(\text{CO}_3)_2$
Enstatite	E, En	MgSiO_3
Forsterite	F, Fo	Mg_2SiO_4
Magnesite	M, Mag	MgCO_3
Talc	Tc	$\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_2$
Tremolite	Tr	$\text{Ca}_2\text{Mg}_5\text{Si}_8\text{O}_{22}(\text{OH})_2$

5.4 Chemography

Fig. 5.3 shows the position of the phases selected for study in the ternary compositional space CaO-MgO-SiO_2 . Also shown for comparison is the bulk compositional field covered by the Flinton Creek ultramafic rocks.

The phases talc, tremolite, dolomite, and magnesite are chemographically external. Thus for the purpose of this study they are always stable. Phase diagrams resulting from this analysis may only be applied to bulk compositions which fall within this quadrilateral.

5.5 Combinatorial Analysis

For the three component system CaO-MgO-SiO_2 , which describes only the solid phases present, Gibb's phase rule ($f+p=c+2$) predicts that a four phase assemblage is univariant, with one degree of freedom ($f=1$). Likewise a five phase assemblage will be invariant. It is assumed that a fluid phase, consisting only of CO_2 and H_2O , was always present during metamorphism of the ultramafic rocks. The combinatorial theorem may be used to predict the maximum number of univariant reactions and invariant points in this portion of the system described. The number of invariant points may be calculated by determining the number of ways to put seven solid phases together five at a time or:

$$\frac{7!}{5!(7-5)!} = 21 \text{ possible invariant points}$$

Likewise to calculate the number of reactions:

$$\frac{7!}{4!(7-4)!} = 35 \text{ possible univariant reactions}$$

The 21 possible invariant points are listed in Table 5.2. Following Schreinemakers (1915-1925), absent phases are listed within brackets.

Figure 5.3 Projection of minerals commonly found in metamorphosed ultramafic rocks into the system CaO-SiO₂-MgO (mole percent), assuming a mixed H₂O-CO₂ fluid is present in excess. Ruled area shows the approximate bulk compositional range of the Flinton Creek Ultramafic Rocks.

Table 5.2**Fluid-bearing Invariant Points in the sub-system CaO-MgO-SiO₂**

(Anth, Dol)	En, Fo, Mag, Tr, Tc	(Dol, En)	Anth, Fo, Mag, Tr, Tc
(Anth, En)	Dol, Fo, Mag, Tr, Tc	(Dol, Fo)	Anth, En, Mag, Tr, Tc
(Anth, Fo)	Dol, En, Mag, Tr, Tc	(Dol, Mag)	Anth, En, Fo, Tr, Tc
(Anth, Mag)	Dol, En, Fo, Tr, Tc	(Dol, Tr)	Anth, En, Fo, Mag, Tc
(Anth, Tr)	Dol, En, Fo, Mag, Tc	(Dol, Tc)	Anth, En, Fo, Mag, Tr
(Anth, Tc)	Dol, En, Fo, Mag, Tr		
(En, Fo)	Anth, Dol, Mag, Tr, Tc	(Fo, Mag)	Anth, Dol, En, Tr, Tc
(En, Mag)	Anth, Dol, Fo, Tr, Tc	(Fo, Tr)	Anth, Dol, En, Mag, Tc
(En, Tr)	Anth, Dol, Fo, Mag, Tc	(Fo, Tc)	Anth, Dol, En, Mag, Tr
(En, Tc)	Anth, Dol, Fo, Mag, Tr		
(Mag, Tr)	Anth, Dol, En, Fo, Tc	(Tr, Tc)	Anth, Dol, En, Fo, Mag
(Mag, Tc)	Anth, Dol, En, Fo, Tr		

Fluid-bearing Invariant Points in the sub-system MgO-SiO₂

(Anth)	En, Fo, Mag, Tc	MEFT [*]
(En)	Anth, Fo, Mag, Tc	MAFT
(Fo)	Anth, En, Mag, Tc	
(Mag)	Anth, En, Fo, Tc	
(Tc)	Anth, En, Fo, Mag	MAFE

(XXX,XXX) Absent Phase

^{*} Name applied by Evans and Trommsdorff (1974)

The ternary system CaO-MgO-SiO_2 contains the binary subsystem MgO-SiO_2 , studied by Evans and Trommsdorff (1974). Since there is one less component to consider, univariant reactions ($f=1$) will have three solid phases in equilibrium while four phases in equilibrium constitute an invariant point.

The elimination of CaO as a component removes dolomite and tremolite from consideration, decreasing the total number of phases to five.

In this subsystem, the total number of possible invariant points is:

$$\frac{5!}{4!(5-4)!} = 5 \text{ possible invariant points.}$$

The number of possible univariant reactions is:

$$\frac{5!}{3!(5-3)!} = 10 \text{ possible univariant reactions.}$$

5.6 Elimination of Possible Invariant Points

The number of invariant points predicted for the system by combinatorial analysis is a maximum amount for a system with no compositional degeneracy. The system chosen for study is degenerate since it has several colinear phases on the MgO-SiO₂ join (see Fig 3.2). Therefore the number of invariant points in the system will be less than the number determined by combinatorial analysis. Non-degenerate invariant points are selected from the total number through a process of elimination.

Invariant points which have one or both calcium bearing phases absent may be eliminated from further consideration since both tremolite and dolomite are chemographically external and must therefore remain stable for all pressures and temperatures. Furthermore, the presence of only one such phase precludes the possibility of reaction between calcium-bearing minerals. The elimination of these invariant points removes eleven of the 21 possible invariant points. The remaining ten invariant points are listed in Table 5.3.

Table 5.3

Fluid-bearing Invariant Points in the sub-system CaO-MgO-SiO₂

(Anth, En)	Dol, Fo, Mag, Tr, Tc	(Anth, Fo)	Dol, En, Mag, Tr, Tc
(Anth, Mag)	Dol, En, Fo, Tr, Tc	(Anth, Tc)	Dol, En, Fo, Mag, Tr
(En, Fo)	Anth, Dol, Mag, Tr, Tc	(Fo, Mag)	Anth, Dol, En, Tr, Tc
(En, Mag)	Anth, Dol, Fo, Tr, Tc	(Fo, Tc)	Anth, Dol, En, Mag, Tr
(En, Tc)	Anth, Dol, Fo, Mag, Tr	(Mag, Tc)	Anth, Dol, En, Fo, Tr

5.7 Log $f_{(H_2O)}$ - Log $f_{(CO_2)}$ Diagrams

It has been demonstrated, see Chapter 3, that the minerals of the Flinton Creek ultramafic rocks do not appear to change with increasing metamorphic grade. Rather, the stable mineral assemblages appear to be related to the proximity of the ultramafic - host rock contact and the presence or absence of veins. This suggests that changes in the chemical potential of one or more components may play a greater role in mineral stability at the peak of metamorphism than changing pressure and temperature. The volatile components, CO_2 and H_2O are considered to be the most likely components whose variation is responsible for the changing mineralogy of the ultramafic rocks. It should be noted that variation in the chemical potential of other less mobile components (ie. SiO_2 , MgO or CaO) might also be responsible for the mineral zonation.

Log fugacity H_2O versus log fugacity CO_2 diagrams were constructed to investigate the effect of changing H_2O and CO_2 chemical potentials (μ) at constant pressure and temperature. Chemical potential is related to fugacity by the equation:

$$\mu_i^{P,T} = (\mu_i^O)^{P,T} + RT \log(f_i) \quad (5.1)$$

Since μ_i^O is constant, at constant P and T , μ - μ diagrams and $\log f$ - $\log f$ diagrams differ only by a scale factor.

5.8 Determination of Univariant Reactions

Once degenerate invariant points have been eliminated, the configuration of the remaining invariant points may be constructed following the method of Schreinemakers (1915-1925).

In this study there are three inert components ($C=3$); therefore each invariant point should be surrounded by five reactions. Compositional degeneracy however, could result in a smaller number of radiating reaction lines.

All possible reactions surrounding each stable invariant point were determined graphically by chemography assuming the system was open to H_2O and CO_2 . In total,

twenty-two different reactions, listed in Table 5.4 were determined. Reactions were balanced either by trial and error or following the method of determinants (Korzhinskii, 1959, pg. 103).

5.9 Construction of Invariant Points in $\log f_{(H_2O)} - \log f_{(CO_2)}$ Space

The properties of $\log f_{(H_2O)} - \log f_{(CO_2)}$ diagrams have been summarized by Korzhinskii (1959). Following the methods of Korzhinskii (1959) and Zen (1966), eight Schreinemakers bundles were constructed.

As the number of phases in the system increases beyond $c+2$, the number of stability types for each invariant point increases (Skippen and Trommsdorff, 1975); therefore, each of the invariant points constructed above has more than one stability level. In a general $(c+m)$ phases system, the number of stability levels equals $m-1$. The system in question has seven solid phases plus a mixed fluid phase, therefore $p=8$. The number of components required to describe this system is five ($c=5$). Thus this study is concerned with a $c+3$ phase system and each invariant point has two stability levels; stable and metastable.

To construct $\log f_{(H_2O)} - \log f_{(CO_2)}$ diagrams, it is necessary to link the various invariant points. This is done by connecting the univariant reactions with the same stability level which surround the invariant points to form a petrogenetic grid. Therefore, before the grid can be assembled, it is necessary to determine the stability level of the various invariant points since this in turn determines the stability level of the surrounding univariant reactions.

Table 5.4 – Univariant Equilibria in a Portion of the System CaO–MgO–SiO₂–H₂O–CO₂

1) F + A = 9E + H ₂ O	Mg ₂ SiO ₄ + Mg ₇ Si ₈ O ₂₂ (OH) ₂ = 9MgSiO ₃ + H ₂ O
2) 2D + 13E + H ₂ O = Tr + 5F + 4CO ₂	2CaMg(CO ₃) ₂ + 13MgSiO ₃ + H ₂ O = Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 5Mg ₂ SiO ₄ + 4CO ₂
3) 2D + 5A = Tr + 32E + 4H ₂ O + 4CO ₂	2CaMg(CO ₃) ₂ + 5Mg ₇ Si ₈ O ₂₂ (OH) ₂ = Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 32 MgSiO ₃ + 4H ₂ O + 4CO ₂
4) 18D + 13A = 9Tr + 32F + 4H ₂ O + 36CO ₂	18CaMg(CO ₃) ₂ + 13Mg ₇ Si ₈ O ₂₂ (OH) ₂ = 9Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 32Mg ₂ SiO ₄ + 4H ₂ O + 36CO ₂
5) Tr + 13M = 2D + 8F + 9CO ₂ + H ₂ O	Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 13MgCO ₃ = 2CaMg(CO ₃) ₂ + 8Mg ₂ SiO ₄ + H ₂ O + 9CO ₂
6) E + M = F + CO ₂	MgSiO ₃ + MgCO ₃ = Mg ₂ SiO ₄ + CO ₂
7) Tr + 5M = 8E + 2D + H ₂ O + CO ₂	Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 5MgCO ₃ = 8MgSiO ₃ + 2CaMg(CO ₃) ₂ + H ₂ O + CO ₂
8) 2D + 5T = Tr + 12E + 4CO ₂ + 4H ₂ O	2CaMg(CO ₃) ₂ + 5Mg ₃ Si ₄ O ₁₀ (OH) ₂ = Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 12MgSiO ₃ + 4H ₂ O + 4CO ₂
9) F + T = 5E + H ₂ O	Mg ₂ SiO ₄ + Mg ₃ Si ₄ O ₁₀ (OH) ₂ = 5MgSiO ₃ + H ₂ O
10) 10D + 13T = 5Tr + 12F + 20CO ₂ + 8H ₂ O	10CaMg(CO ₃) ₂ + 13Mg ₃ Si ₄ O ₁₀ (OH) ₂ = 5Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 12Mg ₂ SiO ₄ + 8H ₂ O + 20CO ₂
11) 8T + 2D = Tr + 3A + 4CO ₂ + 4H ₂ O	8Mg ₃ Si ₄ O ₁₀ (OH) ₂ + 2CaMg(CO ₃) ₂ = Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 3Mg ₇ Si ₈ O ₂₂ (OH) ₂ + 4H ₂ O + 4CO ₂
12) A = T + 4E	Mg ₇ Si ₈ O ₂₂ (OH) ₂ = Mg ₃ Si ₄ O ₁₀ (OH) ₂ + 4MgSiO ₃
13) M + A = 8E + H ₂ O + CO ₂	MgCO ₃ + Mg ₇ Si ₈ O ₂₂ (OH) ₂ = 8MgSiO ₃ + H ₂ O + CO ₂
14) Tr + 4M = 2D + A	Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 4MgCO ₃ = 2CaMg(CO ₃) ₂ + Mg ₇ Si ₈ O ₂₂ (OH) ₂
15) M + T = 4E + CO ₂ + H ₂ O	MgCO ₃ + Mg ₃ Si ₄ O ₁₀ (OH) ₂ = 4MgSiO ₃ + H ₂ O + CO ₂
16) 2D + 2T = Tr + 3M + CO ₂ + H ₂ O	2CaMg(CO ₃) ₂ + 2Mg ₃ Si ₄ O ₁₀ (OH) ₂ = Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 3MgCO ₃ + H ₂ O + CO ₂
17) M + 2T = A + H ₂ O + CO ₂	MgCO ₃ + 2Mg ₃ Si ₄ O ₁₀ (OH) ₂ = Mg ₇ Si ₈ O ₂₂ (OH) ₂ + H ₂ O + CO ₂
18) 4F + 9T = 5A + 4H ₂ O	4Mg ₂ SiO ₄ + 9Mg ₃ Si ₄ O ₁₀ (OH) ₂ = 5Mg ₇ Si ₈ O ₂₂ (OH) ₂ + 4H ₂ O
19) 9M + A = 8F + H ₂ O + 9CO ₂	9MgCO ₃ + Mg ₇ Si ₈ O ₂₂ (OH) ₂ = 8Mg ₂ SiO ₄ + H ₂ O + 9CO ₂
20) 5M + T = 4F + H ₂ O + 5CO ₂	5MgCO ₃ + Mg ₃ Si ₄ O ₁₀ (OH) ₂ = 4Mg ₂ SiO ₄ + H ₂ O + 5CO ₂
21) 2T + 2D = 3M + Tr + CO ₂ + H ₂ O	2Mg ₃ Si ₄ O ₁₀ (OH) ₂ + 2CaMg(CO ₃) ₂ = 3MgCO ₃ + Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + H ₂ O + CO ₂
22) Tr + 13M = 8F + 2D + 9 CO ₂	Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂ + 13MgCO ₃ = 8Mg ₂ SiO ₄ + 2CaMg(CO ₃) ₂ + H ₂ O + 9CO ₂

The stability level of an invariant point in $\log f_{(\text{H}_2\text{O})} - \log f_{(\text{CO}_2)}$ space is determined by the stability of the phases at that invariant point. If thermodynamic data predicts that all minerals present at an invariant point are in stable equilibrium, then the invariant point is stable. However, if changing pressures and temperatures during metamorphism make that mineral assemblage unstable, then the invariant point becomes metastable. Therefore, before a petrogenetic grid can be constructed in $\log f_{(\text{H}_2\text{O})} - \log f_{(\text{CO}_2)}$ space, the pressure and temperature stability range of various mineral assemblages must be determined. This is done by first predicting stable mineral assemblages using the results of experimental work, summarized in thermodynamic databases, and then comparing these results with those actually observed in the field through petrography.

5.10 Thermodynamic Data

Two databases contain the thermochemical data for all the required phases. They are Helgeson, Delany, Nesbitt and Bird (1978) (Table 5.5), and Berman (1988) combined with Berman and Chernosky (1989) (Table 5.6).

Table 5.5 Thermodynamic Database
Helgeson, Delany, Nesbitt, and Bird (1978)

<i>Standard state thermodynamic properties at 1 bar, 298.15K</i>						<i>Coefficients for calculation of heat capacity</i>			
	G (cal/mole)	H (cal/mole)	S (cal/mole/K)	V (cm ³ /mole)		a (cal/mole)	b (x10 ⁻³) (cal/mole/K/K)	c (x10 ⁻⁵) (cal*K/mole)	
Anthophyllite	-2715430	-2888749	128.60	264.40		180.682	60.574	38.462	
Dolomite	-517980	-556851	37.09	64.365		41.557	23.952	9.884	
Enstatite	-348930	-369686	16.20	31.276		24.55	4.74	62.80	
Forsterite	-491564	-520000	22.75	43.79		35.81	6.54	8.52	
Magnesite	-245658	-265630	15.70	28.018		19.731	12.539	4.748	
Talc	-1320188	-1410920	62.34	136.25		82.48	41.61	13.34	
Tremolite	-2770685	-2944478	131.19	272.92		188.222	57.294	44.822	
Carbon Dioxide	-94262	-94054	51.072	0.0		10.57	2.1	2.06	
Water (gas)	-54634	-57796	45.104	0.0		7.30	2.46	0.0	

<i>Standard state thermodynamic properties at 1 bar, 298.15K</i>						<i>Coefficients for calculation of heat capacity</i>			
	G (kJ/mole)	H (kJ/mole)	S (J/mole/K)	V (J/bar)		k ₀ (J/mole)	k ₅ (x10 ⁻³) (J/mole/K/K)	-k ₂ (x10 ⁻⁵) (J*K/mole)	
Anthophyllite	-11361.359	-12086.526	538.06	26.440		755.973	253.442	160.925	
Dolomite	-2167.228	-2329.865	155.18	6.437		173.874	100.215	41.355	
Enstatite	-1459.923	-1546.766	67.78	3.128		102.717	19.832	262.755	
Forsterite	-2056.704	-2175.680	95.19	4.379		149.829	27.363	35.648	
Magnesite	-1027.833	-1111.396	65.69	2.802		82.555	52.463	19.866	
Talc	-5523.667	-5903.289	260.83	13.625		345.096	174.096	55.815	
Tremolite	-11592.546	-12319.696	548.90	27.292		787.521	239.718	187.535	
Carbon Dioxide	-394.392	-393.522	213.69	0.0		44.225	8.786	8.619	
Water (gas)	-228.589	-241.818	188.72	0.0		30.543	10.293	0.0	

Table 5.6 Thermodynamic Database
Berman (1988), Chernosky and Berman (1989)

<i>Standard state thermodynamic properties at 1b, 298.15K</i>					
	G (kJ/mol)	H (kJ/mol)	S (J/mol/K)	V (J/bar)	
Anthophyllite	-11342.582	-12069.032	535.196	26.560	
Dolomite	-2162.354	-2325.248	154.890	6.432	
Enstatite	-1458.181	-1545.552	66.170	3.133	
Forsterite	-2055.023	-2174.420	94.010	4.366	
Magnesite	-1030.709	-1114.505	65.090	2.803	
Talc	-5517.657	-5897.387	261.241	13.610	
Tremolite	-11578.548	-12305.578	551.150	27.268	
Carbon Dioxide	-394.341	-393.510	213.677	0.0	
Water (gas)	-228.538	-241.816	188.720	0.0	

<i>Coefficients for calculation of heat capacity (J/mol/K)</i>						
	ko	k1 (x10 ⁻²)	k2 (x10 ⁻⁵)	k3 (x10 ⁻⁷)	k4 (x10 ⁻³)	k5 (x10 ⁻³)
Anthophyllite	1219.31	-57.665	-347.661	440.090	0.0	0.0
Dolomite	328.48	-25.544	-46.885	79.038	0.0	0.0
Enstatite	166.58	-12.006	-22.706	27.915	0.0	0.0
Forsterite	238.64	-20.013	0.000	-11.624	0.0	0.0
Magnesite	162.30	-11.093	-48.826	87.466	0.0	0.0
Talc	664.11	-51.872	-21.472	-32.737	0.0	0.0
Tremolite	1229.35	-64.019	-320.899	420.881	0.0	0.0
Carbon Dioxide	93.00	-13.409	1.238	0.0	6.336	-2.876
Water (gas)	115.45	-37.999	-28.713	0.0	51.055	-2.062

5.11 Solid Phase Reactions in P-T Space

There are two solid phase reactions in the system:

To plot these univariant lines in P-T space, the free energy of reaction must be expressed as a function of both pressure and temperature and then the free energy of reaction must be set equal to zero. Skippen and Carmichael (1977) derive such an equation to express the pressure and temperature dependence of free energy.

The number of terms in this equation depends on the power series used to describe the change in heat capacities with temperature. Berman *et al.* (1988) uses a seven term polynomial:

$$C_p = k_0 + k_1T^{-1} + k_2T^{-2} + k_3T^{-3} + k_4/T + k_5T + k_6T^2 \quad (5.2)$$

While Helgeson *et al.* (1978) uses the Maier-Kelly equation:

$$C_p = a + bT - cT^{-2} \quad (5.3a)$$

An equivalent form of the Maier-Kelly equation is obtained by setting k_1 , k_3 , k_4 , and k_6 of the Berman *et al.* (1988) equation to zero therefore:

$$C_p = k_0 + k_5T - k_2T^{-2} \quad (5.3b)$$

Using equation 5.2 for C_p (Berman *et al.*, 1988), the Skippen and Carmichael (1977) equation is expressed as follows:

$$\begin{aligned}
 0 = \Delta G_{(T,P)} = \Delta H_{(T_R,1)} - T\Delta S_{(T_R)} + \Delta V_s(P - 1) + RT \text{ Log}K & \quad (5.4) \\
 + \Delta k_0\{T - T_R + T\text{Log}(T_R/T)\} & \\
 + \Delta k_1\{4\sqrt{T} - 2\sqrt{T_R} - 2T/\sqrt{T_R}\} & \\
 + \Delta k_2\{1/T_R^2 - T/(2T_R) - 1/(2T)\} & \\
 + \Delta k_3\{1/2T_R^2 - T/3T_R^3 - 1/6T^2\} & \\
 + \Delta k_4\{\text{Log}(T) - \text{Log}(T_R) - T/T_R + 1\} & \\
 + \Delta k_5\{T \times T_R - T_R^2/2 - T^2/2\} & \\
 + \Delta k_6\{T_R^2 \times T/2 - T_R^3/3 - T^3/6\} &
 \end{aligned}$$

Note that for the first attempt, it is assumed that the solid phases have ideal, end-member compositions. Therefore the equilibrium constant for each reaction equals unity and the logK term drops out of the calculation.

Equation 5.4 was solved for temperature at various pressures by iteration on a Lotus 123 spreadsheet. The resulting equilibrium lines are plotted in Fig. 5.4 for both the Berman (1988) and Helgeson *et al.* (1978) databases.

Once the location of each solid-solid reaction is known in pressure - temperature space, the stability level of the invariant points in the system can be determined. Note that considerable difference exists between in two databases in question concerning the location of the second solid phase reaction:

This difference means the two databases will have different grid topologies on $\log f_{(\text{H}_2\text{O})} - \log f_{(\text{CO}_2)}$ diagrams.

Figure 5.4 Solid phase reactions in a portion of the system $\text{CaO-MgO-SiO}_2\text{-H}_2\text{O-CO}_2$, assuming ideal mineral compositions. Equilibrium lines were plotted using equation 5.4 and the thermodynamic data of Helgeson *et al.* (1978) and Berman (1988) plus Chernosky and Berman (1989). Solid phases were assumed to have ideal compositions.

5.12 Ideal Activity Models

Results presented in the previous section assumed that the solid phases have ideal compositions, listed in Table 5.1. This is an oversimplification since the minerals selected for study are crystalline solutions in which certain cations may substitute for each other in the crystal structure. As indicated by electron microprobe analyses, given in Chapter 4, the most common substitution amongst the minerals in question is the replacement of magnesium in the crystal structure by iron, manganese and nickel. If we assume that this replacement occurs randomly throughout the crystal structure and produces no heat of mixing, then the resulting solution is said to be ideal. The activity of an ion, mixing with others at a specific site, in an ideal solution is expressed by:

$$a_{i, \text{ideal}} = X_i^n \quad (5.5)$$

Where the exponent n refers to the number of identical specific sites in the formula unit.

When mixing occurs in two or more different sites, the total activity of the mineral is the product of the individual ion activities for mixing at each site.

If the solid phases in question are treated as ideal solutions, the activities of the various magnesian-end member components are calculated from the following formulas, see Table 5.7:

Table 5.7

Activity of Species in Solid Phases

$$\begin{aligned} \text{Mg-amphibole} \\ a_{\text{Mg}_7\text{Si}_8\text{O}_{22}(\text{OH})_2} &= [\text{Mg}/7]^7 * [\text{Si}/8]^8 = [6.094/7]^7 * [7.958/8]^8 \\ &= 0.36 \end{aligned}$$

$$\begin{aligned} \text{dolomite} \\ a_{\text{CaMg}(\text{CO}_3)_2} &= [\text{Ca}] * [\text{Mg}] = [0.98] * [0.98] = 0.96 \end{aligned}$$

$$\begin{aligned} \text{orthopyroxene} \\ a_{\text{MgSiO}_3} &= [\text{Mg}] = 0.89 \end{aligned}$$

$$\begin{aligned} \text{olivine} \\ a_{\text{Mg}_2\text{SiO}_4} &= [\text{Mg}/2]^2 = [1.805/2]^2 = 0.82 \end{aligned}$$

$$\begin{aligned} \text{magnesite} \\ a_{\text{MgCO}_3} &= [\text{Mg}] = 0.94 \end{aligned}$$

$$\begin{aligned} \text{talc} \\ a_{\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_2} &= [\text{Mg}/3]^3 * [\text{Si}/4]^4 = [2.913/3]^3 * [3.980/4]^4 \\ &= 0.90 \end{aligned}$$

$$\begin{aligned} \text{Ca-amphibole} \\ a_{\text{Ca}_2\text{Mg}_5\text{Si}_8\text{O}_{22}(\text{OH})_2} &= [3\text{-Ca-Na-k}] * [\text{Ca}/2]^2 * [\text{Mg}/5]^5 \\ &= 1.043 * [1.943/2]^2 * [4.810/5]^5 \\ &= 0.81 \end{aligned}$$

5.13 Solid - Solid Reactions and Ideal Activity Modelling

The activities derived in the previous section may be used to calculate the value of the equilibrium constant for both solid phase reactions in the system, reactions 12 and 14:

$$\begin{aligned}
 12) \text{ Tc} + 4\text{En} &= \text{Anth} & K &= a_{(\text{Anth})} / \{a_{(\text{Tc})} \times a_{(\text{En})}^4\} \\
 & & &= 0.36 / (0.90 \times 0.89) \\
 & & &= 0.64
 \end{aligned}$$

$$\begin{aligned}
 14) \text{ Tr} + 4\text{Mag} &= 2\text{Dol} + \text{Anth} & K &= \{a_{(\text{Dol})}^2 \times a_{(\text{Anth})}\} / \{a_{(\text{Tr})} \times a_{(\text{Mag})}^4\} \\
 & & &= (0.92 \times 0.36) / (0.81 \times 0.78) \\
 & & &= 0.53
 \end{aligned}$$

The equilibrium constant values calculated above were then substituted into equation 5.4 which was solved for P and T by iteration. The results of this calculation are shown in Fig. 5.5.

Figure 5.5 Solid phase reactions in a portion of the system $\text{CaO-MgO-SiO}_2\text{-H}_2\text{O-CO}_2$ assuming non-component mineral compositions. Microprobe analyses of mineral phases present in the Flinton Creek Ultramafic Rocks were used to determine the activity of species present in the solid phases (see Table 5.7).

5.14 Construction of $\text{Log}f_{(\text{H}_2\text{O})} - \text{Log}f_{(\text{CO}_2)}$ Diagrams

Knowing the various invariant point configurations and levels of stability, it is possible to assemble these points together to form qualitative petrogenetic grids. However, using the two thermochemical databases, it is possible to assign values of water and carbon dioxide fugacity, at fixed pressure and temperature, to the axis of these diagrams.

This is done by rearranging equation 5.4 and solving for $\log K$, the equilibrium constant:

$$\begin{aligned} \log K = & \Delta H_{(T_R, 1)}/RT - \Delta S_{(T_R)}/R + (\Delta V_s/RT)(P - 1) \quad (5.6) \\ & + (\Delta k_0/RT) \{ T - T_R + T \log(T_R/T) \} \\ & + (\Delta k_1/RT) \{ 4\sqrt{T} - 2\sqrt{T_R} - 2T/\sqrt{T_R} \} \\ & + (\Delta k_2/RT) \{ 1/T_R - T^2/(2T_R) - 1/(2T) \} \\ & + (\Delta k_3/RT) \{ 1/2T_R^2 - T/3T_R^3 - 1/6T^2 \} \\ & + (\Delta k_4/RT) \{ \log(T) - \log(T_R) - T/T_R + 1 \} \\ & + (\Delta k_5/RT) \{ T \times T_R - T_R^2/2 - T^2/2 \} \\ & + (\Delta k_6/RT) \{ T_R^2 \times T/2 - T_R^3/3 - T^3/6 \} \end{aligned}$$

Using this equation with the thermochemical data contained in the two databases, it is possible to calculate $\log K$ for any reaction at fixed temperature and pressure. If solid phases involved in the reactions are assumed to have ideal end-member compositions, then their activity is equal to one, and the equilibrium constant only involves $f_{(\text{H}_2\text{O})}$ and/or $f_{(\text{CO}_2)}$ terms. Ideal equilibrium constant expressions for the 21 univariant reactions in the system are listed in Table 5.8. Equating these expressions with the value of $\log K$ derived from equation 5.6 above allows the values of $f_{(\text{H}_2\text{O})}$ and/or $f_{(\text{CO}_2)}$ for the reaction to be calculated at fixed temperature and pressure.

Table 5.8

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Equilibrium Constants for Univariant Reactions Assuming Ideal Mineral Compositions

Reaction	K	Log K
1) $F + A = 9E + H_2O$	$f_{(H_2O)}$	$\log f_{(H_2O)}$
2) $2D + 13E + H_2O = Tr + 5F + 4CO_2$	$f_{(CO_2)}^4 / f_{(H_2O)}$	$4\log f_{(CO_2)} - \log f_{(H_2O)}$
3) $2D + 5A = Tr + 32E + 4H_2O + 4CO_2$	$f_{(H_2O)}^4 \times f_{(CO_2)}^4$	$4\log f_{(H_2O)} + 4\log f_{(CO_2)}$
4) $18D + 13A = 9Tr + 32F + 4H_2O + 36CO_2$	$f_{(H_2O)}^4 \times f_{(CO_2)}^{36}$	$4\log f_{(H_2O)} + 36\log f_{(CO_2)}$
5) $Tr + 13M = 2D + 8F + 9CO_2 + H_2O$	$f_{(H_2O)} \times f_{(CO_2)}^9$	$\log f_{(H_2O)} + 9\log f_{(CO_2)}$
6) $E + M = F + CO_2$	$f_{(CO_2)}$	$\log f_{(CO_2)}$
7) $Tr + 5M = 8E + 2D + H_2O + CO_2$	$f_{(H_2O)} \times f_{(CO_2)}$	$\log f_{(H_2O)} + \log f_{(CO_2)}$
8) $2D + 5T = Tr + 12E + 4CO_2 + 4H_2O$	$f_{(H_2O)}^4 \times f_{(CO_2)}^4$	$4\log f_{(H_2O)} + 4\log f_{(CO_2)}$
9) $F + T = 5E + H_2O$	$f_{(H_2O)}$	$\log f_{(H_2O)}$
10) $10D + 13T = 5Tr + 12F + 20CO_2 + 8H_2O$	$f_{(H_2O)}^8 \times f_{(CO_2)}^{20}$	$8\log f_{(H_2O)} + 20\log f_{(CO_2)}$
11) $8T + 2D = Tr + 3A + 4CO_2 + 4H_2O$	$f_{(H_2O)}^4 \times f_{(CO_2)}^4$	$4\log f_{(H_2O)} + 4\log f_{(CO_2)}$
12) $A = T + 4E$	0	1
13) $M + A = 8E + H_2O + CO_2$	$f_{(H_2O)} \times f_{(CO_2)}$	$\log f_{(H_2O)} + \log f_{(CO_2)}$
14) $Tr + 4M = 2D + A$	0	1
15) $M + T = 4E + CO_2 + H_2O$	$f_{(H_2O)} \times f_{(CO_2)}$	$\log f_{(H_2O)} + \log f_{(CO_2)}$
16) $2D + 2T = Tr + 3M + CO_2 + H_2O$	$f_{(H_2O)} \times f_{(CO_2)}$	$\log f_{(H_2O)} + \log f_{(CO_2)}$
17) $M + 2T = A + H_2O + CO_2$	$f_{(H_2O)} \times f_{(CO_2)}$	$\log f_{(H_2O)} + \log f_{(CO_2)}$
18) $4F + 9T = 5A + 4H_2O$	$f_{(H_2O)}^4$	$4\log f_{(H_2O)}$
19) $9M + A = 8F + H_2O + 9CO_2$	$f_{(H_2O)} \times f_{(CO_2)}^9$	$\log f_{(H_2O)} + 9\log f_{(CO_2)}$
20) $5M + T = 4F + H_2O + 5CO_2$	$f_{(H_2O)} \times f_{(CO_2)}^5$	$\log f_{(H_2O)} + 5\log f_{(CO_2)}$
21) $2T + 2D = 3M + Tr + CO_2 + H_2O$	$f_{(H_2O)} \times f_{(CO_2)}$	$\log f_{(H_2O)} + \log f_{(CO_2)}$
22) $Tr + 13M = 8F + 2D + 9CO_2$	$f_{(CO_2)}^9$	$9\log f_{(CO_2)}$

Two $\log f_{(\text{H}_2\text{O})} - \log f_{(\text{CO}_2)}$ diagrams were constructed for each database assuming ideal solid phase compositions. The pressure - temperature values (600C, 5600 bars) and (750C, 6800 bars), selected for each of these diagrams falls along the proposed geothermal gradient for the eastern Elzevir Terrain, on either side of the solid phase reaction $\text{talc} + 4 \text{ enstatite} = \text{anthophyllite}$. This produces two different diagram topologies, one with anthophyllite stable (Fig 5.7 and 5.9), the other with talc and enstatite (Fig. 5.6 and 5.8). Note that pressure - temperature conditions of the 500C, 5600bar diagram are similar to those of staurolite breakdown in the kyanite stability field, conditions which are believed to exist in the Kaladar metasediments, near the maximum P-T conditions of the Flinton Creek Ultramafics.

Figure 5.6

$\text{Logf}(\text{H}_2\text{O}) - \text{Logf}(\text{CO}_2)$ Diagram

$T = 600^\circ\text{C}$, $P = 5600\text{bars}$

Thermodynamic Data - Chernosky and Berman (1989)

- Berman (1988)

Assuming Ideal Mineral Compositions

Figure S.7

 $\text{Logf}(\text{H}_2\text{O}) - \text{Logf}(\text{CO}_2)$ Diagram

T = 750°C, P = 6800 bars

Thermodynamic Data - Chernosky and Berman (1989)
 - Berman (1988)
 Assuming Ideal Mineral Compositions

Figure 5.8
Diagram
 $\text{Logf}(\text{H}_2\text{O}) - \text{Logf}(\text{CO}_2)$

$T = 600^\circ\text{C}$, $P = 5600\text{bars}$
 Thermodynamic Data - Helgeson *et al.* (1978)
 Assuming Ideal Mineral Compositions

Figure 5.9
 $\text{Logf}(\text{H}_2\text{O}) - \text{Logf}(\text{CO}_2)$ Diagram

$T = 750^\circ\text{C}$, $P = 6800\text{bars}$
 Thermodynamic Data - Helgeson *et al.*, (1978)
 Assuming Ideal Mineral Compositions

If the solid phases are assumed to have non-component composition, then the equilibrium constant must be changed to incorporate the previously calculated ideal activity of these phases. These new equilibrium constant expressions are listed in Table 5.9. The procedure used to calculate real-composition $\log f_{(\text{H}_2\text{O})} - \log f_{(\text{CO}_2)}$ diagrams is the same as that followed above. LogK for the reaction is calculated from the thermochemical data and is set equal to the equilibrium constant expression listed in Table 5.9. This equation is then solved for $\log f_{(\text{H}_2\text{O})}$ and/or $\log f_{(\text{CO}_2)}$.

Three $\log f_{(\text{H}_2\text{O})} - \log f_{(\text{CO}_2)}$ diagrams were calculated for each thermodynamic database at P-T conditions along the proposed geothermal gradient for the eastern Elzevir Terrain. These conditions are as follows:

<u>Diagram</u>	<u>Temperature (C)</u>	<u>Pressure (bars)</u>	<u>Similar Local Conditions</u>
5.10, 5.13	530	4700	southern-most pelite, Flinton synclinorium.
5.11, 5.14	560	5100	
5.12, 5.15	600	5600	Kaladar metasediments.

Bold - Chernosky and Berman (1989); Berman (1988)

Italics - Helgeson, Delaney, Nesbitt and Bird (1978)

Diagrams 5.10 and 5.13 represent the lowest possible metamorphic pressure-temperature conditions experienced by the Flinton Creek ultramafic rocks, while diagrams 5.12 and 5.15 represent the highest. Diagrams 5.11 and 5.14 were calculated to show phase relations at intermediate pressure-temperature conditions between 5.10, 5.13 and 5.12, 5.15. These conditions are assumed to be nearest the actual metamorphic conditions reached by the Flinton Creek ultramafic rocks.

Table 5.9 - Equilibrium Constants for Reactions Assuming Non-Component Mineral Compositions

Reaction	Log K
1) $F + A = 9E + H_2O$	$9\log(a_{MgSiO_3}) + \log(f_{H_2O}) - \log(a_{Mg_3Si_2O_7}) - \log(a_{Mg_3Si_2O_7(OH)_2})$
2) $2D + 13E + H_2O = Tr + 5F + 4CO_2$	$\log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) + 5\log(a_{Mg_2SiO_4}) + 4\log(f_{CO_2}) - 2\log(a_{CaMg(CO_3)_2}) - 13\log(a_{MgSiO_3}) - \log(f_{H_2O})$
3) $2D + 5A = Tr + 32E + 4H_2O + 4CO_2$	$\log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) + 32\log(a_{MgSiO_3}) + 4\log(f_{H_2O}) + 4\log(f_{CO_2}) - 2\log(a_{CaMg(CO_3)_2}) - 5\log(a_{Mg_3Si_2O_7(OH)_2})$
4) $18D + 13A = 9Tr + 32F + 4H_2O + 36CO_2$	$9\log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) + 32\log(a_{Mg_2SiO_4}) + 4\log(f_{H_2O}) + 36\log(f_{CO_2}) - 13\log(a_{Mg_3Si_2O_7(OH)_2}) - 18\log(a_{CaMg(CO_3)_2})$
5) $Tr + 13M = 2D + 8F + 9CO_2 + H_2O$	$2\log(a_{CaMgCO_3}) + 8\log(a_{Mg_2SiO_4}) + \log(f_{H_2O}) + 9\log(f_{CO_2}) - \log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) - 13\log(a_{MgCO_3})$
6) $E + M = F + CO_2$	$\log(a_{Mg_2SiO_4}) + \log(f_{CO_2}) - \log(a_{MgSiO_3}) - \log(a_{MgCO_3})$
7) $Tr + 5M = 8E + 2D + H_2O + CO_2$	$8\log(a_{MgSiO_3}) + \log(a_{CaMgCO_3}) + \log(f_{H_2O}) + \log(f_{CO_2}) - \log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) - 5\log(a_{MgCO_3})$
8) $2D + 5T = Tr + 12E + 4CO_2 + 4H_2O$	$\log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) + 12\log(a_{MgSiO_3}) + 4\log(f_{H_2O}) + 4\log(f_{CO_2}) - 2\log(a_{CaMgCO_3}) - 5\log(a_{Mg_3Si_2O_7(OH)_2})$
9) $F + T = 5E + H_2O$	$5\log(a_{MgSiO_3}) + \log(f_{H_2O}) - \log(a_{Mg_2SiO_4}) - \log(a_{Mg_3Si_2O_7(OH)_2})$
10) $10D + 13T = 5Tr + 12F + 20CO_2 + 8H_2O$	$5\log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) + 12\log(a_{Mg_2SiO_4}) + 8\log(f_{H_2O}) + 20\log(f_{CO_2}) - 10\log(a_{CaMgCO_3}) - 13\log(a_{Mg_3Si_2O_7(OH)_2})$
11) $8T + 2D = Tr + 3A + 4CO_2 + 4H_2O$	$\log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) + 3\log(a_{Mg_3Si_2O_7(OH)_2}) + 4\log(f_{H_2O}) + 4\log(f_{CO_2}) - 8\log(a_{Mg_3Si_2O_7(OH)_2}) - 2\log(a_{CaMgCO_3})$
12) $A = T + 4E$	$4\log(a_{MgSiO_3}) + \log(a_{Mg_3Si_2O_7(OH)_2}) - \log(a_{Mg_2Si_6O_{22}(OH)_2})$
13) $M + A = 8E + H_2O + CO_2$	$8\log(a_{MgSiO_3}) + \log(f_{H_2O}) + \log(f_{CO_2}) - \log(a_{MgCO_3}) - \log(a_{Mg_3Si_6O_{22}(OH)_2})$
14) $Tr + 4M = 2D + A$	$2\log(a_{CaMgCO_3}) + \log(a_{Mg_3Si_6O_{22}(OH)_2}) - \log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) - 4\log(a_{MgCO_3})$
15) $M + T = 4E + CO_2 + H_2O$	$4\log(a_{MgSiO_3}) + \log(f_{H_2O}) + \log(f_{CO_2}) - \log(a_{MgCO_3}) - \log(a_{Mg_3Si_6O_{22}(OH)_2})$
16) $2D + 2T = Tr + 3M + CO_2 + H_2O$	$\log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) + 3\log(a_{MgCO_3}) + \log(f_{H_2O}) + \log(f_{CO_2}) - 2\log(a_{CaMgCO_3}) - 2\log(a_{Mg_3Si_6O_{22}(OH)_2})$
17) $M + 2T = A + H_2O + CO_2$	$\log(a_{Mg_3Si_6O_{22}(OH)_2}) + \log(f_{H_2O}) + \log(f_{CO_2}) - \log(a_{MgCO_3}) - 2\log(a_{Mg_3Si_6O_{22}(OH)_2})$
18) $4F + 9T = 5A + 4H_2O$	$5\log(a_{Mg_3Si_6O_{22}(OH)_2}) + 4\log(f_{H_2O}) - 4\log(a_{Mg_2SiO_4}) - 9\log(a_{Mg_3Si_6O_{22}(OH)_2})$
19) $9M + A = 8F + H_2O + 9CO_2$	$8\log(a_{Mg_2SiO_4}) + \log(f_{H_2O}) + 9\log(f_{CO_2}) - 9\log(a_{MgCO_3}) - \log(a_{Mg_3Si_6O_{22}(OH)_2})$
20) $5M + T = 4F + H_2O + 5CO_2$	$4\log(a_{Mg_2SiO_4}) + \log(f_{H_2O}) + 5\log(f_{CO_2}) - 5\log(a_{MgCO_3}) - \log(a_{Mg_3Si_6O_{22}(OH)_2})$
21) $2T + 2D = 3M + Tr + CO_2 + H_2O$	$3\log(a_{MgCO_3}) + \log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) + \log(f_{H_2O}) + \log(f_{CO_2}) - 2\log(a_{Mg_3Si_6O_{22}(OH)_2}) - 2\log(a_{CaMgCO_3})$
22) $Tr + 13M = 8F + 2D + 9CO_2$	$8\log(a_{Mg_2SiO_4}) + 2\log(a_{CaMgCO_3}) + 9\log(f_{CO_2}) - \log(a_{Ca_2Mg_3Si_6O_{22}(OH)_2}) - 13\log(a_{MgCO_3})$

Figure 5.10

$\text{Logf}(\text{H}_2\text{O}) - \text{Logf}(\text{CO}_2)$ Diagram

$T = 530^\circ\text{C}$, $P = 4700\text{bars}$

Thermodynamic Data - Chernosky and Berman (1989)

- Berman (1988)

Assuming Non-Component Mineral Compositions

Figure 5.11
 $\text{Logf}_{\text{H}_2\text{O}} \cdot \text{Logf}_{\text{CO}_2}$ Diagram

$T = 560^\circ\text{C}$, $P = 5100\text{bars}$

Thermodynamic Data - Chernosky and Berman (1989)
 - Berman (1988)
 Assuming Non-Component Mineral Compositions

Figure S.14
 $\text{Logf}(\text{H}_2\text{O}) - \text{Logf}(\text{CO}_2)$ Diagram

$T = 560^\circ\text{C}$, $P = 5100\text{bars}$

Thermodynamic Data
 - Helgeson *et al.* (1978)
 Assuming Non-Component Mineral Compositions

Figure 5.15
 $\text{Logf}_{\text{H}_2\text{O}} - \text{Logf}_{\text{CO}_2}$ Diagram

$T = 600^\circ\text{C}$, $P = 5600\text{bars}$
 Assuming Non-Component Mineral Compositions
 Thermodynamic Data - Helgeson *et al.* (1978)

5.15 The Total-Pressure Surface

The total pressure surface drawn on the $\log f_{(\text{H}_2\text{O})}$ - $\log f_{(\text{CO}_2)}$ diagrams shows the location in $\log f_{(\text{H}_2\text{O})}$ - $\log f_{(\text{CO}_2)}$ space where fugacity values meet the condition that:

$$P_{(\text{Total})} = P_{(\text{H}_2\text{O})} + P_{(\text{CO}_2)} \quad (5.7)$$

$$\text{where } P_{(\text{CO}_2)} = f_{(\text{CO}_2)} / \gamma_{(\text{CO}_2)} = X_{(\text{CO}_2)} (P_{(\text{H}_2\text{O})} + P_{(\text{CO}_2)}) \quad (5.8a)$$

$$\text{and } P_{(\text{H}_2\text{O})} = f_{(\text{H}_2\text{O})} / \gamma_{(\text{H}_2\text{O})} \quad (5.8b)$$

$\gamma_{(\text{H}_2\text{O})}$ and $\gamma_{(\text{CO}_2)}$ are the fugacity coefficients for water and carbon dioxide.

The condition that total pressure be equal to the sum of the water and carbon dioxide partial pressures (equation 5.7) is a major limitation of the T - $X_{(\text{CO}_2)}$ diagram but is not a required condition for $\log f_{(\text{H}_2\text{O})}$ - $\log f_{(\text{CO}_2)}$ diagrams.

Chapter Six - Summary

Metamorphic History of the Flinton Creek Ultramafic Rocks

In this chapter, an attempt will be made to describe the metamorphic conditions which produced the mineral assemblages and textures observed in the meta-ultramafic rocks. Data from the three previous chapters will be used including:

- 1) the petrography of the enstatite-magnesite rock and the anthophyllite-carbonate rock, described in chapter 4.
- 2) field observations concerning the spatial relationship between the two ultramafic rock types, (chapter 3).
- 3) the phase diagrams produced in chapter 5.

Figure 6.1 shows the proposed geothermal gradient for the eastern Elzevir Terrain superimposed on Figure 5.4; a pressure-temperature plot of the two solid phase reactions in the system of interest defined in chapter five. Note that differences between the two thermodynamic databases used to plot the reaction lines result in a considerable difference in the location of the tremolite + 4 magnesite = 2 dolomite + anthophyllite curve in pressure-temperature space. Thermodynamic data from Helgeson *et al.* (1978), show that the assemblage dolomite + anthophyllite is stable at the assumed pressures and temperatures of ultramafic metamorphism while Berman (1988), predicts that tremolite and magnesite will co-exist.

Figure 6.1 also demonstrates that neither thermodynamic database clearly accounts for the observed stability of anthophyllite in the Flinton Creek Ultramafic rocks when ideal mineral compositions are assumed. (Although this statement must be taken with a great deal of caution, since an unknown amount of uncertainty exists in the derived geothermal gradient for the eastern Elzevir Terrain, the inferred pressure-temperature conditions of ultramafic metamorphism, and the thermodynamic data used to plot the reaction curves.) When either the Helgeson *et al.* (1978) or Berman (1988) database is used to plot the reaction talc + 4 enstatite = anthophyllite in P-T space, the

Figure 6.1 Solid phase reactions in a portion of the system $\text{CaO-MgO-SiO}_2\text{-H}_2\text{O-CO}_2$, assuming ideal mineral compositions. Also shown is the inferred pressure-temperature gradient for the Eastern Elzevir Terrain and the minimum, intermediate and maximum metamorphic conditions assumed for the Flinton Creek Ultramafic Rocks. Note that for both thermodynamic databases, the entire P-T range of metamorphism assumed for the Flinton Creek Ultramafic Rocks lies in the talc + enstatite stability field, rather than the anthophyllite stability field. Anthophyllite does not become stable along the inferred P-T gradient until approximately the point of muscovite breakdown (shown by a star). It is unreasonable to assume the Flinton Creek Ultramafic Rocks reached these P-T conditions during metamorphism since muscovite is stable in surrounding pelites. Regionally, muscovite breakdown was observed by Rivers (1976) near Flower Station 80 Km north-east of the study area (see Figure 3.5, isograd 4). Also note that the two thermodynamic databases have a difference of approximately 600°C in the location of the tremolite + 4 magnesite = dolomite + anthophyllite curve.

metamorphic window predicted for the ultramafic rocks from the surrounding pelitic assemblages lies in the talc + enstatite stability field (see Fig. 6.1). Anthophyllite does not become stable along the proposed geothermal gradient until approximately the point of muscovite breakdown to potassium feldspar and sillimanite, the reaction which corresponds to isograd 4 near Flower Station on Figure 2.5 (Rivers, 1976).

Figure 6.2 shows the proposed P-T gradient superimposed over Figure 5.5, the location of the solid-solid reactions in pressure-temperature space when actual mineral compositions are used with an ideal activity model to calculate solid phase activities. Both thermodynamic databases now require that anthophyllite be stable, relative to talc and enstatite, over the entire pressure-temperature range of metamorphism inferred for the Flinton Creek ultramafic rocks. Furthermore, since talc and enstatite are stable on the low-temperature side of the reaction, the mineral pair should be replaced by anthophyllite during prograde metamorphism. This is in agreement with petrographic observations made in Chapter 4, where the overgrowth of talc and enstatite by anthophyllite was described (see Plate 4.10b). The replacement shown in Plate 4.10b was not observed in all thin sections however, since other samples were observed to have talc and enstatite in textural equilibrium without anthophyllite. This is assumed to be due to the metastable persistence of the lower temperature, talc and enstatite assemblage. The thermodynamic data of Berman (1988) is more favourable to this assumption than the Helgeson *et al.* (1978) data since the talc + 4 enstatite = anthophyllite reaction plots at higher temperatures using the Berman (1988) data. This means the reaction occurs closer to the inferred metamorphic P-T range for the Flinton Creek ultramafic rocks, requiring the metastable persistence of talc and enstatite over a smaller temperature range.

The temperature difference separating the tremolite + 4 magnesite = 2 dolomite + anthophyllite reaction curve plotted using the Helgeson *et al.* (1978) data from the curve plotted using the Berman (1988) data is lowered by approximately 50 degrees at 1 bar when actual mineral compositions are used in the calculations (see Fig 6.2);

Figure 6.2 Similar to Figure 6.1 except non-component mineral compositions were used in the calculations. Note that anthophyllite is now stable relative to talc and enstatite for both thermodynamic databases over the entire range of metamorphism assumed for the Flinton Creek Ultramafic Rocks. The point of muscovite breakdown is shown for comparison with Figure 6.1.

however, a difference of approximately 500 degrees still remains between the two reaction lines. As in Figure 6.1, Helgeson *et al.* (1978) requires dolomite and anthophyllite to be stable over the inferred P-T range of metamorphism, while Berman (1988) requires that tremolite and magnesite be in stable co-existence.

In the Flinton Creek ultramafic rocks, tremolite grains in contact with magnesite are typically skeletal, developing irregular-shaped grain boundaries which have a corroded appearance, (see Plate 4.9a). Therefore, based on textural evidence, tremolite and magnesite do not appear to form an equilibrium assemblage; this is contrary to the requirements of the Berman (1988) database. Dolomite and anthophyllite, on the other hand, are often found in textural equilibrium in the Flinton Creek ultramafic rocks. The mineral pair is most common in the anthophyllite-carbonate rock. As discussed in chapter 4, when the ultramafic/host rock contact is approached, the amount of dolomite present in the anthophyllite-carbonate rock increases until the rock essentially becomes a biminerallitic anthophyllite-dolomite rock, with both minerals in stable co-existence. Further petrographic evidence supporting anthophyllite-dolomite stability was found in the sample of enstatite-magnesite rock depicted in Plate 4.9b. In this picture, relict tremolite, which may or may not have been in contact with magnesite, has partially reacted to dolomite and has been surrounded by anthophyllite. Thus based on textural evidence, dolomite-anthophyllite appears to be a stable assemblage whereas tremolite-magnesite is not. Therefore, for the system as defined in chapter 5, the Helgeson *et al.* (1978) database is in agreement with the field evidence; the Berman (1988) database is not.

The disagreement of the Berman (1988) database with field observations during consideration of solid-solid reactions in the system carries over into the discussion concerning the effects of volatiles on mineral stability in the ultramafic rocks since the solid phase reactions determine the topology of the $\log f_{(H_2O)} - \log f_{(CO_2)}$ diagrams used in this discussion. If the solid phase reactions are not in agreement with the petrography, then paragenesis diagrams constructed on the $\log f_{(H_2O)} - \log f_{(CO_2)}$ diagrams will also disagree with the petrographic observations.

Whole-rock chemistry discussed in chapter 3 showed that un-metasomatised ultramafic rocks from the Flinton Creek body typically contain up to 5 mole percent CaO and have a $\text{SiO}_2 / \text{MgO}$ ratio anywhere from 0.4 to 0.6. Therefore, when plotted on a CaO-SiO₂-MgO mole percent diagram, the typical Flinton Creek ultramafic rock may be considered to have a bulk composition which falls within an area immediately surrounding the location of enstatite. This observation can be used in conjunction with mineral paragenesis diagrams drawn on the $\log f_{(\text{H}_2\text{O})}$ - $\log f_{(\text{CO}_2)}$ plots below to further understand the development of metamorphic minerals in the ultramafic rocks.

Two $\log f_{(\text{H}_2\text{O})}$ - $\log f_{(\text{CO}_2)}$ diagrams (Fig. 5.14 and 5.15), constructed using thermodynamic data from the Helgeson *et al.* (1978) database are reproduced at a larger scale as Figures 6.3 and 6.4. The shaded area shows the range of H₂O-CO₂ fugacity over which enstatite and magnesite are stable together while the patterned area shows the fugacity range for the stable co-existence of anthophyllite-magnesite. Depending on the $\text{SiO}_2 / \text{MgO}$ ratio of the ultramafic rock, the paragenesis diagram for the portion of the shaded area labelled A on Figure 6.3 shows that two different mineral assemblages could be stable. Those ultramafic rocks with bulk compositions on the SiO₂-rich side on the enstatite-dolomite tie line have an anthophyllite-enstatite-dolomite mineral assemblage while those with bulk compositions on MgO-rich side should have an enstatite-magnesite-dolomite stable assemblage. Both of these mineral assemblages have been observed in the enstatite-magnesite rock. The observed mineral assemblage in the anthophyllite-carbonate rock (anthophyllite, magnesite and dolomite), agrees with required stable mineral assemblage indicated by the paragenesis diagrams in the patterned area.

Figure 6.3 $\text{Log}f_{(\text{H}_2\text{O})} - \text{Log}f_{(\text{CO}_2)}$ diagram at intermediate metamorphic conditions assumed for the Flinton Creek Ultramafic Rocks. The shaded area shows the CO_2 - H_2O fugacity range over which enstatite + magnesite assemblages are stable. The stippled area shows the fugacity range required for anthophyllite + magnesite assemblages. Note that the stable co-existence of enstatite and magnesite at this temperature and pressure, requires an extremely CO_2 -rich fluid ($X_{\text{CO}_2} > 0.95$). Relative to enstatite + magnesite assemblages, anthophyllite + magnesite assemblages co-exist with a more water-rich fluid phase. This fluid must still be comparatively CO_2 -rich however; since λ_{CO_2} must be greater than 0.90.

Figure 6.4 Similar to Figure 6.3; however, this diagram was calculated at the highest metamorphic pressure and temperature conditions assumed for the Flinton Creek Ultramafic Rocks. Note that stable enstatite + magnesite assemblages require a fluid phase with $X_{\text{CO}_2} > 0.90$. A fluid phase with $X_{\text{CO}_2} > 0.60$ is required for anthophyllite + magnesite stability.

$\text{Log}f_{(\text{H}_2\text{O})} - \text{Log}f_{(\text{CO}_2)}$ Diagram
 $T = 600^\circ\text{C}, P = 5600\text{bars}$
 Thermodynamic Data - Helgeson *et al.* (1978)

On the basis of field evidence, presented in chapter 3, it was suggested that the transition from enstatite-magnesite rock to anthophyllite-carbonate rock was related to changing fluid composition rather than changes in pressure or temperature across the ultramafic lenses. Figures 6.3 and 6.4 support this conclusion by demonstrating that increased water fugacity stabilizes anthophyllite-magnesite assemblages (patterned area) relative to enstatite-magnesite assemblages (shaded area). Therefore the transition between the two rock types could be related solely to changing fluid composition.

A minimum value for the mole fraction of carbon dioxide in the fluid phase required to stabilize enstatite-magnesite and anthophyllite-magnesite assemblages can be obtained for the pressure - temperature range of metamorphism inferred for the Flinton Creek ultramafic rocks from surrounding pelitic mineral assemblages by assuming that the range of possible H_2O and CO_2 fugacities is restricted to lie close to the total pressure surface (where $P_{(Total)} = P_{(H_2O)} + P_{(CO_2)}$). Using this assumption, $X_{(CO_2)}$ values can be read directly from Figure 6.3 and 6.4; these results are summarised in Table 6.1 below.

Table 6.1 - Estimates of Fluid Composition

During Metamorphism of the Flinton Creek Ultramafic Rocks

Pressure (bars)	Temp. (°C)	Minimum $X(CO_2)$ for Enst.-Mag. Stability	Minimum $X(CO_2)$ for Anth.-Carb. Stability
5100	560	>0.95	0.90
5600	600	0.90	0.60

Thus, the fluid phase present during the formation of both the enstatite-magnesite rock, and the anthophyllite-carbonate rock, was relatively rich in carbon dioxide. The distribution of the two ultramafic rock types implies that a gradation in $X_{(CO_2)}$ existed across the ultramafic lens, since the preservation of the enstatite-

magnesite assemblage in the core of the ultramafic lens requires a volatile phase richer in carbon dioxide than that required to produce the stable anthophyllite-carbonate assemblage along the lens margins. It seems unlikely that this concentration gradient could have been maintained throughout the period of regional metamorphism in the area. It seems more likely that an initially CO₂-rich fluid, which permitted the growth of enstatite and magnesite, became more water-rich with time. This water-rich fluid infiltrated from the margins of the ultramafic lens inwards, converting the enstatite-magnesite assemblage into anthophyllite and carbonate. It is possible that sediments from the adjacent Flinton Group provided the source for the CO₂, and later H₂O, in this fluid.

The final petrographic observation which must be accounted for in the metamorphic history of the Flinton Creek Ultramafic rocks is the development of late stage forsterite in the enstatite-magnesite rock. Both textural varieties of olivine (coarse and fine-grained), appear to be vein (fluid) related. Petrography of the fine-grained olivine veinlets suggests that the olivine is replacing pre-existing anhedral magnesite grains which are found in enstatite (see Plate 4.8a and 4.8b). Thus, the following equilibrium is interpreted to have constrained olivine formation:

This is a volatile producing reaction; therefore olivine is produced, along with CO₂, by addition of heat to the system. It seems unlikely that forsterite found in the Flinton Creek Ultramafic rocks was formed simply by the addition of heat for two reasons. First, enstatite and magnesite appear stable throughout the rock away from the olivine veinlets. If forsterite formation was due simply to increased temperature, why does the enstatite-magnesite assemblage not break down throughout the rock. Second, no major thermal event postdating the regional metamorphism has been described in the study area.

If addition of heat to form forsterite in the ultramafic rocks is rejected, then a second alternative is to keep the rocks at high temperature and reduce the pressure (isothermal decompression). Figure 6.5 shows a T-X section constructed at 600 C, using the Helgeson *et al.* (1978) database. The temperature selected for this diagram is based on inferred metamorphic conditions for the Kaladar metasediments, which are assumed to have equilibrated near the maximum temperature reached by the Flinton Creek ultramafic rocks. The proposed geothermal gradient for the Eastern Elzevir Terrain passes through 600C at approximately 5600 bars. If this value is used for the initial pressure, and the fluid composition within the ultramafic lens is allowed to become more water rich during decompression (fluid composition changes to point A on Fig. 6.5), then a pressure drop of at least 1500 bars is required prior to crossing the enstatite + magnesite = forsterite + CO₂ reaction. If the fluid remains the same composition (follows arrow to point B on Fig. 6.5), then a pressure drop of 2500 bars is required.

Figure 6.5 P - X_{CO_2} diagram calculated at $T=600^\circ\text{C}$ using the Helgeson *et al.* (1978) database. Shaded area shows the P - X_{CO_2} range for the stable co-existence of enstatite and magnesite. The star indicates the assumed peak pressure of metamorphism reached by the ultramafic rocks at 600°C . From this point, if the fluid stays the same composition (follows a path to point B), a pressure drop of approximately 2.5 kbar is required to cause the formation of forsterite from enstatite and magnesite. On the other hand, if the fluid becomes more water-rich (follows a path to point A), then a pressure drop of approximately 1.5 kbar is required to produce forsterite-bearing assemblages.

Figure 6.6 shows the $\log f_{(\text{H}_2\text{O})}$ - $\log f_{(\text{CO}_2)}$ diagram which results from a pressure drop of 2000 bars, a value between that required to reach point A and point B above. At this pressure (3600 bars) and temperature (600 C) enstatite + magnesite assemblages are stable at $X_{(\text{CO}_2)}$ values greater than 0.9. Forsterite + enstatite and forsterite + magnesite assemblages are stable at $X_{(\text{CO}_2)}$ values between 0.4 and 0.9. The selective replacement of magnesite and enstatite by forsterite can therefore be explained in two ways:

- 1) Injection of a relatively water-rich fluid along small fractures in the enstatite-magnesite rock. The presence of a CO_2 -rich fluid throughout the rest of the rock preserves the enstatite-magnesite assemblage, while the rock is converted to forsterite-bearing assemblages near the veins.
- 2) Injection of a water-rich fluid along small fractures in the relatively dry enstatite-magnesite rock. The assumption that the rocks are dry is valid since the meta-ultramafic rocks have already passed through the thermal peak of metamorphism. The selective replacement of magnesite and enstatite by forsterite could therefore be due to kinetic barriers, with only those portions of the rock exposed to the fluid having any reaction.

Figure 6.6 $\text{Log}f_{(\text{H}_2\text{O})} - \text{Log}f_{(\text{CO}_2)}$ diagram calculated at P-T conditions which result from an isothermal decompression of 2 kbar from 5600 bars and 600°C. Note that forsterite becomes stable relative to enstatite + magnesite for fluids with $X_{\text{CO}_2} < 0.90$. Forsterite + enstatite assemblages and forsterite + magnesite assemblages require a fluid with $0.90 > X_{\text{CO}_2} > 0.75$. Forsterite + anthophyllite assemblages require a fluid with $0.75 > X_{\text{CO}_2} > 0.40$.

3 OF/DE 3

MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS-1963-A
STANDARD REFERENCE MATERIAL 1010a
(ANSI Z39.48 TEST CHART NO. 2)

Chapter Seven - Conclusions

The petrology of a metamorphosed ultramafic body (the Flinton Creek Mafic-Ultramafic Body) within the Central Metasedimentary Belt of the Grenville Province was investigated. The metamorphosed ultramafic rocks occur in discontinuous, elongate lenses surrounded by mafic schists and gneisses.

Monomineralic reaction zones developed between these two dissimilar rock types during regional metamorphism. The following mineral zones are encountered during progression outwards from the ultramafic rocks into the surrounding mafic rocks:

- 1) tremolite zone
- 2) biotite zone
- 3) chlorite zone

The initial boundary between the mafic and ultramafic rocks was at or near the sharp contact between the biotite and tremolite zone. The presence of the reaction zones implies that the ultramafic rocks were metamorphosed *in-situ*.

Within the ultramafic lenses, two different, essentially biminerallitic, meta-ultramafic rock types were discovered: enstatite-magnesite rock and anthophyllite-carbonate rock. Enstatite-magnesite rock is composed of 40-90 percent coarse-grained, blocky, euhedral enstatite and 60-10 percent coarse to medium-grained, subhedral magnesite. The anthophyllite-carbonate rock is composed of fine to medium-grained, anhedral magnesite or dolomite and medium-grained, subhedral anthophyllite. The anthophyllite needles grow in spherical radiating aggregates or rosettes. The enstatite-magnesite rock was found only in the core of the larger ultramafic lenses and was always surrounded by an envelope of anthophyllite-carbonate rock. This spatial relationship was found to be consistent throughout the ultramafic body. Change in fluid composition, rather than pressure or temperature, during regional metamorphism is believed to be responsible for this mineral zonation.

Ultramafic lenses are surrounded by variably deformed mafic rocks. These are generally composed of amphibole and plagioclase with minor amounts of quartz and biotite. Fine-grained metabasites closely resemble the surrounding Tudor Formation (Lumbers, 1967a) volcanics, while coarser-grained metabasites have textures which suggest that they were derived from mafic plutonic rocks. Stress induced grain-size reduction of the coarser-grained material may be responsible for the grain-size variation in the metabasites.

The ultramafic rocks and surrounding mafic rocks were emplaced into the Flinton Synclinorium and Northbrook Batholith as a tectonic melange after the first period of regional deformation D_1 (Thompson, 1972), and prior to, or during the second period, D_2 (Thompson, 1972).

Bulk compositional differences in the system $\text{CaO-SiO}_2\text{-MgO}$ between enstatite-magnesite rock and anthophyllite-carbonate rock are minimal, and cannot account for the different mineralogy of the two rock types.

Chromium and nickel abundances confirm that the ultramafic rocks had an igneous, rather than sedimentary protolith. Whole-rock chemical data suggest that the ultramafic rocks were originally peridotite ultramafic cumulates, from either a stratigraphic-type of intrusion or the cumulate portions of an ophiolite. Observation of what could be relict igneous layering supports this conclusion.

The mineralogy of surrounding pelites suggests that the Flinton Creek Ultramafic rocks have been metamorphosed to at least the kyanite - staurolite zone, and may be partially in the sillimanite zone. Most of the ultramafic rocks are found in the diopside-microcline zone of the Lessard Formation (Thompson, 1973). Using a pressure-temperature gradient derived for the study area and the upper and lower pelitic isograds which surround the ultramafic rocks yields an upper estimated limit of metamorphic conditions for the Flinton Creek rocks ($T=600^\circ\text{C}$, $P=5.6$ kbar) and a lower estimated limit ($T=530^\circ\text{C}$, $P=4.7$ kbar).

The minerals present in the Flinton Creek Ultramafic rocks are metamorphic in origin. Petrographic observation suggests that several mineral assemblages became unstable during metamorphism, resulting in the formation of new mineral assemblages. Some of the new assemblages formed in response to changing pressures and temperatures of metamorphism, while others formed due to changing fluid composition. The newly formed mineral assemblages, and the older ones they replaced, are as follows:

- 1) Enstatite and talc are replaced by anthophyllite.
- 2) Tremolite and magnesite are replaced by dolomite and anthophyllite.
- 3) Enstatite is replaced by anthophyllite and magnesite.
- 4) Enstatite and magnesite are replaced by forsterite.

Several equilibria may have constrained this new mineral growth:

When real mineral compositions, obtained by electron microprobe analyses, are used in conjunction with an ideal activity model, both the Helgeson *et al.*, (1978) and Berman (1988) thermodynamic databases predict that anthophyllite will be stable versus talc and enstatite over the P-T range estimated for ultramafic metamorphism. It is suggested that co-existing talc and enstatite, without anthophyllite overgrowths, is metastable.

Berman (1988) requires tremolite + magnesite to be in stable co-existence. Helgeson *et al.* (1978) predicts that the alternative mineral assemblage anthophyllite + dolomite will be stable. This latter mineral assemblage is found in abundance in the anthophyllite - carbonate rock. Thus data from Helgeson *et al.* (1978) is compatible with the field evidence while Berman's (1988) data is not.

$\text{Log } f_{(\text{CO}_2)} - \text{log } f_{(\text{H}_2\text{O})}$ diagrams suggest that the transition from enstatite-magnesite rock, found in the cores of the ultramafic lenses, to anthophyllite-carbonate rock, on the margin, may be due to changing fluid composition. While the fluid phase present during the formation of both the enstatite-magnesite rock and the anthophyllite-carbonate rocks was relatively rich in CO_2 , the fluid in equilibrium with the enstatite - magnesite rock had to be richer in carbon dioxide than fluid in the anthophyllite - carbonate rock.

The formation of late-stage forsterite from enstatite and magnesite is believed to be due to isothermal decompression during uplift.

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APPENDIX I

WHOLE-ROCK CHEMISTRY ANALYTICAL PROCEDURES AND DATA

Appendix I

A1.1 Sample Collection and Preparation

Due to the inhomogeneous nature of the ultramafic rocks, large samples (the smallest was 2.5 kg) of unweathered and unfractured rock, with no visible veining or mineral segregations, were collected. Any weathered surface remaining on samples which could not be obtained otherwise, was sawed off in a water cooled rock saw at Carleton University. Samples were crushed to chip size in a jaw crusher and then split and remixed twice. Since the sample was now homogenised, a smaller sample (at least 0.25 kg.) was extracted to be pulverised in a stainless steel shatter-box.

A1.2 Analytical Techniques

Analyses of 11 major and 10 trace elements were carried out on a Philips PW 1410 automated X-ray fluorescence (XRF) unit at the University of Ottawa. The determinations were made on fused glass disks, prepared by mixing 1.300 g of the sample with a flux composed of 3.900 g of lithium tetraborate ($\text{Li}_2\text{B}_4\text{O}_7$) and 0.433 g of lithium carbonate (Li_2CO_3). A Cr X-ray tube was used for sample irradiation, and data reduction was completed using the method of De Jongh (1973). Accuracy, determined by duplicate analyses, and precision, determined by analyses of standards (DR-N(1), MRG-1, PCC-1, DTS-1), are given in Table A1.5b and A1.5a respectively.

A1.3 Analytical Results

Table A1.2 lists the analytical results for samples of ultramafic rocks from the Flinton Creek body as well as one sample from the nearby Actinolite body. Table A1.3 lists the XRF results for metabasites associated with the Flinton Creek Ultramafic rocks. Analytical results for gabbroic samples taken from the Kaladar Complex are listed in Table A1.4. In all cases, major element abundances are given in weight percent, while trace element abundances are in parts per million.

AI.4 Accuracy, Precision and Detection Limits

Accuracy and precision for each element, based on repeated analysis of international standards, are given in Table AI.5a. Accuracy is generally satisfactory where the element is present in sufficient amounts. Recommended values for the standards used in this study were obtained from Abbey (1983).

Results for the two ultramafic standards (PCC-1 and DTS-1), show that the accuracy for Al_2O_3 , Na_2O , TiO_2 and P_2O_5 is unsatisfactory. Nickel and chromium are the only two trace elements analysed in the ultramafic rocks with satisfactory accuracy.

Table AI.1 – Whole-Rock Chemistry Sample Descriptions

AI.1a Clinton Creek Ultramafic Rocks

<u>Number</u>	<u>Enst</u>	<u>Anth</u> [*]	<u>Carb</u>	<u>Talc</u>	<u>Trace Minerals</u>	<u>Comments</u>
101588-C1	----	30 (60)	40	30	magnetite	Fresh sample, partial replacement of anth. by talc, homogeneous, anth rosettes growing in med-grained carb.
101588-C2	65	20	10	----	5% mag.	Severely altered (weathered), enstatite completely replaced by fn-grained chlorite and talc.
101788-C1	----	20 (60)	40	40	magnetite	Little alteration with the exception of replacement of talc after anth., large 1-2cm carbonate alteration patches.
103188-C1	----	50	50	Tr	magnetite	Extremely fresh sample, large carbonate alteration patches greenish alteration mineral formed after anthophyllite
103188-C2	----	60	40	Tr	magnetite pyrite	Extremely fresh sample.
110488-C2	70	10	15	----	5% mag.	Extremely fresh to altered, carb.-poor with coarse grained (2x1cm) interlocking enst. crystals, cut by anth. veinlets.
110488-C4	----	30 (40)	60	10	magnetite	Fresh sample with minor replacement of anth. by talc, homogeneous distribution of anth. rosettes in carb. matrix.
110488-C5	70	5	20	Tr	5% mag.	Fresh sample, extremely coarse-grained (8x4cm), minor carb. veining and talc replacement of enst.
110488-C7	----	10 (70)	30	60	magnetite	Altered sample, strong foliation defined by alignment of anth. needles, most anth. replaced by talc.
110488-C8	----	10 (40)	60	30	magnetite	Highly altered, weakly foliated, most anthophyllite replaced by talc.
110488-C9	----	0 (50)	50	50	magnetite	Extremely altered, anth. completely replaced by fine-grained talc, unfoliated.

* Number in bracket refers to amount of anthophyllite present in rock prior to talcification.

Tr Trace

---- Mineral not identified

AI.1b Metabasites Associated with the Flinton Creek Ultramafic Rocks

<u>Number</u>	<u>Plag</u>	<u>Amph</u>	<u>Biot</u>	<u>Other Minerals</u>	<u>Comments</u>
101788-C2	5	80	15	----	Extremely fresh sample, medium-grained, homogeneous, weak foliation and strong lineation.
110188-C1	50	40	10	----	Extremely fresh sample, very fine grained, schistose with strong lineation, homogeneous, no compositional zonation.
110488-C1	----	90	10	5% chalco	Extremely fresh sample, medium-grained, homogeneous, amphibole grains display interlocking, equigranular texture.
110488-C3	40	35	10	10% quartz 5% epidote	Fresh sample, inhomogeneous with a strong gneissic foliation, leucocratic (qtz. & plag.) and melanocratic (amph. & bi.) zones.
110488-C6	30	40	30	----	Fresh sample, homogeneous with a strong schistosity, medium-grained.

AI.1c Kaladar Complex Meta-Gabbroic Rocks

<u>Number</u>	<u>Plag</u>	<u>Amph</u>	<u>Biot</u>	<u>Other Minerals</u>	<u>Comments</u>
111488-C2	30	65	----	5% garnet	Fresh sample, 2 amphiboles: a fine-grained, green variety surrounds a coarse-grained, black type found as single laths.
111488-C3	50	30	20	----	Medium-grained, unfoliated, plagioclase laths are found intergrown with the finer-grained amphibole matrix.
111488-C4	60	5	35	----	Medium to coarse grained, interlocking plagioclase and amphibole laths.
111488-C5	60	5	25	10% K-spar	Fresh sample, weakly foliated but no compositional zoning.

TABLE A1.2

X-Ray Fluorescence analyses of the Flinton Creek Ultramafic rocks.

Sample Rock Type	1015-C1 Anth-Carb	1015-C2 Enst-Mag	1017-C Anth-Carb	1031-C1 Anth-Carb	1031-C2 Anth-Carb	1104-C2A+ Enst-Mag	1104-C2B+ Enst-Mag	1104-C4 Anth-Carb	1104-C5 Enst-Mag N=2#
SiO ₂	48.73	31.03	48.98	41.38	37.02	45.71	51.79	25.80	39.12
Al ₂ O ₃	1.31	0.89	3.12	0.57	0.84	1.76	1.68	1.12	2.07
Fe ₂ O ₃ *	11.28	8.95	6.51	7.15	6.99	11.51	10.41	9.82	11.54
MgO	26.48	27.86	30.56	32.64	27.57	31.15	33.19	24.48	33.50
CaO	2.94	11.47	3.13	3.25	9.84	1.78	0.43	14.09	1.64
Na ₂ O	0.12	0.39	0.07	0.16	0.15	0.06	0.00	0.17	0.24
K ₂ O	0.01	0.02	0.01	0.10	0.02	0.02	0.01	0.01	0.01
TiO ₂	0.06	0.02	0.09	0.01	0.02	0.07	0.06	0.03	0.06
P ₂ O ₅	0.00	0.00	0.00	0.01	0.00	0.01	0.01	0.00	0.01
MnO	0.11	0.14	0.13	0.11	0.16	0.13	0.15	0.23	0.14
S	0.05	0.06	0.07	0.09	0.10	0.11	0.12	0.05	0.05
Ba	38	51	41	61	40	66	77	45	70
Cr	5884	4364	2747	2091	1575	7617	6255	6323	5156
Zr	<10	<10	<10	<10	<10	<10	<10	<10	<10
Sr	<10	143	22	20	81	31	<10	125	17
Rb	<10	<10	<10	13	<10	<10	<10	<10	<10
Y	<10	<10	<10	<10	<10	<10	<10	<10	<10
Nb	<10	<10	<10	<10	<10	<10	<10	<10	<10
Zn	110	43	127	94	60	99	109	84	75
Ni	1796	1857	2073	1877	1202	2237	2302	1693	1949
V	76	30	40	20	22	65	57	29	56
TOTAL	92.211	81.726	93.359	86.037	83.132	93.752	99.107	76.975	89.406

* total iron as Fe₂O₃

+ two samples, (A and B), taken from different locations in the same outcrop

duplicate analyses, see Table A1.5b

TABLE A1.2, continued
X-Ray Fluorescence analyses of the Flinton Creek Ultramafic rocks.

Sample Rock Type	1104-C7A+ Anth-Carb	1104-C7B+ Anth-Carb	1104-C8 Anth-Carb	1104-C9 Anth-Carb	1114-C1**
SiO ₂	46.22	48.24	28.75	31.84	36.39
Al ₂ O ₃	4.86	1.50	0.76	1.14	1.78
Fe ₂ O ₃ *	9.68	9.05	7.14	6.47	11.48
MgO	29.22	28.45	24.45	25.17	33.60
CaO	2.42	4.47	14.69	12.72	2.04
Na ₂ O	0.00	0.29	0.17	0.10	0.26
K ₂ O	0.01	0.01	0.01	0.01	0.01
TiO ₂	0.04	0.05	0.01	0.04	0.07
P ₂ O ₅	0.01	0.01	0.00	0.00	0.01
MnO	0.10	0.12	0.29	0.17	0.10
S	0.07	0.05	0.08	0.05	0.09
Ba	83	65	71	54	54
Cr	3231	3333	4962	5534	8149
Zr	<10	<10	<10	<10	<10
Sr	45	98	99	107	31
Rb	<10	<10	<10	<10	<10
Y	<10	<10	<10	<10	<10
Nb	<10	<10	<10	<10	<10
Zn	72	101	60	77	77
Ni	1487	1341	1880	1449	2248
V	47	28	11	67	76
TOTAL	93.328	92.976	77.344	78.753	87.346

* total iron as Fe₂O₃

+ two samples, (A and B), taken from different locations in the same outcrop

** sample from Actinolite ultramafic body located south of the Elzevir Batholith

TABLE A1.3

*X-Ray Fluorescence analyses of metabasites associated with
the Flinton Creek Ultramafic rocks*

SAMPLE	1017-C2	1101-C1	1104-C1	1104-C3A+	1104-C3B+	1104-C6 N=2#
SiO ₂	33.95	56.85	39.32	52.87	52.40	47.40
Al ₂ O ₃	21.32	17.93	12.37	14.43	15.02	13.10
Fe ₂ O ₃ *	15.70	7.12	14.72	8.42	8.59	15.84
MgO	12.20	3.92	15.66	7.93	7.72	6.85
CaO	10.45	6.18	11.22	10.15	10.00	6.42
Na ₂ O	1.77	5.71	2.08	0.61	0.64	2.29
K ₂ O	1.82	0.32	0.44	2.36	2.59	2.46
TiO ₂	2.15	0.80	2.95	1.14	1.17	2.91
P ₂ O ₅	0.47	0.19	0.48	0.27	0.27	0.45
MnO	0.24	0.10	0.22	0.13	0.13	0.18
S	0.05	0.08	0.08	0.05	0.04	0.07
Ba	571	150	<10	746	631	146
Cr	281	20	768	535	307	25
Zr	193	123	222	164	163	193
Sr	777	475	176	235	246	129
Rb	41	<10	<10	56	51	88
Y	50	23	27	24	24	62
Nb	<10	<10	20	15	16	<10
Zn	111	81	54	90	104	100
Ni	124	<10	360	144	78	12
V	272	141	307	78	87	356
TOTAL	100.424	99.332	99.811	98.641	98.781	98.056

* total iron as Fe₂O₃

+ two samples (A and B), taken from different locations in the same outcrop

duplicate analyses, see Table A1.5b

TABLE AI.4

X-Ray Fluorescence analyses of Kaladar Complex metagabbroic rocks

SAMPLE	1114-C2	1114-C3	1114-C4	1114-C5 N=2#
SiO ₂	47.55	55.08	58.18	62.42
Al ₂ O ₃	16.76	15.40	16.23	16.37
Fe ₂ O ₃ *	8.70	8.49	6.81	4.91
MgO	12.35	7.29	5.39	3.46
CaO	9.90	6.67	5.58	3.93
Na ₂ O	1.64	3.55	3.88	4.13
K ₂ O	1.72	1.51	2.31	3.48
TiO ₂	0.99	1.19	1.08	0.74
P ₂ O ₅	0.00	0.34	0.28	0.18
MnO	0.16	0.13	0.10	0.07
S	0.05	0.05	0.05	0.05
Ba	402	270	416	414
Cr	525	326	231	119
Zr	24	117	136	215
Sr	379	409	369	320
Rb	26	39	64	104
Y	15	24	29	25
Nb	<10	12	14	12
Zn	88	91	73	72
Ni	243	90	63	49
V	246	169	135	78
TOTAL	100.079	99.900	100.089	99.899

* total iron as Fe₂O₃

duplicate analyses, see Table AI.5b

TABLE A1.5a

Analytical data for standards PCC-1 and DTS-1

	PCC - 1				DTS - 1			
	N=5		RV	ACC %	N=3		RV	ACC %
	AVERAGE	RANGE			AVERAGE	RANGE		
SiO ₂	42.11	0.97	42.10	0.0	40.41	0.66	40.61	0.5
Al ₂ O ₃	0.68	0.02	0.73	7.4	0.22	0.02	0.25	13.3
Fe ₂ O ₃ *	8.46	0.42	8.29	2.0	8.64	0.26	8.73	1.0
MgO	43.31	1.00	43.50	0.4	48.57	0.55	49.80	2.5
CaO	0.55	0.02	0.55	0.4	0.13	0	0.14	7.1
Na ₂ O	0.14	0.25	0.01	1300	0.15	0.44	0.01	1367
K ₂ O	0.00	0.00	0.00	----	0.00	0	0.00	----
TiO ₂	0.00	0.01	0.01	80	0.00	0	0.00	----
P ₂ O ₅	0.00	0.01	0.01	60	0.00	0.01	0.00	----
MnO	0.12	0.01	0.12	3.3	0.13	0	0.12	8.3
S	0.01	0.00	0.01	0.0	0.01	0.01	0.00	----
Ba	44	29	4	1005	47	27	5	833
Cr	2848	58	2800	1.7	4145	48	4200	1.3
Zr	0	0	<10	----	0	0	<10	----
Sr	0	0	0.4	100	0	0	0.4	100
Rb	0	0	0.3	100	0	0	0.1	100
Y	1	2	----	----	0	0	<10	----
Nb	0	0	1	100	0	0	----	----
Zn	62	48	41	51	87	18	46	90
Ni	2429	96	2400	1.2	2294	47	2300	0.2
V	30	8	29	2.1	5	5	11	52
Total	96.128				99.170			

Range Maximum - minimum value

RV Recommended Value, Abbey (1983)

ACC % Accuracy = $\{[(\text{Avg}-\text{RV})(\text{Avg}-\text{RV})]^{0.5} / \text{RV}\} \times 100$ * total iron as Fe₂O₃

---- Not Determined

TABLE A1.5a, continued

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Analytical data for standards DR-N(1) and MRG-1

	DR - N(1)				MRG - 1			
	N=3		RV	ACC %	N=3		RV	ACC %
	AVERAGE	RANGE			AVERAGE	RANGE		
SiO ₂	53.53	0.43	52.98	1.0	38.94	0.52	39.32	1.0
Al ₂ O ₃	17.71	0.06	17.56	0.8	8.42	0.08	8.50	0.9
Fe ₂ O ₃ *	9.63	0.31	9.72	1.0	17.35	0.34	17.85	2.8
MgO	4.51	0.24	4.41	2.3	13.93	0.17	13.49	3.3
CaO	7.10	0.04	7.07	0.5	14.84	0.03	14.77	0.5
Na ₂ O	3.00	0.16	3	0.1	1.08	0.21	0.71	52
K ₂ O	1.71	0.01	1.7	0.8	0.18	0	0.18	0.0
TiO ₂	1.05	0.01	1.09	3.4	3.75	0.02	3.69	1.5
P ₂ O ₅	0.22	0.03	0.25	13.3	0.03	0.01	0.06	56
MnO	0.22	0	0.22	0.0	0.17	0	0.17	0.0
S	0.04	0.01	----	----	0.06	0	0.06	0.0
Ba	407	41	390	4.3	0	0	----	----
Cr	39	3	42	6.3	462	14	450	2.7
Zr	128	2	125	2.4	92	2	105	12.4
Sr	392	5	400	2.0	258	0	260	0.8
Rb	67	4	70	4.3	0	0	8	100
Y	30	1	30	1.1	17	0	16	6.3
Nb	5	1	6	22	17	4	20	16.7
Zn	161	31	145	11.3	198	54	190	4.0
Ni	21	5	22	6.1	154	3	195	21
V	210	5	230	8.7	535	12	----	----
Total	98.912				99.002			

Range Maximum - minimum value

RV Recommended Value, Abbey (1983)

ACC % Accuracy = $\{[(\text{Avg}-\text{RV})(\text{Avg}-\text{RV})]^{0.5} / \text{RV}\} \times 100$ * total iron as Fe₂O₃

---- Not Determined

TABLE A1.5b
Duplicate analyses

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Sample	1114-C5	1114-C5d	1104-C6	1104-C6d	1104-C5	1104-C5d
Rock Type	Gabbro	Gabbro	Amph.	Amph.	Enst-Mag	Enst-Mag
SiO ₂	62.70	62.14	47.36	47.44	39.03	39.21
Al ₂ O ₃	16.44	16.30	13.11	13.08	2.05	2.08
Fe ₂ O ₃ *	4.92	4.90	15.69	15.99	11.62	11.46
MgO	3.49	3.42	6.84	6.86	33.44	33.56
CaO	3.94	3.91	6.42	6.41	1.64	1.64
Na ₂ O	4.13	4.13	2.40	2.13	0.08	0.40
K ₂ O	3.49	3.47	2.45	2.46	0.01	0.01
TiO ₂	0.74	0.73	2.92	2.89	0.06	0.06
P ₂ O ₅	0.18	0.17	0.46	0.44	0.01	0.01
MnO	0.07	0.07	0.18	0.18	0.14	0.14
S	0.05	0.04	0.04	0.09	0.05	0.05
Ba	428	400	154	138	49	91
Cr	120	117	27	22	5150	5161
Zr	215	215	195	191	<10	<10
Sr	318	322	128	129	17	16
Rb	103	105	88	87	<10	<10
Y	24	25	61	62	<10	<10
Nb	12	11	<10	<10	<10	<10
Zn	68	75	91	108	78	72
Ni	55	43	23	<10	1982	1915
V	76	80	348	363	52	59
TOTAL	100.33	99.47	98.03	98.17	89.16	89.65

* total iron as Fe₂O₃

APPENDIX II

MINERAL CHEMISTRY ELECTRON MICROPROBE ANALYSES

Appendix II

AII.1 Electron Microprobe Analysis

The composition of minerals in the enstatite-magnesite rock and anthophyllite-carbonate rock were determined at Carleton University by energy dispersive spectrometry (EDS) using a Cambridge Microscan 5 electron microprobe (Hamster Probe). Operating conditions were set at 20 kV accelerating potential and a specimen current of 9.0 nanoamperes, measured on an amphibole standard (Kakanui hornblende, U.S.N.M. Standard 143965). The average spectrum collection time was 100 seconds (live detector time). Data reduction was completed on a personal computer using the EDDI program of Pringle (1989). In this program, all iron is calculated as FeO.

Results of microprobe analyses for various minerals are listed in the following tables:

<u>Mineral</u>	<u>Table</u>
orthopyroxene	AII.1
magnesite	AII.2
dolomite	AII.3
talc	AII.4
chlorite	AII.5
olivine	AII.6
tremolite	AII.7
magnesio-cummingtonite	AII.8
anthophyllite	AII.9
chromium-magnetite	AII.10

Table All.1
Microprobe analyses of orthopyroxene

	AVERAGE N=9	STD DEV	AVERAGE N=34	STD DEV	AVERAGE N=7	STD DEV
Sample	111587-1C		032188-2B		032188-5A	
Rock Type	Enst+Mag		Enst+Mag		Enst+Mag	
SiO ₂	57.65	0.55	57.35	0.30	57.87	0.43
TiO ₂	0.07	0.04	0.08	0.04	0.08	0.03
Al ₂ O ₃	0.08	0.12	0.24	0.11	0.01	0.01
Cr ₂ O ₃	0.10	0.05	0.09	0.05	0.08	0.04
FeO *	6.05	0.19	6.87	0.37	7.52	0.13
MgO	34.99	0.47	34.78	0.49	34.53	0.35
CaO	0.09	0.03	0.12	0.03	0.12	0.02
MnO	0.14	0.04	0.19	0.05	0.19	0.05
NiO	n/a**	n/a	0.05	0.04	n/a	n/a
TOTAL	99.17	0.93	99.76	0.48	100.39	0.68
<i>Cation proportions on the basis of 6 oxygens</i>						
Si	1.998	0.007	1.986	0.009	1.996	0.005
Al (IV)	0.003	0.005	0.010	0.005	0.000	0.001
Sum Z	2.002	0.012	1.996	0.007	1.996	0.006
Ti	0.002	0.001	0.002	0.001	0.002	0.001
Cr	0.003	0.001	0.003	0.001	0.002	0.001
Fe	0.175	0.005	0.199	0.011	0.217	0.004
Mg	1.808	0.014	1.795	0.021	1.775	0.011
Ca	0.003	0.001	0.004	0.001	0.005	0.001
Mn	0.004	0.001	0.005	0.002	0.006	0.001
Ni	n/a**	n/a	0.001	0.001	n/a	n/a
Sum Ti to Ni	1.991	0.022	2.010	0.014	2.000	0.018
TOTAL	3.997	0.006	4.006	0.007	4.002	0.006
mg#	0.907		0.894		0.886	

* Total iron as FeO

** Not analysed

mg# = $Mg / (Mg + Fe + Mn + Ca + Cr + Ni)$

Table All.1, continued
Microprobe analyses of orthopyroxene

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	AVERAGE N=41	STD DEV	AVERAGE N=14	STD DEV	AVERAGE N=14	STD DEV
Sample	101788-1B		110488-01		110488-C2B	
Rock Type	Enst+Mag		Enst+Mag		Enst+Mag	
SiO ₂	57.93	0.67	57.56	0.52	57.46	0.34
TiO ₂	0.11	0.04	0.09	0.04	0.10	0.04
Al ₂ O ₃	0.14	0.17	0.08	0.11	0.00	0.00
Cr ₂ O ₃	0.14	0.05	0.12	0.02	0.10	0.04
FeO *	6.19	0.60	8.07	0.16	7.25	0.22
MgO	35.08	0.58	33.83	0.26	34.57	0.29
CaO	0.10	0.03	0.14	0.02	0.11	0.02
MnO	0.15	0.05	0.24	0.04	0.24	0.05
NiO	n/a**	n/a	n/a	n/a	n/a	n/a
TOTAL	99.85	0.90	100.12	0.70	99.84	0.49
<i>Cation proportions on the basis of 6 oxygens</i>						
Si	1.996	0.008	1.995	0.006	1.992	0.006
Al (IV)	0.006	0.007	0.003	0.005	0.000	0.000
Sum Z	2.002	0.015	1.999	0.010	1.992	0.006
Ti	0.003	0.001	0.002	0.001	0.003	0.001
Cr	0.004	0.001	0.003	0.001	0.003	0.001
Fe	0.179	0.018	0.234	0.004	0.210	0.006
Mg	1.791	0.068	1.749	0.012	1.786	0.012
Ca	0.004	0.001	0.005	0.001	0.004	0.001
Mn	0.004	0.002	0.007	0.001	0.008	0.004
Ni	n/a**	n/a	n/a	n/a	n/a	n/a
Sum Ti to Ni	1.980	0.090	1.993	0.019	2.006	0.021
TOTAL	3.986	0.065	3.999	0.005	4.005	0.006
mg#	0.904		0.875		0.888	

* Total iron as FeO

** Not analysed

mg# = $Mg / (Mg + Fe + Mn + Ca + Cr + Ni)$

Table All.2. continued
Microprobe analyses of magnesite

Sample	Rock Type	103188-C2		110488-C2B	
		AVERAGE N=9 Anth+Carb	STD DEV	AVERAGE N=11 Enst+Mag	STD DEV
MgO		44.42	0.40	43.66	0.38
FeO *		4.64	0.20	4.66	0.09
CaO		0.55	0.30	0.39	0.03
MnO		0.39	0.13	0.23	0.04
NiO		0.04	0.03	0.08	0.09
TOTAL		50.04	0.45	49.02	0.32
<i>Cation proportions based on 6 oxygens</i>					
Mg		1.864	0.008	1.869	0.005
Fe		0.109	0.005	0.112	0.002
Ca		0.017	0.009	0.012	0.001
Mn		0.009	0.003	0.006	0.001
Ni		0.001	0.001	0.002	0.002
TOTAL		2.000	0.001	2.000	0.001
mg#		0.9320		0.9344	

* Total iron as FeO
mg# = $Mg / (Mg + Fe + Ni + Mn + Ca)$

Table A11.3
Microprobe analyses of dolomite

Sample	AVERAGE N=12	STD DEV	AVERAGE N=17	STD DEV	AVERAGE N=12	STD DEV	AVERAGE N=9	STD DEV	AVERAGE N=21	STD DEV
Rock Type	Enst+Mag		Enst+Mag		Anth+Carb		Enst+Mag		Anth+Carb	
111587-1C	032188-2B		103188-C2		110488-C4					
FeO *	1.16	0.06	0.98	0.30	4.52	0.58	1.45	0.10	1.51	0.20
MgO	21.25	0.14	21.29	0.23	19.29	0.78	20.97	0.20	21.16	0.19
MnO	0.08	0.03	0.24	0.11	0.32	0.08	0.08	0.07	0.28	0.11
SrO	0.01	0.02	0.03	0.06	n/a**	n/a	0.44	0.32	n/a	n/a
CaO	29.25	0.25	29.35	0.17	29.08	0.82	29.23	0.21	29.12	0.24
TOTAL	51.74	0.32	51.89	0.28	53.20	0.44	52.17	0.44	52.07	0.42
Cation proportions based on six oxygens										
Fe	0.030	0.002	0.026	0.008	0.119	0.015	0.038	0.002	0.039	0.005
Mg	0.989	0.006	0.988	0.007	0.900	0.032	0.975	0.005	0.981	0.008
Mn	0.002	0.001	0.006	0.003	0.009	0.002	0.002	0.002	0.007	0.003
Sr	0.000	0.001	0.001	0.001	n/a	n/a	0.008	0.006	n/a	n/a
Total	1.021	0.005	1.021	0.005	1.027	0.030	1.023	0.008	1.028	0.005
Ca	0.979	0.005	0.979	0.005	0.973	0.031	0.977	0.008	0.972	0.005
TOTAL	2.000	0.001	2.000	0.001	2.000	0.000	2.000	0.000	2.000	0.000
mg#	0.9682		0.9681		0.8761		0.9532		0.9547	

* Total iron as FeO

** Not analysed

mg# = Mg/(Mg+Fe+Mn+Sr)

Table All.4

Microprobe analyses of talc.

	Average N=10	STD DEV	Average N=27	STD DEV
Sample	101788-1A		103188-C1	
Rock Type	Enst+Mag		Anth+Carb	
SiO ₂	62.06	0.26	61.86	0.51
TiO ₂	0.14	0.04	0.12	0.03
Al ₂ O ₃	0.15	0.11	1.13	0.25
Cr ₂ O ₃	0.12	0.04	0.16	0.04
FeO *	1.16	0.05	1.21	0.05
MgO	30.45	0.20	30.36	0.28
MnO	0.06	0.03	0.06	0.04
NiO	0.35	0.03	0.29	0.06
TOTAL	94.48	0.39	95.19	0.69

Cation proportions based on twenty-two oxygens

Si	7.965	0.016	7.883	0.035
Al (IV)	0.022	0.017	0.117	0.035
Sum Z	7.977	0.013	8.000	0.000
Al (VI)	0.001	0.002	0.053	0.016
Ti	0.014	0.003	0.011	0.003
Cr	0.012	0.004	0.016	0.004
Fe	0.124	0.005	0.130	0.005
Mg	5.824	0.024	5.767	0.027
Mn	0.007	0.003	0.006	0.004
Ni	0.033	0.010	0.029	0.006
Sum Al(VI) to Ni	6.0141	0.028	6.013	0.018
TOTAL	14.001	0.016	14.013	0.018
mg#	0.971		0.969	

* Total iron as FeO

mg# = $Mg/(Mg+Fe+Mn+Cr+Ni)$

Table AII.5
Microprobe analyses of chlorite

	AVERAGE N=4	STD DEV	AVERAGE N=3	STD DEV	AVERAGE N=15	STD DEV
Sample	111587-1C		032188-5A		032188-5B	
Rock Type	Enst+Mag		Enst+Mag		Contact	
SiO ₂	32.89	1.50	34.03	2.03	31.56	0.58
TiO ₂	0.10	0.03	0.11	0.01	0.12	0.03
Al ₂ O ₃	16.31	1.25	15.51	2.04	17.35	0.39
Cr ₂ O ₃	1.31	0.16	1.50	0.03	1.38	0.12
FeO *	3.44	0.16	3.71	0.31	3.94	0.15
MgO	35.01	0.38	34.91	0.22	32.94	0.22
MnO	0.11	0.04	0.09	0.05	0.05	0.03
NiO	0.41	0.02	0.23	0.02	0.18	0.04
TOTAL	89.56	0.64	90.09	0.13	87.52	0.31
<i>Cation proportions on the basis of twenty-eight oxygens</i>						
Si	6.065	0.222	6.234	0.354	5.963	0.098
Al (IV)	1.935	0.164	1.766	0.236	2.037	0.098
Sum Z	8.000	0.000	8.000	0.000	8.000	0.000
Al (VI)	1.614	0.136	1.585	0.212	1.827	0.036
Ti	0.014	0.004	0.015	0.001	0.017	0.004
Cr	0.191	0.023	0.218	0.004	0.206	0.018
Fe	0.530	0.025	0.568	0.045	0.622	0.025
Mg	9.626	0.057	9.534	0.086	9.277	0.076
Mn	0.017	0.006	0.013	0.007	0.008	0.005
Ni	0.060	0.003	0.033	0.003	0.027	0.006
Sum Al (VI) to Ni	12.052	0.072	11.966	0.129	11.985	0.056
TOTAL	20.052	0.072	19.966	0.129	19.985	0.056
mg#	0.923		0.920		0.915	

* Total iron as FeO

mg# = $Mg/(Mg+Fe+Mn+Ni+Cr)$

Table AII.6
Microprobe analyses of olivine

Sample	Rock Type	111587-1C		032188-5A		050888-2A		101788-1B		110488-C2B		050489-3B	
		AVERAGE N=4	STD DEV	AVERAGE N=5	STD DEV	AVERAGE N=26	STD DEV	AVERAGE N=26	STD DEV	AVERAGE N=46	STD DEV	AVERAGE N=48	STD DEV
		Enst+Mag		Enst+Mag		Enst+Mag		Enst+Mag		Enst+Mag		Enst+Mag	
SiO ₂		41.49	0.42	40.83	0.61	40.79	0.25	40.97	0.64	40.68	0.20	40.95	0.29
FeO *		8.09	0.16	9.16	0.32	8.54	0.17	8.67	0.27	9.32	0.23	9.30	0.19
MgO		50.98	0.31	49.54	0.53	50.07	0.34	49.36	0.46	48.77	0.48	49.37	0.39
MnO		0.16	0.02	0.14	0.03	0.20	0.04	0.14	0.05	0.16	0.04	0.10	0.05
NiO		0.74	0.05	0.43	0.06	0.52	0.05	0.62	0.07	0.37	0.10	0.39	0.07
TOTAL		101.45	0.65	100.10	1.48	100.11	0.61	99.76	0.97	99.29	0.68	100.08	0.64
<i>Cation proportions based on four oxygens</i>													
Si		0.997	0.005	0.998	0.002	0.995	0.004	1.007	0.008	1.003	0.005	1.001	0.004
Fe		0.162	0.003	0.187	0.004	0.174	0.003	0.178	0.006	0.192	0.005	0.190	0.003
Mg		1.826	0.010	1.805	0.008	1.821	0.007	1.802	0.014	1.792	0.009	1.799	0.008
Mn		0.003	0.000	0.003	0.001	0.004	0.001	0.003	0.001	0.003	0.001	0.002	0.001
Ni		0.014	0.001	0.009	0.001	0.010	0.001	0.012	0.001	0.007	0.002	0.008	0.001
TOTAL		3.003	0.004	3.003	0.002	3.005	0.004	2.997	0.008	2.998	0.005	2.999	0.004
mg#		0.910		0.901		0.906		0.903		0.898		0.900	

* Total iron as FeO
mg# = Mg/(Mg+Fe+Ni+Mn)

Table All.7
Microprobe analyses of tremolite

	AVERAGE N=29	STD DEV	AVERAGE N=5	STD DEV
Sample	111587-1C		032188-2B	
Rock Type	Enst+Mag		Enst+Mag	
SiO ₂	57.33	0.49	56.80	0.27
TiO ₂	0.18	0.05	0.13	0.04
Al ₂ O ₃	0.67	0.26	0.32	0.11
Cr ₂ O ₃	0.33	0.08	0.27	0.04
FeO *	1.93	0.16	1.94	0.18
MgO	23.64	0.32	23.48	0.31
MnO	0.12	0.05	0.10	0.04
CaO	13.20	0.26	13.25	0.48
K ₂ O	0.09	0.02	0.06	0.02
Na ₂ O	0.02	0.04	0.00	0.00
TOTAL	97.52	0.69	96.36	0.44
<i>Cation proportions on the basis of twenty-three oxygens</i>				
Si	7.863	0.029	7.888	0.013
Al (IV)	0.106	0.038	0.052	0.017
Sum Z	7.969	0.023	7.941	0.016
Al (VI)	0.003	0.012	0.000	0.000
Ti	0.018	0.005	0.014	0.004
Cr	0.036	0.008	0.029	0.005
Fe	0.221	0.018	0.225	0.021
Mg	4.834	0.053	4.860	0.057
Mn	0.014	0.006	0.012	0.005
Sum Al (VI) to Mn	5.127	0.044	5.140	0.082
Ca	1.940	0.041	1.972	0.071
K	0.015	0.004	0.011	0.003
Na	0.005	0.012	0.000	0.000
Sum Ca + Alkalies	1.960	0.045	1.983	0.070
TOTAL	15.0	0.019	15.064	0.011
mg#	0.947		0.948	

* Total iron as FeO

mg# = Mg/(Mg+Fe+Mn+Cr)

Table All. 7, continued
Microprobe analyses of tremolite

	AVERAGE N=21	STD DEV	AVERAGE N=7	STD DEV
Sample	032188-5B		110488-C2B	
Rock Type	Contact		Enst+Mag	
SiO ₂	57.51	0.40	54.74	0.76
TiO ₂	0.15	0.05	0.18	0.03
Al ₂ O ₃	0.37	0.17	2.44	0.75
Cr ₂ O ₃	0.21	0.04	0.45	0.09
FeO *	2.22	0.16	2.68	0.23
MgO	23.65	0.29	22.62	0.28
MnO	0.12	0.06	0.09	0.04
CaO	13.17	0.30	12.85	0.14
K ₂ O	0.08	0.02	0.09	0.03
Na ₂ O	0.03	0.06	0.32	0.26
TOTAL	97.56	0.57	96.44	0.53

Cation proportions on the basis of twenty-three oxygens

Si	7.835	0.173	7.642	0.099
Al (IV)	0.061	0.028	0.355	0.107
Sum Z	7.896	0.176	7.996	0.009
Al (VI)	0.000	0.000	0.047	0.023
Ti	0.016	0.005	0.019	0.003
Cr	0.023	0.005	0.049	0.010
Fe	0.254	0.018	0.313	0.027
Mg	4.839	0.046	4.706	0.047
Mn	0.014	0.007	0.010	0.004
Sum Al (VI) to Mn	5.146	0.060	5.144	0.019
Ca	1.936	0.046	1.923	0.024
K	0.014	0.004	0.016	0.005
Na	0.008	0.016	0.086	0.070
Sum Ca + Alkalies	1.958	0.050	2.025	0.053
TOTAL	15.000	0.177	15.166	0.064
mg#	0.943		0.927	

* Total iron as FeO

mg# = Mg/(Mg+Fe+Mn+Cr)

Table A11.8

Microprobe analyses of Magnesio-Cummingtonite

Sample	111587-1C							
Analysis	4.2	3.2	8.2	12.2	14.2	10.2	9.2	17.2
SiO ₂	59.28	58.84	58.35	58.08	58.22	57.75	56.66	57.08
TiO ₂	0.04	0.04	0.15	0.16	0.24	0.17	0.13	0.22
Al ₂ O ₃	0.09	0.01	0.14	0.39	0.63	0.14	0.63	0.43
Cr ₂ O ₃	0.06	0.03	0.22	0.28	0.25	0.24	0.31	0.31
FeO*	6.92	5.89	4.87	3.71	3.48	2.77	2.13	1.98
MgO	31.01	30.27	28.56	26.67	25.86	25.41	23.88	23.44
MnO	0.36	0.31	0.27	0.21	0.17	0.15	0.11	0.11
CaO	0.58	2.47	5.71	8.64	9.44	10.55	12.64	13.13
K ₂ O	0.13	0.12	0.10	0.10	0.11	0.10	0.06	0.08
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TOTAL	98.47	97.98	98.37	98.24	98.40	97.28	96.55	96.78
<i>Cation proportions on the basis of twenty-three oxygens</i>								
Si	7.93	7.923	7.872	7.868	7.877	7.908	7.848	7.89
Al (IV)	0.01	0.001	0.022	0.063	0.101	0.022	0.103	0.07
Sum Z	7.946	7.924	7.894	7.931	7.978	7.930	7.951	7.959
Al (VI)	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ti	0.004	0.004	0.015	0.017	0.025	0.017	0.014	0.023
Cr	0.006	0.004	0.024	0.030	0.027	0.026	0.034	0.034
Fe	0.775	0.664	0.550	0.421	0.393	0.317	0.247	0.229
Mg	6.186	6.074	5.743	5.385	5.216	5.186	4.931	4.828
Mn	0.040	0.036	0.031	0.024	0.019	0.017	0.013	0.013
Sum Al (VI) to Mn	7.011	6.782	6.363	5.877	5.680	5.563	5.239	5.127
Ca	0.083	0.356	0.826	1.254	1.368	1.548	1.875	1.944
K	0.022	0.020	0.017	0.017	0.018	0.018	0.010	0.014
Na	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Sum Ca + Alkalies	0.105	0.376	0.843	1.271	1.386	1.566	1.885	1.958
TOTAL	15.062	15.082	15.100	15.079	15.044	15.059	15.075	15.044
mg#	0.883	0.896	0.905	0.919	0.922	0.935	0.944	0.946
ca#	0.012	0.050	0.116	0.178	0.196	0.220	0.266	0.278

* Total iron as FeO

mg# = Mg/(Mg+Fe+Mn+Cr)

ca# = Ca/(Ca+Mg+Fe)

Table All.9

Microprobe analyses of anthophyllite

	AVERAGE N=6	STD DEV	AVERAGE N=18	STD DEV	AVERAGE N=18	STD DEV
Sample	111587-1C		032188-2B		032188-5B	
Rock Type	Enst+Mag		Enst+Mag		Contact	
SiO ₂	55.48	0.27	58.54	0.23	59.69	0.54
TiO ₂	0.08	0.05	0.06	0.04	0.09	0.03
Al ₂ O ₃	0.14	0.07	0.16	0.10	0.17	0.08
Cr ₂ O ₃	0.10	0.03	0.05	0.03	0.07	0.04
FeO *	6.88	0.12	7.17	0.29	6.62	0.48
MgO	30.72	0.26	30.46	0.47	29.85	0.39
MnO	0.35	0.06	0.26	0.02	0.24	0.06
CaO	0.70	0.22	0.47	0.11	0.31	0.07
K ₂ O	0.12	0.02	0.12	0.02	0.14	0.03
Na ₂ O	0.03	0.06	0.06	0.11	0.01	0.03
TOTAL	98.61	0.29	97.36	0.39	97.20	0.96
<i>Cation proportions on the basis of twenty-three oxygens</i>						
Si	7.947	0.024	7.931	0.034	8.049	0.054
Al (IV)	0.019	0.012	0.024	0.016	0.001	0.005
Sum Z	7.967	0.022	7.956	0.030	8.050	0.053
Al (VI)	0.002	0.004	0.001	0.005	0.026	0.013
Ti	0.008	0.005	0.006	0.004	0.009	0.003
Cr	0.011	0.003	0.006	0.003	0.008	0.004
Fe	0.769	0.014	0.813	0.033	0.746	0.053
Mg	6.118	0.053	6.150	0.076	5.999	0.052
Mn	0.040	0.006	0.030	0.002	0.028	0.006
Sum Al (VI) to Mn	6.948	0.045	7.006	0.069	6.816	0.100
Ca	0.100	0.032	0.068	0.016	0.045	0.010
K	0.020	0.003	0.020	0.004	0.024	0.005
Na	0.008	0.015	0.016	0.028	0.003	0.007
Sum Ca + Alkalies	0.128	0.030	0.104	0.030	0.072	0.013
TOTAL	15.043	0.028	15.066	0.034	14.938	0.055
mg#	0.882		0.879		0.885	

* Total iron as FeO

mg# = $Mg/(Mg+Fe+Mn+Cr)$

Table AII.9, continued
 Microprobe analyses of anthophyllite

	AVERAGE N=18	STD DEV	AVERAGE N=46	STD DEV
Sample	103188-C1		110488-C2B	
Rock Type	Anth+Carb		Enst+Mag	
SiO ₂	58.33	0.43	59.12	0.56
TiO ₂	0.07	0.03	0.08	0.03
Al ₂ O ₃	0.22	0.11	0.36	0.20
Cr ₂ O ₃	0.07	0.03	0.06	0.04
FeO *	7.81	0.22	6.98	0.30
MgO	29.85	0.45	30.78	0.27
MnO	0.16	0.04	0.25	0.04
CaO	0.34	0.05	0.36	0.03
K ₂ O	0.13	0.03	0.12	0.02
Na ₂ O	0.07	0.12	0.00	0.00
TOTAL	97.04	0.75	98.11	0.69
<i>Cation proportions on the basis of twenty-three oxygens</i>				
Si	7.944	0.042	7.917	0.045
Al (IV)	0.027	0.022	0.055	0.033
Sum Z	7.970	0.033	7.972	0.024
Al (VI)	0.008	0.015	0.002	0.005
Ti	0.007	0.003	0.008	0.003
Cr	0.008	0.003	0.007	0.005
Fe	0.890	0.023	0.782	0.032
Mg	6.058	0.078	6.144	0.056
Mn	0.018	0.004	0.029	0.005
Sum Al (VI) to Mn	6.989	0.076	6.971	0.051
Ca	0.050	0.007	0.050	0.008
K	0.022	0.005	0.021	0.004
Na	0.018	0.030	0.000	0.000
Sum Ca + Alkalies	0.090	0.033	0.071	0.010
TOTAL	15.049	0.045	15.014	0.037
mg#	0.869		0.883	

* Total iron as FeO

mg# = Mg/(Mg+Fe+Mn+Cr)

Table All. 10

Microprobe analysis of chromium-magnetite

Sample Rock Type	AVERAGE	STD
	N=17	DEV
	1104-C2B Enst+Mag	
TiO ₂	1.16	0.08
Al ₂ O ₃	0.67	0.15
Cr ₂ O ₃	19.03	1.02
Fe ₂ O ₃ *	46.83	1.22
FeO*	27.24	1.05
MgO	2.68	0.64
MnO	0.00	0.00
NiO	0.45	0.10
ZnO	0.21	0.10
TOTAL	98.27	0.60

Cation proportions on the basis of thirty-two oxygens

Ti	0.265	0.018
Al	0.240	0.054
Cr	4.557	0.250
Fe ³⁺	10.672	0.251
Sum Ti to Fe ³⁺	15.735	0.018

Fe ²⁺	6.901	0.292
Mg	1.208	0.280
Mn	0.000	0.000
Ni	0.109	0.025
Zn	0.047	0.022
Sum +2 ions	8.265	0.018

TOTAL	24.000	1.193
-------	--------	-------

fe#	0.835
-----	-------

mg#	0.146
-----	-------

Al/({Fe ³⁺ }+Cr+Al)	0.016
--------------------------------	-------

Cr/({Fe ³⁺ }+Cr+Al)	0.295
--------------------------------	-------

Fe ³⁺ /({Fe ³⁺ }+Cr+Al)	0.690
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* FeO, Fe₂O₃ calculated assuming perfect stoichiometry

Fe# = Fe²⁺ / ({Fe²⁺} + Mg + Mn + Ni + Zn)

Mg# = Mg / ({Fe²⁺} + Mg + Mn + Ni + Zn)

APPENDIX III

GENERALIZED SAMPLE LOCATION MAP

Appendix III

Generalized Sample Location Map

Sample Numbers

MoDaYr-? Petrographic sample
-C? Whole-rock chemistry sample

Date of Sample
Locality Visit
and Collection

Daily Sequential
Station Number

Mo-Month
Da-Day
Yr-Year

-?
-C?

APPENDIX IV

**TABLE OF
MINERAL ASSEMBLAGES**

Appendix IV**Abbreviations used in Table AIV.1 - Table of Mineral Assemblages****Rock Type**

Enst+Mag	enstatite-magnesite rock
Anth+Carb	anthophyllite-carbonate rock
Contact	contact metasomatic rock

Mineral Names

Anth	anthophyllite
Carb	unidentified carbonate minerals, require further microprobe analysis to positively identify carbonate mineralogy.
Chl	chlorite
Dol	dolomite
En	enstatite
Fo	forsterite, I indicates that type I (coarse-grained) olivine is present II indicates that type II ("blebby") olivine is present.
Mag	magnesite
M-C	magnesio-cummingtonite
Op	opaque minerals
Tr	tremolite
Tc	talc, 1 indicates that primary (medium-grained) talc is present. 2 indicates that secondary (coarse-grained) talc is present.
Uk	unknown pleochroic green mineral, chromium-phlogopite?

Mineral Abundances

X	Mineral present
x	Mineral present in minor quantities (less than five percent).
P	Mineral probably present, requires microprobe analysis for positive identification.
p	Mineral probably present in minor quantities.

Table AIV.1
Table of Mineral Assemblages

Sample Number	Rock Type	Mineral Assemblage											
		Anth	Carb	Chl	Dol	En	Fo	Mag	M-C	Op	Tr	Tc	Uk
111587-1C	Enst+Mag			X	X	X	I	X	X	X	X	1	
111587-2	Enst+Mag	X	X	X	P	X		P	X	X		2	
011788-1	Enst+Mag	X	X	X	P	X		X	X	X	X	1	
032188-2B	Enst+Mag	X		X	X	X		X	P	X	X	1	
032188-3B	Enst+Mag	X	x	X	p	X		p	X	X	P	1&2	
032188-5A	Enst+Mag	X	x	X	p	X	I	p		X		1&2	
032188-5B	Contact	X	x	X	p			p	X	X	X	2	
050888-2A	Enst+Mag	x	X	X	P	X	II	P	p	X	p	2	
050888-3	Anth+Carb	X	X	X	P			P	X	X			
050888-8A	Enst+Mag	X	X	X	P	X		P	x	X			
101788-1A	Enst+Mag	X		X		X		X	x	X	x	1	
101788-1B	Enst+Mag	X		X		X	I	X		X		1	
103188-C1	Anth+Carb	X	X	X	P			P	X	X		1&2	X
103188-C2	Anth+Carb	X	X	X	X			X	X	X			X
110488-1	Enst+Mag	X	X	X	P	X		P	X	X			
110488-C2B	Enst+Mag	X		X	X	X	I	X	X	X	X	2	
110488-C4	Anth+Carb	X		X	X			p	x	X			
110488-C5	Enst+Mag	x	x	X		X		p		X		1&2	
110488-C8	Anth+Carb	X	X	X	P			P	x	X		1&2	
110488-C9	Anth+Carb	X	X	X	P			P	x	X		1	X
060488-2B	Enst+Mag	X	X	X	P	X	I	P		X			

END

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FIN