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**Origin and Evolution of Peraluminous Granite and Rare-Element
Pegmatite in the Dryden Area, Superior Province of Northwestern
Ontario**

by

Frederick William Breaks, B.Sc., M.Sc.

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

Department of Earth Sciences
Carleton University
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Ottawa, Ontario, Canada

July 20, 1989

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thesis is available upon request from
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To my parents, Ann and Frederick William, Sr., who provided much encouragement, especially at a time when it was needed the most.

"Pegmatites, those widely distributed rocks that are at once both fascinating and frustrating"

R.H. Jahns (1982)

FRONTISPIECE View of Gullwing Lake looking towards Mica Point
from the summit of the Gullwing Lake pegmatite granite.

The undersigned hereby recommend to the Faculty of Graduate
Studies and Research acceptance of this thesis, submitted by
Frederick W. Breaks, in partial fulfillment of the
requirements for the degree of Doctor of Philosophy

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ABSTRACT

Rare-element pegmatites in the Superior Province of northwestern Ontario occur at the boundaries between high grade gneissic belts [English River (ERS) and Winnipeg River (WRS) Subprovinces] and low to medium grade metavolcanic/sedimentary belts (Uchi and Wabigoon Subprovinces).

The Dryden pegmatite field has been studied by geological mapping (1:63,360 to 1:60), petrography and mineral and rock geochemistry. It comprises two distinct rare-element pegmatite clusters, the Mavis Lake Group (MLG) and Gullwing - Tot Lakes Group (GTC). The MLG is genetically linked with the peraluminous Ghost Lake Batholith (GLB), a late-tectonic, composite plutonic complex emplaced discordantly to regional metamorphic zones. The eastern lobe of the GLB contains sporadic beryl, columbite and cassiterite. The adjacent MLG reveals systematic rare-element mineral zoning (beryl -> columbite+beryl -> tantalite+beryl+spodumene) and an increase in albite-rich replacement features with increasing distance from the batholith.

The GLG exhibits more complex mineral zoning: columbite -> molybdenite + beryl + columbite + spodumene -> beryl + columbite + tantalite + pollucite + spodumene. The source of the GTG is less certain. In the Gullwing Lake dyke, a spodumene- and columbite-bearing pegmatite core is enveloped by albitized muscovite-biotite pegmatitic granite, implying a genetic link.

A five-stage genetic model is proposed for the Dryden Pegmatite Field. Rare elements, boron and fluorine were concentrated during the development of sedimentary basins in the ERS and WRS. High grade Abukuma-regional metamorphism at 2.68 Ga led to fluid-absent breakdown of muscovite and biotite producing S-type peraluminous granitic melt with concentrated rare elements and volatiles; restite unmixing and crystal-melt fractionation in the GLB led to vertical compositional zonation characterized by repletion of Na₂O, Rb, Cs, Li, B, F, MnO, Ga and H₂O and strong depletion of REE, Sr, Ba and Zr.

Fluxing components (e.g B₂O₃, F, Li₂O) inherited from the metasedimentary protolith depressed liquidus and solidus temperatures and shifted residual magma compositions towards increased Na/K as magma ascended to higher crustal levels, being ultimately emplaced in low and medium grade rocks of the boundary zones as <2.56 Ga chemically specialized granites.

Rare element concentration was increased by late- to post-magmatic albitization and greisenization. This resulted in muscovitization of a Li-F-Fe-enriched biotite and primary muscovite which allowed release of Li into an exsolved fluid phase enriched in Li, Rb, Cs, B and F. A holmquistite alteration zone in mafic metavolcanics may represent the missing fractionation link between the GLB (198 ppm Li) and the nearest spodumene pegmatite (8500 ppm).

The post-magmatic stage of rare-element pegmatite evolution led to the development of aureoles characterized by elevated concentrations of rare elements, F and B.

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Chapter One

INTRODUCTION

BACKGROUND AND PURPOSE

Rare-element pegmatites have been known in the Superior Province, northwestern Ontario since the mid-1950's when their discovery was spurred by a sharp increase in lithium demand for atomic energy programs in the United States (Milne, 1962).

In recent years, increasing exploration and academic interest has been directed towards these rocks; they are unique or major repositories of Cs, Ga, Rb, Ta, Be, and Nb, and still represent the most important source of Li.

The major problem addressed by this study is why these Archean rare-element pegmatites in Ontario are principally confined to certain subprovince boundaries, i.e. those between metavolcanic-rich subprovinces (Uchi and Wabigoon) and high grade gneissic belts (English River and Winnipeg River Subprovinces). This distribution characteristic of rare-element pegmatites emerged during a series of 1:63,360 scale, helicopter-supported, geological reconnaissance mapping and metallogenic studies coordinated by F.W. Breaks and W.D. Bond during the 1970's (Breaks et al. 1978). This project covered 44,000 km² of the essentially unmapped English River Subprovince (ERS) and the Winnipeg River Subprovince (WRS). A further lithologic-metamorphic control on pegmatite distribution recognized during this reconnaissance work was their restriction to medium grade supracrustal rocks that are proximal to extensive zones of high to granulite grade,

migmatized clastic metasediments and spatially associated peraluminous granitoid rocks.

This restricted distribution led Breaks et al. (1978) to propose that the rare-element pegmatites are chemically evolved derivatives from S-type granite masses (the latter mapped as homogeneous and inhomogeneous diatexite). In order to comprehensively investigate this model, a geologically favourable study area was necessary. The Dryden area was selected to test the S-type granite-rare-element pegmatite connection because it possesses the following attributes:

- 1) a zoned, rare-element pegmatite swarm within a subprovince boundary zone situated adjacent to a potential parent granite mass;
- 2) a well-defined regional metamorphic facies series, preserved in the boundary zone clastic metasediments, which ranges from low to high grade;
- 3) reasonably well-understood regional temporal relations and chemical characteristics of the various granitoid suites;
- 4) a general absence of the complicating influence of major shear zones;
- 5) good access.

In many pegmatite districts the petrogenetic processes responsible for the elevated concentration of rare elements are difficult or impossible to determine (Cerny et al. 1982b), because the pegmatites are too far from their potential sources or the parental granite is unexposed. The Dryden area, in contrast, provides an excellent opportunity to unravel such rare-element-concentrating

mechanisms as all of the important geological components are excellently exposed and in proximity.

LOCATION, ACCESS AND PHYSIOGRAPHY

The study area (Figures 1 and 2) is bounded by Latitudes 49°44' and 50°05', and Longitudes 92°20' and 93°13', encompassing an area of approximately 2,000 km². Most parts of the area are accessible by Highways 11, 119, 594, and 604 and intersecting township access roads. The village of Dryden is the major centre of population and commerce in the area. Wabigoon, Eagle River and Oxdrift represent much smaller settlements.

The area is one of subdued relief typical of Canadian Shield terrane in northwestern Ontario. Overall relief is 150 m which may locally reach 40 m as in vicinity of the Gullwing Lake Pegmatite and near the eastern end of Mavis Lake. The most prominent physiographic feature is Wabigoon Lake, situated along the southern limits of the study area. Drainage in the area is effected by the English River system, locally through its tributary, the Wabigoon River.

Outcrops are generally distributed homogeneously throughout the area, although the overall percentage is only about five percent owing to a pervasive thin mantle of Pleistocene varved clay deposits which originated from Lake Agassiz (Sado and Carswell 1987). The Hartman end moraine, a major glacial landform, extends through the study area and lies adjacent to the Mavis Lake pegmatite group.

Figure 1. Location Map for the Study Area.

PREVIOUS WORK

Geological investigations in the study area have a fifty year history. A summary of this work, as it applies to Li, Cs, Ta, Mo, W, Sn, U and Th mineralization, peraluminous granitoid rocks, and their host rocks will now be presented.

REGIONAL GEOLOGICAL SURVEYS

Early geological investigations were undertaken by Moorhouse (1939), Burwash (1939), Satterly (1941) and Harding (1950).

The work of Moorhouse (1939) represented the first detailed mapping (scale 1:63,360) of the southwestern portion of the Ghost Lake Batholith and its host rocks (Figure 2).

The first record of lithophile mineralization in the region was reported by Burwash (1939) who briefly described cassiterite associated with tourmaline pegmatite on Partridge (Portage ?) Bay of Eagle Lake, 10 km west of the Ghost Lake Batholith.

Satterly (1941) provided mapping coverage (1:63,360 scale) over the east part of the Ghost Lake Batholith and was the first to document tourmaline replacement in metasediments adjacent to pegmatite dykes. Satterly (1941, p.55) also mentioned a beryl occurrence in Zealand township which probably coincides with the Taylor Beryl Occurrence of this thesis.

Harding (1950) mapped an area centred at Tot Lake at a scale of 1:63,360. Although the Gullwing Lake pegmatitic granite dyke was approximately delineated by Harding's survey, rare-element mineralization was undetected.

Page and Christie (1980) undertook detailed mapping (1:15,840 scale) in Webb and Drope Townships, an area covering rare-element pegmatites of the Gullwing - Tot Lakes Group. This survey included a cursory examination of the Tot Lake rare-element pegmatite and Lateral Lake molybdenite pegmatites.

Ucakuwun (1981) completed a cursory regional examination of various granitoid masses and rare-element pegmatites and focussed upon mineralogical and chemical work. His study, however, did not include the Gullwing, Mica Point, Coates, Taylor No.2 beryl pegmatite, Fairservice East and Northeast Zone spodumene pegmatites that are dealt within this thesis.

MINERAL EXPLORATION HISTORY

Mineral exploration for lithophile elements in the region has mainly been carried out during two distinct time intervals, for three commodities: 1955-1964 for lithium and the late 1960s to 1984 for tantalum and tungsten.

All exploration information in this section was compiled from the Assessment Files Research Office, Ontario Geological Survey, Toronto.

Discovery of spodumene pegmatites near Mavis Lake, Brownridge Township, in 1956 led to investigation by Lun-Echo Gold Mines Limited. A program involving trenching and forty diamond-drill holes totaling 1,968 m established lithium mineralization over a strike length of 670 m and width varying between 1.5 and 11 m. The best intersection recorded 2.76 percent Li_2O over 2.9 m.

During 1964, A. Kozowy and A. Leduchowski discovered a spodumene-pollucite-bearing, complex pegmatite near Tot Lake. J. Donner investigated the mineralization by geological mapping, a magnetometer survey, and diamond drilling (4 holes totalling 732 feet). In 1973, A. Kozowy did a limited amount of trenching.

Columbite mineralization in the Gullwing Lake and Mica Point Pegmatites was discovered in the 1960's by A. Kozowy, and who informed the writer of its possible significance in 1981.

In the late 1960s, tungsten mineralization was discovered by D. Petrunka near Sharpe Lake in Brownridge Township and initial evaluation was undertaken by Noranda Mines Limited (Blackburn and Hailstone, 1983). From 1981 to 1984, a 32-claim block was assessed under option agreement by Sanmine Exploration Incorporated. This company mainly focussed upon surface evaluation and limited diamond drilling.

Interest in the tantalum potential of rare-element pegmatites of the Dryden area commenced in 1979 on the part of Selco Incorporated, who optioned the Mavis Lake spodumene pegmatite property from Mr. R. Fairservice and carried out geological mapping, diamond drilling, and a lithium lithogeochemical survey. In the same year, the Tantalum Mining Corporation of Canada Limited under the direction of D.L. Trueman acquired property containing the Tot Lake complex rare-element pegmatite and undertook a limited diamond drill program. In 1980, property containing the Gullwing Lake complex rare-element pegmatite was optioned from A. Kozowy by Selco Incorporated and Dome Exploration

Canada Limited. Both Selco Incorporated - Dome Exploration Canada limited options were terminated in 1982.

Tantalum Mining Corporation of Canada briefly optioned the Fairservice property in 1982 and terminated work after a magnetometer survey.

PRESENT WORK

The geological mapping component of the present study was one of progressively decreasing mapping scale. The initial phase of work involved 1:63,360 reconnaissance work (Breaks et al. 1976a, b), followed by 1:31,360 scale, more detailed mapping encompassing 520 km² of the Dryden-Eagle River area (Breaks and Kuehner 1984). The latter study delineated and named the Ghost Lake Batholith. The next stage of geological mapping by the writer occurred at the claim-group scale. Detailed mapping of the areas containing the Gullwing Lake pegmatitic granite dyke and Fairservice Property spodumene pegmatites (North and East Zones) was conducted at scales of 1:1400 and 1:4800, respectively. This phase of mapping utilized grids with 200 foot line spacings cut by Selco Incorporated in 1980-1981. The final phase of detailed mapping (1:60 and 1:120 scale) was restricted to the Gullwing Lake, Tot Lake, Taylor and the Fairservice (Mavis Lake) No.1 pegmatites, employing a grid with 5 foot line spacing.

In total about ten months of field residence time was spent by the writer in the Dryden area. In this study, rare-element pegmatites of economic interest were stripped, washed with a Wajax Mark III fire control-type water pump and hose system, and bleached with Javex by the author and assistants when employed by Selco Incorporated in 1981.

This procedure proved effective in revealing petrographic and mineralogical features and controls on rare element mineralization; this also greatly facilitated mineral and bulk sample selection.

Size of the rock samples selected for chemistry varied considerably depending upon grain size of the units. For most medium- to coarse-grained units of the Ghost Lake Batholith sample weight ranged between 4 and 6 kg. Pegmatitic granite and rare-element pegmatite samples were generally more voluminous (except for aplite and fine-grained leucogranite) and varied between 10 and 25 kg. All weathered surfaces were removed by a diamond saw. A Stihl masonry saw with 3/16" thick by 10" diameter carborundum blades allowed bulk channel sampling of specific units within zoned pegmatites and otherwise inaccessible mineral samples.

Modal analyses were mainly undertaken on slab surfaces stained for K-feldspar and plagioclase and aided by a binocular microscope (3X to 20X magnification). At least 500 points were tallied per specimen, utilizing mylar grids with 5 and 6 mm spacings.

Mineral identification was undertaken by petrographic microscope examination and X-ray diffraction at the Geoscience Laboratories of the Ontario Geological Survey. Rock and mineral analyses were also performed in house; analytical methods, precision and accuracy are given in Table A-1.

The minerals analyzed by wet chemical means (K-feldspar, muscovite, spodumene and beryl) were sufficiently coarse and homogeneous to allow rapid mesoscopic elimination of most impurities.

This was followed by sectioning of mineral samples (after removal of weathered surfaces) with a water-cooled 4" diameter saw; these sections were inspected for obvious alteration such as albite replacement in K-feldspar and secondary mica in beryl and spodumene. In the case of K-feldspar, homogeneity of the sample was checked with HF etching and feldspar staining.

The resulting coarsely crushed 20 to 50 gm mineral separate was checked with a 1.25X to 10X binocular microscope and further impurities were removed with metal tongs. Many samples were also examined for impurities in thin section. Small quantities of fine-grained muscovite and lepidolite along cleavages of K-feldspar and spodumene may possibly have been missed by this separation procedure as suggested by high trace levels of Li in some samples (e.g. Appendix 18, analysis KF4 in Pegmatite 9). Purity of most mineral separates was estimated to be around 95 percent.

All sample and photographic localities can be found in Figures 15, 17, 18, 27, 28 29, 31 and 38.

TERMINOLOGY

Important terminology used in this thesis is defined and explained in this section.

ROCK CLASSIFICATION

Plutonic rocks were classified according to the system proposed by the IUGS Subcommittee on the Systematics of Igneous Rocks (Strecke'sen 1976).

Characteristic varietal minerals exceeding one percent were added to rock names in order of increasing relative abundance; for example, a muscovite-biotite trondhjemite contains more biotite than muscovite.

METAMORPHIC GRADE

Division of metamorphic grade follows that proposed by Winkler (1979). Wherever known, metamorphic mineral assemblages are listed in order of increasing relative abundance.

MIGMATITE TERMINOLOGY

Metasedimentary migmatites are classified according to the scheme of Mehnert (1968) and Brown (1973) which was adapted by Breaks et al. (1978) for use in the English River and Winnipeg River subprovinces. This classification, with general petrologic attributes of the major migmatite stages, is presented in Table 1. The following terms are particularly important as directly quoted from Brown (1973, p.374 and 375):

Metatexis

This term refers to "the process of segregation (usually of quartz and feldspar) by metamorphic differentiation and partial fusion. A metatexite is a rock produced by metatexis and in which migmatitic banding is evident".

Diatexis

This process is defined by "high grade anatexis in which fusion may be complete. A diatexite is a rock produced by diatexis and in which there is no continuous migmatitic banding".

Table 1. Classification and General Features of Metasedimentary Migmatites from the English River Subprovince. Terminology after Mehnert (1968), Brown (1973) and Breaks et al. (1978).

Migmatitic Stage	Leucosome:Leucosome + Paleosome Ratio	Diagnostic Migmatitic	General Processes	General Field Features
Proto-Metatexite	<0.1	Locally stromatic. Sedimentary bedding or laminae may be preserved.	Incipient, 'selective' anatexis. Metamorphic differentiation may be important.	Characterized by intercalated wacke and porphyroblastic pelite in which podiform and lenseoid granite leucosomes are confined to pelitic horizons.
Metatexite	0.1 to 0.6	Stromatic and/or phlebitic.	Local to moderate degree of anatexis	Leucosome developed "in situ"; biotite-rich melanosomes are distributed along the paleosome-leucosome interface. Leucosome may also be injected along the foliation or bedding.
Inhomogeneous Diatexite	0.6 to 0.9	Schollen structure especially characteristic. Locally schlieritic or nebulitic.	Anatexis relatively extensive; granitic melt essentially autochthonous.	High concentrations of inclusions of metatexite, unmigmatized wacke and restite clumps and schlieren.
Homogeneous Diatexite	>0.9	Typically massive and homogeneous. Local schollen schlieritic or nebulitic structures.	Segregation of leucosome from paleosome and restite into magmatic masses which were emplaced above the fusion zone.	Commonly holo-leucocratic, hypidiomorphic-granular and seriate. Rare biotite-rich variants (Cl 15 to 30) maybe present.

Leucosome

This refers to the leucocratic component of a metatexite or diatexite migmatite with a plutonic appearance.

Melanosome

The mafic-rich, dark coloured component of a migmatite occurring as selvages in metatexite or irregular clots in diatexite is called the melanosome or restite.

Paleosome

This constitutes the parent rock of leucosome and melanosome components; also called the protolith.

GRANITIC TEXTURESAllotriomorphic-Granular

A plutonic rock textural term in which anhedral minerals predominate over euhedral and subhedral ones.

Hypidiomorphic-granular

A plutonic rock textural term which is characterized by most minerals having a subhedral habit.

PEGMATITE GRAIN SIZE AND LITHOLOGY

The grain size of barren and rare-element pegmatites is defined according to the convention of Cameron et al. (1949, p.16) below. In the description of pegmatitic granite intrusions the grain size terms

adopted for granitic rocks will be used as indicated below.

<u>Grain Size Term</u>	<u>Diameter</u>
fine-	<2.5 cm
medium-	2.5 - 10 cm
coarse-	10 - 30.5 cm
very coarse-	>30.5 cm

Pegmatitic Granite

The term was, to the writer's knowledge, first used by Cerny et al. (1981, p.50-73) in reference to heterogeneous granitic intrusions (typical 0.5 to 10 km²) which are characterized by a complex intermingling of a maximum of six distinctive units (specified below) commonly varying in grain size (fine- to very coarse-grained) and mineralogy.

The term "granite" in pegmatitic granite and in some of its divided units (pegmatitic leucogranite and fine-grained leucogranite) is used in the broad compositional sense. Specific granite compositions in this study are given only where feldspar staining work has been undertaken.

Internal units of pegmatitic granite masses were classified according to the criteria of Cerny et al. (1981, p.50 and p.61) which were modified by Cerny and Meintzer (1988). All minerals of assemblages placed in parentheses are accessory phases; these minerals are commonly distributed inhomogeneously within a given rock unit.

The internal constitution of pegmatitic granite intrusions has been mainly investigated by Cerny et al. (1981) and Cerny and Meintzer (1988). Their classification is therefore useful in petrographically

dividing such intrusions and providing a basis for regional comparison and petrochemical and mineralogical investigations.

Pegmatitic Leucogranite: a rock that tends to be the predominant unit in pegmatitic granite masses and is distinguished by medium- to very coarse-grained, graphic quartz-K-feldspar megacrysts contained within a medium- to coarse-grained matrix of muscovite, quartz and albite and accessory garnet, tourmaline as exemplified in Plate 13a.

Fine-grained Leucogranite: a rock that is fine-grained, equigranular and massive which is composed of quartz, albite and microcline with accessory garnet, tourmaline, and muscovite as exemplified in Plates 5e and 6a.

Sodic Aplite: a rock that is typically finer-grained than fine-grained leucogranite, sodic aplite is commonly layered with potassic pegmatite (Plates 6d and 7d). Most of the rock is composed of albite and quartz; accessory muscovite, phosphates and tourmaline may be present.

Potassic pegmatite: invariably the youngest pegmatitic granite unit, this medium- to coarse-grained rock is characterized by euhedral to subhedral blocky K-feldspar crystals, commonly in contact with quartz-rich masses (Plates 6d and 7d). Medium-grained muscovite and tourmaline are common varietal minerals. Accessories include biotite, garnet, beryl and columbite.

Biotite granite: typically subordinate or absent in most pegmatitic granite intrusions. This fine- to coarse-grained rock locally containing K-feldspar megacrysts, is indistinguishable from

other granites if associated pegmatitic granite units are absent (Cerny and Meintzer, 1988, p. 178).

Rare-element-enriched pegmatite: refers to small pods of pegmatite containing rare-element minerals (spodumene, petalite, lepidolite, beryl, tourmaline, columbite-tantalite group, microlite, wodginite, cassiterite, triphylite-lithiophilite and amblygonite-montebrasite) may occur in some pegmatitic granite bodies according to Cerny and Meintzer (1988, p.180). However as these workers do not give chemical or modal criteria to distinguish "rare-element-enriched pegmatite" from other pegmatitic granite units which may contain rare-element minerals (e.g. beryl-bearing potassic pegmatite at the Dryden Airport; Chapter 3) this term will not be used.

Barren pegmatite

This is a commonly used term in the literature although it has not, to the writer's knowledge, been properly defined. Therefore in this study it will be used for pegmatites whose primary assemblage does not contain any of the rare elements in economic concentrations. Trace levels of <100 ppm of these elements can occur in barren pegmatites particularly those which are close to rare-element pegmatite swarms.

Exocontact Pegmatite

A term used by the writer in reference to pegmatite dykes and small masses which lie in host rocks close to the external contact of a granitic pluton. The pegmatites must have physical, mineralogical or chemical evidence supporting a link with the granite. The term

"exocontact" although common in the literature is not defined by Bates and Jackson (1987).

PEGMATITE GROUPING TERMINOLOGY

Cerny et al. (1981, p.1-2), in their classification of southeastern Manitoba pegmatites, modified the terminology of Kuzmenko (1977). Cerny (1982 a, p.8-9) suggested, in a provisional tone, the following hierarchy: pegmatite province - pegmatite belt - pegmatite district - pegmatite group - pegmatite series. Those terms as used in this study are given below, as directly quoted from Cerny 1982, p.8-9):

Pegmatite District - part of a pegmatite province containing several associated pegmatite fields, separated from others either territorially or geologically.

Pegmatite Field - a large-scale association of diversified granitoids and related pegmatites within a pegmatite province or belt, frequently hosted by a single lithologic unit, generated during a single tectonomagmatic stage and spatially separated from other fields (e.g. Cat Lake-Winnipeg River and Wekusko Lake pegmatite fields in Manitoba).

Pegmatite Group - a swarm of pegmatite bodies, fairly homogeneous to distinctly fractionated in composition, related to a single granitic intrusion within a pegmatite district or field (e.g. Greer Lake group in the Winnipeg River district; Redout-Ross Lakes pegmatite group in the Slave province, Hutchinson 1955)".

GREISEN TERMINOLOGY

Two terms from Scherba (1970) were used to describe certain metasomatic units associated with the GLB:

Autometamorphic Greisen

This is an autometasomatic rock produced within a granite pluton by interaction of late- to post-magmatic fluids with previously consolidated granitic units.

Exogreisen

This is a metasomatic rock produced by interaction of late- to post-magmatic fluids derived from a granitic pluton with adjacent host rocks.

MINERALOGICAL TERMINOLOGY

K-feldspar

In this study, K-feldspars were mainly used to examine trace metal chemical trends (Rb, Cs, Li, Ga and Sr) in the GLB and rare-element pegmatites. As no X-Ray diffraction or optical work was undertaken by the author to determine the mineral species (maximum microcline, intermediate microcline and orthoclase), K-feldspar is used only in a general sense.

Albite

Two types of albite were mesoscopically recognized in pegmatites of the study area:

- a) platy albite, characterized by pronounced elongation parallel to (C10)
- b) equant albite.

Platy albite, according to Smith (1974, p.436-437) has been incorrectly described as having the cleavelandite habit in many pegmatite studies. Unless platy albite is specifically mentioned, all albite in the descriptive sections will be of the equant type.

Columbite-Tantalite Group

Where minerals of this group were encountered in the field and not subsequently specified by laboratory techniques they are referred to as columbite-tantalite. Otherwise specific names will appear in the text (columbite or tantalite) coupled with reference to electron-microprobe analyses given in Appendices 39, 40 and 41.

Chapter Two

REGIONAL GEOLOGICAL SETTING OF THE DRYDEN PEGMATITE FIELD

The purpose of this chapter is to briefly provide the regional geological, stratigraphic, structural and metamorphic setting for rare-element pegmatites constituting the Dryden Pegmatite Field.

REGIONAL GEOLOGY

Rare-element pegmatites in the Dryden area of northwestern Ontario were first described by Mulligan (1965, p.63-64) who termed their localization the "Dryden Pegmatite Field". Regionally, this pegmatite field is mostly situated within mafic metavolcanic host rocks of the Wabigoon subprovince, situated near its interdigitating boundary with metasediments of the Wernipeg River Subprovince (WRS) as defined by Breaks et al. (1978). The position of this subprovince boundary, shown in Figures 2 and 4, is controversial and is largely dependent upon the lithologic, structural, metamorphic and geophysical criteria utilized (Breaks et al., 1985, p.158; Beakhouse 1985, p.73-75).

Regional zonation of pegmatites within part of the Dryden Pegmatite Field was first recognized by Breaks (1980, p.6) as outlined in Table 2 and is described in detail within an ensuing section. The original assessment was subsequently modified (Breaks and Kuehner 1984) in order to accommodate new data and, as indicated in Table 2, the pegmatite field was divided into two distinct groups ten kilometres apart:

Figure 2. Distribution of Rare-Element Pegmatites (triangles) in the Dryden Field in relation to Regional Geology. Each triangle Represents either a Single Pegmatite or a Pegmatite Swarm.

Table 2. Petrological and Chemical Features of Rare-Element Pegmatites from the Dryden Field.

AREA/OCCURRENCE	PEGMATITE STRUCTURE	CHARACTERISTIC PEGMATITE MINERAL ASSEMBLAGE	GEOCHEMICAL ASSOCIATION	EXTENT OF LATE STAGE REPLACEMENT ALBITIZATION
I. Davis Lake Pegmatite Group				
Dryden Airport Occurrence, Zealand Township	Unzoned, internal, usually barren potassic pegmatite	Garnet - Muscovite - Tourmaline - Albite - Quartz - Blocky K-feldspar - (Line-green Beryl)	8-(Be)	Absent
Contact Beryl Occurrence, Zealand Township	Autometamorphic greisen and exogreisen vein system at Ghost Lake Batholith contact	Beryl - Tourmaline - Biotite - Muscovite - Albite - Holmquistite - Actinolite - Pyrite - (Cassiterite-Chrysoberyl-Tren. lite)	Be-B-F-Li (Sn, Rb)	Local albitization of -4, -6 and -7 Albite exocontact dykes with Tourmaline and Beryl also present
Taylor Occurrence, Concession VII, Zealand Township	Unzoned potassic pegmatites developed marginal to Ghost Lake Batholith	Beryl - Tourmaline - Albite - Quartz - Graphitic K-feldspar - Blocky K-feldspar - (Columbite - Garnet - Muscovite - Green Apatite)	8-Be-(Sn-Nb-Ta)	Incipient. Small pods of fine-grained albite (Chromium Mica - Quartz - Albite)
Petrunka Occurrence, Brownridge and Zealand Townships	Unzoned, tourmaline-enriched fracture- and pillow selvage - controlled replacement in Brownridge mafic metavolcanics	Quartz - Plagioclase - Biotite - Scheelite - Tourmaline - (Pyrite - Holmquistite - Fluorite?)	W-B-F- (Li-Sn)	Not observed
Fairservice Property, Brownridge Township North and South Zone	Unzoned to crudely zoned external pegmatites. Contains randomly oriented, green primary spodumene phenocrysts (Type 1 spodumene of Heinrich, 1975)	Beryl - Green muscovite - Albite - Spodumene - Blocky K-feldspar - Quartz - (Blue Apatite - Garnet - Tourmaline - Tantalite)	Li-Be-Ta-Nb-B- (Sn-Rb)	More extensive albite replacement units: 10 to 60 percent of pegmatite (- Garnet - Tantalite - White Beryl - Green Muscovite - Quartz-Platy Albite)
East Zone	Unzoned external pegmatites	Primary mineralogy largely replaced. Remnants of Blocky K-feldspar, Spodumene, and Quartz still discernible	Ta-Nb-Be-(Li)	Pervasive albitization. Most of pegmatite replaced by White Beryl - Tantalite - Green Muscovite - Quartz - Platy Albite assemblage

Table 2, continued

II. Gullwing Lake - Tot Lake Pegmatite Group

Mica Point Pegmatitic Granite, Drope Township	Crudely zoned, external potassic pegmatite	Muscovite - Biotite - Albite - Quartz - K-feldspar - (Molybdenite - Columbite)	Nb>Ta - (Mo)	Incipient
Gullwing Lake Pegmatitic Granite, Webb Township	Complex external pegmatitic granite dyke with subordinate core zones of rare-element pegmatite	Spodumene - Muscovite - Albite - Quartz - Blocky K-feldspar and Albite - Green Muscovite - Quartz - Blocky K-feldspar. Contains green, type 1 spodumene of Heinrich (1975)	Li-Nb>Ta-(Mo-Be)	More extensive albitization developed in two stages: 1) early medium-grained, equigranular Albite - Biotite - Columbite - Muscovite - Quartz) 2) later, fine-grained Lithian Muscovite - Albite Complex (Microcline-Columbite - Beryl - Garnet - Muscovite - Sphalerite Platy Albite)
Coates Pegmatite Webb Township	Unzoned, external potassic pegmatite	Muscovite - Albite - Quartz - Graphite K-feldspar - (Molybdenite - Pyrite - Garnet - Samarskite - Sphalerite Chalcocopyrite - Beryl)	Mo-(Nb>Ta)	Incipient replacement by fine-grained albite along edges of K-feldspar. Most molybdenite and samarskite associated with albitized areas.
Pegmatite 215 Metres Southwest of Coates Pegmatite, Webb Township	Unzoned, external potassic pegmatite	Muscovite - Albite - Quartz - Blocky K-feldspar - (Columbite-Tantalite - Garnet - Molybdenite)	Nb>Ta - (Mo)	Incipient, film-like albitization and associated Columbite in coarse blocky K-feldspar.
Tot Lake Pegmatite, Webb Township	Complex external pegmatite. Contains pink, Type 2 spodumene of Heinrich (1975)	Pink Spodumene - Pollucite - Quartz - Albite - Blocky K-feldspar - (Fluorapatite - Elbaite - Garnet)	Li-Cs-Rb-Ta>Nb- (Be-F-P)	Extensive albitization developed in two stages: 1) sodic apatite (Blue Apatite - - Tourmaline - Quartz - Albite) 2) Lithian Muscovite-Albite Complex (-White Beryl) - Garnet - Columbite - Tantalite - Sericite - Lithian Muscovite-Quartz-Platy Albite)

Accessory minerals and subordinate elements are placed in parentheses.

1. Mavis Lake Pegmatite Group and
2. Gullwing - Tot Lakes Pegmatite Group.

REGIONAL STRATIGRAPHY

Stratigraphic relationships in the lithologically diverse supracrustal succession exposed near the subprovince boundary between Eagle River and Sioux Lookout have been investigated by numerous workers including Pettijohn (1930, 1934, 1935, 1936, 1939), Hurst (1933), Moorhouse (1939), Satterly (1941), Johnston (1969, 1972), Walker and Pettijohn (1971), Turner and Walker (1973), Trowell et al. (1981) and Blackburn et al. (1985). Distribution of the important supracrustal units is portrayed in Figure 3. Pettijohn (1939) and Satterly (1941) proposed that the Dryden area is underlain by a homoclinal succession composed five lithostratigraphic units (Table 3). The succession proposed by Pettijohn (1939) was deduced to face south to southeast and to be terminated against the Wabigoon Volcanics (Upper Wabigoon Volcanics of Blackburn et al. 1985) by the Wabigoon Fault (see Structural Geology section of this chapter).

This succession may, however, be overly simplistic. Blackburn in Trowell et al. (1977), Blackburn (1979b), and Blackburn et al. (1982) have observed numerous exposures of grade wacke-pelitic clastic metasediments which exhibit top reversals and isoclinal fold closures. Blackburn et al. (1982, p.56) suggested that the sequence is tightly to isoclinally folded about subhorizontal axes, at least in the region near the Wabigoon Fault.

As a corollary, it is possible that the narrow unit of Thunder River Volcanics, which closely follows the northern shore of Wabigoon

Lake between the east boundary of Zealand Township and the town of Dryden (Figure 10, back pocket), is a macroscopic equivalent of these folds, in which case the distinction of Zealand Sediments from the Thunder Lake Sediments is rendered tenuous.

The Mavis Lake rare-element pegmatite group is emplaced almost exclusively within mafic metavolcanics of the Central Volcanic Belt (Figure 3). Stratigraphic affiliation of the mafic metavolcanic rocks hosting most of the Gullwing - Tot Lakes rare-element pegmatite group is, however, less certain. The host rocks possibly belong to the Northern Volcanic Belt of Trowell et al. (1980), because an apparently contiguous intermediate pyroclastic unit to the north provides a physical connection with this belt (Figure 20). The nearest mafic metavolcanic rocks to the south of this pegmatite group, on the other hand, are situated beyond the Abram Group metasediments which have been interpreted to lie in faulted contact with the Central Volcanic Belt (Trowell et al. 1980).

PLUTONIC ROCKS

The Winnipeg River Subprovince (WRS), in proximity to the study area, consists dominantly of four suites of granitoid rocks as classified by field relations, petrological and chemical attributes and radiometric age. These major suites have been summarized in Table 4 and have been described in detail by Breaks et al. (1978) and Breaks and Bond (in press). It should be noted that most of the rare-element and U-Th mineralization of the region is associated with only two of the granitoid suites, i.e. the peraluminous granitoid and potassic

Figure 3. Regional Stratigraphic Units in the Vicinity of the Dryden Pegmatite Field (modified after Trowell et al. 1980)

Table 3. Stratigraphic Units of the Dryden Area and Lateral Equivalents in the Sioux Lookout Area.

<u>DRYDEN AREA</u>		<u>SIOUX LOOKOUT AREA</u>	
<u>Pettijohn (1937)</u>	<u>Turner and Walker (1973)</u>	<u>Irowell, Blackburn and Edwards (1980)</u>	<u>U/Pb Zircon Age²</u>
Wabigoon Volcanics Wabigoon fault	Southern Volcanic Belt	Southern Volcanic Belt - Upper Wabigoon Volcanics Wabigoon fault	
Zealand Sediments ¹ Thunder River Volcanics Thunder Lake Sediments	Southern Sedimentary Belt Minnitaki Group	Southern Sedimentary Belt Minnitaki Group	2713±2 Ma clast age
Brownridge Volcanics	Central Volcanic Belt	Central Volcanic Belt Neepawa Group	2733±1 Ma
		<hr/> Little Vermilion fault <hr/>	
Brownridge Sediments	Northern Sedimentary Belt Abram Group Little Vermilion Formation Daredevil Formation Ament Bay Formation	Northern Sedimentary Belt Abram Group Little Vermilion Formation Daredevil Formation Ament Bay Formation	
	Northern Volcanic Belt (local intercalation of Patara Sediments)	Northern Volcanic Belt Patara Group (mixed volcanics and metasediments) Botham Bay Volcanics	2903±16 and 2698±4 Ma clast ages
			2703±3 Ma quartz porphyry intrusion

1. includes the correlative Warclub Sediments southwest of Eagle Lake (Trowell et al. 1980)
2. from Davis and Trowell (1982) and Davis et al. (1988).

plutonic suites. The characteristics of the gneissic and sodic plutonic suites are summarized in Table 4.

Peraluminous Granitoid Suite

This group of granitoid rocks was initially mapped as homogeneous and inhomogeneous diatexites (Table 1) and later reclassified as the peraluminous granitoid suite (Plate 1a). This suite is believed to have been derived by regionally extensive anatexis of clastic metasediments, mainly in the English River subprovince (ERS) and locally in the WRS under low pressure, high grade metamorphic conditions during the Kenoran thermo-tectonic event (Breaks et al. 1978), dated at 2.68 ± 20 Ga (Krogh et al. 1976). Peraluminous granitoid rocks are widespread in the ERS, accounting for a cumulative area of approximately 3400 km² between the Ontario-Manitoba boundary and Longitude 89°00'W.

Most of the peraluminous granitoid suite is contained in the Churchill Lake (2200 km²) and the Wapese Lake (635 km²) Batholiths, shown on Figure 4. Dykes, stocks and occasional small batholiths of more fractionated granitoids, usually enriched in muscovite and black tourmaline, are widely intrusive into low and medium grade supracrustal rocks within the boundary zone between the ERS and the Uchi Subprovince.

Distribution of peraluminous granitoids is more restricted in the WRS, perhaps due to a paucity of precursor metasediments. Most of these granitoids are contained in the Ghost Lake Batholith (Figure 4).

Table 4. Characteristics of the Granitoid Suites in the Winnipeg River and English River Subprovinces.

GRANITOID SUITE	FORM	COMPOSITION	PETROLOGICAL CHARACTERISTICS	AGE	EXAMPLES	ASSOCIATED MINERALIZATION
Gneissic	Large, segmented bodies, exclusively in WRS mainly west of longitude 93°00'.	Trondhjemite tonalite and granodiorite, lesser diorite, quartz diorite	Banded (10-50cm), highly deformed, heterogeneous, interlayered tonalite, trondhjemite, granodiorite and quartz diorite; allotriomorphic equigranular, usually fine to medium grained, grey to white on weathered surfaces	Mol resolved; field relations suggest >3 008 Ga near Lac Seul; gneisses near Kenora are 2 760 ± 0 040 Ga (Gower and Clifford 1981)	Tonalite gneisses near Kenora, gneisses associated with domal structures in Vermilion Bay area	None known
Sodic Plutonic	Large batholiths and stocks; simple and complex; in both WRS and ERS	Trondhjemite, tonalite and granodiorite, minor diorite, quartz diorite, granite locally abundant	Moderately to weakly deformed, usually foliated and/or lineated, commonly medium-grained, hypidiomorphic and allotropic equigranular, locally K-feldspar porphyritic, more restricted colour index range (1-20) than gneissic suite, grey to white weathered surfaces	3 008 ± 0 005Ga (Sen Bay Pluton, Krogh et al. 1976) to 2 630 ± 0 010 Ga (Dallies Batholith, Gower and Clifford 1981)	Chamberlain Narrows, Bluffy Lake, Roddy Lake and Dallies Batholiths; Otatagan Lake, Camp Lake and Lateral Lake Stocks	None known May host molybdenite pegmatites
Peraluminous Granitoid Suite	Large stocks and small batholiths, also as more fractionated dikes and small stocks intruding supracrustal rocks in subprovince boundary zones	Granite greatly predominant, granodiorite, also as more trondhjemite, and alkali feldspar granite locally abundant	Massive to weakly foliated, locally cataclastic, medium grained to pegmatitic, but most commonly coarse grained, allotriomorphic equigranular, locally K feldspar porphyritic, colour index usually <10, white on weathered surfaces	2 681 ± 0 020 Ga (leucosome from Lac Seul, Krogh et al. 1976)	Churchill Lake, Wapiti Lake, Wenasaga Lake, Ghost Lake and Sharpe Lake Batholiths; Root Bay and Zealand Stocks	Rare-metal pegmatites; U-bearing calc-silicate garnets and tourmaline-rich sheets; U mineralization
Potassic Plutonic	Large batholiths, mainly in WRS small stocks common east of longitude 93°00', abundant apophyses	Dominantly granite and granodiorite, alkali feldspar granite occurs, usually as late pegmatitic units	Massive to flow-foliated, non-recrystallized, hypidiomorphic equigranular to K-feldspar porphyritic, medium-grained to pegmatitic, colour index usually <8 light pink on weathered surfaces	2 635 ± 0 05Ga (Lount Lake Batholith, Beakhouse and McMill 1980) to 2 560 ± 0 040 Ga (discordant pegmatite dyke, Krogh et al., 1976)	Lount Lake, Tetu Lake and Gullwing Lake Batholiths; Burma Lake Stock	Molybdenite-bearing pegmatites; U mineralization

Figure 4. Distribution of Peraluminous (stippled) and Potassic Granitoid Rocks and Rare-Element, U, Th and Mo Mineralization in the English River and Winnipeg River Subprovinces.

Peraluminous granitoids typically are massive, leucocratic, allotriomorphic to hypidiomorphic granular, with a wide range in grain size (1 mm to 30 cm) and K-feldspar/total feldspar ratio (Figure 5). Granite is most common; granodiorite is subordinate. Less abundant trondhjemite, quartz syenite, and quartz monzonite are locally important. The peraluminous granitoids have a distinctive accessory mineralogy which includes any or all of garnet, cordierite (Plate 1a), sillimanite, tourmaline, muscovite and rare beryl, dumortierite and andalusite. Rafts of metasedimentary paleosome and biotite-rich melanosome, interpreted as remnants of the anatectic process, are commonly present.

Potassic Plutonic Suite

This suite, emplaced between 2.56 and 2.635 Ga (Krogh et al. 1976; Beakhouse and McNutt 1980) represents the culmination of Archean plutonic activity in the ERS and WRS. These rocks underlie approximately 40 percent of the WRS but are comparatively uncommon in the ERS (Table 4).

Compositionally, the potassic plutonic suite is predominantly granite; granodiorite is subordinate (Table 4). Members of this suite are typically uniform in composition and texture over large areas and are composed of plagioclase, quartz, potassium feldspar, biotite and magnetite. Hornblende is rare to absent. Colour index tends to be lower than in the sodic plutonic suite.

The potassic plutonic suite ranges from thin tongues to stocks and huge batholiths. Two major batholiths, which share a common boundary, dominate the western part of the WRS (Breaks and Bond in

Figure 5. Modal Variation of Quartz, Plagioclase and K-feldspar in Peraluminous Granitoid Rocks of the English River Subprovince of Ontario (Breaks and Bond, in press).

PLATE 1

- a. Garnet-biotite-cordierite granite typical of the perlauminous granitoid suite in the ERS. Note the abundant subhedral cordierite megacrysts near the 2.6 cm diameter coin. Part of a 2 m wide dyke which sharply truncates its metatexite metasedimentary host. Outcrop located on eastern Lac Seul at latitude 50°36'45" and longitude 91°50'00".
- b. K-feldspar porphyritic variant of the potassic plutonic suite as exemplified by an exposure in the Lount Lake Batholith on the Jones Road (54°30'00", 93°57'15") about 16 km northeast of the village of Jones. Coin diameter = 1.8 cm.
- c. Pegmatitic mass typical of the potassic plutonic suite which exhibits euhedral to subhedral blocky K-feldspar (as marked by 2.4 cm coin) within irregular quartz-rich patches. Located on Highway 596, 26 km northwest of Kenora (49°49', 94°47").

PLATE 1

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press). These batholiths, the Tetu Lake and the Lount Lake (Figure 4), together form an elongate mass at least 2600 km² in area, oriented east, parallel to the general strike of the WRS. The Tetu Lake Batholith occupies 525 km² and is predominantly granite (ss). Its contacts, gradational over several kilometers, are marked by interdigitation of the granitic units with the host sodic or gneissic granitoid unit. Contacts are defined subjectively on the basis of the predominant plutonic unit. Subtle changes in texture and colour index imply that the batholith is composed of numerous, roughly contemporaneous intrusive units. The Lount Lake Batholith, covering 2060 km², is one of the largest discrete batholiths in the WRS. It is grossly elliptical, being 120 km long and averaging 24 km wide. Approximately 80 percent of the batholith is composed of remarkably uniform, massive, K-feldspar porphyritic granodiorite (Plate 1b). The centre of the batholith consists of porphyritic granite, which is interpreted to be the youngest plutonic unit. The compositional variability of the Lount Lake Batholith is illustrated in Figure 6.

The Gullwing Lake Batholith represents the largest mass of potassic plutonic suite rocks in the study area (Figure 2).

Chemically, these rocks are metaluminous to weakly peraluminous and are distinct from the peraluminous granitoid suite (Chapter 7). It is pointed out here that even though this batholith shares a common geographic name with the Gullwing Lake pegmatitic granite dyke, these plutons are of different relative ages and are chemically distinctive (Chapter 7).

Figure 6.

K-Na-Ca Variation (weight percent) in Potassic Plutonic Suite Rocks in the Winnipeg River Subprovince, Typified by the Lount Lake Batholith (Breaks et al. 1978). Dots: Early Plutonic Units; Crosses: Late Plutonic Units.

Pegmatites related to the potassic plutonic suite occur widely as *in situ* pods such as in biotite granite of the Gullwing Lake Batholith or as exocontact dykes intruded into the sodic and gneissic granitoid suites (Plate 1c). Normally these pegmatites were emplaced later than the peraluminous granitoid suite. These rocks are mineralogically simple relative to rare-element pegmatites of the study area and consist essentially of quartz, albite, K-feldspar, biotite and magnetite.

Pegmatitic dykes are common in amphibolitic mafic metavolcanics near the eastern end of the Gullwing Lake Batholith. These are intruded and locally albitized by biotite-muscovite-bearing, peraluminous dykes such as the Gullwing Lake pegmatitic granite, which contain spodumene and columbite mineralization (see Chapter 4). This intrusive relationship suggests that peraluminous intrusive rocks such as the Gullwing Lake pegmatitic granite were emplaced after 2.56 Ga, which is the youngest emplacement age for the potassic plutonic suite in the WRS (Krogh et al. 1976).

STRUCTURAL GEOLOGY

INTRODUCTION

No detailed structural studies have been undertaken in the study area. Reconnaissance work by Blackburn in Trowell et al. (1977) suggests the presence of three ductile deformation events in the region. The earliest phase of folding was inferred from top reversals in graded wacke-pelites in the Thunder Lake sediments although fold closures were not observed (Blackburn et al. 1982, p.62-63).

The two ensuing fold events relate to "major dextral shear that utilized the Wabigoon Fault as a decollement plane" (Blackburn et al. 1982, p.63).

Near the fault, just south of Thunder Lake (Figure 10), interfering fold relations were observed between:

- 1) early, tight to isoclinal, west-plunging Z-shaped minor folds, commonly with sheared limbs, and,
- 2) later, more open, southwesterly-plunging Z-folds.

The latter fold system has apparent macroscopic equivalents in the Thunder Lake Anticline and Syncline (Figure 10).

GULLWING-TOT LAKE PEGMATITE GROUP

These rare-element pegmatites are concentrated along the southwestern flank and nose of the Lateral Lake Antiform (Figure 20). This doubly plunging structure is centred around the Lateral Lake Stock (Page and Christie 1980).

Top determinations by Page and Christie (1980) in Abram clastic metasediments (Figure 3) and intercalated pyroclastic metavolcanics

suggest that these rocks overlie the mafic metavolcanics which envelop the Lateral Lake Stock. However, the contact between these rocks is not exposed and is postulated by Page and Christie (1980) to be a fault. This cannot be confirmed as top-determinations in the mafic metavolcanics are lacking. The presence of a fault, however, is suggested by the sudden termination of the Gullwing Lake pegmatitic granite dyke at a topographic low marking this area and by the steepening of host rock foliation dips (51° to 86°) immediately south of this contact.

Well developed mineral lineations pervade all supracrustal units in this area; they have a persistent southwest trend and plunges varying between 15° and 55° . This lineation is most impressively developed in pyroclastic breccia of intermediate composition, situated 2 km southwest of the Gullwing Lake dyke where L:W ratios of stretched clasts up to 200:1 were observed. In mafic metavolcanics, which continuously rim the Lateral Lake Stock, and in clastic metasediments of the Abram Group the lineation is more subtle, defined by the orientation of hornblende, biotite or cummingtonite crystals.

Stereonet analysis suggests that the western nose of the antiform at Gullwing Lake has a conical geometry (Figure 7). Mesoscopic lineations which are probably parallel with the macroscopic fold axis have an average trend and plunge of 235° and 28° SW, respectively.

Foliations exhibit a dramatic flexure in strike from southwest to northeast between the known extremities of the Gullwing - Tot Lakes

Figure 7. Lower Hemisphere Schmidt Equal Area Projection of Planar and Linear Fabric Elements Associated with the Southwest Part of the Lateral Lake Antiform.

pegmatite group, and exhibit reversal of dip directions (Page and Christie 1980).

Foliations are generally more poorly developed in the mafic metavolcanics which host most of the rare-element pegmatites in the area. Preservation of primary structures in the metavolcanics is confined to rare occurrences of intact pillow selvages.

Rare-element and barren garnet-biotite-muscovite pegmatites tend to be emplaced at high angles to host rock foliation and the contact between the host mafic metavolcanics with Abram metasediments (Figure 21, back pocket). They exhibit two general strike directions in the Gullwing-Tot Lakes area: 015° to 037° and 120° to 175° . Dips are commonly steep, in contrast to most rare-element pegmatites of the Mavis Lake group (Tables 10 and 12).

It is possible that such pegmatites were simultaneously emplaced along a pre-existing conjugate fracture system, as suggested at the Gullwing Lake pegmatitic granite, which has dyke segments in both groups.

MAVIS LAKE PEGMATITE GROUP

Rare-element pegmatites of this group are structurally confined to the northern limb of the west-plunging Thunder Lake Syncline (Figure 10, back pocket). The host Brownridge volcanics contain an east to northeast-striking penetrative foliation (Figures 17 and 18), and rare isoclinal folds, which have axial planes concordant with the foliation. Dips are steep (70° to 88°) to the north and northeast. Mineral lineations are rare and plunge approximately down-dip. Most

pegmatites in the North Zone (Fairservice Property) tend to strike sub-parallel to host rock foliations (Figure 16, back pocket). However, as the East and Northeast Zones are approached an increasing proportion are discordant, oriented at high angles to host foliations (Pegmatites 11, 12, 17, 18 in Figure 17).

In contrast to the post-tectonic pegmatites of the Gullwing - Tot Lakes area, many of the Mavis Lake pegmatite group show mild effects of late-tectonic deformation. This is particularly evident at the North Zone (Fairservice Property) in pegmatites emplaced near the southeastern contact of the Ghost Lake Batholith, in the form of:

- 1) pegmatitic apophyses folded due to flattening in the contact strain aureole of the Ghost Lake Batholith;
- 2) ductile shearing subsequent to crystallization of youngest zones of some pegmatites (Plate 2a);
- 3) pull-apart structures (Jahns, 1953), particularly involving acicular minerals such as tourmaline;
- 4) mildly strained pegmatite contacts (Plate 2b).

METAMORPHISM

Introduction

Zones of regional metamorphism in the ERS and the WRS are arranged subparallel to boundaries with adjoining Uchi and Wabigoon Subprovinces (Figure 8) as delineated by Breaks et al. (1978), Thurston and Breaks (1978), and Breaks and Bond (in press). The bilateral regional metamorphic pattern between Dryden and Lake St. Joseph (Figure 8) indicates that metamorphic grade increases from the low grade, metavolcanic-rich Wabigoon and Uchi Subprovinces northwards

PLATE 2

- a. Protomylonite texture in the spodumene - K-feldspar - quartz - albite zone of Fairservice Pegmatite No.4.
A = albite, K= K-feldspar, Q=quartz, S=spodumene. Coin diameter = 2.4 cm.
- b. Shallowly-dipping, deformed contact between highly albitized spodumene pegmatite and underlying metasomatized holmquistite-biotite-tourmaline mafic metavolcanics at Fairservice Pegmatite 2. Note presence of tectonic flames protruding from the host into the more competent pegmatite. Coin diameter = 2.4 cm.

PLATE 2

Figure 8. Generalized Distribution of Metamorphic Zones within the English River, Winnipeg River and parts of the Uchi and Wabigoon Subprovinces in Northwestern Ontario (after Breaks et al. 1978).

and southwards, respectively, into the high grade to granulite WRS and ERS.

In other areas, original relationships are now complicated by post-metamorphic fault displacement and widespread emplacement of 2.56 to 2.635 Ga potassic plutonic granitoid suite rocks. In the case of the Wabigoon Subprovince - WRS boundary, there is a general dearth of pelitic bulk compositions that are sensitive enough to record a progressive metamorphic sequence of assemblages. Fortunately in the Dryden-Eagle River area most of these problems are not a factor and there is a relatively intact facies series in which Zealand and Thunder Lake metasediments vary from low grade to high grade over a distance across strike of about 10 km (Figure 9).

The purpose of this section is to describe the field and general petrographic characteristics of the various regional metamorphic zones adjacent to the Ghost Lake Batholith and the Mavis Lake pegmatite group.

EAGLE RIVER - DRYDEN AREA

Metamorphic studies have been undertaken only in two small parts of the study area, by Bartlett (1978) in Eagle River area and less extensively by Gorzynski (1978) around Thunder Lake. Continuity of metamorphic zones and isograds in the intervening region was delineated on a reconnaissance basis by Breaks and Kuehner (1984) over a strike length of 50 km.

Isograds were documented in the field by the initial appearance of index minerals or critical mesoscopic mineral assemblages in

Figure 9. Distribution of Regional Metamorphic Zones and Isograds in Relation to the Ghost Lake Batholith.

widespread metapelitic units. The metamorphic zones listed in Table 5 are bounded by five isograds established on the basis of petrography and staining. Mineral assemblage data are presented in Table 6 (back pocket).

Zoning in the ERS and WRS is consistently characteristic of the low pressure intermediate facies series of Miyashiro (1961) and corresponds to Bathozone 2 of Carmichael (1978). It is marked by an absence of kyanite and widespread coexistence, in intermediate metamorphic zones, of cordierite and andalusite. Cordierite is ubiquitous in the medium, high and granulite grade metapelites of the ERS (Breaks and Bond, in press).

Regional metamorphism was caused by the Kenoran Orogeny which occurred at 2.681 ± 20 Ga, based upon U/Pb zircon dating of leucosome derived from partial melting of clastic metasediments in the English River Subprovince (Krogh and Harris 1976).

Chlorite-Biotite Zone

The lowest metamorphic grade is developed within a 800 m to 1.6 km wide band in the Zealand metasediments, situated adjacent to the Wabigoon Subprovince boundary (Figure 9).

Metamudstones, which are fine-grained and well foliated, contain chlorite-biotite-muscovite which collectively define the foliation. The metawackes, which are more widespread, are also fine-grained but exhibit a greater range in grain size, containing well preserved detrital grains of plagioclase and quartz with diameters up to 2 mm. These rocks commonly have assemblages with little or no muscovite.

Table 5: Metamorphic Zones in the Eagle River - Dryden Area

<u>Metamorphic Zone</u>		
<u>Grade</u>	<u>Eagle River Area</u>	<u>Dryden Area</u>
Low	Chlorite-Biotite	Chlorite-Biotite
Medium	Andalusite Sillimanite-Muscovite	Andalusite Staurolite-Biotite Garnet-Biotite (Cumingtonite) Sillimanite-Muscovite
High	Sillimanite-K-feldspar (absence of metatexis) Sillimanite-K-feldspar-Cordierite (presence of metatexis)	

Andalusite Zone

This zone is also continuous across the southern part of the study area (Figure 9), and varies in width from 425 m to 3.2 km. This zone is characterized by extensive development, in metamudstone units, of light pink, medium- to coarse-grained andalusite porphyroblasts, typically 1 cm across, which protrude prominently on weathered surfaces (Plate 3a). This mineral composes up to 20 percent of such units. In some exposures andalusite is oriented parallel to an S_2 foliation that it has overgrown; this has a low inclination to the composite S_0 - S_1 planar fabric (Plate 3a).

Metawacke units are slightly more recrystallized than their counterparts in the chlorite-biotite zone. Mineralogical changes upon entry into the andalusite zone are reflected in the appearance of garnet and disappearance of chlorite. Cordierite also makes its initial appearance near the andalusite-in isograd (Figure 9). This mineral is conspicuously developed as light grey-green ellipsoidal poikiloblasts, up to 1.5 cm long, composing up to 15 percent of metamudstone units. Pseudo-hexagonal poikiloblasts are far less common and are almost completely replaced by dark green chlorite (Plate 3b).

In relatively rare, possibly magnesium-rich metamudstone units the assemblage chlorite-cordierite-biotite is apparent at locality G8-5 (Figure 38, back pocket). Outcrops to the south contain chlorite-biotite-muscovite metamudstone. The spatial association of these assemblages suggests the continuous reaction:

$$\text{Chlorite} + \text{Muscovite} + \text{Quartz} = \text{Cordierite} + \text{Biotite} + \text{H}_2\text{O}.$$

2

MICROD

Iron-rich metapelites such as those interbedded with magnetite-chert banded iron formation at locality F24-8 contain the assemblage biotite-garnet-hornblende-cummingtonite-clinopyroxene-magnetite-plagioclase which may be indicative of the low pressure reaction of Froese (1980):

Hornblende + Plagioclase = Garnet + Clinopyroxene + Cummingtonite.

Staurolite Zone

This zone was observed only on the southwestern shoreline of Thunder Lake, approximately 400 m north of the garnet-in and andalusite-in isograds (Figure 9). At station 32-14-3 (Figure 38, back pocket) metapelite contains 10 percent subhedral staurolite poikiloblasts up to 3 x 4 mm, apparently in two generations:

- 1) staurolite with internal inclusion trails parallel to S_1 biotite-muscovite foliation in microlithons separated by S_2 crenulation cleavage planes and;
- 2) staurolite poikiloblasts post-tectonically overgrowing microfolds associated with development of the S_2 cleavage.

Staurolite also occurs sporadically in the andalusite zone as ragged inclusions up to 1 x 1.5 mm within cordierite poikiloblasts. This apparently unstable form of staurolite, now present downgrade from the staurolite isograd at Thunder Lake, may be equivalent to the first generation of staurolite described above.

Sillimanite-Muscovite Zone

The first sillimanite zone was delineated in two areas adjacent to the southern contact of the Ghost Lake Batholith (Figures 9 and 10):

- 1) Ardis Lake area in northwestern Aubrey Township, and,
- 2) Dryden area in van Horne and Zealand Townships.

Within the intervening area, between Aubrey Creek and the Oxdrift Road, the first appearance of sillimanite could not be conclusively traced because of sporadic exposure (Figures 9 and 10). However, it is likely that the first sillimanite isograd does not continue through the study area but is truncated by the Ghost Lake Batholith. This is suggested by proximity of andalusite-cordierite metamudstones to the Ghost Lake Batholith near the south end of Peganeis Lake (Figure 10). The first sillimanite zone likely developed through inversion of andalusite to sillimanite. The isograd may actually be a band as relict andalusite occurs over a 1200 m wide interval in the Eagle River area, and coexists with a progressive coarsening sillimanite. Bartlett (1978) recognized andalusite rimmed by plagioclase in metapelite containing cordierite and sillimanite in Temple Township 1.5 km north of locality 32-5-20 (Figure 38, back pocket) which in part may explain its metastable persistence into the higher grade sillimanite-muscovite zone.

In the central part of the zone sillimanite is readily discernible in metamudstone units as coalescing coarse white to grey fibrous aggregates which protrude above the weathered surface (Plate 3c) and compose up to 15 percent of the rock. Sillimanite here coexists with cordierite, biotite and muscovite.

Sillimanite - K-feldspar Zone

High grade metamorphism in the study area is marked by the coexistence of sillimanite and K-feldspar. This zone is almost

exclusively adjacent to the western contact of the Ghost Lake Batholith in Sanford, Eton, and Rugby Townships (Figure 9).

Two subzones are mappable, based upon presence or absence of leucosomes:

- 1) Sillimanite - K-feldspar Zone (non-migmatitic), and
- 2) Sillimanite - K-feldspar - Cordierite - Garnet Zone (migmatitic).

Sillimanite - K-feldspar Zone (Non-Migmatitic)

This sub zone occupies a small area immediately north of the sillimanite-muscovite zone in Aubrey and Temple Townships (Figure 9 and 10) adjacent to the extreme southwestern contact of the Ghost Lake Batholith. Metamudstone units are marked by coexistence of sillimanite and K-feldspar in the absence of migmatitic leucosome. These rocks are porphyroblastic, fine-to coarse-grained, and well foliated, quartz veins 2 mm to 4 cm in width, concordant to foliation are common. These veins, which have been deformed into tight to isoclinal folds, may represent a former fluid phase developed through metamorphic prograde devolatilization reactions.

Sillimanite tends to occur in aggregates up to 1.3 x 2.3 cm and coexists with cordierite, biotite, K-feldspar and occasional garnet. Adjacent to the Ghost Lake Batholith at station G8-1 (Figure 38, back pocket) fibrolitic sillimanite occurs in nodules with muscovite and quartz, hosted by numerous 3 to 15 m wide granitoid sills which are similar to plutonic unit GLB-3 (Chapter 3).

Sillimanite - K-feldspar - Cordierite - Garnet Zone (Migmatitic)

This subzone is characterized by metamudstone units which have locally yielded protometatexite and metatexite migmatites (Table 1), similar to those encountered in the ERS (Plates 20a and 20b; Breaks et al. 1978). They are generally porphyroblastic, well foliated, fine-to coarse-grained and exhibit rough weathered surfaces compared to intercalated metawackes (Plate 3d). The assemblage sillimanite - K-feldspar - biotite - garnet - quartz - plagioclase is common.

In contrast, metawackes are generally fine-grained, equigranular and weakly foliated. These monotonous rocks typically consist of the relatively simple assemblage almandine - biotite - quartz - plagioclase and invariably exhibit no evidence of partial melting. Cordierite poikiloblasts occasionally develop in metawackes and compose up to five percent of the rock. This high grade zone contains less than five percent leucosome, consisting of dykes and local masses of medium-grained to pegmatitic, massive, white, inequigranular granite. Both the *in situ* and petrographically identical intrusive granitoid rocks commonly contain sillimanite, garnet, biotite, muscovite, cordierite and tourmaline. The latter two minerals also occur conspicuously in symplectitic intergrowths with quartz. Blue dumortierite is a rare but notable accessory, occurring in a pegmatitic leucogranite dyke at E31-9 (Figure 39, back pocket).

At localities F7-8 and E9-1 leucosome layers 3 to 8.5 cm wide are continuously flanked by 3 to 30 mm thick selvages enriched in biotite, cordierite and sillimanite.

Metamorphic Conditions

The observed progression of isograds in the Eagle Lake area indicates a low pressure - high temperature gradient of regional metamorphism of the Abukuma type (Figure 11). This inferred gradient is strikingly similar to those in other parts of the Superior and Churchill Province as revealed in studies by Ermanovics and Froese (1978), Bailes and McRitchie (1978), Breaks et al. (1978) and Poulsen (1980).

The medium grade zones have an approximate load pressure of 3 kbar, constrained by the widespread coexistence of andalusite and cordierite and by positioning of the staurolite-in isograd within the andalusite zone (Figures 9 and 10). In the high grade metamorphic zones the position of the PT gradient is constrained below the 3.5 kbar invariant point defined by intersection of univariant curves $\text{muscovite} + \text{plagioclase} + \text{quartz} = \text{K-feldspar} + \text{Al}_2\text{SiO}_5 + \text{H}_2\text{O}$ of Thompson (1974) with the muscovite-granite solidus of Huang and Wyllie (1973). Evidence for this restriction lies in the coexistence of sillimanite and K-feldspar in metapelites devoid of anatectic leucosome.

Bartlett (1978, p.45 to 47) estimated a temperature of $700^\circ\text{C} \pm 50^\circ\text{C}$ for the sillimanite-K-feldspar zone based upon Fe-Mg exchange in garnet-hornblende and garnet-clinopyroxene pairs in an amphibolite dyke situated at the village of Eagle River. This estimate assumed a load pressure of 3 kbars.

Figure 11. Path of Regional Metamorphism for the Dryden-Eagle River Area (shaded bar). P-T Conditions for Granulite Facies Assumed to be at Depth in Study Area Estimated from the Work of Chipera and Perkins (1988) on the Eastern Lac Seul Granulite Zone of the ERS. Numbers Refer to Works of: 1 = Hemley (1967); 2 = Brown (1971); 3 = Holdaway (1971); 4 = Hirschberg and Winkler (1968); 5 = Richardson (1968); 6 = Thompson (1974); 7 = Newton et al. (1974); 8 = Huang and Wyllie (1973); 9 = Kerrick (1972); 10 = Vielzeuf and Holloway (1988); 11 = Le Breton and Thompson (1988).

Timing of Ghost Lake Batholith Emplacement vs Regional Metamorphism

The general pattern of metamorphic zones immediately adjacent to much of the Ghost Lake Batholith is notably asymmetrical (Figure 9). High grade migmatitic metasediments are restricted to the western and northwestern contact areas. The eastern boundary and almost the entire southern contact is in contrast enveloped by the sillimanite - muscovite zone. Also apparent is the low-angle discordance of the first and second sillimanite isograds against the southern contact of the batholith. This truncation implies that the more evolved plutonic units (GLB-3, -4, -5, -6 and -7; Chapter 3) were emplaced later than the regional metamorphism. This interpretation is corroborated by the sharp intrusive contacts of these plutonic units against adjacent supracrustal rocks, as at the Contact Beryl Occurrence where biotite-rich contact metasomatic selvages overprint a hornblende lineation in previously deformed, amphibolitic mafic metavolcanic rocks (see Chapter 3). The western to northwestern contact, although not exposed, is probably an intrusive contact between the GLB and metasedimentary migmatites of the sillimanite-K-feldspar-cordierite-garnet zone. This implies that a deeper level of the batholith is exposed in this area, given the abundance of inhomogeneous diatexite and restite patches, and a more primitive degree of chemical fractionation (Chapter 7) relative to rocks at the eastern end of the pluton. The low proportion of petrographically identical granitoid leucosome both of an in situ and intrusive origin in the metasedimentary migmatites militates against the derivation of inhomogeneous diatexites of GLB-1 at the present exposure level.

Chemical reasons for the discordance between the level of batholith emplacement and regional metamorphic grade of the host rocks will be examined in Chapters 7 and 9.

Metamorphic Setting of the Rare-element Pegmatites

The Mavis Lake rare-element pegmatite group occurs within the first sillimanite zone, similar to the Black Hills pegmatite district of South Dakota (Shearer et al. 1986), and some pegmatites in the Sparow Lake area of the Yellowknife pegmatite district in the Northwest Territories (R. Kretz, personal communication, 1989). Cerny (1982b, p.448-449) reported that rare-element pegmatite fields in general are hosted by Abukuma-type metamorphic sequences and specifically lie within the andalusite-cordierite-muscovite zone. Only barren pegmatites were said to occur above the andalusite-out isograd (Cerny 1982b, p.449), which is brought into question by the conflicting examples given above.

Less is known regarding metamorphic context of the Tot-Gullwing Lakes pegmatite group. Regional metamorphism in this area is generally medium-grade, judging by overall lack of chlorite in mafic metavolcanics, widespread presence of muscovite in metapelites and some metawackes, coupled with lack of migmatization. Isograds in this area are impossible to delineate as metamudstones of the Abram Group in contact with the host mafic metavolcanics are very poorly exposed east of the Webb-Drope Township boundary (Page and Christie 1980). Clastic metasediments are probably within the first sillimanite metamorphic zone as indicated by coexistence of sillimanite with muscovite in some metapelites.

Chapter Three

GEOLOGY AND PETROLOGY OF THE MAVIS LAKE PEGMATITE GROUP

INTRODUCTION

The geology of this pegmatite group is dominated by the Ghost Lake Batholith (GLB), a peraluminous, two-mica granitoid complex. This batholith represents the most plausible source for the Mavis Lake rare-element pegmatite group, on the basis of petrological features and chemical characteristics, the latter of which will be documented in Chapter 7. The purpose of this chapter is to establish the geological setting and to examine the petrological features of the GLB and adjacent rare-element pegmatites.

GEOLOGICAL SETTING

The Mavis Lake pegmatite group is almost entirely contained within the Brownridge volcanics and apparently confined to the north limb of the west-plunging Thunder Lake Syncline (Figure 10, back pocket). Dimensions of the pegmatite group, which strikes 090° to 105°, are 8 km in length (between the Dryden Airport and just east of Mavis Lake), and breadth varying between 0.8 and 1.4 km.

A well-developed horizontal zonation of pegmatites is apparent over this strike length. Characteristic mineral assemblages, internal structures, degree of albitization and chemistry change progressively with increasing distance from the parental Ghost Lake Batholith (Table 2).

GHOST LAKE BATHOLITH (GLB)

This plutonic mass, approximately 280 km² in area, has an

elongation concordant with the east- to northeast-striking regional structural trends. It is exposed for about 40 km parallel to strike between Ghost Lake and the village of Eagle River and has a breadth varying between 2.5 and 13 km. In view of the strong possibility of a genetic linkage with pegmatites, at least with the Mavis Lake rare-element pegmatite group, it was recently mapped at a scale of 1:31,680 (Figure 10, back pocket).

The overall geological setting and petrological characteristics are strikingly similar to those of other S-type granitoid masses abundant in the English River subprovince (ERS), as documented by Breaks et al. (1978, 1985) and Breaks and Bond (in press).

The GLB is distinguishable, however, from virtually all other plutonic masses in the region by virtue of a unique accessory and varietal mineral suite comprising muscovite, cordierite, sillimanite, Mn-enriched garnet, tourmaline and rare dumortierite and beryl. Host rocks of the GLB (Figure 2) consist of wacke, pelite and ironstone belonging to the Zealand and Thunder Lake metasedimentary rocks along most of the southern contact, derived metatextitic metasedimentary migmatite along the western contact and granitoid rocks along the north to northeastern contact.

Collectively the supracrustal rocks envelop approximately 75 percent of the batholith. Along the southern contact, the GLB lies in abrupt contact with low- to medium-grade metasediments. The western and northwestern contact, not exposed, is probably an intrusive contact with metatextitic (see Table 1) metasedimentary migmatite, given the rapid change from diatexite in the pluton to metatextite

which contains a much lower content (<5 percent) of the leucosome component.

Mafic metavolcanics of the Central Volcanic Belt are only locally in contact with the GLB, extending for 10 km from west-central Zealand Township to Ghost Lake in Brownridge Township (Figure 2). These rocks are extremely important, however, as they are host to most of the rare-element pegmatites.

Granitoid host rocks consist of the Gullwing Lake Batholith along the northern and northeastern extremities of the GLB (Figure 10, back pocket) and the Broadtail Lake Stock along the arcuate west to northwest trending contact in Zealand and Wainwright Townships. The Broadtail Lake Stock is massive to foliated and locally gneissic biotite or magnetite-biotite tonalite (sodic plutonic suite of Breaks et al. 1978), intruded by massive medium- to coarse-grained and occasionally pegmatitic biotite granite and alkali-feldspar granite (potassic plutonic suite of Breaks et al. 1978) which most likely originated from the Gullwing Batholith. A small stock of porphyritic biotite granite forms the country rock at the eastern extremity of the GLB at Ghost Lake, where it appears locally as inclusions.

Petrological features of the sodic and potassic plutonic suites are summarized in Table 4.

Petrology of the Ghost Lake Batholith

Eight distinct plutonic units were delineated within the batholith. Their general distribution, petrological and geological

features are summarized in Figure 12 and Table 7.

GLB-1: this unit composes most of the batholith (Figure 12) and consists of white, homogeneous diatexite of granitic to granodioritic composition. These rocks are massive to weakly foliated, generally hypidiomorphic-granular (Plates 4a and 4b), inequigranular to rarely porphyritic and coarse-grained to locally pegmatitic rocks (see Figure 14 later in this chapter and Table 7). A subordinate rock type, characterized by cordierite megacrysts, is distributed along part of the northwest contact (Figure 12) and is gradational into typical GLB-1.

GLB-1 grades into inhomogeneous diatexite where metasedimentary paleosome inclusions, mafic-rich restite patches and metasedimentary migmatite rafts individually or collectively exceed ten percent of a given outcrop (Plates 4c and 4d). Inhomogeneous diatexite variants of GLB-1 are mainly confined to two narrow, curvilinear, en-echelon zones. The larger more northerly zone (Figure 12) appears to represent a more highly mobilized easterly continuation of a prominent re-entrant of metatextitic metasedimentary migmatite.

Biotite and muscovite are the commonest varietal minerals in GLB-1; cordierite and sillimanite only locally exceed one percent. Accessory minerals include apatite, tourmaline and zircon. Muscovite is largely of replacement origin as suggested by widespread ragged intergrowth with plagioclase, K-feldspar, cordierite and biotite. North of the larger diatexite unit, biotite commonly prevails over muscovite. Cordierite also appears locally near the northwest contact (Plate 4e). Rectangular cordierite crystals up to 0.5 x 1.6 cm,

Figure 12. General Distribution of Intrusive Units GLB-1 to -8 in the Ghost Lake Batholith.

Table 7. Summary of Petrological Features of the Plutonic Units of the Ghost Lake Batholith.

INTRUSIVE UNIT	ROCK TYPE	PETROGRAPHIC FEATURES		INTRUSIVE RELATIONS
GLB-1	Muscovite-biotite and biotite-muscovite granite; rare granodiorite and quartz syenite. Locally cordierite-bearing	Massive to weakly foliated, coarse-grained inequigranular, generally hypidiomorphic-granular. Accessory sillimanite, garnet, apatite and zircon		
GLB-2	Muscovite-biotite granite to granodiorite	Massive, occasionally weakly foliated, fine- to medium-grained, allotriomorphic-granular. Distinctive medium-grey colour		Occurs as small masses and dykes cross-cutting GLB-1
GLB-3	Muscovite-biotite and muscovite-biotite granite to rare quartz monzonite	Massive to weakly foliated and occasionally lineated, medium-grained, allotriomorphic-granular. Unique presence of fibrolite - quartz - muscovite faserkiesel. Accessory garnet		No intrusive relations with adjacent GLB-1 or -2 observed. GLB-3 cut by dykes and small masses of GLB-4 and -7. GLB-3 contains layers and small pods of gradational, coarse-grained to pegmatitic granite
GLB-4	Muscovite-biotite and biotite-muscovite granite to granodiorite	Massive coarse-grained, vaguely layered with fine-grained apite, hypidiomorphic-granular. Accessory garnet and tourmaline. Distinctive plumose intergrowths of quartz and muscovite		Observed as veins and dykes cutting GLB-1, -2 and -3
GLB-5	Garnet-muscovite-tourmaline granodiorite to granite	Massive to occasionally layered, fine-grained, allotriomorphic-granular, typically equigranular		Noted only in eastern part of Ghost Lake Batholith and intruded by GLB-6 and -7
GLB-6	Garnet-biotite-tourmaline-muscovite granite to granodiorite	Massive, coarse-grained hypidiomorphic-granular with distinctive randomly oriented bladed aggregates of biotite and green muscovite		Cross-cuts GLB-5
GLB-7	Muscovite-tourmaline granite	Massive, coarse-grained hypidiomorphic-granular may be associated with fine-grained, layered sodic apite. Accessory garnet, biotite, apatite, beryl and columbite-tantalite group		Cross-cuts GLB-1, -2, -3 and -4. Youngest granitic unit in Ghost Lake Batholith. May be correlative with exocontact, pegmatitic dykes having tourmaline halos
GLB-8	Tourmaline-quartz veins	Small veins concentrated towards eastern part of Ghost Lake Batholith. Relatively common in metasedimentary rocks between Dryden and Thunder Lake. Local metasomatic tourmaline halos in host rocks		

PLATE 4

- a. Well-foliated biotite granite: homogeneous diatexite of GLB-1 at locality E21-4. Pen length = 15 cm.
- b. Hypidiomorphic-granular, coarse-grained texture typical of cordierite-muscovite-biotite granite from unit GLB-1 at locality C20-3. Quartz and stained feldspars (K-feldspar = yellow, plagioclase = white, quartz = grey and biotite = black).
- c. Typical exposure of inhomogeneous diatexite variant of GLB-1 composed of abundant granite leucosome, biotite-rich restite and lesser metawacke inclusions. This unit is intruded by a narrow, slightly boudinaged, grey dyke of weakly foliated, muscovite-biotite granite (GLB-2). Locality E19-8 at Oxdrift. Coin (2.4 cm diameter) marks the upper contact between the two units.
- d. Detail up of cordierite-sillimanite-biotite restite in biotite granite (GLB-1) at locality E21-4. Coin diameter = 2.4 cm.
- e. Cluster of subhedral to anhedral cordierite crystals in cordierite-muscovite-biotite granite of unit GLB-1 at locality C17-4. Note the marginal replacement of cordierite by a dark green, aphanitic aggregate of chlorite and muscovite.
- f. Grey, weakly foliated muscovite-biotite granite (GLB-2) at locality E19-8. Note the presence of coarse muscovite porphyroblasts which subtly interrupt the foliation.

PLATE 4

occurring as partial to complete pseudomorphs of dark blue-green chlorite, were noted at several localities (C17-4A and C20-3 for example, Figure 38, back pocket). In many places, light mauve cordierite relics are megascopically discernible.

Muscovite tends to predominate over biotite along the south to southeast limits of GLB-1. South of the larger unit of inhomogeneous diatexite, sillimanite is a widespread accessory, commonly visible in hand specimen as pale brown fibrous aggregates. Red-brown garnet, typically less than 4 mm diameter, is uncommon in GLB-1 relative to GLB-3 to -7. It is most conspicuously developed in GLB-1 at the northern extremity of the batholith at Rugby Lake (Figure 10, back pocket), as subhedral crystals up to 5 mm diameter mantled by abundant biotite.

GLB-2: this scarce plutonic unit post-dates GLB-1, occurring as narrow dykes up to 3 m wide and small masses scattered throughout the batholith (Plate 4c). It is massive to subtly foliated, fine- to medium-grained, generally equigranular and allotriomorphic-granular. It is usually grey on the fresh surface, however some specimens are light pink. Biotite is the chief accessory and always exceeds muscovite in abundance. Muscovite occurs as 4-5 mm diameter clots which may be porphyroblastic in origin, as they have overgrown a disruption of subtle foliation defined by oriented thin muscovite and biotite laths (Plate 4f). An example of a GLB-2 dyke cutting discordantly across foliation of GLB-1 diatexite can be observed on Highway 17 at Oxdrift (Plate 4c). This dyke is post-dated in turn by narrower lens-shaped dykes of a possibly deeper level equivalent of

GLB-7. The latter consist of white, massive, coarse-grained, apatite-muscovite-biotite granite.

GLB-3: is confined to a slender unit immediately contiguous to the southwest contact of the batholith. The main rock type is a white, massive to weakly foliated and occasionally lineated, generally fine- to medium-grained, allotriomorphic-granular granite containing biotite and muscovite as the predominant accessories (Plate 5a).

GLB-3 grades throughout into coarse-grained to pegmatitic pods and layers of biotite and garnet-biotite granite. The coarser rock is also allotriomorphic-granular and is typified by thin randomly-oriented biotite blades up to 0.5 mm x 1.5 cm and anhedral light pink to white K-feldspar up to 2 x 3 cm.

Many exposures contain porcellaneous white or hematite-stained quartz-muscovite-fibrolite faserkiesel (Plate 5b), which truncate the foliation. Larger, lensoidal muscovite-quartz-sillimanite nodules up to 3 x 4 x 2 cm also occur in the batholith, for example near the intersection of Highway 594 and the Minnitaki Road. These greatly resemble sillimanite nodules in the Bancroft area, described by Breaks and Shaw (1972). Quartz-muscovite knots also occur, for example at locality F13-1. These aggregates consist of coarse, random anhedral muscovite, up to 1 cm diameter, surrounded by a thin 2-4 mm wide mantle of medium-grained quartz. The host exhibits a biotite-muscovite mineral lineation. A 0.5-1 cm halo exhibiting biotite depletion everywhere surrounds the quartz-muscovite may be a product of local dealkalization (Losert 1968).

PLATE 5

- a. Unit GLB-3 showing muscovite-biotite granite with typical medium-grained, allotriomorphic-granular texture at locality F11-6. Stained feldspars (K-feldspar = yellow, plagioclase = white), quartz = grey, black = biotite and red-brown = garnet.
- b. Muscovite, sillimanite-quartz faserkiesel in unit GLB-3 near the southern batholith contact at locality F13-1. Coin diameter = 2.4 cm.
- c. Part of a pegmatitic leucogranite dyke (GLB-4) transgressing the foliation of GLB-3 on Highway 17 adjacent to the Dryden cemetery at locality F21-1. Coin (1.8 cm diameter) marks the contact between the two units.
- d. Plumose intergrowths of muscovite and quartz (dark areas) in typical pegmatitic leucogranite at locality E29-1. K = K-feldspar megacrysts and A = part of a garnet-muscovite aplite layer. Coin diameter = 1.8 cm.
- e. Garnet-tourmaline-muscovite fine-grained leucogranite (GLB-3) intruded by a dyke of tourmaline-muscovite potassic pegmatite. Locality E31-9 near eastern end of the Dryden Airport. Coin (left centre) diameter = 1.8 cm.

PLATE 5

Temporal relations between GLB-3 and GLB-1 and -2 are unknown as these units were never observed in contact. A roadside exposure on Highway 17 at the Dryden cemetery, which represents the most easterly exposure of GLB-3, however, does exhibit the later emplacement of dykes of GLB -4 and -7 (Plate 5c).

GLB-4, -5, -6, -7 and -8: these units, in a complex relationship, constitute the eastern portion of the GLB in Zealand and Brownridge Townships. GLB-4, -5, -6, -7 and -8 are mainly confined to an elongate area situated adjacent to the south and east contact (between the Dryden Airport and Ghost Lake).

Individual delineation of these various plutonic units would probably be effected only by very detailed mapping. This segment of the batholith is considered to represent a higher level than the area dominated by GLB-1, suggested by a paucity of metasedimentary paleosome and restite clots, striking differences in accessory mineralogy, and sharp intrusive contacts with medium-grade (first sillimanite zone) metasediments. Muscovite tends to be the chief varietal mineral accessory; biotite is typically of accessory status (less than one percent) and is commonly absent. Garnet contents reach their highest in this part of the batholith (3 percent). Tourmaline is particularly abundant and locally reaches levels of 10 to 20 percent. This mineral, in particular, suggests a relatively high concentration of boron, at least, in the eastern part of the GLB; it is also abundant in nearby, exocontact pegmatite dykes.

GLB-4, the most abundant of these plutonic units, consists of pegmatitic leucogranite which is in places interlayered with

subordinate amounts of aplite, as at localities E29-1 and F22-5 (Figure 38). Here anhedral megacrysts of graphic K-feldspar and quartz up to 15 x 15 cm are common, set in a matrix of albite, quartz, muscovite, biotite garnet and tourmaline. Wispy light pink aplite bands contain up to 20 percent euhedral, red-brown garnet and 1-2 percent biotite. Plumose aggregates of muscovite and quartz are commonplace (Plate 5d).

Fine-grained leucogranite, GLB-5, is widespread near the batholith contact south of the Dryden airport. It is massive to locally layered, commonly equigranular, and granodiorite in composition. Accessories, in order of increasing abundance, are garnet, green muscovite and tourmaline. Typical examples are shown in Plates 5e and 6a.

Locally, gradational into a megacrystic variant of GLB-5 is notable as at locality E31-4 (Plate 6b). These rocks are characterized up to five percent by subhedral poikilitic tourmaline megacrysts which locally exceed 1 cm in length. Tourmaline contains round quartz inclusions 0.1 to 1 mm in diameter. However, no inclusions of the adjoining K-feldspar, albite, medium-grained quartz, or garnet occur in the tourmaline.

Muscovite-rich potassic pegmatite, GLB-6, is restricted to outcrops near the eastern end of the Dryden Airport and at the Contact Beryl Occurrence (Figures 10 and 14). It is similar to potassic pegmatite (GLB-7) and may be a possible variant of this unit as intrusive relations between the two could not be found. It is in places interlayered with GLB-5 (Plate 6c) but also occurs as masses

PLATE 6

- a. Garnet-muscovite-tourmaline fine-grained leucogranite from locality E31-4. Coin diameter = 1.8 cm.
- b. Variant of fine-grained leucogranite characterized by poikilitic tourmaline megacrysts at locality E31-4.
- c. Muscovite-rich potassic pegmatite (GLB-6) interlayered with fine-grained leucogranite at locality E31-9, near the Dryden Airport. Hammer length = 34 cm.
- d. Garnet-muscovite-tourmaline potassic pegmatite (GLB-7) and layered sodic aplite at locality E31-9. Note the well developed blocky K-feldspar, marked by 2.4 cm diameter coin, in contact with a quartz-rich pod.

PLATE 6

containing inclusions of fine-grained leucogranite (Plate 7c).

GLB-6 is inequigranular, comprising subhedral to anhedral megacrysts of K-feldspar 2 to 16 cm across in a matrix of finer-grained quartz, albite, green muscovite garnet, biotite and tourmaline (Plate 7a). The main petrologic distinction from GLB-7 is the abundance of green muscovite in randomly oriented, bladed aggregates, occasionally intergrown with biotite.

Although mainly developed within the eastern part of the batholith, dykes of GLB-6 have been noted up to 15 km west of Dryden airport (station E21-4, Figure 38). Here it occurs as a 0.25 x 20 m lens intruding GLB-1 and GLB-4. This dyke is characterized by substantial green muscovite (15 percent) and black tourmaline (10 percent). Fine-grained red garnet is the most abundant accessory; sparse apatite is also present. Comb structure, mainly evident as 0.5 x 6 cm tourmaline oriented normal to contacts, is developed along the dyke margins and grades into a central quartz-rich core zone. This structure suggests emplacement of magma into tension fractures (Rossovskiy and Konovalenko 1978).

GLB-7 occurs as small masses and dykes of potassic pegmatite. It is notable in being the only internal unit of the Ghost Lake Batholith which contains rare-element minerals (see Dryden Airport and Contact Beryl Occurrences, below). This unit varies from zoned pegmatites exhibiting well developed blocky K-feldspar crystals, in contact with a quartz-rich core (Plate 6d) to masses consisting principally of more irregularly-shaped K-feldspar megacrysts (Plate 7d). Both types may be layered with sodic aplite (Plates 6d and 7d).

PLATE 7

- a. Muscovite-rich potassic pegmatite immediately adjacent to orbicular aplite, containing megacrysts K-feldspar in a muscovite-rich matrix at locality E31-2. Coin diameter = 2.4 cm.
- b. Layer in fine-grained leucogranite which is truncated by muscovite-rich potassic pegmatite at locality E31-2. Coin diameter = 2.4 cm.
- c. Orbicular aplite at locality E31-4, comprising tourmaline-muscovite-albite orbicules set in a K-feldspar-quartz-muscovite-albite matrix. Coin diameter = 2.4cm.
- d. Potassic pegmatite featuring elongate K-feldspar megacrysts oriented normal to banding in adjacent sodic aplite, at locality E31-2. Coin diameter = 2.4 cm.
- e. Crystal aggregate of columbite (near 2.4 cm wide coin) contained in albite of garnet-tourmaline-muscovite potassic pegmatite at locality E31-2.

PLATE 7

Potassic pegmatite may have appreciable black tourmaline (up to 20 percent). The most abundant accessories are garnet, green muscovite, and tourmaline.

GLB-8, the youngest plutonic unit of the Ghost Lake Batholith, consist of very minor tourmaline-quartz veins that cut GLB-1 as at locality E21-3 and GLB -5, -6, and -7 in the eastern end of the batholith. These veins are similar to that shown in Plate 9f, which may be an exocontact equivalent of GLB-8.

Petrographic Features of GLB-5, -6, and -7 at the Contact Beryl Occurrence are best revealed in the vicinity of locality E31-4 (Figure 38). Here a complex melange of GLB-5, -6 and -7 and altered equivalents provides a rare glimpse of internal contact relations among these chemically fractionated rocks (see Chapter 7) at the only exposed contact of the batholith. This area was extensively stripped and washed of overburden by Sanmine Exploration Incorporated in 1984 as part of an exploration program in search for gold and tungsten.

Because of the complexity of plutonic units in the eastern part of the GLB, and the discovery by the author of an apparently unique occurrence of holmquistite in an alteration zone spatially linked with the batholithic contact, this 90 by 170 m area was mapped in detail (Figure 13).

Potassic pegmatite (GLB-7), muscovite-rich potassic pegmatite (GLB-6) and fine-grained leucogranite (GLB-5) represent the most abundant plutonic units. Minor sodic aplite, orbicular aplite and

Figure 13. Detailed Geology of Ghost Lake Batholith-Central Volcanic Belt Contact Zone at the Contact Beryl Occurrence. A and B are Described in Table 8.

pegmatitic leucogranite (GLB-4) are also evident. Primary units and their modes assemblages are listed in Table 8. Modal analyses of layered and orbicular aplite were not undertaken.

Fine-grained leucogranite (GLB-5) occurs as layers 0.5 to 2 m in thickness, which are fine- to medium-grained, equigranular, and locally have a laminated appearance. This unit is characterized by 10-15 percent disseminated black tourmaline and negligible muscovite. Accessory, fine-grained, ruby-red garnet is normally present in quantities of 1 to 2 percent.

Muscovite-rich potassic pegmatite (GLB-6) is distinguished by large homogeneous to locally graphic (with quartz) megacrysts of K-feldspar embedded in a matrix composed of quartz, albite, tourmaline and abundant muscovite (up to 35 percent). The K-feldspar megacrysts compose up to 27 percent of the rock, are up to 16 cm in diameter, and commonly coalesce (Plate 7a).

Muscovite-rich potassic pegmatite (GLB-6) occurs as masses grossly interlayered with orbicular aplite and fine-grained leucogranite (GLB-5). Field relations indicate that this unit is later than GLB-5: narrow layers of fine-grained leucogranite with internal laminations have been broken and engulfed by the muscovite-rich potassic pegmatite (Plate 7b). Furthermore, coarse-grained elongate muscovite crystals have clearly crystallized normal to the contact of laminated fine-grained leucogranite (Plate 7b).

Orbicular aplite, found only at the Contact Beryl Occurrence, is a 0.6 by 2 m unit intercalated with GLB-5 and 6. It is conspicuous

owing to ovoid proto-orbicules (Leveson 1966) that attain a maximum size of 1.5 by 4 cm and compose about five percent of the unit (Plate 7c). They display delicate concentric internal laminations composed of fine-grained tourmaline. There is an outermost shell, 1-5 mm wide, consisting mainly of green muscovite and quartz. Muscovite tends to be oriented with the (001) cleavage approximately normal to the boundary of the orbicules. The coarser-grained matrix is composed of tourmaline, muscovite, K-feldspar, quartz and plagioclase.

Potassic pegmatite (GLB-7) occurs as 0.5 to 2 m thick layers; some contain white elongate blocky K-feldspar up to 0.3 by 1 m which oriented perpendicular to the contact with layered sodic aplite to produce a comb structure (Plate 7d). In other parts of the Contact Beryl Occurrence the potassic pegmatite is more equigranular and the K-feldspar is deep-pink (see Plate 8a, next section).

Commonly, GLB-7 is paired with sodic aplite which may vary from a 1 cm wide layer to a multi-layered rock up to 50 cm thick. The latter is intricately folded in places (Plate 7d). The delicate internal laminations are produced by concentrations of fine-grained green muscovite, red garnet or black tourmaline.

Accessory minerals in GLB-7, listed in order of increasing abundance include columbite (Plate 7e), green beryl, garnet, tourmaline and green muscovite.

Post Magmatic Replacement Units

Replacement units related to the Ghost Lake Batholith show their most impressive development at the Contact Beryl Occurrence. These consist of albite-rich replacement of potassic pegmatite, muscovite-

rich potassic pegmatite, and fine-grained leucogranite. Greisen alteration also appears as local internal units (autometamorphic greisen of Shcherba 1970, p.139) and in a more extensive exogreisen vein system (Shcherba, *ibid*). These units appear to most closely conform to the "amphibolite greisen formation" in the classification of Shcherba (*ibid*).

Albite-rich replacement units: these rocks are incipiently developed in the pegmatitic granites in two settings: (1) along contacts between the batholith and amphibolitized mafic metavolcanics, and (2) in an exocontact dyke network. In the first setting, replacement occurs as a 10-100 cm wide garnet-muscovite-tourmaline-quartz-albite border unit (albitized potassic pegmatite containing 73 percent albite). Textures indicate a progressive conversion of blocky K-feldspar in a beryl-tourmaline-muscovite potassic pegmatite into a tourmaline-quartz-albite rock as the batholith contact is approached (Plate 8a). This disappearance of K-feldspar is complemented by a biotite-rich halo in the adjacent amphibolitic host rocks, which also suggests a metasomatic origin for this albite-rich unit (Plate 8b). The exocontact dykes (Plate 8c) are 10-30 cm wide, and are abundant in one locality (8, Figure 13). These rocks normally consist of a tourmaline-quartz-albite assemblage which grades along strike into beryl-bearing albitite (Table 8), or interconnects with pyrite-tourmaline-quartz veins, locally containing small pods (up to 50 by 100 cm) of approximately 90 percent beryl. Skeletal tourmaline intergrowths with albite are another notable textural feature in the albite-rich dykes (Plate 8d).

PLATE 8

- a. Textural evidence supporting the metasomatic transformation of potassic pegmatite (P), containing pink K-feldspar, into an albite-rich rock (A) which contains subordinate tourmaline, quartz and K-feldspar remnants at locality E31-2. Coin diameter = 2.4 cm.
- b. Biotite-rich metasomatic selvage developed in amphibolite next to albitized potassic pegmatite of the Ghost Lake Batholith at locality E31-2.
- c. Transition of an albite-beryl-muscovite autometamorphic greisen (G) into bifurcating beryl-tourmaline-albite exocontact dykes (D) at locality B in Figure 11. Hammer length = 34 cm.
- d. Skeletal tourmaline, cored by albite, distributed within an albite-rich replacement unit at Locality A in Figure 14. Coin diameter = 1.8 cm.
- e. Detail of autometamorphic greisen unit in Plate 8e showing alternation of coarse-grained layers of white beryl, albite and muscovite layers with finer-grained green muscovite-dominated layers. Arrow points to a beryl grain. Pencil = 15 cm length.

PLATE 8

Table 8. Modal Analyses of Pegmatitic Granites and Related Metasomatic Suites from the Contact Beryl Occurrence.

	PRIMARY PEGMATITIC GRANITIC UNITS (Locality A)				REPLACEMENT UNITS (Locality B)			
	Pegmatitic ¹ Leucogranite GLB-4	Fine-grained ² Leucogranite GLB-5	Muscovite- Rich Potassic Pegmatite GLB-6	Potassic ¹ Pegmatite GLB-7	Autometamorphic Griesen ¹	Exocontact Beryl Albitite Dike ¹	Contact Albitized Potassic Pegmatite ¹	Partially Albitized, Fine-Grained Leucogranite ²
Quartz	32	35	29	24		4	22	26
Plagioclase	19	23	57	31	5	57	73	49
K-Feldspar	43	27	11	41				21
Muscovite	6	14	X	3	85		X	X
Tourmaline	1	1	2	1		8	5	4
Garnet		X	X				X	X
Beryl				X	10	32		
Cassiterite					X			
Chrysoberyl					X			
TOTAL	100	99	99	100	100	101	100	100

X - present in trace amounts

1 Analyses undertaken on clean fresh outcrop surfaces covering a 3 x 3 metre area and 5 cm grid spacing for the coarse-grained primary units and 0.5 x 0.5 m area (1 cm grid spacing) for the relatively finer-grained replacement units.

2 thin section modal analysis

interacted chemically with amphibolitic mafic metavolcanic host rocks, resulting in complementary 1-5 cm wide, brown weathering selvages enriched in K_2O , Li, Rb, Cs, F and Sn (Chapter 7). Most fluids were probably channeled in the host rocks to locality A in Figure 13 which represents the beginning of the greisen vein system.

Autometamorphic greisen: Autometamorphic greisens occur as local lenses up to 4.5 m by 0.4 m, as at locality B (Figure 13, Plate 8e). These units are enclosed within and are gradational into a small exocontact pegmatitic granite lens (3 m by 14 m) composed of albitized, muscovite-rich potassic pegmatite which envelops a core of fine-grained leucogranite. The eastern end of the muscovite-rich potassic pegmatite unit grades into albite-rich dykes with black tourmaline selvages, that also contain appreciable beryl (Table 8). The western end of this pegmatitic pod grades into a similar muscovite-rich greisen, which contains appreciable pyrite, tourmaline and biotite.

The greisen pods are strongly foliated and lineated, fine to coarse grained, and yellow-green due to the presence of 85 percent muscovite. They contain white and light grey, euhedral to subhedral, beryl crystals up to 2.5 cm long. Lesser amounts of anhedral albite and accessory chrysoberyl and cassiterite are also present (Table 8). Chrysoberyl ($BeAl_2O_4$), which occurs as 1-2 mm yellow-green anhedral grains, is a relatively rare mineral in pegmatitic rocks and is apparently restricted to desilicated or contaminated pegmatites (Burt 1982, p.139).

Exogreisen vein system: forms a bifurcating vein system, mostly

conformable to host foliation, which is in contact with pegmatitic granite at two points (localities A and B in Figure 13). This system is traceable for at least 0.65 km eastward from Locality A and varies in width from 1.5 to 10 km.

The exogriesen vein system begins as a 60 m long by 2-3 m wide, highly mafic, actinolite-biotite-tourmaline zone (Unit 4, Figure 13, and Plates 9a and 9b), which is tangential to the batholith contact at Locality A. Most of the vein system however is a rusty-weathering tourmaline-pyrite-muscovite unit (Unit 5, Figure 13), which is well foliated, fine to medium grained and has sharp boundaries with its host.

The vein system mainly straddles the lithologic contact between black hornblende-plagioclase amphibolite and green, mafic to intermediate, actinolite-plagioclase-bearing metavolcanics. A subordinate branch crosses the metavolcanics, establishing a direct link with the pegmatitic granites at locality A. The mineralogy of this branch changes systematically with increasing distance from the batholith contact, from (1) biotite - actinolite - tourmaline - (quartz - albite) (CI 90-95), to (2) albite - quartz - tourmaline (CI 80-90), to (3) tourmaline - pyrite - albite - holmquistite - chlorite - actinolite - biotite (CI 70) and finally to (4) tourmaline - chlorite - albite - chlorite - pyrite (CI 50-60).

The remainder of the vein system is composed of an anastomosing network of smaller veinlets in actinolite-rich, mafic to intermediate, metavolcanic rocks or in a holmquistite-biotite-actinolite-plagioclase

PLATE 9

- a. Actinolite-biotite-tourmaline rich exogreisen zone at Locality A in Figure 14, the point where this unit leaves the Ghost Lake Batholith (lower right). Extension fractures contain quartz-albite-beryl veins.
- b. Detail of The exogreisen zone in Plate 9a revealing actinolite (marked by 2.6 cm diameter coin), light brown biotite and radiating aggregates of black tourmaline.
- c. Interconnected holmquistite-biotite-plagioclase veinlets near the mafic exogreisen assemblage of Plate 9b. This veinlet system is post-dated by a tourmaline-beryl-albite dyke in the lower right. Coin diameter = 2.4 cm.
- d. Porphyroblastic intergrowths of coarse-grained anthophyllite, actinolite, holmquistite and biotite in a medium-grained chlorite-biotite-plagioclase rock along the southern contact of unit 4 (Figure 13) at its westernmost exposure. Coin diameter = 2.6 cm.
- e. Exocontact dyke composed of quartz, albite and tourmaline, situated 30 m from the Ghost Lake Batholith contact at locality E31-2. Hammer length = 34 cm.
- f. Zoned exomorphlic selvage developed along a black tourmaline-albite exocontact dyke situated 20 m from the Ghost Lake Batholith at locality E31-2. Arrow points to an inner zone rich in black tourmaline which is enveloped by a more continuous biotite-rich zone. Note the presence of skeletal tourmaline in the dyke.
- g. Tourmaline replacement of medium grade clastic metasediments adjacent to a narrow, discordant quartz at Dryden (locality F24-8). Note the overprinting, by tourmaline, of the host foliation (parallel to 15 cm long pen) and porphyroblasts of andalusite and cordierite (near arrow).

PLATE 9

rock (Plate 9c). The veinlets conceivably developed during the same fluid transmittal event as the larger units of the vein system, since they also contain holmquistite and biotite. The veinlets occur up to 5.5 m from the larger veins.

Both scales of vein system are cross-cut by tourmaline-quartz-albite dykes (10-30 cm in width), which locally contain light blue to white beryl. Some dykes exhibit boudinage.

Holmquistite distribution: the lithium-bearing amphibole holmquistite [$\text{Li}_2(\text{Mg}, \text{Fe}^{2+})_3\text{Al}_2\text{Si}_8\text{O}_{22}(\text{OH})_2$] at the Contact Beryl Occurrence varies in size from fine-grained up to 1.5 by 7 cm. It is readily recognized in the field, where replacement of the precursor amphibole has been relatively complete (type locality 4 below), by a distinctive violet cast. More commonly, however, the distinctive violet colours are muted by a complex intergrowth of the holmquistite with biotite, chlorite, tourmaline, actinolite and anthophyllite.

Holmquistite was observed in four specific settings within and near the exogreisen vein system:

- 1) with biotite-actinolite-plagioclase and plagioclase-biotite-Fe-oxide rocks in a system of closely spaced, anastomosing, 1-3 cm wide veinlets hosted in actinolite-rich mafic to intermediate metavolcanics (Plate 9c);
- 2) with actinolite, anthophyllite, biotite and chlorite in porphyroblastic segregations developed in a medium- to coarse-grained biotite-plagioclase rock of possible clastic metasedimentary origin (Plate 9d);

- 3) with pyrite, biotite, tourmaline and albite in a unit traceable for at least 650 m within the main exogreisen vein system; and
- 4) in thin biotite-holmquistite sheets, oriented at a high angle to foliation and situated 2 m from the main exogreisen vein system.

Type 3 holmquistite, in the main exogreisen zone, has a well-developed mineral lineation concordant to the regional lineation. In type 1 and 2 occurrences, holmquistite is randomly oriented on foliation surfaces and probably developed after the regional lineation. Type 4 holmquistite is obviously later than this deformation event, however, its temporal relations with types 1 and 2 are uncertain. Multiple generations of holmquistite have also been suggested elsewhere. For example, Morgan and London (1987) proposed three holmquistite-generating events in exomorphic haloes around the Tanco Pegmatite, west of the study area in the Bird River Greenstone Belt of Manitoba.

Several planar to slightly boudined albite-rich dykes (Plate 9c) post-date the formation of types 1, 2 and 3 holmquistite. These garnet-tourmaline-quartz-albite dykes vary from 10 to 30 cm in width and locally contain light blue to white beryl. Biotite-rich selvages to these dykes, 1 to 4 cm in width, overprint the type 1 holmquistite-bearing veinlet system.

Zealand Stock

This 0.8 by 2.7 km ovoid pluton in Concession VII of Zealand Township was first mapped by Satterly (1941). The pluton contains a suite of intrusive units similar to GLB-4, -5, -6 and -7 of the Ghost

Lake Batholith. In view of its proximity to the inferred western extension of the axial trace of the Thunder Lake syncline (Figure 11) this stock may represent a phacolithic apophysis from the nearby GLB. The Zealand Stock and the adjacent GLB is considered to represent the most plausible source area for rare-element pegmatites in the nearby Mavis Lake pegmatite group, in the view of the relatively advanced degree of chemical fractionation and sporadic presence of rare-element mineralization, discussed in ensuing sections (this chapter and Chapter 7).

Barren Exocontact Granitoid Dykes

Coarse-grained to pegmatitic dykes enriched in tourmaline and/or muscovite, plagioclase and quartz (Plate 9e), possibly related to the Ghost Lake Batholith and Zealand Stock, are particularly prevalent south of these plutons between Dryden and Thunder Lake. Near the GLB contact exocontact dykes tend to be rich in albite and skeletal tourmaline and are characterized by zoned exomorphic selvages (Plate 9f). These dykes are almost invariably barren of rare-element minerals; however at locality F24-8 near the Dryden High School sparse, 5 mm wide, white beryl were found in several muscovite-rich dykes.

The immediate wall rocks, medium grade clastic metasediments, commonly exhibit narrow haloes of tourmaline replacement, first described in the area by Satterly (1941, p.25-26). He indicated that tourmaline replacement is very common in Zealand Township (lots 15 to 19, Concession VI and VII). This phenomenon also occurs adjacent to tourmaline-bearing quartz veins, where replacement of andalusite and cordierite porphyroblasts and enclosing matrix is evident in some

metapelitic units (Plate 9g). Tourmalinization is additionally commonplace adjacent to virtually all rare-element pegmatites of the Mavis Lake pegmatite group and will be described more fully in a later section. The above features are interpreted to represent boron-degassing of a relatively high level of the Ghost Lake Batholith and may be accompanied by appreciable fluorine (up to 1.8 percent; Table 45, Chapter 7) as in biotite-rich mafic rocks at the Contact Beryl Occurrence.

Modal Variation of the Ghost Lake Batholith and Zealand Stock

Sixty modal analyses are presented in Appendices 3 to 7. A salient feature of the modal variation of quartz, plagioclase and K-feldspar shown in Figure 14 is the wide range in K-feldspar/total feldspar ratios, resulting in a compositional variation between granite and leucocratic, albite quartz diorite. This variation compares closely with the general range for peraluminous granitoid rocks in the ERS (Figure 5).

GLB-1, the most abundant plutonic unit, is almost exclusively confined to the granite field of Figure 1, with minor exceptions represented by cordierite-bearing granodiorite and biotite-muscovite quartz syenite. In contrast, rocks constituting most of the eastern lobe of the Ghost Lake Batholith (GLB -4, -5 and -6) exhibit relative enrichment in plagioclase lying in the granodiorite field and plagioclase-rich portion of the granite field. Only a minor overlap with the main GLB-1 field is evident in Figure 14.

GLB-3, almost exclusively granite, forms the most restricted compositional field of any major plutonic unit from the batholith and

Figure 14. Modal Variation of Quartz-Plagioclase-K-feldspar in Plutonic Units of Ghost Lake Batholith, Zealand Stock and Related Exocontact Pegmatites.

overlaps with the main GLB-1 field and that of the eastern lobe units (GLB-4, -5, -6, and -7). GLB-2 appears to be predominantly granite, however granodiorite compositions also appear.

Specimen F16-4 (Appendix 6) contains lowest quartz content of any rock from the Ghost Lake Batholith and plots in the granite field of Figure 14. Such examples of quartz impoverishment, coupled with Al_2O_3 enrichment may represent zones of feldspar accumulation.

The few modal analyses in Appendix 8 of exocontact granitoid dykes related to the batholith appear to define two compositional trends related to the general grade of regional metamorphism of the host rocks. Dykes and masses emplaced in medium-grade metasediments (F22-1, F22-6D and D34-3) exhibit a marked tendency toward lower K-feldspar/total feldspar ratios, and a tendency toward quartz and muscovite enrichment relative to dykes emplaced within the high-grade metasedimentary migmatite terrain situated immediately west of the batholith in Sanford Township (F9-1 and F9-2 in Appendix 8). The latter also tend to show biotite exceeding muscovite and sporadic presence of cordierite-quartz symplectitic aggregates which are apparently unique to dykes emplaced in high-grade host rocks. Significant differences in trace lithophile element content of whole rock and mineral samples (Chapter 7) also distinguish the two groups.

RARE-ELEMENT PEGMATITES OF THE MAVIS LAKE GROUP

Detailed petrographic and geological descriptions of all known pegmatite types within the Mavis Lake pegmatite group are presented in this section, as a function of increasing distance from the Ghost Lake Batholith.

Dryden Airport Beryl Occurrence

Lime-green primary beryl is sporadically evident in some potassic pegmatites near the southeastern end of the main Dryden Airport runway and was discovered by A. Pryslak in 1979 (A. Pryslak, Selco Incorporated, personal communication, 1980). The large elevated outcrop here consists of a complex assortment of units GLB -4, -5, -6 and -7 of the Ghost Lake Batholith and is identical to the Greer Lake, Manitoba, beryl-garnet-cordierite pegmatitic granite except for absence of cordierite (P. Cerny, University of Manitoba, personal communication, 1980; Cerny et al. 1981). Beryl is entirely confined to unit GLB-7, which intrudes fine-to medium-grained, massive to vaguely banded garnet muscovite-tourmaline fine-grained leucogranite (GLB-5) and composes most of the outcrop (specimen E31-9A in Appendix 7 and Plate 5e). In addition to euhedral beryl crystals measuring up to 3.0 by 8.5 cm on outcrop surface, the potassic pegmatite contains blocky K-feldspar, green muscovite, albite and black tourmaline (Plates 6d and 10a). Occasional quartz-rich pockets in contact with well-developed blocky K-feldspar up to 15 x 15 cm tend to be interspersed towards the centres of the potassic pegmatites. No columbite-tantalite group minerals have been identified.

PLATE 10

- a. Yellow-green, euhedral beryl crustal with tourmaline, muscovite and albite at the edge of a quartz-rich core zone in potassic pegmatite (GLB-7) at locality E31-9. Coin diameter = 2.4 cm.
- b. Muscovite-beryl-tourmaline-quartz-K-feldspar assemblage from Taylor Pegmatite dyke 2. B=beryl, K= K-feldspar, A= albite, T = tourmaline and Q = quartz.
- c. Small pocket of quartz-albite aplite (Ap) replacing muscovite-beryl-tourmaline-albite-quartz K-feldspar pegmatite unit of the Taylor Pegmatite dyke No. 2. Dashed line represents the cryptic border between fine-grained albite and coarser-grained, primary albite (A). K= stained K-feldspar, Q = quartz, T = tourmaline, M = muscovite.
- d. Partially vitreous emeralds in a beryl-apatite-albite-quartz-tourmaline assemblage, developed adjacent to the contact of Taylor Pegmatite dyke 2 with metaultramafic rocks.

PLATE 10

Contact Beryl Occurrence

This minor beryl occurrence was discovered by the writer in 1983; geology and modal variation have been described in a preceding part of this chapter. The importance of the occurrence lies in the association of beryl with several primary and replacement units of the Ghost Lake Batholith. Chemical and modal analyses of the important rock units are given in Tables 8 and 45.

Taylor Beryl Occurrence

This occurrence of rare-element mineralization, located on Lot 17, Concession VII of Zealand Township, is composed of two pegmatitic dykes situated 340 m south from the Ghost Lake Batholith contact (Figure 10, back pocket). Date of discovery is unknown, however it is believed to have been found circa 1940 by J.G. Taylor who submitted specimens of beryl to Satterly (1941, p.55) which purportedly came from the same area. Investigation of the mineralization was apparently undertaken only by a few shallow blast pits.

Detailed maps of both dykes are presented in Figures 15 and 16. These pegmatites, which are approximately 60 m apart are herein designated as Dyke 1 and Dyke 2. Both dykes occur in contact with a 45 m wide talc-rich metaultramafic unit which was delineated for 1.2 km by Satterly (1941), and is intercalated with low-to medium-grade mafic metavolcanics.

Pegmatite dyke 1: this pegmatite dyke was emplaced concordant to the steeply-dipping northern contact between metaultramafic and mafic metavolcanic units for 52 m along strike and varies between 2.4 and 8 m in breadth (Figure 15). Dips along the southern,

interdigitating contact vary between 75°S and 76°N. Along the southern contact in the metaultramafic unit, conspicuous concentrations of phlogopite, anthophyllite and holmquistite up to 1 m in width attest to the intensity of contact metasomatism. Adjacent portions of the dyke exhibit increased levels of black tourmaline, blue apatite and occasional red-brown garnet (Table 9).

Four main pegmatitic units are recognizable in Dyke 1:

- 1) graphic quartz and K-feldspar - (tourmaline) pegmatite,
- 2) garnet - tourmaline - albite - quartz - (K-feldspar) pegmatite,
- 3) garnet - tourmaline - quartz - albite aplite, and
- 4) quartz-rich core zones

Graphic quartz - K-feldspar pegmatite (unit 1) occupies the largest part of Dyke 1. Block-shaped quartz - K-feldspar aggregates up to 0.7 by 1 m, embedded in a medium-grained tourmaline-quartz-albite matrix, typify this unit. Small quantities (<1 percent) of euhedral black tourmaline are disseminated throughout the graphic zone, but in lower concentrations than in that of the tourmaline pegmatite unit.

Euhedral, white to faintly green and locally vitreous prisms of beryl up to 1 x 3 cm occur sporadically in this zone, particularly within the small blast pit near the southern dyke contact (Figure 15). A modal analysis of material from this area is presented in Table 9.

Garnet-tourmaline-albite-quartz pegmatite (unit 2) contains the coarsest and greatest abundance of tourmaline in Dyke 1, with maximum dimensions of single crystals up to 3 cm in diameter and 14 cm in length. Locally tourmaline exceeds 20 percent. Light pink albite constitutes the most abundant mineral (Table 9); K-feldspar is

Table 9. Modal Analyses from Beryl and Spodumene Pegmatites at Taylor Occurrence and Fairservice Property Of Mavis Lake Group.

TAYLOR BERYL PEGMATITES				NORTH ZONE OF FAIRSERVICE PROPERTY				NORTHEAST ZONE	
Mineral	Albite-Rich Zone Sample No.	Beryl Sample No.	Spodumene- Rich Zone Sample No.	Albite Sample No.	Quartz Sample No.	Spodumene Zone Sample No.	K-feldspar-Rich Spodumene Zone Sample No.	Permit No.	No.
Quartz	1	1	1	47	37	29	23	15	
Albite	1	1	1	4	7	26	13	8	
Spodumene	4	1	1	5	1	55	55	3	
Green Muscovite				4	7	7	3	X	
Beryl				1	1		X	1	
Albite	1	1	1						
Spodumene	1	1	1						
Quartz	1	1	1						
Albite				1	49	39	7	X	
Spodumene								53	
Albite									
								100

Figure 15. Detailed Geology of Part of Pegmatite Dyke 1, Taylor Beryl Occurrence and Location of Analyzed Samples. Mapped Area Located on Figure 38, Back Pocket.

comparatively rare. Red-brown euhedral garnets up to 7 mm diameter and fine-grained, film-like green muscovite aggregates are minor accessories. Rare light-green beryl and columbite-tantalite group minerals were recognized in one portion of the largest unit 2 (Figure 15).

Euhedral crystals of columbite-tantalite up to 3 x 8 mm (located by arrow in Figure 15) are associated with an albite-rich portion of tourmaline pegmatite, marked by replacement of K-feldspar by fine-grained albite along cleavages and outer crystal contacts. Several smaller possible columbite-tantalite grains between 1 and 3 mm diameter were also noted near the larger crystal.

Aplite (unit 3) is either intimately interlayered, or in contact, with tourmaline pegmatite units. These light pink rocks are massive to vaguely banded and consist essentially of quartz and albite. Garnet and tourmaline in the 1-4 mm diameter range are disseminated throughout but may be especially concentrated along contacts with tourmaline pegmatite units.

Unit 4 is distributed as small quartz-rich pods up to 1 x 3 m and is mainly confined to or in contact with Unit 1 (Figure 15). Unit 4 contains 70-90 percent massive quartz and sporadic, moderately pink, perthitic, blocky K-feldspars up to 13 x 19 cm. K-feldspar usually exhibits incipient albitization guided by cleavages and outer crystal surfaces. Rare subhedral white beryl crystals up to 3 cm across occur in some core zones.

Pegmatite dyke 2: this beryl-bearing dyke is of particular significance in that it represents the only known emerald occurrence in the Superior Province of Ontario. The pegmatite is interpreted as an intersecting system of dykes (Figure 16) entirely contained within the same metaultramafic unit which hosts Dyke 1. The host rocks are rusty-brown weathering, massive ultramafics containing a relict, medium-grained, equant mineral which represent serpentine pseudomorphs possibly after olivine. Beryl occurs only in the northeast-striking dyke which is crudely zoned in that beryl is significantly concentrated along the east- to moderately southeast-dipping contact (beryl-muscovite-K-feldspar quartz-tourmaline-albite assemblage) while the remainder of the pegmatite here is impoverished in beryl and K-bearing minerals (quartz-tourmaline-albite-apatite assemblage). The contact has to have been warped possibly by pegmatite intrusion and is featured crenulations in the adjacent metasomatic aureole (Figure 16); having fold axes of variable orientation but distinct from the regional lineation attitude (055°/311°).

Intense metasomatism is evident along this contact, where phlogopite-rich rock is conspicuous at distances up to 10 cm from the contact. The ultramafic host has been converted into a medium-grained, highly schistose, black rock composed of 95 percent phlogopite with minor plagioclase and randomly oriented, glossy black tourmaline prisms. Inclusions of similar metasomatic rocks occur within the pegmatite and are either identical to the contact zone or are more plagioclase-rich with purplish lenticles of fine-grained holmquistite.

Metasomatism is also apparent along the western contact of the NNE-striking dyke. Here, coarse-grained tourmalinite, is impressively developed over a 10 cm width, is composed of 80 percent black medium-grained tourmaline, minor blue-green apatite, quartz, biotite and albite (Table 9).

The beryl-bearing assemblage (unit 4a, Figure 16) is fine-to medium-grained and light orange except along the southeast contact where black tourmaline predominates (Plate 10b). Muscovite is present in minor quantities as faint green subhedral books up to 2 x 7 cm. Pink subhedral blocky K-feldspar attains maximum dimensions of 12 x 14 cm. Anhedral K-feldspar masses also occur in the tourmaline-rich area, interstitial to and containing inclusions of euhedral tourmaline, apatite and occasionally quartz.

Quartz is typically milky-white but locally possesses striking dark smoky-grey colour. This mineral may have in part formed early, because euhedral to subhedral crystals occur as inclusions in tourmaline-K-feldspar segregations.

Tourmaline is prevailingly euhedral, glossy black, fine-to medium-grained with maximum diameter of 2.5 cm. Very fine-grained tourmaline inclusions occur in some beryl crystals.

Albite is commonly pink to pink-grey, equigranular, subhedral, and 1-1.5 cm long. A striking grey-green albite is developed in the tourmaline-rich border unit along the southeast contact. This type of albite occurs in coarse, anhedral, irregularly shaped crystal aggregates typically 3 x 6 cm, interstitial to relatively early quartz, tourmaline, beryl and apatite.

Most albite in the two main mineral assemblages appears to be of primary origin. Local albitic replacement, however, has been noted in three settings:

- 1) discontinuous development along edges and cleavage traces of blocky K-feldspars in the beryl zone;
- 2) local veining of euhedral beryl by fine-grained, equigranular albite;
- 3) small circular to ovoid pockets of fine-grained quartz-albite-green (chromian?) mica enclosed by the primary assemblage beryl-K-feldspar-quartz-tourmaline-albite.

The small pockets of albitic replacement vary in size from 5 cm in diameter to 4 x 10 cm (Plate 10c). These pockets contrast strongly with the surrounding primary assemblage owing to the greatly diminished grain size, virtual lack of muscovite and tourmaline, and sporadic appearance of green mica. In some cases, euhedral tourmaline crystals can be clearly observed to be truncated by the albitic pockets. Also, euhedral beryl crystals display incipient albitic replacement in proximity to these pockets. Euhedral outlines become noticeably corroded and penetrated by albite. There appears to be a correlation between discontinuously developed opaque emerald coloration in beryl and albitization disposed in the small pockets. Rare quantities of columbite (analyses 409A and B, Appendix 39) up to 1 mm were detected on polished slabs of the albitic pockets.

Apatite, 0.5 mm to 2.5 cm, is a minor but widespread accessory in Dyke 2. It is particularly concentrated in the tourmaline-rich zone along the southeast contact where modal abundance reaches five percent. It apparently crystallized early, occurring as euhedral, green-blue inclusions in beryl, tourmaline, quartz, albite and K-feldspar.

Beryl is disseminated mainly within a 0.3-1 x 4 m area and coincides with high tourmaline abundance. Beryl is predominately euhedral, opaque white to faint green, with maximum observed diameter of 1.8 cm and a maximum length of 2.3 cm. About ten percent of the beryl population, however, is a vitreous emerald-green variety (Plate 10d). The emeralds tend to be finer grained than the white to faint green beryls and are localized along the warped, southeast dipping contact or, more rarely, embedded in phlogopite-rich contact metasomatic rocks.

Most emeralds examined on polished slab surfaces are cloudy and variably fractured. In rare instances however vitreous areas up to 0.6 x 1.5 cm are apparent (Plate 10d) and if it were not for presence of interconnecting fractures, these emeralds would possibly constitute gem material.

The remaining parts of Dyke 2 consist of unit 4b (Figure 16), a fine-to medium-grained albite-rich pegmatite composed mainly of the assemblage quartz-tourmaline-albite-(garnet). Beryl, K-feldspar and muscovite are virtually absent. Subhedral bluish-white equigranular albite up to 2.5 cm in diameter constitutes the dominant mineral. Albite also occurs with quartz in pods of fine-grained aplite. Black tourmaline up to 2.5 cm and typically 0.5 cm in diameter is disseminated throughout this zone. Trace amounts of fine-grained garnet occur sporadically in patches.

Fairservice Property Spodumene Pegmatites

Spodumene is present in rare-element pegmatites of the Mavis Lake area, 4.4 km from the Ghost Lake Batholith contact (Figure 10). Most spodumene pegmatites are confined to claims owned by Mr. R. Fairservice and are situated in four main areas herein denoted as the North, South, East and Northeast Zones (Figure 17, back pocket). Detailed mapping by Pryslak (1981) and the writer has delineated eighteen distinct spodumene pegmatitic lenses (numbered in Figure 17) which are generally concordant to host rock foliation at surface. Contacts, however, are typically undulating in the limited vertical sections observed and also in diamond drill sections (A. Pryslak, personal communication, 1981). In the North Zone, for example, two spodumene pegmatite lenses exhibit variable dip values of 42° - 75° N and 49° S- 80° N. Spodumene pegmatites sharply discordant to host rock foliation occur only at one locality (Pegmatites 11 and 12, Table 10 and Figure 17).

There is no appreciable change in primary mineral assemblages of the Mavis Lake spodumene pegmatites as a function of increasing distance from the Ghost Lake Batholith i.e. traversing from the North to East Zone (Figure 17). However, the degree of post-magmatic albitization increases significantly to the east. In the North Zone, spodumene pegmatites contain 10 to 60 percent of albite-rich replacement units whereas in the East Zones, pegmatite sheets contain only a very low percentage of recognizable primary assemblage units.

North Zone: a detailed 1:480 scale geology of the North Zone is given in Figure 18, backpocket). Internal zonation within individual Mavis Lake area pegmatites is generally only crudely developed and

Table 10. Dimensions and Attitudes of Spodumene Pegmatites from the Fairservice Property of the Mavis Lake Group.

Area	Pegmatite	Length (m)	Maximum Width (m)	Strike	Dip
NORTH ZONE	No. 1	78	15	0930-1170	450 - 500 N (north contact)
	No. 2	280	15	0930-1100	600 N to vertical (south contact)
	No. 3	67	10	0750-1030	620-800N (locally 550S) along north contact
	No. 4	185	35	0900-1000	250-450S to 480-600W along south contact
SOUTH ZONE	No. 5		12.2*	0820-1100	620S - vertical (north contact)
	No. 6	>11	6	0350-1300	400-770S to 800W (south contact)
	No. 7	>75	15	1040	720N (contact with inclusion)
	No. 8	18	3	1180	820S (south contact)
	No. 9	107	4	0880-1030	500W
	No. 10	>15	3	0810-0940	400-700S (north contact)
	No. 11	>46	6	0900	420-740N (north contact)
	No. 12	>50	0.6	0600	450N*
	No. 13	60	6	0900	820N
	No. 14	erosional remnants of flat lying sheet traceable for 320 m	0.3	0800	?
EAST ZONE	No. 15	75	0.3	0080	280S*
	No. 17	33	2	0040	550N*
NORTHEAST ZONE	No. 18	215	18	1150	800W
	No. 19			1600	600W
	No. 20			1550	500-700 SE
	No. 21			1600	350 SE
	No. 22			1550	300SE
	No. 23			0450	?
	No. 24			1350	600NE

*asterisks denote data obtained from diamond drill hole sections of Prystak (1981)

often obscured by late-stage albitization. The best example of internal zonation is provided by Number 1 pegmatite from the North Zone of the Fairservice Property (Figure 19). This 12 by 76 m pegmatite with $\text{Li-Be-Ta} > \text{Nb}$ contains the following zones, from the contacts inwards:

- 1) potassic feldspar-enriched spodumene pegmatite (Plate 11a),
- 2) spodumene-enriched albite-quartz pegmatite (Plate 11b), and
- 3) quartz-rich core zone (Plates 11c, d, 12a, b).

A characteristic texture occurring solely in the core zone features medium green spodumene megacrysts (up to 4 x 8 cm) partially to completely enveloped by fine-grained albite coronas up to 2 cm wide (Plate 12c). These coronas, which contain a few muscovite books and rare tantalite, probably relate to late stage albitic replacement. Glacier-blue to white euhedral beryl constitutes a common accessory in the core zone (Plate 11d). This beryl, which achieves maximum dimensions of 3 x 7 x 2 cm is considered primary. It is typically isolated in quartz and not in contact with any albite. Such beryl is much coarser and exhibits a greater degree of euhedralism than secondary, white, anhedral beryl associated with albite-rich replacement units in the same pegmatite (Plate 12d).

The quartz-rich core zone is discontinuously developed as a series of irregular pod-like domains up to 2 x 9 m which are typically, although not strictly, confined to the central part of Pegmatite 1. Most pods are dominated by light to medium grey massive quartz (70 to 80 percent), with ancillary amounts of beryl, light-green spodumene, green muscovite and blue apatite. A subtle local

Figure 19. Detailed Geology of Part of Pegmatite 1, North Zone of Fairservice Spodumene Pegmatites Showing Localities of Analyzed Samples.

PLATE 11

- a. K-feldspar-rich spodumene pegmatite zone in Fairservice Pegmatite No.1. The dark areas represent spodumene-quartz graphic intergrowths. Compass diameter = 7 cm.
- b. Albite-spodumene-quartz pegmatite zone, developed peripherally to the quartz-rich core zone of Fairservice Pegmatite No.1. S = spodumene, Q = quartz, A = albite and K-feldspar.
- c. Light green spodumene, marked by 2 cm diameter coin, completely enveloped by albite - quartz corona in the quartz-rich core zone of Fairservice Pegmatite No.1.
- d. Light-blue, euhedral crystal of beryl occurring adjacent to spodumene megacrysts in the quartz-rich core zone of the Fairservice Pegmatite 1. Coin diameter = 2.4 cm.

PLATE 11

foliation is mainly defined by preferred orientation of lenticular aggregates of fine-grained green mica.

Blocky K-feldspar up to 33 x 41 cm are conspicuously developed in the core zone, occurring as crystal aggregates (Plate 12a) or as single euhedral crystals (Plate 12b). Both types of K-feldspar display marginal albitic replacement which is best seen in stained K-feldspar crystals (Plate 12a).

Conspicuous masses of fine-grained, yellow-green mica accompanied by minor orange garnet, black tourmaline and albite are sporadically developed in the core zone. In some cases, these aggregates apparently replace portions of blocky K-feldspar (Plate 12b).

Tantalite crystals and clusters (analysis 406, Appendix 39), the latter up to 0.5 cm x 1.5 cm, are most conspicuous in the core zone of Pegmatite 1, where they commonly reside in albitized margins of K-feldspar aggregates and spodumene phenocrysts.

Outer margins of the core zone in places grade into the albite-spodumene-quartz pegmatite unit. The quartz core zone boundary is however more typically marked by a metasomatic halo (<1 m in width) in which all primary minerals are variably replaced by a fine-grained green mica-garnet-quartz-albite assemblage.

Enveloping albite-spodumene-quartz pegmatite contains the highest quantities of spodumene observed in the Mavis Lake area. This mineral varies between 29 and 48 percent in Pegmatites 1 and 4 (Table 9) and achieves a maximum length of 1 m (parallel to the c-axis) and

PLATE 12

- a. Quartz-rich core zone showing an irregular aggregate of K-feldspar (stained orange) with subordinate discontinuous peripheral replacement by light pink albite at the Fairservice Pegmatite 1. Subhedral, white beryl (B) and tantalite (T) are associated with the albite replacement areas. Coin diameter = 1.8 cm.
- b. Euhedral blocky and irregularly-shaped K-feldspar megacrysts from the quartz-rich core zone of the Fairservice Pegmatite 1. Coin diameter = 2.4 cm. M = green muscovite-rich masses, Q = quartz, S = spodumene, K = K-feldspar and A = albite.
- c. Part of a spodumene megacryst, marked by 1.8 cm diameter coin, containing intergrown droplets of quartz and enveloped by a quartz-muscovite rich corona. Fairservice Pegmatite No.2.
- d. Albite-rich replacement unit in K-feldspar-rich spodumene pegmatite at the Fairservice Pegmatite No.1. Two crystals of white, secondary beryl are indicated by arrows. Coin diameter = 2.4 cm.
- e. Tourmaline porphyroblasts and holmquistite-rich segregations in mafic metavolcanics adjacent to albite-rich unit of Fairservice Pegmatite No. 4. Coin diameter = 2.4 cm.

PLATE 12

width of 9 x 12 cm. In some coarse spodumene crystals remnants of hiddenite, the vitreous emerald-green spodumene variety, were recognized. In coarse-grained spodumene-rich pegmatite at Pegmatite 6, pale green, white and light orange spodumene were noted. The latter colour may relate to iron absorption from groundwater. In very rare instances, white to pale-green vitreous spodumene crystals up to 3 x 7 cm are, partially to completely enveloped by fine-grained aplite, as at the western end of Pegmatite 2.

Quartz contents can also be appreciable, reaching 47 percent in that part of albite-spodumene-quartz pegmatite nearest the quartz-rich core zone in Pegmatite No.1.

Pockets of sodic aplite up to 1 x 3 m occur scattered throughout the spodumene-rich zone and are probably primary as no replacement textures are discernible. In Pegmatites 2 and 6, pods and layers of similar aplitic material show no evidence of albite replacement even where in sharp contact with spodumene. Blue apatite is a common accessory which is, as at the western end of Pegmatite 2 and in Pegmatite 1 accompanied by triphylite altered to the heterosite-purpurite group which were coated by brown Fe-Mn oxides, possibly from groundwater.

K-feldspar-rich spodumene pegmatite is well developed in Pegmatite 1 and the western part of Pegmatite 2, and to a minor extent in Pegmatite 4. In Pegmatite 1 it represents the outermost primary zone, exhibiting an asymmetrical distribution obvious (Figure 19). This unit is distinguished by coarse-grained, light pink to white K-feldspar which constitutes about 55 percent of the mode. A pronounced

depletion of spodumene (7 percent) relative to adjacent spodumene-rich pegmatite is notable. Spodumene is principally confined to an interstitial relationship with the coalescing K-feldspar. A conspicuous texture not encountered in any other pegmatitic unit of the Mavis Lake spodumene pegmatites is a symplectite of green spodumene and quartz (Plate 12c).

Slender spodumene phenocrysts in Pegmatite 2 up to 1 m in length and 12-19 cm width contain fine-grained, granular quartz intergrowths composing about 30 percent of the host. The spodumene-quartz symplectites are usually enclosed by a thin shell of quartz and green muscovite, the total assemblage being completely enveloped by the coarse K-feldspar aggregate.

Albitization, as shown by fine-grained albite along K-feldspar boundaries and fractures, is only slightly developed in this zone. Patches up to 10 x 50 cm consisting mainly of fine-grained albite and green muscovite, with accessory red-brown garnet and white anhedral beryl (up to 1 x 2 cm) may also be of replacement origin. A few megascopic columbite-tantalite group grains were noted within these albitic patches and in other parts of the K-feldspar-rich pegmatite.

Irregular albite-rich replacement zones typify all spodumene pegmatites of the Mavis Lake area, although the proportion of these zones to primary assemblages can vary considerably from one pegmatite to another. In Pegmatite 1, approximately 10 percent of the exposed dyke consists of intensely albitized rocks within which no vestige of primary mineralogy has survived. This unit consists of a fine-grained white beryl-green muscovite-quartz-albite assemblage (Plate 12d) in

masses up to 4.6 x 9.2 m which may contain accessory garnet and ilmenite. Pegmatite 6 comprises about sixty percent of similar rock (Figure 18). Holmquistite masses and tourmaline porphyroblasts commonly occur in mafic metavolcanics in contact with albitized parts of spodumene pegmatites (Plate 12e) on the Fairservice property.

There is some evidence to indicate that albitization increases upwards towards the apical portions of spodumene pegmatites in the Mavis Lake area. Pegmatite 5 presents a 7 m vertical section of the upper part of a tapering dyke. The extreme upper portion of this dyke and apophyses are pervasively albitized and associated with dense tourmaline replacement of the host rocks, which is among the most extensive in the Mavis Lake area. Fracture-controlled holmquistite deposition is associated with the zone of heaviest tourmaline replacement. The extent of albite replacement gradually diminishes with increasing depth, giving way to primary assemblages containing spodumene, partially enveloped by fine-grained green mica. The full vertical extent of the albitization could not, however, be assessed because of the limited section available.

Purple mica mesoscopically identified as lepidolite was not observed associated with any albite-rich replacement units of the Mavis Lake spodumene pegmatite dykes. A scant amount of lepidolite was, however, noted to locally replace some spodumene towards the eastern end of Pegmatite 4.

South Zone: this zone consists of four known spodumene pegmatites (Figure 17, Table 10). The largest is a relatively flat-lying sheet which dips 28° south and comprises mainly aplite (50-60

percent). The fine-grained, inequigranular, massive to vaguely foliated aplite is composed of quartz and albite accompanied by an unidentified dark-brown mineral (possibly a weathered phosphate) which composes 15 percent of the rock and minor black tourmaline and green muscovite. No columbite-tantalite or beryl were observed within the aplite units. This unit is quite susceptible to weathering such that only weakly coherent specimens are available at surface. Primary spodumene pegmatite is commonly interlayered with the aplite. Contacts between these units are invariably sharp; no replacement of spodumene and blocky K-feldspar is obvious even where the aplite lies in contact with these minerals.

The spodumene pegmatite consists of the assemblage K-feldspar-albite-quartz-spodumene-green muscovite. Pale pink to medium blue, blocky K-feldspars up to 20 x 20 cm size (about 5 percent) are embedded in an albite-quartz-spodumene matrix having an average grain size of 2-4 cm. Pale green spodumene reaches maximum dimensions of 3 x 40 cm.

East Zone: rare-element pegmatites at the East Zone (Figure 17, back pocket) have been emplaced in a southward-facing mafic metavolcanic sequence comprising mainly fine-to medium-grained massive mafic units and minor pillowed flows. A 93 to 230 m wide serpentinized metaultramafic unit originally mapped by Satterly (1941), is also a host.

These pegmatites consist mainly of shallowly dipping, thin sheets and one possibly interconnecting dyke that have been subjected to pervasive albitization. Only outlines of crystals which have been

completely replaced by fine-grained, lime-green mica can verify the former presence of spodumene (similar to Plate 19a).

The pegmatitic sheets occur as a linear series of ovoid erosional remnants which can be intermittently traced for up to 320 m (Pegmatite 14, Figure 17 and Table 10). These sheets strike parallel to the host rocks but their shallow southeast dips are discordant to the generally steep northwest dips (68° - 78°) of the metavolcanic sequence.

Individual sheets vary significantly in dip values, as exemplified by Pegmatite 16. At the southwest end steep southeast dips between 65° - 76° are evident, compared to 38° at the northeast end.

Pegmatite 14 is typical of the intensely albitized sheets; only ten percent of the dyke contains recognizable spodumene relicts occurring haphazardly arranged laths typically 0.5×4 cm. Anhedral to subhedral pink to blue blocky K-feldspar relicts are even scarcer and range from 5-17 cm long. Most of Pegmatite 14 consists of 20-25 percent fine grained white to smoky quartz (typically 1.5×1.5 cm) embedded in a fine-grained, felty mass of white to light-pink platy albite, accompanied by scant black tourmaline.

Beryl occurs sporadically in Pegmatites 13, 14 and 15 as subhedral white to bluish-white crystals up to 17×20 mm embedded within quartz and platy albite. Euhedral to subhedral white beryl 3-5 cm long also is present within quartz-rich segregations in Pegmatite 15 and occasionally in contact with fine-grained white platy albite.

Tantalite (analysis 403, Appendix 39) up to 2 x 4 x 7 mm, is locally associated with the quartz-platy albite replacement assemblage, as in Pegmatite 16, which may also carry accessory fine-grained garnet and green muscovite.

A salient feature of the East Zone is the widespread occurrence of thin (usually <2.5 cm) tourmaline-rich sheets and interconnecting veins which are associated with the pegmatite sheets. These possibly represent boron-fluorine degassing structures. Intense tourmaline replacement of underlying massive mafic metavolcanics is especially conspicuous near the northwestern end of Pegmatite 16, in two modes of occurrence:

- 1) adjacent to 1-4 cm wide quartz-tourmaline-albite veins oriented subperpendicular to the pegmatitic sheet, and
- 2) in 8 - 13 cm thick tourmaline-rich layers oriented subparallel to the sheet.

Scheelite was identified by ultraviolet examination in vein systems underlying Pegmatite 16 and in similar veins between Pegmatites 13 and 14, near the main dykes of the East Zone. In the case of Pegmatite 16, it is clear that the tungsten mineralization is genetically associated at least with the tourmaline-rich veins occurring with spodumene-bearing rare-element pegmatites of the East Zone.

Northeast Zone: this zone consists of a 214 m long spodumene pegmatite which strikes 135° and dips 60° to the northeast. This attitude is approximately normal to regional foliation trends in the

host massive to pillowed mafic metavolcanics. Width at surface is variable between 4.6 m and 18 m.

This dyke was previously examined in 1956 by Milestone Mines Limited which undertook surface blast-pit work and one diamond drill hole (33.6 m). The drill hole intersected 1.8 m of spodumene pegmatite which assayed 1.20 percent Li_2O (Assessment Files Research Office, Ontario Geological Survey, Toronto).

The only primary assemblage observed consists of muscovite-tourmaline-K-feldspar-albite-spodumene-quartz pegmatite which is considerably less coarse than comparable primary assemblages from the North and South Zones. Spodumene is usually light green (rarely white), forming blades up to 3 x 17 cm and typically 1 x 2 cm. It ranges in abundance from 20 to 53 percent (Table 9). Green muscovite plates and books up to 0.5 cm thickness and euhedral black tourmaline up to 4 mm across (typically 1-2 mm) constitute the most common accessories. Some tourmaline prisms form a lineation plunging approximately down dip. Green-blue apatite is comparatively rare, occurring as 1-3 mm euhedral crystals enclosed within quartz or spodumene. No beryl or columbite-tantalite group minerals were identified.

Fine-grained aplite occurs as pockets within the primary spodumene unit. This white to pale-pink rock is distinguished by sharp contacts and higher apatite contents than the enclosing pegmatite in addition to enrichment in 5-10 percent of a weathered, dark brown, unidentified mineral (possibly triphylite). No evidence

of a replacement origin for the aplite could be discerned on careful examination of several polished and stained slabs.

The Northeast Zone spodumene pegmatite may represent an intermediate stage between the exclusively beryl-bearing external pegmatites of the Taylor Occurrence type and spodumene pegmatites characteristic of the North Zone. This is suggested by a lesser distance from the Ghost Lake Batholith contact (4.6 km) and by a degree of chemical fractionation which is intermediate between the Taylor and the North Zone rare-element pegmatites (see Chapter 7).

Tungsten Mineralization in the Mavis Lake Pegmatite Group

Recent investigations by Sanmine Exploration Incorporated in Brownridge and Zealand Townships have disclosed scheelite mineralization within a 1500 x 500 m area (R. Redmond, personal communication, 1983). This zone, which parallels the 095° trend of the Brownridge volcanics/Zealand sediments contact of Satterly (1941), is situated within the beryl zone of the Mavis Lake pegmatite group. This major contact is associated with a notable scarp, probably of erosional derivation.

Scheelite mineralization occurs mainly in amphibolitized, pillowed and occasionally amygdaloidal mafic metavolcanics and to a lesser extent in intercalated ultramafic and sulfidic layers. Massive mafic metavolcanic flows also occur on the property but generally do not appear to be favourable hosts for scheelite mineralization. Scheelite has thus far been delineated within two distinct settings:

- 1) in calc-silicate parallel to the deformation fabric of the host rock and

- 2) in tourmaline-rich sheets up to 0.3 thickness which variably transect foliation trends.

Calc-silicate domains, composed mainly of the assemblage quartz - calcite - diopside - epidote - garnet, and locally containing vesuvianite and axinite, carry low levels of scheelite.

The highest mesoscopically estimated abundance of scheelite occurs in generally flat-lying (11-19° dip) undulating, tourmaline-rich sheets up to 0.3 m in thickness. Scheelite and tourmaline are accompanied by ancillary quartz, plagioclase, pyrite and rare holmquistite. This assemblage is also developed in several spatially associated fracture sets with relatively steep dips (69°-85°) and along 0.7-5 cm wide pillow selvages; the latter exhibit intense replacement by tourmaline and biotite in the vicinity of the scheelite mineralization. Scheelite, in masses up to 5 by 7.5 cm, is predominantly anhedral, poikilitic and honey-brown on clean weathered surfaces. Analysis of a 3.5 m channel sample (along strike) selected from a flat-lying biotite-tourmaline sheet containing the highest observed scheelite content, is given below (in ppm unless otherwise stated):

<u>W</u>	<u>Be</u>	<u>Cs</u>	<u>F</u>	<u>Li</u>	<u>Sn</u>	<u>Au</u>
950	45	40	9,600	246	193	<2 ppb

Levels of tungsten are similar to those previously documented in the area by Noranda Mines Limited, who reported channel samples up to 0.1 percent WO₃ (Blackburn and Hailstone 1983, p.12-13), and by Selco Incorporated who obtained grades up to 0.09 percent WO₃ (A. Pryslak,

personal communication, 1983). It is apparent, however, that large bulk samples of several tonnes will be required to accurately assess the grade of zones which consist of "stacked" flat-lying scheelite-bearing tourmaline-rich sheets. Limited vertical observation indicates that these sheets are, at surface, spaced about 0.75 m apart and are separated by essentially barren pillowed mafic metavolcanics.

Similar occurrences of scheelite have also been verified by the author within the North and East Zones of the Fairservice Property. At the North Zone, small quantities of scheelite are disseminated within calc-silicate pods and layers in mafic metavolcanics situated within up to one metre of spodumene pegmatite. The occurrence of axinite in these calc-silicate domains suggests that boron was introduced from nearby albitized spodumene pegmatites. Boron metasomatism is also evident by increase in intensity of tourmaline replacement of mafic metavolcanics as the rare-element pegmatites are approached.

Chapter Four

GEOLOGY AND PETROLOGY OF THE GULLWING LAKE - TOT LAKE PEGMATITE GROUP

Introduction

This rare-element pegmatite group, situated in an 0.8 x 9 km area between the northeast end of Gullwing Lake and Tot Lake (Figure 20), is mainly confined to a medium grade, highly deformed, dominantly southwest-striking mafic metavolcanic unit which envelops the Lateral Lake Stock in the core of the Lateral Lake antiform (Breaks et al. 1978).

Most pegmatites of this group are oriented sub-perpendicular to the curvilinear contact between the mafic metavolcanic host unit and Lateral Lake Stock. Exceptions to this are the columbite-bearing Mica Point pegmatitic granite, concordantly emplaced in the mafic metavolcanic unit where it deflects north to northeast around the western terminus of the Lateral Lake Stock, and by the Drope Township columbite-molybdenite pegmatite exposed at the extreme southern known limits of the pegmatite group. The latter pegmatite was emplaced concordant to foliation of clastic metasedimentary host rocks.

The relationship of rare-element pegmatites of this group to molybdenite-rich granitic pegmatites of the Lateral Lake area, which commence approximately 4.8 km east of the Tot Lake pegmatite, is presently uncertain. Although these molybdenite-rich pegmatites were not included in the current study, a genetic relationship could very well be involved as molybdenite is an accessory in most of the rare-element pegmatites of the Gullwing-Tot Lakes group.

Six rare-element pegmatites and pegmatitic granites have been

Figure 20. General Geology of the Gullwing-Tot Lakes Pegmatite Group Showing Distribution of Pegmatites with Rare-Element Mineralization. Location of 1:4800 Geological Maps (Figures 21 and 22) Covering the Southwestern Part of the Group are also given.

discovered to date (Table 11) which exhibit several differences from those developed in the Mavis Lake pegmatite group:

- 1) most pegmatites are significantly discordant to the general 070° structural trend of the host rocks;
- 2) internally zoned pegmatites exhibit more complex chemical specialization i.e. Li-Cs-Rb-Be-Ta>Nb and Li-Nb>Ta-Be-Mo;
- 3) there is no evidence of associated scheelite mineralization;
- 4) tourmaline replacement of host rocks is absent and there is a less extensive development of contact metasomatic assemblages;
- 5) regional zonation of rare-element pegmatite types is not as clearly developed;
- 6) a parental granitic source area is not clearly apparent.

GEOLOGICAL SETTING

The general features of the supracrustal and felsic intrusive rocks within the Gullwing-Tot Lakes pegmatite group will now be described in this section.

SUPRACRUSTAL ROCKS

The mafic metavolcanic unit which hosts most of the rare-element pegmatites varies in width from 0.6 to 1 km (Figure 20). These rocks are moderately foliated and/or lineated, fine-to medium-grained, amphibolites of medium grade, as revealed by the general assemblage dark green to black hornblende-plagioclase (-biotite-garnet-actinolite).

Mafic metavolcanic rocks are stratigraphically overlain in the western to southwestern part of the area by clastic metasediments and intermediate pyroclastic metavolcanics (Page and Christie 1980). These

Table 11: Dimensions and Attitudes of Rare-Element Pegmatites from Gullwing-Tot Lakes Group.

<u>Pegmatite</u>	<u>Length (m)</u>	<u>Width (m)</u>	<u>Attitude</u>
Mica Point Nb-Ta Pegmatitic granite	250	29	0° - 150°/vertical
Gullwing Lake Li-Nb-Ta-Mo Pegmatitic granite	421	24-46	0180°/520° - 700E to 0200°/550° - 740W
Nb-Ta-Mo pegmatitic granite dyke near Ghost Lake Road	4.6	1.5	0900°/650S
Coates molybdenite- samaraskite pegmatitic granite	55	2-3	bifurcating dyke system 0180°/ vertical to 1200°/75S
Columbite-molybdenite pegmatites near Coates pegmatite	100 & 120	3-4	bifurcating dyke system 0050°/vertical and 0800°/750S
Tot Lake Li-Cs- Rb-Ta-Nb-Be Pegmatite	48	6 maximum	7450°/700° - 750NE

metasediments consist of garnet-biotite and garnet-muscovite-biotite metawacke with subordinate garnet-biotite-muscovite and sillimanite-biotite-muscovite metamudstone. Feldspathic biotite metawacke, quartz-rich biotite metawacke, and amphibole-garnet-biotite metamudstone are considerably less abundant. Intermediate metavolcanics are contained within two southwest-striking units (Figure 20) and possibly represent part of a local pyroclastic center mapped by Page and Christie (1980). These fragmental rocks are mainly heterolithic and composed of stretched, ovoid clasts up to 0.3 by 0.6 m which range in composition from mafic to intermediate. These rocks are predominantly tuff breccia to pyroclastic breccia with minor dacitic tuff to lapilli tuff; epidote alteration is evident in most exposures. Intercalation of the more extensive pyroclastic unit with southward facing clastic metasediments implies subaqueous deposition of these pyroclastics.

FELSIC INTRUSIVE ROCKS

Five pegmatitic granite masses occur in supracrustal rocks underlying the Gullwing-Tot Lakes group, as listed below in order of decreasing size (Figures 21 and 22, back pocket):

- 1) Coates Bay Stock,
- 2) a bifurcating dyke system near the Coates Bay Stock,
- 3) Gullwing Lake,
- 4) Mica Point, and,
- 5) RT pegmatitic granite dyke.

The Coates Bay Stock is a 1.6 x 3.2 km pluton mapped by Page and Christie (1980). This pluton mainly consists of pegmatitic leucogranite which is subtly zoned with respect to accessory

mineralogy. Muscovite is the predominant accessory within the 60 to 200 m wide marginal zone and is accompanied by fine-grained red garnet. This muscovite zone is succeeded by biotite towards the center of the stock.

A bifurcating, irregularly-shaped pegmatitic leucogranite dyke situated about 300 m northeast of the Coates Bay Stock (Figure 22, back pocket) is quite similar to the outer zone of this pluton in composition and by presence of plumose aggregates of fine-grained green muscovite and quartz (Plate 13a). Coarse muscovite books and fine-grained red garnet are the main accessory minerals.

This pegmatitic granite dyke appears to have been emplaced along northwest- and northeast-striking joint systems which account for its branching aspect.

The RT pegmatitic granite dyke, which is situated 170 m west of the Gullwing Lake dyke (Figure 21), is composed of muscovite-biotite granite petrographically identical to that in the latter. There is, however, no development of a quartz-rich core zone or any significant areas of albitization as is found in the Gullwing Lake dyke.

The Mica Point and Gullwing Lake dykes, described in detail below, are classified as pegmatite granite plutons in this study because of the following reasons:

- 1) most of these dykes contain an assortment of units (e.g. 5H, J and K in Figure 26) that more closely resemble those typical of pegmatitic granite complexes (such as GLB-5, -6 and -7 in the

Dryden Airport area of the Ghost Lake Batholith; compare Plates 6c and 13f) than units of rare-element pegmatites.

- 2) in the Dryden Field accessory biotite typically occurs in pegmatite granite units whereas it is not found in spodumene-bearing rare-element pegmatites.

The Gullwing Lake dyke is essentially a pegmatitic granite complex with local (5 x 13 m to 25 x 30 m), interspersed, much coarser-grained core units. The latter compare petrographically and mineralogically more closely with rare-element pegmatites such as Fairservice Pegmatite No. 1 (compare Plates 11b and 13d). As such, the Gullwing Lake dyke could represent a transitory type between pegmatitic granite and rare-element-enriched pegmatite classes of Cerny and Meintzer (1988, p.178-180).

A profusion of narrower steeply dipping pegmatitic granite dykes have invaded supracrustal rocks in the area around the above plutons. These dykes are usually less than 4.5 m in width and consist of two distinct types as reflected by cross-cutting relations and accessory mineralogy:

- 1) pale to medium pink granite- aplite with garnet-biotite (-muscovite) pegmatite-aplite (Unit 5b).
- 2) white to pale pink granite with garnet-green muscovite (-biotite) pegmatite (Unit 5c or 5d).

The first group of pegmatite dykes strikes predominately between N5E and N30E, rarely north to northwest. Accessory mineralogy is mainly restricted to fine-grained red garnet and biotite with rare

PLATE 13

- a. Pegmatitic leucogranite from a granite dyke near the Coates Bay Stock, which exhibits well developed plumose aggregates of green muscovite and quartz. Coin diameter = 2.4 cm.
- b. Biotite - muscovite-potassic pegmatite unit of the Gullwing Lake dyke which lies adjacent to the albitization I unit. Note the presence of yellow-green beryl below the 1.8 cm wide coin. Photograph taken 7 m east of the quartz-K-feldspar core in the South Zone.
- c. View of part of the blocky quartz-K-feldspar unit near the eastern contact of the North Zone, Gullwing Lake pegmatitic granite. Hammer = 58 cm length.
- d. Unaltered light green spodumene megacrysts relatively enveloped by a matrix of quartz and blocky K-feldspar. North Zone of Gullwing Lake dyke. Compass diameter = 7 cm..
- e. Herringbone aggregates of muscovite and quartz in contact with a platy albite-lithian muscovite-green muscovite replacement assemblage (albitization II unit). Note that the spodumene megacrysts have been completely replaced by fine-grained albite green mica. North Zone of Gullwing Lake dyke. Hammer length - 18 cm.
- f. Potassic pegmatite which composes most of the Gullwing Lake dyke. Photograph located 3 m east of the blocky K-feldspar-quartz core at the Central Zone. Note the presence of slender, bladed aggregates of muscovite-biotite. Arrow points to a small quartz-rich pocket in contact with blocky K-feldspar (K). Compass diameter = 7 cm.

PLATE 13

magnetite. These pegmatites are virtually devoid of muscovite, platy albite and rare-element minerals.

Many of the wider dykes of this group exhibit a well-developed aplite-pegmatite layering parallel to contacts as exemplified in those which form part of the host rocks along the eastern and southern contacts of the North Zone of the Gullwing Lake rare-element pegmatitic granite dyke (Figure 21).

The second group essentially exhibits two general strike directions i.e., N15E to N30E and N5W to N60W. These dykes contain ubiquitous accessory fine-grained red garnet, grey to green muscovite, and in rare instances exhibit scant biotite. In several cases, especially in the wider dykes of this group, a podiform quartz-rich core zone, contains coarser orange blocky K-feldspar, muscovite, local blue to white platy albite replacement masses and sparse, fine-grained platy columbite-tantalite. The last mineral occurs in sparse quantities in six of the narrow pegmatitic dykes of this group (Figure 21, back pocket). Pockets of aplite are also apparent. A younger age for group 1 dykes may be inferred from cross-cutting relations at the South and North Zones of the Gullwing Lake dyke (see below); the latter as based upon the similarity in orientation and some accessory minerals is interpreted to be a larger scale member of this group.

In the southwest part of the Gullwing-Tot Lakes pegmatite group, near the Coates Bay Pluton, there is a general tendency for green muscovite-bearing, narrow white pegmatites (group 2) to be hosted by clastic metasediments. In contrast, pink pegmatites of group 1 which

contain garnet and biotite are predominantly confined to intermediate pyroclastic rocks.

RARE-ELEMENT PEGMATITES AND PEGMATITIC GRANITES

The following section provides detailed descriptions of those pegmatites and pegmatitic granites in the Gullwing-Tot lakes group which contain rare-element mineralization.

MICA POINT PEGMATITIC GRANITE

This pegmatitic granite dyke, which marks the known northwestern limits of the Gullwing-Tot Lakes pegmatite group (Figure 20), was previously mapped by Harding (1950) and Page and Christie (1980). However, the presence of rare-element mineralization was not realized until A. Kozowy brought specimens of columbite to the writer in 1981. Most of this 18 to 27 m wide dyke is composed of light pink (weathered surface), medium to coarse-grained garnet-biotite granite, as exemplified at Mica Point. Farther north, in the vicinity of several shallow blast pits, a much coarser biotite-muscovite potassic pegmatite, identical to that surrounding the South and Central Zones of the Gullwing Lake and RT dykes, assumes prominence within the central portions of the dyke. This unit is flanked to the west by a pegmatitic leucogranite characterized by coarse-grained, white, graphic quartz-K-feldspar crystals up to 1 x 3 m. Near the western contact near shoreline of Gullwing Lake, this unit is succeeded by a fine-grained, grey, equigranular muscovite-biotite granite that may contain isolated euhedral blocky K-feldspar crystals.

Columbite (390, Appendix 40) occurs very locally within a 0.8 x 1 m area of the biotite-muscovite potassic pegmatitic which is partly

exposed by a shallow blast pit. It is distributed as euhedral inclusions (up to 8 x 8 mm) in muscovite books or along contacts between muscovite-biotite books and albite up to 1.3 cm diameter). Secondary fine-grained muscovite occurs as a replacement product in mica books contacting and coating the columbite and is, in places, associated with hematite staining.

Muscovite and biotite are intergrown in books up to 3 cm thickness or in blades up to 5 x 80 cm in the pegmatitic leucogranite unit. Biotite is characteristically intergrown as thin, 1-2 cm wide stripes within silver coloured muscovite. Blocky K-feldspar crystals up to 20 x 21 cm are developed within the central potassic pegmatite or in small quartz-rich pods erratically distributed throughout this unit. Fine-grained, lime-green mica coats the surfaces of some K-feldspar crystal faces.

No beryl was observed in the Mica Point dyke. The only additional accessory minerals noted were euhedral, red-brown garnet (usually less than 5 mm diameter), fine-grained black tourmaline and pyrite.

GULLWING LAKE SPODUMENE-COLUMBITE-BERYL-MOLYBDENITE PEGMATITIC GRANITE

Introduction

This zoned, two-mica pegmatitic granite (Plate 13b) is of critical importance in understanding the genesis of rare-element mineralization in the Dryden Field, as it contains the only known association of spodumene with a pegmatitic granite in the study area. Normally, spodumene occurs in rare-element pegmatites, such as near Tot Lake and Mavis Lake. Investigation of the Gullwing Lake dyke was

assisted by an appreciable local relief (55 m; Figure 23); virtually all other known rare-element pegmatites in the Dryden area provide less than 6 m of vertical section.

The Gullwing Lake dyke is a 412 m long curvilinear mass striking 160° and dipping from 30° - 70° E to 60° - 75° W. The opposing dip directions may be related to joint-controlled emplacement of the pegmatite dyke. These two dip directions are also reflected by the numerous narrow white dykes of garnet-green muscovite white pegmatitic granite in the immediate area, which may be related to the Gullwing Lake dyke.

Several granitic and pegmatitic units were documented by detailed mapping (Figures 24 in back pocket, 26 and 28). These were classified according to whether they are of primary or late replacement origin.

The Gullwing Lake pegmatitic granite exhibits the metal association Li-Nb>Ta-Mo-Be which is unique to the Dryden pegmatite field. An association of molybdenum with complex Nb-Ta minerals appears elsewhere in the Gullwing-Tot Lakes group, at the Coates molybdenite occurrence, in two intersecting pegmatite dykes a short distance north of the Coates molybdenite occurrence and at the Drope Township columbite-molybdenite occurrence near the Ghost Lake Road (Figures 21 and 22); however these occurrences do not have elevated lithium levels.

Columbite mineralization (7 analyses in Appendix 40) in the Gullwing Lake dyke is developed within three discrete localities,

herein termed the North, Central, and South Zones. These areas were mapped in detail at 1:60 scale (Figures 24, 26 and 28) in order to document the nature of zoning and geologic controls on distribution of the rare-element minerals. Locations of these zones are given in Figure 21 and all analyzed sample sites are shown in Figures 25, 27, and 29.

Several systematic changes occur in the core units in direction from the South Zone to the North Zone:

1. general increase in pegmatite grain size of the blocky K-feldspar - quartz - albite - muscovite zone, which implies increase in volatile concentration,
2. appearance of spodumene only in the two topographically lower zones and a greater abundance and larger grain size in the lowest zone,
3. increase in abundance and general size increase of columbite towards the lowest zone, and
4. appearance of the lithian muscovite-platy albite replacement assemblage only in the lowest zone.

North Zone

As indicated in Figure 21 (back pocket) this part of the Gullwing Lake dyke lies isolated 50 m from the main mass. Continuity at depth between the North and Central Zones is suggested by presence of a shallow southerly dip along the irregular upper contact of the North Zone (Figure 24) with its host rocks. These rocks consist of

Figure 25. Sample Location Map of North Zone of Gullwing Lake Pegmatitic Granite. Lithologic Codes and Geological Symbols as in Figure 24. Location of Analyzed Mineral or Rock Samples = ● (Bio = Biotite, KF = k-feldspar, LM = Lithian Muscovite, MUS = Muscovite, CT = Columbite, GL = Metasomatic Selvedge and GWMV 1 to 6 = Incremental Sampling Across a Metasomatic Selvedge from Pegmatite Contact).

lineated and foliated amphibolites, layered aplite-pegmatite granitic dykes (Unit 5b) and subordinate, biotite-bearing intermediate metavolcanic units. The earlier group 1 granitic dykes are albitized close to the Gullwing Lake (Figure 24). Several features contrast with the Central and South Zones (described later):

- 1) absence of pegmatitic granite units at the North Zone (5H, J and K in Figures 24 and 26); a mineralogical link between the North and Central Zones is suggested by unit 5B in Figures 24 and 26.
- 2) abundance of amphibolite and intermediate metavolcanic inclusions; some of these can be approximately fitted together, although 1.5 m of lateral offset of a 2a unit is suggested in two adjacent inclusions near the mid-point of the North Zone.
- 3) structural continuity across the pegmatite dyke i.e. regional lineation within the metavolcanic xenoliths and adjacent host rock is essentially similar.
- 4) intense metasomatism of metavolcanic xenoliths and host rocks.

The North Zone is crudely rectangular on surface, averaging about 25 m in width and 30 m long. Strike of the pegmatite varies between 015° and 025° with westerly dips of 55° to 66° .

Most of the North Zone is composed of a coarse-grained pegmatite composed of quartz and blocky K-feldspar (Plate 13c). This primary assemblage is the coarsest encountered in the Dryden Pegmatite Field; single irregular quartz masses reach 6 x 9 m.

An irregular marginal rim of equigranular albite and bladed biotite - muscovite replacement complex has completely isolated this primary zone from contact with metavolcanic xenoliths or host rocks (Figure 24).

The spodumene-blocky K-feldspar-quartz-albite-muscovite zone composes only about 10 percent of the North Zone and is confined to three small pods situated towards the lower levels of the pegmatite (Figure 24). It is possible that the spodumene-bearing unit constitutes the upper part of a more extensive, continuous zone expanding at depth in similar fashion to that documented by Vlasov (1968, p.51). Spodumene phenocrysts, intergrown with orange blocky K-feldspar, quartz, grey muscovite and albite (Plate 13d), are randomly oriented, light-green, euhedral and measure up to 4 x 40 x 100 cm. by 100 cm. The true colour is commonly masked on fresh crystals by orange to brown iron-staining. Pseudomorphic replacement by an aphanitic aggregate of albite and green mica can also make recognition of spodumene difficult (Plate 13e).

Coarse blocky K-feldspars envelop the large spodumene phenocrysts and thus have crystallized later. Muscovite in the spodumene zone tends to occur in herringbone intergrowths with quartz (Plate 13e), which are principally situated along the flanks of the spodumene zone.

Central Zone

The Central Zone is at least 35 m in strike length and is tapered relative to the North Zone, varying between 7 and 20 m in width. Asymmetrical zonation is vividly exhibited (Figure 26).

White potassic pegmatite (unit 5J, Figure 26) which encompasses most of the Gullwing Lake dyke between the east side of the Central Zone and the South zone, completely enclosing the latter is characterized by bladed aggregates of muscovite and biotite that in places exceed 1 m in length and are commonly less than 5 cm in width. This rock type has a hypidiomorphic-granular texture. Small quartz-rich pockets up to 10 x 30 cm abut against small blocky K-feldspar crystals (Plate 13f), a textural feature diagnostic of potassic pegmatite (Cerny and Meintzer 1988, p.178). Accessory quantities of green beryl, black tourmaline and columbite are locally present.

Grey, muscovite-biotite fine-grained leucogranite is the second most abundant unit at the Central Zone (Figure 26). It occurs along the west side of the central spodumene-bearing core zone. A scalloped contact with the albitization is apparent, likely developed via outward replacement of marginal unit by the secondary assemblage (Plate 14a).

Two mineral assemblages are present in the quartz-rich core zone.

- 1) blocky K-feldspar-quartz-albite-muscovite
- 2) quartz-spodumene-blocky K-feldspar-albite-muscovite

The latter assemblage is restricted to five relatively small pockets. Spodumene is altered to fine-grained to aphanitic albite-white mica intergrowth (Plate 13e). This replacement occurred during the albitization II replacement stage and is intimately associated with white to blue radiating masses of platy albite, partially albitized

Figure 26. Detailed Geology of Central Zone of Gullwing Lake Pegmatitic Granite.

Figure 27. Sample Location Map of Central Zone of Gullwing Lake Pegmatitic Granite. Lithological Codes and Geological Symbols as in Figure 26. Location of Analyzed Mineral Samples = ● (KF = K-feldspar, MUS = Muscovite and CT = Columbite. □□□ = Location of Rock Channel Sample.

3

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PLATE 14

- a. Scalloped contact between the albitization I replacement complex and a marginal unit of the grey, fine-grained leucogranite in the Central Zone of Gullwing Lake dyke. Location is one metre west of the blocky K-feldspar-quartz core. Compass diameter = 7 cm.
- b. Large, euhedral columbite crystal occurring at the contact of a quartz pod, blocky K-feldspar (K) and a platy albite-muscovite replacement intergrowth (P). Coin diameter = 1.8 cm. Central Zone of the Gullwing Lake dyke
- c. Partial, peripheral replacement of rusty brown weathering biotite by a grey aggregate of randomly-oriented, fine-grained muscovite. Also note the subtle association of muscovite rims with albite (A) as at the 2 cm diameter, formed via replacement of large, blocky K-feldspar (K), a ragged remnant which is indicated by the arrow. South Zone of the Gullwing Lake dyke. Coin diameter = 1.8 cm.
- d. Localization of columbite minerals (arrows), below 1.8 cm diameter coin, to the contact between marginally albitized (A) portion of a large blocky K-feldspar (K) and a quartz-rich (Q) pocket. South Zone of the Gullwing Lake dyke. Coin diameter = 1.8 cm.
- e. Columbite (arrows) disseminated within a large, light-orange blocky K-feldspar and associated thin veinlets of albite. Coin diameter = 1.8 cm. South Zone of Gullwing Lake dyke.

PLATE 14

orange blocky K-feldspar, orange garnet and rare faint green beryl. Columbite (analysis 397, Appendix 40) up to 3 x 3 cm (Plate 14b) tends to be distributed along the contact between the core unit and the albitization I assemblage. It also occurs as fine-grained plates disseminated within the platy albite-green mica replacement of the spodumene-bearing primary assemblage.

South Zone

The South Zone is similar to the Central Zone in zonation pattern except for a more symmetrical distribution of pegmatite granite around the quartz-rich core (Figure 28). Absence of spodumene is an important distinction of this locality from the North and Central Zones. Dimensions of the quartz-blocky K-feldspar-muscovite core zone and enclosing albitization I shell ranges between 5 and 12 m in width. Length is at least 25 m, the southern part of which is under shallow gravel overburden. The albite-quartz-muscovite replacement zone shows a subtle transition from the muscovite-biotite potassic pegmatite. Progressive conversion of bladed aggregates of biotite and muscovite into fine-grained secondary muscovite is apparent (Plate 14c) as the albitization I zone is approached. Complete destruction of these biotite books has occurred in the albitization shell. Fine-grained columbite-tantalite is disseminated within this replacement zone. Coarser crystals are especially notable along the western limits of the quartz-blocky K-feldspar-muscovite core, reaching up to 1 x 2 cm (Plate 14d). At the southern limits of the core zone, a large blocky K-feldspar contains abundant columbite (analysis 398, Appendix 40) associated with incipient, film-like albite replacement (Plate 14e).

Figure 28. Detailed Geology of South Zone of Gullwing Lake Pegmatitic Granite.

Figure 29. Sample Location: Map for South Zone of Guliwing Lake Pegmatitic Granite. Lithological Codes and Geological Symbols as in Figure 28. Location of Analyzed Mineral Samples = ● (Bio = Biotite, CT = Columbite, KF = K-feldspar and Mu = Muscovite). □ — □ = Location of Rock Channel Sample.

Late Stage Albite-Rich Replacment Units

Replacement of the primary assemblages took place in two events termed albitization I and II. The more extensive albitization I replacement stage is characterized by relatively coarse, equigranular albite replacing medium-to coarse-grained blocky K-feldspar (Plate 15a) and may contain subordinate remnants of biotite, muscovite, quartz and K-feldspar. A variant of this albitization is a unit characterized by the intense development of randomly oriented bladed aggregates of muscovite and biotite (Plate 15b). Biotite is unusual in albitization zones of rare-element pegmatites of the study area. This unit tends to be best developed between the upper pegmatite contact and the mid-point of the eastern contact. Molybdenite and platy columbite (analysis 393, Appendix 40) are commonly intergrown with the micas in this unit.

Distribution of the albitization I zones is clearly controlled by contacts with host rocks and metavolcanic xenoliths (Figure 24, back pocket). The host rocks themselves exhibit a zonality of alteration (i.e. biotite-rich selvages) which is never developed away from the albitization I units.

At the North Zone, a biotite-rich alteration selvage is continuously present along the pegmatite contact (Figure 24). The selvage is modally variable and its composition depends upon the host rock being replaced. Intermediate metavolcanic units were the most reactive judging by the relatively wide alteration zones (up to 1 m width). These are characterized by enrichment in biotite (50 to 95 percent) accompanied by plagioclase and quartz. These units are

PLATE 15

- a. Part of an albitization I unit (AB I) exhibiting marginal replacement of a large blocky K-feldspar (K) by albite and bladed aggregates of biotite - muscovite. Arrow points to K-feldspar remnants. Coin diameter = 1.8 cm. North Zone of Gullwing Lake dyke.
- b. Bladed muscovite-biotite aggregate albite replacement unit with K-feldspar remnants (K). Compass diameter = 7 cm. North Zone of Gullwing Lake dyke.
- c. Contact metasomatic rock adjacent to an albitization I unit showing intense development of plagioclase poikiloblasts and biotite enrichment in the host rocks. Coin diameter = 1.8 cm. North Zone of Gullwing Lake dyke.
- d. Contact metasomatic zone developed between amphibolitic mafic metavolcanics and the albitization I replacement unit, showing areas of the selvages enriched in albite (A) and biotite (B). Hammer length = 30 cm. North Zone of Gullwing Lake dyke.
- e. Narrow platy albite-rich vein of the albitization II replacement unit (AB II) stage cross-cutting and locally overgrowing the equigranular albite replacement unit (AB I) at the North Zone of Gullwing Lake dyke. Coin diameter = 1.8 cm.
- f. Fine-grained, purple lithian muscovite-platy albite aggregate which has replaced a blocky K-feldspar in the spodumene-K-feldspar-quartz core of the North Zone of the Gullwing Lake dyke. Coin diameter = 1.8 cm.

PLATE 15

locally porphyroblastic, with poikiloblasts of plagioclase (Plate 15c) and biotite up to several cm in diameter. Mafic metavolcanic host rocks exhibit narrower alteration zones (usually less than 15 cm width) with less biotite and greater quartz and plagioclase contents (Plate 15d).

A plagioclase-rich alteration assemblage commonly lies between the biotite selvages and the albitized pegmatite (Plate 15d). In certain cases relict pillow selvages and regional mineral lineations in the amphibolite host as well as portions of older, granitic dykes (unit 5a in Figure 24) truncated by the North Zone pegmatite can be traced into the plagioclase-rich selvages. The selvedge contains 50-80 percent plagioclase, suggesting an overall net increase in this mineral relative to its typical abundance of 40-50 percent in nearby amphibolites not affected by the metasomatism. This suggests an ingress of Fe and Mg into the albitization I zone, which when combined with K, Al and Si released through destruction of blocky K-feldspar could have formed the prominent bladed aggregates of biotite and muscovite.

Albitization II is a later stage of replacement specifically confined to the three spodumene zones and is characterized by radiating aggregates of platy, white to blue platy albite. Rarely do both albitization stages coexist (Plate 15e) the platy albite is seen to subtly penetrate the coarser equigranular albite type. Platy albite replaces all primary minerals including blocky K-feldspar, spodumene, and quartz. Faint greenish to white, anhedral to subhedral beryl up to 4 x 6 cm is associated with platy albite in the most

westerly spodumene-bearing unit of the North Zone. Light to dark purple micas were tentatively identified as lithian muscovite based upon Li_2O contents of (3.00 to 3.40 weight percent) which are not quite in the lepidolite range (3.5 to 4.5 weight percent; Hawthorne and Cerny, 1982, p.85). Lithian muscovite occurs only at the North Zone where it appears:

- 1) as fine-grained replacement of spodumene or blocky K-feldspar (Plate 15f),
- 2) relatively coarse intergrowths with platy albite,
- 3) in quartz-rich pockets associated with platy albite, and
- 4) incipient marginal replacement of muscovite books.

Columbite distribution is closely related to the albite-rich replacement units. All mesoscopic columbite crystals have been located on Figure 24. This mineral occurs as single euhedral crystals up to 2 x 5 cm or more rarely in clusters. An area of high columbite abundance (0.5-1 percent) developed within the southwest portion of the North Zone immediately above the largest spodumene zone. A second area of concentration is situated towards the northeast extremity of the North Zone. Columbite here (analysis 394, Appendix 40) is disseminated within a unit rich in fine-grained green mica and albite poikiloblasts several cm in diameter and composes one-half to one percent of the rock. In the lithian muscovite-platy albite replacement assemblage columbite minerals are invariably fine-grained, relatively inconspicuous and are enriched in Nb_2O_5 relative to columbite in the albitization I units (see Figure 96 in Chapter 8).

Rust spots commonly surround the fine-grained columbite in albitization II units which facilitates their recognition.

Sphalerite occurs as dark brown, euhedral crystals up to 0.8 x 1.3 cm in very local parts of the spodumene core in the North Zone. A secondary origin is indicated by its intricate intergrowth with light purple lithian muscovite, muscovite, quartz and platy albite (map unit 5E in Figure 24).

DROPE TOWNSHIP COLUMBITE-MOLYBDENITE PEGMATITE

The southwestern limit of the Gullwing-Tot Lake pegmatite group is defined by a small potassic pegmatite dyke situated near the southeastern corner of Drope Township (Figure 21). This fine-grained pegmatite is located very close to the Ghost Lake Road approximately 260 m due west of the Webb Township boundary.

The pegmatite is white weathering, enriched in green muscovite (10-20 percent) and was emplaced concordant to foliation of enclosing andalusite-muscovite metapelites. A 10 cm wide tourmaline-rich halo is developed within the host metasediments. Asymmetric boudinage has affected this dyke; quartz fills the spaces between boudins.

Columbite mineralization (analysis 401, Appendix 40) is restricted to a few subhedral grains up to 7 x 8 mm associated with green muscovite, quartz, albite, and accessory red garnet. Molybdenite up to 4 mm across occurs in the same association but was not observed to contact the columbite.

COATES MOLYBDENITE-SAMARSKITE-BERYL PEGMATITE

This pegmatite, situated 2200 m ENE from the Gullwing Lake dyke (Figure 21), is named after Cosmo Coates, an early prospector in the region who most likely discovered the mineralization (Harding 1950). Four blast pits of two differing investigation periods are obvious. An early, extensively overgrown pit was possibly excavated by Coates and three relatively recent blast pits were noted by the writer in 1975, with no record of the party who undertook this investigation.

The Coates pegmatite is a T-shaped dyke system, composed of a main arm at least 55 m in length and 3 m wide, striking 018° with vertical dip and a subordinate 15 m long, 2 m wide dyke trending 120° with dip of 75°S . These arms intersect near the eastern limits of the exposed pegmatite. A conspicuous light-orange to pink coloration contrasts this dyke from all other rare-element pegmatites in the group. This orange colour is mainly due to iron oxide staining which is especially prevalent in vicinity of the westernmost blast pit.

Along the contacts of the north striking dyke, mafic metavolcanic host rocks have been altered in a zone 5 cm wide containing abundant biotite porphyroblasts. These metasomatic selvages may have been generated via release of potassium by late stage albitic replacement of K-feldspar. High trace levels of some Li, Rb, Cs, Be, Sn and F are also notable (see Table 40 in Chapter 7).

Sulphide and rare-element mineralization is apparently confined to the north-striking arm. This medium- to coarse-grained potassic pegmatite contains a high abundance of both homogeneous, perthitic blocky K-feldspar (10 x 20 cm maximum) and graphic quartz-K-feldspar intergrowths (17 x 17 cm maximum) distributed in no discernible zonal

pattern. Tabular, smoky-grey massive quartz segregations up to 20 cm thickness are arranged en echelon in a "step-ladder" distribution subperpendicular to the dyke contact. Incipient to moderate replacement of K-feldspar by fine-grained, equigranular, grey-to pink-grey albite (Plate 16a) occurs adjacent to the quartz segregations. These replacement units occur as masses up to 0.2 m wide or as interconnecting 0.5 - 1 cm wide veins controlled in most cases by cleavage in the K-feldspar host. The albitic replacement is important because since samarskite, rare beryl and some molybdenite (Plate 16a) are contacted by or contained within these areas. Red-brown garnet less than 1 mm diameter also occurs sporadically in some albite veins.

Molybdenite occurs mostly in the quartz-rich pockets and ladder veins as single, commonly pseudohexagonal crystals between 2 mm and 2 cm diameter or in aggregates up to 3.5 cm x 9 cm. More rarely is molybdenite found in garnet-bearing aplite pockets. Very local ferrimolybdenite staining may be associated with the molybdenite mineralization.

Samarskite is apparently very localized in the Coates pegmatite, developed only in the late stage albitic replacement units. This mineral occurs generally as glossy-black, brittle, euhedral crystals up to 7 mm x 2.5 cm, surrounded by rusty haloes. A partial analysis is presented in Appendix 41.

Beryl is extremely rare; only one euhedral, faintly green crystal measuring 1.4 x 3 cm was recognized and possibly is of secondary origin as it is associated with replacement albite.

COLUMBITE-MOLYBDENITE PEGMATITE NEAR THE COATES OCCURRENCE

Two garnet-muscovite potassic pegmatite dykes are situated at locality 30-17-1, 200 m southwest of the Coates molybdenite pegmatite (Figure 21). These 3-4 m wide dykes form an intersecting system with strikes of 005° and 080° and dips between 75° and vertical.

The east-striking dyke contains anhedral to subhedral columbite grains which vary from 3 x 4 mm to 2 x 3 cm. Columbite (analysis 400, Appendix 40) is associated with thin seams of fine-grained replacement albite which penetrate edges and interiors of pink-to blue blocky K-feldspars. The K-feldspar megacrysts can be very coarse and elongate with maximum dimensions of 25 x 112 cm. The coarsest columbite crystal occurs within a K-feldspar rich aggregate and lies in contact with veinlets of albite replacement.

Silver-coloured, medium-grained muscovite is particularly abundant in both dykes. Accessory minerals include fine-grained garnet, apatite and molybdenite.

PLATE 16

- a. Part of a large orange blocky K-feldspar from the Coates Pegmatite, showing lenticular quartz veins and attendant albitic replacement areas. Note presence of blue-grey molybdenite (arrow) which can clearly be linked to an albitic replacement vein.
- b. View of the Tot Lake Pegmatite at the southeastern end, revealing the longitudinal zoning, marked by couplets comprising of spodumene-rich (S) and aplitic (A) layers intruded by a central quartz vein.
- c. Detail of Plate 16b, showing alternating spodumene-rich (S) and aplite (A) layers.
- d. Polished slab of a spodumene-rich layer from unit 2, Tot Lake Pegmatite revealing the maximum zonal spodumene content (78 percent) found in the Dryden Pegmatite Field.
- e. Transverse layered zone (unit 1) near Trench 3 (Figure 30), which contains spodumene megacrysts within a finely layered matrix composed of light green, fine-grained aplite alternating with layers of coarser-grained quartz-albite-K-feldspar-spodumene. Coin diameter = 2 cm.

PLATE 16

TOT LAKE SPODUMENE-POLLUCITE-COLUMBITE-TANTALITE-BERYL PEGMATITE

Introduction

The Tot Lake rare-element pegmatite, which is located situated in Webb Township 1.3 km at 210° from the most southerly part of Tot Lake, is the most easterly pegmatite of the Gullwing-Tot Lakes group (Figure 20).

The area surrounding this dyke is covered by extensive Pleistocene glacio-lacustrine clay and is considerably less elevated than the western part of the pegmatite group.

The Tot Lake pegmatite striking 145° and dipping 70-75° northeast is sharply discordant to foliation trends of its host rock. It is exposed for about 48 m, achieving a maximum width of 6 m at the southwestern end and constricting to about 1 m at its mid-point (Figure 30, back pocket).

Host rocks consist mainly of low-grade, well-foliated, fine-grained mafic metavolcanics. The assemblage biotite-chlorite-amphibole-plagioclase-quartz is typical. A unit of massive mafic metavolcanics, in width, containing medium-grained, hornblende porphyroblasts lies in sharp contact with the fine-grained variety near the southeastern end of the dyke. The extreme southeastern and mid-points of the pegmatite lies in contact with metaultramafic rocks.

This diverse pegmatite dyke is the most complex in the Dryden Field due to internal zonation, local alignment of sub-horizontal spodumene phenocrysts normal to dyke contacts and late stage, fracture-controlled moderate to intense albitization. A small

pollucite zone is a unique feature and attests to extreme chemical fractionation (Cerny et al. 1985).

Primary and Secondary Zones and Associated Structures

Ten different, commonly gradational units were delineated by 1:60 scale detailed mapping (Figure 30, back pocket). All chemical sample locations are given in Figure 31. Approximately seventy percent of the primary pegmatitic assemblages have been variably altered by late stage albitization and lepidolitization.

Primary Zones: five zones containing primary assemblages are situated at the southeastern end of the pegmatite and towards the dyke center near the northwestern end (Figure 30). These zones are listed below; minerals in each unit are given in general order of decreasing abundance:

<u>Unit #</u>	<u>Primary Assemblage</u>
1)	Spodumene-K-feldspar-albite-quartz
2)	Spodumene-green muscovite-albite-quartz
3)	Spodumene-albite-green muscovite-quartz
4)	K-feldspar-quartz-beryl-spodumene
5)	Pollucite-Spodumene-K-feldspar-quartz-albite-green muscovite-fluorapatite-tantalite

Two contrasting types of primary layering are recognizable:

- 1) longitudinal layering oriented parallel to external pegmatite contact
- 2) transverse layering with high inclination to external pegmatite contact

Figure 31. Sample Location Map for Tot Lake Pegmatite. Geological Units same as in Figure 30.

Longitudinal layering is exposed near the southeastern end of the dyke (Figure 30, Plate 16b). It is defined by alternation of 50-60 cm wide layers of unit 2 and 3. Unit 2 is conspicuous by its high content of pink spodumene (up to 78 percent in Table 12 and Plates 16c and 16d), and low quartz. The c-axes of spodumenes trend approximately normal to the contacts between unit 2 and 3 with shallow plunges.

Unit 3 is a finer-grained rock enriched in albite and green muscovite and notably impoverished in spodumene (less than 5 percent) relative to unit 2.

These mineralogically contrasting layers appear to have developed in a coupled fashion as suggested by sharply bounding contacts of Unit 3 against unit 1. A gradational, slightly scalloped contact, on the other hand, occurs between units 2 and 3 (Plate 16c).

The transverse-layered zone is developed at the opposite end of the Tot Lake Pegmatite in unit 1 and its albitized counterpart (unit 7) around Trench 3 (Figure 30). It is defined by much thinner layers (1-3 cm wide) alternating from light orange spodumene pegmatite to finer-grained, apple green aplite (Plates 16e and 17a).

The transverse-layered zone is a part of unit 1 because it grades into spodumene-rich pegmatite (to the southeast of Trench 3) characterized by randomly oriented, "giant" megacrysts of spodumene which achieve maximum dimensions of 1 x 14 x 38 cm (Plate 18a). Towards the northeast, the spodumene megacrysts gradually diminish in abundance to the point where they become isolated from the main area

Table 12: Modal Analyses of the Pollucite Zone, Columbite Zone, and Spodumene-Rich Units from Longitudinal and Transverse-Layered Zones at the Tot Lake Pegmatite.

	1	2	3	4	5	6	7	8	9
Quartz	12	4	41	43	4	3	1	2	3
Albite	17	16	47	12	25	30	42	32	43
Pollucite	32	-	-	-	-	-	-	-	-
Altered Pollucite	5	-	-	-	-	-	-	-	-
Spodumene	27	78	7	35	33	44	13	24	28
Green Muscovite	7	3	5	8	-	1	-	1	-
Fine-grained White Mica	1	-	X	-	-	-	-	-	-
K-feldspar	X	-	-	-	38	23	40	29	26
Columbite-Tantalite	X	-	X	-	-	-	-	-	-
Beryl	-	-	-	2	-	-	4	1	-

1. Pollucite zone in Trench 2
2. Spodumene-rich layer near southeastern end
3. Columbite zone near Trench 3
4. Altered spodumene zone near northwest end
5. -
9. Modes of five successive coarser grained spodumene-rich layers from transverse-layered spodumene pegmatite-aplite zone near Trench 3.

X = present in accessory quantities

PLATE 17

- a. Detail of the transverse-layered unit from the Tot Lake Pegmatite, located 2 m northwest of Trench 3 in Figure 30, revealing the rhythmic alteration of layers of fine-grained albite-rich aplite and coarser-grained quartz (Q), albite (A), K-feldspar (K) and spodumene (S)

PLATE 17

PLATE 18

- a. Largest spodumenes in the Dryden Pegmatite Field as found in unit 1, near Trench 3 of Tot Lake Pegmatite. Hammer = 58 cm..
- b. Example of an albitization front (A) in part of the transverse layered spodumene zone near Trench 3. The front, as marked by the 2 cm wide coin, is oriented approximately normal to the primary layering.
- c. View of the pollucite zone of the Tot Lake Pegmatite near the northeastern pegmatite contact in Trench 2, showing pink, euhedral spodumene blades within a matrix dominated by milky-white pollucite. Coin diameter = 2 cm.
- d. Slab from the pollucite zone illustrating textural relations amongst the primary minerals spodumene (S), pollucite (P) and quartz (Q) and a fine-grained secondary replacement assemblage of green muscovite and albite. Slab is oriented normal to the view of the zone in Plate 18c.
- e. Spodumene-K-feldspar-quartz pegmatite (unit 8) showing randomly oriented, pink spodumene blades (arrow) enveloped by irregular, K-feldspar megacrysts (K) and masses of quartz (Q). Small pockets of fine-grained aplite (A) occur in the upper right corner. Coin diameter = 2.4 cm.

PLATE 18

of concentration and embedded in a matrix of layered spodumene pegmatite-aplite (Plate 18b).

Modal analyses of spodumene pegmatite layers from five successive pegmatite-aplite couplets are given in Table 12. Spodumene, K-feldspar and albite are the dominant phases which exhibit considerable interlayer variation. Spodumene ranges from 13-44 percent. The feldspars show more restricted variation; albite usually dominates over K-feldspar (Table 12). Quartz is notably low (1-4 percent). Accessory minerals include green muscovite and beryl.

The aplite layers are compositionally uniform and consist of 90-95 percent albite with minor quartz and green muscovite. The transverse layering gradually diminishes to the northeast which is in part possibly a function of increasing albitization described in the following section.

Secondary replacement units: five secondary units (Figure 30, backpocket) were developed by late stage metasomatic activity which variably replaced primary mineral assemblages with albite, green mica and minor purplish mica (megascopically identified as lepidolite) as detailed below:

Unit #

6. Incipiently to moderately albitized spodumene-albite-quartz-K-feldspar pegmatite. Replacement by albite-green muscovite-lepidolite-white beryl.

7. pervasively albitized spodumene-albite-quartz-K-feldspar pegmatite. Replacement by fine-grained albite - green muscovite - white beryl.
8. Quartz-albite K-feldspar-spodumene (altered to green muscovite and albite)-columbite-garnet-pink to white beryl.
9. Sodic aplite: albite-quartz-blue tourmaline- green muscovite-apatite-garnet.
10. Holmquistite veins.

These secondary units occur in two distinct manners:

- 1) Irregular pods of fine-grained sodic aplite, and
- 2) fracture-controlled replacement related to a central branching quartz-vein system.

The first type of replacement, distinctive in virtue of a bluish cast on wet surfaces, is featured by six small pods of sodic aplite up to 0.5 x 1.5 m entirely enclosed within unit 1 or lying across contacts of units 1 with 5 and 2 with 3. Two additional pods were mapped in unit 7, the host rocks of which were later pervasively altered by fracture-controlled albitization. The podiform type of albitization is of minor importance but does suggest that two stages of replacement developed in the Tot Lake pegmatite.

The second style of alteration is much more widespread and is distinctively associated with a late fracture system (now mainly filled with quartz) which tends to lie subparallel to dyke contacts (Figure 30). The degree of alteration is variable, ranging from incipient to moderate albitization as seen towards the southeastern end, with primary textural preservation, to complete transformation

into a fine-grained albite - quartz - green muscovite rock as at the northwestern end (Plate 18b).

The largest quartz vein, confined near the dyke centre, reaches a maximum width of 0.5 m and is traceable continuously for about 30 m. Although albitization is undoubtedly related to the fractures, as observed at the northwestern end of the pegmatite, unaffected primary mineral assemblages abut against the central quartz vein at the southeastern end of the exposure.

In one case, fracturing was guided partially by the southwestern dyke contact, in the vicinity of Trenches 1 and 2. There, an irregularly albitized zone related to a quartz-tourmaline vein shows strongly embayed contacts with unit 1 (Figure 30), and presence of relict spodumene. Deposition of a fine-grained black tourmaline "film" is noteworthy at the contact between hornblende - porphyroblastic mafic metavolcanic metaultramafic host rocks, near the southeastern end of the dyke.

The incompleteness of the metasomatic process is apparent near the northwestern end of the exposure where spodumene phenocrysts and fine transverse primary layering are overprinted approximately at right angles by an albitization front (Plate 18b). Altered equivalents of the spodumene and relict layering are still apparent on wet outcrop surfaces of the altered zone. The albitization zones are arranged parallel and symmetrical to quartz-filled fractures.

Diagnostic minerals and their textural relations: Important diagnostic minerals and textural relations in the various primary and secondary units will now be described in detail. As these minerals, with the exception of pollucite, occur in numerous zones they are individually described below for the sake of brevity.

Spodumene is a major mineral only in primary units 1, 2 and 4, being relegated to accessory status only in units 3 and 4. A great range in grain size is evident, from grains 1 x 3 cm in unit 3 to "giant" crystals 38 x 14 x 1 cm developed in unit 1. Spodumene mainly occurs in the form of blades elongated to the b and c directions. Color is almost invariably pink in rare cases, vitreous lilac crystals (kunzite variety) were noted, especially in the quartz-rich spodumene-green mica assemblage between the pollucite-rich zone and meta-ultramafic host rocks near the end of Trench 2. According to Heinrich (1975, 1978) lilac to pink colorations are indicative of Type 2 zonal spodumene in which manganese acts as the chief chromophore in conjunction with iron content lower than that of Heinrich's *ibid*) green Type 1 phenocrystic spodumene. Type 2 zonal spodumenes characteristically have low Fe₂O₃ contents, commonly in the range 0.01-0.03 percent (Heinrich 1978).

Euhedral spodumene blades commonly form an intricate boxwork texture (Plates 18c and 18d) with interstices filled by later-formed albite, K-feldspar, quartz and pollucite. These blades partially penetrate or are completely enclosed by irregular aggregates of K-feldspar (Plate 18c). Textural relations suggest the following crystallization progression of major mineral phases:

Unit 1: Spodumene --> Albite --> K-feldspar --> Quartz

Unit 5: Fluorapatite --> Spodumene --> Albite --> K-feldspar -->
Pollucite --> Quartz

Spodumene typically alters to an albite-green muscovite intergrowth. The extent of alteration ranges from incipient development of a lens-shaped corona of albite - green muscovite with a surviving core of spodumene to complete replacement by this assemblage (Plate 19a). A similar range in variability also applies to lepidolite replacement of spodumene, however this type of alteration is restricted to spodumenes in the pollucite zone (Plate 19b).

K-feldspar is evident as white, discrete blocky to irregular perthitic megacrysts up to 30 cm diameter (Plate 18c) composing 10 to 50 percent of units 1 and 5. It dominates only in unit 4, the "orange K-feldspar zone", which is composed mostly of an aggregate of blocky K-feldspar (80-90 percent) with minor interstitial quartz and rare white beryl. Ucakwun (1981) determined that most K-feldspars from the Toi Lake pegmatite are maximum microcline.

Pollucite occurs as coarse anhedral to rarely subhedral white crystal aggregates filling interstices between bladed spodumene phenocrysts in Unit 5, and composes about 32 percent of the mode (Table 14). Rare textural relations show subhedral pollucite with crystal faces adjacent to smoky quartz and enveloping relatively early euhedral spodumene (Plate 18d). This evidence indicates that pollucite was the last major rare-element mineral to crystallize in the pollucite zone. Pollucite characteristically develops during the last stage of magmatic crystallization in lithium-sodium pegmatites

PLATE 19

- a. Extensive replacement of spodumene by a very fine-grained aggregate of albite and green mica from the columbite zone (unit 8) in Trench 3. Arrow points to relict spodumene core. A = coarser albite, B = beryl, Q = quartz and T = columbite-tantalite group.
- b. Lepidolite (L) replacement of pink spodumene (S) adjacent to pollucite (P) in Trench 2. Coin diameter = 2 cm.
- c. Mica-rich replacement veinlets (M) in pollucite (P) from Trench 2. Smoky quartz = Q.
- d. Garnet mass, probably enriched in spessartine, enclosed in a small aplite replacement mass (A) which occurs in the pollucite zone in Trench 2 of the Tot Lake Pegmatite. S = spodumene, P = pollucite, B = white beryl and Q = quartz.
- e. Abundant, euhedral columbite crystals (Appendix 41, analysis 408) in unit 8 at Trench 3. Distribution of columbite is strongly controlled by the margins of completely albitized, blocky to irregular K-feldspar. Note the presence of lenticular, partially to totally replaced spodumenes, some of which possess relict spodumene cores (arrow). Coin diameter = 2.4 cm.

PLATE 19

and is an important mineralogical indicator of tantalum enrichment (Ginzburg and Feldman 1978). Incipient to pervasive replacement by a fine-grained yellow-green mineral aggregate is typical for pollucite from the Tot Lake pegmatite. Early-stage replacement is commonly arranged in a polygonal manner (Plate 19c). Ucakuwun (1981, p.105) indicated that this interconnecting vein alteration consists of an unidentified microcrystalline mineral. Irregular masses of fine-grained calcite commonly replace pollucite as well.

The Tot Lake rare-element pegmatite represents one of only two known pollucite occurrences in Ontario, the other being the spodumene-pink elbaite -tantalite dykes near Lily Pad Lake in the Fort Hope area (Wallace, 1978).

Beryl is a relatively rare mineral in the Tot Lake pegmatite, found as sporadic subhedral to anhedral crystals in the columbite zone, in the orange K-feldspar zone, in the transverse-layered zone and in altered parts of the spodumene zone (unit 1). In the columbite-zone it occurs as white to light pink crystals interstitial to randomly oriented, almost completely replaced spodumene phenocrysts (Plate 19a). White euhedral beryl was also noted with a similarly altered spodumene near the northeastern end of Trench 2. The largest beryl occurs in the transverse-layered unit near Trench 3 as an 8 cm wide basal section.

Spessartine is rare in the Tot Lake pegmatite and tends to occur as orange, irregular masses, up to 5 x 6 cm, associated with pockets containing a replacement assemblage albite, green muscovite, and fine-grained dark blue tourmaline in the pollucite zone (Plate 19d). It

has also been noted interspersed with quartz in the columbite zone (unit 8) near the northwestern end. An analysis by Ucauwun (1981, p.104) compiled in Table 53, probably from this zone, indicated the highest MnO content (39.29 percent) of any garnet in the Dryden area.

Tourmaline is also uncommon in this pegmatite and occurs mainly in a narrow (<1 cm), fine-grained zone which follows for 10 m along the contact between a quartz vein and host rocks at the southeastern end. Fine-grained, dark green tourmaline is also developed sporadically in pockets of sodic aplite which occurs adjacent to the pollucite zone in Trench 2. Ucauwun (1981, p.95-99) interpreted tourmaline in the Tot Lake pegmatite as secondary in origin, on the basis of its association with replacement sodic aplite which is supported by this study.

Holmquistite is developed as fine-grained, unoriented acicular purplish crystals veins and on joint surfaces of the immediate wall rocks. The massive, coarse-grained porphyroblastic mafic metavolcanics that are the principal host, contain holmquistite for distances up to 1 m from the dyke contact.

Fluorapatite identified by X-ray diffraction, appears as a very rare, early mineral, occurring as pale blue euhedral crystals up to 0.5 x 1.2 cm in pollucite and also partially enveloped by spodumene in Trench 2. Fine-grained possible manganese oxide commonly coats the crystal faces.

Lepidolite was found only in the pollucite zone, as an alteration product of spodumene (Plate 19b).

Columbite-Tantalite group minerals are discernible only in the zones of intense albitization (unit 8 = Columbite Zone). Locally, these steeply black euhedral to subhedral crystals up to 1 x 2 cm (analysis 408, Appendix 41) compose 0.2 percent of the dyke (Plate 19e). The columbite zone is a quartz-rich unit containing completely albitized blocky K-feldspars with deposition of columbite to a large extent on the margin of these megacrysts. Less commonly, this mineral group may contact highly altered spodumene and scarce pale pink beryl (Plate 19a).

Tantalite (analyses 407A and B, Appendix 41) also occurs in the pollucite zone in Trench 2. However, this observation was confined to one euhedral 4 x 10 mm black crystal which contacts a small pocket of grey platy albite, situated between horizontally aligned pink spodumene, interstitial pollucite (altered to fine-grained green mica) and grey quartz.

X-ray characterization of these minerals was undertaken by Ucauwun (1981, p.109-115) who apparently analyzed material only from the northeastern end of the pegmatite. His work indicated that the columbite-tantalite is structurally disordered or partly ordered ("pseudo-ixiolite" of Nickel et al. 1963). Electron microprobe analyses revealed complex zoning, a trend towards Mn enrichment and tantalum generally exceeding niobium (Ucauwun 1981, p.113). Chemical data on Tot Lake pegmatite columbite-tantalite group minerals from this study will be dealt with in Chapter 8.

Sphalerite was documented by Ucakuwun (1981, p.64, 117) as a brown secondary mineral filling cavities up to 2 x 5 cm. This mineral occurs only within a slightly albitized, quartz-K-feldspar-spodumene-muscovite assemblage which is situated adjacent to the central quartz vein near Trench 2.

Chapter Five

REGIONAL AND INTERNAL ZONAL VARIATION OF RARE-ELEMENT PEGMATITES

INTRODUCTION

The concept of internal and regional zonation of pegmatite mineral assemblages has been extensively addressed in the literature. This section will compare and contrast pegmatite mineral zoning in the Dryden Field in light of a brief summary introduction of the pertinent literature.

ZONAL CLASSIFICATION SCHEMES

The landmark report of Cameron et al. (1949) revealed that systematic changes in mineral assemblages occurred in rare-element and spatially associated barren pegmatites within the confines of a single pegmatite district. They proposed that a maximum of eleven mineral assemblages may be developed as, for example, in the Black Hills pegmatite district. This classification was revised down to nine assemblages by Norton (1983) and was derived from more recent petrographic data distributed on a wider geographic basis than the original study. Minerals in each assemblage are listed in order decreasing relative abundance:

- 1) Plagioclase-quartz-muscovite
- 2) Plagioclase-quartz
- 3) Quartz-perthite-plagioclase, with or without muscovite
- 4) Perthite-quartz
- 5) Quartz-plagioclase-perthite-spodumene (petalite)-amblygonite
(pollucite may be present)

- 6) Quartz-spodumene (petalite?)
 - 7) Quartz-microcline (or perthite)
 - 8) Quartz
 - 9) Lepidolite or lithian mica - plagioclase - quartz - microcline
- Assemblages 3 and 5 have been divided by Norton (1983, p.857-858) to account for vertical variation in perthite content.

Heinrich (1953) apparently first used the term "regional zoning" in connection with pegmatite fields. The arrangement of pegmatites around granite masses with systematic changes in mineral assemblages, was one criterion that helped convince many workers such as Hutchinson (1955), Jahns (1955), Nedumov (1964) and Varlamoff (1972) that magmatic processes dominantly controlled pegmatite emplacement.

In an extensive, 30 year field study of Central and West African pegmatites, Varlamoff (1972) postulated in an important paper that barren and rare-element pegmatites occur in predictable zonal successions in a given pegmatite district, irrespective of the age of the associated orogeny. His generalized classification of pegmatite types shown in Figure 32 and Table 13 (back pocket), is basically a depth-zone scheme whereby the exact vertical positioning of barren and mineralized pegmatite swarms with respect to the nearby granite roof margin is dependent upon the depth of crystallization of this granite. It should be noted that Varlamoff's scheme does not include the minerals pollucite or petalite; these minerals should fall into his pegmatite types 6 or 7. Relatively high level granites are characterized by little spatial separation of pegmatite types above the cupola, as most are contained within the granite. Examples

Figure 32. Distribution of Barren and Rare-Element Pegmatites in Relation to Cupola of Parental Granites in Kivu Region, Zaire in Comparison to Mavis Lake Pegmatite Group. Diagram Modified Varlamoff (1972).

include the Maniema region of Zaire (Varlamoff 1972, p.206) and the MNW petalite-beryl-amblygonite-tantalite-cassiterite pegmatite in the Georgia Lake area (Breaks 1980).

Ginzburg et al. (1979) developed a more comprehensive depth zone classification of pegmatites which involved a greater diversity of types than was included by Varlamoff. This classification comprises four major categories, listed below with depth ranges:

- I. Pegmatite formation of shallow depths = Mirolitic pegmatites (1.5 -3.5 km);
- II. Pegmatite formation of intermediate depth = Rare metal pegmatites (3.5-7 km);
- III. Pegmatite formation of great depths = Mica-bearing pegmatites (7 to 11 km);
- IV. Pegmatite formation of maximal depths (>11 km).

Varlamoff's classification is approximately equivalent to categories I and II of Ginzburg et al. (1979).

Evidence for a genetic lineage between zoned pegmatite swarms and proximal granites is most convincingly provided by divisions I and II of the Ginzburg classification, although this is more common in Phanerozoic than in Precambrian terranes (Cerny 1982a).

REGIONAL ZONATION IN THE DRYDEN PEGMATITE FIELD

Pegmatites of the intermediate level (Ginzburg et al. 1979) commonly exhibit an almost complete vertical separation of rare-element pegmatites from the granite cupola contact. The Kivu region of Zaire (Varlamoff 1972, p.208) in this respect compares closely with the observed zoning in the Mavis Lake pegmatite group which includes

internal pegmatites of the parental Ghost Lake Batholith (Table 13). Here the succession of barren and rare-element pegmatites in Varlamoff's classification is very similar in mineral assemblages (Figure 32) and dimensions to those of the study area. This succession of study area pegmatites is also quite similar to classical zonation described in the Yellowknife-Beaulieu district (Hutchinson 1955), Preissac-Lacorne area of Quebec (Mulligan 1965) and the Issia et Touvre area of the Ivory Coast (Varlamoff 1972).

Varlamoff (1972, p.208) indicated that the maximum dimensions of pegmatite types 6 and 7 from Kivu are 10 m in thickness and 100 m in length, similar to those of spodumene pegmatites from the Mavis Lake group. The vertical distance between the apical contact of the granite and pegmatite types 7 and 7-8 was estimated to be 250-350 m by Varlamoff (1972).

Internal pegmatites of the Ghost Lake Batholith comprise types 1 to 5, although beryl-bearing potassic pegmatites (type 5) can either occur as internal or exocontact pegmatites, as exhibited respectively at the Dryden Airport and Contact Beryl Occurrences.

The volatile-enriched apical parts of granites interpreted by Varlamoff (ibid) as parental to rare-elemental pegmatites have notable concentrations of muscovite and tourmaline relative to deeper levels of the granite, which are typified by biotite-tourmaline and biotite pegmatites. The complex zone comprising GLB-4 to -8 at the Dryden Airport is characterized by muscovite-and tourmaline-enrichment and thus is considered to represent the highest exposed level in the batholith and the source region of rare elements for the nearby

mineralized pegmatites of the Mavis Lake group. This is suggested by the following features:

- 1) presence of rare-element minerals beryl and columbite within the batholith itself;
- 2) presence of an exogreisen vein system enriched in holmquistite, testifying to accumulation of high lithium concentrations in this part of the batholith;
- 3) high degree of chemical fractionation relative to deeper level internal pegmatites (Chapter 7);
- 4) virtual absence of accessory biotite, predominance of primary muscovite and tourmaline.

The Gullwing-Tot Lakes pegmatite group presents a less ideal zoning pattern than the Mavis Lake group. The west to east zonal progression of pegmatites apparently suggests an abrupt jump in the Varlamoff classification. This is demonstrated by the hiatus between the Gullwing Lake two-mica pegmatitic granite, bearing core zones of spodumene-columbite-beryl pegmatite (a variant of type 3 pegmatite), and the Tot Lake spodumene-pollucite-beryl-columbite-tantalite pegmatite (Varlamoff type 7 pegmatite). Pegmatites between these are either barren of rare-element minerals, or contain beryl-molybdenite and columbite-molybdenite mineralization (Varlamoff type 5 pegmatite).

This abnormal zoning trend may reflect a multi-granite source for the rare-element pegmatites, perhaps combined with a strong structural control which gained dominance over a simple, vertically disposed pegmatite swarm emplaced above a single granite roof zone.

MODAL VARIATION OF INTERNAL ZONES OF LITHIUM-RICH PEGMATITES

Modal variation on a global basis of major mineral phases in spodumene and petalite-bearing pegmatites is depicted in Figure 33, based upon data compiled from Stewart (1978, p.974) and Tables 9, 12 and 14. Figure 33 is identical to that utilized by Stewart (1978, p.974) in which modal spodumene is converted into eucryptite. This allows representation of the other members of the lithium aluminosilicate family (spodumene and petalite) on the eucryptite-quartz sideline. This diagram includes homogeneous spodumene pegmatites and internal units from zoned Li-rich pegmatites.

Stewart (1978, p.974) determined that Li_2O contents of homogeneous pegmatites fall within narrow limits. This is exemplified by the well-studied King's Mountain spodumene pegmatite which ranges between 1.23 and 1.85 Li_2O at the 95 percent confidence limits (2σ) according to Kesler (1964) and Stewart (1978). Stewart (1978) pointed out that Li_2O contents of lithium-rich homogeneous pegmatites and the zoned Etta pegmatite commonly fall within these limits (Figure 33) and that all pegmatite compositions that he studied plot in the sub-triangle petalite-quartz-total feldspar.

It should also be noted that most compositions amongst the homogeneous pegmatites are largely limited to an ovoid area outlined by dashed lines in the eucryptite-quartz-total feldspar diagram. This field is elongated approximately parallel to the quartz-feldspar sideline and most bulk compositions contain feldspar in excess of quartz (Stewart 1978, p.974). However, departures from these modal limits are obvious for many zonal compositions of spodumene pegmatites

Figure 33. Eucryptite-Quartz-Total Feldspar Modal Variation in Homogeneous and Zoned Lithium-Rich Pegmatites.

Table 14. Modal Analyses of Petalite and Spodumene Pegmatites from the Literature and Unpublished Data of the Author.

	1	2	3	4	5	6	7	8	9	10	11	12	13
Quartz	24	32	33	41	27	30	38	36.7	10	60	25	35	70
Plagioclase		27)			29	25	)	)		15	10	35	1
K-feldspar	2	14)	44		15	20	)	8.0	70	2	50	1	3
Spodumene		20	18	21	22	20	16	49.7		3	10	25	25
Petalite	71												
Muscovite	1	6	5	6	7	5	8	2.1	2	20	5	3	1
Beryl									3	<1			
Tourmaline													
Amblygonite	2							3.4					
Lithian-muscovite													
Biotite									15				

1. MNW Pegmatite, Georgia Lake area (49°14'20", 88°00'00"). Modal analyses of the quartz-petalite zone. Spodumene-quartz intergrowths counted as original petalite. Unpublished data of the author.
2. King's Mountain spodumene pegmatite, North Carolina (Kesler 1961).
3. King's Mountain spodumene pegmatite, North Carolina (Browning et al. 1961).
4. Val D'Or spodumene pegmatite, Quebec (Browning et al. 1961).
5. Peg Claims spodumene pegmatite, Maine (Sundelius 1963).
6. Georgia Lake spodumene pegmatite (Milne 1962).
7. Mateen spodumene pegmatite, South Dakota (Browning, Clemmons, and McVay 1961).
8. Leinster Spodumene Pegmatites, southeastern Eire. Average of 5 modal analyses in Leucke (1981, p.211). K-feldspar and albite not distinguished in modal analyses of Leucke (1981) but was tallied as alkali feldspar.
9. Etta Pegmatite, South Dakota (Norton 1964 for 9 to 13, inclusive). Microcline-biotite zone.
10. Etta Pegmatite, quartz-muscovite-albite zone.
11. Etta Pegmatite, perthite-quartz-spodumene zone.
12. Etta Pegmatite, quartz-cleavelandite-spodumene zone.
13. Etta Pegmatite, quartz-spodumene zone.

from the Dryden Field (Tables 9 and 12, Figure 33). These compositions fall mainly within the spodumene-petalite-total feldspar sub-triangle. These relate to higher contents of Li_2O (up to 5.27 percent) and hence of the calculated eucryptite end-member. These zones exceed the maximum limits of Li_2O concentration of 2.2 percent that theoretically can accumulate within a magma initially crystallizing in the albite field of the synthetic albite-quartz-eucryptite- H_2O system at $p\text{H}_2\text{O} = 2$ kbars (Stewart 1978, p.977). This limit corresponds to about 18.5 percent eucryptite or 27 percent spodumene.

The spodumene-rich Leinster Pegmatite of Luecke (1981) and the petalite-rich core zone of the MNW Pegmatite also plot within the petalite-spodumene-total feldspar sub-triangle. Thus there are exceptions to the compositional restrictions suggested by Stewart (1978, p.974), which are not confined to the Dryden Field.

Extreme spodumene contents of 77 percent occur in certain layers of the Tot Lake Pegmatite. This zonal composition represents the only one plotting within the eucryptite-spodumene-total feldspar sub-triangle of Figure 33. Petrogenetic mechanisms capable of generating such extreme spodumene contents have not been adequately addressed in the literature. Explanations arising from this study will be reserved for Chapter 9.

Considering the internal pegmatite zones from the study area, only the albite-quartz-spodumene pegmatite zones of Fairservice Pegmatites 1 and 4 contain an excess of quartz over feldspar (Table 9). This zone appears to closely match the internal zonal assemblage

6 of Cameron et al. (1949) revised by Norton 1983). Modal analyses of the quartz core zone from Fairservice pegmatite 1 were not taken, however its composition would plot near the quartz apex as only approximately ten percent of combined blocky K-feldspar, spodumene and green muscovite is present.

The presence of a silica-rich supercritical fluid phase has been deduced to play an important role in the evolution of complex rare-element pegmatites (Jahns and Burnham 1969, Stewart 1978). The pegmatite zones 6 to 9 of Cameron et al. (1949) revised by Norton (1983) have been interpreted by Stewart (1978, p.978) to result from a domination by an alkali-bearing supercritical fluid phase. Such zones according to Stewart (1978, p.978) are characterized by modal quartz exceeding total feldspar. This suggests that evolution of quartz-rich spodumene pegmatite zones in Fairservice Pegmatites 1 and 4 may have involved an appreciable vapour component.

Chapter Six

GEOCHEMICAL CYCLE OF LITHIUM

INTRODUCTION

Lithium is the most abundant rare-element in the majority of pegmatites of the Dryden Field. As a more complete set of abundance data exists in the literature for lithium in geological environments than for the other rare elements, it is thus useful to briefly examine its geochemical cycle. This literature survey serves as a background for the ensuing chapter on the chemistry of granites and pegmatites in the Dryden area.

As indicated in Figure 34, this cycle has a number of important segments. The fundamental processes controlling lithium distribution in each segment will be discussed in this chapter.

GEOCHEMISTRY OF LITHIUM

Lithium is generally grouped with the "rare elements", a geochemical classification that pertains to lithophile elements that are normally highly dispersed in crustal materials; only in rare circumstances do these elements achieve levels of concentration conducive to their economic extraction. Lithium is a member of the alkali group (Table 15). With the exception of Li, the alkali metal atoms are relatively large; these exclusive of Li have, in recent years, been included in the LIL or large ion lithophile geochemical category (Condie 1976). Characteristically all of the alkali elements are readily oxidized since loss of one valence electron is easily achieved as indicated by the relatively low ionization potentials (Table 15). It may also be noted that decrease in electronegativity potential progresses with increasing atomic number such that Cs is

Figure 34. The Geochemical Cycle of Lithium. Diagram Based upon Information Compiled by the Author.

Table 15. Some Chemical Properties of the Alkali Elements (after Ahrens 1952; Pauling 1948; Heier and Billings 1972).

	<u>Lithium</u>	<u>Sodium</u>	<u>Potassium</u>	<u>Rubidium</u>	<u>Cesium</u>
Atomic Number	3	11	19	37	55
Atomic Weight	6.940	22.991	39.100	85.48	132.91
Ionic Radius(Å)	0.68	0.97	1.33	1.47	1.67
First Ionization Potential	5.363	5.12	4.318	4.159	3.87
Electro-negativity	1.0	0.9	0.8	0.8	0.7
Observed Coordination Number	4,6	6,8	8-12	8-12	12

Table 16. Comparison of Some Chemical Properties of Cations Capable of Substituting for Lithium.

	<u>Ionic Radius (Å)</u>	<u>First Ionization Potential</u>	<u>Electronegativity</u>
Li ⁺	0.68	5.39	0.95
Mg ²⁺	0.66	7.64	1.2
Fe ²⁺	0.74	7.87	1.7
Al ³⁺	0.51	5.98	1.5

able to form bonds having the most ionic character of the alkali element group. Alkali elements have thus been described as "oxyphilic" due to strong attraction for oxygen, forming ionic bonds with this element group in silicate structures and consequently exhibiting pronounced concentration in the upper crust during geochemical evolution of the earth. Lithium however departs markedly from all other members of the alkali element group in its anomalously low ionic radius, a factor which has a great bearing on its distribution patterns.

The crystal chemistry of Li is mainly influenced by its small ionic radius compared to other alkali elements. Only limited substitution of Li for Na, the second smallest alkali element, occurs thus greatly restricting entry of Li into feldspar structures. The ionic radius of Li is much closer to Mg^{2+} , Fe^{2+} and Al^{3+} , (Tables 15 and 16) and hence a major proportion of Li in crustal materials occurs in diadochic substitution for these elements in the octahedral of ferromagnesian minerals as predicted by the classical Goldschmidt Rules (Goldschmidt 1937, 1954). Lithium mainly occurs in high concentrations in silicates and phosphates, as found in pegmatites and evaporite deposits (Table 17).

According to Rankama and Sahama (1950) and Howie (1955) mica minerals offer the most desirable lattice sites for the substitution of Mg by Li. Absolute lithium abundance and the Li/Mg ratio increases systematically in the series orthopyroxene-clinopyroxene-amphibole-muscovite-biotite. In peraluminous granites of the Cornwall area of England ion microprobe analyses by Wilson and Long (1983) revealed the

following series of increasing Li concentration: topaz < feldspar < quartz < tourmaline < muscovite < biotite.

SUBSTITUTION SCHEMES OF LITHIUM FOR OTHER ELEMENTS

Several substitution schemes have been proposed to account for Li concentrations in igneous and sedimentary minerals:

<u>Substitution</u>	<u>Mineral</u>	<u>Reference</u>
1) $\text{Li}^+ + \text{Al}^{3+} = 2\text{Mg}^{2+}$	Holmquistite	Whittaker (1969)
2) $2\text{Li}^+ + \text{Si}^{4+} = \text{Al}^{3+} + \square_{\text{VI}} + \text{Al}^{3+}$	Diocathedral micas	Hawthorne and Cerny (1982, p.72)
3) $3\text{Li}^+ = \text{Al}^{3+} + 2\square_{\text{VI}}$	Diocathedral micas	Hawthorne and Cerny (1982, p.72)
4) $\text{Li}^+ + \text{Si}^{4+} = \text{Fe}^{2+} + \text{Al}^{3+}$	Triocathedral micas	Hawthorne and Cerny (1982, p.75)
5) $\text{Li}^+ + (\text{Mg}^{2+}, \text{Fe}^{2+}) = \text{Fe}^{3+} + \square_{\text{VI}}$		R. Kretz, personal communication, 1989

$\square_n$ = vacancy of coordination number n.

The first substitution scheme is considered the most important by Ronov et al. (1970) and greatly influences partitioning of Li in ultramafic, mafic and most felsic igneous rocks. Substitution schemes 2, 3, and 4 are important situations where Li accumulates in residual pegmatitic melts or in Li-enriched two-mica granitoid rocks with primary muscovite. These substitutions are also probably of major importance in Mg-deficient clay minerals such as the kaolinite species.

LITHIUM DISTRIBUTION IN TERRESTRIAL MATERIALS

Igneous Rocks

In most crustal igneous rock types, lithium abundances generally exhibit limited dispersion about the mean overall crustal value, currently estimated to be 12 ppm (Heier and Billings 1972 and Table 18). This is evidenced by the limited range in mean values of

Table 17: Minerals that contain lithium as a major element. The most abundant minerals are underlined. Compiled from Heier and Billings (1972), Hawthorne and Cerny (1982), London and Burt (1982) and G.Y. Chao (personal communication, 1989)

I. SILICATES

Petalite	$(\text{Li}, \text{Na})\text{AlSi}_4\text{O}_{10}$
<u>Spodumene</u>	$\text{LiAlSi}_2\text{O}_6$
Eucryptite	LiAlSiO_4
Virgilite	$\text{LiAlSi}_5\text{O}_{12}$
Bikitaite	$\text{LiAlSi}_2\text{O}_6 \cdot \text{H}_2\text{O}$
Holmquistite	$\text{Li}_2(\text{Mg}, \text{Fe}^{2+})_3\text{Al}_2\text{Si}_8\text{O}_{22}(\text{OH})_2$
<u>Elbaite</u>	$(\text{Na}, \text{Ca}^{2+})(\text{Mg}, \text{Li}, \text{Al}, \text{Fe})_3(\text{Al}, \text{Mg}, \text{Fe})_6$
<u>Zinnwaldite</u>	$\text{K}_2\text{Fe}_2^{2+}\text{Li}_2\text{Al}_2\text{Si}_6\text{Al}_2\text{O}_{20}(\text{OH}, \text{F})_4$
Polyolithionite	$\text{K}_2\text{Li}_4\text{Al}_2\text{Si}_8\text{O}_{20}(\text{F}, \text{OH})_4$
Trilithionite	$\text{K}_2\text{Li}_3\text{Al}_3\text{Si}_6\text{Al}_2\text{O}_{20}(\text{F}, \text{OH})_4$
Ephesite	$\text{Na}_2\text{Li}_2\text{Al}_4\text{Si}_4\text{Al}_4\text{O}_{20}(\text{OH}, \text{F})_4$
Taeniolite	$\text{K}_2\text{Mg}_4\text{Li}_2\text{Si}_8\text{O}_{20}\text{F}_4$
<u>Lepidolite</u>	$\text{K}_2(\text{Li}, \text{Al})_6(\text{Si}, \text{Al})_8\text{O}_{20}(\text{F}, \text{OH})_4$
Masutomilite	$\text{K}_2(\text{Li}, \text{Al}, \text{Mn})_6(\text{Si}, \text{Al})_8\text{O}_{20}(\text{F}, \text{OH})_4$
Cookeite	$\text{Li}_2\text{Al}_8[\text{Al}_2\text{Si}_6\text{O}_{20}(\text{OH})_{16}]$
Manandonite	$\text{LiAl}_4\text{Si}_3\text{B}_3\text{O}_{10}(\text{OH})_8$
Bityite	$\text{CaLiAl}_2[\text{Si}_2\text{AlBeO}_{10}](\text{OH})_2$
Balipholite	$\text{BaMg}_2\text{LiAl}_3\text{Si}_4\text{O}_{12}(\text{OH})_8$
Baratovite	$\text{KCa}_7(\text{Ti}, \text{Zr})_2\text{Li}_3\text{Si}_{12}\text{O}_{36}\text{F}_2$
Emeleusite	$\text{Na}_4\text{Li}_2\text{Fe}_2^{3+}\text{Si}_{12}\text{O}_{30}$
Hsianghualite	$\text{Ca}_3\text{Li}_2\text{Be}_3(\text{SiO}_4)_3\text{F}_2$
Liberite	$\text{Li}_2\text{BeSiO}_4$
Sogdianite	$(\text{K}, \text{Na})_2\text{Li}_2(\text{Li}, \text{Fe}^{3+}, \text{Al})_2\text{ZrSi}_{12}\text{O}_{33}$
Neptunite	$\text{KLiNa}_2\text{Fe}_2\text{Ti}_2\text{O}_2\text{Si}_6\text{O}_{22}$
Mangan-neptunite	$\text{KNa}_2\text{Li}(\text{Mn}, \text{Fe}^{2+})\text{Ti}_2\text{Si}_8\text{O}_{24}$
Brannockite	$\text{KSn}_2\text{Li}_3\text{Si}_{12}\text{O}_{30}$

II. PHOSPHATES

<u>Amblygonite</u>	$(\text{Li}, \text{Na})\text{AlPO}_4(\text{F}, \text{OH})$
Montebtasite	$(\text{Li}, \text{Na})\text{AlPO}_4(\text{OH}, \text{F})$
<u>Lithiophilite</u>	LiMnPO_4
Triphylite	LiFePO_4
Ferrisicklerite	$\text{Li}(\text{Fe}^{3+}, \text{Mn}^{2+})\text{PO}_4$
Lithiophosphate	Li_3PO_4
Palermoite	$(\text{Li}, \text{Na})_4\text{SrAl}_3(\text{PO}_4)_8(\text{OH})_9$
Tavorite	$\text{LiFe}^{3+}\text{PO}_4(\text{OH})$
Tancoite	$\text{HLiNa}_2\text{Al}(\text{PO}_4)_2(\text{OH})$
Bertossaite	$(\text{Li}, \text{Na})_2\text{Ca}_4(\text{PO}_4)_4(\text{OH}, \text{F})_4$

III. FLUORIDES

Colquiraitite	CaLiAlF_6
Cryolithionite	$\text{Na}_3\text{Li}_3\text{Al}_2\text{F}_{12}$
Griceite	LiF

Table 18: Estimates of Alkali Metal Distribution in the Earth (ppm unless otherwise indicated)

	<u>Lithium</u>	<u>Sodium</u>	<u>Potassium</u>	<u>Rubidium</u>	<u>Cesium</u>
Mantle	1.4	9,500	1,100	4.2	0.10
Crust	12	2.5%	1.24%	37	1.7
Oceanic Crust	10	2.0%	1,400	2.5	1.0
Continental Crust	13	2.22%	1.6%	48	3.0
Percent of Total Earth Element Concentration in Continental Crust	4%	?	7-31%	24%	?
Reference	Heier & Billings 1972	Cocco et al., 1972	Heier & Billings 1972c	Cocco et al., 1964	Cocco et al., 1972a

selected lithic groups presented in Table 19). Exclusive of alkalic/peralkalic phonolitic extrusive rocks, the mean values of most plutonic and volcanic igneous rocks fall between 2 and 50 ppm. This observation prompted Taylor (1965) to conclude that lithium is not a particularly useful element to utilize as an overall index of fractionation for most types of igneous rocks. Note also in Table 19 that many mafic to ultramafic groups contain comparable levels of lithium to most classes of calc-alkalic and alkalic granitoid rocks. Enrichment of Li is however notable for the low calcium granites (Table 19).

Lithium is useful as an indicator of extreme fractionation where concentration in crustal igneous material exceeds 100 ppm (Taylor 1964, p.142-143), as exemplified in Table 20. Lithium in these rock types is accompanied by copious amounts of Rb, Cs and F. Extreme enrichment of Li along with F, P, B and Mn is recorded in some rare-element pegmatites, agpaitic nepheline syenites and ongonite dykes, the last being a topaz-bearing quartz keratophyre considered by Kovalenko et al. (1970, 1972, 1973a, b) to be a subvolcanic analogue of Li-F rare-element pegmatites. Ongonites contain the highest fluorine abundance of any natural granitic melt. The highest Li contents known in any volcanic rock occurs in macusonite, an aphyric, peraluminous rhyolitic obsidian from Peru (3440 ppm: London 1988).

In comagmatic plutonic rock series, an increase in the Li/Mg ratio with differentiation has been documented by numerous works (Strock 1936; Nockolds and Allen 1953, 1954, 1956) even though the absolute level of Li may drop towards the latest, most fractionated

Table 19: Lithium abundance data in various igneous rock-

<u>I. Plutonic Rocks</u>		
<u>Rock Type</u>	<u>Lithium Abundance (ppm)</u>	<u>References</u>
Ultramafic rocks olivine and olivine-pyroxene	2 ± 1.5	Beus (1964)
Ultramafic rocks (pyroxene-amphibole)	15 ± 5	Beus (1964)
Gabbros	18 ± 5	Beus (1964)
Diorites	27 ± 3	Beus (1964)
Granodiorites	20	Heier and Adams (1964)
High-Calcium Granites	24	Turekian and Wedepohl (1961)
Low calcium granites	40	Turekian and Wedepohl (1961)
Syenites	10	Heier and Adams (1964)
Nepheline Syenites	20	Heier and Adams (1964)
<u>II. Volcanic Rocks</u>		
Average Basalt	10	Taylor (1964)
Oceanic Tholeiitic Basalt	9 ± 6	Engel et al. (1965)
Alkalic Basalts from seamounts and islands	11 ± 5	Engel et al. (1965)
Andesites	12	Heier and Adams (1964)
Average Trachyte	27	LeMaitre (1962)
Phonolites, Mont Dore, France	70 ± 30	Letolle (1965)
Rhyolite	50	Heier and Adams (1964)
New Zealand (average and range)	36, 9-100	Ewart et al. (1968)

members. This also appears to represent the case of uranium pegmatites of the English River and Winnipeg River subprovinces. Breaks and Bond (in press) have shown that the steep positive slope of Mg-Li curves for five uranium pegmatites contrasts strongly with Li-Cs-Rb-Be rare-element pegmatites of the Tot Lake type revealing increasing Li levels at constant Mg. In these environments the residual concentration of lithium is so substantial that minerals with lithium as a major element are formed. In these pegmatite types a curious bimodal distribution of Li is often in evidence (Stewart 1978) which is not explained but may relate to interlayering of a spodumene-rich pegmatite phase and Li-poor aplitic phase, as seems to be the case at the Kings Mountain spodumene pegmatite deposit. The experimental work of Stewart (1978) in the synthetic lithium-enriched granite system suggests that a maximum level of 2.2 percent Li_2O can accumulate as a result of magmatic crystallization.

Lithium in Subaerial and Subaqueous Environments

Fractionation in these environments contributes to considerable variation in Li abundances. Enrichment is especially manifest in rocks composed of voluminous amounts of clay minerals, i.e. mudstones, as shown in Table 21 where the average marine shale exhibits a six-fold increase in Li over the average crustal value. According to the work of Keith and Degens (1959) the average Li contents of marine shales exceed fresh water equivalents by a factor of 1.7 and exhibit a range more than twice as great. This apparent surficial control on Li content in shale, coupled with a similar relation for boron, has been touted as a geochemical indicator of a marine sedimentary environment. Ronov et al. (1970) indicated that average Li content of platform

Table 20. Alkali Metal Contents of Highly Fractionated Igneous Material (in ppm).

	<u>Li</u>	<u>Na</u>	<u>K</u>	<u>Rb</u>	<u>Cs</u>	<u>References</u>
Cornwall Granites, England	635	22,600	42,900	420	47	Bowler (1958, 1959)
Stanningle Granites, Erzgebirge Germany	7,360	2,220	74,700	5743	524	Tischendorf (1977)
Riebeckite Younger Granites, Nigeria	260	35,800	36,000	510	1-5	MacLeod et al. (1971)
Muscovite-Biotite Granite, Blue Tier Batholith, Tasmania	360	2,440	38,200	1035	ND	Groves (1975)
Ongonite Dykes, Mongolia	1,750	30,300	24,000	1950	ND	Kovalenko et al. 1970, 1972, 1973a, b; Kovalenko 1973
Spodumene Pegmatites, Superior Province, Northwestern Ontario	3,641	46,180	18,000	2615	1463	Breaks and Bond in press
Aphyric, peraluminous rhyolite obsidian, Macusani, Peru	3,440	30,340	29,300	1160	580	London et al. (1989)

ND= not determined

Table 21. Lithium Content of Various Sedimentary Rocks.

<u>Rock Type</u>	<u>Lithium Abundance (ppm)</u>		<u>References</u>
	<u>x</u>	<u>range</u>	
Sandstones	38	7-93	Heier and Billings (1972)
Pelagic Sediment	59		Goluberg and Arrhenius (1958)
Marine Shales	76	4-400	Heier and Billings (1972)
Fresh-water Shales	67	20-155	Heier and Billings (1972)
Argillaceous Sediments	66	17-207	Heier and Billings (1972)
Average Pelite	56		Shaw (1956)
Platform Sediments	51		Ronov et al. (1970)
Geosynclinal Sediments	37		Ronov et al. (1970)
Dolostone	15		Heier and Billings (1972)

sediments is significantly higher than those characterizing geosynclinal basins. This difference was related to more extensive and prolonged weathering of parent materials, in addition to a tendency for constituent detritus to be subjected to several sedimentary cycles as contrasted to the considerably less weathered, more rapidly deposited geosynclinal sediments.

Exceptionally high Li contents (80-1500 ppm) are sometimes encountered in brines from salars (salt-pan deposits) in arid climatic zones, and these are currently being considered as sources for low cost, recoverable Li (Ericksen et al. 1978).

Lithium shows substantial enrichment in spilitized basalts relative to unaltered equivalents as shown by Vatin-Perignon et al. 1979. Lithium has an average of 80 ppm and a range of 20 - 200 ppm in spilitic basalts which, according to these workers, is a good indicator of spilitization.

Lithium in the surficial weathering zone: distribution of Li in the surficial environment segment of the geochemical cycle commences in the weathering zone and in this regard studies by Butler (1953, 1954), McLaughlin (1955), Lebedev (1957), and, Ronov et al. (1970) are especially pertinent. The work of Ronov et al. (1970) in particular demonstrates that a zonal distribution of Li is evident in the soil profile over various rock types and does not strictly relate to the original concentration of Li in the parent materials. Rather, differences in physicochemical and climatic conditions determine the ultimate Li content (Ronov et al. 1970, p.77):

- 1) Alkaline Stage - initial weathering is typified by disintegration of parent rock materials and leaching of Na, K, Ca, Mg and Li, producing alkaline conditions in the soil solution. The weathered profile here consists of primary mineral relicts and newly developed clay minerals (chlorite, smectite, illite);
- 2) Neutral to Weakly Acid Stage - more extensive weathering is reflected in elution of most alkali and alkali-earth elements (Table 22) and pH conditions gradually progress towards weakly acid values. This change promotes alteration of clay minerals of stage 1 to those of the kaolinite group.

It is important to note that Li significantly departs from other alkali element members in being preferentially concentrated during the second stage. An increase in Li occurs upwards in the kaolinite-dominant part of the weathering crust (Table 22), but substantial depletion is encountered in the laterite zone.

Lithium content in clays increases from glacial (65 ppm) to arid (83 ppm) to humid (100 ppm) conditions (Ronov et al. 1970, p.90). Ronov's results also allow deductions regarding the relation of Li to major cations in various clay minerals. Although montmorillonite clay species appear to favor substitution of significant levels of Li for Mg, Horstmann (1957), Ronov et al. (1970) and Tardy et al. (1972) indicate strong Li depletion compared to various other clay groups (Table 23). This is also substantiated by upward increase in Li/Mg ratios in the various weathering crusts (Table 22). However, a strong Li-Al correlation is more important than Li-Mg in clay horizons within

Table 22: Distribution of Lithium in Weathering Crusts developed on Amphibolite and Nepheline Syenite, Ukrainian Shield (after Ronov et al. 1970). Lithium Abundance in ppm, K, Al and Mg in Weight Percent

<u>I. Crust on Amphibolite Zone in Weathering Profile</u>							
	Li	K	Mg	Al	Li/Mgx10 ³	Li/Alx10 ³	Depth (m)
Laterite Zone (gibbsite and goethite)	35	0.06	0.28	23.95	12.5	0.15	95.5- 96.3
	45	0.07	0.24	14.25	18.8	0.32	96.3- 96.6
	38	0.10	0.10	1.98	38.8	1.92	97.9- 98.9
Goethite- Kaolinite Zone	118	0.11	0.05	18.96	236	0.62	98.9-100.9
	68	0.05	0.17	14.23	40.0	0.40	100.9-104.2
	60	0.06	0.10	13.50	60.0	0.44	107.5-110.7
	50	0.07	0.06	12.42	83.5	0.40	113.3-116.6
Disintegration Zone Parent Rock	20	0.27	1.49	8.32	1.3	0.24	130.3-132.3
	15	0.16	1.45	9.93	1.0	0.13	132.7-133.0
	67	0.25	4.75	8.23	1.5	0.82	133.5
<u>II. Crust on Nepheline Syenite</u>							
Kaolinite Zon	115	0.17	0.25	18.58	46.0	0.62	27.5- 29.0
	110	0.51	0.40	9.05	27.5	1.21	29.0- 30.5
Disintegration Zone	40	5.02	0.48	10.87	0.43	8.3	31.0- 32.0
	35	5.73	0.76	11.37	0.44	4.5	38.0- 41.0
	15	5.63	0.41	8.10	0.28	3.7	53.0- 56.0
	40	5.47	0.31	12.92	0.36	12.9	62.0- 64.5
	55	5.17	0.77	9.35	0.32	7.2	70.5- 74.5
Parent Rock	50	5.88	0.52	8.25	9.6	0.11	70.5

weathering crusts as studied in areas such as the Ukrainian Shield (Ronov et al. 1970), U.S.A. (Goldich 1938; Horstmann 1957) and Hong Kong (Brock 1943).

According to Ronov et al. (1970) most Li in weathering profiles is contained in clay minerals and is extracted by these minerals at all stages in the sedimentary cycle. This extraction is reflected in the balance of Li between the lithosphere and hydrosphere. Ocean water contains only about 0.03 percent of crustal Li in contrast to 2 percent of the total crustal Na (Ronov et al. 1970, p.98).

Transportation of Li-bearing kaolinite into the marine environment produces significant results:

- 1) Kaolinites of continental regions are readily converted into illites and montmorillonites which are low in Al_2O_3 and enriched in K and Mg; there are no data on the fate of Li here.
- 2) The Li-Al correlation, typical of continental clay deposits is diminished with increasing salinity. Substitution of Li for Mg becomes more important, as most readily observed in clays of salt lagoons (Ronov et al. 1970, p.92).

Lithium distribution in clay mineral groups: rocks composed of clay minerals contain the highest lithium contents of all sedimentary rocks. In most clay mineral groups, the exact site of lithium is not exactly known (Tourtelot and Meier, 1976), however there are three general possibilities:

- 1) Li is held in ion-exchange positions;
- 2) Li substitutes for major cations in structural sites;
- 3) Li resides in impurities

The amount of Li held in ion-exchange positions on the surfaces of clay minerals is considered to be relatively small. This is in part related to Li being the most easily replaced member of the alkali element group, the order of attraction for clay minerals decreasing in the sequence Cs>Rb>K>Na>Li (Grim 1968). Most of the lithium content of various clay mineral groups is considered to be held in strongly bound non-exchangeable structural sites or as impurities, as shown by very low Li concentrations in leachates recorded from kaolinite samples containing exceeding high lithium (Tourtelot and Meier, 1976).

Kaolinite group minerals have the lowest cation exchange capacity of any major clay mineral (Table 23), kaolinites surprisingly contain among the highest lithium values (Table 24). Because magnesium contents in most kaolinites are extremely low (Weaver 1975), it is considered that Li substitution for octahedral Al, as outlined earlier, is important, as demonstrated by Ronov et al. (1970, p.85) in clays developed in weathering crusts.

The cation-exchange capacity of kaolinites is considered by Grim (1968, p.189, 193) to be mainly reliant upon broken crystal edges. This effect was discussed by van der Marel (1958) who demonstrated a high degree of positive correlation between cation-exchange capacity and surface area of kaolinite. A maximum of about 17 mequiv/100 gm was recorded.

Some of the highest Li contents in clay minerals occur in flint clays from the eastern United States, which are defined by Keller as "a sedimentary, microcrystalline clay (rock) composed dominantly of kaolin which breaks with a pronounced conchoidal fracture and resists

Table 23: Cation-Exchange Capacities of Various Clay Minerals (Heaver 1975)

<u>Clay Species</u>	<u>Cation Exchange Capacity (Milliequiv./100 gms)</u>
Illite	10 - 40
Smectite	122, 115, 130
Illite-Smectite	12 - 77
Palygorskite	20 - 36
Kaolinite	7 - 9

Table 24: Lithium Contents of Various Clay Mineral Groups

<u>Clay Mineral or Groups</u>	<u>Lithium Content (ppm)</u>	<u>References</u>
Kaolinite	171	Horstmann (1957)
Dickite	42	Horstmann (1957)
Nacrite	61	Horstmann (1957)
Flint Clay	160-5100	Tourtlot and Tourtelot (1978)
Smectite Group	4-70	Tardy et al. (1972)
Smectite Group	56	Horstmann (1957)
Hectorite	100-1000	Glanzman et al. 1978
Bentonite	45	Horstman (1957)
K-Bentonite	48	Horstmann (1957)
Illites	48-120	Tardy et al. *1972)
Red Clay	58	Horstmann (1957)

stacking in water". This material can contain up to 5100 ppm Li (Table 24). However it was determined by Tourtelot and Brenner-Tourtelot (1978) that the exceptionally high quantities of Li in these kaolinite group clays are mainly due to impurities of cookeite, a lithium chlorite (Table 18) and not to extensive substitution of Li for Al.

Illite group clay minerals have very little Li distribution data (Table 24), however Li contents should be generally high due to substitution for Mg and the higher cation-exchange capacity which should enable greater amounts of Li to be held by adsorption. Shales which are moderately enriched in Li usually contain high amounts of illite.

Smectite group clay minerals have a very high cation exchange capacity (Table 23 but usually contains less Li than most kaolinites, perhaps leading to the conclusion that adsorption of Li is relatively insignificant and, secondly, that substitution of Li for Mg may not always be very high. No publications seem to offer any explanation for the puzzling inability of most smectites to concentrate much Li; an exception is hectorite, a trioctahedral smectite (Glanzman et al. 1978) which can contain up to 5100 ppm.

Lithium in marine sedimentary and diagenetic environments:

Kaolinites constitute the chief clay species formed in humid continental surficial environments and a significant quantity of Li is probably transported from such environments to marine basins, contained in detrital kaolinite. The work of Hirst (1962), who found

that the Li/Al ratio did not vary in a study of argillaceous and sandy sediments of the Gulf of Paria, suggested that most of the Li content is contained in clay structures, a deduction partly corroborated by the poor adsorption capacity of Li.

Kaolinites, being larger than other clay species, are generally deposited closer to shore (Dunoyer De Segonzac 1970). This observation, bearing in mind the higher Li contents of kaolinite clay species, helps explain the sharp decrease of Li in sediments deposited from the mouth of rivers such as the Mississippi River as documented by Welby (1958). Ronov et al. (1970), p.90, 93), also documented a similar relationship for sediments of the Russian Platform in which a decrease in Li from continental (94 ppm) to coastal marine (89 ppm) and pelagic (64 ppm) environments is generally seen. Li/Mg ratios also exhibit a steady decrease in this sequence.

Kaolinite is however, unstable under alkaline conditions and ultimately undergoes systematic conversion to illite by addition of K and Mg, probably due to contact with saline solutions and relatively high temperature (1000-2000 C) conditions during deep diagenesis (Dunoyer De Segonzac 1970, p.291). Weaver (1961b) considers that very little illite forms in the marine environment until after burial. Ronov et al. (1970) suggests that Li is retained during conversion of kaolinite to illite but gives no data for this generalization.

Illites of the Al-rich, disordered type (IMd form) are presently forming in continental environments from weathering of K-feldspars (Weaver 1975. p.19) and are considered the most common type of illite present in marine clays (Dunoyer de Segonzac 1970, p.314). It may be

suspected however that relatively little Li is contained in these illites as feldspars in general do not allow significant partitioning of Li (Heier and Adams 1963, p.297). Some detrital illite of the most stable 2M form is also being contributed from cold, dry continental climatic zones characterized by a lack of rapid weathering. These illites develop through degradation of biotites, chlorites, and muscovites. The fate of Li contents of these primary minerals, which can be appreciable (see below) is not known; however, cold dry climate characterized by an absence of hydrolysis would be conducive to negligible loss in Li through elution and hence most Li is probably ultimately transported in this illite type to marine environments.

Smectite clay species also undergo systematic "illitization" via diagenetic aggradation. This process involves systematic adsorption of Mg, K and Na from the interstitial fluid phase into interlayer positions and ultimately into octahedral and tetrahedral layers (Milliot et al. 1966). Two major pathways for this process have been recognized (Dunoyer de Segonzac 1970, p.311):

- 1) aggradation in a Mg-rich environment, which initially produces mixed-layer minerals of the chlorite-smectite type, and;
- 2) aggradation in a Na, K-rich environment which initially produces mixed-layer minerals of the illite-smectite-type.

Ultimately these two contrasting mixed-layer groups become completely unmixed at temperatures around 400°C, respectively recrystallizing to chlorite and illite (Dunoyer de Segonzac 1970, p.311). Thus the two chief retainers of Li in the sedimentary-surficial environment, montmorillonite and kaolinite, suffer complete destruction as conditions of diagenesis merge with those of burial

metamorphism. The subsequent fate of Li is probably dictated by entry into Mg, Al and possibly Fe octahedral positions of illites and chlorites, although data specifically bearing on distribution of Li in these minerals in this transitional environment are lacking in the literature.

Distribution of lithium in metamorphic and anatectic environments: fate of Li in the metamorphic segment of the geochemical cycle is much less known than for igneous, surficial, and sedimentary environments as there is a paucity of published work concerning distribution of this element as a function of metamorphism. The situation has, however, been remedied somewhat for Rb, K and Na in the years subsequent to publication of the extensive review paper of Heier and Adams (1963).

Shaw (1954, 1956) found that average Li contents of metapelites of the Littleton Formation, New Hampshire, do not change significantly in his low, medium and high grade zones. However, it must be pointed out that Shaw's "high grade zone" is actually a sillimanite-muscovite zone which does not exhibit anatexis.

Several studies, however, have demonstrated a general decrease in Li contents with increasing metamorphism. Sighinolfi and Gorgoni (1978) encountered a depletion pattern of Li in metasedimentary rocks of the Western Italian Alps, marked by an 80 percent decrease of Li contents in progressing from amphibolite to granulite facies.

Metapelites and metawackes of the Lake St. Joseph area in the English River subprovince of northwestern Ontario exhibit a systematic decrease in average Li contents from pre-anatectic, low and medium

grade zones to high grade and granulite areas (Table 25 and Breaks and Bond, 1982). This depletion trend also involves Rb, Cs, and U. Evans (1964) documented a substantial decrease in Li contents of pelitic hornfelses in contact with mafic and ultramafic rocks from Connemara, Ireland compared to unhornfelses pelites. Virtually no Li was found in pelitic xenoliths suggesting that vapor phase transport or diffusion may be important mechanisms to be considered. A similar pattern of depletion was also documented for Ba, Rb, La, Cs, Si, Al, K, Na, and H₂O. Kesler's (1976, p.49) hypothesis of leaching of loosely bonded Li from pelitic metasediments via volatiles emanating from the Cherryville Batholith, Kings Mountain, North Carolina, may thus have some validity in effecting pre-anatectic concentration of this element.

There are presently no known published data regarding partitioning patterns of Li between coexisting biotite-muscovite and muscovite-chlorite pairs in metapelites and metawacks as a function of increasing metamorphic grade and type of metamorphism. It is known from mineral chemistry studies on muscovites and biotites from granitoid rocks (Table 26) that substantial trace levels of Li can be partitioned in such minerals. In particular, the Li content of muscovites in such metasedimentary rocks presents a major unknown variable in application of a model of anatectic, S-type derivation for rare-element pegmatites. In two-mica granitoid rocks of the Preissac-Lamotte-Lacorne region of Quebec, Siroonian et al. (1959) showed that primary muscovites contain Li levels comparable to those of coexisting biotites (Table 26). Muscovites from various rare-element pegmatites not containing biotite curiously show lower average Li contents (Table

Table 25: Average and Range for Li, Rb, Ca, Ba, and U Contents (in ppm) of Metawacks and Metapelites from Lake St. Joseph area, English River Subprovince, as a Function of Metamorphic Grade.

<u>Zone</u>	<u>N</u>	<u>Li</u> <u>X range</u>	<u>Rb</u> <u>X range</u>	<u>Cs</u> <u>X range</u>	<u>Ba</u> <u>X range</u>	<u>U</u> <u>X range</u>
<u>Medium Grade</u>						
Metapelites	7	66 30-165	91 70-120	11 9-15	394 210-620	2.9 2.4 - 3.9
Metawacks	9	45 10-185	68 20-160	11 3-30	326 180-480	3.0 1.2 - 6.6
<u>High Grade</u>						
Metapelites	17	25 10-60	62 4-110	9 4-21	332 220-460	2.8 1.6 - 4.4
Metawacks	39	11 5-30	53 10-110	6 1-9	287 120-780	2.7 1.6 - 4.0
<u>Granulite</u>						
Metapelites	9	14 5-20	61 40-80	<6 <2-7	558 220-1200	1.8 1.2 - 2.3
Metawacks	9	10 6-25	42 20-60	<5 <5-25	379 220-780	1.0 0.9 - 2.3

n = number of samples

Table 26. Lithium Content (ppm) in Muscovite and Biotite from Various Granite and Pegmatite Rocks.

Rock Type	Location	Biotite	Muscovite	References
Biotite-muscovite granite, Lacorne Complex	Quebec	3120*	2740*	Siroonian et al. (1959)
Biotite muscovite granite, Lamotte Complex	Quebec	2080*	1480*	Siroonian et al. (1959)
Biotite muscovite granite, Preissac Complex	Quebec	1530*	2040*	Siroonian et al. (1959)
Two mica granite	Brittany, France	2900 (range not given)	775 (range not given)	Carron and Lagache (1972)
Hercynian muscovite biotite granite	Northern Portugal	1508 (970-1900)		De Albuquerque (1973)
Hercynian muscovite	Northern Portugal		193 (115-450)	De Albuquerque (1973)
Muscovite biotite granite (halite Pluton)	Nova Scotia	902 (175-1740)	223 (66-820)	Smith et al. (1982)
Royal tonalite-muscovite pegmatite	Near Sandy Creek, Far Fall, Ontario		530 (single analysis)	Breaks and Bond (in preparation)
Muscovite biotite-potassic pegmatite	Gullwing Lake dyke, Ontario	5650*	2270*	This study
Stodolene pollicite-biotite columbite-tantalite pegmatite	Tot Lake pegmatite, Ontario			This study
Biocrite biotite pegmatite	Barque Township, Ontario	20012*		Unpublished data of the author
Granite biotite pegmatite (Davidson occurrence)	Emeryville Lake area, Ontario	510 (single analysis)		Unpublished data of the author

* Asterisks indicate co-existence of biotite and muscovite

26). According to Gordiyenko (1970) the entry of lithium into the muscovite lattice is governed by the substitution sequence:

Lithium distribution in anatectic rocks: the concentration of Li in average metapelitic rocks is 6 times above the average crustal value of Heier and Billings (1972). This enrichment in Li and other rare-elements suggests that metapelitic rocks in the anatectic environment could play a crucial role in determining the evolution of Li-enriched granitoid rocks and pegmatitic affiliate. There appears to be an extremely common association on a global scale between two-mica granitoid rocks and rare-element pegmatites in high-grade metasedimentary belts of Archean to Proterozoic age. Several workers such as Stewart (1978) and Norton (1973, 1975) here contended that partial melting of Li-enriched metasedimentary sequences is the only feasible process to account for the formation of spodumene/petalite pegmatites.

Holland and Lambert (1972) suggested that anatectic fractionation of plutonic gneisses and metasedimentary rocks would produce a leucosome component depleted in Li and a corresponding biotite-rich restite contrastingly enriched in Li. These workers assumed that all Li is partitioned into the biotite of the restite or melanosome component (i.e. mafic selvages in metatexites, mafic clots in diatexites). Hence ultrametamorphic conditions of granulite facies would be necessitated in order to release most of the Li to advanced-stage anatectic melts. These workers neglected the possible implications of pre-anatectic vapor phase transport and the likelihood

that, in pelitic metasediments at least, muscovite could contain significant levels of Li which would be rendered available to the anatectic melt above the second sillimanite isograd, by reactions similar to:

This reaction assumes that significant quantities of lithium are hosted in the muscovite structure.

In the Lake St. Joseph area a drastic drop in Li levels is obvious in surviving metawacke and metapelitic paleosome from migmatitic rocks exhibiting low amounts of partial melting. These high-grade metapelites show high concentrations of biotite (30-50 percent) yet the rock as a whole contains less than half of the Li present in its low-to medium-grade counterparts. In this case it may be concluded that greater than 50 percent of the Li was released to the leucosome and restite migmatitic components through muscovite breakdown which may have been augmented by vapor phase stripping of loosely bound Li prior to anatexis.

In order to release to anatectic melts that Li which has substituted for Mg in biotite, dehydration of this mineral under granulite metamorphic conditions is required. Extraction of Li here is possibly governed by reactions similar to:

Chapter Seven

PETROCHEMISTRY

ZEALAND CLASTIC METASEDIMENTS

INTRODUCTION

Norton (1973) and Stewart (1978) postulated that clastic metasediments serve as concentrators of rare elements which are directly released into granitic melts via anatexis. Although lithium-rich metasediments were hypothesized by Stewart (1978), he presented no data on the initial Li concentration required in the protolith nor on the melting mechanism that would be necessary for the anatectic fractionation model to be effective. More generally, virtually no data are available concerning the distribution of rare elements in clastic metasediments of rare-element pegmatite districts. If the anatectic derivation of pegmatites has any significance then it is crucial to examine the distribution of the rare-elements Li, Cs, Be and associated volatiles F and B in clastic metasedimentary rocks under a variety of metamorphic conditions. The Dryden area represents an ideal one in which to investigate the distribution characteristics of such elements since clastic metasediments are widespread and vary from a low grade of metamorphism to high grade and migmatitic.

The sampling procedure involved selection of 1-2 kg of reasonably fresh material from metawacke (59 samples) and metapelite (71 samples) units distinguished on the basis of differences in metamorphic assemblages and grain size (see section on metamorphism, Chapter 2). Avoidance of sampling near leucosome veins and/or intrusive granite dykes was always exercised. A uniform geographical distribution of samples from most metamorphic zones was attained

(Figures 9, 10 and 38). However, specimens of metapelites and metawackes from the chlorite-biotite zone and the sillimanite-K-feldspar zone (non-migmatitic) are comparatively scarce. As such the data from these zones must be treated more tentatively.

MAJOR ELEMENT VARIATION

Nineteen Zealand metasediments were analyzed for major and selected trace elements (Tables 27 and 28). Rocks classified in the field as wackes and pelites generally exhibit a chemical similarity to analyses of greywacke and argillite-slate groups in the literature. According to Pettijohn (1963) an excess of Na_2O over K_2O is one of the prime chemical characteristics of metawackes. Three analyses from the study area (Table 27) demonstrate this feature, plotting within the centre of Pettijohn's greywacke field (Figure 35). These analyses also compare closely (Table 27) with the average greywacke of Pettijohn (1963, p.515) and the average metawacke from the English River Subprovince (Breaks and Bond, in press).

In contrast, the metapelites consistently exhibit $\text{Na}_2\text{O}:\text{K}_2\text{O}$ ratios less than unity and mostly fall within the argillite-slate field of Pettijohn (1963). Many analyses are comparable (Table 28 and 29) to the average Precambrian lutite of Nanz (1953), the average metapelite of Shaw (1954) and the average metapelite from the English River Subprovince (Breaks and Bond, in press).

There is no obvious variation in metapelite bulk composition as a function of metamorphism. An $\text{Fe}(\text{total})\text{-Mg-Na+K}$ variation diagram for the study area metapelites (Figure 36) shows that most fall within the field for the English River Subprovince.

Figure 35. Na₂O vs. K₂O Variation in Weight Percent of Zealand Clastic Meta-Sediments of Different Metamorphic Grade. Stars Represent Average Wackes and Pelites from the Literature in Tables 27 and 28.

Table 27: Chemical Analyses of Zealand Metawackes Compared to Average Wackes from the Literature. Major Elements in Weight Percent; Trace Elements in ppm. Sample Localities on Figure 38.

	Chlorite- Biotite Zone 32-9-1A	Andalusite F24-8A	Zone 32-5-7A	Average Wackes 4 5	
SiO ₂	65.7	60.1	67.7	66.73	66.75
Al ₂ O ₃	16.2	16.0	16.2	15.53	13.45
Fe ₂ O ₃	1.26	4.67	1.38	3.83	3.54
FeO	4.03	6.60	2.98	0.87	1.60
MgO	2.22	1.89	2.13	2.49	2.15
CaO	2.28	3.06	3.40	2.67	2.54
Na ₂ O	2.87	3.38	2.58	3.35	2.93
K ₂ O	2.22	2.03	1.87	1.94	1.99
TiO ₂	0.54	0.93	0.48	0.57	0.63
P ₂ O ₅	0.11	0.12	0.10	0.12	0.16
S	0.09	0.19	0.34	0.08	0.07
MnO	0.09	0.17	0.12	0.07	0.12
CO ₂	0.94	0.53	0.37	0.12	1.24
H ₂ O ⁺	1.10	0.60	0.46	0.53	2.42
H ₂ O ⁻	0.48	0.61	0.41	0.40	0.55
TOTAL:	100.10	100.90	100.50	99.83	100.23
B					
Ba	470	560	320	297	
Be	<1	<1	4		
Co	<10	14	17	16	
Cr	173	160	750	169	
Cs	11	10	6	7	
Cu	24	44	87	33	
F					
Li	34	44	16	18	
Nb					
Ni	23	18	56	42	
Pb	32	23	22	27	
Rb	90	90	60	58	
Sc					
Sn	<3	<3	<3		
Sr	260	190	290	313	
Ta					
Th	10	<10	20	2.6	
U	4	<1	16		
V	100	250	90	88	
Y					
Zn	66	210	108	76	
Zr	200	200	200	154	

4. Average metawacke from the English River Subprovince (77 analyses in Breaks and Bond, in press)

5. Average of 61 greywackes of various ages (Pettijohn 1963, p.515).

Table 28, continued

	SILLIMANITE-MUSCOVITE ZONE				SILLIMANITE-K-FELDSPAR ZONE			
	F22-1	G4-1	32-5-4	Non-Migmatitic F6-1	Migmatitic G8-1	F9-6	F7-8	
SiO ₂	63.3	57.5	57.4	57.6	55.2	61.5	59.3	
Al ₂ O ₃	17.9	20.3	19.2	19.00	20.7	18.2	18.8	
Fe ₂ O ₃	1.68	2.29	1.75	2.08	2.36	1.06	1.52	
FeO	4.73	6.92	6.52	6.86	7.61	6.28	6.65	
MgO	3.31	3.95	3.62	3.67	3.98	3.16	3.34	
CaO	1.44	0.79	0.79	1.07	1.08	1.33	1.54	
Na ₂ O	1.10	0.86	0.98	1.43	1.58	1.69	1.71	
K ₂ O	3.30	3.51	3.82	4.33	3.36	3.24	3.74	
TiO ₂	0.63	0.77	0.82	0.84	0.88	0.71	0.78	
P ₂ O ₅	0.07	0.07	0.16	0.07	0.09	0.12	0.03	
S	0.02	0.06	0.10	0.08	0.16	0.11	0.18	
MnO	0.05	0.08	0.09	0.09	0.01	0.08	0.01	
CO ₂	0.15	0.34	1.10	0.15	0.10	0.78	0.09	
H ₂ O*	1.62	1.73	1.22	1.96	1.83	0.90	1.55	
H ₂ O ⁺	0.16	0.19	0.42	0.31	0.23	0.32	0.25	
TOTAL	99.46	99.36	98.00	99.54	99.26	99.50	99.58	
R	240	40	1400	6	170	660	10	
Ba	600	610	<1	825	660	<1	700	
Be	8	2	21	4	3	27	3	
Co	13	17	21	25	28	214	21	
Cr	134	183	217	202	205	14	196	
Cs	20	20	31	10	80	20	20	
Cu	26	26	46	55	86	46	40	
F	530	430		400	610	480	480	
Li	86	46	94	76	590	27	66	
Nb	30	35		35	35	35	35	
Ni	34	56	62	86	99	84	80	
Pb	18	12	18	19	15	20	17	
Rb	120	120	120	170	260	90	130	
Sr	17	25		22	25	22	22	
Sn	3.8	3.1	7	2.5	5.8	3	2.5	
Sp	190	130	110	190	160	150	220	
Ta	100	130		130	130	130	130	
Th	20	10	10	10	20	<10	<10	
U	3	3	11	4	3	<1	1	
V	110	145	200	165	170	150	145	
Zn	6	5		6	4	3	3	
Zr	62	51	13*	112	46	87	110	
Σ	115	115	100	150	130	200	120	

Table 28, continued

	ANGALUSITE ZONE		SILLIMANITE-MUSCOVITE ZONE		SILLIMANITE-K FELDSPAR ZONE	
	G8.2	G10.1	F22.1	G4.1	G8.1	F7-A
Y			9.1	7.5	9.5	7.2
La	30	32	39	3.9	49	36
Ce	63	67	77	82	97	71
Nd	25	27	29	33	40	29
Pr	6.4	6.8	7.8	8.4	10	7.5
Sm	4.4	4.6	5.1	5.7	7.1	5.1
Eu	1.1	1.1	1.1	1.2	1.3	1.1
Gd	3.1	3.1	3.5	3.9	4.6	3.5
Tb	0.4	0.39	0.44	0.47	0.56	0.43
Dy	2	2	2.1	2.1	2.6	1.9
Ho	0.37	0.34	0.35	0.31	0.4	0.29
Er	0.96	0.80	.78	0.66	0.79	0.55
Tm	0.15	0.12	0.11	0.09	0.09	0.07
Yb	0.97	0.80	0.73	0.6	0.61	0.45
Lu	0.15	0.12				
ΣREE	138.0	146.17	167.12	175.64	214.22	156.96
La/YbN	20.0	26.1*	34.8	42.5	51.5	53.0
Eu/Eu	1.03	0.90	0.59	0.58	0.71	0.89

Table 29. Average Composition of Zealand Metapelites from various Metamorphic Zones Compared to Metapelites from the Literature.

	Andalusite Zone	Sillimanite Muscovite Zone	Sillimanite K-Feldspar (non-migmatitic)	Sillimanite K-Feldspar (migmatitic)	Average Metapelites		
					1	2	3
SiO ₂	59.1	59.4	56.4	60.4	57.42	56.30	61.54
Al ₂ O ₃	18.8	19.7	19.9	18.5	19.88	17.24	16.95
Fe ₂ O ₃	1.99	1.91	2.22	1.29	6.46	5.09	3.90
FeO	5.96	6.06	7.24	6.47	1.59	3.83	2.56
MgO	3.42	3.63	3.83	3.25	3.94	2.54	2.52
CaO	1.17	1.01	1.08	1.44	1.40	1.00	1.76
Na ₂ O	1.54	0.98	1.51	1.70	1.96	1.23	1.84
K ₂ O	3.89	3.54	3.85	3.49	3.57	3.79	3.45
TiO ₂	0.72	0.74	0.86	0.75	0.78	0.77	0.82
P ₂ O ₅	0.10	0.10	0.08	0.08	0.12		
S	0.03	0.06	0.12	0.15	0.10		
nO	0.08	0.07	0.05	0.05	0.09		
CO ₂	0.60	0.53	0.13	0.04	0.28	0.84	1.67
H ₂ O ⁺	1.78	1.52	1.90	1.23	1.38	3.31	3.47
H ₂ O ⁻	0.33	0.26	0.27	0.29	0.51		0.28
SO ₃							1.98
FeS ₂							1.18
TOTAL	99.51	99.51	99.44	99.13	99.48	100.00	100.48

1. Average metapelite from the English River Subprovince (51 analyses in Breaks and Bond, in press)

2. Average of 33 Precambrian lutites (Hanz, 1953, p.57)

3. Average of 155 metapelites (Shaw 1954, p.928).

Figure 36. Na+K-Total Fe Variation (weight percent) in Zealand Metapelites. English River Subprovince Field Based Upon 35 Analyses in Breaks and Bond (in press).

DISTRIBUTION OF SELECTED LITHOPHILE TRACE ELEMENTS

Partial analyses of 130 Zealand clastic metasediments for B, Be, Cs, F, Li, and Sn are tabulated in Appendix 2. Mean abundances and ranges from Appendix 2 have been summarized in Table 30.

Of the trace elements determined, lithium, cesium and boron exhibit the most significant variation. Pronounced enrichment in lithium is recorded in several metapelitic rocks (Table 30) representing all metamorphic zones between the chlorite-biotite and the non-migmatitic sillimanite-K-feldspar (non-migmatitic). Average lithium values for shales in the literature vary between 42 ppm and 66 ppm (Strock 1936, Shaw 1954, Horstman 1957 and Cameron and Garrels 1980). The average lithium abundances of the sillimanite-muscovite zone (107 ppm) and the sillimanite-K-feldspar zone (275 ppm) are significantly higher than the reported averages. Average values for the chlorite-biotite zone (50 ppm) and the andalusite zone (63 ppm) compare with the average shale values.

It should be noted that several analyses of metapelites reveal very high lithium concentrations (375 and 590 ppm in Appendix 2) which have only rarely been reported in the literature. Only Shaw's (1954) values of 500 ppm and 630 ppm in metapelites from the Littleton Formation in New Hampshire are comparable.

Lithium distribution in the study area metapelites shows a strong relationship with metamorphic grade. Average values of 63, 107 and 275 ppm for the andalusite, sillimanite-muscovite, and non-migmatitic sillimanite-K-feldspar zones contrast with a sudden drop to 47 ppm and contraction of range to 27-76 ppm in the migmatized blank

Table 36. Average Abundance and Range of L, Cs, F, Li and Sn in Zealand Metawackes and Metapelites in Relation to Metamorphic Grade.

Metamorphic Zone	Li	Range	Ca	Range	Sn	Range	F	Range	B	Range	n
METAWACKES											
Chlorite-Biotite	28	22-34	<10	<10-11	3.0	1.9-4.0	277	200-400	12	5-17	6
Andalusite	67	16-235	<17	<10-100	2.9	1.1-4.1	413	370-830	21	<5-120	14
Sillimanite-Muscovite	119	24-355	<17	<10-50	4.1	1.5-6.7	689	300-2140	46	<5-160	7
Sillimanite-K-feldspar (non-migmatitic)	80	51-108	<10	-	2.35	2.3-2.4	445	350-540	23	14-32	2
Sillimanite-K-feldspar (migmatitic)	38	11-94	<10	<10-20	2.4	1.5-4.8	540	310-1000	<5	<5-8	12
Inclusions in Ghost Lake Batholith	36	26-48	<10	<10-20	2.4	1.0-5.3	569	440-770	<7.5	<5-24	10
METAPLITES											
Chlorite-Biotite	50	16-114	<57	<10-150	4.2	3.0-5.3	477	370-600	61	33-98	4
Andalusite	63	20-210	<10	<10-31	3.1	1.9-5.3	515	460-740	65	5-230	25
Sillimanite-Muscovite	107	37-375	<27	>10-90	3.6	1.4-7.0	520	370-710	204	5-560	10
Sillimanite-K-feldspar (non-migmatitic)	275	64-590	28	10-80	3.7	2.3-5.8	595	400-720	49	6-170	6
Sillimanite-K-feldspar (migmatitic)	47	27-76	<11	<10-20	2.3	1.4-3.1	540	410-630	<8	<5-16	11
Inclusions in Ghost Lake Batholith	46	29-67	<10	<10-10	1.6	1.3-1.8	513	340-650	<5	-	3

n = number of samples

metasedimentary terrane west of the Ghost Lake Batholith. Inclusions of metapelitic rocks from this batholith have an average and range virtually identical to that of the migmatitic metapelites. It is not understood why Li has a lower average (50 ppm) in the chlorite-biotite zone relative to the ensuing three higher grade metamorphic zones (Table 32), although this value may reflect inadequate sampling as only four analyses were performed.

Cesium is also enriched in metapelites. However, a less confident appraisal of data can be made as cesium concentrations of this study commonly fall below the 10 ppm analytical sensitivity which is well above its mean crustal abundance (clarke value) of 3 ppm (Heier and Adams 1963). There is also considerably less comparative data in the literature on the distribution of cesium in metasediments. Metapelites in the chlorite-biotite, andalusite, sillimanite-muscovite and non-migmatitic sillimanite-K-feldspar zones have, respectively, maximum Cs contents of 150, 31, 90, and 80 ppm (Table 30). These values are significantly above the ranges given for shales by Carney (1952, Cs < 2 to 18 ppm) and Horstman (1957, Cs < 2 to 14 ppm). The maximum cesium contents in the migmatitic metasediments and Ghost Lake Batholith inclusions drops considerably, to 3 and 7 times clarke, respectively.

Metawackes have lower Li and Cs contents and a generally more limited range than the metapelites (Table 30). Direct comparison with metawackes in other areas is difficult because of the paucity of Li, Cs, and Be data in the literature. Taylor (1965, p.142) gives an average Li value of 30 ppm for greywackes. Mean Li content of

sandstones is 38 ppm (range 7-93 ppm) according to Heier and Billings (1973). No data dealing with Li abundance in Archean metawackes were found in the literature. Mean Li abundances in the Zealand metawackes are either comparable to the above averages (chlorite-biotite, migmatitic sillimanite-K-feldspar zones, and metawacke inclusions in the Ghost Lake Batholith) or exhibit substantial enrichment. Both the mean abundance and range show a marked increase in non-migmatitic metawackes of the andalusite (67 ppm), sillimanite-muscovite (119 ppm) and non-migmatitic sillimanite-K-feldspar (80 ppm) zones.

Maximum Cs contents in the low and medium grade metawackes vary between 3.7 and 33 times the clarke value of 3 ppm. An abrupt decrease in the maximum cesium content is obvious upon entry into the non-migmatitic and migmatitic sillimanite-K-feldspar zones and in inclusions from the Ghost Lake Batholith.

Argillaceous rocks contain the highest boron levels of any exogenic rock group (Christ and Harder 1973). Harder (1959b) reported a mean value of 100 ppm boron for argillaceous sediments while Shaw and Bugry (1960) have documented a mean content of 160 ppm boron. These averages indicate a substantial concentration of boron above its clarke value of 10 ppm (Taylor 1962) during the sedimentary cycle. Studies involving variation of boron in metasedimentary rocks from a single regional metamorphic terrane containing low, medium and high grade zones could not be found in the literature. Only minor losses in boron are said to occur under conditions of diagenesis and low grade metamorphism (Harder 1961). According to Christ and Harder (1973, p.M-3) boron contents in high grade metamorphic rocks are much

lower than those of low grade, although little of their data are drawn from metapelites and metawackes which have undergone anatexis.

Metapelites from the study area exhibit significant boron variation as a function of increasing metamorphic grade. The non-migmatitic zones show in Table 30 a rise in mean boron content for the chlorite-biotite (61 ppm) and andalusite (65 ppm) zones to the sillimanite-muscovite (204 ppm) zone. The high grade zones in contrast exhibit an abrupt drop in mean boron content which suggests loss into a vapor phase in the non-migmatitic sillimanite-K-feldspar zone (49 ppm B). A drastic degree of depletion is evident in the migmatitic sillimanite-K-feldspar zone (mean boron <8 ppm) which is comparable to that of the lower continental crust (9.3 ppm: Truscott et al. 1986). A similar metamorphic zonal pattern is also apparent for mean boron in the Zealand metawackes (Table 30), which also reveals a pronounced drop in the transition from medium grade (21 to 46 ppm boron) to the migmatitic sillimanite-K-feldspar zone (<5 ppm mean boron).

Fluorine contents in the Zealand metasediments are more erratically distributed and do not exhibit significant variation with regional metamorphic grade. Average values for metapelites do not depart significantly from the clarke value of 544 ppm. However, the metapelites as a group are somewhat depleted in fluorine relative to the abundances in average shales of 800 and 940 ppm (Fleischer and Robinson 1963). Metawackes in the non-migmatitic zones generally contain lower average F contents than in metapelites from the same zone (Table 30), except in the case of the sillimanite-muscovite zone,

where metawackes contain higher average F (689 ppm) than is found in the metapelites (520 ppm). Fluorine contents of metawackes from the migmatitic sillimanite-K-feldspar zone (540 ppm) and in metawacke inclusions from the Ghost Lake Batholith (569 ppm) are comparable to those of the equivalent metapelites; they are higher than in metawackes of the chlorite-biotite (277 ppm) andalusite (413 ppm) and non-migmatitic sillimanite-K-feldspar (445 ppm) zones.

It should also be noted that the greatest F enrichment in single samples of in Zealand metasediment is apparent in metawackes from the sillimanite-muscovite zone (2140 ppm) and non-migmatitic sillimanite-K-feldspar zone (1000 ppm).

It is obvious from Table 30 that average Li, Cs, F and B in the low and medium grade zones increase in the direction of increasing metamorphic grade and towards the GLB contact. Although the reason for this is not clearly understood, it does not appear to be a result of exomorphic dispersion from the GLB. The plutonic unit here (GLB-3) does not show evidence of albitization, nor do exomorphic mineral segregations occur along the batholith contact or in adjacent metasediments. This post-magmatic process could account for redistribution and concentration of these elements however it only has been documented for host rocks adjacent to chemically evolved units in the eastern lobe of the GLB. This pattern contrasts with that of the supracrustal rocks hosting the Mavis Lake pegmatite group (see Exomorphic Chemical Dispersion in Chapter 7).

Rare Earth Element Chemistry

Rare-earth element analyses of six metapelites are given in

Table 28. These samples, of varying metamorphic grade, exhibit similar steep chondrite-normalized curves marked by total REE contents between 138 and 214 ppm and La/Yb_N ratios between 20 and 53. The latter is a significant feature because Taylor and McLennan (1984) considered La/Yb_N ratios to be relatively low in Archean metasediments. The La/Yb_N ratio in their average Archean shale, plotted for comparison in Figure 37, is only 6.8, and indicates a significant enrichment in HREE relative to the study area metapelites. Coupled with the more strongly fractionated nature of the Zealand metapelites is the presence of small negative Eu anomalies in five of the six analyzed samples (Figure 37); the ratio Eu/Eu^* in Table 28 varies between 0.58 and 1.03. Archean metapelites do not normally possess such patterns, according to Taylor and McLennan (1984, p.52).

Figure 37. Chondrite-Normalized REE Curves for Zealand Metapelites Compared to the Average Archean Shale of Taylor and McLennan (1984).

DISTRIBUTION OF Ba, Cs, Li, Rb AND U IN CLASTIC METASEDIMENTS OF THE LAKE ST. JOSEPH AREA

Variation in selected lithophile trace elements (Li, Cs, Rb, Ba and U) in metasediments as a function of regional metamorphic zoning has also been studied elsewhere in the English River Subprovince (Figure 8 and Breaks 1982, Table 25). The Lake St. Joseph area (Breaks et al. 1978) is situated 115 km of the Dryden Pegmatite Field. A metamorphic facies series is developed in this subprovince near its interface with the metavolcanic-rich Uchi Subprovince and exhibits a continuous range from medium to granulite grade. Lithophile element variation in this area is pertinent to the Dryden Pegmatite Field since it constitutes a similar lithologic-metamorphic terrane and also contains rare-element pegmatites (Root Lake Pegmatite Field, Mulligan 1965, Breaks and Bond in press and Figure 4).

Mean values of Li, Cs, Rb and U all exhibit varying degrees of depletion in the high grade zones relative to the medium and low grade zones. Loss of Li is greatest amongst the elements studied. Metapelites and metawackes from the high grade migmatitic zones respectively show a 66 and 76 percent decrease in abundance relative to that of the medium grade zone. Depletion within the high grade zone relative to the medium grade zone is also notable for Rb and to a much lesser extent Cs and U. Uranium does not record any appreciable depletion in either rock group except at granulite conditions, where losses of 38 and 66 percent relative to the medium grade zone are documented for metapelites and metawackes, respectively.

SUMMARY

In summation, there has not been adequate research conducted

upon distribution of rare elements, boron and fluorine in metasediments spatially associated with Archean rare-element pegmatite fields. Results from this study indicate that substantial trace level enrichment of Li and Cs is present in some clastic metasediments of the Dryden area. It has also been established that considerable depletion of Li, Cs, Rb, and B has occurred in migmatized metasediments of high grade metamorphic zones relative to lower grade zones in the ERS and WRS. The most drastic depletion documented is of Li and B. Boron is of potential significance as a complexing agent for rare elements in anatectic melts of the areas studied, given its increased abundance in highly fractionated plutonic units of the GLB (see Chapter 7) and widespread presence in metasomatic rocks in the Dryden and Lake St. Joseph area.

GHOST LAKE BATHOLITH

MAJOR ELEMENT VARIATION

In Appendices 3 to 8 inclusive, 58 analyses of the Ghost Lake Batholith and related exocontact granitic dykes are presented. These appendices also include trace elements, selected element ratios, Barth mesonorms and modal analyses. Localities for all analyzed rock and mineral specimens from this batholith are given in Figure 38.

Harker diagrams (Figures 39, 40, and 41) reveal variation with SiO_2 of Al_2O_3 , CaO , K_2O , Na_2O , total Fe as FeO, P_2O_5 , TiO_2 , molecular $\text{Al}_2\text{O}_3/\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O}$ and trace elements Rb, Sr, Zr and total REE. Of these species only Al_2O_3 exhibits a well-defined linear array, showing negative correlation with SiO_2 . CaO , FeO (total), TiO_2 and P_2O_5 show no significant change with increasing SiO_2 .

Data for K_2O , Na_2O , TiO_2 and molecular $\text{Al}_2\text{O}_3/\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O}$ can be divided into several overlapping fields, for GLB-1, -2, -3, and a composite field for the later, more fractionated plutonic units (GLB - 4, -5, -6 and -7). These four fields may individually express either flat, positive, negative or no correlation with SiO_2 .

Trace elements Sr, Rb, Zr and total REE display similar lack of overall trend against SiO_2 (Figure 41). The plutonic units again occupy relatively distinct, albeit overlapping, data fields for Sr, Rb and Zr.

Regular decrease of CaO , Al_2O_3 and Na_2O with increasing SiO_2 is conventionally regarded as resulting from fractional crystallization dominated by plagioclase (Barker and Arth 1976). The complex

Figure 39. Harker Variation Diagrams for Al_2O_3 , CaO , K_2O and Na_2O in Ghost Lake Batholith, Zealand Stock and Related Barren Exocontact Dykes.

Figure 40. Harker Variation Diagrams for FeO (Total), P₂O₅, TiO₂ and Molecular Al₂O₃/CaO+K₂O+Na₂O for Ghost Lake Batholith, Zealand Stock and Related Barrer Exocontact Dykes.

Figure 41. Harker Variation Diagrams for Sr, Rb, Zr and Total REE in Ghost Lake Batholith, Zealand Stock and Related Barren Exocontact Dykes.

interelement variation in the Ghost Lake Batholith is however suggestive of a complicated evolutionary history as only Al_2O_3 shows a well-defined negative correlation with increasing SiO_2 . This will be examined in more detail in the final chapter dealing with possible petrogenetic processes.

All GLB samples are peraluminous in the sense of Shand (1950) in that the molecular ratio $\text{Al}_2\text{O}_3/\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O}$ exceeds unity (Figure 42). More recently this expression has been called the "aluminum saturation index" by Zen (1986). The peraluminous nature of the GLB rocks is also reflected in the presence of substantial Barth mesonormative corundum contents (4 to 7 percent, Appendices 3 to 7) and absence of normative diopside which is in contrast to metaaluminous granitoid rocks (Chappell and White 1982).

Figure 42 depicts variation in molecular $\text{Al}_2\text{O}_3/\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O}$ against fractionation (measured by the K/Rb ratio) of internal plutonic rocks and barren exocontact pegmatites. The great majority of analyses fall within an elongate field which reveals a gradual decrease in molecular $\text{Al}_2\text{O}_3/\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O}$ with increasing fractionation, i.e. diminishing K/Rb ratio. This correlation suggests that a preferential concentration of Na_2O and K_2O relative to Al_2O_3 occurred in the most fractionated parts of GLB-3 and in the eastern lobe of the batholith (GLB-5, -6 and -7). If alkali loss to a volatile phase had been an important mechanism (Tuttle and Bowen, 1958, p.89-91; Norton 1970) in the origin of the late GLB units an opposite trend of increasing molecular $\text{Al}_2\text{O}_3/\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O}$ should be evident. This is only suggested for the barren exocontact pegmatites,

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MicroD

Figure 42. Molecular $\text{Al}_2\text{O}_3 / \text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}$ vs. K/Rb for the Gullwing Lake Batholith, Zealand Stock and Various Internal Units of the Ghost Lake Batholith.

which are among the most peraluminous rocks in the area. These rocks are mainly concentrated in the field outlined in Figure 42.

Variation of $Al_2O_3/CaO+Na_2O+K_2O$ in magnetite-biotite granites of the adjacent Gullwing Lake Batholith is significantly different than the GLB. These potassic granitoid suite rocks, which constitute the only other pegmatite-generating pluton in the study area, are metaluminous to weakly peraluminous (Figure 42) and exhibit a limited degree of chemical fractionation relative to most units of the GLB.

The leucocratic character of the Ghost Lake Batholith is reflected chemically by very low contents of FeO, Fe_2O_3 , MgO, and TiO_2 . Total iron contents (as FeO) rarely exceed one percent. MgO exhibits similarly low abundances, having a maximum value of 0.72 percent and rarely exceeding 0.30 percent. The low content of mafic components is readily apparent in the AFM diagram of Figure 43, where most analyses tightly cluster near the total alkalis apex.

The most consistent petrochemical variation is exhibited by Na_2O and K_2O , which in Figures 43 and 48 exhibit a well-defined inverse relationship over a range of ratios between 0.1 to 2.4. Most plutonic units plot within discrete, partly overlapping fields suggesting the fractionation sequence GLB-1-2-(3,4)-(5,6), which correlates with a progressively decreasing average K/Rb ratio. The difference between GLB-3 and -4 and GLB-5 and -6 is not obvious in Figure 43; the group involving GLB 3 and 4 has a lower Na_2O/K_2O ratio than GLB-5 and -6, and thus is less fractionated as supported by relatively higher K/Rb ratios (Appendix 7). GLB-1 contains a maximum K_2O content of 8.20

Figure 43. AFM Variation for the Ghost Lake Batholith, Zealand Stock and Peraluminous Granitoid Rocks of the English River Subprovince.

percent, whereas the highest Na_2O level of 5.61 percent occurs in GLB-5. The Zealand Stock has a slightly higher maximum Na_2O content of 6.15 percent. Its data field broadly overlaps with GLB-3, -5 and -6.

Exocontact pegmatites emplaced in medium grade host rocks and which are barren of rare-element minerals exhibit a trend toward decreasing K_2O at essentially constant Na_2O (field in Figure 44), although these analyses overlap in part with the Ghost Lake Batholith sodium-potassium trend.

Calcium varies between 0.16 and 2.36 percent; approximately seventy percent of the samples contain one percent CaO and less, thus comparing to the average alkalic or low calcium granites in Table 31. Highest calcium contents tend to occur in GLB-2 which has a range of 1.00 to 2.36 percent. The lowest calcium levels are in units GLB-5, -6, -7, the Zealand Stock and exocontact pegmatites emplaced in medium grade rocks.

Major element evolution of the Ghost Lake Batholith is best revealed in the CaO - Na_2O - K_2O diagram (Figure 45). Variation of the study area rocks shows a significant departure from the calc-alkalic trend of Nockolds and Allan (1953) and the trondhjemite trend of Barker and Arth (1970). The distinctively low calcium content of this batholith results in a curvilinear fractionation path subparallel to the K_2O - Na_2O sideline, that has been named the "low calcium granite trend" in this study. This trend also differs greatly from those characterizing the potassic (Figure 6), sodic and gneissic granitoid suites of the English River Subprovince (Breaks and Bond in press).

Figure 44. Na_2O vs. K_2O Variation for Intrusive Units of the Ghost Lake Batholith, Zealand Stock and Related Barren Exocontact Dykes. Analyses in Weight Percent.

Figure 45. Variation of CaO, Na₂O and K₂O in Intrusive Units and Related Exocontact Pegmatites of Ghost Lake Batholith and Zealand Stock. Calc-alkaline Trend after Nockolds and Allan (1953). Trondhjemitic Trend from Barker and Arth (1976). Analyses in Weight Percent.

Figure 46. Barth Mesonormative Proportions of Quartz, Albite and Orthoclase for the Ghost Lake Batholith. Dashed Lines and Stars respectively Correspond to Cotectic Field Boundaries and Ternary Minima and Eutectics from Luth et al. (1964). Encircled Stars Represent H_2O -Saturated Minima at 1 kbar with Varying Levels of Fluorine in which $M1 = 1\% F$, $M2 = 2\% F$ and $M4 = 4\% F$ (after Manning 1981).

Figure 47. General Fields of Barth Mesonormative Quartz-Albite-Orthoclase for Plutonic Units of Ghost Lake Batholith and Zealand Stock. Extent of Chemical Fractionation is given by the Average K/Rb Ratio: GLB-1 (282), GLB-2 (232), GLB-3 (110), GLB-4 (180), GLB-5 (78), GLB-6 (94) and Zealand Stock (112).

Table 31: Chemical Analyses of Various Calc-alkaline and Low Calcium Granites from the Literature which are comparable to the Ghost Lake Batholith

	GLB-1	1	2	3	4	5	6	7	8	9	10
SiO ₂	73.77	72.08	73.86	74.22	74.1	76.75	72.53	74.04	74.66	74.4	73.9
Al ₂ O ₃	15.27	13.86	13.75	13.61	13.88	12.95	14.77	14.67	15.55	14.3	15.4
Fe ₂ O ₃	0.10	0.86	0.78	2.03*	1.66*	0.4	1.41*	0.84*	0.17	0.19	0.5
FeO	0.58	1.67	1.13			0.5			0.42	0.53	0.3
MgO	0.17	0.52	0.26	0.27	0.32	0.01	0.34	0.18	0.02	0.16	0.3
CaO	0.90	1.33	0.72	0.71	0.74	0.35	0.69	0.50	0.42	0.56	0.8
Na ₂ O	2.21	3.08	3.51	3.48	3.27	3.85	3.27	4.06	4.29	3.44	4.5
K ₂ O	5.85	5.46	5.13	5.06	4.91	4.4	4.89	4.50	3.08	4.72	3.5
TiO ₂	0.26	0.37	0.20	0.20	0.13	0.03	0.20	0.07	0.03	0.12	0.02
P ₂ O ₅	0.05	0.18	0.14		0.23		0.03	0.14	0.15	0.17	
MnO	0.09	0.06	0.05	0.04	0.06	0.02		0.02	0.07	0.02	0.04
CO ₂	0.17						0.02	0.07			
H ₂ O ⁺	0.28	0.53	0.47			0.4	1.47	0.79	0.66	0.68	
H ₂ O ⁻	0.07				0.03			0.04	0.06		
F						0.06		0.11		0.04	0.3
TOTAL	99.77	100.0	100.18	99.58	99.30	99.75	99.60	99.92	99.60	99.47	99.56

*Total iron as Fe₂O₃

- GLB Average muscovite-biotite granite, unit GLB-1 of Ghost Lake Batholith
- 1. (20 analyses, this study)
 1. Average calc-alkaline granite (Nockolds 1954)
 2. Average alkalic granite (Nockolds 1954)
 3. Average low calcium granite (Turekian and Wedepohl, 1961, p.188)
 4. Average muscovite-biotite granite from Halifax Pluton, Nova Scotia (Smith et al. 1982, p.307)
 5. Average biotite-muscovite granite from Blue Tier Batholith, Tasmania (Mount William Sheet in Groves and McCarthy 1978, p.15).
 6. Average of 1148 two-mica granites and leucogranites in France (Le Fort 1981, p.10557)
 7. Manaslu biotite-muscovite granite, central Nepal. Average of 36 analyses from core of pluton (Le Fort 1981, p.10557).
 8. Muscovite granite from Harney Peak, South Dakota (Huang and Wyllie 1981, p.10516).
 9. Prosperous lake granite, near Yellowknife, N.W.T. Unpublished analysis of the author.
 10. Muscovite-albite plumbitic apogranite (Beus 1982, p.412).

Barth mesonormative proportions of quartz, albite and orthoclase from the Ghost Lake Batholith, Zealand Stock and related exocontact, barren granitic pegmatites have been plotted in the haplc-granite system

$\text{SiO}_2\text{-NaAlSi}_3\text{O}_8\text{-KAlSi}_3\text{O}_8\text{-H}_2\text{O}$ in Figure 46. The $\text{Na}_2\text{O-K}_2\text{O}$ compositional variation, described previously, is reflected in this diagram by a considerable scattering of data points, trending from near the middle of the quartz-orthoclase sideline towards the albite apex.

Most of the data fall within several overlapping fields expressive of a fractionation trend in the direction of increasing $\text{P}_{\text{H}_2\text{O}}$. This is revealed by an overall trend of increasing albite, decreasing orthoclase and slightly decreasing quartz (Figure 47). This trend is in the direction of generally increasing chemical fractionation as indicated by decreasing K/Rb and increasing $\text{Na}_2\text{O/K}_2\text{O}$ ratios.

COMPARISON OF $\text{Na}_2\text{O-K}_2\text{O}$ VARIATION OF THE GHOST LAKE BATHOLITH WITH OTHER SUITES IN THE ENGLISH RIVER AND WINNIPEG RIVER SUBPROVINCES

Figure 48 depicts generalized fields of $\text{Na}_2\text{O-K}_2\text{O}$ variation for important granitoid suites of the ERS and WRS, based upon all analyses in Breaks and Bond (in press). Although several plutonic units of the GLB bear a similarity in major element composition with the K-feldspar megacrystic variant of the potassic plutonic suite, the trend of $\text{K}_2\text{O-Na}_2\text{O}$ variation is quite different. The GLB exhibits a steeper negative slope for $\text{Na}_2\text{O-K}_2\text{O}$ variation compared to that of both variants of the potassic plutonic suite. The sodic plutonic suite lies outside the GLB trend towards high Na_2O , mainly a function of lower K_2O contents characteristic of this suite.

The characteristic wide range of $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios in the GLB also reveals a distinct departure from S-type granitoids of the Lachlan Fold Belt (LFB) of Australia. Chappell and White (1982, p.92) who indicated that low levels of Na_2O , CaO and Sr are diagnostic of peraluminous granitoid rocks there. It seems apparent, however, that Archean S-type granitoid rocks, at least in the study area, have the capacity to fractionate toward high Na_2O , which moves the various compositional fields away from the LFB field in Figure 48. Only some of the less fractionated units (GLB-2 and a very small part of the GLB-1 field) compare with the LFB S-Type granitoids.

Regional and secular chemical variation in the S-type granitoids may play as yet an undefined role in the apparent K_2O and Na_2O differences between the GLB and the Paleozoic LFB granitoid rocks. This possibility has not received due attention in the literature as few examples of Archean S-Type granitoid rocks have been described. Chemical features identical to the GLB have however also been documented for granites in a very young orogenic belts. Le Fort (1981) has described similar $\text{Na}_2\text{O}-\text{K}_2\text{O}$ and REE patterns in Eocene muscovite granite from the Manaslu Pluton of Nepal. This peraluminous pluton, which occurs within the Main Central Thrust, is likely an S-type granite generated during continental collision as the higher temperature Tibetan Slab was thrust over the Indian Plate (LeFort 1981).

Therefore, caution is recommended in utilizing the LFB as a "type locality" for S-type granitoids, particularly when dealing with Archean counterparts.

Figure 48. Fields of K₂O-Na₂O Variation in the Ghost Lake Batholith Units GLB-1, -2, -3, -5 and -6 in Comparison to Sodic and Potassic Suite Granitoid Rocks from the English River and Winnipeg River Subprovinces. Diagram is based upon 225 Analyses in this Study and in Breaks and Bond (in press).

TRACE ELEMENT VARIATION

Significant petrochemical variation is shown within the Ghost Lake Batholith by certain trace elements, in particular B, Ba, Li, Rb, Sr, Zr and REE. Trends involving enrichment of B, Be, Ga, Li, Nb and Rb and depletion of Sr, Zr, Ba and REE can be observed within the batholith as one progresses from the main northwestern mass of GLB-1 in Eton and Sanford Townships towards the narrow eastern lobe and, secondly, south and southwest towards the area of GLB-3 in Aubrey Township as revealed in the hand-contoured plots of Figures 49 and 50 (in pocket).

Boron, which is an important volatile element in the rare-element pegmatites of the Mavis Lake group, undergoes enrichment from 9-18 ppm in GLB-1 to a maximum of 2500 ppm in vicinity of the Dryden Airport. Trends involving fluorine (60 to 540 ppm) are not as pronounced. A maximum level of 560 ppm fluorine is recorded in the Zealand Stock. Very high fluorine contents (650 ppm to 1.8 percent) are however observed in tourmaline-rich masses at the southeastern batholith contact (see Contact Beryl Occurrence), and near this contact at the Taylor beryl pegmatite dyke 2 and the Petrunka scheelite property (Tables 39 and 45). In addition, minor fluorine enrichment (570-750 ppm) is notable in some exocontact pegmatites (devoid of rare-element minerals) emplaced in medium grade metasediments near Dryden (Appendix 8).

This enrichment suggests that no significant quantity of fluorine was retained by the highly fractionated GLB plutonic units exemplified at the Contact Beryl Occurrence. The tourmaline-rich contact and exocontact units which contain holmquistite and have the highest fluorine contents are interpreted to have formed from a B-F-Li enriched vapour phase (see Chapter 9).

Be, Ga and Nb exhibit more limited concentration ranges, however distinct enrichment trends toward the east and southwest are again discernible. Maximum values of Be (24 ppm) and Nb (55 ppm) are recorded in GLB-3 (Appendix 6). In contrast, levels of Be and Nb in the main mass of GLB-1 are typically 1 ppm and less (Figure 49).

Lithium, which ranges from <3-30 ppm in GLB-1 is generally enriched in GLB-3 and in the eastern lobe of the Ghost Lake Batholith where concentrations locally exceed 100 ppm. A maximum value of 198 ppm was recorded near the Dryden Airport (Appendix 7). Even more anomalous values may occur in exocontact pegmatites, as exemplified by 225 ppm in specimen F22-1 near Dryden (Appendix 8). According to Taylor (1965), lithium contents in granites which exceed 100 ppm indicate extreme fractionation.

Tin exhibits a slight tendency for increased concentration in GLB-3 and GLB-5, -6 and -7 relative to GLB-1, which commonly contains <0.8 ppm and a maximum level of 2.1 ppm Sn. GLB-3 shows a higher range of 3.6 - 10.2 ppm. GLB 5, 6, and 7 in vicinity of the Dryden Airport exhibit levels of 2.7 - 13.8 ppm. Maximum Sn levels are recorded in exocontact pegmatite dykes close to the southern batholith contact (32.7 - 55 ppm, Figure 49). The Ghost Lake Batholith can thus

locally be classed as a stanniferous granite (i.e. one with >10 ppm Sn, in the sense of Flinter 1971).

Ghost Lake Batholith intrusive units yield strong, approximately linear trace element continua in Figures 51 and 52. Ba, Rb and Sr exhibit well-defined inverse concentration trends reflected by rapidly decreasing Ba and Sr as a function of moderately increasing Rb. These trends compare favourably with models of fractional crystallization of granitic melts, developed by McCarthy and Hasty (1976), based upon the relationship of cumulus and intercumulus phases.

Fractionation trends in the Ghost Lake Batholith closely parallel the composite field of the Main granite (coarse-grained variety) -Bobbejaankop granite- Lease granite of the Bushveld Igneous Complex (McCarthy and Hasty 1976, p.1357) and the Blue Tier Batholith of Tasmania (Groves and McCarthy 1978, p.16), both of which contain syngenetic cassiterite mineralization which was concentrated in late stage units. The extremely low Sr and Ba contents (10 ppm and less) that typify rocks formed from crystallization of residual liquids in these two tin-bearing complexes are also found in some rocks of the Ghost Lake Batholith exposed in the Dryden Airport area (Appendix 7).

Total rare earth elements exhibit no systematic variation with SiO₂ (Figure 41). This lack of correlation contrasts with REE behavior in trondhjemite-tonalite-granodiorite suites, where a clear decrease in total REE occurs with increasing SiO₂ (Arth et al. 1978, Martin 1987). However, if fractionation of the GLB is measured by the K/Rb ratio, a rapid decrease in total REE (273 ppm to 6.2 ppm) correlates with increasing fractionation (Figure 53). Behavior of REE

Figure 51. Petrochemical Variation (ppm) of Rb and Ba for Ghost Lake Batholith, Related Barren Exocontact Pegmatites, Zealand Stock and Rare-Element Pegmatites. Comparative Trends from Blue Tier Batholith of Tasmania (Groves and McCarthy 1978) and Main Granite of Bushveldt Igneous Complex (McCarthy and Hasty 1976).

Figure 52. Petrochemical Variation (ppm) of Rb and Sr for Ghost Lake Batholith, Related Barren Exocontact Pegmatites, Zealand Stock and Rare-Element Pegmatites. Comparative Trends from Blue Tier Batholith (Tasmania) and Main Granite of Bushveldt Igneous Complex.

in the Ghost Lake Batholith is complex and will be described in more detail (Chapter 7) in comparison to rare-element pegmatites.

Most element ratios indicate advanced fractionation within two small areas of the Ghost Lake Batholith (Figure 50). These areas occur as narrow, strip-like domains which closely parallel the southern contact of the batholith near the Dryden Airport in Zealand Township (including the Zealand Stock) and near the southwestern terminus in Aubrey Township. The former is the more chemically fractionated as indicated by the lowest K/Rb, Ba/Rb and Al/Ga and highest K/Ba, Ca/Sr, Rb/Sr and Na/K ratios developed in the Ghost Lake Batholith.

K-Rb variation in peraluminous granitoid rocks of the Dryden area is illustrated in Figure 54. This group displays a trend of decreasing K/Rb (387 to 70) with decreasing K. This is antipathetic to normal upper crustal trends of Shaw (1968), who documented a slight increase in K/Rb with decreasing K. The GLB and related rocks are thus distinguished by an unusual departure from the chemical coherence normally exhibited by K and Rb.

Th/U ratios in the GLB show no apparent zonal pattern although it should be noted that many values significantly exceed the Canadian Shield averages of 4 and 6.2 respectively of Shaw et al. (1976) and Fahrig and Eade (1968). Significant enrichment of U was noted only in rare instances, as at sample localities D18-1A (280 ppm), E17-4A (30 ppm) and E19-2 (28 ppm).

The average Th/U ratio in the GLB (7.2) is quite similar to that in the nearby Mavis Lake pegmatite group (8.2). These averages are

Figure 53. Variation of Total REE Contents of Ghost Lake Batholith Plutonic Units as a Function of Increasing Fractionation (K/Rb ratio).

Figure 54. K-Rb Variation for Ghost Lake Batholith and Related Rocks.

higher than those of the potassic granitoid suite plutons (range of mean values 0.5 to 6.4 in Breaks and Bond, in press) which are the only other plausible progenitor for granitic pegmatites. The high Th/U ratios of the GLB thus constitute an additional unique chemical attribute which suggests linkage of this granitoid mass at least with rare-element pegmatites of the Mavis Lake group.

PETROCHEMISTRY OF FINE-GRAINED LEUCOGRANITE

Chemical comparison between specific pegmatitic granite units from different pegmatite fields has not been commonly undertaken due to a lack of data and to possible confusion caused by the variation in unit terminology, prior to the classification of pegmatitic granites by Cerny et al. (1981). The largest body of chemical data that presently exist is for fine-grained leucogranites from the Winnipeg River pegmatite district (Cerny et al. 1981, Goad and Cerny 1981); this in part, relates to ease in sampling relative to the associated, coarse-grained pegmatitic leucogranite and potassic pegmatite units.

Because the trace element characteristics of the fine-grained leucogranite may have genetic and mineral exploration significance (Cerny et al. 1981) it is worthwhile to compare and contrast such rocks of the GLB with those of the Winnipeg River Pegmatite District.

Cerny et al. (1981, p.144-146) and Goad and Cerny (1981) indicated a relatively primitive chemistry for fine-grained leucogranite, in comparison to other internal units of pegmatitic granites in the Winnipeg River pegmatite district.

Bulk compositions of fine-grained leucogranite from the study

area are comparable to those from the Greer Lake and Tin Lake pegmatitic granites of Goad and Cerny (1981, p.182). There is, however, a tendency for Na_2O to exceed K_2O in fine-grained leucogranite from the Dryden Field. This is also borne out by 23 analyses of Cerny et al. (1981, p.69) which, when plotted in the haplogranite system, scatter around $\text{Ab}_{36} \text{Or}_{32} \text{Qtz}_{35}$. Nevertheless, there is a significant overlap in the compositional fields from the two pegmatite areas.

Other significant differences include higher boron (0.65 to 0.81 percent B_2O_3) and lower H_2O^+ (0.10 to 0.14 percent) in fine grained leucogranite from the Ghost Lake Batholith. These relate to the presence of tourmaline (2.8 to 4.5 percent) and a paucity of muscovite (0.4 to 1.0 percent), which contrast with the fine-grained leucogranite from the Winnipeg River Pegmatite District.

Trace elements also reveal some differences in degree of fractionation of fine-grained leucogranite from the Winnipeg River Pegmatite District relative to that in the Dryden Pegmatite Field (Table 32). Abundances of Ga, Li, Sn, Sr, U and Zr are higher in Winnipeg River district fine-grained leucogranites whereas Ba and Th are generally lower. An overlap in average Rb levels is obvious (Table 32). Barium exhibits a pronounced depletion in the study area counterparts that coupled with intermediate Rb levels results in very low Ba/Rb ratios of 0.03-0.07. These values are slightly lower than for most fine-grained leucogranite from the Winnipeg River district

Table 32.

Average and Range of Selected Trace Elements and Element Ratios for Fine-Grained Leucogranite from the Ghost Lake Batholith Compared to those from Pegmatitic Granites of the Winnipeg River Pegmatite District (from Cerny et al. 1981). Analyses in ppm.

	<u>Ghost Lake Batholith</u>	<u>Lin Lake</u>	<u>Osis Lake</u>	<u>Greer Lake</u>	<u>Eaglenest Lake</u>
Ba	17 (10-20)	217 (31-416)	8 (3-18)	112 (12-150)	70 (0-110)
Be	4 (3-6)	2.0 (1-3)	0.8 (0.3-1.2)	3.8 (1.2 - 10)	5.6 (0.7-13)
Ga	28 (23-33)	30 (22-46)	48 (41-57)	53 (32-73)	58 (56-60)
Li	32 (16-52)	31.8 (11-59)	63.3 (47-74)	58 (29-86)	79.7 (27-140)
Rb	323 (270-390)	245 (128-338)	137 (103-193)	562 (340-1050)	536 (293-817)
Sn	3.9 (2.7-4.9)	8.7 (0-13)	13 (10-15)	13.6 (5-25)	13 (11 - 15)
Sr	12 (12-13)	35.4 (3-55)	25.7 (13-36)	18.2 (11-28)	10.3 (5-15)
Th	17 (10-20)	10.3 (0-30)	6.7 (4-12)	7.4 (0-25)	7 (0-18)
U	3.7 (1-7)	28 (13-40)	23 (8-52)	15 (8-30)	6 (4-8)
Zr	37 (30-40)	103 (61-140)	38 (17-74)	44 (11-58)	53 (38-71)
ΣREE	25.7 (21.7-29.6)	92.8		40.9	
Al/Ga	2883 (2380-3460)	2351	1674	1526	1258
Ba/Rb	.05 (.03 - .07)	0.89	.06	0.20	0.13
K/Ba	1712 (917-2805)	279	5402	767	338
K/Rb	78 (70-91)	176	132	87	75
Rb/Sr	26 (22-32)	10.9	6	32.1	52.8
Th/U	6.5 (2.9 - 10)	0.37	0.29	2.0	1.2
Associated Rare-Element Mineralization	Li, Be, Rb Ta, Nb, (Sn)	Be, (Li, Nb) Ta	Li, Rb, (Cs), Be, Sn, Ta Nb	Be, Nb > Ta, (Li)	Be, Nb > Ta, (Li)

(Ba/Rb=0.06 to 0.89 in Table 32); of the fine-grained leucogranites of this district that form the Greer Lake stock (Cerny et al. 1981) is most similar to that from the GLB.

FUSION AND CRYSTALLIZATION CONDITIONS FOR THE GHOST LAKE BATHOLITH MAGMA

Pressure and temperature conditions for fusion of the progenitor and subsequent crystallization of the Ghost Lake Batholith have been estimated from experimentally derived phase relationships in the haplogranite system and selected reaction curves involving critical minerals such as cordierite and muscovite.

Fusion Conditions

Anderson and Cullers (1978) have adapted the haplogranite system to estimate pressure conditions under which fusion and crystallization developed in comagmatic plutonic units of the mid-Proterozoic Wolf River Batholith, north Wisconsin. Application of the Anderson-Cullers Ab/An-pressure grid to the Ghost Lake Batholith is given in Figure 55.

Pressure conditions for fusion were roughly estimated by using rocks interpreted to represent the most chemically primitive compositions selected on the basis of highest K/Rb ratios (i.e. greater than 300) and at the suspected deepest level of the GLB, i.e. adjacent to the western contact. Ten analyses utilized here are of biotite granite from unit GLB-1 (D14-6b, D18-1b, D18-2a, D20-1b, D22-1a, E14-1a, E17-4a, E17-8b, E25-4 and E25-7). These rocks form a cluster lying below the Ab/An ratio line of 2.9 in Figure 55. It must be cautioned that a disparity exists between this normative Ab/An ratio from this diagram and the average CIPW Ab/An ratio of 4.4 for

Figure 55. Anderson-Cullers Normative Ab/An vs. Total Pressure Grid for the Estimation of Fusion Pressures for GLB-1.

these rocks. The reason for this difference is not understood; essentially similar pressure values would be registered if the data field was centred about the CIPW Ab/An ratio of 4.4 in Figure 55. This diagram suggests fusion conditions ranging between 2 and 5 kbar, corresponding to a crustal depth of 7 - 18 km.

Intersection of the interpreted P-T metamorphic gradient for the study area with the muscovite and biotite dehydration curves in Figure 56 can theoretically produce, by fluid-absent partial melting of metapelites and metawackes, peraluminous granitic melts at crustal depths of 12 to 15 km. Fluid-absent anatexis was first described by Burnham (1967, p.68) and is characterized by the derivation of H₂O required for melting entirely from dehydration of protolith minerals muscovite, biotite and hornblende. In contrast, fluid-present anatexis involves the introduction of H₂O from a source external to that in the protolith (e.g. Wickman 1987) as potentially guided by fractures or shear zones. The lower depth estimate is theoretically derived from the intersection of the biotite dehydration curve (Figure 56, curve 4) with the metamorphic P-T gradient extrapolated to granulite grade for the Dryden area. This reaction possibly occurred at depth below the sillimanite-K-feldspar-cordierite-garnet zone of the study area; granulite conditions necessary for biotite dehydration are not apparent in metasediments of the Dryden area. Nevertheless the above depth interval broadly compares with the 7 to 18 km range derived from the Anderson-Cullers grid in Figure 55.

Presence of magmatic cordierite in GLB-1 also suggests a maximum depth of 17 km (Figure 56), since the experimentally determined blank

Figure 56. Relationship of Metamorphic P-T Gradient for the Eagle River Area and Stability Fields for Cordierite- and Primary Muscovite-Bearing Granitic Melts. P-T Curves Compiled from: 1 = Holdaway (1971), 2 = Huang and Wyllie (1981), 3 = Kerrick (1972), 4 = Le Breton and Thompson (1988).

cordierite stability field in peraluminous granites by Clemens and Wall (1981) is confined to relatively low pressures. Garnet is virtually absent from most of GLB-1, which according to the experimental melting of natural pelites by Green (1977) suggests magma generation depths less than 25 km. In contrast, garnet is a widespread accessory mineral in the petrochemically advanced plutonic units (GLB-4, -5, -6, and -7) and its stability in these is possibly a function of increasing manganese contents in residual melts (see Chapter 8). Green (1977) suggested that garnets with 20-25 mole percent of the spessartine component can be stable at pressures below 3 kbars.

Crystallization Conditions

The H₂O-saturated solidus curve for the muscovite granite from the 1.7 Ga Harney Peak granite of South Dakota (Huang and Wyllie 1973, curve 2) is particularly appropriate to the estimation of crystallization conditions for the more chemically fractionated units of the Ghost Lake Batholith (GLB -4, -5, -6, and -7) which are interpreted to lie at higher crustal levels than GLB-1 from which the conditions of fusion were estimated. This peraluminous granitic pluton bears a parental relationship with spodumene and beryl pegmatites of the Southern Black Hills Pegmatite Field (Orville 1960, Norton 1975). Presence of primary muscovite in GLB -5, -6, and -7 suggests a restriction of crystallization conditions (within the sillimanite field) between 3.5 and 9 kbar at solidus temperatures of 650° to 700°C (Figure 56).

The shift of normative albite-quartz-orthoclase ratios towards the albite apex in Figure 47 may be in part dependant upon the progressive concentration of fluxing agents (F, B, Li and P) in residual melts that became increasingly sodic with fractionation (Hudson and Arth 1983, Pichavant and Manning 1984). The important works of Manning (1982), Manning and Pichavant (1983, 1984) and Pichavant (1981, 1987) in this respect have particular relevance to the GLB and pegmatitic granites composing most of the Gullwing Lake, RT and Mica Point dykes. The high concentrations of boron in the late plutonic units GLB -5, -6 and -7 and in contact metasomatic units have previously been noted in this chapter. Significant concentration of fluorine is also implied by its substantial enrichment in contact metasomatic units (Contact Beryl Occurrence, Chapter 4) and by fluorine levels in muscovite from the GLB itself (300 to 1900 ppm F in Appendices 29, 30 and 31).

Chemical and mineralogical data given in Chapters 2, 3, and 7 suggest that B and F was an important volatile agent in the late plutonic units of the GLB, exocontact dykes and local greisen zones. The effect of fluorine on the crystallization of these plutonic units is less certain as little of this element is contained in the GLB (Maximum 540 ppm) in GLB-6, Appendix 7). Fluorine has a high affinity for the melt phase ($K_D=5$ to 10, Hards 1976), however, it may not have been concentrated in low calcium granitic rocks as the formation of important mineral hosts such as fluorite would expectedly be inhibited, due to the low concentrations of calcium. Therefore, it is possible that fluorine was mobilized into a fluid phase which may account for its higher concentration (Table 45, 650 ppm to 1.78 percent F) in metasomatic selvages such as at the Contact Beryl Occurrence.

CHEMISTRY OF INTERNAL UNITS OF RARE-ELEMENT PEGMATITES

Analytical data for primary pegmatite zones and albite-quartz-muscovite replacement units are presented in Tables 33, 34, 36 and 37. Comparative major element bulk compositional data for spodumene- and petalite-bearing pegmatites from the literature have been compiled in Table 35.

PRIMARY PEGMATITE UNITS

Relative to the Ghost Lake Batholith, rare-element pegmatite bulk and zonal compositions exhibit advanced chemical fractionation. This is revealed by generally high SiO_2 contents (up to 82 percent), high normative corundum and very low Ca, Mg and total Fe. More revealing, however, are interelement ratios such as K/Ba , K/Cs , K/Rb , Ba/Rb , Rb/Sr and Al/Ga . These express profound fractionation compared to other granitoid rock suites of the English River Subprovince (Breaks et al. 1978, Breaks and Bond, in press).

Mavis Lake Pegmatite Group

The petrochemical data base for this pegmatite group is relatively complete (Tables 33 and 34). It can be seen that the Taylor beryl pegmatites are more poorly fractionated than spodumene-beryl-tantalite pegmatites from the Fairservice Property and comparable to or slightly more fractionated than the late plutonic units of the Ghost Lake Batholith (GLB -4, -5, -6, and -7). This is best revealed in the Ba-Rb and Rb-Sr diagrams (see below: Figures 62 and 63).

The Taylor beryl pegmatite No.2 contains higher Mg, Ca, Ti, and total Fe than other rare-element pegmatites of the Dryden Field.

Enrichment in these elements is in part evident in high tourmaline contents (up to 80 percent, Table 9); such elements may have entered into the tourmaline structure during late stage metasomatic replacement in which the ultramafic host rocks underwent chemical exchange with a pegmatite fluid which resulted in a K₂O-rich selvage along the contact (see Table 39, analyses C, D and E). This is suggested by the pronounced depletion in the contact metasomatic selvage of appreciable Mg, Ca, Ti, and Fe relative to their concentrations found in the less altered ultramafic host rock (Table 39).

The majority of lithium-rich pegmatites described in the literature have Na₂O content that exceeds K₂O and Na₂O/K₂O ratios between 1 and 5 (Figure 57). Zonal and bulk compositions of the Mavis Lake Group, however, deviate slightly to appreciably from this range. Spodumene pegmatites exhibit Na₂O/K₂O ratios less than unity, whereas the Taylor beryl pegmatites are highly sodic with Na₂O/K₂O greater than 5 and closely compare to the San Luis spodumene pegmatite in Argentina (Herrera, 1968).

The Na₂O/K₂O ratios of the rare-element pegmatites are similar to those of the high sodic part of the Ghost Lake Batholith trend (Figure 57), but Na₂O+K₂O contents in the pegmatites are appreciably lower than in the GLB. The subordinate trend for barren, exocontact pegmatites related to the batholith (Figure 57) shows a closer similarity to many lithium-rich pegmatites. The closest bulk composition to the GLB trend in Figure 57 is the Gullwing Lake two-mica pegmatitic granite which plots at the high Na+K end of the Li vs

Figure 57. Na₂O vs. K₂O Variation of Bulk Compositions and Zones of Rare-Element Pegmatites from the Dryden Field and Those Compiled from the Literature. Analyses (weight percent) are from Tables 33, 34 and 35.

Na+K curve for rare-element pegmatites (Figure 58). If zonal compositions are excluded this diagram reveals a well defined inverse correlation between Li and Na + K.

Generally high normative corundum levels (0.8 to 11.7 percent) in the Dryden pegmatites overlap with the Ghost Lake Batholith range of 1.3 to 7.5 percent. Highest normative corundum contents occur in the tourmaline-rich pegmatite from the Taylor beryl dyke No.2 (11.7 percent) and the albite-spodumene-quartz zone from the Fairservice Pegmatite No.1 (9.3 percent). The mild to strongly peraluminous character of most pegmatite units is also revealed by molecular $Al_2O_3/CaO+Na_2O+K_2O+Li_2O$ ratios which range from 1.04 to 2.39 (Table 33).

Chemical variation within primary zones of spodumene pegmatites of the Mavis Lake Group was able to be evaluated only at Pegmatite No.1 on the Fairservice Property. Two spatially related zones situated peripheral to the quartz-rich ore zone were analyzed (Table 33):

- 1) beryl-K-feldspar-muscovite-albite-quartz-spodumene pegmatite (analyses 169 and 171)
- 2) beryl-muscovite-spodumene-albite-quartz-K-feldspar (analysis 170)

Differences between these two zones are minor (compare analysis 169 with 171 in Table 33), but higher SiO_2 and LiO_2 and lower Al/Ga, K/Cs, K/Rb, Rb/Sr, and molecular $Al_2O_3/CaO+Na_2O+K_2O+Li_2O$ (in sample 169) suggest that the albite-spodumene-quartz zone is the more

Figure 58. Li vs. Na+K Variation, in Atomic Percent, for Bulk and Zonal Compositions of Rare-Element Pegmatites from Dryden Field and Literature. Analyses are from Tables 33, 34 and 35.

fractionated. However, higher Rb/Sr, K/Ba, and lower Ba/Rb (in sample 170) infer a greater degree of fractionation for the K-feldspar-quartz-spodumene zone. This unit is also exceptional in containing the highest K₂O (5.87 percent in Table 33) of the analyzed pegmatite samples (Figures 57 and 59).

Most studies of zoned pegmatites indicate from field relationships that the quartz-rich units are invariably the last to crystallize (Cameron et al. 1949). This is also considered to be the case for the albite-spodumene-quartz core zone at Pegmatite 1, by virtue of its much higher quartz content (80 percent) and gradational relationship with adjacent spodumene-rich pegmatite.

Aplite in the Fairservice spodumene pegmatites 1, 2, and 9 was inferred in Chapter 4 to be primary. An analysis of a 1.24 m channel sample across such a unit from Pegmatite 2 is given in Table 33 (sample 173). Comparison to the composition of the replacement stage aplite from the Fairservice spodumene pegmatites, given in Table 36, reveals a general major element similarity, although most replacement aplites have lower Na₂O and higher K₂O than the primary aplite.

Among the trace elements Rb is the most distinctive, being depleted in the primary aplite (190 ppm) relative to the great majority of replacement aplite bodies (190-3700 ppm) in Table 36. Higher levels of Li, Cs, Ga, Be and Sn also tend to occur in the replacement aplites. The greater degree of chemical fractionation of these aplites is generally apparent from the lower Ba/Rb and K/Rb and higher Rb/Sr and K/Ba.

Gullwing-Tot Lakes Pegmatite Group

Evaluation of petrochemical trends in the Tot-Gullwing Lakes pegmatite group is more difficult as these pegmatites are more complexly zoned and only four analyses of primary units were performed, three samples of which come from the Tot Lake pegmatite.

Bulk composition of the muscovite-biotite-albite-K-feldspar potassic pegmatite from the Gullwing Lake dyke is given in Table 34. Most of the pegmatite consists of this unit and it conceivably represents the composition from which the spatially associated Central and South Zone core assemblages were derived. This strongly peraluminous unit is enriched in F, Rb, Li, Cs, Be and Sn in comparison to late plutonic units of the Ghost Lake Batholith. Advanced chemical fractionation is also evident in low Ba, Sr, Zn and total REE and in most interelement ratios (Table 34).

On the basis of chemistry and accessory mineralogy the Gullwing Lake biotite-muscovite-albite pegmatitic granite shows similarity to an "apogranite". According to Beus et al. (1982, p.410) an apogranite is a "strongly albitized granite (35-70 percent albite) containing abnormally high values of one or more rare-elements such as Be, Nb, Ta, Li, Zr, and REE related to the presence of specific rare-element minerals which typically occur in the apical parts of granite intrusions forming separate cupolas, apophyses, dykes and metasomatic bodies along the contacts and tectonic blocks within the upper part of the intrusive body". A close major element similarity between the Gullwing Lake pegmatitic granite and the muscovite-albite plumbitic apogranite type of Beus (1982) is evident (Tables 34 and 31). This is

Table 34. Petrochemistry, Modes, CIPW Norms and Element Ratios of Primary Units from the Gullwing Lake and Tot Lake Rare-Element Pegmatites.

	GULLWING PEGMATITE			TOT LAKE PEGMATITE		
	250	62	251	182-12	182-14	251
SiO ₂	72.90	63.20	69.90	63.20	72.90	69.90
Al ₂ O ₃	15.90	19.00	19.10	26.70	19.90	19.10
Fe ₂ O ₃	2.11		0.0	0.00	0.00	0.0
FeO		0.24	0.16			0.16
MgO	0.47	0.0	0.0	0.08	0.13	0.0
CaO	0.75	0.39	0.27	0.24	0.34	0.27
Na ₂ O	5.01	2.24	6.01	1.92	3.95	6.01
K ₂ O	1.97	1.24	3.06	0.43	1.24	3.06
TiO ₂	0.12	0.14	0.31			0.31
P ₂ O ₅	0.01	0.11	0.09			0.09
S	0.02	0.01	0.01			0.01
MnO	0.10	0.11	0.04			0.04
CO ₂	0.14	0.19	0.11			0.11
H ₂ O ⁺		0.85	0.18			0.18
H ₂ O ⁻		0.13	0.07			0.07
Li ₂ O	0.31	1.68	0.25	5.27	1.81	0.25
Cs ₂ O		6.06				
Rb ₂ O						
F	0.12		0.50			0.50
TOTAL	99.93	95.57	100.31	97.84	98.46	100.31

B	<5	60	25
Ba	50	70	20
Be	8	9	21
Cs	90		610
F	1150	520	80
Ga	36	56	54
Li	1450	7800	1180
Mb	65	7	75
Pb	850	<10	20
Rb	5	1920	4510
Sc	20	36	1
Sn	15	18	22.1
Sr	<30	40	55
Ta	10	30	270
Th	1	20	130
U	1	0.9	3
V	18	2	2
Y	9	1	2
Zn	43	36	36
Zr	35	1	19

Table 34, continued

	ELEMENT RATIOS					
	2341	1798	0.05	0.006	1875	
Al/Ga	0.06	0.04	4	5.2	0.004	
Ba/Rb	9.4	107	32	60	7.4	
Rb/Cs	57	147	119	1029	82	
Rb/Sr	372	0.2	22	30	1270	
K/Ba	182	5	6	6	42	
K/Cs	19	0.02	0.09	0.09	6	
K/Rb	1.9	30	6.5	2.8	0.6	
Mg/Li	10	0.6	3.9		0.27	
Th/U	2.3	0.18				
Nb/Ta	>2.2	1.364	1.235	1.351	1.455	
Al ₂ O ₃ /Na ₂ O·K ₂ O +CaO·Li ₂ O·Rb ₂ O (Molecular)	1.255					
NORMATIVE MINERALS						
Quartz	31.8	41.0	25.5	35.1	19.2	
Albite	44.2	22.3	15.8	34.2	53.4	
Anorthite	3.5	0.2	1.0	0.2	0.9	
Orthoclase	11.4	8.3	2.0	7.2	19.1	
Eucryptite	0.3	16.6	46.6	15.6	2.1	
Corundum	5.0	11.0	7.0	7.3	4.9	
Apatite		0.2			0.2	
Ilmenite					0.2	
Rutile	0.1				0.1	
Magnetite	2.4	0.4			0.2	
Ferrosillite						
Enstatite	1.4		2.0	0.4		
MODES						
Quartz		12	4			
Plagioclase		17	16			
K-feldspar		X				
Spodumene		27	78			
Pollucite		32				
Altered Polucite		5				
Green Mica		7	3			
White Mica		1				

X = percent in accessory quantity

Sample Description:

- 250 : Large bulk sample of muscovite-biotite-albite pegmatitic granodiorite adjacent to the South Zone, Gullwing Lake Pegmatite
- 62 : Large bulk sample of pollucite-spodumene pegmatite in Trench 2 from unit 5 of the Tot Lake Pegmatite
- TR2-12 : Spodumene-rich layer from unit 3 of the Tot Lake Pegmatite, containing the highest spodumene content in the Dryden Pegmatite Field
- TR2-14 : Aplite intercalated with spodumene-rich layers from unit 2 of the Tot Lake Pegmatite
- 25J : Matrix between large spodumene phenocrysts from transverse layered unit, trench 3 of Tot Lake pegmatite

reflected in the presence of accessory beryl and columbite, in addition to trace level enrichment in Li, Nb, Ta, Sn, Be, Cs and F. This apogranite type is distinguished by fluorine up to 0.3 percent, BeO up to 600 ppm, SnO₂ up to 800 ppm, Ta₂O₅ up to 100 ppm and a Nb₂O₅/Ta₂O₅ ratio between 0.2 and 2 (Beus, 1982, p.411). Lithium content of the Gullwing Lake pegmatitic granite (1150 ppm, Table 34 actually exceeds the 40 to 900 ppm range given by Beus (ibid) and suggests a transition into the lepidolite-amblygonite-albite apogranite type which contains Li up to 5000 ppm.

Tischendorf (1977) coined the term "metallogenetically specialized granite" based upon an extensive review of barren and mineralized granites chiefly of Phanerozoic age. These granites are mainly characterized by elevated levels of the elements Li, Rb, Be, Sn, W, Mo and F (Appendix 42). A modified form of this term, i.e. "specialized granite" will be used in the ensuing text in general reference to perlauminous pegmatitic granites will have chemical compositions similar to that in Appendix 42 (e.g. analysis 250 in Table 34). The term "fertile granite" as used by Cerny et al. (1988) has not, to the writer's knowledge, been chemically defined.

Four zone compositions from the Tot Lake pegmatite are given in Table 34. These analyses indicate the most extreme degree of chemical fractionation in the Dryden Field as testified by high cesium concentrations, which ultimately lead to the appearance of pollucite, and by various interelement ratios. The pollucite zone exhibits the lowest K/Rb, K/Cs, Ba/Rb, Rb/Cs, Al/Ga and highest Rb/Sr (Table 34). According to Cerny et al. (1985, p.388) "pollucite is the ultimate

expression of Cs accumulation in the very last residua of granite fractionation".

Bulk cesium content (Table 34) exceed one percent only in the pollucite zone (6.6 percent Cs_2O). Solodov (1966) estimated that a Cs_2O content of one percent is required to initiate pollucite crystallization from a given pegmatitic magma.

Extreme Li_2O contents (up to 5.27 percent) are also characteristic in certain zones of the Tot Lake pegmatite, and occur in spodumene-rich layers of unit 2 (Figure 59). This unit is interlayered with and gradational into spodumene-poor aplite (unit 3) suggesting a genetic relationship between the two units. High Li contents of unit 3 relative to replacement stage aplites (Tables 34 and 37) also make this aplite unique. Replacement stage aplite in the Tot Lake Pegmatite (Table 37, analysis 248) contains only 220 ppm Li, much higher Na_2O and lower K_2O and Al_2O_3 than the primary aplite of unit 3.

Profiles of trace element distribution and interelement ratios across the southeastern end of the Tot Lake Pegmatite are given in Figures 60 and 61 and are respectively located at Trench 2 and 3 in Figure 30. This petrochemical section, which crosses mostly primary pegmatite units, was constructed on the basis of partial analyses from Breaks and Bond (in press). The chemical trends suggest crystallization symmetrically inward from the dyke walls as exemplified by increase in Li, Cs, Rb and K/Ba, Rb/Sr and trough-like distribution patterns for Mg/Li and Ba/Rb.

Figure 59. Li-Na-K Variation (atomic Percent) in Primary and Replacement Pegmatite Units and Contact Metasomatic Selvedges from the Dryden Pegmatite Field.

Figure 60. Profiles of Ba, Be, Cs, Li, Rb, Sr, Th and U Variation (ppm) in Trench 2, Tot Lake Li-Cs-Rb Pegmatite.

Figure 61. Profiles of K/Ba, Rb/Sr, K/Cs, K/Rb, Mg/Li and Ba/Rb Variation in Trench 2, Tot Lake Li-Cs-Rb Pegmatite.

Rb-Sr and Ba-Sr Variation in Rare-Element Pegmatites

Rare-element pegmatites from the study area and elsewhere in the Superior Province of Ontario collectively define distinctive trends on Ba-Rb and Rb-Sr diagrams (Figures 62 and 63). These pegmatite trends lie at high angles to the strong linear trends, with negative slope, that characterize the Ghost Lake Batholith. McCarthy and Hasty (1976) modelled trends similar to that of the GLB, based upon fractional crystallization influenced by cumulus/intercumulus mineralogy. Rare-element pegmatites, in contrast, show a positive Rb-Sr correlation and no correlation between Ba-Rb. Both pegmatite data fields intersect the batholith trends only at the most fractionated end.

Crystallization Conditions of Lithium-Rich Pegmatites

Application of experimentally determined stability fields for peraluminous granites, pegmatites and critical Li-rich mineral reactions allows an estimate of P-T conditions for the crystallization of rare-element pegmatites of the study area.

Figure 64 gives normative ratios of eucryptite: albite: quartz for bulk and zonal compositions of rare-element pegmatites, and lithium-rich volcanic and subvolcanic analogues as compiled from the literature and from the Dryden area. These analyses have been taken from Tables 33, 34 and 35. The following literature analyses which were also used were not tabulated in this study: ongonite (Kovalenko 1977), Spor Mountain rhyolite (Webster et al. 1987), Harding spodumene pegmatite (Burnham and Jahns 1962) and Macusanite (London et al. 1988). Phase relations for the water saturated, synthetic lithium-aluminosilicate system of Stewart (1978) at a pressure of $p_{H_2O} = 2$

Figure 62. Ba-Rb Variation Diagram for Bulk and Zonal Compositions of Pegmatites from Dryden Field and other Superior Province Fields. Analyses (ppm) from Tables 33, 34 and 35.

Figure 63. Rb-Sr Variation Diagram for Bulk and Zonal Compositions from Dryden Field and other Superior Province Fields. Analyses (ppm) from Tables 33, 34 and 35.

kbar are presented in this diagram. Stewart's work defined a eutectic at 640°C for this system composed of 39 percent petalite, 34 percent albite and 27 percent quartz (equivalent to quartz:albite:spodumene of 42:34:24 and quartz:albite:eucryptite of 50:34:16). Spodumene- and petalite-bearing rocks show no distinctive groupings in Figure 64, and curiously none of the petalite pegmatites plot in the petalite-spodumene-albite sub-triangle.

If the experimental data on the lithium-rich haplogranite system has reasonable application to magmatic crystallization trends in natural pegmatitic systems, then an approximate upper limit of 32 percent spodumene (equivalent to 2.6 weight percent Li_2O) should be realized in the bulk compositions of such pegmatites (Stewart 1978). This is suggested by most bulk compositions in Figure 64 containing Li_2O contents below the 2.6 percent isopleth. According to this diagram Li_2O content (2.6%) exceeds that at the eutectic (1.90% Li_2O) due to the curved nature of the field boundary between albite and the lithium-aluminosilicates (Figure 64). The isopleth for the minimum Li_2O content required to initiate spodumene crystallization estimated at 1.07 percent by Solodov (1966) is also shown in Figure 64. In theory, fractional crystallization of spodumene involved in the various zones of a lithium-rich pegmatite should be marked by a progression of normative quartz:albite:eucryptite ratios towards a eutectic composition in the synthetic lithium aluminosilicate system. It is inferred from Figure 64 that fractional crystallization from a melt played a minor role in the evolution of many pegmatitic bulk compositions. Figure 64 indicates that only 4 rare-element pegmatite bulk compositions from the literature and this study plot near the

Figure 64. C.I.P.W. Normative Eucryptite-Albite-Quartz Variation in Lithium-Rich Pegmatites in the Synthetic Lithium Aluminosilicate System after Stewart (1978). Analyses from Tables 33, 34 and 35.

Table 35, continued

	NORMATIVE MINERALS													
	Quartz	Albite	Anorthite	Orthoclase	Eucryptite	Corundum	Apatite	Ilmenite	Magnetite	Ferrosilite	Enstatite	Calcite	Rutile	
	47.6	23.6	0.4	9.6	11.1	6.0	0.5	0.4	0.5	0.1				
	35.7	27.4		17.3	15.4	4.2								
	32.4	40.5		11.6	12.9	2.8								
	46.1	17.8	0.4	11.2	17.5	6.3	0.2	0.2						
	53.7	9.7		3.6	25.7	6.8	0.2	0.5						
	30.3	17.0		43.5	1.8	6.2								
	37.8	25.8		19.2	13.7	3.5								
	38.4	29.8		14.2	12.7	4.9								
	36.2	15.2		15.1	12.4	6.6								
	38.0	30.5		7.1	30.3				0.2					
	38.7	43.5		1.0					2.1					
	6.1			5					0.1					
	20.7			5										
	4.0													
	1.6													

SAMPLE DESCRIPTION

1. 9 m channel sample taken by F.W. Breaks across westernmost exposure of the MMW Pegmatite, Georgia Lake Pegmatite Field (Breaks 1980).
2. Muscovite-biotite-albite granite from MMW Stock and taken by F.W. Breaks one metre north of contact with MMW Pegmatite.
3. Hirvikallio petalite pegmatite, Finland. Average of 2 analyses in Vesasalo (1959).
4. Bikita petalite pegmatite, Zimbabwe (Gallager 1962).
5. Tanco petalite-polucite pegmatite, Manitoba (London et al. 1988, p.361).
6. Roadhouse River spodumene pegmatite, northwestern Ontario. Average of 18 analyses (Breaks et al., 1985, p.20).
7. 6.1 m channel sample taken by F.W. Breaks across the North Zone spodumene pegmatite of Nama Creek Mines Limited (Pye 1965, p.90-95), Georgia Lake Pegmatite Field.
8. 3.5 m channel sample taken by F.W. Breaks across the southernmost exposure of the Brink spodumene pegmatite (Pye 1965, p.62-64), Georgia Lake Pegmatite Field.
9. Calculated average composition of Georgia Lake spodumene pegmatites (Milne 1962, p.122).
10. King's Mountain spodumene pegmatite, North Carolina (Kessler 1961, p.9).
11. Peg Claims spodumene pegmatite, Maine (Sundelius 1963, p.102).
12. Leinster spodumene pegmatite, Ireland. Average of five analyses calculated from Luecke (1981, p.215).
13. San Luis spodumene pegmatite, Argentina (Herrera 1968, p.27)
14. North Lacorne spodumene pegmatite, Quebec (Derry 1950).

eutectic (dashed area in Figure 64). Some scatter is to be expected for the normative ratios rather than strong clusterings as is common for finer-grained granitic rocks plotted in the haplogranite system, due to the difficult task in obtaining representative bulk samples. Nevertheless, this scatter is inferred to reflect a predominance of processes other than fractional crystallization in the evolution of rare-element pegmatites.

Zonal and bulk compositions show an even greater normative variation in Figure 64, than that predicted by magmatic crystallization. These include the Leinster and Nama bulk compositions and spodumene-rich zones from the Tot Lake and Fairservice Pegmatite No.1.

Application of Figure 64 to zones of the Fairservice Pegmatite No.1 suggest that fractional crystallization may have played a role only in the formation of the K-feldspar-spodumene-albite zone (FAIRSERVICE 170). This unit contains low normative eucryptite and quartz relative to the adjacent quartz-spodumene-albite zone. Although the albite-quartz-spodumene zone (FAIRSERVICE 171 in Figure 64) lies on the cotectic line between the quartz and spodumene fields, it contains substantially higher normative quartz and eucryptite, than FAIRSERVICE 170, as reflected by the proximity of 171 to the 830°C quartz-eucryptite binary eutectic. Such high amounts of normative quartz (56.9 percent, Table 33) is according to Stewart (1978, p.977-978) suggestive of deposition from a silica-rich gas phase.

Water-saturated solidus and liquidus relations for the Harding spodumene pegmatite (Burnham and Jahns 1962; Vaughan 1963) are

juxtaposed with the silica-saturated stability fields for eucryptite, spodumene and petalite (London 1984) in Figure 65, to form a grid for estimating P-T conditions of crystallization for Li-rich pegmatites. In Figure 65, this suggests a crystallization interval of 640 to 610°C for primary spodumene in the study area, i.e. above the reaction petalite $\rightarrow$ spodumene + quartz since petalite does not occur in the Dryden Pegmatite Field. Entrapment conditions of fluid inclusions in spodumenes from other pegmatite fields indicate a wider crystallization interval of 500-650°C for this mineral at pressures of 3-5 kbars (London 1984, 1986).

Transformation of spodumene to either virgilite ($\text{LiAlSi}_5\text{O}_{12}$) or hexagonal β -spodumene constrains the upper temperature limit of spodumene-or petalite-bearing pegmatites since virgilite apparently occurs only in subvolcanic peraluminous magmas (French et al. 1978) and β -spodumene is not found in nature (Stewart 1978). P-T conditions for the formation of lithium-rich pegmatitic magma evidently did not surpass the virgilite stability curve; this is suggested by the position of the Harding Pegmatite solidus and liquidus curves (Figure 65). The bulk sample of the Harding Pegmatite is probably the most representative of any pegmatite body (Burnham and Nekvasil 1986). However, it should be noted that the bulk Li_2O content of the Harding Pegmatite (0.65 weight percent in Table 35) is approximately one half of that typical for homogeneous, unzoned Li-rich pegmatites (1.5 weight percent, Stewart 1978). Thus, the liquidus and solidus curves for the Harding Pegmatite will probably lie at higher temperatures than those for the Fairservice Pegmatite No.1, since increased lithium levels would be expected to shift these curves to slightly lower

Figure 65.

Petrogenetic Grid for Lithium Aluminosilicate-Bearing Granitic Rocks. H_2O -Saturated Liquidus (L) and Solidus (S) Curves for Harding Spodumene Pegmatite from Burnham and Jahn³ (1962) and Vaughan (1963). Lithium Aluminosilicate Stability Fields, Magmatic and Subsolidus P-T Crystallization Trends for Harding, Tanco and Bikita Pegmatites from London (1984). Muscovite Stability Curve after Chatterjee and Johannes (1974).

temperatures. Wyllie and Tuttle (1964) observed an 80°C decrease in the haplogranite solidus temperature upon addition of 2 weight percent Li_2O ; liquidus temperatures decreased 25°C and 50°C for 1 and 23 weight percent Li_2O , respectively (Martin and Henderson 1984).

Upper pressure limits for crystallization of spodumene pegmatites from the study area cannot be accurately constrained at present. If crystallization trends are similar to the Harding spodumene pegmatite, an approximate upper total pressure limit of about 5 kbars at 600°C would be expected based upon the calorimetric experimental work of London (1984, 1987) as noted in Figure 65.

The Tot Lake spodumene-pollucite pegmatite may represent the lowest pressure conditions for crystallization of any in the study area. As well, this pegmatite may represent a transitional type between Kuzmenko's (1976) magmatogenic pegmatite class 2 (primary spodumene-bearing) and class 4 (petalite+pollucite-bearing). Pollucite usually occurs with petalite and rather more uncommonly with primary spodumene (Cerny and Ferguson 1972) as at Tot Lake. In the latter association, pollucite is apparently restricted to the lowest pressure primary spodumene pegmatites (Melentyev 1961). In this context it would be very useful to have experimentally determined limits for pollucite stability; some initial work in this respect was undertaken by Henderson and Manning (1984), with inconclusive results.

ALBITE-QUARTZ-MUSCOVITE REPLACEMENT UNITS

Petrochemistry of replacement units is respectively presented in Tables 36 and 37 for the Mavis Lake and Gullwing-Tot Lakes pegmatite groups. Variation among the major components is very limited as

revealed by a tight clustering of analyses in the Si-Al-Na+K+Li diagram towards the Si apex (Figure 66A). The greatest variation is evident in the Ca-Na-K plot (Figure 66B) marked by a limited dispersion of analyses along the Na-K sideline near the Na apex. Only sample Fairservice 168, containing a relatively high content of green muscovite, departs from the main cluster of analyses. A considerable relative decrease in Li levels from those of primary pegmatite compositions is notable in the Li-Na-K diagram (Figure 59).

Replacement processes have apparently effected no significant fractionation of K over Rb, relative to the spodumene-bearing primary pegmatitic units, even though absolute levels of both elements vary greatly. K/Rb ratios between 4 and 23 in the replacement aplites nevertheless indicate extreme fractionation. This is also revealed by severe losses in Li and increase in Na (Figure 59). Ba/Rb, Rb/Sr, K/Cs and K/Ba ratios are similar to those in the primary pegmatite units (Tables 33 and 34).

Nb/Ta ratios in the primary spodumene-bearing pegmatites (0.5 to >1.7) partially overlap with those in the replacement units (0.34 to >1.2). However, there is a tendency for lower ratios in the latter which is indicative of a higher degree of fractionation in the replacement units. Generally higher concentrations of tantalum (<30 to 380 ppm) are evident in the replacement units, a factor to be considered in their economic evaluation.

Figure 66. Si-Al-Total Alkalies and Ca-Na-K Variation (atomic percent) in Replacement Stage Albitic Rocks from the Dryden Pegmatite Field.

Table 36. Petrochemistry, CIPW norms and Element Ratios of Albitic Replacement Units from the Fairservice Property, Mavis Lake Pegmatite Group.

Pegmatite No. Sample No.	MORTH ZONE				EAST ZONE			
	1	2	3	4	13	14	15	247
SiO ₂	74.50	72.80	76.00	71.10	71.70	73.68	73.30	70.80
Al ₂ O ₃	16.80	17.90	15.00	17.50	18.00	16.40	17.10	18.40
Fe ₂ O ₃	0.02	0.13	0.13	0.05	0.13	0.13	0.13	0.13
FeO	0.13	0.13	0.13	0.13	0.13	0.13	0.13	0.13
MgO	0.07	0.07	0.09	0.03	0.01	0.03	0.03	0.03
CaO	5.53	6.29	4.75	9.40	7.05	5.45	6.82	7.88
Na ₂ O	2.41	1.97	5.36	2.27	1.54	3.30	1.17	1.12
K ₂ O	0.02	0.01	0.02	0.03	0.02	0.02	0.02	0.02
TiO ₂	0.05	0.16	0.07	0.14	0.05	0.08	0.01	0.01
P ₂ O ₅	0.00	0.01	0.00	0.01	0.01	0.01	0.13	0.30
MnO	0.02	0.03	0.03	0.03	0.03	0.03	0.04	0.04
CO ₂	0.15	0.14	0.15	0.22	0.25	0.18	0.10	0.27
H ₂ O ⁺	0.91	0.38	0.36	0.32	0.43	0.13	0.36	0.66
H ₂ O ⁻	0.07	0.49	0.17	0.09	0.43	0.13	0.36	0.14
Li ₂ O	0.17	1.08	0.29	0.14	0.50	0.05	0.43	0.05
TOTAL	99.44	99.62	98.03	98.18	98.75	99.98	98.85	98.39
B	39	16	15	100	108	35	41	400
Ba	20	50	40	60	30	40	30	20
Be	255	45	60	170	55	90	230	90
Cs	70	30	90	60	30	20	60	330
F	220	100	50	60	90	140	210	700
Ga	25	35	34	27	24	37	53	50
Li	800	1100	5000	1330	132	670	845	225
Nb	100	75	35	210	120	100	75	95
Rb	1800	1150	3700	1800	1000	190	1760	2600
Sc	<1	<1	<1	<1	<1	<1	<1	<1
Sn	66	70	59	29	16	5.7	69	40
Sr	18	18	17	16	17	17	15	25
Ta	140	100	<30	380	280	140	120	280
Th	10	10	10	10	<10	<10	10	10
U	1	5	3	2	3	8	2	5
V	2	3	<1	2	<1	<1	4	3
Zn	<2	<2	<2	<2	<2	<2	<2	<2
Zr	30	22	12	24	20	30	30	30

Table 36, continued

Al/Ga	3562	2438	2790	2944	3313	2507	1660	2650	3219	2924	1950
Ba/Rb	0.01	0.01	0.01	0.03	0.03	0.21	0.02	0.003	0.007	0.03	0.008
Rb/Cs	26	38	41	30	33	9.5	29	23	32	35	8
Rb/Sr	100	64	106	113	59	11	117	84	181	62	104
K/Ba	1000	327	1112	398	426	93	628	320	1370	324	465
K/Cs	286	545	494	398	426	187	314	213	304	324	29
K/Rb	11	14	12	13	13	20	11	9.5	9	9	4
Th/U	10	2	5	2.5	<3.3	<1.3	5	5	5	1.3	2
Nb/Ta	0.71	0.75	>1.2	0.55	0.43	0.71	0.63	0.44	0.64	0.63	0.34
Na/K	2.1	2.8	0.4	1.5	3.9	18.6	2.1	4.1	1.5	5.2	6.3
Al ₂ O ₃ /CaO+Na ₂ O+K ₂ O+Li ₂ O+Rb ₂ O	1.422	1.264	1.794	1.354	1.173	1.049	1.443	1.326	1.293	1.334	1.250

(Molecular)

NORMATIVE MINERALS

Quartz	32.4	27.9	35.8	35.0	27.7	13.0	31.4	22.5	27.1	26.8	18.2
Albite	47.1	56.2	22.6	43.1	60.6	81.7	49.1	62.8	49.0	60.9	70.1
Anorthite								0.6		0.1	
Orthoclase	14.3	11.6	32.3	17.2	9.1	2.6	13.5	9.0	19.5	6.9	6.6
Eucryptite											
Corundum	6.1	3.8	9.0	4.5	2.5	1.6	5.7	4.9	4.2	5.0	4.6
Apatite	0.1	0.1	0.1	0.2	0.1	0.7	0.1	0.1	0.2	0.3	0.5
Ilmenite											
Ferrosilite		0.1				0.2					
Hypersthene		0.1				0.2					

SAMPLE DESCRIPTION

- 172: 0.82 m channel sample across white beryl - green muscovite - quartz - albite unit, Pegmatite 1.
- 248: 3 m channel sample across albitized spodumene pegmatite, west end of pegmatite 1.
- 168: 2.6 m channel sample across albitized spodumene pegmatite, Pegmatite 2.
- 175: 0.9 m channel sample across spodumene - green muscovite - K-feldspar-quartz-albite unit. This unit contains highest mesoscopic tantalite contents in Mavis Lake Group.
- 176: 1 m channel sample across albitized spodumene pegmatite (Pegmatite 2).
- 151: 2.3 m channel cut across albitized unit between spodumene pegmatite and north contact. Unit contains platy tantalite, white beryl, green muscovite, quartz, and albite. (Pegmatite 4).
- 150: 2.15 m channel cut across albitized zone, Pegmatite 6. Unit contains platy tantalite, beryl, green muscovite, quartz, and albite.
- 244: 1.3 m channel cut across beryl-tantalite-green muscovite-quartz-albite unit, Pegmatite 13.
- 245: 0.63 m channel cut across tantalite-quartz-albite dyke which contains spodumene totally replaced by green muscovite. Pegmatite 14.
- 246: 1.7 m channel cut across green muscovite-quartz-albite dike with spodumene totally replaced by green muscovite (Pegmatite 15).
- 247: Albite-rich pod which contains minor tantalite, green muscovite, quartz, and green tourmaline (600 m north of Pegmatite 14).

Table 37, continued

	NORMATIVE MINERALS	
Quartz	35.5	11.2
Albite	36.5	14.9
Anorthite	2.6	2.4
Orthoclase	16.6	2.6
Eucryptite		
Corundum	7.3	2.1
Apatite		0.8
Ilmenite	0.1	0.2
Rutile		0.3
Magnetite		0.1
Ferrosillite		0.1
Hypersthene	0.9	
Hematite	0.6	
Enstatite	0.9	
Al/Ga	1373	4982
Ba/Rb	0.09	0.11
Rb/Cs	7.3	14
Rb/Sr		19
K/Ba	269	191
K/Cs	170	286
K/Rb	23	20
Mg/Li	5	6
Th/U		
Nb/Ta	6.9	6.7
Na/K	4.9	0.2
Al ₂ O ₃ /CaO+Na ₂ O+	1.11	7.5
K ₂ O+Li ₂ O+Rb ₂ O+Cs ₂ O	1.112	1.06
(molecular)	1.580	1.024

250: large bulk sample of albitization I unit, North Zone of the Gullwing Lake Pegmatite

249: 2.6 m channel sample across albitization I unit, at the South Zone of the Gullwing Lake Pegmatite.

191: 0.65 m channel sample across albitization II unit, Central Zone of the Gullwing Lake Pegmatite.

248: Green muscovite-quartz-albite unit replacing spodumene-K-feldspar pegmatite unit (unit I) of Tot Lake Pegmatite

EXOMORPHIC GEOCHEMICAL DISPERSION

INTRODUCTION

Late-to-post magmatic replacement processes have variably affected the most chemically fractionated units of the Ghost Lake Batholith and all rare-element pegmatites studied. This is recognized by replacement of primary pegmatitic minerals such as blocky K-feldspar, spodumene, beryl, quartz and albite by secondary assemblages consisting mainly of albite and quartz. This pegmatite "reprocessing" event led to a significant egress of elements such as Li, K, B, Rb, Cs and F, which could not be accommodated in great quantities within the albite-rich replacement unit. Consequently, these "incompatible" elements were transmitted into the adjacent mafic metavolcanic host rocks. In addition to constituting an important exploration tool (Trueman 1978) the chemistry of metasomatic assemblages developed proximal to variably-replaced rare-element pegmatites potentially can shed some light on the composition of late stage fluids emanating from these magmatic/hydrothermal systems (Morgan and London 1987).

MINERALOGICAL FEATURES OF EXOMORPHIC DISPERSION

Exomorphic dispersion is most readily recognized in mafic metavolcanic units of the study area by elevated concentration of minerals not typically widespread in these rocks. Most prominent are black tourmaline, biotite and holmquistite. These minerals may occur individually or co-exist within pods, layers, replaced pillow selvages, porphyroblasts and as selvages adjacent to quartz and granitic veins (Plate 9g).

Tourmaline, the most conspicuous exomorphic mineral, is restricted to the Mavis Lake pegmatite group. Here prominent porphyroblasts generally occur within 3 m of pegmatite contacts (Plate 12e). The metasomatic development of tourmaline may have been induced by prior destruction of a primary tourmaline-bearing pegmatite assemblage via albitization and consequent diffusion of boron-rich volatile complexes into the host rocks (Beus et al. 1986). Field and petrographic evidence suggest that exomorphic dispersion occurred in two stages which can be distinguished on structural grounds:

- 1) a relatively early tourmaline-rich selvage (boron metasomatism)
- 2) a later biotite-rich selvage without tourmaline (potassium-rare alkali metasomatism).

The early boron-rich exomorphic event is locally evident by a black tourmaline-rich selvage adjacent to narrow dykes emanating from the batholith (Plate 9f). Mild boudinage of the dyke and selvage suggests that the metasomatism occurred prior to this deformation. The ensuing potassium-rich exomorphic pulse produced wider, biotite-rich selvages which occurred subsequent to this boudinage. This is suggested by the regular, planar aspect of these selvages in contrast to the scalloped nature of the tourmaline-rich ones.

A similar sequence 4.6 kms to the east occurs along the north contact of spodumene-beryl-tantalite Pegmatite No. 4, North Zone of the Fairservice Property. Here the metasomatic selvages include holmquistite which can be seen mesoscopically as randomly oriented needles overprinting tourmaline. Holmquistite also occurs as fine-grained, late fracture-fillings with albite in massive hornblende-

porphyroblastic mafic metavolcanics adjacent to pegmatite No. 5 Fairservice Property and the Tot Lake Pegmatite and in metawackes near the north contact of Pegmatite No. 1 (Fairservice Property). Holmquistite is a specific indicator of lithium exomorphism (Heinrich 1965) and in most situations constitutes a practical exploration pathfinder for highly fractionated Li-rich pegmatites (London 1986). Cerny et al. (1981) documented the presence of holmquistite up within 20 m of rare-element pegmatites of the Bernic Lake Group and Maskwa Lake Series.

Shearer et al. (1984, p.2578-79) speculated that a similar progression of exomorphic pulses affected clastic metasedimentary host rocks adjacent to the Tip Top spodumene pegmatite in the Black Hills Pegmatite District of South Dakota. Morgan and London (1987), on the basis of petrographic and chemical work, proposed that three episodes of metasomatic alteration occurred in amphibolitic mafic metavolcanics adjacent to the Tanco Pegmatite:

- 1) $B \pm Li$ metasomatism (tourmalinization),
- 2) $K-Rb-Cs-F \pm Li$ metasomatism (biotite replacement) and
- 3) propylitic alteration.

SAMPLING AND DATA ANALYSIS

Detailed sampling of mafic metavolcanics was undertaken only around spodumene-beryl-tantalite pegmatites constituting the North Zone of the Fairservice Property. Results of 209 partial analyses for Li, Cs, Rb, B, Be, K_2O , F and Sn are given in Appendix 9; the sample localities can be found in Figure 18.

Elsewhere, chemical profiles were used to evaluate exomorphic dispersion around the Taylor, Gullwing Lake and Tot Lake Pegmatites. These profiles also allow comparison with the Fairservice spodumene pegmatites (Figures 68, 69, and 72; Tables 39 to 43 and Appendix 9).

Background levels for K_2O , Be, Li and Sn were assumed to approximate the average contents of these elements in mafic metavolcanics from the Sioux Lookout greenstone belt (Table 38). Average levels of Cs, F, Rb and B in these rocks are presently unknown and have been assumed to correspond to those in the average Canadian Shield tholeiite of Baragar and Goodwin (1969) and Condie and Baragar (1974). It should be noted that the average K_2O abundance of the Lateral Lake mafic metavolcanics analyzed by McCarter (1981) is being about three times higher than that of Trowell (in preparation) who sampled on a more regional basis. McCarter's sampling was mainly concentrated around the Lateral Lake granitic pegmatite-hosted Mo-Bi deposit. Metasomatic biotite and muscovite derived from fluids which emanated from nearby albitized pegmatites could have contributed to higher K_2O in some of his specimens. Accordingly, the average K_2O content of Trowell (in preparation) has been used.

PETROCHEMISTRY OF METASOMATICALLY AFFECTED MAFIC METAVOLCANICS

Major Element Variation

Bulk analyses of metasomatic selvages developed along pegmatite contacts in the Mavis Lake and Gullwing-Tot Lakes Groups are given respectively in Tables 39 and 40. The most striking petrochemical change associated with the alteration is a marked increase in K_2O relative to mafic metavolcanics of the Sioux Lookout greenstone belt

TABLE 38: Average and Range of Selected Elements in Mafic Metavolcanics from the Canadian Shield.
Clarke Values are Given in Column D (ppm unless otherwise indicated)

	A		B		C	D
	Average	Range	Average	Range	Average	Clarke Value
B						g
Ba	71	30-460	214	40-940	105	390
Be			< 1	< 1-4		2.0
CaO	8.27%	4.92-13.60%	11.03%		9.39%	4.66%
Co			47	40-59		29
Cr			410	<5-860	187	122
Cs					0.4	2.6
F						544
Ga			15	10-15		19
K ₂ O	0.27%	0.00-1.54%	0.71%	0.22-1.68%	0.32%	1.84%
Li	16	4-28	18	8-30		18
Na ₂ O	2.61%	0.00-5.70%	1.99%	1.54-2.75%	2.11%	2.27%
Nb						20
Ni			149	55-337	103	99
Rb					8	78
Sn			<3	<3-5		2.1
Sr			163	60-500	164	384
Ta						1.7
Zn	<70	<10-184	156	91-320	110	76
Zr			96	80-200	105	162
La					7.2	34.6
Ce					15	66.4
Sm					2.5	7.02
Tu					1.3	2.14
Tb					1.2	1.18
Yb					2.2	3.1
Lu					0.37	0.54
Y			24	15-30	20	31.0

- A. Average of 46 mafic metavolcanics from the Sioux Lookout Greenstone Belt (Trowell, in preparation)
 B. Average of 10 mafic metavolcanics from the Lateral Lake area (McCarter 1981)
 C. Average Archean tholeiite from the Canadian Shield (Baragar and Goodwin, 1969, and Condle and Baragar, 1974).
 D. Clarke values from Ronov and Yaroshevski (1967).

Table 39, continued

	ELEMENT RATIOS									
	Al/Fe	2870	269	1057	3433	1205	4080	2575	5221	4566
Ba/Rb	0.02	2.69	0.04	0.04	3.6	0.045	1.2	0.06	0.24	1.1
Ca/Sr	308	409	1505	1490	1490	1116	3207	290	181	124
K/Ba	676	47	352	13.8	13.8	340	16	134	48	25
K/Cs	53	936	37	41.5	41.5		12	25	17	25
K/Rb	15	136	13.6	49.8	49.8	17.2	22	8.1	11.7	20
Mg/Li	8.9	71	41.8	17480		40.4	59.0	9.75	5/5	157.8
Na/K	1.05	6.7	0.03			0.09		0.11	0.81	4.0
Rb/Cs	3.5	7.3	2.70	0.8	0.8		0.56	3.1	1.45	6.89
Rb/Sr	51.6	0.2	153	0.25	0.25	88	3.6	47	4.3	0.5
										4458
										0.67
										597
										34
										233
										910
										25
										27
										19
										5.7
										0.2
										1.8
										0.23

SAMPLE DESCRIPTION AND LOCATION

- A Biotite-rich altered mafic metavolcanics at contact of Ghost Lake Batholith (near locality E31-2)
- B Relatively unaltered amphibolitic mafic metavolcanics one metre from Sample A
- C Biotite-rich altered metabasaltic at contact with the main body of dyor pegmatite dyke 2
- D Identical sample to A, also from contact of Taylor pegmatite dyke 2
- E Serpentinized metabasaltic rock located 5 m west of Taylor beryl pegmatite 2
- F Pillow breccia in brownish mafic metavolcanics replaced by black tourmaline and minor biotite. Sample immediately underlies shallowly-dipping tourmaline-rich sheet containing schreibelite mineralization
- G Located in Iceland Trench (Latitude 49.8°N Longitude 32.7°W)
- H Melnikite-biotite mafic metavolcanics at north contact at south-western beryl pegmatite No. 2 North Zone of Fairweather Property
- I Tourmaline-biotite mafic metavolcanics 1.22 metres north of pegmatite No. 2 contact
- J Biotite-tourmaline mafic metavolcanics 5.18 metres north of pegmatite No. 2 contact
- K Biotite-tourmaline mafic metavolcanics 0.7 metres north of pegmatite No. 2 contact
- L Melnikite-biotite mafic metavolcanics 0.7 metres north of pegmatite No. 4 contact

Table 40. Petrochemistry and Selected Element Ratios of Metasomatically Altered Host Rocks in the Gullwing- Tot Lakes Pegmatite Group.

	Intermediate Metavolcanics		Mafic Metavolcanics				Coates Pegmatite
	GL-1	GL-4A	GL-2	GL-3A	GL-5A	GL-5B	CP-1
SiO ₂	39.3	48.2	53.3	49.9	55.9	38.3	45.7
Al ₂ O ₃	18.1	16.8	20.0	17.6	17.8	18.2	14.4
Fe ₂ O ₃	19.5*	2.82	9.83*	2.96	9.40*	4.40	1.88
FeO		10.4		9.12		13.0	8.16
MgO	6.31	4.81	3.13	4.58	3.68	6.26	7.28
CaO	0.09	0.29	0.54	0.10	1.51	0.24	7.95
Na ₂ O	0.23	1.10	2.77	0.09	3.44	0.31	0.96
K ₂ O	8.82	6.39	6.18	7.20	4.18	8.62	4.70
TiO ₂	0.93	0.96	0.44	0.70	0.39	0.96	0.71
P ₂ O ₅		0.34		0.19		0.45	0.01
S		0.00		0.00		0.00	0.00
MnO		0.62		0.54		0.82	0.29
CO ₂		0.38		0.17		0.16	3.04
H ₂ O*	1.70	2.46	2.54	2.46	1.70	2.44	1.68
H ₂ O-		0.28		0.28		0.26	0.20
TOTAL	94.98	96.0	98.73	96.0	97.20	94.4	97.0

*Total iron as Fe₂O₃

B		55		<5		<5	
Ba	280	90	150	120	120	70	60
Be		6		5		6	20
Ce		1170		470		1185	
Cs	1100	750	600	790	640	1140	705
Co	31			28		39	44
Cr		57		31		87	208
Cu		7		10		8	14
F		9000		10200	18400	10400	10400
Ga		97		103	120	27	
Li	11000	2950	2200	3060	1960	4600	1450
Mo		<10		<10	20	<10	<10
Nb	950	435	490	605	430	925	13
Nd		505		170		345	45
Ni		72		53		97	105
Pb		53		79		69	28
Rb	6890	5280	3750	5500	3450	7580	4890
Sc		30		25	35	35	35
Sn	320	109	100	85.6	60	157	30.6
Sr		30		30		35	75
Ta	100	35	54	55	48	60	<10
Tb		110		75		80	20
U		18		23		38	<1
V		160		120		130	230
W		<50		<50		<50	<50
Y		105		45		95	16
Zn		1670		1040		2560	270
Zr		295		335		455	14
Y		156		69		168	17
La		314		120		352	28
Ce		940		350		914	29
Nd		403		138		428	13
Sm		135		1.4		149	5
Eu		2.3		39		3.4	1.5
Gd		96		17.3		114	9
Dy		39.5		17.3		49.4	3.5
Yb		10.5		5.9		13.4	2.4
ΣREE		1940.3		720.6		2023.2	89.4
La/Ybn		19.6		13.4		17.2	7.5
Eu/Eu*		0.06		0.1		0.08	0.69

unassociated with rare-element mineralization (Figure 67). Potassium metasomatism correlates with generally lower CaO contents; Na₂O levels are approximately similar.

K₂O levels in selvages from the Gullwing Lake pegmatitic granite range between 4.18 and 8.82 percent, the highest within the study area. Otherwise, K₂O contents of both pegmatite groups are similar, clustering approximately midway along the Ca-K sideline (Figure 67) and distinctly different from unmineralized mafic metavolcanics of the region.

Calcium exhibits drastic depletion, especially near the Gullwing Lake pegmatitic granite. Here, levels fall within a 0.09 to 1.51 percent range, well below the regional ranges for mafic metavolcanics of the Sioux Lookout area (4.92 to 11.1 percent) and Lateral Lake area (7.95 to 14.2 percent).

Potassic metasomatism causes slightly higher total alkalis which results in selvedge compositions plotting near or within the calc-alkaline field in the AFM diagram (Figure 67). Only about 15 percent of mafic metavolcanics from the Sioux Lookout greenstone belt match AFM ratios of the contact metasomatic selvages.

Similarity of variation patterns of Cs, Li, Rb and K₂O are generally evident in the exomorphic chemical profiles illustrated in Figures 68, 69, and 70 (in pocket). Mineralogically, this relationship can generally be related to diadochic incorporation of Cs, Li and Rb in biotite.

Figure 68. Comparison of Exomorphic Chemical Profiles Adjacent to Beryl and Spodumene Pegmatites of the Mavis Lake Group.

Figure 6-. Exomorphic Chemical Profiles in Mafic Metavolcanics Adjacent to Spodumene Pegmatites of Gullwing-Tot Lakes Pegmatite Group.

One of the best means of chemical discrimination between metasomatized and unaffected mafic metavolcanics is provided by the Mg-Li diagram. A considerable disparity between these two rock groups is evident in Figure 71. Metasomatic selvages adjacent to rare-element pegmatites exhibit Mg/Li ratios approximately 230 times lower than mafic metavolcanics not known to host such pegmatites. The latter show a distinct cluster which is approximately bisected by the average crustal Mg/Li ratio of 2300. It should also be noted that Mg contents in both groups are very similar, which provides further chemical evidence of the metasomatic addition of lithium to mafic metavolcanics hosting rare-element pegmatites.

Minor and Trace Element Variation

Positive lithochemical anomalies involving B, Be, Cs, F, Li, Rb and Sn envelop all rare-element pegmatites studied. The relative mobility of these elements has been mainly evaluated by comparing aureole width on the pegmatite hanging wall side at an arbitrary enrichment value of five times background level (Table 41; Figures 68, 69, and 70).

Lithium generally forms the widest exomorphic haloes around spodumene pegmatites of the study area. This observation concurs with the data of Beus et al. 1968) who considered lithium to be one of the most mobile exomorphic elements. Lithium aureoles in the Dryden Field can be up to 400 m in width at the 100 ppm level (Table 41).

B.M. Shmakin (personal communication 1980) has indicated that lithium disperses up to 100 metres from pegmatites in the U.S.S.R.

Figure 71. Li-Mg Variation in Metasomatic Selvedges Adjacent to Rare-Element Pegmatites Compared to Unaffected Mafic Metavolcanics.

TABLE 41:

Comparison of Exomorphic Chemical Aureole Radii in Rare-Element Pegmatite Groups of the Dryden Field

	I		II	
	GULLWING - TOT LAKES		MAVIS LAKE	
	PEGMATITE GROUP		PEGMATITE GROUP	
	Gullwing Lake	Tot Lake	Jaylor Pegmatite 1	Fairservice Pegmatite 2
	Pegmatite	Pegmatite		
K ₂ O	7.3m	6.1m	1.4m	1.25m
B	none	7.6m	3.75m	18-37m
Be	1.5m	none	<14 cm	5 cm
Cs	10m	50m	3.5m	26-62m
F	1.8m	0.6m	0.5m	3-6m
Li	1.2m	29m	3.45m	120m
Rb	3.7m	23.8m	2.2m	9-37m
Sn	2.4m	1.5m	4.25m	50cm

Extreme exomorphic concentrations of Li occur in both pegmatite groups of the study area. Most notable are those adjacent to the Gullwing Lake Pegmatite (2.4 percent Li_2O) and Fairservice Pegmatite No. 4 (1.48 percent Li_2O). In certain circumstances, low exomorphic concentrations of K_2O in mafic metavolcanics apparently favour formation of holmquistite. Otherwise, lithium is preferentially incorporated into exomorphic biotite. This is revealed by a chemical comparison of metasomatic selvages which are developed around spodumene pegmatites. Biotite-rich assemblages lacking holmquistite can contain up to 2.4 percent Li_2O and 8.82 percent K_2O as seen near the apical part of the Gullwing Lake Pegmatite (North Zone). In contrast, holmquistite-rich metasomatic assemblages apparently have lower Li_2O (1.48 percent) and extremely low K_2O (0.20 percent) as determined near Pegmatite No. 4 on the Fairservice Property at Mavis Lake.

Field observations in the study and those by B.M. Shmakin (personal communication 1980) in the U.S.S.R. indicate that holmquistite formation is restricted to within 3 m of a rare-element pegmatite. Cerny et al. (1981, p.101) have identified holmquistite up to 20 m from pegmatites in the Cat Lake-Winnipeg River Field of southeastern Manitoba.

All rare elements determined define distinctive elongate anomalies which are oriented east-west. Dispersion of Li, B and F is extensive compared to K_2O , Rb, Ca, Be and Sn. Such anomaly shapes characterize steeply-dipping rare-element pegmatites (Trueman and Cerny 1982, p.488-89).

The width of lithium aureoles is greatest adjacent to spodumene pegmatites of the Mavis Lake Group (Table 41). Hand-contoured plots of various exomorphic elements for the North Zone of the Fairservice property are presented in Figure 70 (in pocket). This area of detailed sampling represents a small portion of an extensive 100 ppm Li anomaly reported by Pryslak (1981). His work indicates a length of at least 3,050 m and a variable width of 120 m to 400 m. The northern and western terminations of this anomaly are presently not established.

The pattern of lithium distribution culminates in a 15-60 by 595 metre anomaly in which concentrations exceed 500 ppm. This anomaly corresponds to a close grouping of spodumene Pegmatites No. 2, 3 and 4 (Figure 70).

Figure 68 presents in detail the variation of selected elements in mafic metavolcanics situated within 3 metres of an extensively albitized portion of Pegmatite No. 4. This profile illustrates the similarity in curve shapes of K_2O , Cs, Rb, Li and the narrowness of the Be and Sn aureoles. Boron has the most variable abundance curve and does not correlate well with any of the other elements. Spodumene pegmatites of the Gullwing-Tot Lakes group tend to have less extensive exomorphic aureoles (Table 41). Even adjacent to highly fractionated pegmatites such as the Tot Lake Pegmatite the Li aureoles are only 30 m in width compared to 60-120 m for such aureoles on the Fairservice property. However, concentrations of many elements can be considerably elevated above their background values even where the metasomatic selvages are relatively narrow. Highest exomorphic

Table 42: Variation of Selected Elements and Ratios in Mafic Metavolcanics
Within 100 metres of the Gullwing Lake Pegmatite Contact (North Zone)

	<u>CONTACT</u>	<u>4.3 m</u>	<u>7.3 m</u>	<u>14 m</u>	<u>29m</u>	<u>44.2 m</u>	<u>71.6m</u>	<u>91.44</u>
CaO	2.69%	11.7	11.5	13.1	12.8	12.2	12.9	13.4
K ₂ O	5.65%	2.57	2.62	0.45	0.43	0.44	0.29	0.23
Na ₂ O	1.34%	0.66	0.64	2.18	2.07	2.50	1.87	2.18
B	<5	5	5	<5	<5	<5	<5	<5
Be	12	1	1	<1	<1	1	<1	<1
Cs	1060	20	10	<10	10	<10	<10	20
F	1.27%	230	410	450	270	520	160	190
Li	2880	75	106	68	32	40	45	27
Sn	31	1.1	1.4	1.3	1.1	1.1	0.8	<0.8
Sr	95	265	315	235	95	255	90	100
Rb	3020	20	20	<10	<10	<10	<10	10
Ca/Sr	210	314	295	396	957	340	1018	951
K/Cs	44	1067	2175	>374	357	>365	>241	191
K/Rb	15.2	1067	1087	>374	>357	>365	>241	191
Rb/Cs	2.8	1	2	<1	<1	2	018	951
Rb/Sr	31.8	0.08	0.08	<0.04	<0.1	<0.04	<0.1	0.1
K/Cs	44	1067

Figure 72. Variation of Selected Elements and Ratios in Metasomatic Selvages Developed in Mafic Metavolcanics Adjacent to North Zone of Gullwing Lake Pegmatitic Granite.

concentrations of Li, Nb, Rb, Sn, Ta, F, K₂O, Zn, Zr, Y and total REE in the Dryden Field occur within 10 cm of the Gullwing Lake Pegmatite (Tables 42 and 43, analyses GL-1 and GL-5B).

Profiles of exomorphic elements adjacent to the Tot Lake Pegmatite are given in Figure 69 and Table 44. Most notable is the 50 m wide Cs aureole which is considerably greater than any other in the Dryden Field.

This aureole represents the only example where cesium mobility has apparently surpassed that of lithium and thus may be an important factor in exploration for Li-Cs-Ta pegmatites in the area. The host rocks of this pegmatite contain the highest exomorphic cesium concentration (1.6 percent) in the Dryden Field. K/Cs ratios reveal an anomalous zone at least 123 m in width where ratios vary between 2.5 and 672. A more complete examination of K/Cs, K/Rb and Rb/Cs variation around the study area pegmatites is currently hampered by a 10 ppm analytical sensitivity by XRF for Rb and Cs (Appendix 1).

A detailed presentation of K₂O, Cs, Li, Rb and Ba variation within a 12 cm wide biotite-rich contact selvage on the Gullwing Lake Pegmatite (similar to that in Plate 15c) is given in Table 43 and graphically portrayed in Figure 72. Lithium has the lowest rate of abundance decrease with increasing distance, relative to other pathfinder elements. Enrichment of Cs relative to Rb within 7 cm of the contact is reflected by a marked drop in Rb/Cs, and K/Cs ratios which are considerably lower than the average Canadian Shield tholeiite (Rb/Cs = 80; K/Cs = 6640). K/Rb values are also considerably depressed (average Canadian Shield tholeiite K/Rb=332)

Table 43: Variation of Selected Elements and Ratios
in a Biotite-Rich Metasomatic Selvege Developed in Mafic
Metavolcanics as a Function of Increasing Distance from
Contact of Gullwing Lake Pegmatite (North Zone)

Distance from Pegmatite Contact	<u>1 cm</u>	<u>3 cm</u>	<u>5 cm</u>	<u>7 cm</u>	<u>9 cm</u>	<u>11 cm</u>
%K ₂ O	4.91	2.73	1.34	1.01	0.86	0.80
Ba	200	120	110	70	80	90
Cs	560	300	88	34	25	22
Li	2200	1300	780	560	470	400
Nb	110	10	20	10	20	<10
Rb	3250	1170	480	340	240	210
Sn	3	12	10	12	15	15
Ta	5	<5	<5	<5	<5	<5
RATIOS						
K/Ba	204	189	101	120	89	74
K/Cs	73	75	126	247	285	302
K/Rb	12.5	19.4	23.2	24.7	29.7	32.6
Ba/Rb	6.2	10.3	22.9	20.6	33.3	42.9
Rb/Cs	5.8	3.9	5.5	10	9.6	9.6

Table 44: Variation of Selected Elements and Ratios in Mafic Metavolcanics as a Function of Increasing Distance from the Tot Lake Pegmatite. Concentrations in ppm Unless Otherwise Indicated.

Distance from Pegmatite Contact	0m	1.4m	2m	5.5m	8.2m	10.7m	13.7m	35m	47m	81m	110m	123m	174m	203m
%CaO	6.41	8.35	6.11	5.26	12.3	10.7	10.7	11.3	12.2	12.2	12.3	12.2	11.90	11.50
%Na ₂ O	0.80	0.94	1.70	1.28	1.37	2.39	1.87	1.42	1.82	1.66	0.71	1.75	1.18	1.45
%K ₂ O	4.78	0.57	0.32	2.23	0.11	0.15	0.57	0.34	0.35	0.34	0.81	0.19	0.28	0.19
B	25	110	28	1110	<5	<5	49	13	<5	<5	7	<5	7	<5
Be	2	2	1	1	1	<1	1	1	2	1	<1	1	1	1
Cs	1.6%	600	480	2760	20	50	80	10	20	<10	10	10	<10	<10
F	5930	550	380	940	<50	170	520	80	710	80	210	<50	340	120
Li	1900	955	810	1480	160	192	275	48	30	14	29	15	15	19
Mn	1600	1100	1070	1220	930	1100	1210	1520	1270	1410	1440	1320	1690	1670
Rb	4820	420	240	200	20	30	130	10	10	<10	30	<10	<10	<10
Sn	23.9	3.0	1.5	4.1	2.3	1.0	2.9	1.9	1.2	1.0	3.0	<0.8	16.8	2.0
Sr	90	65	85	60	105	115	135	75	115	115	70	115	70	60
N	<50	<50	<50	<50	<50	<50	<50	<50	<50	<50	<50	<50	<50	<50
RATIOS														
K/Cs	2.5	7.9	5.5	6.7	46	25	59	282	145	>282	672	158	>232	>158
K/Rb	8.2	11.3	11.1	93	47	42	36	282	291	>291	224	>158	>232	>158
Rb/Cs	0.3	0.7	0.5	0.07	1	0.6	1.6	1	0.5	-	3	<1	-	-
Rb/Sr	54	6.5	2.8	3.3	0.2	0.3	1.0	0.1	0.09	<0.08	0.4	<0.08	<0.14	<0.17
Ca/Sr	377	912	510	622	832	661	563	1070	753	753	1248	753	1207	1361
Na/K	0.15	1.5	4.7	1.95	11.1	14.2	2.9	3.7	4.6	4.4	0.8	8.2	3.8	6.8

and are comparable to those in the most chemically fractionated K-feldspars from the Gullwing Lake pegmatite.

Beryl pegmatites generally exhibit much narrower aureoles for most exomorphic elements studied as than compared to spodumene pegmatites (Table 41; Figure 69). This, in part, relates to their smaller size, lesser amount of albitization and more primitive chemical constitution.

The comparison in exomorphic aureole size for beryl and spodumene pegmatites is best made in the Mavis Lake group as the Gullwing-Tot Lakes group does not contain strictly beryl-bearing pegmatites such as at the Taylor Occurrence. The Fairservice spodumene pegmatites reveal more extensive exomorphic aureoles for Li, Rb, Cs, and B than the beryl-bearing pegmatites. K_2O and Be aureoles are comparable. Tin forms an exceptional aureole around the Taylor beryl pegmatite No. 1 which is about eight times wider than those adjacent to the Fairservice spodumene pegmatites (Figure 68).

Relationships of the exomorphic metals Li and Sn with volatiles F and B contained in mafic metavolcanics of the Dryden Field are presented in Figure 73. It is plausible that fluorine involved in anionic complexes could have mobilized cations such as lithium and tin during post-magmatic alteration of granites (Bailey 1977). Little is known regarding the role of boron in rare-element transport. Sn-F exhibits the best possible correlation, followed by that between Li and F (Figure 73). Boron, however, has no evident correlation with Sn or Li and is concluded to have had a negligible role in the exomorphic transport of rare elements into the host mafic metavolcanics. Only

Figure 73. Li and Sn vs. F and B Variation (ppm) in Mafic Metavolcanics which Host Rare-Element Pegmatites in the Mavis Lake and Gullwing-Tot Lakes Group.

negligible levels of exomorphic boron were found in mafic metavolcanics here and yet the Li-F and Sn-F correlations are similar to those in the Mavis Lake Pegmatite Group.

Tantalum-niobium exomorphism: these elements were concluded by Beus et al. (1968) to be immobile during transfer of fluid phase components from rare-element pegmatites undergoing albitization, and consequently should not be enriched in metasomatized wall rocks. Zagorsky et al. (1979, p.110), however, indicated that "patchy low contrast anomalies in Ta and Nb occur at distances of not more than one metre from the tantalite-lepidolite-petalite veins in the U.S.S.R.". Portions of the mica-rich selvages at the contact of these pegmatites contain up to 650 ppm Ta and 135 ppm Nb; however, values fall to background levels 15 to 30 cm away.

Local exomorphic enrichment of Ta and Nb is supported by the present study but only adjacent to the Gullwing Lake pegmatitic granite. Biotite-rich metasomatic selvages from the North Zone contain anomalous Ta (35-100 ppm) and Nb (430-950 ppm) within 0.5 m of the contact (Table 40). Nb/Ta ratios in these selvages range from 9 to 15.4 (Table 40) and are similar to those in columbites from the adjacent albitization I zone (7 to 13.9 in Appendix 40). The lower end of the selvedge Nb/Ta range approaches the bulk Nb/Ta ratio of 6.9 in the albitic replacement unit of this zone (Table 36).

Nb/Ta ratios in the metasomatic selvages may possibly be utilized as a rapid means of assessing economic potential, which in the case of the Gullwing Lake pegmatitic granite is low due to great predominance of Nb over Ta. The possible exploration applicability of

this relationship in other rare-element pegmatite fields remains to be tested.

Rare-Element Pegmatites and Contact Metasomatic Selvedges of the Ghost Lake Batholith

Enrichment in Be, Cs, F, Li, Rb, Sn and K are apparent in biotite-rich selvedges developed in metavolcanic rocks at the only exposed contact of the eastern lobe of the Ghost Lake Batholith (Table 45, Plate 9f). Most notably enriched are cesium and rubidium, which respectively show concentrations of 1890 and 235 times greater than in the average Canadian Shield tholeiite in Table 38. Fluorine also exhibits considerable enrichment (1.78 percent) which suggests that this element may have played an important role in the genesis of the youngest and most chemically evolved plutonic units of the batholith (Table 45). These chemical data also lend further credence to the GLB as a rare-element source and underscores the importance of sampling of all potassic metasomatic selvedges in regional exploration programs.

In summary, the main features characterizing exomorphic chemical dispersion in the Dryden Pegmatite Field are:

- 1) potassium and boron metasomatism are spatially associated in the Mavis Lake Group whereas potassic alteration is greatly predominant in the Gullwing-Tot Group. The added presence of boron may be due to the relative proximity of the Mavis Lake rare-element pegmatites to a boron-rich parental granite;
- 2) boron exhibits a poor correlation with Li and Sn and probably was not a factor in the volatile complexing and subsequent mobilization of such metals during albitization of the pegmatite source;

Table 45. Partial Chemical Analyses of Contact Metasomatic Selvedges and Greisen from Contact Beryl Occurrence, Zealand Township. Concentration in ppm unless otherwise indicated.

ALTERATION TYPE	Sample No.	E	Li	Rb	Cs	Be	Sn	Nb	La	W	K ₂ O	Na ₂ O
<u>CONTACT METASOMATIC ROCKS</u>												
Albite-biotite selvedge	88FWB-1	4100	1860	2476	702	17	147	14	<10	<50	4.48%	0.53%
Albite-biotite selvedge	Gl884-1	2900	1490	2460	595	10	65.9	21	11	<50	2.72%	
<u>AUTOMETAMORPHIC GREISEN</u>												
Cassiterite-Chrysoberyl-albite-beryl-green muscovite greisen	Gl888-20	1800	303	3370	195	1000	440	65	91	<50	7.42%	1.48%
Pyrite-muscovite-biotite-plagioclase	Gl888-24		1050	1915	670		102	16	<10	<50		
<u>EXOGREISEN</u>												
Plagioclase-rich rock with porphyroblasts of holmquistite, actinolite, biotite and chlorite	Gl888-5A		1960	128	170	2	2.3	<5	<10	<50	0.55%	3.66%
	Gl888-5B		1340	128	170	2	2.3	7	<10	<50	0.57%	2.81%
	Gl888-8A		2450	270	340	2	2.9	<5	<10	<50	1.09%	2.78%
	Gl888-9A		1020	190	220	2	1.6	6	<10	<50	0.80%	3.60%
	Gl888-9B		1480	210	235	2	1.5	<5	<10	<50	3.85%	0.76%
Albite-tourmaline-muscovite rock	Gl888-22	650	178	169	26	<1	3	5	<10	<50	4.9%	1.54%
Actinolite-biotite-tourmaline rock in mafic-rich, marginal sub-unit	88FWB-14	1.30%	1270	1950	915	9	52	<5	<10	<50	6.01%	0.42%
	Gl884-3	1.78%	1400	1860	755	6	142	7	<10	<50		

- 3) most of the important exomorphic elements (Li, Rb, Cs, Sn and Be) show a distinct correlation with K_2O ;
- 4) Li, Cs and Rb aureoles enveloping spodumene pegmatites are more extensively developed where concordant to host rock strike (i.e. Mavis Lake Group) in contrast to those which are highly discordant (Gullwing-Tot Group).

RARE EARTH ELEMENT GEOCHEMISTRY OF PERALUMINOUS GRANITE-PEGMATITE SYSTEMS

INTRODUCTION

Rare-element pegmatites are among the most chemically fractionated igneous systems in Archean rocks of the Superior Province (Goad and Cerny 1981; Breaks and Bond, in press). It is striking that fractionation processes operative in peraluminous granite-pegmatite systems within the crust can reduce total REE contents of rocks ultimately to levels which are comparable to those in many so-called primitive rocks.

Cerny et al. (1981, p.146) were among the first to point out that flat REE patterns featuring drastic depletion in total REE, coupled with large negative Eu anomalies, are characteristic of "fertile" granites and consanguineous rare-element pegmatites. Earlier, Emmermann et al. (1975) had studied REE variation in Hercynian S-type peraluminous granites of North Schwarzwald, Germany and documented identical chemical characteristics. Similar REE variation patterns are evident in the Ghost Lake Batholith and derived barren and rare-element mineralized pegmatites.

The flattening effect on chondrite patterns of a greater rate of LREE depletion relative to HREE has been ascribed to the influence of residual biotite and apatite (Puchett and Emmerman 1976, Emmerman et al. 1976, Hanson 1978). These minerals should also contribute to the observed negative Nd and pronounced negative Eu anomalies. The negative Nd anomalies are apparently characteristic of peraluminous granites and derived pegmatites and yet little mention has been made in the literature of their possible significance. Analyses in this

work, Cerny et al. (1981, p.67), Goad and Cerny, 1981, p.1985) and Kistler et al. (1981, p.10604) demonstrate these patterns. Kistler et al. (ibid) suggested that Nd and HREE depletion in the Jurassic Ruby Mountains Pluton of Nevada, relative to the average Archean sediment, is due to residual garnet left in the magma source area after anatexis.

The ensuing discussion deals REE variation in the plutonic units of the Ghost Lake Batholith, in barren and rare-element exocontact pegmatites and in contact metasomatic selvages of the Ghost Lake Batholith. REE variation in the host metasediments has already been described in this chapter.

GHOST LAKE BATHOLITH

REE analyses are presented in Table 46 and Figures 74 and 75. Considerable variation in REE patterns is evident in Figures 74 and 75; total REE contents vary from 6.2 to 273 ppm, La/Yb_N from 2.4 to 337 and Eu/Eu* from <0.1 to 1.98. As previously pointed out in this chapter there is no significant correlation of total REE with increasing SiO₂. However, total REE contents of peraluminous granites decrease rapidly with increasing fractionation (Cerny and Meintzer 1986), evident in the Ghost Lake Batholith when total REE levels are plotted against K/Rb (Figure 54).

Highest total REE and La/Yb_N values are generally found in plutonic units GLB-1 (29.8 ppm to 273 ppm, average La/Yb_N=100.4) and GLB-2 (128 to 178 ppm, average La/Yb_N=69.0) as shown in Figure 75. In GLB-1 however, 3 analyses with positive Eu anomalies are marked by considerably lower REE (29.8 to 54.0 ppm) and generally shallower

TABLE 46: Rare-Earth Elements (ppm) in Plutonic Units of the Ghost Lake Batholith, Barren and Rare-Element Pegmatites and Contact Metasomatic Selvedges

1.	GHOST LAKE BATHOLITH									
	G/LB 1									
	G/LB 1	G/LB 2	G/LB 3	G/LB 4	G/LB 5	G/LB 6	G/LB 7	G/LB 8	G/LB 9	G/LB 10
La	7.4	11	40	13	47	42	67	125	13	67
Ce	13	21	81	45	92	81	125	13	11	125
Pr	1.3	2.3	9.8	3.0	11	8.9	13	13	34	13
Nd	4.3	8.2	31	9.3	34	32	48	48	6.6	48
Sm	1.1	1.7	6.1	2.1	6.6	6.8	9	9	0.95	9
Eu	0.64	0.80	0.86	0.87	0.99	0.95	0.85	0.85	0.54	0.85
Gd	0.94	1.5	3.9	1.3	3.8	4.8	5.8	5.8	0.27	5.8
Tb	0.13	0.21	0.48	0.16	0.39	0.54	0.63	0.63	0.27	0.63
Dy	0.51	0.98	1.9	0.49	1.2	2	2.4	2.4	0.44	2.4
Ho	0.07	0.17	0.27	0.06	0.13	0.27	0.32	0.32	0.055	0.32
Er	0.14	0.43	0.55	0.10	0.16	0.44	0.55	0.55	0.06	0.55
Tm	<0.05	<0.05	0.06	<0.05	<0.05	<0.05	0.06	0.06	0.06	0.06
Yb	0.12	0.29	0.44	0.10	0.09	0.26	0.42	0.42	0.06	0.42
Lu	<0.05	<0.0	0.07	<0.05	<0.05	<0.05	0.06	0.06	0.06	0.06
Y	2.1	4.4	6.8	1.6	3	6.9	9	9	103.1	9
Zr	29.8	48.7	174.4	54.0	195.7	180.1	273	273	376	273
La/Yb	39.6	24.5	58.8	84.9	337	104.7	103.1	103.1	0.52	103.1
Eu/Lu*	1.98	1.53	0.68	1.55	0.52	0.52	0.35	0.35	0.33	0.35
K/Rb	290	265	271	349	324	376	272	272	94	272

	G/LB 2									
	G/LB 3									
	G/LB 3	G/LB 4	G/LB 5	G/LB 6	G/LB 7	G/LB 8	G/LB 9	G/LB 10	G/LB 11	G/LB 12
La	19.2	10.1	179.1	1028.1A	131.9A	131.9B	131.9C	131.9D	131.9E	131.9F
Ce	9.6	6.2	1.0	5	3.1	1.5	2.6	2.6	2.6	2.6
Pr	18	13.0	2.0	10	7.6	2.9	5.9	5.9	5.9	5.9
Nd	1.9	1.5	0.23	1	0.94	0.34	0.67	0.67	0.67	0.67
Sm	7.2	4.9	0.77	3.8	3.0	1.1	2.5	2.5	2.5	2.5
Eu	1.9	1.5	32	1	1.4	0.45	0.95	0.95	0.95	0.95
Gd	0.22	0.12	<0.05	0.42	<0.05	<0.05	0.05	0.05	0.05	0.05
Tb	1.7	1.3	0.27	0.21	<0.05	0.12	0.19	0.19	0.19	0.19
Dy	0.26	0.23	0.05	0.21	<0.05	0.24	0.43	0.43	0.43	0.43
Ho	1.5	1.1	0.19	1.2	1.4	0.24	0.07	0.07	0.07	0.07
Er	0.24	0.16	<0.05	0.24	0.78	<0.05	0.07	0.07	0.07	0.07
Tm	0.59	0.38	0.07	0.67	0.11	0.18	0.41	0.41	0.41	0.41
Yb	0.07	<0.05	<0.05	0.08	0.11	0.20	0.06	0.06	0.06	0.06
Lu	0.49	0.31	0.06	0.46	0.76	0.19	0.19	0.19	0.19	0.19
Y	0.06	<0.05	<0.05	0.06	0.11	0.06	0.06	0.06	0.06	0.06
Zr	8.5	5.0	1.0	6.7	9.7	3.1	5.5	5.5	5.5	5.5
La/Yb	43.8	30.8	6.2	25.1	21.7	34.6	15.9	15.9	15.9	15.9
Eu/Lu*	12.8	13.0	10.9	7.0	2.7	4.8	4.1	4.1	4.1	4.1
K/Rb	0.38	0.27	0.42	1.3	0.1	0.06	0.33	0.33	0.33	0.33
	137	115	100	181	91	73	94	94	94	94

Table 46, continued

II. BARREN AND RARE-ELEMENT PEGMATITES							
	Barran Exocontact Pegmatites	F22-1	F22-5D	Taylor Beryl Pegmatite 2	Fairservice Spodumene Pegmatite 1 (Bulk Composition)	Gullwing Lake Pegmatitic Granite	Tot Lake Pegmatite, Pollucite Zone
La	0.61	0.61	1.0	1.6	0.18	6.9	0.24
Ce	1.1	1.1	1.9	3.4	0.24	17	0.18
Pr	0.11	0.11	0.24	0.39	<0.05	2.4	<0.05
Nd	0.33	0.33	0.81	1.3	0.11	8	0.11
Sm	0.10	0.10	0.29	0.65	<0.05	2.7	<0.05
Eu	<0.05	<0.05	0.05	0.06	<0.05	0.06	<0.05
Gd	0.11	0.11	0.30	0.69	<0.05	2.2	<0.05
Tb	<0.05	<0.05	0.07	0.17	<0.05	0.36	<0.05
Dy	0.10	0.10	0.49	0.88	<0.05	1.7	<0.05
Ho	<0.05	<0.05	0.07	0.11	<0.05	0.82	<0.05
Er	<0.05	<0.05	0.23	0.33	<0.05	0.82	<0.05
Tm	<0.05	<0.05	<0.05	0.05	<0.05	0.13	<0.05
Yb	<0.05	<0.05	0.32	0.41	<0.05	0.94	<0.05
Lu	<0.05	<0.05	<0.05	0.06	<0.05	0.14	<0.05
Y	0.47	0.47	2.7	4.7	0.18	13	0.18
ΣREE	<2.8	<2.8	<5.9	10.1	<1.08	43.7	<1.2
La/YbN	-	-	-	-	>21.4	-	-
Eu/Eu*	0.52	0.52	-	0.28	-	0.08	-
K/Rb	37	37	74	24	8	19	5

III. CONTACT METASOMATIC SELVEDGES

	Ghost Lake Batholith Contact	Taylor Beryl Pegmatite 2	Fairservice Spodumene Pegmatites, North Zone	Gullwing Lake Spodumene Pegmatite	Coates MoS ₂ Pegmatite
La	14	22	24	314	28
Ce	34	16	30	940	29
Pr	4.7	-	43	350	914
Nd	22	8	23	403	13
Sm	6.6	2	8	135	5
Eu	1.4	1.0	2.6	2.3	1.5
Gd	6.7	3	8	96	9
Tb	1.3	-	-	39	114
Dy	8.1	2.1	3.5	39.5	3.5
Ho	1.7	-	-	17.3	49.4
Er	5	-	-	-	-
Tm	0.74	-	-	-	-
Yb	4.7	1.6	1.5	10.5	2.4
Lu	0.69	-	-	-	-
Y	40	9	14	156	17
ΣREE	111.6	55.7	122.9	1940.3	89.4
La/YbN	8.8	8.9	114.7	720.6	7.6
Eu/Eu*	2.55	1.24	9.9	13.2	0.69
			1.04	0.10	0.08

Figure 74. Chondrite-Normalized Rare Earth Element Curves Arranged in a West to East Section Across the Ghost Lake Batholith.

Figure 75. REE Chondrite Plots for Ghost Lake Batholith Units GLB-2 and -3.

5

MicroD

chondrite patterns ($\text{La/Yb}_N=24.5$ to 84.9). Two of these rocks contain appreciable cordierite (C15-5 and C20-3). Cordierite was not recognized in D18-2A, however the high similarity of its REE pattern to C15-5 and C20-3 suggests that it is also cordierite-bearing and should thus be included in GLB-1A.

Concomitant with decreasing RREE is a fundamental change in the shape of the chondrite-normalized patterns. This is revealed in a west-to-east profile across the batholith (Figure 74). The most chemically primitive rocks, in the west central part of the batholith (D14-1, D14-6B, D18-2A, D22-1A) generally exhibit relatively steep curves (La/Yb_N 49.8 to 337), which either have a small negative or positive Eu anomaly ($\text{Eu/Eu}^*=0.52$ to 1.55). This pattern, which changes to a flat to gently sloping one ($\text{La/Yb}_N=2.4$ to 10.9) in units of the eastern lobe of the batholith is primarily caused by pronounced depletion of LREE in GLB-4, -5 and -6 relative to GLB-1 and -2. Abundances of HREE are approximately the same.

A further characteristic is the pronounced negative Eu anomalies in plutonic units GLB -5 and -6 (Figure 45) where Eu/Eu^* ratios of <0.1 to 0.33 are the lowest in the batholith. These values correspond to the most fractionated units and generally the lowest ΣREE levels of 8.2 to 34.6 ppm (Table 46). The modest negative Eu anomalies seen in GLB-1 suggest plagioclase-controlled retention of Eu in the zone of magma generation. A relatively calcic plagioclase would be expected to persist in the restite along with sillimanite, biotite, cordierite and possibly garnet under conditions of partial melting of a

metasedimentary protolith (Winkler 1979) and such a segregation could have prevented most of the Eu from entering into the separating melt.

Explanation of the small positive europium anomalies in three samples of GLB-1 (C15-5, C20-3, and D18-2A) is more difficult. Most strongly peraluminous granitoid suites do not exhibit this type of Eu anomaly. Similar REE patterns and total REE contents however suggest a common origin for all the GLB-1 samples. It is possible that cordierite could retain Eu however no REE analyses of this mineral were performed on this mineral and no REE data are available in the literature.

GLB-2 samples display similar REE patterns to those of GLB-1 (Figures 74 and 75). Sample E19-5D, hosted in GLB-1 at the western end of the batholith, has a relatively steep pattern ($\text{La/Yb}=33.6$) and a small positive Eu anomaly. Sample D28-1B, however, which is situated near the most highly fractionated part of the GLB, has higher total REE (179.3 ppm), a flatter pattern ($\text{La/Yb}_N=106.1$) and a negative Eu anomaly.

GLB-3 exhibits flatter curves ($\text{La/Yb}=12.8$ to 13.0) than GLB-1 or -2, and greatly resembles GLB-6 from the Dryden Airport area (Figure 74), except for a less pronounced negative Eu anomaly.

Deeper levels of the Ghost Lake Batholith, potentially closer to the zone of peraluminous magma generation, are probably not exposed. However, peraluminous S-type granitoid rocks compatible with low pressure granulite metamorphic conditions occur widely within the metasedimentary migmatite-dominated English River Subprovince (Breaks

et al. 1985) and may represent an analogue of deeper levels of the Ghost Lake Batholith.

The REE pattern for garnet-cordierite-biotite granite from the Churchill Lake Batholith (one analysis in Breaks and Bond in press) which is situated 100 kms northeast of the Ghost Lake Batholith, exhibits a striking similarity to that of most GLB-1 samples, exclusive of the cordierite-bearing variants with positive Eu anomalies. The Churchill Lake Batholith, however, has a slightly higher Σ REE (289 ppm) than most samples of GLB-1.

Goad and Cerny (1981) studied the distribution of REE in the two predominant units of pegmatitic granite masses in the Winnipeg River pegmatite district of Manitoba. They documented that, in the case of units from the same intrusion, pegmatitic leucogranite almost invariably contains higher RREE than fine-grained leucogranite. The Σ REE patterns of both are, however, quite similar in shape and show very low La/Yb_N ratios. A larger, more conspicuous negative Eu anomaly is apparent in the fine grained leucogranite.

The most drastic depletion in Σ REE is shown by the potassic pegmatite unit of Goad and Cerny (ibid); this conclusion is tentative as data in the literature are considerably scarcer than for pegmatitic leucogranite or fine grained leucogranite (Goad and Cerny, ibid). These workers presented one analysis with a total of 15.8 ppm, for 8 REE in potassic pegmatite from the Greer Lake pegmatitic granite. These concentrations are slightly higher than those in barren, exocontact potassic pegmatite dykes related to the Ghost Lake batholith which contain 2.8 to 5.9 ppm Σ REE (Table 46). However, yet

more extreme depletion of Σ REE is recorded in pegmatitic leucogranite and fine-grained leucogranite from the Ghost Lake Batholith, relative to the Greer Lake intrusion (Figure 74 and Goad and Cerny 1981, p.182).

RARE-ELEMENT PEGMATITES AND CONTACT METASOMATIC SELVEDGES

As rare-element pegmatites are among the most REE-depleted igneous systems known in the Archean, it is pertinent to examine possible depositories for these elements. Very little consideration has been given in the literature towards ascertaining the ultimate fate of the REE, which either are apparently incompatible in rare-element pegmatite systems or have been subsequently removed during post-magmatic alteration. Muecke and Clarke (1981) recognized extreme depletion of REE in the peraluminous South Mountain Batholith complex of Nova Scotia. The ultimate destination of these elements, which are comparably depleted to REE in the Dryden Pegmatite Field, was suggested by these workers (Muecke and Clarke 1981, p.144) to be in "REE-rich mineral deposits yet to be found". In this section an alternative will be considered: contact metasomatic selvages.

REE abundance data for four rare-element pegmatites and eight contact metasomatic selvages are presented in Table 46 and Figures 76 and 77. Total REE contents are variable but considerably higher than in the associated rare-element pegmatites (compare Figures 76 and 77). Contact metasomatic selvages of the Gullwing Lake Pegmatite contain the highest total REE (721 to 2032 ppm) which represents a concentration factor of 16 to 46 times that of the associated muscovite-biotite pegmatitic granite unit (43.7 ppm total REE).

Figure 76. Chondrite-Normalized REE Curves for Rare-Element Pegmatites of Dryden Field.

Figure 77. Chondrite-Normalized REE Curves for Contact Metasomatic Selvedges from the Dryden Pegmatite Field.

Total REE contents are lower in selvages of the Mavis Lake Group (55.7 to 126.4 ppm), however much higher enrichment factors are generally apparent (5.5 to >117 times) because the associated rare-element pegmatites are more highly depleted than in the case of the Gullwing Lake pegmatitic granite.

Other niches for these mobilized REE may be present. These potentially include skarn assemblages in the amphibolitic mafic metavolcanic rocks surrounding pegmatites of the Mavis Lake Group and tourmaline-rich metasomatic segregations containing accessory biotite and apatite. No examination of REE distribution in these segregations was undertaken in the present study.

Experimental data strongly infer that REE are mobilized into an aqueous phase through fluoride complexing during late magmatic stages (Flynn and Burnham 1978, Alderton et al. 1980, Muecke and Clarke 1981). Muecke and Clarke (1981) found that extreme depletion of REE in quartz-phengite greisen zones was partially explained by moderate enrichment in associated fluorite veins. In the Dryden Field a relationship between REE mobilization and high fluorine is suggested in the contact metasomatic selvages. These rocks contain the highest fluorine contents in the Dryden Field (1010 to 18,400 ppm in Tables 41 and 42) which show a steep positive correlation with total REE.

In Figure 77 it may be seen that the Mavis Lake Group selvages have small or negligible positive or negative Eu anomalies (Eu/Eu^* 0.92-1.24) and lower La/Yb_N (7.4-9.9) than those for the Gullwing-Tot Lakes Group. The latter pronounced negative Eu anomalies; the Eu/Eu^* (0.06 - 0.69) ratio varies from moderate to extreme. Higher La/Yb_N

ratios (7.6 to 19.6) are generally evident. Where REE curves for pegmatites have been defined a similarity in REE variation pattern to the associated contact metasomatic selvage can be observed (Figures 76 and 77: Taylor and Gullwing Lake Pegmatites).

It is evident from the above data that the contact metasomatic selvages represent important "collectors" of REE mobilized during late stage magmatic-hydrothermal evolution of rare-element pegmatites.

An alternative explanation of the LREE depletion in the late units of the GLB could relate to early precipitation of accessory minerals such as monazite and allanite with very high solid/melt distribution coefficients. A sequence of chondrite patterns virtually identical to that of the GLB was described by Miller and Mittlefehldt (1982) for Late Cretaceous, strongly peraluminous granite plutons of southeastern California. These workers convincingly ascribed the flattening of REE curves, and deepening negative Eu anomaly with increasing fractionation, as resulting from the compatible behaviour of the LREE in relatively early granitic units of a comagmatic series which precipitated accessory monazite and allanite. The distribution coefficients for LREE between solid and liquid in this case were estimated by Miller and Mittlefehldt (ibid) to range from 1000 to 10,000. Trivial quantities of these minerals may thus profoundly influence REE patterns and, if overlooked due to very small grain size, can lead to incorrect petrogenetic interpretations (Miller and Mittlefehldt, ibid).

As monazite and allanite have not been recognized in the GLB plutonic units, and apatite is modally very low, fluoride complexing

of REE into an aqueous phase is the preferred explanation for the appreciable increase of these elements in contact metasomatic selvages, and the drastic depletion in the late plutonic units of the GLB.

**Origin and Evolution of Peraluminous Granite and Rare-Element
Pegmatite in the Dryden Area, Superior Province of Northwestern
Ontario**

by

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**A thesis submitted to the
Faculty of Graduate Studies and Research
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the degree of Doctor of Philosophy**

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Chapter Eight

MINERAL CHEMISTRY

INTRODUCTION

Chemical variation was examined in several minerals from the Ghost Lake Batholith, barren and peraluminous rare-element pegmatites and potassic plutonic suite rocks. Minerals analyzed are K-feldspar, muscovite, biotite, garnet, cordierite, beryl, spodumene and columbite-tantalite. This information has allowed a more definitive investigation into peraluminous granite-pegmatite fractionation than would have been possible from lithochemistry alone.

K-FELDSPAR

Extensive work by Gordiyenko (1983, 1976, 1971) and Cerny et al. (1981) has shown that variation of Cs, Li and Rb contents of K-feldspar can be used to divide the various mineralogical types of rare-element pegmatites and to distinguish these from those which are barren. To effect this discrimination for K-feldspar chemistry Gordiyenko (1971) developed the Na_2O vs K/Cs diagram and Cerny et al. (1981) showed that the K/Rb vs Cs diagram was also effective.

Concentrations of Na_2O (in most samples), K_2O , Ba, Cs, Li, Rb and Sr, determined in this study are reported in Appendices 10 to 28 for 213 K-feldspar samples from the Dryden Field. Analyses for CaO, Al_2O_3 , Ga and Pb in some samples are also included. This set includes 83 K-feldspar analyses uniformly covering the Ghost Lake Batholith and associated exocontact pegmatite dykes.

Ucakuwun (1981, p.74-80) determined obliquity (Δ) values of K-feldspars from most rare-element pegmatites and pegmatite granites in the Dryden area. His work revealed three species of K-feldspar:

1. microcline (maximum or near maximum ($\Delta \leq 0.90$) orthoclase
2. intermediate microcline ($\Delta \leq 0.90$) > orthoclase
3. orthoclase (<0.80) > intermediate microcline

The data of Ucakuwun (ibid) is summarized below:

	Average Δ	range	n
Dryden Airport pegmatitic granite	0.93	0.91 - 0.95	5
Taylor beryl pegmatite no. 1	0.95	0.91 - 1.01	4
Fairservice spodumene pegmatites			
North Zone	0.83	0.71 - 0.96	10
South Zone	0.88	0.56 - 0.96	10
Mica Point pegmatite granite	0.965	0.93 - 1.00	2
Tot Lake spodumene-pollucite pegmatite	0.95	0.84 - 0.98	5

Structural state determination of the analyzed K-feldspar samples was not undertaken as the writer was primarily concerned with revealing fractionation patterns by examining the variation of the trace elements and the inter-element ratios. It is not possible to generally apply the structural data of Ucakuwun (1961 p.74-80) to the K-feldspars sampled in this study even for those determinations from the same rare-element pegmatite body as several of these contain more than one species (e.g. orthoclase, intermediate microcline and maximum microcline at Tot Lake).

K-feldspar is the main accumulator of the rare alkalis Rb and Cs in most granitoid rocks; micas are also important where they are abundant. In granitic plutonic rocks concentrations of Rb and Cs in K-feldspar increase and the ratio of K to these elements have been shown by McCarthy and Hasty (1976) to diminish with degree of

fractional crystallization. Ba and Sr, on the other hand, tend to diminish in concentration in the fractional crystallization modelling of McCarthy and Hasty (ibid).

GHOST LAKE BATHOLITH AND EXOCONTACT PEGMATITES

Relatively high pressure, deep level, poorly fractionated pegmatites (mica-bearing pegmatites of deep levels, Ginzburg et al. 1979) favour accumulation of small radius cations such as Sr, Ba and Pb in K-feldspar reflected by low K/Ba and Rb/Sr, and high K/Rb, Ba/Rb and K/Cs ratios. These pegmatites, suggested by Ginzburg et al. (1979) which crystallize at depths between 7 and 11 km have relatively little potential for hosting highly fractionated rare-element mineralization. In the Dryden area, such pegmatites may correspond to:

1. internal pegmatites within the north-western part of the Ghost Lake Batholith;
2. exocontact garnet-biotite-(muscovite) pegmatites situated immediately west to northwest of the Ghost Lake Batholith; and,
3. pegmatites associated with potassic plutonic suite masses such as the Gullwing Lake Batholith (see ensuing section).

The data given in Appendices 10 to 12, also presented in a hand-contoured map (Figure 78, back pocket) reveals fractionation trends in the coarse-grained to pegmatitic units of the Ghost Lake Batholith. Four distinct chemical domains are evident:

1. poorly fractionated northwestern main batholith mass (GLB-1; Appendix 10),
2. moderately fractionated southwestern batholith lobe (GLB-3; Appendix 11),

3. moderately fractionated part of eastern batholith lobe involving unit GLB-4 (Appendix 12) and
4. highly fractionated part of eastern batholith lobe (GLB-7; Appendix 12)

Average Cs, Li and Rb contents and ratios K/Ba/Sr and Rb/Sr increase and Ba and Sr contents and ratios K/Cs, K/Rb and Ba/Rb decrease in the above sequence. The most dramatic change occurs in the Ba/Rb ratio which decreases from 6.2 to 0.006 within the Ghost Lake Batholith intrusive sequence. Taylor and Heier (1958) originally proposed on the basis of work undertaken in Precambrian basement rocks of South Norway, that the Ba/Rb ratio of K-feldspars should be a sensitive indicator of magmatic fractionation. However, because of the erratic nature of this variable (some pegmatites show Ba enrichment; Appendix 10) K/Rb ratios were chosen as a more reliable indicator of fractionation in the Mavis Lake pegmatite group (Figures 78 and 79).

Lithium concentration in most K-feldspars of the Ghost Lake Batholith is limited and generally below 10 ppm. Variation of lithium however does follow the observed fractionated trends in the batholith rising to a maximum of 74 ppm in GLB-3 and to 65 pm close to the Dryden Airport beryl occurrence.

In addition to the pluton-scale variation of the above elements and ratios, systematic chemical trends can also be observed in K-feldspars from smaller areas containing GLB-1, -4 and -7 such as locality E21-4 (Figure 80 and Appendices 10, 12). The Rb content of K-feldspar from GLB-6 at this locality (770 ppm) is less than one-half the average of the main occurrence of potassic pegmatite (GLB-6 and

Figure 79. Fractionation Trends within the Eastern Lobe of the Ghost Lake Batholith, Zealand Stock and Related Exocontact Pegmatites Based upon the K/Rb Ratio of K-feldspar.

Figure 80. Variation in Ba, Rb, Sr, Ba/Rb, K/Ba and K/Rb of K-feldspars from Ghost Lake Batholith Units GLB-1, -4 and -6.

-7), 14.5 km to the east in the vicinity of the Dryden Airport. Ba and Sr contents are, however, comparable.

GLB-3 is, next to GLB-7, the most fractionated plutonic unit of the Ghost Lake Batholith that was sampled. It should be noted however that K-feldspars from GLB-5 were not analyzed.

Exocontact Pegmatites of the Ghost Lake Batholith

Data on K-feldspars from exocontact muscovite and muscovite-biotite pegmatites are presented in Appendices 13 to 15, divided according to metamorphic grade of the host rocks and locality. K-feldspars from pegmatites emplaced in low to medium grade supracrustal rocks are more chemically fractionated than their counterparts from the high grade metamorphic zones situated north and northwest of the Ghost Lake Batholith. This trend is especially evident in the high Rb and Cs contents and Rb/Sr ratio, and low K/Rb and K/Cs ratios, seen in the former. These are the most extreme of any K-feldspar in the Dryden area, exclusive of the Mavis Lake pegmatite group.

MAVIS LAKE PEGMATITE GROUP

Relative to the higher pressure, deeper level, poorly fractionated mica-bearing class of pegmatites moderately to highly fractionated, lower pressure pegmatites of the intermediate level of Ginsburg et al. (1979; 3.5 km to 7 km depth) are characterized by K-feldspars with higher concentrations of Rb and Cs in K-feldspars and ratios Rb/Sr and K/Ba and lower in Ba, Sr and ratios K/Cs, K/Rb and Ba/Rb (Shmakin 1971). Pegmatites of this category are thought to include rare-element pegmatites such as in the Dryden Field and some internal pegmatites of the Ghost Lake Batholith (GLB-7). Rb and Cs

contents of these internal pegmatitic granite units respectively average 1678 and 341 ppm (Appendix 12). Levels of these elements are highest where beryl mineralization is involved, as the Dryden Airport Beryl Occurrence (Rb=1813 ppm, Cs=935 ppm).

Exocontact pegmatites containing rare-element mineralization exhibit much higher average Rb contents (1980 - 9680 ppm), increase systematically in the Mavis Lake Group with increasing distance from the Ghost Lake Batholith contact, as indicated by the decreasing K/Rb ratio (Figure 79; Table 47, Appendices 16 to 19).

Cesium also exhibits significant variation, ranging from 62 to 1150 ppm in K-feldspar from the Mavis Lake Group. Its variation, however, is not as systematic as that of Rb. Figure 81 indicates that although K-feldspars from some beryl-bearing rare-element pegmatites (Dryden Airport and Taylor Pegmatite 2) are poorly fractionated relative to the Fairservice spodumene pegmatites in terms of K/Rb ratios, the former contain exceptionally high Cs levels. Three analyses exceed 1000 ppm which are slightly higher than the maximum Cs content for the Fairservice Property K-feldspars. Cesium correlates negatively with K/Rb in the Mavis Lake pegmatite group as three distinct, en-echelon data clusters, all of which are notably removed from the lower cesium GLB field. Only the Taylor beryl pegmatite No. 1 overlaps with the GLB.

Ba-Rb variation in the GLB exhibits a well developed inverse fractionation trend very similar to the petrochemical one (compare Figures 62 and 82) which shows rapidly diminishing Ba (7780 to <10 ppm) against a more modest increase in Rb (190 to 2200 ppm). The diagram is marked by a series of overlapping GLB unit fields Table

Table 47. Average Ba, Cs, Li, Rb, and Sr and Selected Element Ratios of K-feldspar from Mavis Lake Rare-Element Pegmatite Group and Ghost Lake Batholith. Averages Calculated from Analyses in Appendices 10 to 19.

I. Ghost Lake Batholith													
	Ba	Cs	Li	Rb	Sr	Ba/Rb	K/Ba	K/Cs	K/Rb	Rb/Cs	Rb/Sr	N	
1. GLB-1	1655	<10	<4	290	360	6.2	1328	>14000	384	>38	1.2	24	
2. GLB-3	124	<24	23	719	38	0.22	1273	>4844	148	35	>33	13	
3. GLB-4	304	14	<7	596	87	0.60	1314	>8858	186	>54	14	13	
4. GLB-7	29	341	35	1678	33	0.021	4533	2119	69	>53	61	13	
5. Exocontact Pegmatites in High Grade Metasedimentary Migmatites	2180	<7	<4	255	342	9.4	96	>16265	410	-	1.1	10	
6. Exocontact Pegmatites in Medium Grade Metasediments	78	55	<10	1204	30	0.09	2782	3054	109	30	63	9	
7. Exocontact Pegmatites near northeast contact of Ghost Lake Batholith	643	9	<4	527	220	1.4	199	9218	224	42	3.4	3	
II. Dryden Airport Beryl Pegmatite	25	935	46	1813	40	0.014	4648	121	59	2.0	60	4	
III. Taylor Beryl Occurrence													
1. Pegmatite 1	83	62	34	1980	46	0.042	1412	1878	55	35	44	6	
2. Pegmatite 2	112	480	20	2958	59	0.12	1057	299	36	8	51	5	
IV. Fairservice Spodumene Pegmatites													
1. Pegmatite 1	30	135	171	6955	80	0.004	4228	713	16	70	97	8	
2. Pegmatite 2	30	183	130	9661	157	0.004	4043	726	15	48	49	7	
3. Pegmatite 3	70	215	43	6835	168	0.011	1945	618	17	36	41	2	
4. Pegmatite 4	31	244	114	8497	185	0.004	4095	551	13	40	46	7	
5. Pegmatite 5	30	295	130	9680	208	0.003	3846	448	12	37	47	2	
6. Pegmatite 6	38	260	17	9110	202	0.004	3101	447	12	38	45	5	
7. Pegmatite 9	30	570	319	9335	226	0.003	3826	286	13	20	43	4	
8. Pegmatite 10	20	520	7	8540	150	0.002	5520	212	13	16	57	2	
9. Pegmatite 18	418	225	16	5835	360	0.07	272	500	19	26	14	4	

N = number of analyses

Figure 81. K/Rb vs. Cs (ppm) for K-feldspar from Ghost Lake Batholith and Related Exocontact Pegmatites, Zealand Stock and Mavis Lake Rare-Element Pegmatite Group.

Figure 82. Variation of Ba and Rb (ppm) in K-feldspar from Ghost Lake Batholith, Zealand Stock and Mavis Lake Rare-Element Pegmatite Group.

suggesting the fractionation sequence GLB1-(3-4)-(5,6,7). Rare-element pegmatites of the Mavis Lake Group are detached from the GLB trend towards higher Rb levels. The most notable feature however is the chemical hiatus between the GLB, the Taylor beryl pegmatite No. 2 and the spodumene pegmatites near Mavis Lake (Fairservice property). This discontinuity is also apparent in the muscovite chemistry (discussed later in the chapter); petrogenetic interpretations of this behaviour of Ba, Rb and Sr in K-feldspar and muscovite will be discussed in Chapter 9.

Ahrens and Liebenberg (1945) stated that a Li content exceeding 5 ppm in feldspar is exceptional. The maximum content of Li documented in K-feldspar (390 ppm) was reported in pegmatite from the Kola Peninsula by Borovik-Romanova and Kalita (1958) and Borovik-Romanova and Sosedko (1958). This study has revealed a maximum of 865 ppm Li in a rubidium-rich K-feldspar from Pegmatite 9, in the South Zone of the Fairservice spodumene pegmatites (Appendix 18). This high Li concentration may reflect accessory levels of a mineral such as lepidolite or lithian muscovite which was not detected during purification of the K-feldspar sample.

It is apparent that definite correlation exists in the study area of increased levels of Li in K-feldspar with presence of Li-bearing minerals such as spodumene. In the rare-element pegmatites of the Mavis Lake Group, a range of 26 to 865 ppm is evident for North and South Zones of the Fairservice property and more than one-half of the analyzed K-feldspar contain between 114 and 319 ppm Li (Appendices 17 and 18).

Rb-Sr VARIATION IN K-FELDSPAR FROM THE GHOST LAKE BATHOLITH AND MAVIS LAKE PEGMATITE GROUP

Within the Ghost Lake Batholith Rb/Sr ratios increase significantly from less than unity, within the northern poorly fractionated portion (GLB-1), to values exceeding 77 in the southwest lobe and towards the eastern batholith terminus (maximum Rb/Sr = 68) as shown in Figure 83. Rb/Sr ratios reach even greater levels for several exocontact pegmatites between Dryden and Thunder Lake (maximum Rb/Sr = 179).

A major inflection in the regional variation curve occurs in GLB-7 near the Dryden Airport, where pegmatite K-feldspar have a trend of gradually increasing Sr (40-120 ppm) with rapidly advancing Rb levels (1940 - 10,500 ppm) which correlates with increasing distance from the field of GLB-7 (Figure 83).

In many respects the Rb-Sr variation in K-feldspars from the GLB and adjacent Mavis Lake pegmatite group is very similar to that of Ba-Rb (Figures 86 and 83); chemical discontinuities exist between the spodumene pegmatites on the Fairservice Property and the GLB trend.

A possible cause of the observed inflection in the Rb-Sr trend in K-feldspars from the Mavis Lake group may be related to an increase in radiogenic ^{87}Sr in pegmatites which contain high Rb levels (Clarke and Cerny 1987). K-feldspars from the Fairservice Property spodumene pegmatites contain 0.5 to one percent Rb; these pegmatites exhibit the highest Sr levels (50 to 500 ppm in Figure 83).

POTASSIC PLUTONIC SUITE PEGMATITES

In the study area pegmatites of this association are probably mainly derived from the Gullwing Lake Batholith and are thus largely

Figure 83. Variation of Rb and Sr (ppm) in K-feldspar from Ghost Lake Batholith, Zealand Stock and Mavis Lake Rare-Element Pegmatite Group.

situated to the north and northeast of the Ghost Lake Batholith. These pegmatites are poorly fractionated as revealed by high Ba, Sr, K/Rb, K/Cs and Ba/Rb, and low Rb, Cs, K/Ba and Rb/Sr in K-feldspar (Appendix 20; Figure 79, back pocket). K-feldspar chemical trends of these pegmatites compare closely with the most poorly fractionated part of the Ghost Lake Batholith (Appendix 10).

GULLWING LAKE-TOT LAKE PEGMATITE GROUP

The fractionation trend in this pegmatite group, based upon K/Rb ratios in K-feldspar, is more ambiguous than that of the Mavis Lake group (Figures 79 and 84; Table 48; Appendices 21 to 28).

Fractionation appears to culminate in the western part of the Gullwing-Tot Lakes group and increases towards the Gullwing Lake pegmatite granite from three different directions (Figure 84). A further fractionation trend is suggested in the middle to eastern part of the pegmatite group where K/Rb ratios decrease in an easterly direction from 32 (Coates pegmatite) to 9 at the Tot Lake pegmatite. This trend is more uncertain on account of the lack of exposed pegmatites in the intervening, drift-covered area.

Rare-element pegmatites of the Gullwing-Tot Lakes group define an orderly progression of increasing cesium contents in K-feldspar as a function of decreasing K/Rb ratio (Figure 85). Pegmatitic granite at Mica Point and in the Coates Bay Stock with little or no rare-element mineralization is poorly fractionated. Cesium contents are generally below 100 ppm and the K/Rb ratio exceeds 70 (Figure 85). These pegmatite granites are succeeded, in order of increasing Cs and decreasing K/Rb, by the blocky K-feldspar-quartz zone of the Gullwing Lake pegmatite, the spodumene-quartz core of the same pegmatite and

Figure 84. Fractionation Trends in Gullwing-Tot Lakes Rare-Element Pegmatite Group Based upon K/Rb Ratios in K-feldspar. Geological Units are the same as in Figure 20.

Table 48. Average Ba, Cs, Li, Rb and Sr and Selected Element Ratios of K-feldspar from Rare-Element Pegmatites and Pegmatitic Granites of the Gullwing-Tot Lakes Group. Averages Calculated from Analyses in Appendices 21 to 28 and Unpublished Data of the Author.

	<u>Ba</u>	<u>Cs</u>	<u>Li</u>	<u>Rb</u>	<u>Sr</u>	<u>Ba/Rb</u>	<u>K/Ba</u>	<u>K/Cs</u>	<u>K/Rb</u>	<u>Rb/Cs</u>	<u>Rb/Sr</u>	<u>n</u>
Mica Point Pegmatitic Granite	82	63	11	1143	46	0.075	2094	2165	93	27	26	6
Gullwing Lake Pegmatitic Granite												
Spodumene Unit, North Zone	295	515	18	6840	80	0.043	412	308	18	16	88	12
Blocky K-feldspar-Quartz Unit, North Zone	136	69	17	2416	38	0.042	825	1670	44	38	73	13
Spodumene Unit, Central Zone	1674	250	22	3992	48	0.54	190	823	31	25	83	5
Blocky K-feldspar-Quartz Core, South Zone	193	80	3	2613	43	0.08	586	1376	43	33	66	3
Coates Molybdenite Pegmatite	80	93	12	3603	57	0.023	1424	1611	32	49	64	3
Tot Lake Pegmatite	36	1282	53	10800	157	0.003	3687	88	9	9	70	19
Coates Bay Pegmatitic Granite	<40	37	23	1103	<16	<0.37	4917	3042	103	31	>88	3
Pegmatitic Granite Dyke 230 m NE of Coates Bay Pluton	85	45	7	1875	15	0.046	1372	2667	58	45	130	2

n = number of analyses

Figure 85 K/Rb vs. Cs (ppm) Variation in K-feldspar from Pegmatitic Granites and Rare-Element Pegmatites of Gullwing-Tot Lakes Group.

the Tot Lake pegmatite. The blocky K-feldspar-quartz zone of the Coates pegmatite lies slightly off this trend (Figure 85, Appendices 22-27). The lower end of the Gullwing Lake-Tot Lake Pegmatite Field closely parallels the Rush Lake Li-Rb-Cs-Be-Ta>Nb-Sn pegmatite field (Cerny et al. 1981).

The Tot Lake pegmatite thus represents the most chemically fractionated rare-element pegmatite in the Dryden Pegmatite Field, with maximum levels of Rb (11,700 ppm) and Cs (2410 ppm) in K-feldspar. This pegmatite shares some chemical characteristics with the Bernic Lake pegmatite group, the most chemically fractionated pegmatites of southeastern Manitoba, as indicated by overlap of respective fields in the K/Rb vs Cs diagram in Figure 86.

Cesium contents of the Tot Lake pegmatites are very similar to the 750 to 3000 ppm range of the Tanco Pegmatite in the Bernic Lake Group (Cerny et al. 1981), the only disparity being the slightly higher K/Rb ratios at Tot Lake (7.9 - 10.3). Nevertheless, presence of a rare-element pegmatite approaching the degree of chemical fractionation similar to the Tanco Pegmatite, an important producer of Ta, Li and Cs lends significant economic potential to the Gullwing-Tot Lakes group.

In terms of K/Rb, K/Cs, Rb/Cs and Rb/Sr ratios, the spodumene core in the North Zone of Gullwing Lake Pegmatite represents the second most chemically fractionated part of the Gullwing-Tot Lakes group. Other average ratios in Table 48 are less useful as a guide to fractionation in the pegmatite group.

Rb-Sr variation in K-feldspar from this pegmatite group is quite similar to that of the Mavis Lake group (compare Figures 83 and 86).

Figure 86. Variation of Rb and Sr (ppm) in K-feldspar from Pegmatitic Granites and Rare-Element Pegmatites of Gullwing-Tot Lakes Group.

Possible parental granitic plutons are not extensively exposed in the vicinity, which may imply why evidence of a negatively sloping Rb-Sr variation curve analogous to that of the Ghost Lake Batholith is largely missing. Only the Mica Point pegmatitic granite has a trend similar to the GLB although the range in Rb/Sr ratios is much more restricted.

Gullwing Lake Pegmatitic Granite

A more detailed examination of K-feldspar chemistry undertaken on this pegmatitic granite than for any other in the Gullwing-Tot Lakes group, because of the complexity of this granitic body and the significant vertical exposure which allowed assessment of pegmatite fractionation with increasing height.

Vertical and lateral variation of selected elements and ratios in blocky K-feldspar from the North Zone are presented in Figures 87 and 88 and Appendix 23. The lateral chemical profile (Figure 87) indicates very little occurs across the pegmatite; most of the profile involves the blocky quartz-K-feldspar zone.

Vertical variation of Cs and Rb in blocky K-feldspar exhibits a more noticeable trend of increased fractionation with depth (Figure 88, Appendix 23) as indicated by an increase in Cs, corresponding to a considerable decrease in K/Cs ratio (2292 to 183). Rb and K/Rb values also reflect this trend (see also Appendix 23). Sample KF-10 is the most fractionated blocky K-feldspar from the Gullwing Lake Pegmatite, exhibiting values of many of the chemical indicator values of Makagon (1974) that supposedly indicate the presence of pollucite in complex pegmatites.

Figure 87. Lateral Variation of Ba, Cs, Li, Rb, and Ratios Ba/Rb, K/Cs, K/Ba, K/Rb and Rb/Cs in Blocky K-feldspar from North Zone of Gullwing Lake Pegmatitic Granite.

Figure 88. Vertical Variation of Ba, Cs, Li and Rb and Ratios Ba/Rb, K/Cs, and K/Rb in Blocky K-feldspar in North Zone of Gullwing Lake Pegmatitic Granite.

Figure 89. Vertical Variation of Ba, Cs, Li and Rb and Ratios Ba/Rb, K/Ba, K/Cs, K/Rb and Rb/Cs in Blocky K-feldspar from Spodumene-K-feldspar-Muscovite-Albite-Quartz Core at Central Zone of Gullwing Lake Pegmatitic Granite.

Fractionation in the Central Zone exhibits a similar pattern to that of the North Zone. Levels of Cs and Rb in blocky K-feldspar (Appendix 24) progressively increase with depth, corresponding to a decrease in the ratios K/Cs, K/Rb and Rb/Cs. These trends are portrayed in Figure 89. The drop of Ba content in K-feldspar is more notable in the central zone as revealed in a considerable decrease (Figure 90) of Ba/Rb ratio.

The South Zone is the least fractionated of the three known columbite-bearing parts of the Gullwing Lake pegmatitic granite. A lower extent of fractionation is revealed by lack of spodumene, virtual lack of platy albite-green muscovite replacement assemblages, and relatively low Rb and Cs contents of blocky K-feldspar (Appendix 26). The only rare-element pegmatites with comparable K/Rb ratios and Cs values to the Gullwing Lake pegmatitic granite are some spodumene pegmatites from the Fairservice Property (Figure 81).

Lithium contents of K-feldspar are generally lower in the Gullwing-Tot Lakes pegmatite group than in the Mavis Lake Group. This element is usually below 30 ppm and attains its highest concentration of 220 ppm and average (53 ppm) in the Tot Lake Pegmatite (Appendix 28).

MUSCOVITE AND BIOTITE

INTRODUCTION

Muscovite and biotite are important hosts for most rare elements and associated lithophile elements which, in some instances, may exhibit toward similarly high concentrations in co-existing rare-element pegmatites and parental granites. In particular, Ta, Nb, Sn,

W, Li, Be, Rb, Cs and F may attain high abundances in muscovite in magmatic-contact metasomatic environments (Beus et al. 1968) which in some cases correlate with the chemical character of the observed rare-element mineralization.

CHEMICAL CLASSIFICATION OF MUSCOVITE AND BIOTITE

Sixty muscovite and eleven biotite analyses from the study area are presented in Appendices 29 to 37. Of these, thirty analyses are relatively complete and are accompanied by structural formulae. In Figure 90 the complete mica analyses have been plotted in the $(Al^{3+}, Fe^{3+}, Ti^{3+})-(Fe^{2+}, Mn^{2+})-Mg^{2+}$ and $Li^{+}-(Al^{3+}, Fe^{3+}, Ti^{3+})-(Fe^{2+}, Mg^{2+}, Mn^{2+})$ diagrams of Foster (1960a,b). Muscovites from the Gullwing Lake pegmatitic granite contain the highest phengite component, and exhibit normal to the $Li-(Fe^{2+}+Mn+Mg)$ sideline, which almost passes into the lithium ferrian muscovite field of Sun Shihua (1982). All other muscovites plot closer to the $Al+Fe^{3+}+Ti$ apex.

None of the analyzed muscovites can be classified as lithian muscovite according to the concentration range criterion of 9,000 to 30,000 ppm Li suggested by Luecke (1980). This is further demonstrated in Figure 93 where it can be observed that the study area muscovites do not fall within the Tanco lithian muscovite field (Cerny et al. 1981).

Biotite from the Gullwing Lake pegmatitic granite is characterized by high Li, F and Al^{VI} relative to biotites from metaluminous granitic suites (Dodge 1969). On the basis of Figure 90 this biotite can be classified as a lithium-rich siderophyllite, which plots close to the protolithionite field.

Figure 90. Chemical diagrams of micas from Dryden Pegmatite in the $(Al^{3+}, Fe^{3+}, Ti)-(Fe^{2+}, Mn)-Mg$ and $Li-(Al^{3+}, Fe^{3+}, Ti)-(Fe^{2+}, Mn)$ Diagrams after Foster (1960a,b) and Sun Shihua (19820).

MUSCOVITE

As in the case of K-feldspar, the variation in trace elements and interelement ratios suggests a progression of fractionation from potential parental granitic masses into adjacent rare-element pegmatites.

Ghost Lake Batholith

Muscovite analyses from GLB-1, -3, -4 and -7 with related exocontact pegmatite dykes are presented in Appendices 29 and 30 and in Table 49 and Figure 91 (back pocket). Muscovite analyses are not as numerous or uniformly distributed as K-feldspar data within the Ghost Lake Batholith; muscovite is uncommon in most of GLB-1, the most extensive plutonic unit. The data indicate a dual trend of increasing chemical fractionation in a southwesterly and easterly direction, as particularly expressed by increasing Cs, Li, Rb and Sn and decreasing K/Rb. These trends corroborate the petrochemical and K-feldspar chemical trends.

Average concentrations of Be, Rb, W and Nb increase in the sequence GLB 1-4-3-7 as earlier inferred on basis of K-feldspar chemistry. Cs, Li and Sn trends are unclear, although average values of these metals in GLB-4 are more fractionated. An overall significant difference can, nevertheless, be noted between GLB-1 and GLB-7 in Cs (70 vs 228 ppm), Li (47 vs 649 ppm), and Sn (70 vs 317 ppm).

Tungsten in muscovite exhibits a moderate increase from GLB-1 in the sequence GLB 1-4-3-7 where average values rise from 160 ppm to 225 ppm (Table 49). Although data on tungsten distribution in micas are rather scanty in the literature, concentrations of 5 to 50 ppm are

Table 49. Average Be, Cs, Li, Nb, Rb, Sn, Ta and W and K/Cs, Rb/Cs and Nb/Ta for Muscovite from the Dryden Pegmatite Field.

	Be	Cs	Li	Nb	Rb	Sn	Ta	W	Rb/Cs	K/Rb	K/Cs	Nb/Ta
I. MAVIS LAKE GROUP												
Ghost Lake Batholith												
GLB-1	6.5	70	47	115	630	70	60	160	9.4	136	1212	2.0
GLB-3	13	162	222	130	1199	125	<72	208	21	70	1764	>2.5
GLB-4	7	39	320	99	990	190	<38	173	39	72	3314	>5.2
GLB-7	14.5	228	649	149	1965	317	<55	225	17	47	335	>5.2
Exocontact Pegmatite Dykes	13	109	518	145	1700	114	<56	155	22	57	1182	>3.6
Taylor Beryl Pegmatite #2	19	480	443	76	3370	794	100	160	7	25	177	0.8
Fairservice Spodumene Pegmatites												
Pegmatite 1	19	131	540	109	7063	729	69	149	55	12	629	1.9
Pegmatite 6	19	280	610	80	9420	734	90	155	34	8	283	0.9
II. GULLWING-TOT LAKES GROUP												
Gullwing Lake Pegmatite:												
North Zone	12	335	2470	190	5320	227	55	142	18	16	305	3.8
Central Zone	--	378	2060	374	6806	476	73	--	20	17	275	5.3
South Zone	10.3	231	1974	410	4344	251	46	--	27	21.5	589	10.3
Drope Township Columbite Pegmatite	--	44	--	160	1490	40	71	--	34	60	2037	2.3
Columbite Pegmatite near Coates Pegmatite	--	340	660	370	7360	80	72	--	22	12	251	5.1
Tot Lake Pegmatite	17	1093	518	55	13100	177	30	155	15	6.2	74	1.8

n= number of samples

considered common (Evans and Krauskopf 1974). Ivanova (1969) documented tungsten levels in muscovite between 14 and 150 ppm from two-mica granites and 41-120 ppm from granite in a tungsten ore zone from East Transbaikal, U.S.S.R.

In the light of Ivanova's study, it is apparent that W is anomalously enriched in some muscovite of the Ghost Lake batholith; the maximum level of 300 ppm approaches the extreme value of 510 ppm reported by her from greisen.

The highest W levels in muscovite are attained situated near the extreme southwest end of the Ghost Lake Batholith (270 and 30 ppm) and in the vicinity of the Dryden Airport (225-235 ppm). Several exocontact pegmatite dykes in the latter area exhibit even higher W levels (235-265 ppm) as shown in Figure 91. The mafic metavolcanics immediately adjacent to the southern contact of the Ghost Lake Batholith in Zealand Township contain scheelite-tourmaline mineralization at the Petrunka Occurrence.

Sn contents of muscovite exhibit strong enrichment within the eastern lobe of the Ghost Lake Batholith, reaching a maximum level of 684 ppm in potassic pegmatite at locality E31-2 near the Dryden Airport. Even higher Sn levels are contained in muscovites of nearby rare-element pegmatites of the Mavis Lake group (639 to 870 ppm; Appendices 29 and 30) providing further evidence of genetic connection between the Ghost Lake Batholith and this rare-element pegmatite group.

Tantalum enrichment trends in the GLB are more difficult to assess as some of the data fall below the 30 ppm sensitivity limit. Nevertheless, several muscovite specimens from this batholith (GLB-3, -4 and -7) and exocontact pegmatites are anomalously high in Ta and Nb/Ta ratios in muscovite from a granitic complex (Appendices 29 and 30). These values significantly exceed the 14.2 ppm average Ta content and average Nb/Ta ratio (5.2) of 1249 muscovite analyses of muscovite from magmatic rocks (Florke et al. 1978b), and serve to further emphasize the rare element character of such GLB plutonic units.

Mavis Lake and Gullwing-Tot Lakes Pegmatite Groups

Chemistry of primary muscovite reveals some significant differences in element associations developed in these two rare-element pegmatite groups, which broadly correlate with the observed mineral assemblages (Table 2). A general trend of higher average Cs, F, Li, Zn and Nb/Ta and lower B, Rb, Sn and Sr is apparent in the major pegmatites of the Gullwing-Tot Lakes Group relative to those of the Mavis Lake Group (Tables 33-36). Of all the rare elements in muscovite, lithium exhibits the greatest variation. The highest lithium concentrations (1080 to 3280 ppm) occur in muscovite from the Gullwing Lake pegmatitic granite. On the other hand, relatively low Li contents (475 to 560 ppm) characterize muscovite from the Tot Lake pegmatite. This is surprising in view of the extremely fractionated nature of the Tot Lake pegmatite as shown by its very low K/Rb and K/Cs ratios in K-feldspar and by the presence of pollucite, an indicator of extreme chemical fractionation (Cerny et al. 1985). Although the Tot Lake muscovite compare closely in many ways with

those of the Tanco Pegmatite (Cerny et al. 1981) as apparent in the Cs vs K/Rb plot of Figure 92, a marked disparity in the two fields is obvious in the Li vs K/Rb plot (Figure 93). The average Li contents of most muscovite groups in Table 49 exceed that of Tot Lake. This may reflect crystallization of muscovite in the Tot Lake pegmatite from a melt already strongly depleted in lithium due to prior crystallization of abundant spodumene.

Cesium variation is also significant and is best portrayed in the Cs vs K/Rb diagram of Cerny et al. (1985; Figure 92). Cesium abundances exhibit a greater range in the Gullwing-Tot Lake group (100-1130 ppm in Appendices 34 to 36) than in the Mavis Lake Group. The high Cs concentrations in Tot Lake muscovite closely compare with those from the Tanco pegmatite (Figure 92) and are consistent with the presence of pollucite in both pegmatites.

Cesium levels in muscovite from the Gullwing Lake pegmatitic granite are lower (110 to 630 ppm) than the Tot Lake Pegmatite and suggest on Figure 92 increasing fractionation in the order: South zone-North zone-Central zone.

Cesium in muscovite from the Mavis Lake pegmatite group are generally lower than those from the Gullwing-Tot Lakes Group. Those from spodumene pegmatites on the Fairservice property (Appendix 33) exhibit a tight clustering for Pegmatite 1 (110 to 160 ppm Cs). One sample from Pegmatite 6 is more fractionated (280 ppm; K/Rb = 8.4), but data are insufficient to infer a fractionation trend in spodumene pegmatites from the Fairservice property.

Figure 92. K/Rb vs. Cs (ppm) Variation for Micas from the Dryden Pegmatite Field. Cat Lake-Winnipeg River Pegmatite Field Data from Cerny et al. (1981).

Figure 93. K/Rb vs. Li (ppm) Variation for Micas from Dryden Pegmatite Field. Cat Lake-Winnipeg River Pegmatite Data from Cerny et al. (1981).

The Taylor beryl pegmatites do not reveal any meaningful trend either. Data of Ucakuwun (1981), probably from Taylor Pegmatite No. 1, reveal the most primitive muscovite in the Dryden Field (Cs = 19 ppm; K/Rb = 97 to 110). These muscovites are surpassed in degree of fractionation by most of those of the GLB (Figure 92). In contrast, muscovite from Taylor beryl Pegmatite No. 2 contain considerably higher Cs (390 to 570 ppm) and Rb (3290 to 3450 ppm) contents exceed most of those from the GLB (Appendices 29, 30 and 32) and other rare-element pegmatites from the Mavis Lake and Gullwing-Tot Lakes groups. Higher Zn contents in muscovite from the Gullwing-Tot Lakes group (260-1760 ppm) correlate with the sporadic appearance of sphalerite whereas this mineral has, to date, not been observed in rare-element pegmatites of the Mavis Lake group. Zinc contents of muscovite in the latter are lower, ranging between 67 and 260 ppm.

The volatile elements boron and fluorine show significant variation in muscovite from the Mavis Lake group, and may reflect basic differences in composition of the volatile phase associated with rare-element pegmatites in the two groups. Muscovite from the Mavis Lake group have boron contents between 155 and 285 ppm as opposed to the 21-44 ppm range in the Gullwing-Tot Lakes group (Appendices 34 to 36). Fluorine, in contrast is relatively enriched in the latter, averaging 8380 ppm to the Gullwing Lake pegmatitic granite and reaching a maximum of 1.42 percent in the Tot Lake Pegmatite. Fluorine is much lower in the Mavis Lake group, ranging between 1440 and 2260 ppm in Fairservice Pegmatites 1 and 6, and 850-1090 ppm in Taylor Beryl Pegmatite 2.

The higher boron contents correlate with the spectacular quantities of black tourmaline commonly occurring in metasomatic assemblages associated with rare-element pegmatites of the Mavis Lake group. Metasomatic assemblages of the Gullwing-Tot Lakes group, in contrast, show a marked paucity of tourmaline.

It can be concluded from the K-feldspar and muscovite chemistry that significant levels of certain rare-elements and associated elements such as Li, Rb, Cs, F, Sn and W underwent enrichment in the later intrusive units of the Ghost Lake Batholith, and in pegmatitic granites of the Gullwing-Tot Lakes pegmatite group. The GLB and Mavis Lake pegmatite group provide a better illustration of a 'parent-progeny' link on geological and mineralogical evidence than the Gullwing-Tot Lakes group. Although strong fractionation trends are developed in muscovites from the parental GLB, chemical discontinuities are obvious in the relative positioning of the Taylor Pegmatite No. 2 and the Fairservice spodumene pegmatites (see Figures 92 and 93). Petrogenetic reasons for these chemical discontinuities will be dealt with in Chapter 9.

Tantalum and Niobium in Muscovites from Rare-Element Pegmatites.

Tantalum and niobium contents in muscovites from granitic rocks have been advocated by Russian workers, such as Beus et al. (1968), as useful chemical indicators in the exploration for rare-element pegmatites.

Micas are the most ubiquitous rock-forming minerals in granites and pegmatites that accumulate Ta and Nb (Florke et al. 1978 a,b). Systematic variation in concentrations of these elements, coupled with

other indices of chemical fractionation (K/Cs, K/Rb, Rb/Cs etc.) in a given pegmatite or district exploration target area may provide useful information on rare-element specialization type to be expected and geographic direction(s) of fractionation trend(s).

Tantalum analyses are sparse in recent published trace element studies of North American granitoid rocks and minerals, even in such important contributions as Cerny et al. (1981). In view of the meagre quantity of data relating to Ta and Nb distribution in muscovite and biotite from North American pegmatite fields and parental granitoid masses, comparisons of the Dryden Pegmatite Field can only be made with available Russian examples. The importance of Ta and Nb determinations has been demonstrated in studies such as those of Beus (1966), Beus et al. (1968), Nekrasov (1970), Odikadze (1967) and Sitnin (1966).

Beus et al. (1968) regarded concentrations of Ta in muscovite exceeding 20 ppm as highly significant. Statistically, the probability of encountering such levels in muscovite originating from micaceous and ceramic-type pegmatites which are barren of rare-element minerals is only 0.07 percent according to calculations by means of the Laplace integral (Beus et al. 1968). In contrast Ta values exceeding 20 ppm will constitute between 51 and 98 percent of observations in muscovites from rare-element pegmatite fields. Odikadze (1967) considers that Nb/Ta ratios less than 5 in muscovite from pegmatites is a very reliable criterion of tantalum mineralization, where Ta contents exceed 30 ppm.

On a regional basis, sampling of muscovites from pegmatites can provide valuable information as to whether a potential rare-element pegmatite field or a parental granitoid mass has been encountered. For example, in the Gullwing Lake area, virtually all muscovite pegmatite dykes in vicinity of the Gullwing Lake pegmatitic granite contain anomalous Ta enrichment (31 to 260 ppm, unpublished data of the author) according to the criterion set by Beus et al. (1968). The full areal extent of this enrichment has not been evaluated in the Gullwing Lake area but is at least 3 by 10 km.

The Ta and Nb contents of muscovites and biotites in granitoid rocks from this work are compared in Table 50 with those from the U.S.S.R. It is apparent that the Ta contents in muscovites from the various pegmatite groups in the Dryden area are highly anomalous. Average values of Ta exceed the proposed "cut-off" of 20 ppm by 2 to 3 times. Maximum Ta concentration in muscovite for the Dryden area appears to be about 260 ppm. The meagre quantity of Ta in muscovites from ceramic-type pegmatites of the U.S.S.R. (3.5 ppm) is also very notable and similar low levels of Ta have not been encountered in muscovites from the Dryden area.

BIOTITE

Biotite occurs only in the northwestern part of the Gullwing-Tot Lakes pegmatite group, at the Gullwing Lake, Mica Point and RT pegmatitic granite dykes.

According to Hawthorne and Cerny (1982, p.81-82) biotite of their internal pegmatite type is "normally found in the deep-seated pegmatites of the ceramic and mica-bearing types and in simple barren

Table 50. Average Ta, Nb and Nb/Ta in Muscovite and Biotite from Dryden Pegmatite Field Compared to Data from Literature.

	<u>Ta</u>	<u>Nb</u>	<u>Nb/Ta</u>	<u>n</u>	<u>Reference</u>
<u>I. Muscovite</u>					
Gullwing Lake Batholith	<56	131	>2.3	20	This work
Gullwing Lake Li-Nb>Ta-Mo Pegmatite					
North Zone	55	190	3.8	6	This work
Central Zone	73	374	5.1	5	This work
South Zone	49	432	8.8	12	This work
Tot Lake Li-Ta>Nb-Rb-Be Pegmatite	30	55	1.8	1	This work
Taylor Beryl Pegmatite 2	100	76	0.8	2	This work
Fairservice Spodumene Pegmatite 1	69	109	1.9	8	This work
Miscellaneous and ceramic pegmatites barren of rare-element minerals (average of 20 analyses from U.S.S.R.)	3.5 ± 0.8	-	-	-	Beus et al. (1968); page 72 of translation
Rare-element bearing pegmatite fields of Eastern Transbaikalia (average of 48 analyses)	38 + 5	-	-	-	Beus et al. (1968); page 72 of translation
Field of rare-element pegmatites with tantalite-columbite containing 50 - 100 g/ton Ta ₂ O ₅ (average of 80 analyses)	28 ± 4	-	-	-	Beus et al. (1968); page 72 of translation
<u>II. Biotite</u>					
Gullwing Lake Li-Nb>Ta-Mo Pegmatite	113	1111	10.1	8	This work
Mica Point Nb>Ta Pegmatite	200	1380	6.9	1	This work
Unreplaced quartz-microcline pegmatite from Dzirula Massif (U.S.S.R.)	7	58	8		Odikadze (1967)
Albitized and greisenized pegmatites with columbite, manganotantalite and cassiterite	6	27	4.5	?	Odikadze (1967)
Two-mica Hercynian granites	13	140	8.6	3	Odikadze (1967)

n = number of analyses

to Be, Ti, Nb>Ta, REE (U,Th)-bearing pegmatites" of high grade metamorphic terranes. These biotites are typified by low Li, Rb, F, Cs and high Ti, Mg and Ba contents (Gordiyenko et al. 1976).

Occurrences of biotite associated with well fractionated rare-element pegmatites such as at Gullwing Lake have only rarely been described in the literature. Cerny et al. (1981, p.101-102) encountered lath-shaped biotite in the outer zone of pegmatites of the Shatford Lake, Axial, Greer Lake, and Cat Lake groups in Manitoba. Analyses of nine biotites from the Shatford Lake Group show a similarity in Li and Cs contents to biotites from the Gullwing Lake Pegmatite (Figures 92 and 93) but with generally higher K/Rb ratios. Biotite also appears in rare alkali-enriched dykes where it is concluded by Hawthorne and Cerny (1982, p.81-82) to have formed through contamination rather than from a primary melt.

Two types of pegmatite biotite are evident in the study area:

- 1) primary biotite in muscovite-biotite pegmatitic granite which envelop a quartz-blocky K-feldspar core, containing rare-element mineralization, at the Central and South Zones of the Gullwing Lake pegmatitic granite, Mica Point and RT pegmatitic granites,
- 2) secondary biotite in albitization I replacement zones of K-feldspar-quartz potassic pegmatite barren of rare-element minerals, as in the North Zone of the Gullwing Lake pegmatite granite.

Ten analyses of biotite for Ba, Cs, Li, Nb, Sn and Ta and major elements K₂O and TiO₂ are given in Appendix 37. One complete biotite analysis (Biotite 4472) from muscovite-biotite pegmatitic granite near

the South Zone, Gullwing Lake dyke is presented in Appendix 34, with an analysis of the co-existing muscovite (MU4471).

The most conspicuous chemical feature of the biotite is the high Li content (2,500 to 16,000 ppm) which results in a field virtually distinct from muscovite in the Li vs K/Rb plot (Figure 91). This field has similar K/Rb ratios to muscovites of the Central and South Zones (Gullwing Lake pegmatitic granite) and exhibits only minor overlap with these fields.

The maximum Li contents of Gullwing Lake Pegmatite biotites exceeds even that of the rare alkali biotite (1.07 percent) belonging to the exomorphic reaction class of Hawthorne and Cerny (1982, p.77, 80-81). Maximum Rb and Cs contents of the Gullwing Lake biotite are however, lower than in the rare alkali biotite, which contains 1.7 percent Rb and 1.04 percent Cs. Cesium levels, which vary between 530 and 1600 ppm, are higher in the Gullwing Lake North Zone albitization zones than in the muscovite-biotite pegmatitic granite. The latter fall within a tight cluster which generally exhibits higher Cs contents but lower K/Rb ratios than most muscovites from this rock type.

Average Ta and Nb levels in biotite, given in Table 50, are at least twice those in muscovite from various parts of the Gullwing Lake dyke. This feature results from the substitution of these elements for Ti^{4+} ; titanium is much higher in biotite (0.96 to 1.25 percent TiO_2) than in muscovite (0.02-0.22 percent in Appendices 29 and 32 to 34). These concentrations are considerably higher than in any biotites from various barren and peraluminous granitoids of the

U.S.S.R. (Table 50) as studied by Odikadze (1967). Even biotite from albitized-greisenized pegmatites containing mangano-columbite, manganotantalite and cassiterite (Table 50) contains very low Ta and Nb levels compared to that in the Gullwing Lake pegmatitic granite. The contents of Ta-Nb in biotite of the study area is also considerably greater than those in the average biotite (Ta = 10.1 ppm; Nb = 91 ppm) from magmatic rocks (Sitnin 1968 and Odikadze 1967, respectively).

The usefulness of biotite as a chemical indicator mineral is limited however as most rare-element pegmatites in the Dryden Field and elsewhere in the Superior Province (Mulligan 1965) do not contain biotite.

SPODUMENE

Nineteen analyses of Dryden Field spodumenes are given in Table 51. These correspond to the Type 1 phenocrystic (Fairservice pegmatites and Gullwing Lake dyke) and Type 2 zonal (Tot Lake) spodumenes of Heinrich (1975). For comparative purposes two analyses of secondary spodumene from the MNW Pegmatite (Figure 30) are included in Table 51.

Type 1 green phenocrystic spodumenes from Fairservice North Zone Pegmatite No. 1 exhibit a wider range in Na₂O than the Gullwing or Tot Lake Pegmatites (Figure 94). Most analyses, however, are below 0.5 percent and higher Na₂O levels may indicate intergrowths with very fine-grained albite but was not confirmed by X-ray diffraction. Na₂O contents of Tot Lake spodumenes show a restricted range similar to those of the Gullwing Lake pegmatitic granite.

Table 51. Analyses of Spodumene from the Dryden Pegmatite Field.

I. Type I: PHENOCRYSTIC SPODUMENE

MAVIS LAKE PEGMATITE_1

	0cm	1.2m	2.2m	13.4m	SP1	SP2	SP3	SP4	SP5	SP6	GULLING LAKE PEGMATITE	
											SP1	SP2
SiO ₂	62.7	63.1	62.0	60.5	61.8	60.5	61.3	61.5	61.2	60.6	60.4	58.4
Al ₂ O ₃	27.7	27.6	28.6	31.7	28.4	29.8	30.6	29.3	29.0	30.5	30.6	33.1
Fe ₂ O ₃	0.38	0.34	0.30	0.34	0.46	0.39	0.10	0.37	0.41	0.33		0.43
FeO	0.21	0.29	0.15	0.29	0.29	0.29	0.21	0.29	0.21	0.29	0.29	0.29
MgO	0.13	0.14	0.12	0.16	0.16	0.14	0.13	0.14	0.13	0.11	0.11	0.18
CaO	0.05	0.06	0.05	0.10	0.08	0.07	0.09	0.07	0.15	0.06	0.05	0.06
Na ₂ O	0.42	0.49	1.73	1.35	1.73	0.57	0.27	0.28	0.28	0.20	0.19	0.24
K ₂ O	0.02	0.03	0.04	0.01	0.02	0.07	0.02	0.04	0.04	0.04	0.02	0.04
TiO ₂	0.02	0.03	0.04	0.01	0.02	0.07	0.02	0.04	0.04	0.04	0.02	0.04
P ₂ O ₅	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.02	0.01
Li ₂ O	4.00	4.91	5.39	6.63	6.00	5.70	5.43	6.59	5.50	6.06	6.46	6.09
MnO	0.09	0.08	0.07	0.07	0.08	0.09	0.08	0.09	0.09	0.13	0.21	0.15
CO ₂	0.14	0.09	0.08	0.12	0.12	0.09	0.10	0.12	0.14	0.10	0.21	0.22
H ₂ O ⁺	0.13	0.04	0.08	0.30	0.09	0.13	0.14	0.13	0.24	0.14	0.36	0.87
H ₂ O ⁻	0.05	0.06	0.10	0.14	0.09	0.09	0.08	0.09	0.07	0.08	0.10	0.11
TOTAL	96.03	97.29	98.73	101.75	99.35	97.94	98.03	99.02	97.48	98.68	99.32	100.93

TRACE ELEMENTS (PPM)

B	10	<10	<10	10	10	20	<10	10	<10	<10	40	70
Ba	95	30	6	11	1	2	6	2	2	2	3	7
Be	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5
Cr	68	88	105	86	115	120	110	82	92	105	105	135
Ga	6	5	5	6	6	5	11	6	5	5	8	13
Nb	20	30	10	30	10	<10	60	10	20	20	220	890
Rb	13	13	14	13	13	13	13	13	13	14	36	75
Sr	29	24	24	59	26	34	37	34	35	28	1545	1299
Zn	2159	1662	1444	1954	1309	1316	1445	1894	1671	1540		
Al Ga												

Table 51, continued

11 TYPE 2 ZONAL SPODUMENE										
SP1	SP2	SP3	SP4	SP5	SP6	SP7	81F08° 219	81F08° 220	ASIER PETALITE MMW-1	MMW-2
SiO ₂	61.2	61.8	59.4	58.7	59.7	60.7	58.30	58.50	61.1	60.3
Al ₂ O ₃	28.9	28.6	32.0	32.7	30.6	30.5	25.40	25.70	29.9	30.7
Fe ₂ O ₃	0.18	0.14	0.09	0.14	0.07	0.05	-	-	0.014	0.009
FeO	0.29	0.21	0.21	0.21	0.21	0.15	0.24	0.24	0.07	0.13
MgO	0.12	0.17	0.18	0.18	0.15	0.15	0.12	0.0	0.17	0.15
CaO	0.10	0.11	0.16	0.14	0.08	0.09	0.89	0.66	0.07	0.05
Mn ₂ O	0.14	0.15	0.22	0.28	0.20	0.19	8.25	9.35	0.32	0.32
K ₂ O	0.04	0.05	0.06	0.04	0.06	0.06	3.31	3.16	0.19	0.10
TiO ₂	0.03	0.01	0.02	0.02	0.02	0.01	0.33	0.11	0.01	0.01
P ₂ O ₅	0.02	0.01	0.01	0.01	0.01	0.00	0.02	0.01	0.00	0.00
Li ₂ O	6.52	6.85	6.63	6.63	6.63	6.46	0.017	0.013	6.70	6.72
MnO	0.15	0.18	0.17	0.15	0.15	0.14	0.08	0.11	0.01	0.01
CO ₂	0.04	0.05	0.18	0.18	0.18	0.16	0.28	0.31	0.14	0.14
H ₂ O ⁺	0.18	0.20	0.27	0.31	0.34	0.30	1.39	0.70	0.36	0.32
H ₂ O ⁻	0.08	0.06	0.11	0.09	0.08	0.10	0.15	0.14	0.06	0.12
TOTAL	97.99	98.60	99.30	99.70	99.79	99.79	99.06	98.78	99.11	99.08

TRACE ELEMENTS (PPM)										
B							57	70		
Ba	70	50	40	40	10	80	230	140	30	20
Be	275	4	5	7	13	9	7	7	3	3
Cr	<5	<5	<5	<5	<5	<5	<5	19	<5	<5
F							210	380		
Ga	160	120	140	120	145	96	94	105	20	25
Li							58	78		
Nb	6	8	11	7	8	5	7	4	7	9
Rb	60	100	120	100	100	170	3010	2510	30	40
Sn							91	76.4		
Sr	13	13	12	13	12	13	55	60	13	12
Th							80	60		
U							<1	<1	10	8
Zn	28	20	38	30	40	18	45	95	7924	6508
Al/Ga	957	1263	1162	1413	1195	1689	1432	1297		

* spodumene totally replaced by albite and green mica

1. Unpublished data of the author from MMW Pegmatite, Georgia Lake Field, Ontario

FeO contents (0.15 - 0.29 percent) are similar in all analyzed spodumenes from the Dryden Field and apparently have no correlation with spodumene type and colour. Fe₂O₃ levels (Figure 94) are more variable; green phenocrystic spodumene from the Fairservice and Gullwing Lake pegmatites have Fe₂O₃ contents which range from 0.21 to 0.29 and are higher than in pink zonal spodumene of the Tot Lake Pegmatite (0.05 - 0.18 percent). Ucauwun (1981, p.89) documented some higher Fe₂O₃ contents, up to 0.64 percent, in spodumenes from the Fairservice North and South Zones as in Figure 95. The Tot Lake spodumenes exhibit FeO exceeding Fe₂O₃, the reverse of type 1 green phenocrystic spodumene. Maximum FeO contents of 0.31 percent are found in this pegmatite.

In Figure 94 the data fields for type 1 phenocrystic spodumenes (Fairservice and Gullwing Lake) and zonal spodumene are almost mutually exclusive; including Ucauwun's data results in minor overlap. Of the trace elements, Ga and Rb exhibit widest variation in spodumene (Table 51). Phenocrystic green spodumene from Fairservice Pegmatite 1 and Gullwing Lake Pegmatite have comparable respective ranges of 68 to 120 ppm and 105 to 135 ppm gallium. Pink zonal spodumene from the Tot Lake Pegmatite shows a tendency towards gallium enrichment although the concentration range of 94 to 160 ppm overlaps with that in phenocrystic spodumene.

Al/Ga ratios also exhibit significant variation. In phenocrystic spodumene these range from 1309 to 2159 (Mavis Lake) and 1299 to 1545 (Gullwing Lake). Zonal spodumene from Tot Lake illustrates a tendency for greater chemical fractionation in generally

Figure 94. Fe_2O_3 - Na_2O and MnO - Na_2O Variation (weight percent) in Spodumene from the Dryden Pegmatite Field. Diagram also includes Data of Ucauwun (1981) and an Unpublished Analysis of the Author (MNW Pegmatite, Georgia Lake Field of Ontario).

lower Al/Ga ratios of 957 to 1689; there is, however, some overlap with the other spodumene populations.

White secondary spodumene from the MNW Pegmatite is unique in the drastically lower Ga contents (20-25 ppm) and higher Al/Ga ratios (6508-7924) relative to the chemical data of Cerny et al. 1985, p.401. These workers suggest an Al/Ga ratio range of 1000 to 3000 for spodumene-quartz intergrowths after petalite. Although data on gallium content in various spodumene types are meagre, the results of this study suggest that significant variation within and amongst Superior Province pegmatite fields does exist.

Rubidium levels of 220-690 ppm in spodumene samples from the Gullwing Lake pegmatitic granite are much higher than those elsewhere in the Dryden Field. These concentrations correlate with elevated K_2O and H_2O^+ contents and may relate to aphanitic intergrowths with replacement micas or inclusions of the same. Ucakuwun (1981, p. 85) noted the partial replacement of spodumene by a mesh of fine-grained muscovite. Complete replacement of spodumene by a mica-albite assemblage leads to even higher levels of Rb (2510 - 3010 ppm), H_2O^+ (0.70 to 1.39 percent) and K_2O (3.16 to 3.31 percent) as at the Tot Lake Pegmatite (Table 51). There is also a significant increase in Na_2O (to 8.25 - 9.35 percent) relative to the unaltered spodumenes from the Fairservice Pegmatite No. 1, and losses in Li_2O , Al_2O_3 and SiO_2 . No significant change in the Al/Ga ratio (1297-1432) has, however, occurred in the replaced spodumenes relative to the unaltered ones.

BERYL

Studies by Cerny (1975) and Cerny and Turnock (1975) documented significant alkali element variation in beryl and showed that such variation represents a reasonable means of assessing fractionation levels of the host pegmatite. The most important chemical relationship is a progressive decrease in Na/Li ratio with rapidly increasing Cs in the direction of advancing chemical fractionation (Cerny 1975). This relationship is valid for both primary beryl and late secondary beryl from replacement zones (Cerny et al. 1981, p.113) of the Tanco Pegmatite and Greer Lake Pegmatite in the Winnipeg River District.

In Table 52, partial complete analyses for 20 beryls from the Dryden Field are presented, including eight partial analyses compiled from Ucauwun (1981, p. 101).

Green beryl from potassic pegmatite (GLB-7) of the Ghost Lake Batholith at the Dryden Airport has similar chemical characteristics to the exocontact Taylor pegmatites. This is reflected in the low Cs (827 ppm), Na₂O (0.48 percent) and Li (966 pm) contents documented by Ucauwun (1981, p.101).

The pattern of Na/Li vs Cs fractionation in pegmatites of the Mavis Lake group (Figure 95) reveals generally higher Cs contents in the Fairservice spodumene pegmatites than in the beryl pegmatites (Dryden Airport and Taylor Occurrences). However this is opposite to the normal fractionation pattern of decreasing Na/Li with increasing Cs as documented for either late beryl or multigeneration beryl from individual pegmatites such as the Tanco Pegmatite (Cerny et al. *ibid*).

Table 52. Analyses of Beryl from the Dryden Pegmatite Field.

Tot Lake Beryl-Pollucite-Spodumene Pegmatite										
Gullwing Lake Pegmatitic Granite										
	GWL-1	GWL-2	GWL-33Z	BIFWB-194	BIFWB-195	GPE-14	GPE-22	GPE-27	GPE-28	GPE-29
SiO2				57.90	60.60					
Al2O3				26.70	21.10					
Fe2O3				13.06	17.78					
FeO				0.21	0.21	0.07	0.06	0.05	0.02	0.02
MgO				0.15	0.17	0.02	0.02	0.01	0.02	0.01
CaO				0.12	0.07	0.11	0.10	0.06	0.02	0.09
Na2O	1.69	1.78	1.04	1.01	0.86	1.45	1.75	1.48	1.30	1.87
K2O				0.09	0.20	1.38	0.09	0.07	0.08	0.10
Rb2O			0.03	0.13	0.12	0.27	0.11	0.10	0.11	0.10
Cs2O	0.54	0.69	0.62	1.06	2.13	1.56	1.99	1.10	1.55	1.07
Li2O	0.65	0.74	0.42	0.80	0.75	0.89	1.38	1.25	0.51	1.06
TiO2				0.09	0.20					
P2O5				0.03	0.03					
MnO				0.02	0.02					
CO2				0.17	0.17					
H2O+				1.05	0.75					
H2O+				0.17	0.13					
Na/Li	4.2	3.9	4.0	2.10	1.80	2.6	2.0	1.9	4.1	2.8
Fair Service Beryl-Tantalite-Spodumene Pegmatites										
	Dryden Airport Beryl Pegmatite			Iaxlor Beryl Pegmatites			North Zone			East Zone
	APG-10	ZP-04	Dyke-1 ZP-08	FWB-83 -336	Dyke-2 FWB-83 -338	B-92	B-93	B-94	B-95	B-335
SiO2						63.0	63.0	62.9	63.5	
Al2O3						19.4	19.3	19.5	19.5	
Fe2O3										
FeO	0.25	0.43	0.55			0.21	0.14	0.14	0.07	
MgO	0.03	0.23	0.29			0.16	0.15	0.15	0.12	
CaO	0.02	0.01	0.01			0.06	0.04	0.04	0.04	
Na2O	0.48	0.97	1.08	0.72	0.40	1.27	1.34	1.14	1.11	0.91
K2O	0.03	0.08	0.06		0.07	0.10	0.03	0.09	0.02	
TiO2						0.06	0.03	0.03	0.02	
P2O5						0.00	0.03	0.00	0.02	
BeO						13.06	14.17		16.94	
MnO						0.01	0.01	0.01	0.01	
Co2						0.13	0.15	0.15	0.12	
H2O+						1.42	1.55	1.68	1.41	
H2O-						0.04	0.05	0.07	0.05	
Cs2O	0.09	0.17	0.18	0.23	0.24	0.27	0.19	0.31	0.17	0.39
Li2O	0.21	0.44	0.48	0.44	1.10	0.21	0.29	0.24	0.25	0.20
Rb2O	0.04	0.03	0.02	0.07	0.01	0.06	0.06	0.002	0.05	0.02
TOTAL	3.7	3.5	2.65	5.9	9.5	99.84	100.53	103.40	103.40	7.2
Na Li						7.3	7.8	7.8	7.3	

Table 52, continued

GWB-1 and 2:	Faint green beryl intergrown with platy albite in albitization II unit adjacent to relic spodumene at North Zone, Gullwing Lake pegmatitic granite
GWL-337:	lime green beryl from muscovite-biotite potassic pegmatite unit granite which envelopes quartz-rich core zone, South Zone, Gullwing Lake Pegmatite
81FWB-194 and 195:	White beryl from albitized spodumene-K-feldspar-quartz unit at Tot Lake pegmatitic granite
GPE-14, 22, 29:	White to colourless beryl from quartz-K-feldspar-spodumene-muscovite zone of Tot Lake Pegmatite. Analyses by Ucauwun (1981, p.101)
APG-10:	Light green beryl (10 cm in length) from potassic pegmatite of Ghost Lake Batholith (GIB-7). Beryl is in contact with blocky K-feldspar, quartz and green muscovite. Analysis by Ucauwun (9181, p.101).
ZP-04, 2P-08:	White to pale green euhedral beryl, 1 to 3 cm in diameter, 8 cm maximum length likely from Taylor beryl dyke 1. Analyses by Ucauwun (181, p.101).
FWB 83-336	White euhedral beryl, 3.1 x 2.7 cms, from incipiently albitized, rudimentary core zone. Taylor beryl dyke 1.
FWB-83-338:	White to light green euhedral beryl from tourmaline - K-feldspar - albite-quartz zone of Taylor beryl dyke 2. Composite sample of several small crystals (less than 1 cm diameter).
B-92 to 95:	White to faint blue euhedral beryl from quartz-rich core zone in adjacent spodumene-rich pegmatite
B-335:	White subhedral beryl, 10 x 15 mm, from albitized spodumene pegmatite dyke 15, East Zone, Fairservice Property.

Na/Li vs Cs variation in such beryls (Figure 95) interestingly reveals that those from spodumene pegmatites, exemplified by Fairservice Pegmatite 1, are chemically more primitive than those from pegmatites containing no spodumene (Taylor Pegmatites No. 1 and 2). The former express a range of Na/Li from 7.2 to 9.5, and relatively low Cs contents of 985 to 2900 ppm. These compositions fall near the least fractionated segment of the multigeneration beryl trend for the Tanco Pegmatite (Cerny et al. 1981, p.113). Late white beryl from completely albitized rare-element pegmatites at the East Zone, Fairservice Property (Pegmatite 15) reveals only a minor increase in Cs (3700 ppm) and a very similar Na/Li ratio of 7.2 (Table 52, sample B-33).

In contents beryl from the Taylor pegmatites exhibit lower Cs (1598 to 2150 ppm) and Na₂O (0.40 to 1.08 percent) while Li contents tend to be higher (995 to 2208 ppm).

Beryls from the Gullwing-Tot Lakes Group are, with exception of sample 337 (Table 52), are associated with albite-rich replacement units. Fractionation patterns reveal the expected decrease in Na/Li and increase in Cs with increasing mineralogical complexity. Beryl contained in the platy albite-lithian muscovite replacement complex in the North Zone of the Gullwing Lake dyke plots near the Greer Lake Pegmatite compositional field of Cerny et al. (1981, p.113). Cesium contents of 5100 and 6500 ppm and Na/Li ratios of 3.9 to 4.2 are much lower than those in the highly fractionated Tot Lake Pegmatite, that are comparable to those in the Bernic Lake compositional field of Cerny et al. (1981, p.11).

Figure 95. Cs vs. Na/Li for Early and Late Beryl from the Dryden Pegmatite. Compositional Fields for Late Beryl and Multigeneration Trends for the Tanco Pegmatite and Greer Lake Pegmatites 11 and 12 from Cerny et al. (1981, p. 113).

COLUMBITE-TANTALITE GROUP

Fine-grained columbite-tantalite group minerals occur sparsely disseminated in rare-element pegmatites of the Dryden Field. This mineral also occurs locally in potassic pegmatite (GLB-7) of the Ghost Lake Batholith (see petrography of the Contact Beryl Occurrence in Chapter 3). Localized concentrations of columbite-tantalite reach 0.2 percent as in unit 8 of the Tot Lake Pegmatite and in the Fairservice Pegmatite No. 2.

Appendices 39, 40 and 41 present thirty electron microprobe analyses and unit cell contents of columbite-tantalite from all important pegmatites of the Dryden Field. These tables include nine analyses from Ucauwun (1981, p.113).

Cerny and Ercit (1985) and Cerny et al. (1985, p.408-410) have extensively studied the chemistry of Ta and Nb-bearing phases in rare-element pegmatites. They advocated use of the $Ta/(Ta+Nb)$ vs $Mn/(Mn+Fe)$ diagram (Figure 96) to present compositional trends in these minerals, where the oxides of other elements such as CaO , Y_2O_3 , La_2O_3 , SnO , Ce_2O_3 , MgO and TiO_2 are low.

The fractionation trend for columbite-tantalite minerals from the Dryden Pegmatite Field is evident in Figure 96. The $Mn/(Mn+Fe)$ ratio must exceed a value of 0.6 before a steep rise in $Ta/(Ta+Nb)$ commences. Mineral compositions thus proceed in the series: ferrocolumbite --->manganocolumbite ----->manganotantalite. In this regard there is a close resemblance to the Cross Lake Pegmatite Field of northern Manitoba (Cerny et al. 1985).

Figure 96. FeNb_2O_6 - FeTa_2O_6 - MnNb_2O_6 - MnTa_2O_6 Variation (mole percent) in Columbite-Tantalite Minerals from the Dryden Pegmatite Field. Analyses 1 to 9 from Ucakuwun (1981, p. 113).

Individual pegmatite groups in the Dryden Field reveal some differences in their columbite-tantalite minerals. The Gullwing-Tot Lakes Group exhibits a wider compositional span than the Mavis Lake Group (Figure 96). Relatively primitive ferrocolumbite compositions occur in the more poorly fractionated Mica Point and Gullwing Lake pegmatitic granites.

Ferrocolumbite characterize the albitization I zones of the Gullwing Lake pegmatitic granite. The second stage of albitization here resulted in only a modest Ta enrichment and no significant increase in $Mn/(Mn+Fe)$, as revealed in sample 394 (Appendix 40).

Columbite-tantalite group minerals from the Tot Lake Pegmatite document a trend toward much higher accumulations of Ta and Mn (Figure 96). A compositional gap is however apparent and corresponds to most of the manganocolumbite part of Figure 96.

Ferrocolumbite is abundantly present in a pervasively albitized, blocky K-feldspar-spodumene quartz unit near Trench 3 (Unit 8 in Figure 30). Normally highly replaced pegmatitic rocks such as the Tot Lake Pegmatite are characterized by low Nb/Ta ratios in the rock and columbite-tantalite group minerals (Florke et al. 1978b). Specimen 408 (Figure 96 and Appendix 41) contradicts this trend; reasons for the depletion in Ta and Mn relative to other analyses from this pegmatite are unknown. It may be possible that Mn-Ta depletion occurred with extensive albitization in the columbite zone of the Tot Lake Pegmatite.

Variation in columbite-tantalite minerals from the Mavis Lake Group is notably more restricted (Figure 96). These define a main trend from manganocolumbite to manganotantalite, exclusive of ferrocolumbite (specimen 409) from the Taylor Pegmatite No. 2. Tantalum concentrations and Ta/(Nb+Ta) ratios achieve maximum values of 77.4 percent and 0.85, respectively for spodumene pegmatites of this group.

More primitive compositions occur in Taylor Pegmatite No. 2. Two grains from a small replacement pocket (similar to Plate 10c) nevertheless exhibit a wide compositional span (Figure 96).

CORDIERITE AND GARNET

Electron microprobe analyses of one cordierite from the Ghost Lake Batholith and four garnets from plutonic units GLB-5 and -7, the Zealand Stock, and several rare-element pegmatites are given in Table 53. Three additional analyses are from Ucauwun (1981, p.103).

Cordierite is considered by White et al. (1986) to be a diagnostic indicator of S-type granites. In the Ghost Lake Batholith this mineral is restricted to the northwestern part of the pluton (Figure 12). Analysis of a cordierite crystal from locality C17-3 reveals no chemical zoning and a Mg/(Fe+Mg) ratio of 0.41. These values compare with cordierite analyses from the Pakwash Lake area of the ERS and are somewhat higher in the Bundarra plutonic suite and the Strathbogie Batholith, LFB of Australia (Table 53). Clemens and Wall (1981) remarked that in general cordierites from S-Type granitoids have similar Mg/Mg+Fe ratios of around 0.5. The Pakwash Lake cordierite contains a lithium concentration of 122 ppm which likely

Table 53. Analyses of Cordierite and Garnet from the Ghost Lake Batholith, Zealand Stock and Related Pegmatites with Comparative Data from other Peraluminous Granitic Suites.

CORDIERITES										GARNETS			
Bundarra Plutonic Suite Flood and Shaw 1975										Strathgowie Batholith (Phillips et al. 1981)			
Pamash Lake Area										E-J1-4 (Core)			
C17-3 (Core)										E-J1-4 (Core)			
C17-3 (Core)										E-J1-4 (Core)			
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C17-3 (Core)													

Table 53, continued

GARNETS					
	Taylor Pegmatite Dye 1 (wt.-%)	Taylor Pegmatite Dye 2 (wt.-%)	Soudanese Pegmatite 6 Fairweather Property (wt.-%)	Tot. Late Soudanese- Pegmatite (wt.-%)	
SiO ₂	35.71	30.09	36.72	35.56	
Al ₂ O ₃	20.02	20.64	20.83	20.74	
Fe ₂ O ₃	28.23	20.93	0.00	2.53	
FeO*	14.32	21.30	35.91	36.29	
MnO	0.61	0.35	0.02	0.03	
MgO	0.42	0.10	0.20	0.04	
CaO					
Na ₂ O					
K ₂ O					
P ₂ O ₅					
TiO ₂	0.08			0.04	
Cr ₂ O ₃					
LO ₂					
H ₂ O*					
H ₂ O ⁻					
TOTAL	99.29	94.04	101.76	100.03	
STRUCTURAL FORMULAE					
Si ⁴⁺	2.984	2.915	3.010	2.960	2.960
Al ^{IV}	1.972	2.006	1.989	2.021	2.035
Al ^{VI}					
Ti ⁴⁺					
Fe ³⁺	0.076	0.063	0.043	0.006	0.003
Mg					
Fe ²⁺	1.979	1.443	1.230	0.543	0.176
Mn	1.014	1.507	1.691	2.453	2.770
Na					
Ca	0.002	0.025	0.029	0.022	0.075
K					
O	12	12	12	12	12
Mn/Mn ²⁺	0.34	0.50	0		
O ⁹⁴					

Table 53, continued

SAMPLE DESCRIPTIONS

C17-3	Euhedral to subhedral cordierite varies from 2 x 5 mm to 2 x 2 cm parallel to C-axis. Partially replaced along margin to fine-grained dark green chlorite.
Pawash Lake	Large aggregate of vitreous, unaltered, deep blue cordierite from garnet-cordierite biotite granite leucosome in metatectic metasedimentary migmatite near Ear Falls, Ontario. Unpublished data of the author.
Bundarra Plutonic Suite, S.E. Australia	Euhedral, up to 2 cm length from muscovite-biotite granite.
Strathgogie Batholith, S.E. Australia	Euhedral, type A cordierite of Phillips et al. (1981), 0.5 to 10 mm, with scarce micro-inclusions of biotite, zircon, and titanite.
Dryden Airport	Brown to red-brown, almandine garnet, 0.1 to 1 cm diameter from aplitic patches of pegmatitic granite (Uchumun 1981, p. 103).
C11-4	Red-brown euhedral almandine garnet, 5 mm diameter from muscovite-garnet-tourmaline fine-grained leucogranite near Contact Beryl occurrence.
C11-5	Large 1.5 cm diameter red-brown almandine garnet from garnet-muscovite-biotite granodiorite, Zealand Stock.
Taylor Beryl Pegmatite #1 (equivalent to 2P Pegmatite of Uchumun 1981)	Brown-red almandine garnet euhedral, 0.1 - 0.5 cm in diameter, associated with tourmaline, quartz, and K-feldspar.
Taylor Beryl Pegmatite #2	Red-brown euhedral garnet, 4 mm diameter, in garnet-muscovite-tourmaline albite-rich pegmatite.
Fairservice Spodumene Pegmatite #6	Mass of euhedral to subhedral orange-bessartine garnets 2 mm to 1.2 cm diameter, associated with green muscovite, albite, and white beryl at periphery of an albite-rich blocky K-feldspar crystal.
C11-6 Pegmatite	Unzoned orange-bessartine garnet associated with poikilitic quartz, muscovite, and albite. K-feldspar, euhedral 0.2 - 0.4 mm in diameter (Uchumun 1981, p. 103).

substituted for Mg. No data in the literature exists for distribution of lithium in cordierite, however this mineral could represent an important host of lithium and therefore is of some potential importance in the modelling of restite unmixing processes (see Chapter 9).

Garnets in the GLB are abundant until only in evolved units (GLB-3, -4, -5, -6, and -7). Garnet never co-exists with cordierite in GLB-1, being usually absent from this unit (Table 53). The main chemical variation in garnet is one of progressively increasing Mn/(Mn+Fe) ratios in the rock series: GLB -->beryl pegmatite -->tantalite-beryl-spodumene pegmatite -->tantalite-beryl-spodumene-pollucite pegmatite. Contents of Ca and Mg are very low.

Chapter Nine

PETROGENESIS OF PERALUMINOUS GRANITOID ROCKS AND RARE-ELEMENT PEGMATITES OF THE DRYDEN AREA

INTRODUCTION

Substantial progress has been made in granite and pegmatite petrogenesis during the last fifteen years. Much of the impetus for the recent research was instigated by the works of White and Chappell (1974, 1977). They postulated that two fundamental types of crustal magma were involved in the evolution of the Paleozoic Lachlan Fold Belt (LFB), in the Tasman Orogen of eastern Australia:

- 1) S-Type magmas derived from sedimentary or supracrustal protoliths (peraluminous)
- 2) I-Type magmas derived from igneous or infracrustal protoliths (metaluminous to weakly peraluminous).

Petrogenetic studies of peraluminous granites and pegmatites, mainly focussed upon Phanerozoic rocks have increased since this work, (e.g. Le Fort 1981; Ludington 1981, London 1986; Burnham and Nekvasil 1986; Shearer and Papike 1987; Wall and Clemens 1987 ; London et al. 1988, 1989).

Despite their abundance in certain districts such as the English River and Quetico Subprovinces in northwestern Ontario only a few accounts of Archean S-type peraluminous granitoid rocks have appeared in the literature (e.g. Goad and Cerny 1981, Percival and Williams 1989). Granitoid rocks of probable I-Type occur widely in the Winnipeg River and Wabigoon Subprovinces (Beakhouse 1983); they include rocks of the potassic plutonic suite which in bulk composition

resemble some units of the GLB. However, based upon differences in age of emplacement, geological setting, accessory mineralogy and some important chemical features (e.g. molecular $\text{Al}_2\text{O}_3/\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O}$) the peraluminous GLB is interpreted as an S-type granite.

Five important processes can potentially influence compositional variation in granitoid rocks:

1. restite unmixing
2. liquid-state thermogravitational diffusion
3. crystal-melt fractionation during partial melting and crystallization
4. hybridization (contamination with crustal fluids and/or rocks)
5. exsolution of a volatile-enriched phase.

It is possible that many granite suites evolved via a succession or interplay of two or more of these mechanisms, although thorough documentation of multi-process granites is rare in the literature.

These processes, where applicable, will be discussed in this chapter in terms of their relevance to the petrogenesis and emplacement of the Ghost Lake Batholith (GLB) and its proximal rare-element pegmatites.

TECTONO-MAGMATIC ENVIRONMENTS OF S-TYPE GRANITOID ROCKS

Before examining the various mechanisms that can possibly produce peraluminous granitoid rocks and rare-element pegmatites, it is useful to examine the tectono-magmatic environments for peraluminous granitoid rocks of younger orogenic belts in order to assess if similarities with the study area rocks exist.

S-type peraluminous granites in the Lachan Fold Belt (LFB) of Australia are high level, possess contact aureoles and were generated within an intracontinental rift setting (Chappell 1984; White et al. 1986); they may have been accompanied by local accumulations of cordierite-bearing S-type volcanics (Wyborn et al. 1981). In contrast, Archean peraluminous granites from the Superior Province as exemplified by the Churchill Batholith (Figure 4) are deeper level since they lack contact aureoles and are predominantly associated with high grade metasedimentary migmatite belts (English River and Quetico Subprovinces).

Chemically evolved peraluminous granitoid rocks such as the GLB may, however, locally occur at higher crustal levels, i.e., in low and medium grade host rocks of the boundary zones of the migmatites with metavolcanic-rich subprovinces. Even these well fractionated granites, however, do not possess contact aureoles. S-Type volcanics are unknown in Archean terranes.

Since the ERS and WRS peraluminous granitoid rocks reflect a deeper crustal emplacement level, they potentially may "image their sources" in the sense of Chappell (1979, p.400) even more closely than those of the LFB. These rocks may thus harbour key evidence for the unravelling of tectonic processes involved in the development of Archean high grade belts.

The author suggests that it is still premature to advocate application of plate tectonic models in the northwestern Superior Province of Ontario to explain the origin of high grade gneissic belts such as the ERS, in view of the virtual lack of a geochronological

framework for anatectic and intrusive events or a regionally comprehensive structural synthesis (also see Park 1988).

Nevertheless, some geological and chemical similarities of ERS peraluminous granitoid rocks with those of younger orogenic belts (as exemplified by the Himalayas; Le Fort 1981) are worthy of mention and may imply similar tectonomagmatic environments:

1. wide range Na/K ratios
2. plutons commonly have high length: width ratios
3. long axis of the plutons are concordant with major shear zones
4. cordierite is a widespread varietal mineral

An extensive review of Phanerozoic orogenic granitoid rocks in the context of plate tectonics was undertaken by Pitcher (1982) who suggested the following divisions:

- 1) Pacific-type
- 2) Andinotype
- 3) Hercynotype
- 4) Caledonian-type

Peraluminous granitoid rocks of the ERS and WRS most closely compare to those of the Hercynotype category in terms of metamorphic and geological setting, mineralogy, chemistry and metallogeny.

Pitcher (1982) indicated that Hercynotype granites occur in tectonically thickened crust in two environments:

- 1) continental collision zones
- 2) encratic, ductile shear zones

Those in continental collision terranes are commonly hosted in

regionally metamorphosed rocks and occur as foliated, syntectonic plutons.

Two morphological categories of peraluminous granitoid rocks are obvious in the ERS and WRS (Figure 3):

- 1) extensive, equant masses which occur within and near granulite centres
- 2) bodies elongated parallel to, and situated near or at the subprovince boundaries

The first group, which is more extensive than the second group, is exemplified by the Churchill Lake Batholith. It is composed of granite with the varietal minerals cordierite, biotite and orthopyroxene. The restriction of orthopyroxene to this granite group, coupled with association of these rocks with the granulite metamorphic highs, suggest derivation of these rocks through extensive partial melting of metasediments at a deeper crustal level than the granitoid rocks associated with the subprovince boundaries.

The second group, on the other hand, comprises smaller plutons such as the Ghost Lake, Wenasaga Lake and Sharpe Lake Batholiths, Root Bay Stock and a remarkably elongate granite confined within the Sydney Lake Fault Zone (Figure 3). These plutons, which are petrographically similar to the first group of granitoid rocks are hosted in medium grade metasediments and/or metatexite migmatite equivalents with a low percentage of leucosome (Table 1). Such plutons are typically disharmonious in the sense of Walton (1955) and lie in discordant contact with metasediments of both metamorphic zones, as exemplified by the Ghost Lake Batholith in the WRS and the Sharpe Lake Batholith

in the ERS. These contact relations suggest a more distal character for this granitic material, relative to plutons of the granulite centres.

PERALUMINOUS GRANITE-PEGMATITE GENESIS RELATED TO MAJOR ARCHEAN FAULTING

Major fault zones commonly mark subprovince boundaries in the Superior Province of northwestern Ontario (Wilson 1971, Breaks et al. 1978, Stone 1981). Fault-controlled emplacement of pegmatitic granites and rare-element pegmatites has been advocated by several workers (Beus 1962, Beus 1968, Temnikov 1972, Koval 1975, Strong and Hanmer 1980, Cerny 1982 a,b). Cerny et al. (1981, p. 147) remarked that "the restriction of the Winnipeg River pegmatitic granites to major regional fault systems seems to be significant in their genesis and evolution....". Trueman (1980) defined two ages of regional faulting in the Bird River greenstone belt, in the same region of southeastern Manitoba. Many rare-element pegmatites are situated near relatively early fault systems, such as the one extending from Ryerson Lake on the Manitoba-Ontario boundary west to Lac du Bonnet, forming the contact between the Booster Formation and the Bernic Lake Formation (Cerny et al. 1981, Map ER-80-1). This fault truncates F_2 fold axes (Cerny et al. 1981, p. 134) and also apparently cuts some pegmatitic granite intrusions, such as the Osis Lake pluton and, more conclusively, an unnamed pegmatitic granite stock at the northwestern end of Ryerson Lake where left-lateral movement is indicated (Cerny et al. 1981, Map ER80-1). Furthermore, major faulting along the Bird River greenstone belt - WRS boundary transects the Greer Lake pegmatitic granite (Cerny et al. 1981, p. 138). From the field

evidence it can be inferred only that the latest movements on these fault systems post-date regional metamorphism and emplacement of pegmatitic granite stocks.

Peraluminous granitoids that reveal a definite relationship to major ductile shear zones do exist within the Superior Province and are situated reasonably close to rare-element pegmatite districts. A good example is located 25 km northeast of the Winnipeg River pegmatite district. Geological and structural work by Breaks et al. (1978) and Stone (1981) identified a mylonitic, medium-grained to pegmatitic, cordierite-garnet-muscovite peraluminous granite which is completely confined to the Sydney Lake Fault (Figure 3). The extreme elongation of this body [130 km by 0.3 - 2.2 km, the length including 8 km in Manitoba as mapped by McRitchie (1971, p. 22-26 and Map 69-4)] renders it unlike any other peraluminous granitic mass in the ERS. Mechanisms governing the preferential distribution of this mass within a major ductile fault zone plausibly include frictional fusion and/or anatexis aided by ingress of fluids, as inferred in the genesis of similar leucogranites in Brittany (Strong and Hanmer 1980) and in the Himalayas (Le Fort 1981). Elsewhere Chackowsky and Cerny (1984) described a unique petalite-bearing pegmatitic granite in the Sachigo Subprovince at Red Sucker Lake, Manitoba whose emplacement was apparently controlled by a regionally extensive, non-subprovince boundary ductile fault zone.

If chemical fractionation processes in such fault-confined peraluminous granitoid rocks as those in the Sydney Lake Fault effected accumulation of rare elements, it is curious that rare-

element pegmatites are not more common in this zone and in nearby country rocks. Only one rare-element pegmatite is situated near this linear granite mass: the Sandy Creek beryl occurrence just north of Ear Falls (Breaks and Bond, in press). This apparent scarcity of rare-element pegmatites here may merely be a function of a lack of mineral exploration.

Peraluminous pegmatitic granites of rare-element specialization in southeastern Manitoba are typically late- to post-kinematic (Trueman and Cerny 1981) which suggests that movement on the major fault systems was not significant in the genesis or emplacement of some of these rocks. Field and geochronological investigations have established that movement along the Lake St. Joseph Fault, situated along the ERS-Uchi Subprovince boundary to the east of the Sydney Lake Fault, had effectively ceased by $2,689 \pm 4$ Ma (G.M. Stott, personal communication 1989). However, scant U/Pb zircon geochronological data from the ERS and WRS suggest that emplacement of the youngest rare-element pegmatitic granite in the Dryden Field is substantially later than this ductile deformation event. This succession may also be inferred from field relations at the post-tectonic Gullwing Lake pegmatitic granitic dyke which post-dates the potassic plutonic suite. If the age of the latter is approximately equivalent to the 2560 ± 40 Ma age determined by Krogh et al. (1976) for a pegmatite member of this suite situated 85 km to the northeast of the Gullwing Lake dyke, it follows that emplacement of the Gullwing Lake dyke was considerably younger than the cessation of movement along the subprovince boundary faults.

The age of metamorphism for the granulite zones of ERS and WRS is interpreted from U/Pb zircon work from Krogh et al. (1976). They dated leucosome pods in metasedimentary migmatite along the ERS-WRS boundary near the eastern end of Lac Seul; the age of anatexis was estimated as 2681 ± 20 Ma and marks the earliest known peraluminous granites in northwestern Ontario.

If the preceding interpretation regarding the approximate age of the rare-element mineralization in the Gullwing Lake dyke is valid, it is suggested that a time interval of at least 120 Ma may have been available for the chemical-evolution of peraluminous granites.

Crustal level of emplacement of peraluminous granitoid rocks may be an additional factor to consider in explaining the geographic distribution of rare-element pegmatites in the ERS. Deeper crustal levels of high and granulite grade do not contain rare-element pegmatites. However, if peraluminous, S-type magmas can achieve higher crustal levels as represented by the medium grade supracrustal rocks in the boundary zones, then a better possibility exists for concentration of rare elements. It is plausible that rare elements formed complexes with F and possibly B, that allowed these elements to progressively fractionate towards the highest parts of a given peraluminous granite pluton.

A higher crustal level for the ERS and WRS in Manitoba relative to Ontario may partially explain the greater abundance of metavolcanics (Bird River Belt) and the presence of lower pressure, petalite-bearing, rare-element pegmatites in Manitoba (e.g. Tanco pegmatite, Cerny et al. 1981).

Important research by Le Fort (1981) and France-Lanord and Le Fort (1988) in the Himalayan-Indus suture zone has cast light on the origin of cordierite-bearing, S-type leucogranites in very young (14 to 25 Ma) continental collision zones. This field work and chemistry (rock, mineral and oxygen, deuterium and hydrogen isotopes) led to a model for the derivation of peraluminous melts within an intra-continental thrust zone. Anatexis was induced in the Main Central Thrust (MCT) by thrusting of a relatively hot sialic slab (Higher Himalayan Crystallines - HHC) over a low grade volcano-sedimentary pile (Midlands Formation), and was facilitated by the sudden release of fluids from the Midlands Formation. Formation of syn- to late-tectonic peraluminous granites resulted and spanned a 10 Ma interval.

Among the granitoid rocks of the ERS, the elongate granite within the Sydney Lake Fault (Figure 3) bears a striking shape similarity to cordierite leucogranite plutons in the MCT (e.g. Manaslu Pluton, Le Fort, 1981).

If a continental collision mechanism can be applied to Archean terranes, it follows that crystallization of the youngest and undeformed members of the peraluminous granites within a given thrust zone should provide a lower age limit for the collision event.

In the study area there is no obvious field evidence supporting anatexis aided by ingress of fluids into fault zones or of fault-controlled emplacement of the GLB and any rare-element pegmatite. The nearest regional fault zone, the Wabigoon Fault, which exhibits mylonitic deformation 2 to 3 km south of the GLB and adjacent Mavis Lake pegmatite group, does not contain pegmatitic granite intrusions.

RESTITE UNMIXING MODEL

This petrogenetic model has been advocated mainly by Australian geoscientists working on granitoid rocks of the Circum-Pacific rim. While general acceptance of the model has met with limited success, and even strong criticism (e.g. Clemens et al. 1985), there appears to be a more widespread application of this model to S-Type granitoid rocks than to those of I-Type derivation (e.g. Le Fort 1981, Kistler et al. 1981, Price 1983).

The basic concept behind this petrogenetic model is that anatectic granitoid magmas produced in the crust are composed of varying proportions of leucocratic melt and restite (melanosome), the latter occurring as refractory xenocrysts and polymineralic clots. Emplacement of such granitic magmas to higher crustal levels is possible only after the extent of partial melting passes the 20-35 percent value, termed the "rheologically critical melting percentage" (Arzi 1978, p.175) also called the "critical melt fraction:" (van der Molen and Patterson 1979, p.300). Due to the high viscosity and yield strength characteristics of these melts, Chappell et al. (1987) indicated that separation of restite from such magmas would be inhibited, and thus these granites would tend to "image their sources" (Chappell 1979, p.400).

Chappell and White (1982) ascribed chemical variation in such granites (schematized in Figure 97) as due to varying degrees of separation of restite from melt during emplacement and solidification. The initial two lines of evidence used by these workers are linear data arrays in Harker diagrams (Figure 97), and the invariable blank

Figure 97. SiO_2 vs. Element Plot of White and Chappell (1977) Schematically Illustrating the Chemical Variation Derived by Partial Melting of a Source Rock (S) Producing Restite (R) and Liquid (L).

presence of mafic inclusions and clots and refractory minerals such as plagioclase with high anorthite content and cordierite-sillimanite intergrowths within the Australian granitoid rocks. Compositions of the source rock (S), restite (R) and an initial, restite-free liquid (L) theoretically lie along a line. Increase in partial melting converts more of the source rock into leucosome and restite, and if no separation of these migmatitic components occurs, overall chemical composition tends to progress along this line towards that of the source (S). This variation is expressed as a linear series of granite compositions (1 to 5) which progressively reflect greater incorporation of restite material (Figure 97). The array of data points cannot continue past the source composition (S), as this would represent the theoretical magma composition resulting from complete melting of the protolith. However, some variation in the slope of the data points is evident in rock suites. White and Chappell (1977) interpret the negatively-sloping data arrays as minimum melt compositions. Chappell and White (1987, p.1130) established the minimum melt SiO_2 composition for a given granitic suite by "the intersection, on diagrams, of various refractory components such as P_2O_5 and MgO with the SiO axis" at zero percent for these components. Essentially nil quantities of MgO , P_2O_5 , Cr and Ni would be expected in minimum melts. In contrast, many granitic suites exhibit essentially flatly sloping data arrays which do not intersect the abscissa (compare Moruya and Jindabyne suites from LFB in Chappell and White, 1987, p.1131). These trends were interpreted by Chappell and White (1987, p.1130-1131) to be non-minimum melt compositions. Such were thought by these workers to have formed from a different source

to that which produced minimum melt compositions. The non-minimum compositions apparently required "a deficiency of one of the haplogranite components, quartz, K-feldspar or plagioclase in the source" (Chappell and White, 1987, p.1131) and hence required higher temperatures before extensive melt was formed.

One major obstacle to recognition of restite unmixing trends in chemical variation diagrams is the extent to which superimposed igneous processes (e.g. fractional crystallization; exsolution of an aqueous fluid) have modified the original magma composition. These processes can potentially cause considerable compositional scatter in Harker variation diagrams.

APPLICATION OF THE RESTITE UNMIXING MODEL TO PETROGENESIS OF PERALUMINOUS GRANITOID ROCKS OF THE GHOST LAKE BATHOLITH AND ENGLISH RIVER SUBPROVINCE

The possibility of restite-unmixing involvement in the genesis of peraluminous granitoids of the ERS and GLB is suggested by the widespread presence of sillimanite-cordierite-biotite-enriched mafic segregations. To a lesser degree this process is inferred by chemical variation, geological context and petrographic features. Figures 43 and 98 are total FeO-MgO-Na₂O+K₂O and Al₂O₃-SiO₂ plots depicting northwestern Ontario peraluminous granitoid rocks, non-migmatized metawackes and metapelites possibly representative of source material. This diagram includes a restite selvage from metatextitic metasedimentary migmatite which may reflect in situ derivation of it and the coexisting leucosome (Sample Ell-1 in Figure 98). If anatexis can be assumed to be the dominant process in the origin of granitic layers and pods in their metapelitic hosts (Plate 20a, p.445), such

Figure 98. Application of Restite Unmixing Model Al_2O_3 - SiO_2 Diagram of White and Chappell (1977) to Peraluminous Granitoid Rocks of Ghost Lake Batholith and English River Subprovince.

migmatites containing a low percentage of granitic melt could represent the initial stage of restite unmixing.

In the Al_2O_3 - SiO_2 diagram ERS peraluminous granitoids, petrographically similar to those of the GLB, occur in a broadly elongate field. This crudely linear data array lies approximately parallel to the restite unmixing line suggested for the ERS S-type granitoid rocks and spans about two-thirds of the compositional range between the restite-free initial leucosome composition (77.5% SiO_2 and 14.2% Al_2O_3 in Figure 98) and that of the plausible source rock composition represented by the average low and medium grade ERS metapelite. The SiO_2 range for ERS peraluminous granitoid rocks (60.9 to 75.3 percent, Breaks and Bond in press) is similar to the 63.4 to 74.5 percent range for SiO_2 indicated by Chappell and White (1982, p.90) for 316 S-type granitoid rocks from the Lachan Fold Belt (LFB).

Granitoid rocks from GLB-1, the unit most likely from field features to exhibit restite-unmixing, exhibits a range in SiO_2 (70.5 to 77.4 percent) and Al_2O_3 (13.71 to 17.3 percent content) that is more restricted than that of the LFB and ERS S-type granite counterparts. In fact, most analyses of the GLB exceed 70 percent SiO_2 and there is a tendency to higher SiO_2 content as several surpass 76 percent. Biotite-rich peraluminous granitoid rocks characterized by lower SiO_2 levels (between 60 and 65 percent, in Figure 98), that occur widely in the ERS (Breaks and Bond, in press), were not found in the GLB.

Variation in total Fe, Mg and Na+K for the ERS and GLB peraluminous granitoid rocks also shows a coherent linear distribution

between mafic-free leucosome at the Na+K apex and metapelitic protoliths (Figure 43). The less mafic GLB rocks reveal a more restricted range in total iron and magnesium contents.

Analyses of a granite leucosome layer and a co-existing restite selvage from metatexite metasedimentary migmatite situated 0.5 km west of the GLB on Highway 17 are plotted in Figure 98 (sample Ell-1: restite = 24.0% Al_2O_3 , 45.7% SiO_2 and leucosome = 15.5% Al_2O_3 , 73.2% SiO_2). This migmatite could provide the approximate upper and lower SiO_2 - Al_2O_3 limits of restite-unmixing variation for the GLB if the granitic material from both the migmatite and the batholith had a similar origin. Disparities are however obvious as linear field of GLB-1 extends to higher SiO_2 and lower Al_2O_3 than the migmatitic leucosome. In addition the GLB-1 field lies at the lower end of the leucosome-restite tie-line; under ideal conditions of restite-unmixing a more extensive distribution of data points along the tie-line should be expected (White and Chappell 1977).

Harker variation plots for Na_2O , K_2O and TiO_2 for all plutonic units of the GLB (Figures 39 and 40, Chapter 7) exhibit chaotic overall trends and a characteristic wide range in $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios (Figure 44). Chappell and White (1982, p.99) state that although a high degree of covariance is obvious in binary variation diagrams for LFB I-type granites, a poorer correlation is usual for S-type granitoid rocks, which they ascribe to more heterogeneous source regions.

Partial melting of metagreywackes has been advocated by many workers (e.g. White et al. 1986; Miller 1985; Arth and Hanson 1975) as

a mechanism capable of generating copious quantities of peraluminous magma. A restite unmixing model for the ERS, utilizing metawackes as the sole protolith composition in the magma source area, does not however accord with the data. Figure 98 reveals that most analyzed low and medium grade wackes from the ERS and Zealand metasediments are removed from the suggested ERS unmixing line and are completely overlapped by ERS peraluminous granitoids. If wackes predominated as a source rock composition, the peraluminous granitoids derived from them should not extend to the higher Al_2O_3 and lower SiO_2 contents beyond the wacke field as complete anatexis in a closed system should compositionally match that of the protolith.

In conclusion, applicability of the restite unmixing model to peraluminous granitoids of the ERS and to unit GLB-1 of the Ghost Lake Batholith is vaguely suggested by the approximately linear data arrays in SiO_2 - Al_2O_3 and AFM diagrams, coupled with field evidence indicating a widespread distribution in GLB-1 of cordierite-sillimanite-biotite restite segregations and identical, peraluminous accessory and varietal minerals in granitoid rocks from both areas. The bulk composition range of the granites is, however, inconsistent with this model, implying that additional processes may have played a role in the genesis of GLB-1. The ensuing discussion will now introduce other petrogenetic processes that could have brought about composition variation in peraluminous granitoid rocks of the study area.

THERMOGRAVITATIONAL DIFFUSION (TGD)

The role of TGD in the fractionation of peraluminous granitoid rocks is, as yet, undefined. This process has been convincingly

demonstrated for high silica, calc-alkaline rhyolitic ignimbrites by Hildreth (1979, 1981). The theoretical basis for this mechanism was developed by Shaw et al. (1976) and basically is a modification of the Soret effect in which convection-aided diffusion of chemical components in a high silica magma chamber occurs in response to thermal and gravitational gradients.

Application of this process to granite petrogenesis is, however, still in its infancy; few supportive studies have emerged to date (e.g. Ludington 1981; Whalen 1983). This progress has been hindered somewhat by experimental results obtained on andesitic, dacitic and rhyolitic liquids (Lescher et al. 1982) which reveal some contrasting depletion and enrichment trends compared to those of natural rocks. The most serious disparities in the experimental work on silica-rich melts (> 70 percent) vis-a-vis application to natural rocks such as the GLB are:

- 1) Mo, Nb, HREE and Mn migrate to zones of low SiO₂ in the laboratory experiments
- 2) enrichment factors for most major and trace elements vary considerably in natural rocks whereas there is little variation in the laboratory
- 3) the Na/K ratio varies significantly in many highly fractionated peraluminous granitic rock plutons such as the GLB whereas investigated melts show little variation.

The GLB shows enrichment in Rb, Na, Mn, Li, Nb and Ga and depletion in Al, Ba, Sr, K, Mg, LREE and RREE in late plutonic units relative to the early ones (Chapter 7). Only Al₂O₃ exhibits correlation with SiO₂ (Figures 39, 40 and 41).

Experimental work is clearly needed to evaluate the effect of granite compositional variation (peralkaline, metaluminous and peraluminous) on element distribution under TGD conditions. Until this is undertaken the influence of TGD in peraluminous granites such as the GLB cannot be properly evaluated. Nevertheless, this mechanism has attracted the attention of some pegmatite researchers. Cerny and Meintzer (1985, p. 39) suggested that "thermogravitational diffusion and/or other forms of liquid fractionation enhance the effects of advanced crystal/melt partitioning, and they promote the extreme accumulation of rare elements in the caps of pegmatitic granites and in the surrounding aureoles of mineralized pegmatites".

CRYSTAL-MELT FRACTIONATION

This magma compositional variation category includes fractional crystallization and fractional partial melting. Fractional crystallization is the model most widely applied to explain compositional variation in granitic rocks. The presence of fractional crystallization in various granitic suites in the literature is supported by field evidence of igneous sedimentation structures (mineral layering, cross-bedding and rhythmic variation in crystal size: Emeleus 1963; Wilshire 1978; Smith 1975) and by chemical variation. The latter is reflected in extensive trace element variation, most notably in Ba, Sr and Rb, termed "cryptic fractionation" by Robb (1986, p.280).

This variation is most effectively revealed in binary plots involving an element with a high bulk distribution coefficient (D) such as Rb plotted against low D elements (Sr or Ba). Linear trends

in log-log plots are characteristic. The modelling of fractional crystallization and of the batch partial melting mechanisms of Shaw (1970) by McCarthy and Hasty (1976), McCarthy and Robb (1978) and Robb (1986) have revealed contrasting trends:

- 1) steep negative trends of increasing Rb against increasing or decreasing Ba or Sr (fractional crystallization)
- 2) much flatter negative trends exhibiting increasing Sr or Ba with decreasing Rb (partial melting)

McCarthy and Robb (1978) applied their modelled in situ fractional crystallization trends of Na_2O , K_2O , Rb, Ba and Sr to the Nelspruit porphyritic granite of South Africa. They concluded that the cumulate character of this granite can be chemically discerned although with more scatter than the modelled trends possibly due to uncertainties in crystal/melt partition coefficients and to the inhomogeneous distribution of cumulus biotite. According to these workers precipitation of plagioclase, quartz and biotite causes progressive enrichment of Ba and Rb and depletion of Sr in the intercumulus melt; crystallization of successfully derived intercumulate liquids should yield Ba-Rb curves of positive slope.

Appearance of K-feldspar as a liquidus phase produces a marked inflection of Ba-Rb variation (steep negative slope) in the McCarthy-Robb model, as this mineral strongly removes Ba and Sr from the melt while Rb is preferentially partitioned into the intercumulus melt.

Application of the modelled element distribution patterns of these workers to the Ghost Lake Batholith infers that the "K-feldspar-plagioclase-quartz-biotite fractionation branch" of McCarthy and Robb

(1978, p.22)) was important. This is suggested by the close resemblance of the steep negative slopes of Ba-Rb and Rb-Sr distribution in the modelled plots compared to the Ghost Lake Batholith (Figures 51 and 52. There is continuous variation of these element pairs involving all plutonic units of the GLB with no notable chemical gaps. The study area plots also show an absence of Ba-Rb variation which suggests an absence of the plagioclase-quartz-biotite fractionation branch of McCarthy and Robb (1978).

Fractional crystallization apparently was able to effect maximum Rb concentrations of 430 ppm in potassic pegmatite and drastic depletions of Ba and Sr to less than 10 ppm. The more extreme Rb levels in rare-element pegmatites of the study area cause these compositions to fall off the GLB trend in Figures 51 and 52. This suggests that chemical evolution processes other than fractional crystallization (such as exsolution of an aqueous fluid) were important in the pegmatite genesis. Implications of the Ba-Rb and Rb-Sr variation will be discussed later in this chapter.

CONTAMINATION WITH EXTRANEEOUS CRUSTAL FLUIDS

Fractional partial melting can also produce compositional variation in anatectic magma. However progressive anatexis would expectedly lead to magmas with increasing mafic and decreasing felsic components (Winkler 1978). This should lead to the youngest rocks being the most mafic within a comagmatic complex produced by this process. This is opposite to that observed in the GLB. The effect of crustal contamination in the genesis of peraluminous granites has been most widely discussed by British workers (e.g. Simpson et al. 1979;

Plant et al. 1983). These geologists developed a model to explain metalliferous mineralization (Cu-Mo, Mo-W-F, Sn-Be-Li-U-K-Rb-Th-F) in granites of the Scottish Caledonides. The mineralization process commenced with interaction of metamorphic fluids and ultimately involved contact with circulating meteoric waters as the granitic magma rose to epizonal levels. These processes led to hydration of the magma, which modified primary isotopic compositions (e.g. $\delta^{18}\text{O}$ and initial $\text{Sr}^{87}/\text{Sr}^{86}$), and ultimately to hydrothermal alteration.

Extensive crustal contamination was inferred for peraluminous granites of the French Hercynian and the Himalayan orogenic belt (Allegre and Hart 1980). This is best illustrated in their $\text{Sr}^{87}/\text{Sr}^{86}$ vs $\text{Nd}^{143}/\text{Nd}^{144}$ plot in which these granites occupy distinct fields near the $\text{Nd}^{143}/\text{Nd}^{144}$ abscissa and far to the right of the mantle line (Allegre and Hart 1980, p. 142).

Much less work on the possible role of various crustal materials has been undertaken in the evaluation of Archean specialized granites. Cerny et al. (1987) suggested on the basis of isotopic data that the Lac du Bonnet Batholith originated as a hybrid partial melt formed from LIL-depleted tonalite and <25 percent greenstone belt metasediments.

Recognition of chemical signatures produced by interaction of granitic rocks of the study area with meteoric fluids is not tenable without requisite isotopic data and thus is beyond the scope of this thesis.

EXSOLUTION OF A SUPERCRITICAL FLUID PHASE IN GRANITIC MAGMAS

This process is presumed to be of importance in effecting petrographic and chemical variation in granitic magmas that have achieved water vapour saturation (Jahns and Burnham 1969). The Jahns-Burnham model contends that bubbles of exsolving supercritical fluid are a "selective scouring and collecting agent" at least for the alkali metals, SiO_2 and B_2O_3 as they rise down temperature gradients in the volatile-saturated rest magma (Jahns and Burnham 1969; Holloway 1976; Jahns 1982, p.315).

In a review of the many possible processes that may be operative in specialized granitic magmas Eugster and Wilson (1985, p. 95) consider the separation of a supercritical fluid to be "the most effect fractionation step" involving the halogens, Sn, W, Mo, As, Ta, Nb, Cu, Zn and Pb.

In the GLB, evidence supporting the former presence of an exsolved fluid phase is mainly restricted to the eastern lobe of the pluton containing the late units GLB-4 to GLB-8. The importance of an exsolved fluid phase in the evolution of these chemically fractionated units is suggested by the following summary of petrographic/petrochemical features:

- 1) quartz-rich pockets in potassic pegmatite with euhedral blocky K-feldspars and occasional beryl
- 2) rhythmic potassic pegmatite-aplite layering
- 3) metasomatic selvages in mafic metavolcanics at the batholith contact that are enriched in volatiles B_2O_3 , F, H_2O and CO_2 relative to unaffected host rocks

- 4) an overall increase in grain size from chemically primitive to highly fractionated plutonic units, sympathetic with increasing concentrations of the volatile-bearing minerals tourmaline and primary muscovite
- 5) the holmquistite alteration zone that is evidence of Li-F-B volatile discharge into mafic metavolcanic host rocks
- 6) presence of miarolitic cavities in some exocontact albite-rich dykes.

The role of supercritical fluid exsolution in the fractionation of the GLB and associated rare-element pegmatites will be examined further in this chapter in the discussion of the proposed model.

MODELS OF RARE-ELEMENT PEGMATITE GENESIS

INTRODUCTION

Several models of rare-element pegmatite petrogenesis have been proposed over the last fifty years. The purpose of this section is to review the essential attributes of genetic models of possible relevance to Archean terranes. Subsequently, a mechanism to explain the physical and chemical connection between peraluminous granitoid rocks and rare-element pegmatites in the Dryden area will be presented. For a more extensive literature review the reader is referred to Jahns (1955) and Cerny (1982 a,b).

Several general petrogenetic mechanisms have possible applicability to the derivation of rare-element-specialized granites and pegmatites. As summarized in Jahns (1955, p.1056-1067) and Cerny et al. (1981, p.144-147), these are:

- 1) fractionation of I-type juvenile magmas which may be contaminated by interaction with supracrustal rocks (Cerny et al. 1981)
- 2) partial melting of restite which resulted from the previous anatectic extraction of I-type granitic melts (Collins et al. 1982; Clemens et al. 1986; Whalen et al. 1987)
- 3) derivation of Li-rich granitic melts from early anatexis of metasediments and/or felsic volcanics (Norton 1973, 1981; Stewart 1978)
- 4) non-magmatic derivation of pegmatites: replacement pegmatites formed by lateral secretion or metamorphic differentiation (A.M. Smith in Roy et al. 1939; Ramberg 1949, 1952, 1956; Barth 1952; Reitan 1956, 1959; Kretz 1967)

- 5) anatexis of clastic metasediments followed by crystal-melt fractionation of peraluminous S-type magmas (Breaks et al. 1978; Drury 1979; Goad and Cerny 1981)

FRACTIONATION OF I-TYPE GRANITE MAGMAS AS A SOURCE OF PEGMATITES

Granitic masses belonging to the potassic plutonic suite constitute a significant part (45 percent) of the WRS of northwestern Ontario (Breaks et al. 1978) and adjacent Manitoba. They are exemplified by the Lount Lake, Tetu Lake and Gullwing Lake Batholiths.

Recently Beakhouse (1983, p.237-265 and in press) has chemically and isotopically characterized representative granites in the WRS and adjacent Wabigoon Subprovince. A three-stage model of magmatic evolution for the WRS was proposed (p. 261) whereby the potassic plutonic suite formed as I-type granitic melts representing the end-product in a 500 Ma history of Archean crustal evolution. These calc-alkaline granitic rocks were interpreted to have formed by partial melting of older crustal rocks of tonalitic composition. Because of their proximity to the Dryden Field (Figures 2 and 4) and the observation that plutons such as the Gullwing Lake Batholith contain internal and exocontact pegmatites (similar to Plate 1b) this group of granites must be considered as a potential source for rare-element pegmatites. This possibility is further suggested by the presence of Be-Sn-Nb-Ta-Y-REE-F-Zr-Th-U rare-element pegmatites within the biotite granite and in the exocontact zone of the Lac du Bonnet Batholith (Cerny et al. 1981, Cerny et al. 1987). Rocks of the potassic plutonic suite in Ontario are, however, less specialized in terms of pegmatite mineralization than the Lac du Bonnet Batholith; associated pegmatites contain only U-Th or Mo-Bi mineralization (Breaks 1982).

6

MICRO

Comparison of the largest I-type granite mass in the study area, the Gullwing Lake Batholith (GWLb), with the Ghost Lake Batholith and its rare-element pegmatite aureole (Mavis Lake group) reveals the following:

- 1) GWLB is metaluminous to weakly peraluminous relative to the more strongly peraluminous GLB (Figure 42)
- 2) the GWLB has some distinctly different accessory and varietal minerals (magnetite, hornblende and allanite)
- 3) there is different specialization in lithophile mineralization (U-Th in GWLB vs Li-Rb-Be-Ta>Nb-B in GLB).

Breaks (1982) showed that mineralized pegmatites of the potassic granitoid suite are not as extensively fractionated as rare-element pegmatites. His work revealed that the respective K-Rb, K-Ba and Rb-Sr fields do not overlap, which also suggests a genetic distinction. In the Gullwing Lake - Tot Lake pegmatite group, where the main parental granite source of the rare-element mineralization is unexposed, peraluminous granitoid dykes post-date those of the potassic plutonic suite. For the above reasons potassic suite rocks are rejected as a source for rare-element pegmatites in the Dryden Field.

Several plutons within the Dryden Field belong to the sodic plutonic suite, composed of trondhjemite, tonalite and minor diorite and quartz diorite (Breaks et al. 1978). These rocks are inadequate sources for rare elements as they never contain rare-element mineralization nor typically have associated chemically evolved pegmatites (Breaks and Bond, in press).

RARE-ELEMENT-ENRICHED GRANITIC MELTS FORMED BY PARTIAL MELTING OF I-TYPE RESIDUE

This petrogenetic mechanism has been proposed by Collins et al. (1982) as a means of forming anorogenic or A-type granites (as defined by Loiselle and Wones 1979). These granites have been theorized to originate from partial melting, under granulite facies conditions on the lower crust, of a residue which previously resulted from anatectic extraction of I-type melts (Collins et al. 1982). This model is experimentally supported by Clemens et al. (1986). As these granites can concentrate elements such as Li, Be, Ga, Nb, Sn, W, Zr, REE and F (Nurmi and Haapala 1986) they must be considered as a potential source of rare-element pegmatites. Such a genetic connection is suggested for the Pike Peak Batholith and internal Y-REE-Nb-F pegmatites of the South Platte district in Colorado (Cerny 1982a).

A-Type granites were initially inferred to occur exclusively in anorogenic rift settings such as in the Lachan Fold Belt of Australia (Collins et al. 1982). However, Whalen et al. (1987) more recently suggested a greater diversity of tectonic setting (subduction zones and transcurrent faults).

Compositions are dominantly metaluminous and peralkaline (White et al. 1982; Clemens et al. 1986), however peraluminous examples are known (Anderson et al. 1980; Anderson and Thomas 1985). The age of emplacement of A-type granites is dominantly post-Archean and they are particularly abundant, on a worldwide basis, between 170 Ma and 1300 Ma (Nurmi and Haapala 1986). White et al. (1982) suggested that Archean A-type granites exist, on the basis of geological setting and very limited petrochemistry (especially the high Ga/Al ratio relative

to LFB S- and I-type granites). They cite the Moolyella pluton, with exocontact cassiterite-tantalite pegmatites, in the Pilbara Province of Western Australia as an example. White et al. (ibid) suggested that fractionation of A-type metaluminous granites into peraluminous compositions is a possible process, that may best be recognized in areas where S-type granites are absent.

In the Superior Province of Manitoba Cerny et al. (1987) suggested that the eastern part of the 2.6 Ga Lac du Bonnet Batholith composed of highly fractionated leucogranite has A-type chemical characteristics (enrichment in Zr, REE, Y, Th, U, F, Be and Nb>Ta).

Peraluminous granites and pegmatites of the study area do not however have chemical or petrographic features indicative of an A-type granite; the low REE, Y and Zr contents are especially distinctive.

Al/Ga ratios, supposedly a distinctive feature of A-type granites (Collins et al. 1982), are similar in the GLB and A-type suites from the LFB (Mumbulla and Gabo suites, White et al. 1982). It is possible, however, that partial melting of a lower crustal zone containing a high amount of S-type restite accumulated through a previous large scale anatectic event could result in A-type magmas.

DERIVATION OF LITHIUM-RICH GRANITIC MELTS FROM EARLY ANATEXIS OF SUPRACRUSTAL ROCKS

This model was proposed by Norton (1973, 1981) and advocated by Stewart (1978) as a mechanism capable of generating Li-rich melts directly by anatexis. As such it represents an alternative to lithium concentration by crystal-melt fractionation in a parental magma, which these workers felt was an ineffective means to produce the Li levels

necessary to initiate spodumene crystallization in consanguineous rare-element pegmatites. They pointed out that the major factor militating against fractional crystallization as an effective Li-concentrating process is the disparity in concentration factor (of about 70 times) between the most fractionated, potential parental pegmatitic granite unit and the nearest spatially related spodumene- or petalite-bearing Li-rich pegmatite. This contrast is also marked, but less extreme, between the GLB and Mavis Lake pegmatite group; fractional crystallization would be required to concentrate Li 43 times between the most fractionated plutonic units (198 ppm) and the nearest spodumene pegmatite (8600 ppm in Fairservice No. 1 pegmatite).

According to this model, Li-rich pegmatitic melts are syn-metamorphic and form directly as an early product from anatexis of Li-rich metasediments. Ensuing progressive anatexis would produce larger volumes of melt, considered by Stewart (1978) to be represented by the spatially related parental granites with more dilute Li concentrations (<100 ppm).

This theory is in clear discord with the known chemistry of lithium in anatectic environments involving clastic metasediments (Chapter 6). The early Li-rich pegmatites envisaged by Stewart (1978) would, according to partial melting experiments (Winkler 1978), necessarily be water-saturated, relatively low temperature minimum melts formed in muscovite-biotite metapelites by muscovite dehydration. Such granitic melts are probably exemplified by Plate 20a and likely would be Li-poor as available analyses indicate that most Li is contained in the biotite-rich restite (see discussion on Li

PLATE 20

- a. Birth of an S-type granite - podiform granite enclosed entirely in a porphyroblastic garnet-cordierite metapelite from the sillimanite-potassium feldspar zone in the Wegg Lake area of the ERS (Latitude 50°34'50" and Longitude 93°33'20'). Note the large garnet (G) in the process of engulfment as a xenocryst in the granite. Biotite-rich restite = λ , cordierite with intergrown sillimanite = C.

PLATE 20

partitioning in migmatitic components, later in this chapter). In the high grade sillimanite-K-feldspar zone of the ERS, Li levels are low in granites plausibly representing minimum melts (7 to 45 ppm and averaging 15 ppm; 25 analyses in Breaks and Bond, in press).

Support for this model additionally suffers from a paucity of Li abundance data from metasediments in rare-element pegmatite districts and also from an absence of estimates as to what level of Li would be required for the direct anatectic segregation of Li-rich pegmatites.

Lastly, it is known, from localities where parent granite-rare-element pegmatite relationships are evident, that such pegmatites are late. Rare-element-specialized peraluminous granites such as the Leinster Batholith of Eire (Steiger and von Knorring 1974, Moore-Lewy 1986) and the MNW Stock in the Georgia Lake Field near Lake Nipigon (Breaks 1980) contain internal, Li-rich pegmatites, suggesting that such pegmatites segregated from the parental granite at a late stage.

In the GLB early segregation of Li-rich melts according to Stewart's model should be observed in the GLB-1 unit having the earliest relative age and which best approximates minimum melt composition. The host metasediments enveloping GLB-1 also contain the highest Li contents observed in the study area. However, spodumene pegmatites are unknown in such rocks lying adjacent to the western end of the GLB.

Cerny et al. (1981, p. 145) also rejected this model, on the basis of the pronounced disparity between REE patterns in the Bird River greenstone belt metasediments and nearby pegmatitic granites.

However, these granitoid rocks probably represent the upper parts of highly fractionated peraluminous magmatic systems and hence may not reveal a complete comagmatic series extending to deeper plutonic levels, as seen to advantage in the GLB. The inferred lower levels of the batholith (GLB-1) do have REE patterns similar to the Zealand metapelites (Figures 37 and 74). The use of REE as a means of rejecting this model is thus not entirely valid.

However, for the reasons discussed in this section, a model invoking an early anatectic derivation of lithium-rich pegmatites from similarly enriched metasediments is not of practical application in the study area. This model may be relevant to the anatexis of supracrustal sequences containing units anomalously enriched in Li (e.g. evaporites and rare-element rhyolites). Such a case, however, has not to the writer's knowledge been documented in the literature.

REPLACEMENT PEGMATITES FORMED BY LATERAL SECRETION (METAMORPHIC DIFFERENTIATION)

This model involving non-dilational replacement is a viable means of forming some granitic pegmatites. Kretz (1967) provided compelling evidence for the lateral secretion formation of small biotite-plagioclase-quartz-K-feldspar pegmatites in the Northern Indian Lake of Manitoba. This process apparently involved a solid-state diffusion of components from the tonalitic host to form K-feldspar-rich pegmatite. The important evidence supporting the model favoured by Kretz (1967, p.30-41) includes:

- 1) continuity of host rock foliation and inclusions through the pegmatites indicating that no dilation and injection of pegmatite material occurred,
- 2) presence of mafic selvages along the pegmatite-host rock contact which exhibit depletion in minerals which are enriched in the pegmatite (K-feldspar, quartz and plagioclase), and
- 3) similarity in plagioclase and K-feldspar compositions.

This model is not applicable to rare-element pegmatites of the Dryden area as the great majority of these occur in compositionally unfavourable host rocks (mafic metavolcanics) that are deficient in several components (K_2O , SiO_2 , Li and Be) necessary to form important pegmatite minerals. In addition the selvages of metasomatized mafic metavolcanics as exemplified at the North Zone of the Gullwing Lake dyke (Figure 24, Plate 15d) can exhibit enrichment of albite rather than its depletion as would be expected in the lateral secretion model. Furthermore there is evidence for dilational emplacement of many rare-element pegmatites in the study area which will be discussed below.

The process may possibly have application to rare-element pegmatites hosted in granitic rocks of appropriate composition, i.e. rare-element-enriched granites; the writer knows of no documentation of pegmatites derived by lateral secretion from such granites.

PETROGENESIS OF THE PERALUMINOUS S-TYPE GRANITE-RARE-ELEMENT PEGMATITE ASSOCIATION IN THE DRYDEN FIELD

Introduction

Unequivocal examples of a parental granite-rare-element pegmatite lineage are not common in the Canadian Shield. Furthermore, parental granites that are exposed generally represent an amalgamation of chemically evolved, pegmatitic granite units (e.g. Goad and Cerny 1981, p. 189). In contrast, asymmetrical chemical and lithologic variation described in the GLB suggests that plutonic levels are exposed beneath the more highly fractionated zone containing most of the late plutonic units GLB-4 to -8. If correct this presents a unique opportunity to trace the fractionation of rare elements and associated volatiles upwards within the GLB. This section will deal with the origin and evolution of this peraluminous granite-pegmatite system.

Advanced Crystal-Melt Fractionation of Anatectic Melt Followed by Separation of a Lithium-Enriched Pegmatitic Magma-Volatile System

This petrogenetic mechanism, as proposed here to explain the origin of the Ghost Lake Batholith and spatially associated rare-element pegmatites, is a modification of the concept originally advanced by Breaks et al. 1978. The model consists of five stages as outlined below and detailed in Figure 99.

1. extensive anatexis of metapelite and metawacke
2. unmixing of restite in a rising anatectic magma
3. fractional crystallization
4. exsolution of a supercritical aqueous fluid in the parent granite

Figure 99. Flow Chart Depicting Origin and Evolution of the Peraluminous Granitoid-Rare-Element Pegmatite Association in the Dryden Area.

5. emplacement and chemical evolution of exocontact, rare-element pegmatites.

Although the model is complex, it is supported by field relationships and by chemical data of clastic metasedimentary and peraluminous granitoid rocks of the Dryden area and the ERS. Its complexity may also relate to the protracted time interval governing the chemical evolution of these peraluminous granitoid rocks (possibly up to 120 Ma; see above).

Initial concentration of the rare elements, B and F was effected by clay minerals during the development of regionally extensive sedimentary basins. Such basins in the study area underwent sedimentation after 2713 ± 2 Ma ago (Davis et al. 1988). High temperature-low pressure regional metamorphism of these clastic metasediments occurred at 2.68 Ga. High grade metamorphism (to granulite grade at depth) led to the decomposition, in metapelites and metawackes, of muscovite followed by biotite with the consequent production of copious volumes of S-type peraluminous granitic melts. The dehydration of the micas probably occurred under fluid-absent conditions (Burnham 1967) whereby all water required for anatexis was provided by these minerals. There is no field evidence supporting vapour present melting, anatexis in which water is introduced into the zone of anatexis from an external source.

The necessity for biotite dehydration under granulite conditions in the fluid-absent partial melting scenario is supported by the work of Miller (1985). He estimated that a metapelitic source rock containing 20 to 30 percent muscovite cannot, under high grade

conditions (i.e. below the biotite dehydration curve), produce more than 10 to 15 percent granitic melt. Metasedimentary migmatites with similarly low amounts of in situ granitic leucosome are widespread in the sillimanite-K-feldspar zone of the ERS (high grade zone of Figure 8) and in the WRS just west of the GLB. These migmatites are interpreted by the writer to essentially reflect anatexis induced by muscovite breakdown (Plates 3d and 20a). If most in situ leucosome originated by anatexis, the low volumes characteristic of this zone (less than 15 percent, Breaks and Bond in press) are considerably below those values considered by Arzi (1978) and van der Mollen and Patterson (1979) to be required for the formation of granitic plutons and their subsequent mobilization above the zone of anatexis.

The strong spatial association between peraluminous granitoid plutons and metasedimentary migmatite belts in the northwestern Superior Province suggests the possibility of two anatectic processes:

- 1) fluid-present partial melting of muscovite-bearing metapelite and metawacke (Miller 1985);
- 2) fluid-absent partial melting (Burnham 1967) of muscovite-biotite metapelite and metawacke in which water is derived by dehydration of muscovite succeeded by biotite (Miller 1985) induced by a metamorphic P-T gradient passing through amphibolite grade to granulite conditions.

Clemens and Vielzeuf (1985), Vielzeuf and Holloway (1988) and LeBreton and Thompson (1988) contend that fluid-present anatexis is unlikely to produce large volumes of granite especially in the light of the limited upward crustal mobility of H₂O-saturated, "wet"

granitic magmas (see also Cann 1970). This process may, however, be a factor if water from an external source can be introduced into the partial melting zone (Wickham 1987) through shear zones, (Strong and Hanmer 1980) or continental collision zones the latter involving rapid, tectonically-induced introduction of wet sediments to the zone of anatexis (Le Fort 1981; England and Thompson 1984).

The greatest proportion of peraluminous granitoid rocks in the ERS is concentrated within and adjacent to granulite metamorphic zones (Figure 3). Biotite dehydration probably played a key role in the melting process as it is by far the most abundant hydrous mineral in the metasediments of these zones (Breaks and Bond, in press). Amphibole dehydration does not represent an important source of water because it is restricted, in the ERS granulites, to rare amphibolitic units (Breaks 1988).

The anatectic process released rare elements and the volatile-complexing elements boron and fluorine, principally from micas in the metasedimentary protolith, into potassium-rich peraluminous granitic melts, represented by plutonic unit GLB-1. Significant amounts of primary restite (Chappell and White 1988), enriched in biotite, cordierite and possibly orthopyroxene, remained incorporated in the anatectic melt. As the melt volume surpassed the critical melting percentage (Arzi 1978, van der Mollen and Patterson 1979) the gravitationally unstable mass underwent buoyant emplacement to a higher crustal level. Emplacement was possibly effected as crystal-silicate melt mixture which post-dated the peak of regional

metamorphism, as testified by its discordance to the boundary between medium and high grade host rocks.

Fractionation of Li and associated rare elements in peraluminous, restite-bearing S-type granitic magmas during their ascent above the zone of fusion is in general poorly understood. Logically the amount of Li entering the melt phase should depend upon the protolith bulk composition, the intensive variables controlling anatexis and, in particular, the restite mineralogy. Anatexis of metasediments in the granulite zone under Abukuma-type metamorphic conditions would allow any of the following minerals as restite phases: orthopyroxene, cordierite, biotite, garnet, sillimanite and plagioclase. Modelling of Li fractionation between primary restite and leucosome, however, is hampered by a lack of knowledge of the relevant distribution coefficients. Some idea of lithium distribution between co-existing leucosome and restite can be obtained from low leucosome-bearing metasedimentary migmatites in the sillimanite-K-feldspar metamorphic zone of the ERS as exemplified in Plate 20a. Breaks and Bond (in press) report that K-feldspar-sillimanite-plagioclase-biotite restite contains more than seven times the lithium concentration of that in the leucosome (restite = 51 ppm, leucosome = 7 ppm). It follows from this very meagre data that restite produced under conditions of incipient anatexis has a high capacity to retain lithium, probably because of its high biotite content.

Fluid-absent melting of biotite-rich metapelite under Abukuma granulite conditions will produce large volumes of granitic melt (Vielzeuf and Holloway 1988) but likely will leave a primary restite

with less biotite than under high grade conditions. This would qualitatively allow more Li to enter the melt phase. Cordierite may also play an important role in Li fractionation, as this mineral is ubiquitous in peraluminous granitoid rocks of the ERS (Breaks and Bond, in press) and has some capacity to accommodate Li (Table 53).

A further factor that may have important implications on Li contribution to the melt phase is the possible mineralogical modification of primary restite to secondary restite as the crystal-silicate melt ascends to higher crustal levels. Secondary restite, as defined by Chappell et al. (1988) is a chemical/mineralogical modification of the restite which was initially produced in the anatectic magma source region. This transformation could be effected through volatile interaction with restite biotite and cordierite, inducing replacement by muscovite. Such a reaction plausibly could release Li to the melt as muscovite is not able to retain as much Li as biotite (Chapters 6 and 8).

The possible role in the Li melt budget of the GLB involving secondary restite arising from muscovite replacement needs a thorough quantitative assessment which is beyond the scope of this thesis.

It has been inferred from chemical and petrographic data that crystal-melt fractionation played a role in the evolution of the GLB. This process was partly responsible for the notable chemical asymmetry in the GLB (Figures 49, 50 and 78), which is characterized by enrichment in Na₂O, B₂O₃, Mn, H₂O, F, Cs, Rb, Li, Ga, Be and Sn and depletion of CaO, K₂O, MgO, total REE, Ba, Sr, and Zr towards the interpreted cupola zone at the eastern end of the batholith.

The most compelling chemical evidence for fractional crystallization involves the previously mentioned strong linear continua for Ba-Rb and Rb-Sr (Chapter 7). These trends are similar to those of perfect fractional crystallization as modelled by McCarthy and Hasty (1976) based upon cumulus-intercumulus phase relationships in granites of the Bushveldt Complex. They inferred (p.1356) that the Main granite, the Bobbejaankop granite (cassiterite and fluorite-bearing) and the aplitic Lease granite "formed by in situ fractional crystallization of a single parent magma" as based upon congruent data population in the Ba-Rb and Rb-Sr diagrams. However the Rb (750 ppm), Ba and Sr levels in the natural examples differ from those suggested by the models involving perfect fractional crystallization. This observation led these workers to suggest that a significant amount of trapped intercumulus liquid was responsible for the deviation. The overall similarity of the Bushveldt Complex granites with the GLB trend in the Ba-Rb and Rb-Sr diagrams suggests that a separation of cumulus phases from intercumulus melt may have occurred in the GLB. Most rare-element pegmatites and with some barren, exocontact pegmatites plot off the GLB trend and towards Rb-enrichment (Figures 51 and 52). This suggests that processes other than perfect fractional crystallization may have been operative in their petrogenesis. Increasing the amount of intercumulus melt does not seem to adequately account for the bulk and zonal enrichment of Rb (310 to 4510 ppm) because this, according to McCarthy and Hasty (ibid), tends to decrease the Rb content and extent of Ba-Sr depletion in the final crystallized mineral assemblage.

Na₂O and K₂O are the only major element pairs that exhibit a negative linear correlation similar to the trace element pairs Ba-Rb and Rb-Sr (Figures 44, 51 and 52). Na₂O variation in the Harker diagrams (Figure 39) is less regular. While GLB-1 exhibits trends of slightly decreasing Na₂O or K₂O with increasing SiO₂, the remaining fields (GLB-2, GLB-3 and a composite of GLB-4 to -7) lie mostly outside GLB-1. The Na₂O vs SiO₂ plot, in particular, reveals an abrupt change in slope marked by increasing Na₂O levels from GLB-1 --> GLB-2 --> GLB-3 --> GLB-4. This trend may also reflect the increased influence of processes other than fractional crystallization in the chemical evolution of the GLB. For example, increasing Na₂O could be a function of increasing fluorine in residual melts which "ponded" in the apical zone or may be related to late or post-magmatic autometasomatism. These processes will be examined below.

Behaviour of volatiles in the late- to post-magmatic history of the Ghost Lake Batholith: the fourth stage of the proposed petrogenetic model commenced during the late magmatic history of the GLB (Figure 99). This involved collection of residual pegmatitic granitic melts accompanied by an exsolved aqueous phase along the margins of an inferred roof zone in the eastern lobe of the GLB, which possibly included the Zealand Stock. The apical zone is now represented by the youngest plutonic units which consist of a diverse, late-tectonic admixture of pegmatitic leucogranite, fine-grained leucogranite, potassic pegmatite and layered aplite.

By virtue of increased concentrations of fluxing components (H₂O, B₂O₃, F and possibly Li₂O) the pegmatitic granite melts

experienced a retardation in crystallization due to significantly depressed liquidus and solidus temperatures (Bankwitz 1974) and a decrease in melt viscosity with depolymerization (Burnham 1979). London et al. (1988) has documented a liquidus temperature of 650° C (at vapour saturation) and a solidus temperature of 450° C in fractional crystallization experiments with macusanite.

Crystallization of GLB-4 to -7 thus outlasted that of GLB-1 to -3. Presence of volatile components such as F and B₂O₃ may also have importantly shifted residual magma compositions towards increasing Na/K ratios (see Chapter 7).

Autometasomatic Processes in Pegmatitic Granites: these processes played a role in effecting petrological and chemical diversification in some of the pegmatitic granites in the Dryden Field. These processes include albitization and local greisenization; microclinization, which is a common alteration type in many apogranite complexes (Beus and Zalashkova 1964), was not observed. In the GLB and related exocontact dykes such as the Contact Beryl Occurrence (Chapter 3), timing of the replacement is interpreted as have occurred after crystallization of GLB-5, -6 and -7. This inference is based upon textural evidence, as exemplified in Plate 8c, where ragged replacement of primary minerals such as K-feldspar by albite is obvious in the transformation of potassic pegmatite into tourmaline albitite. Exocontact dykes which can be traced into pegmatitic potassic pegmatite and/or fine-grained leucogranite reveal complete replacement by albitite or muscovite-rich greisen rock (Plates 8c and 8d).

Timing of the autometasomatic replacement with respect to tectonic events is somewhat variable in the study area. At the Contact Beryl Occurrence these alteration assemblages post-date regional deformation structures such as mineral lineations but pre-date late-tectonic boudinage structures (Plate 9f). In contrast, autometasomatism of the Gullwing Lake dyke is post-tectonic (Chapter 4).

These processes are important in that mobilization of rare elements and volatiles such as F and B may result from exchange reactions between a supercritical aqueous fluid and pre-existing primary minerals (Eugster 1985). Both albitization and greisenization released such elements to the aqueous phase as a result of the destruction of K-feldspar, biotite, muscovite, plagioclase and less common minerals such as columbite and beryl.

Release of rare elements such as Li via autometasomatism may have a direct bearing on the appearance of spodumene in certain pegmatitic granites. Critical field evidence for this is exposed in the Gullwing Lake dyke. Textural and chemical relations suggest that post-magmatic Na-Li-F autometasomatism occurred in the muscovite-biotite pegmatitic granite unit, revealed by muscovitization of biotite and albitization of blocky K-feldspar (Plate 14c).

In parental pegmatitic granites there is an apparent upper concentration limit of the lithium that can be retained by the melt phase. Cerny (1982b, p. 422) pointed out that low Li contents (<100 ppm) generally prevail in most pegmatitic granites; those containing Li minerals are rare (Chackowsky and Cerny 1986). Extremely high

concentrations are necessary to initiate crystallization of independent Li minerals (e.g. 5000 ppm Li for spodumene, Solodov 1966). Concentrations of Li below this value can only achieve isomorphous entry into K-feldspar or accessory mica and tourmaline. It follows that most Li was probably partitioned as an incompatible element into a fluid phase coexisting with pegmatitic melt. Therefore, some process in addition to fractional crystallization must assume prominence in order to transcend the observed lithium concentration discontinuity between the batholith and spodumene pegmatite.

The Li-rich siderophyllite variety of biotite was established in Chapter 8 to be present in the pegmatitic granite unit and in some of the albitization I units enveloping the muscovite-spodumene-blocky K-feldspar-quartz core zone. Progressively increasing replacement of siderophyllite by muscovite is most obvious within the albitization I zone (Plate 14 , Figures 25 and 26). The capacity of primary muscovite to isomorphously capture Li is considerably less than that of siderophyllite, as revealed in the analyses of coexisting primary muscovite and biotite from beyond the alteration zone. It follows that significant amounts of Li, Rb, Cs, Nb, Ta and F were released by the development of secondary albite and muscovite as there is not the large increase in secondary muscovite required to isomorphously accommodate these elements. The ultimate destination of these elements may then have been the late, silica-rich core zones formed at higher levels in the pegmatite magma chamber, where Li attained the concentrations necessary in a fluid phase to crystallize spodumene.

The post-magmatic "reprocessing" event has been suggested as a factor in the evolution of Li-F apogranites in several areas as, for example, by Beus (1982) and Rossofskiy and Mogarovski (1988). However, Archean representatives of such rocks have not been previously recognized.

Emplacement and chemical evolution of rare-element

pegmatites: this represents the final, fifth stage in the proposed model involves migration of pegmatitic melt-fluid systems from the parent granite roof zone into structurally favourable sites in the supracrustal host rocks. After crystallization of pegmatitic granite units GLB-4 to -7, a residual pegmatite melt coexisting with fluid remained. It outlasted the consolidation of these units on account of higher concentrations of solidus-depressing components (F, B and Li) as well as other rare elements. The silicate melt and co-existing fluid phase collected in apical protuberances of the batholith surface. These highly fractionated systems were subsequently transferred along conduits in the host rock to amenable structurally dilatant zones above the apex of the batholith.

Several field observations which bear evidence of a dilational accommodation of the rare-element pegmatites and pegmatitic granites in the Dryden area are summarized below. A good documentation of criteria allowing discrimination between pegmatites formed by replacement and those produced by dilation of host rocks by the injection of pegmatite material is given by Kretz (1968), based upon his work in the Yellowknife Pegmatite District. A few pegmatites of the study area share similar structural features with those documented by Kretz (1968, p.39-57; asterisks below):

	<u>Taylor</u>	<u>Fairservice</u>	<u>Gullwing dyke</u>	<u>Tot</u>
1. V-shaped protrusion pegmatite into host rocks*	X	X	X	X
2. stoped host rock inclusions			X	X
3. orientation of elongate minerals normal to wall rocks (comb structure)		X		X
4. layering approx- imately normal to wall rocks				X
5. schist bridges*			X	

A feature in apparent conflict with dilation and injection of pegmatite is obvious in Figure 24 at the North Zone of the Gullwing Lake dyke. Host rock inclusions exhibit no evidence of rotation as evidenced by parallelism of their mineral lineations with those of the wall rocks. This relationship could be explained by considering the inclusions to be screens which maintained attachment to the wall rocks at depth during pegmatite emplacement and hence were not forcibly rotated. Other features at this pegmatite suggest dilation-injection: "V"- to "U"-shaped protrusion of pegmatite into the wall rocks and abrupt truncation and 17 m separation of a pre-pegmatite granitic dyke along the east side of the North Zone (unit 5a in Figure 24).

The exodus of a melt-fluid phase from the region of autometasomatized pegmatitic granites of the GLB is demonstrated by the holmquistite-bearing alteration zone described in Chapter 3. This zone may represent the actual locus of disengagement of the melt-fluid phase system from the roof zone of the batholith en route to an evolving exocontact pegmatitic aureole. Evidence for the former

presence of a silicate melt accompanying a fluid phase is provided by occurrence of small exocontact pegmatitic granite masses within the alteration zone, as at the Cointact Beryl Occurrence (locality B in Figure 14). Such a zone containing holmquistite and rare-element pegmatites has, never, to the author's knowledge, been previously documented in any other rare-element pegmatite field. As such it may represent the "missing link" in Li abundance between a parent granite and related Li-rich pegmatites.

The regional zonation in the Mavis Lake group, described in Chapters 3 and 5, is a reflection of vertical fractionation of rare-elements as the melt-fluid systems migrated to increasingly higher levels along the conduit structures. Beryl-bearing pegmatitic granites at the Contact Beryl Occurrence (locality B in Figure 14) represent the lowest level and least fractionated rare-element mineralization (Be, Nb>Ta and Sn) with respect to the batholith roof surface. Higher level pegmatites (North Zone of the Fairservice property) are characterized by a different chemical specialization (Li, Rb, Cs and Ta>Nb) which is distinguished by appearance of spodumene and tantalite. F and B show no apparent concentration gradients and are anomalously abundant in all the rare-element pegmatites, particularly in their exomorphic aureoles and in the scheelite-tourmaline sheets on the Petrunka property.

Crystallization in the upper parts of the regionally zoned Mavis Lake group and the Gullwing-Tot Lakes group is represented respectively by Fairservice No. 1 pegmatite and the Tot Lake pegmatite. The former is inferred to have a consolidation history involving a transition from fractional crystallization to precipitation of minerals from a

supercritical fluid phase. The Tot Lake pegmatite also has a complex crystallization history, however it differs from the Fairservice No. 1 pegmatite in the presence of rhythmic aplite-pegmatite layering and a more restricted development of a silica-rich supercritical fluid phase.

Fairservice No. 1 Pegmatite: is inferred to have formed by fractional crystallization only in the K-feldspar-rich spodumene zone. Relatively coarse, coalescing K-feldspar crystals crystallized from a potassic melt similar in bulk composition to potassic pegmatite the GLB except for higher Li_2O (Table 33 and Appendix 7). This was followed by a much finer-grained, interstitial intergrowth of spodumene and quartz.

The adjacent quartz > spodumene >> albite and K-feldspar zone is contrastingly depleted in feldspars. It is inferred to have formed from supercritical aqueous fluid containing predominantly SiO_2 , Al_2O_3 and Li_2O . This is suggested by the high SiO_2 contents (77.8 to 82.4 percent, Table 33; silica is highly enriched in the exsolved gas phase according to the experimental work of Stewart 1978, p.978) and by Li_2O concentrations of 1.2 to 4.0 percent as calculated from modal spodumene in Table 10 and from analyses in Table 33. Further support for precipitation of the quartz-spodumene-albite zone from a silica-rich aqueous fluid is inferred from phase relations in the synthetic albite-quartz-eucryptite- H_2O system of Stewart (1978). As previously indicated in Chapter 5 Stewart (ibid) determined that 2.2 percent Li_2O represents the maximum that can concentrate in a melt by fractional crystallization which commences in the feldspar field or Li-poor part of the quartz field. Significantly higher Li_2O concentrations in many

analyses of the quartz-spodumene-albite zone of the Fairservice property and require precipitation from the agency of an aqueous fluid. Analyses such as Fairservice 169 and 171 (Table 33) plotting in the quartz field of Figure 64 well removed from the ternary eutectic, further suggest the agency of a silica-rich aqueous fluid. Deposition of the quartz-rich core zone which represents the third primary unit, is a logical extension of the same process which produced the adjacent quartz-spodumene-albite zone, as indicated by identical mineralogy and a gradational contact. However, quartz exhibits a substantial increase from 28-47 percent in the quartz-spodumene-albite zone to 90-95 percent in the adjoining core unit. In the Jahns-Burnham model exsolving bubbles of supercritical aqueous fluid can selectively accumulate SiO_2 , alkalis and B_2O_3 as they separate from the residual magmas. Crystallization of these components normally produces quartz, K-feldspar, muscovite and tourmaline in the core zones (Jahns 1982). Given appropriate concentrations of Li and Be this zone will include spodumene and primary beryl as is found at the Fairservice pegmatite No. 1. London (1988) recently suggested alternatively that exsolving aqueous fluid may actually increase the Li, Rb and Cs contents of the rest magma by extensive removal in the vapour of the major components such as K_2O and Na_2O .

Tot Lake Pegmatite: is one of the most fractionated igneous systems known in the Superior Province and is comparable in most chemical attributes with the Tanco Pegmatite of southeastern Manitoba (Cerny et al. 1981). Both pegmatites contain pollucite, a mineralogical indicator of extreme fractionation; however in the case of the Tot Lake Pegmatite this mineral coexists with primary Mn-enriched spodumene rather than that of secondary spodumene Mn-rich spodumene with its characteristically negligible Fe_2O_3 as in the Tanco Pegmatite (Cerny and Ferguson 1972) and the MNW Pegmatite (Figure 94).

Very few pegmatites of comparable size reveal such a complexity of internal zones and overprinting albitic replacement units as does the Tot Lake pegmatite. Of particular petrogenetic interest are the layers exhibiting high spodumene concentrations. These layers occur in two forms as summarized from Chapter 4:

- 1) longitudinally-oriented spodumene-rich pegmatite, interlayered with aplite, which ultimately grades into the pollucite-spodumene-blocky K-feldspar-albite-quartz zone
- 2) transverse-oriented, smaller scale layers of spodumene-K-feldspar albite pegmatite and muscovite-albite aplite

The high spodumene concentrations in the longitudinal zones of the Tot Lake Pegmatite are far above the maximum spodumene content of 27 percent that can accumulate by fractional crystallization (Figure 64). In addition, the bulk composition of an aplite-pegmatite pair, which was not analyzed but must lie along the tie line in Figure 64, is notably removed from the eutectic.

Both types of layered zones are best explained in the context of the Jahns-Burnham model as products of the resurgent boiling of a Li-enriched melt. Substantial amounts of Li can be partitioned into the exsolved supercritical fluid phase such that the usual order of alkali concentration in the fluid, i.e. $\text{Na} > \text{K} > \text{Li}$ (Stewart 1978, p.978) is reversed, i.e. $\text{Li} > \text{Na} > \text{K}$. This process led to narrow layers of spodumene (77 percent) and Li_2O (5.27 percent) contents (Figure 30), among the highest of any known pegmatite zone in the world. The rhythmically layered aplite-pegmatite, oriented at a high angle to the pegmatite contact contains the largest spodumenes of any pegmatite in the study area occurring as randomly-oriented megacrysts in a finely-layered matrix. The matrix is composed of many 1-3 cm wide layer couplets of spodumene-rich pegmatite and aplite (Plate 17a). These observations suggest that the melt achieved a concentration of lithium sufficient to precipitate large spodumene phenocrysts at an early stage.

The gradation from pegmatite characterized by randomly oriented spodumene (Plate 18a) intergrown with albite, K-feldspar, quartz and beryl into the transverse-layered unit is marked by a significant decrease in the large bladed spodumenes. The spodumene phenocrysts diminish in abundance away from the area of highest concentration to Trench 3 (Figure 30; Plate 18a), where they are more sporadically distributed and completely enveloped by the rhythmic aplite-pegmatite layering (Plate 16e).

This change is interpreted as a commencement of episodic or rhythmic resurgent boiling, effected by repeated segmentation of a

volatile-saturated, Li-rich melt into spodumene pegmatite (variably enriched in Li, Na and K) representing the aqueous fluid and muscovite-albite aplite representing the volatile-depleted residual melt. This process ended with fracturing of the pegmatite roof due to volatile pressure build-up which led to the subsequent escape of the silica-rich aqueous fluid. The fugitive aqueous fluid crystallized to form the central quartz vein network (Plate 16b) which not only cuts across transverse-layered unit but also has imposed albitization upon adjacent primary units.

The Jahns-Burnham model, although applicable to the Tot Lake complex pegmatite, was evidently operative here at much lower SiO_2 contents than those which normally are associated with silica-enrichment trends in the exsolved vapour phase (Jahns and Burnham 1969; Stewart 1978). This is shown by the transverse-layered zone where quartz contents are very low (one to four percent, Table 12) in the pegmatitic layers.

The final primary zone to crystallize in the Tot Lake Pegmatite is characterized by abundant pollucite. Despite its proximity to the longitudinal pegmatite-aplite zone, the pollucite zone does not contain textural evidence of resurgent boiling. Alternatively, it possibly represents a rest melt which became extremely enriched in cesium after significant amounts of lithium were extracted by a previous period of resurgent boiling. This trend is supported texturally by a well defined paragenetic sequence featuring late fractional crystallization of pollucite (Chapter 4; Plate 18c. This

is supported by a bulk composition which plots near the ternary eutectic in the normative albite-quartz.

Cesium strongly partitions into the melt phase and its concentration there is enhanced in the presence of a high fluorine content (Cerny et al. 1985). Cesium concentration in the pollucite zone may thus have been initially enhanced by fluoro-complexes, suggested by the early crystallization of fluorapatite (see Chapter 4).

Post-magmatic replacement in rare-element pegmatites and exomorphic dispersion: the final aspect of pegmatite evolution in the Dryden Field was dominated by replacement processes (albitization, greisenization and lepidolitization). This occurred at the post-magmatic stage as indicated by textural evidence. It is unlikely that the post-magmatic alteration was synchronous with its counterpart in the Ghost Lake Batholith (see above). The probably much lower solidus temperatures in the rare-element pegmatites due to significantly higher Li_2O contents (see Chapter 7) implies that the GLB passed into the post-magmatic stage while Li-enriched pegmatites in the exocontact aureole were still unconsolidated.

Replacement commenced with the reaction of a silica-rich aqueous phase which induced albitization in all primary pegmatite units. This is typified by marginal replacement of spodumene megacrysts by albite in the quartz-rich core units (Plate 12a). Ultimately, virtually complete replacement of such spodumenes resulted (Plate 19a).

Intervention of a vapour phase rich in SiO_2 and NaO_2 may also explain

the commonly observed restriction of albite-replacement zones of the margins of the quartz-rich core units.

Post-magmatic replacement of primary lithium- and potassium-bearing minerals resulted in the egress of K, Li, Rb, Cs, B and F, which largely were incompatible with respect to the albite-quartz-dominated replacement assemblage. Considerable amounts of such elements were thus transferred via the fluid phase into mafic metavolcanics where chemical and mineralogical (biotite, tourmaline and holmquistite) anomalies were developed.

Chapter Ten

SUMMARY AND CONCLUSIONS

Peraluminous, S-type granitoids are ubiquitous in parts of the Superior Province in northwestern Ontario. Although studies of the origin and evolution of peraluminous granitoid rocks have been extensive in recent years, most work has been focussed upon Phanerozoic examples and little has been known of the Archean representatives.

This study describes the geology, petrography, rock and mineral chemistry and origin of an Archean peraluminous granitic pluton, the Ghost Lake Batholith, and two nearby rare-element pegmatite groups (Mavis Lake group and Gullwing-Tot Lakes group) in the Dryden area of north-west Ontario. The area affords a possibly unique opportunity, in Archean terrane at least, to examine lithologic and chemical fractionation trends upwards within such a pluton, its pegmatitic granite cupola zone and exocontact aureole of rare-element pegmatites. The six kilometre wide pegmatitic aureole (Mavis Lake group) is characterized by systematic mineral zoning (i.e., beryl --> columbite and beryl --> tantalite, beryl, and spodumene) which is sympathetic with an increase in albitization and development of exomorphic aureoles. The Gullwing-Tot Lakes pegmatite group has a more nebulous and complex regional zoning (columbite --> beryl, molybdenite, columbite and spodumene --> beryl, columbite, tantalite, pollucite and spodumene). Source region for this group is unclear but may be related to pegmatitic granites of the Coates Bay pluton.

The moderate to strong peraluminous character of the GLB is indicated mineralogically by cordierite, biotite, garnet, tourmaline, sillimanite and primary muscovite and chemically by high $\text{Al}_2\text{O}_3/\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O}$ (molar 1.023-1.715) and Barth mesonormative corundum (3.2-8.4 percent).

Bulk chemical characteristics of the parental granite include high SiO_2 (70.5 to 77.4 percent), variable $\text{Na}_2\text{O}/\text{K}_2\text{O}$ (0.25-2.5), low Ca, Mg and Fe. Strong fractionation in the pluton is revealed by K/Rb (70-387), K/Ba (14-2805), Al/Ga (2380-7020), Ba/Rb (0.03-9.1), Rb/Sr (0.1-33), Th/U (0.3-35), Ca/Sr (6-360), Ce/Yb_N (2.6-247), Eu/Eu* (<0.06-1.98), and ΣREE (1.8-80 x chondrite). Bulk analyses of the associated rare-element pegmatites generally exhibit more advanced chemical fractionation than the GLB in terms of K/Rb (8-24), K/Ba (119-503), Al/Ga (1456-2341), Ba/Rb (0.02-0.15), Rb/Sr (21-122), Th/U (10-50), Ce/Yb_N (1-2) and ΣREE (0.3-12.9x chondrite) although many of these parameters overlap.

A five stage model is proposed to account for the origin of the peraluminous granite-rare-element pegmatite association; this process of chemical fractionation may have been operative for at least 120 Ma:

- 1) extensive anatexis of Li-enriched clastic metasediments
- 2) unmixing of restite in a rising anatectic granitic magma
- 3) crystal-melt fractionation
- 4) exsolution of a supercritical aqueous fluid in the parent granite

5) emplacement and chemical evolution of exocontact, rare-element pegmatites

The GLB is a mass composed of eight co-magmatic plutonic units (GLB-1 to -8) which is inferred to have formed by extensive, fluid-absent anatexis of biotite-rich metapelites and metawackes under granulite facies conditions. This process released rare elements, F and B which were previously held in micas, to the melt. Emplacement was effected as a late tectonic crystal-melt mass discordant to the andalusite and sillimanite-muscovite regional metamorphic zones.

Restite unmixing may have played a role in the chemical and petrological evolution of the most primitive unit, GLB-1, given the abundance of restite segregations and potassic compositions close to the thermal minimum in the haplogranite system. This process was perhaps accompanied by crystal-melt fractionation, which became the more important cause of effecting compositional diversity in the later plutonic units, especially GLB-3 to -7. As anhydrous minerals dominated the early crystallization history (plagioclase, K-feldspar and quartz), accumulation of H_2O , B_2O_3 and F gradually increased in the remaining melt. Ultimately volatile saturation resulted and an aqueous fluid exsolved. Resurgent boiling on a recurring basis led to a diverse assemblage (GLB-4 to -8) of pegmatitic granite and layered aplite which contains sporadic beryl, columbite, and cassiterite.

Late- to post-magmatic autometasomatic "reworking" of primary plutonic units by albitization and mild greisenization aided the concentration of rare-elements, F and B into a Li-rich melt and an aqueous fluid system. Transfer of this metalliferous volatile phase

plus pegmatitic melt from the GLB apical roof zone is suggested by a holmquistite-bearing alteration zone in the host mafic metavolcanics of a type not previously documented. This zone may represent the missing 'parent-progeny' link in Li abundance between a granite and its Li-rich exocontact pegmatites. Disengagement of pegmatite melt from the batholith roof is supported by the presence of pegmatitic granites (with accessory beryl, cassiterite and chrysoberyl) within the holmquistite alteration zone. These pegmatites are concluded to be the lowest level of exocontact rare-element mineralization in the regionally zoned Mavis Lake Group.

The more highly fractionated Li-rich pegmatitic melt-fluid system, because of its lower solidus temperature and possibly lower viscosity, remained mobile longer than pegmatitic granites in the GLB migrating into dilatant structures.

The last stage of rare-element pegmatite evolution involved post-magmatic alteration and development of aureoles enriched in holmquistite, biotite and tourmaline.

An important problem addressed by this study concerns the petrogenetic and structural controls governing the linear regional distribution of rare-element pegmatites in the Superior Province of Ontario. The confinement of these pegmatites to curvilinear boundary zones between some subprovinces which contrast in lithology and metamorphic grade has been known for some time. As prominent fault zones commonly coincide with these boundaries Cerny and Trueman (1978) suggested that rare-element pegmatite emplacement is controlled by these structures. It should be pointed out that fault-controlled

pegmatite localization may also occur within the interiors of some greenstone-rich belts as in the Sachigo Subprovince of Manitoba (Cerny and Meintzer 1988).

While fault control may be an important controlling factor in the regional distribution of peraluminous granites and rare-element pegmatites in other areas, the Dryden Pegmatite Field, in contrast, bears no evidence of such control.

This study provides an alternative explanation for the regional distribution of rare-element pegmatites in subprovince boundary zones of northwestern Ontario. The proposed S-type granite-rare-element pegmatite model is strongly dependent upon protolith lithology and the extent to which derived anatectic granitic melts can move upwards in the crust.

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APPENDICES

APPENDIX 1: Analytical methods, Detection Limits and Precision
for Major, Minor and Trace Elements

APPENDICES 2-41: Analytical Data of Rocks and Minerals.

NOTES:

- 1) Sample localities for analyzed samples are given in Figures 15, 16, 18, 19, 23, 25, 27, 29, 31 and 38.
- 2) Concentrations of major and minor elements given in weight percent; all others are in parts per million (ppm) unless otherwise indicated.
- 3) Blank spaces indicate elements not analyzed for.

APPENDIX 1: Analytical Methods, Detection Limits and Precision for Major, Minor and Trace Elements

Analytical Procedures

Whole rock and mineral chemical analyses were performed by staff of the Geoscience Laboratories, Ontario Geological Survey in Toronto under the supervision of Dr. Chris Riddle.

Major elements were determined by X-Ray fluorescence on fused discs. Precision for mid-range values (MRV) at the 95 percent confidence level (2 times the standard deviation) is tabulated below.

FeO and minor constituents CO₂, S, P₂O₅, H₂O⁺, H₂O⁻ and MnO were analyzed by methods indicated below, with precision at ten times the detection limit.

The trace elements, B, Ba, Be, Cs, F, Ga, Li, Mo, Nb, Rb, Sn, Sr, Th, U, W and Zr were determined by a variety of techniques with precision also at ten times detection limit.

Rare earth elements and Y were analysed by the Inductively Coupled Plasma (ICP) spectrometer on samples prepared from total acid digestion of the rock pulp. Precision at ten times detection limit is given below.

Electron microprobe analyses of complex Nb-Ta oxide minerals, garnet and cordierite were performed by Robert Barnett, Geology Department, University of Western Ontario. A MAC 400 EMP was employed, utilizing an acceleration voltage of 20 KV, sample current of 0.04 amps and a Kriesel control automatic system (V4C-M6SP01) with

Magic IV correction procedures. Standards used for EMP analyses were minerals of known composition.

TABLE A-1: Analytical Methods, Detection Limits, and Precision for Major, Minor and Trace Elements

<u>MAJOR CONSTITUENTS</u>				
Element Limit	Technique	Range %	Precision at Mid-Range Value %	Detection %
SiO ₂	XRF	30-80	± 2.0	
Al ₂ O ₃	XRF	0-20	0.4	
Fe ₂ O ₃	XRF	0-15	0.6	
MgO	XRF	0-20	0.3	
CaO	XRF	0-10	0.5	
Na ₂ O	XRF	0-10	0.1	
K ₂ O	XRF	0-15	0.15	
TiO ₂	XRF	0-1	0.15	
P ₂ O ₅	XRF	0-3	0.16	
MnO	XRF	0-1	0.015	
FeO	Volumetric		0.2	0.02
CO ₂	LECO-IR		0.06	0.05
S	LECO-IR		0.02	0.03
P ₂ O ₅	Colorimetric		0.1	0.03
H ₂ O+	AAS Gravimetric		0.2	0.1
H ₂ O-	AAS Gravimetric		0.1	0.05

<u>TRACE ELEMENTS</u>				
Element	Technique	Detection PPM	Optimum Range (PPM)	Precision at 10x Detection Limit
B	NAA	1		
Ba	AA Flame	10	10-1000	16
Be	ICP/OES	1	1-100	± 1
Cs	XRF	3	3-1000	3
F	Colorimetric	40	40-1000	40
Ga	XRF	5	5-1000	5
Li	AA Flame	3	3-200	3
Mo	AA Flame	3	3-400	3
Nb	ICP/OES	5	5-500	10
Rb	XRF	5	5-1000	5
Sn	AA Hydride	0.8	0.8-10	1
Sr	XRF	5	5-1000	5
Ta	XRF	10	10-5000	20
Th	XRF	10	10-5000	15
U	Fluorimetric	1	1-10000	4
W	ICP/OES	50	50-500	100
Zr	XRF	5	5-1000	10

RARE-EARTH ELEMENTS

Element	Technique	Detection (PPM)	Optimum Range (PPM)	Precision at 10x Detection Limit
Y	ICP/MS	0.02	0.02-200	0.02
La	ICP/MS	0.05	0.05-200	0.05
Ce	ICP/MS	0.05	0.05-300	0.05
Pr	ICP/MS	0.05	0.05-200	0.05
Nd	ICP/MS	0.18	0.18-200	0.18
Sm	ICP/MS	0.15	0.15-100	0.15
Eu	ICP/MS	0.07	0.07- 20	0.07
Gd	ICP/MS	0.14	0.14-100	0.14
Tb	ICP/MS	0.03	0.03- 20	0.13
Dy	ICP/MS	0.13	0.13- 20	0.13
Ho	ICP/MS	0.03	0.03- 20	0.03
Er	ICP/MS	0.10	0.10- 20	0.10
Tm	ICP/MS	0.03	0.03- 20	0.03
Yb	ICP/MS	0.11	0.11- 20	0.11
Lu	ICP/MS	0.04	0.04- 20	0.04

ANALYTICAL METHODS

AA Flame	Atomic Absorption Spectrophotometry (Flame)
AA Hyride	Atomic Absorption Spectrophotometry (Hydride)
AAS Gravimetric	Atomic Absorption Spectrophotometry (Gravimetric)
Colorimetric	Colorimetric Spectrophotometry
Fluorimetric	Fluorine Analysis by Technician Analyzer II
ICP/MS	Inductively Coupled Plasma/Mass Spectrometer
IXCP/OES	Inductively Coupled Plasma/Optical Emission
Spectrometry	
LECO-IR	Infrared Analysis by LECO Determinator
NAA	Neutro Activation Analysis
XRD	X-Ray Fluorescence

Appendix 2. B, Be, Cs, Li and Sn (ppm) of Metapelites and Metawackes from various Metamorphic Zones of the Dryden-Eagle River area. Locations of Analyses given in Figure 38.

A. METAPELITES

Chlorite-Biotite Zone

Sample Number	B	Be	Cs	F	Li	Sn
F24-6	98	2		600	114	3.4
32-9-1B	150	4		16		3.0
32-12-2b	52	1	10	460	45	4.9
32-12-8a	33	1	<10	370	25	5.3
Average	62	2	<57	477	50	4.2

Andalusite Zone

F14-1B	46	<1	10	570	36	2.5
F20-1A	41	<1	<10	510	90	3.2
F22-2	32	2	20	730	44	3.5
F22-6A	66	<1	<10	570	52	2.2
F24-4	90	2	<10	550	210	1.9
F24-4C	150	<1	30	480	138	2.9
F24-5B	34	1	10	740	73	4.2
F24-7	30	1	10	600	102	2.1
F24-8b		<1	10		74	3.0
F32-4	15	<1	10	180	26	2.2
G8-2	60	2	30	460	136	2.5
G8-5	54	3	10	490	37	2.6
G10-1	50	3	10	390	47	2.2
G10-2	230	3	20	560	29	2.0
G10-3	170	2	10	470	20	2.7
32-5-1		8	31		61	3.0
32-8-10c	51	<1	<10	470	38	4.6
32-8-11a	60	<1	<10	530	46	5.3
32-8-12b	21	<1	10	480	40	3.1
32-8-14a	37	<1	10	580	38	3.3
32-8-14b	36	1	<10	450	34	4.5
32-8-14c	50	<1	<10	510	36	2.9
32-14-1	5	<1	20	480	99	4.4
32-14-2b	50	<1	<10	520	38	3.6
32- 5-7b		4	12		38	5
Average	65	<1.9	<10	515	63	3.1

Appendix 2, continued

Sillimanite-Muscovite Zone

F-18-1c	110	<1	90	770	375	6.1
F18-4b	260	1	<10	530	37	2.5
F20-2	560	2	10	510	118	3.4
F22-1	240	8	20	530	86	3.8
F27-1b	525	<1	<10	440	53	3.6
F27-4	5	2	50	440	114	1.4
G4-1	40	2	20	430	46	3.1
32-5-4		<1	31		94	7.0
C35-3	27	<1	10	660	62	2.3
D34-2	68	<1	20	370	88	2.8
Average	204	<2	<27	520	107	3.6

Sillimanite-K-feldspar Zone (non-migmatitic)

F6-1	6	4	10	400	76	2.6
F7-2	43	1	10	470	245	2.3
F7-4	25	2	20	630	152	3.9
F9-1	40	1	20	740	520	3.9
G8-1	170	3	80	610	590	5.8
F9-4	12	1	30	720	64	3.8
Average	49	2	28	595	275	3.7

Sillimanite-K-feldspar Zone (migmatitic)

C15-2a	<5	<1	<10	490	44	2.1
C15-3a	<5	<1	10	480	34	3.1
D12-3a	<5	2	<10	470	62	1.8
E8-1b	12	4	10	490	76	2.2
E9-1	7	3	10	610	39	1.9
E9-3	16	1	10	410	35	1.4
E9-6		<1	14		27	3.0
E11-1	<5	2	10	620	38	2.2
F5-3	9	<1	<10	720	56	2.6
F7-8	10	3	20	480	66	2.5
E15-3b	5	1	<10	630	35	2.1
Average	<8	<1.8	<11	540	47	2.3

Inclusions in Ghost Lake Batholith

C15-5	<5	4	10	550	67	1.8
D20-1a	<5	<1	<10	340	29	1.3
E14-1c	<5	<1	<10	650	42	1.7
Average	<5	<3	<10	513	46	1.6

Appendix 2, continued

B. METAWACKES**Chlorite - Biotite**

Sample Number	B	Be	Cs	F	Li	Sn
32-8-8	17	<1	10	200	28	3.1
32-9-1a		<1	11		34	
32-12-1	<5	<1	10	330	22	4.0
32-12-2a	17	<1	<10	380	30	4.0
32-12-15	15	<1	<10	400	26	2.0
32-8-9a	15	<1	<10	350	26	1.9
Average	12	<1	<10	277	28	3.0

Andalusite Zone

F18-3	19	2	10	440	36	2.2
F20-1b	42	<1	<10	460	105	2.3
F24-5a	23	3	50	760	84	5.3
F24-8a		<1	10		44	
F26-2b	<5	<1	<10	500	128	1.1
F26-3a	5	3	100	830	104	3.9
F26-4	12	<1	40	430	235	1.8
F30-2	17	<1	<10	420	23	2.9
32-5-7a		4	6		16	
32-8-10a	12	<1	<10	370	23	2.2
32-8-12a	12	<1	10	330	24	2.9
32-12-11	120	<1	30	320	50	4.1
32-12-12a	<5	<1	30	570	37	4.1
32-14-2a	20	<1	10	350	28	2.3
Average	21	<1.6	<17	413	67	2.9

Sillimanite - Muscovite Zone

F18-16	31	1	<10	300	24	1.5
F18-2a	<5	<1	50	2140	355	10.7
F20-2b	98	<1	<10	550	75	2.1
F27-3b	<5	<1	10	550	154	3.2
E29-4	15	7	20	380	77	6.7
F30-3	<5	<1	<10	350	109	2.7
D26-1c	160	<1	10	550	39	1.8
F6-1c	14	1	<10	540	51	2.3
F7-1c	32	<1	<10	350	108	2.4
Average	23	<1	<10	445	80	2.35

Appendix 2, continued

Sillimanite- K-feldspar (migmatitic) Zone

C11-1b	<5	<1	10	320	17	2.1
C13-3d	<5	1	<10	630	24	2.3
C15-1a	<5	2	10	460	58	2.4
C17-4b	<5	<1	<10	690	23	1.9
E13-2	<5	<1	<10	530	28	1.7
E13-4	8	3	<10	430	94	4.8
E13-6a	<5	<1	<10	360	14	1.5
E15-2a	<5	<1	10	490	36	2.4
E17-2a	<5	<1	20	1000	78	2.6
E5-4	<5	1	<10	310	11	1.7
F5-4c	<5	<1	10	550	22	1.9
E17-1a	<5	<1	10	700	46	3.0
Average	<5	<1	<10	540	38	2.4

Inclusions within the Ghost Lake Batholith

C17-26	<5	<1	<10	770	38	5.3
C17-5a	24	<1	20	670	31	4.0
D14-7	<5	<1	10	680	30	1.4
D16-1b	<5	<1	<10	570	26	1.0
D18-2b	<5	<1	10	590	42	1.9
D21-1b	<5	<1	<10	470	32	1.7
D22-1b	<5	<1	10	490	30	1.70
E17-3	<5	<1	<10	530	36	1.8
E17-8a	11	2	<10	480	42	3.2
E19-1b	<5	1	10	440	48	1.8
Average	<7.5	<1	<10	569	36	2.4

Appendix 3, continued

	E17-4A	E17-8B	E19-2	E25-3B	E25-4	E25-7	E27-1	E27-3	E27-4B
SiO ₂	75.00	71.20	70.40	74.70	74.30	72.20	74.60	74.50	75.10
Al ₂ O ₃	14.70	16.90	17.30	14.60	15.20	15.90	15.20	15.00	14.50
Fe ₂ O ₃	0.20	0.25	0.30	0.0	0.0	0.0	0.03	0.02	0.33
FeO	0.45	0.73	0.64	0.56	0.45	0.49	0.58	0.73	0.64
MgO	0.14	0.25	0.25	0.09	0.12	0.11	0.24	0.14	0.14
CaO	0.75	0.92	1.25	0.65	1.12	0.99	0.79	0.65	0.95
Na ₂ O	2.00	1.72	2.77	1.97	2.51	2.02	2.63	2.50	2.41
K ₂ O	5.81	6.35	5.10	6.53	5.53	6.44	5.42	4.92	5.42
TiO ₂	0.35	0.36	0.09	0.29	0.25	0.23	0.29	0.26	0.30
P ₂ O ₅	0.10	0.11	0.05	0.10	0.09	0.11	0.10	0.10	0.09
S	0.01	0.0	0.01	0.0	0.0	0.0	0.01	0.01	0.01
MnO	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.02	0.01
CO ₂	0.17	0.19	0.25	0.22	0.10	0.26	0.18	0.19	0.11
H ₂ O+	0.37	0.45	0.36	0.23	0.13	0.20	0.35	0.73	0.23
H ₂ O-	0.04	0.07	0.09	0.07	0.07	0.05	0.01	0.05	0.0
Total	100.16	99.57	95.91	99.92	99.94	99.04	100.43	99.95	100.27

Appendix 3, continued

TRACE ELEMENTS (ppm)

	A17-1	C16-1	D14-1	D14-6B	D15-1B	D16-2A	D20-1B	D22-1A	D24-1A	E15-1
B	10	12	15	13	10	9	15	10	31	16
Ba	370	1060	1240	1160	1210	1090	1090	1940	420	720
Be	<1	<1	2	<1	<1	<1	<1	<1	<1	<1
Co	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5
Cr	19	6	19	<5	<5	<5	<5	<5	<5	5
Cs	10									
F	260	400	260	260	210	90	90	60	130	60
Ga	14	16	15	13	12	12	12	11	12	13
Li	30	6	8	4	6	3	3	3	20	12
Mo	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10
Nb	<1	1	4	<1	<1	<1	<1	<1	4	<1
Pb	30	47	33	37	41	49	43	51	32	40
Rb	200	140	150	140	150	150	120	210	370	140
Sc	2	2	4	3	2	2	2	2	2	3
Sr	100	275	<0.6	<0.6	<0.6	<0.6	1.2	1.1	2.2	1.4
Th	20	10	70	30	235	190	235	200	80	185
U	6	1	3	3	30	20	<10	50	30	20
V	7	6	30	6	3	280	5	3	2	13
Y	7	3	7	7	4	7	5	6	4	5
Zn	26	20	36	18	21	16	14	3	6	4
Zr	75	40	45	155	85	60	30	60	12	95

	E17-4A	E17-8B	E19-2	E25-3B	E25-4	E25-7	E27-1	E27-3	E27-4B
B	19	14	13	13	16	11	100	300	280
Ba	600	1460	580	800	890	1000	770	220	730
Be	<1	<1	2	2	<1	1	1	2	2
Co	<5	<5	<5	<5	<5	<5	<5	<5	<5
Cr	<5	<5	<5	<5	<5	<5	<5	<5	<5
Cs									
F	400	440	420	420	370	410	360	400	300
Ga	15	17	20	16	16	12	17	21	17
Li	12	9	16	20	12	8	10	26	27
Mo	<10	<10	<10	<10	<10	<10	<10	<10	<10
Nb	1	1	4	4	<1	1	2	8	3
Pb	54	42	52	67	41	48	38	30	33
Rb	160	140	150	230	160	170	160	150	190
Sc	3	3	3	3	2	3	3	3	3
Sn	2.0	2.1	1.6	2.9	2.7	3.2	2.0	3.5	2.7
Sr	260	365	260	165	210	235	185	65	125
Th	40	30	20	20	30	30	60	20	20
U	30	6	28	6	5	2.8	5	6	6
V	6	13	11	4	5	6	5	4	7
Y	16	5	8	5	3	5	8	7	5
Zn	26	27	26	24	20	24	19	36	38
Zr	110	70	60	25	30	135	360	50	120

Appendix 3, continued

ELEMENT RATIOS

	A17-1	C18-1	D14-1	D14-6B	D18-1B	D18-2A	D20-1B	D22-1A	D28-1A	E14-1A	E15-1
K/Rb	213	337	271	350	307	349	367	324	181	262	286
K/Ba	115	45	33	42	38	49	43	37	160	80	56
Al/Ga	5340	5040	5120	5990	6890	6090	6450	7090	6440	20900	5870
Na/K	0.4	0.55	0.4	0.3	0.5	0.75	0.2	0.1	0.2	0.5	0.4
Ca/Sr	71	38	26	26	33	31	31	8	42	39	32
Ba/Rb	1.85	7.6	9.3	8.3	8.1	7.1	9.1	8.6	1.1	3.3	5.1
Rb/Sr	2.0	0.5	0.5	0.6	0.6	0.4	0.5	1.05	4.6	0.9	0.6
Th/U	3.3	10	23	10	10	0.07	12	17	15	1.5	1.5
Al ₂ O ₃ : CaO+Ba ₂ O + K ₂ O	1.397	1.356	1.435	1.363	1.367	1.430	1.569	1.329	1.293	1.396	1.476

	E17-4A	E17-4B	E19-3	E25-3B	E25-4	E25-7	E27-1	E27-3	E27-4B
K/Rb	301	376	282	236	325	316	250	272	237
K/Ba	80	46	73	65	52	54	55	156	62
Al/Ga	5190	5370	4590	4940	5040	7020	4740	3790	4570
Na/K	0.3	0.3	0.5	0.3	0.4	0.3	0.5	0.5	0.4
Ca/Sr	21	15	35	25	35	30	31	75	56
Ba/Rb	3.5	10.4	6.9	3.5	6.4	5.9	4.3	1.5	3.5
Rb/Sr	0.6	0.4	0.6	1.4	0.7	0.7	0.97	2.3	1.5
K/Sr									
Th/U	1.3	5	0.7	3.3	7.5	4.75	12	3.3	3.3
Al ₂ O ₃ : CaO+Ba ₂ O + K ₂ O	1.585	1.282	1.71	1.449	1.503	1.535	1.423	1.529	1.475

* = average value

Appendix 3, continued

MODES

	A17-1	C15-1	D14-1	D14-6B	D15-1B	D15-2A	D20-1B	D22-1A	D25-1A	E14-1A
Quartz	35	22	37	30	29	27	36	25	19	25
K-feldspar	32	35	25	44	39	41	34	51	64	62
Plagioclase	29	39	25	23	30	26	19	12	14	27
Biotite	3	3	5	2	1	4	2	1	2	2
Tourmaline										
Almandine										
Muscovite										
Chlorite	x	<1	3	1	1	2	9	7	1	x
Total	100	99	101	100	100	100	100	99	99	100
Colour Index	4	3	5	2	1	1	2	1	2	2
Plagioclase Composition	An20	An21		An15	An15	An21	An25	An10	An13	An17

	E15-1	E17-4A	E17-5B	E19-2	E25-3B	E25-4	E25-7	E27-1	E27-3	E27-4B
Quartz	31	31		21	30	35	25	34	34	29
K-feldspar	34	43		36	42	29	43	27	34	38
Plagioclase	31	19		31	22	27	23	30	25	26
Biotite	2	4		5	3	5	4	1	2	6
Tourmaline										
Almandine								x	0.3	x
Sillimanite				1			x		x	
Muscovite	2	3		6	3	4	2	9	4	2
Chlorite										
Total	100	100		100	101	100	100	100	99	100
Colour Index	2	4		5	3	5	4	1	2	6
Plagioclase Composition	An17	An15	An24	An21	An17	An21	An22	An15	An14	An19

Appendix 3, continued

BARTH MESONMIS

	A17-1	C15-1	D14-6B	D18-1A	D18-2A	D20-1B	D22-1A	D28-1A	E14-1A
Albite	21.0	29.9	19.4	25.6	15.9	14.0	11.1	17.0	25.4
Anorthite	3.2	4.8	1.1	1.7	1.1	3.4	0.0	0.0	2.1
Orthoclase	3.1	33.3	34.4	33.1	36.9	34.0	48.7	47.9	31.3
Quartz	38.4	25.6	37.1		37.9	41.9	34.2	28.2	33.9
Apatite	0.2	0.2	0.2	0.3	0.3	0.2	0.1	0.3	0.2
Sphene	0.5	0.5	0.5	0.5	0.9	0.1		0.5	0.4
Calcite	0.2	0.4	0.6	0.8	0.3	0.3	0.3	0.1	0.6
Corundum	4.2	4.2	4.9	4.9	5.3	5.5	4.3	4.5	5.0
Biotite	2.3	1.4	1.6	0.5	1.5	0.5	1.1	1.5	1.1
Magnetite				0.1					

	F15-1	F17-4A	F17-9B	F19-2	F25-3B	F25-4	F25-7	F27-1	F27-3	F27-4B
Albite	21.5	15.3	15.8	25.4	17.1	22.9	15.6	23.9	23.0	22.0
Anorthite	2.6	0.9	1.3	3.7	0.0	3.3	1.6	1.1	0.5	2.5
Orthoclase	27.5	34.2	37.0	29.5	38.4	32.2	36.3	31.1	25.5	31.6
Quartz	39.6	35.8	34.2	31.3	36.5	34.7	33.0	35.5	35.6	36.7
Apatite	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2	0.2
Sphene	0.6	0.5	0.5	0.2	0.6	0.6	0.5	0.6	0.6	0.6
Calcite	0.5	0.5	0.5	0.7	0.6		0.7	0.5	0.5	0.3
Corundum	4.8	5.5	7.5	6.6	5.9	4.3	5.6	5.1	6.0	4.2
Biotite	2.3	1.5	2.2	2.9	1.2	1.5	1.5	2.1	2.1	1.6
Magnetite		0.2	0.3	0.3						0.4

Appendix 4. Chemical Analyses, Selected Element Ratios, Modal Analyses and Barth Mesonorms of GLB-1A.

	C15-5	C17-2A	C17-4A	C20-3	C20-4
SiO ₂	76.40	74.60	72.50	73.30	77.40
Al ₂ O ₃	14.90	15.30	15.20	15.40	13.70
Fe ₂ O ₃	0.0	0.0	0.16	0.0	0.0
FeO	0.16	0.40	2.10	0.73	0.64
MgO	0.0	0.10	0.91	0.32	0.23
CaO	1.94	0.93	1.25	0.45	0.67
H ₂ O	3.73	2.27	2.17	1.94	1.54
K ₂ O	2.80	5.66	3.90	6.06	5.33
TiO ₂	0.24	0.21	0.21	0.35	0.24
P ₂ O ₅	0.12	0.14	0.07	0.11	0.12
S	0.01	0.01	0.01	0.01	0.0
MnO	0.01	0.01	0.02	0.01	0.01
CO ₂	0.05	0.23	0.20	0.22	0.35
H ₂ O ⁺	0.18	0.13	0.43	0.14	0.38
H ₂ O ⁻	0.10	0.12	0.09	0.09	0.13
Total	100.62	100.11	99.72	99.57	100.52

	C15-5	C17-2A	C17-4A	C20-3	C20-4
B	12	13	17	10	17
Ba	400	630	370	1020	1030
Be	1	1	5	<1	<1
Co	<5	5	5	5	<5
Cr	5	5	13	5	<5
CS					
F	200	230	330	300	260
Ga	15	13	22	14	12
Li	4	8	16	10	6
Hf	<10	<10	<10	<10	<10
Hb	1	<1	9	<1	<1
Pb	47	46	29	37	27
Rb	80	170	160	190	160
Se	1	2	4	2	2
Sn	<0.5	0.5	<0.5	<0.5	1.4
Sr	165	160	135	265	300
Th	<10	30	30	20	<10
U	6	4	4	2	<1
V	3	3	25	11	9
Y	3	6	6	4	3
Zn	8	16	56	22	14
Zr	15	90	145	65	10

Appendix 4, continued

ELEMENT RATIOS

	C15-5	C17-2A	C17-4A	C20-3	C20-4
K/Rb	290	276	207	265	277
K/Ba	74	75	49	69	44
Al/Ga	5260	6240	3660	5930	6050
Na/K	1.3	0.4	0.5	0.3	0.4
Ca/Sr	85	42	45	14	30
Ba/Rb	3.9	3.7	2.3	5.3	6.3
Th/U	<1.7	7.5	7.5	10	--
Rb/Sr	0.5	1.1	1.2	0.7	0.5
Al ₂ O ₃ : CaO+Na ₂ O + K ₂ O*	1.175	1.326	1.512	1.357	1.441

* molecular ratio

MODELS

	C15-5*	C17-2A*	C17-4A	C20-3	C20-4
Quartz	34	24	32	32	36
K-feldspar	17	45	21	40	47
Plagioclase	43	29	42	21	24
Biotite	1	3	4	3	1
Cordierite			11	2	2
Muscovite	1	1		2	1
Total	100	99	100	100	99
Colour Index	1	<1	15	5	4
Plagioclase Composition	An ₂₃	An ₁₉	An ₂₅	An ₂₀	An ₂₀

* Specimens contain biotite-cordierite-quartz restite clots that were not included in the modal and chemical analyses

BARTHELEMY MODELS

	C15-5	C17-2A	C17-4A	C20-3	C20-4
Albite	33.7	20.7	20.1	18.2	14.5
Anorthite	7.5	1.4	3.7	0.8	0.0
Orthoclase	16.4	33.2	18.2	34.4	30.7
Quartz	37.8	36.6	41.3	35.9	45.1
Apatite	0.3	0.3	0.2	0.2	0.3
Sphene	0.5	0.4	0.5	0.7	0.1
Calcite	0.2	0.6	0.6	0.6	1.0
Corundum	3.3	5.5	6.8	5.9	5.9
Biotite	0.3	1.2	7.5	2.7	2.2
Magnetite			0.2		

Appendix 5. Chemical Analyses, Selected Element Ratios, Modal Analyses and Barth Mesonorms of GLB-2.

	D28-1B	D30-4A	F13-4C	F19-4	F19-5A	E27-4A
SiO ₂	74.60	73.60	75.40	74.30	71.90	73.00
Al ₂ O ₃	14.20	14.40	14.30	15.40	16.20	14.30
Fe ₂ O ₃	0.24	0.16	0.39	0.09	0.32	0.41
FeO	1.13	1.29	0.81	0.56	1.05	1.29
MgO	0.24	1.17	0.20	0.10	0.49	0.31
CaO	1.03	1.09	1.41	1.00	2.32	1.11
Na ₂ O	2.81	2.43	2.55	3.24	3.95	2.36
K ₂ O	4.72	5.15	3.44	4.04	2.40	4.98
TiO ₂	0.37	0.46	0.44	0.31	0.47	0.44
P ₂ O ₅	0.10	0.11	0.07	0.08	0.10	0.09
S	0.01	0.01	0.01	0.01	0.01	0.01
MnO	0.01	0.02	0.01	0.00	0.01	0.01
CO ₂	0.17	0.21	0.19	0.26	0.34	0.17
H ₂ O ₊	0.37	0.42	0.40	0.12	0.19	0.57
H ₂ O ₋	0.05	0.09	0.10	0.13	0.09	0.04
Total	100.09	99.61	100.12	99.75	99.94	99.09

	D28-1B	D30-4A	F13-4C	F19-4	F19-5A	E27-4A
B	21	11	9	14	5	20
Ba	560	600	700	2460	1210	530
Be	<1	1	1	1	1	1
Co	<5	<5	<5	<5	<5	<5
Cr	<5	<5	<5	<5	<5	<5
Cs						
F	160	250	160	350	500	350
Ga	19	20	14	16	14	14
Li	26	22	32	14	22	40
Mn	<10	<10	<10	<10	10	<10
Nb	5	9	5	<1	1	4
Pb	26	35	21	19	14	29
Rb	200	210	130	90	80	190
Sc	4	4	3	2	3	3
Sn	0.9	2.5	0.9	0.8	0.9	1.0
Sr	110	110	170	690	835	125
Th	50	40	40	<10	20	60
U	4	4	1	2	1	5
V	14	14	9	8	23	14
Y	11	11	5	3	4	9
Zn	46	49	16	23	40	46
Zr	185	205	150	80	105	225

Appendix 5, continued

ELEMENT RATIOS

	D29-1B	D30-4A	E13-4C	E19-4	E19-5A	E27-4A
K/Rb	196	204	199	376	249	218
K/La	70	71	46	14	17	60
Al/Ga	3690	3820	5410	5100	4770	4210
Th/U	10	40	45	20	12	12.5
Rb/Sr	1.5	1.9	0.9	0.1	0.1	1.5
Ca/Sr	67	71	60	10	20	6.4
Ba/Rb	2.4	2.9	5.4	27.3	15.1	2.8
Na/K	0.6	0.5	0.6	0.8	0.6	0.45
Al ₂ O ₃ : CaO+Na ₂ O +K ₂ O	1.225	1.246	1.312	1.333	1.218	1.288

MODELS

	D29-1B	D30-4	E13-4C	E19-4	E19-5B	E27-4A
Quartz	30	33	34	24	28	26
K-feldspar	31	34	27	31	10	37
Plagioclase	30	25	33	41	56	24
Biotite	4	5	3	3	5	6
Almandine			X			
Muscovite	1	X	2	1	2	5
Fe-oxides				X		
Total	100	101	99	100	101	100
Colour Index	4	5	3	3	5	6
Plagioclase Composition	An14	An21	An24	An15	An26	An22

BARTH MEASUREMENTS

	D29-1B	D30-4A	E13-4C	E19-4	E19-5A	E27-4A
Albite	25.7	22.4	23.4	29.5	35.4	21.3
Anorthite	2.0	1.7	3.4	1.6	6.6	2.3
Orthoclase	26.4	29.2	21.9	23.6	12.0	28.2
Quartz	36.9	36.5	41.6	36.6	34.1	37.2
Apatite	0.2	0.2	0.1	0.2	0.2	0.2
Sphene	0.4	1.0	0.9	0.7	1.0	0.9
Calcite	0.5	0.6	0.5	0.7	1.1	0.5
Corundum	4.2	4.7	5.1	5.6	5.2	4.8
Biotite	3.2	3.2	2.1	1.4	3.7	3.5
Magnetite	0.3	0.2	0.4	0.1	0.3	0.4

Appendix 6. Chemical Analyses, Selected Element Ratios, Modal Analyses and Barth Mesonorms of GLB-3.

	F9-2	F9-3	F10-1	F11-1	F11-2	F11-6A	F11-6B	F11-1B	F11-1A	F16-3A	F16-3C	F16-4	F16-1A
SiO ₂	74.50	73.00	74.70	74.80	73.60	73.00	73.90	75.30	73.90	75.20	75.20	65.70	74.70
Al ₂ O ₃	15.50	15.50	15.20	14.90	14.80	15.50	15.30	15.00	15.70	15.20	14.90	19.90	14.80
Fe ₂ O ₃	0.0	0.04	0.05	0.03	0.26	0.0	0.13	0.24	0.32	0.07	0.0	0.21	0.00
FeO	0.46	0.50	0.44	0.59	0.97	0.56	0.40	0.40	0.40	0.44	0.73	0.89	0.64
HgO	0.0	0.07	0.0	0.36	0.22	0.0	0.03	0.05	0.01	0.05	0.01	0.21	0.05
CaO	0.81	0.94	0.71	1.03	1.14	0.40	0.74	0.77	0.67	0.79	0.77	1.28	0.83
Na ₂ O	3.60	2.76	4.12	3.39	3.21	2.76	4.16	4.90	4.49	3.95	4.16	3.73	3.32
K ₂ O	4.45	5.01	3.87	4.00	4.36	5.30	3.64	3.72	4.14	3.76	3.21	6.26	4.35
TiO ₂	0.20	0.45	0.17	0.27	0.35	0.03	0.03	0.13	0.16	0.19	0.19	0.36	0.42
P ₂ O ₅	0.09	0.12	0.14	0.11	0.10	0.14	0.14	0.13	0.34	0.11	0.09	0.10	0.13
S	0.0	0.01	0.0	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.9	0.0
MnO	0.02	0.01	0.02	0.02	0.01	0.13	0.02	0.02	0.02	0.02	0.02	0.02	0.01
CO ₂	0.06	0.45	0.15	0.27	0.36	0.24	0.15	0.12	0.23	0.11	0.22	0.36	0.11
H ₂ O ⁺	0.29	0.27	0.16	0.40	0.32	0.20	0.15	0.12	0.24	0.15	0.35	0.28	0.31
H ₂ O ⁻	0.05	0.14	0.05	0.05	0.11	0.10	0.09	0.10	0.10	0.07	0.09	0.13	0.03
Total	99.85	99.33	99.62	100.33	99.64	96.23	96.93	100.01	100.35	100.12	99.97	99.43	99.70

Appendix 6, continued

	F9-2	F9-3	F10-1	F11-1	F11-2	F11-6A	F11-6B	F13-1B	F13-1A	F16-3A	F16-3C	F16-4	F18-1A
B	130	47	12	10	10	27	17	17	19	17	12	12	16
Ba	240	290	90	470	520	120	150	180	120	260	110	560	260
Be	2	<1	4	3	3	3	7	6	7	4	7	24	6
Co	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5	<5
Cr	<5	<5	<5	<5	6	<5	<5	<5	<5	<5	<5	<5	<5
Cs								20	20	20	30	10	20
F	310	280	310	390	390	370	470	500	490	390	390	280	380
Ga	25	22	26	15	21	26	22	20	22	23	20	10	23
Li	44	15	52	35	39	74	130	132	130	79	56	34	70
Mo	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10
Nb	9	<1	3	6	4	7	15	11	10	2	6	55	3
Pb	35	39	35	30	31	27	46	32	26	32	28	45	34
Rb	270	180	280	240	200	110	350	370	430	280	270	250	310
Sc	2	2	2	2	2	2	1	1	1	2	2	1	2
Sn	4.8	2.1	6.3	6.1	3.6	7.7	6.7	10.1	7.4	5.2	8.0	2.0	5.9
Sr	35	75	35	95	105	55	50	35	35	55	40	13	75
Th	20	<10	10	30	40	20	20	20	20	20	10	40	20
U	5	5	2	3	6	2.4	2	2	2	2	2	4	3
V	4	3	3	8	12	3	2	2	2	2	2	1	3
Y	3	4	3	5	4	5	4	2	2	3	3	2	3
Zn	26	30	32	47	46	25	46	33	44	33	40	38	34
Zr	25	45	25	125	170	40	35	30	40	35	35	9	50

ELEMENT RATIOS

	F9-2	F9-3	F10-1	F11-1	F11-2	F11-6A	F11-6B	F13-1B	F13-1A	F16-3A	F16-3C	F16-4	F18-1A
K/Rb	157	231	115	135	173	83	90	93	90	112	99	109	109
K/Rs	154	143	357	71	66	353	204	193	256	120	242	93	139
Al/Ga	1240	3740	3100	500	3740	1160	3690	3990	3790	3500	3950	10550	3410
Na/K	0.5	0.5	1.0	0.5	0.7	0.5	1.1	1.0	1.0	1.0	1.2	0.6	0.7
Ca/Sr	167	90	146	79	79	52	107	154	196	103	139	709	90
Ba/Rb	0.3	1.6	0.3	2.0	2.6	0.2	0.4	0.4	0.3	0.9	0.4	2.2	0.5
40K/Sr	2.7	2.4	2.0	2.5	1.9	9.3	7.6	10.6	12.3	5.1	6.5	6.3	6.4
Pb/Pt	4	12	4	30	6.7	5	10	10	10	10	5	10	6.7
137Ba/138Ba	1.055	1.033	1.020	1.022	1.021	1.044	1.028	1.027	1.020	1.025	1.027	1.033	1.062

Appendix 6, continued

MNR 5

	F9-2	F9-3	F10-1	F11-1	F11-2	F11-6A	F11-6B	F11-1B	F11-1A	F16-1A	F16-3C	F16-4	F19-1A
Quartz	30	34	26	27	33		34	25		29	27	5	32
K-feldspar	32	29	30	34	29		26	25		25	25	36	30
Plagioclase	34	29	39	34	33		31	42		40	39	43	32
Biotite	1	4	1	4	3		2	1		3	2	5	2
Almandine	X	X	X				X	X		X	X	X	
Sillimanite	1	3		1				X		1			1
Muscovite	1	2	2	X	2		7	2		X	7	5	3
Total	99	101	100	100	100		100	95		100	100	100	100
Colour Index	1	4	3	4	3		2	1		3	2	5	2
Plagioclase Composition	An12	An17	An9	An15	An17	An3	An7	An10	An4	An10	An10	An17	An13

BARTON RE SUBMITS

	F9-2	F9-3	F10-1	F11-1	F11-2	F11-6A	F11-6B	F11-1B	F11-1A	F16-1A	F16-3C	F19-1A	F16-4
Albite	32.7	25.3	37.3	30.7	29.3	25.7	37.2	35.1	40.2	35.7	37.2	30.3	33.5
Anorthite	2.3	0.0	1.0	1.6	1.2	0.0	1.6	1.7	0.2	1.2	1.1	1.1	1.9
Orthoclase	25.9	29.3	22.4	22.3	23.4	30.2	21.5	21.6	23.9	21.7	18.2	25.2	35.5
Quartz	33.1	35.6	33.0	36.4	36.1	35.6	32.5	35.1	29.1	36.4	35.5	35.7	17.6
Apatite	0.2	0.3	0.3	0.2	0.2	0.1	0.3	0.3	0.3	0.2	0.2	0.3	0.2
Sphene	0.4	0.2	0.4	0.6	0.7		0.1	0.3	0.4	0.4	0.4	0.2	0.8
Calcite	0.2	1.3	0.4	0.8	1.1	0.7	0.4	0.3	0.6	0.3	0.6	0.3	1.0
Corundum	4.1	6.2	4.3	4.2	5.3	6.1	4.3	4.3	4.1	4.4	4.6	4.7	6.9
Biotite	1.0	1.4	1.0	2.4	2.6	1.4	0.5	0.5	0.5	1.1	1.6	1.6	1.0
Magnetite			0.1		0.3		0.3	0.3	0.1	0.1			0.2

Appendix 7. Chemical Analyses Selected Element Ratios, Modal Analyses and Barth Mesonorms of GLB-4, -5, -6, -7 and the Zealand Stock.

	C35-2	E21-4A	E29-1	E31-1	E31-4	E31-7	E31-8A	E31-8B	E31-9A	E31-9B	F22-5	F27-3
SiO ₂	75.40	74.50	74.80	75.60	73.40	66.80	75.80	75.80	76.70	74.50	72.00	74.70
Al ₂ O ₃	14.90	15.60	15.00	14.80	15.40	19.20	14.90	14.60	15.00	15.00	15.90	14.90
Fe ₂ O ₃	0.0	0.32	0.0	0.53	0.68	1.14	0.06	0.19	0.11	0.49	0.10	0.33
FeO	0.40	0.40	0.40	0.18	0.80	1.37	0.37	0.64	0.51	0.18	0.64	0.64
HgO	0.0	0.0	0.0	0.0	0.0	0.13	0.0	0.01	0.00	0.0	0.13	0.18
CaO	1.05	0.42	0.52	0.39	0.50	1.05	0.73	0.59	0.49	0.56	0.89	0.65
Na ₂ O	3.57	3.27	4.97	4.64	5.61	6.15	4.50	3.58	2.92	4.25	6.08	2.51
K ₂ O	4.25	5.13	3.62	3.41	2.71	2.53	2.91	4.13	3.39	3.62	3.66	5.61
TiO ₂	0.35	0.12	0.11	0.23	0.07	0.16	0.24	0.22	0.30	0.19	0.12	0.08
P ₂ O ₅	0.10	0.14	0.16	0.11	0.12	0.11	0.11	0.09	0.07	0.10	0.12	0.09
S	0.01	0.01	0.01	0.01	0.01	0.0	0.01	0.01	0.0	0.0	0.01	0.0
MnO	0.07	0.15	0.01	0.04	0.27	0.25	0.02	0.03	0.01	0.06	0.03	0.02
CO ₂	0.18	0.10	0.27	0.14	0.31	0.34	0.14	0.07	0.05	0.07	0.10	0.15
H ₂ O ₊	0.12	0.16	0.27	0.14	0.10	0.54	0.38	0.55	0.13	0.35	0.24	0.56
H ₂ O ₋	0.09	0.06	0.06	0.05	0.05	0.10	0.02	0.06	0.06	0.07	0.05	0.10
Total	100.42	100.60	100.20	100.27	99.56	99.57	100.49	100.60	99.76	99.47	100.14	99.97
GLB Unit	4	7	4	5	5	Zealand Stock	Zealand Stock	Zealand Stock	5	6	4	4

Appendix 7, continued

	C35-2	E21-4A	E22-1	E31-1	E31-4	E31-7	E31-9A	E31-9B	E31-9A	F31-9B	F22-5	F27-3
B	20	920	15	2300	2500	900	53	21	2000	600	86	210
Ba	240	70	20	20	20	40	40	130	10	30	90	250
Bc	2	2	2	3	6	5	3	2	3	3	3	2
Co	5	5	5	5	5	5	5	5	5	5	5	5
Cr	5	5	5	5	5	5	5	5	5	5	5	5
Cs												
F	270	460	290	410	310	560	420	220	200	580	470	380
Ga	16	21	12	33	29	37	26	21	23	19	13	13
Li	7	5	25	16	52	95	36	59	28	195	30	22
Mo	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10	<10
Nb	1	2	5	1	15	30	3	4	1	40	1	3
Pb	45	13	13	17	16	14	20	23	20	18	19	20
Rb	150	430	300	390	270	230	210	270	310	320	140	210
Sc	2	2	2	2	3	5	2	2	2	10	2	3
Sn	0.8	6.3	3.6	4.9	4.2	13.3	5.6	6.2	2.7	7.1	6.5	2.8
Sr	135	35	12	12	12	21	15	20	13	16	35	55
Th	70	20	20	20	10	<10	10	10	20	10	20	20
U	2	5	1	2	1	2	3	8	3	3	7	1
V	3	4	3	3	3	2	2	3	2	2	2	3
Y	9	7	2	14	10	9	4	4	11	4	4	6
Zn	10	16	42	23	26	64	27	36	23	36	26	37
Zr	50	35	5	40	40	20	13	45	30	1	16	50

ELEMENT RATIOS

	C35-2	E21-4A	E22-1	E31-1	E31-4	E31-7	E31-9A	E31-9B	E31-9A	F31-9B	F22-5	F27-3
Al/Rb	237	99	100	73	70	91	115	127	91	94	217	204
K/Fa	145	606	1502	1415	917	525	604	264	2405	1002	300	171
Al/La	4900	3940	6630	2280	2610	2750	3040	3690	3460	2040	6480	6040
Rb/K	0.5	0.55	1.3	1.3	2.4	2.3	1.6	0.3	0.8	1.1	1.6	0.5
Ca/Sr	56	96	312	234	300	360	350	212	271	252	181	95
Ba/Rb	1.6	0.2	0.07	0.05	0.07	0.2	0.2	0.5	0.03	0.09	0.6	1.2
Mo/Sr	1.1	12.3	25	12.5	22.5	11	14	13.5	24	20	4	3.4
Th/U	35	4	>10	2.9	10	<5	3.3	1.2	6.7	3.3	2.9	6.7
Al/203: CaO-Ha2O + K2O *	1.194	1.336	1.152	1.232	1.231	1.302	1.205	1.279	1.606	1.259	1.023	1.360

* molecular ratio

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MICROD

Appendix 8. Chemical Analyses, Selected Element Ratios,
Modal Analyses and Barth Mesonorms of Barren,
Exocontact Pegmatite Dykes.

	D34-3	E9-1	E9-2	F22-1	F22-60	30-17-2
SiO ₂	76.40	72.80	73.30	78.80	74.90	78.50
Al ₂ O ₃	15.00	15.10	15.80	13.20	16.20	12.20
Fe ₂ O ₃	0.12	0.0	0.00	0.0	0.13	0.0
FeO	0.64	0.64	0.40	0.89	0.40	0.48
MgO	0.01	0.22	0.01	0.17	0.01	0.02
CaO	0.89	1.12	1.31	0.27	0.49	0.16
Na ₂ O	3.03	2.68	3.27	3.10	3.28	3.15
K ₂ O	2.74	5.08	4.19	1.98	2.93	3.48
TiO ₂	0.24	0.31	0.51	0.37	0.19	0.47
P ₂ O ₅	0.10	0.12	0.14	0.13	0.12	0.08
S	0.01	0.01	0.0	0.01	0.01	0.01
MnO	0.17	0.01	0.02	0.04	0.06	0.07
CO ₂	0.12	0.18	0.21	0.10	0.09	0.15
H ₂ O ⁺	0.65	0.16	0.12	0.76	0.62	0.23
H ₂ O ⁻	0.09	0.09	0.10	0.10	0.18	0.06
Total	100.20	98.51	99.38	99.92	99.61	99.46

Appendix 8, continued

	D34-3	E9-1	E9-2	F22-1	F22-60	30-17-2
B	220	<5	7	69	34	13
Ba	80	1210	980	40	170	50
Be	2	<1	2	11	7	3
Co	<5	<5	<5	<5	<5	<5
Cr	<5	<5	<5	<5	<5	<5
Cs	<10	<10	10	120	10	N.D.
F	100	70	130	750	570	220
Ga	30	13	17	39	39	41
Li	16	8	16	255	25	14
Mo	<10	<10	<10	<10	<10	<10
Nb	15	<1	1	35	35	110
Pb	15	36	25	<10	10	11
Rb	190	120	280	450	330	1470
Sc	7	3	2	1	2	2
Sn	3.2	0.8	2.9	32.7	12.8	88.0
Sr	40	225	180	12	20	17
Th	20	40	20	10	10	40
U	3	7	3	<1	5	2
V	4	10	5	4	1	5
Y	10	7	3	1	2	3
Zn	18	20	32	70	24	94
Zr	15	135	60	<1	8	25
Ta	<30	<30	<30	35	<30	50

ELEMENT RATIOS

	D34-3	E9-1	E9-2	F22-1	F22-60	30-17-2
K/Rb	120	324	124	37	74	22
K/Ba	284	35	36	411	347	644
Al/Ga	2650	6120	4930	1840	2200	1580
Na/K	1.05	0.5	0.7	1.5	1.1	0.8
Ca/Sr	154	36	52	150	176	69
Ba/Rb	0.4	9.3	3.5	0.09	0.2	0.03
Rb/Sr	4.75	1.6	1.6	34.6	16.5	86.5
Th/U	6.7	5.7	6.7	>10	2	20
Al ₂ O ₃ : CaO+Na ₂ O + K ₂ O	1.573	1.266	1.287	1.709	1.715	1.263

Appendix 8, continued

MODES

	D34-3	F9-1	F9-2	F22-1	F22-6D	30-17-2
Quartz	35	30	29	46	39	
K-feldspar	11	35	27	x	5	
Plagioclase	36	30	42	33	29	
Biotite	1	5	3	3	1	
Tourmaline				x		
Almandine	1			x	x	
Muscovite	16	1		18	27	
Apatite				x		
Total	100	101	101	100	101	
Colour Index	2	5	3	3	1	

BARTH RESONANCES

	D34-3	E9-1	E9-2	F22-1	F22-6D	30-17-2
Albite	27.8	24.8	29.9	29.7	30.1	29.0
Anorthite	2.1	2.5	2.4	0.0	0.4	0.0
Orthoclase	15.5	29.5	24.6	10.4	17.2	27.8
Quartz	44.7	34.8	34.7	50.8	42.0	43.1
Apatite	0.2	0.3	0.3	0.2	0.3	
Sphene	0.5	0.7	1.1		0.4	
Calcite	0.3	0.5	0.6	0.3	0.3	0.3
Corundum	7.0	4.8	5.6	6.7	8.4	3.2
Biotite	1.6	2.2	0.9	2.6	0.9	1.2
Magnetite	0.1					

Appendix 9. Partial Chemical Analyses of Brownridge Metavolcanics Adjacent to Spodumene Pegmatites of Fairservice Property (North Zone). Locations given in Figure 18.

Sample No	CaO	MgO	K ₂ O	B	Be	Li	Mn	Ba	SiO ₂	Si
1	7.39	1.07	1.74	4600	10	390	2560	1390	8.4	100
2	7.74	2.96	1.38	86	2	160	1620	200	2.3	315
3	7.90	4.21	0.13	11	2	10	990	20	2.3	120
4	8.01	4.06	0.81	12	2	150	980	170	1.5	450
5	7.11	0.92	2.47	450	20	1230	1400	3030	12.2	265
6	14.4	2.51	0.41	<5	1	<10	1730	30	1.6	235
7	16.2	1.11	0.41	7	12	10	1950	20	8.6	300
8	9.63	3.36	0.20	7	1	<10	1370	<10	2	230
9	8.73	3.98	0.65	10	1	<10	1310	30	1.9	340
10	9.60	1.72	0.19	<5	2	<10	3670	10	2.5	30
11	9.13	2.24	0.27	13	2	<10	2370	10	2.6	130
12	13.6	0.24	0.42	13	1	10	1480	10	2.4	630
13	6.80	3.50	1.41	260	4	170	1380	130	10.9	210
14	12.9	0.45	0.32	<5	1	<10	1560	10	2.6	545
15	6.98	4.34	0.86	88	2	780	910	350	2.4	275
16	8.76	3.68	0.10	5	1	<10	1180	<10	3.4	365
17	7.84	2.06	0.42	9	2	110	1090	150	5.5	370
18	7.93	2.35	0.69	5	2	160	1190	300	3.1	375
19	11.4	0.69	1.18	1200	45	400	2430	1280	64.0	95
20	7.48	3.81	0.20	70	1	20	1450	30	4.1	495
21	8.51	2.88	0.43	30	2	70	1950	20	3.2	150
22	13.6	1.85	0.46	13	2	10	1960	30	3.8	400
23	13.1	1.88	0.39	18	1	<10	1340	10	3.4	580
24	8.04	3.55	0.21	6	2	10	1360	<10	2.2	310
25	9.59	3.03	0.38	<5	1	70	1030	110	1.8	295
26	7.68	2.97	0.41	94	1	80	970	70	3.6	185
27	9.67	3.41	0.39	7	1	20	1130	20	12	425
28	8.91	2.75	0.20	11	1	<10	2380	<10	2.5	90
29	6.89	2.52	1.15	1200	1	400	1690	330	2.2	155
30	8.54	0.57	0.15	15	1	50	1530	30	2.0	50
31	8.61	1.84	0.10	15	1	20	1060	10	1.9	330
32	10.3	0.26	1.54	900	6	280	1840	820	3.5	265
33	8.71	2.45	1.23	<5	1	60	1180	60	4.1	155
34	18.1	6.92	2.5	10	2	110	1670	210	3.3	95
35	8.06	2.39	0.60	260	2	130	1750	200	1.9	190

Appendix 9, continued

36	6.58	3.55	0.99	125	1	60	410	690	1460	110	2.5	130
37	8.33	2.96	0.20	105	2	10	210	695	1850	<10	1.8	190
38	8.32	4.20	0.19	12	2	10	500	122	1380	10	2.7	290
39	7.61	2.57	0.38	7	1	40	170	119	1230	30	4.9	325
40	9.76	3.41	0.24	68	2	10	370	152	1710	10	2.7	185
41	8.83	3.32	0.21	<5	1	<10	250	63	1280	10	1.6	140
42	8.31	4.38	0.14	<5	2	<10	460	54	1340	<10	2.5	465
43	8.05	3.76	1.01	8	3	100	440	82	1460	50	2.7	355
44	7.68	4.63	0.11	6	4	<10	1010	161	1120	10	4.1	205
45	6.24	4.31	0.71	11	3	20	930	110	1360	10	2.1	360
46	8.56	3.97	0.32	<5	2	10	370	93	1390	10	1.6	415
47	13.2	0.98	0.18	9	1	<10	1130	330	1570	10	1.6	275
48	6.77	1.76	0.24	5	2	30	310	570	1450	40	1.6	205
49	6.07	4.55	0.55	<5	2	60	920	66	1320	20	2.5	280
50	8.85	4.17	0.32	55	2	<10	440	36	1290	10	2.2	285
51	8.02	2.73	0.55	400	2	40	250	805	1910	40	2.5	155
52	8.19	3.37	0.29	42	2	10	190	480	1840	<10	2.2	165
53	6.00	4.03	0.13	47	10	10	1210	3420	1680	<10	8.7	225
54	6.92	0.68	1.14	4000	14	360	2420	3680	2460	530	37.5	40
55	10.7	0.62	1.17	125	15	180	6500	2750	2850	680	35.0	225
56	10.1	1.31	0.82	1200	2	170	870	1650	1920	330	2.6	279
57	12.7	1.14	0.25	130	1	10	570	380	2710	20	2.0	215
58	11.5	1.02	0.17	400	2	10	820	430	1930	20	2.5	255
59	9.72	1.89	0.18	125	2	<10	180	680	1880	10	3.1	140
60	8.79	3.08	0.56	780	2	30	380	860	1470	40	1.6	145
61	9.15	2.89	0.60	39	3	20	430	675	1970	40	2.8	245
62	7.79	4.73	0.17	33	2	10	380	143	1230	<10	1.6	410
63	9.86	0.68	0.90	5800	17	200	7300	410	1230	350	32.0	40
64	3.19	0.86	1.22	1.82	2	410	2620	1200	410	340	9.0	210
65	8.42	2.87	0.47	76	2	20	300	75	1890	20	2.7	125
66	9.21	2.11	0.38	2100	2	90	640	435	1370	60	4.4	270
67	5.78	4.43	0.05	40	2	<10	540	91	1330	10	2.4	210
68	7.39	4.24	0.13	60	2	10	510	88	1500	<10	2.3	265
69	8.21	3.31	0.10	49	1	<10	300	490	1110	10	2.3	375
70	9.00	2.61	0.57	53	1	30	510	250	1430	40	2.1	295
71	12.6	1.89	0.38	35	1	<10	150	60	1980	20	1.4	285
72	7.96	1.07	2.19	3900	21	870	5900	2850	2340	1870	45.0	265
73	13.4	1.24	0.21	37	2	30	800	136	1230	30	2.4	165
74	11.5	2.12	0.21	41	1	10	500	120	1170	50	3.1	470
75	14.1	0.65	0.23	80	3	10	550	41	3110	<10	3.4	115
76	9.28	3.25	0.13	<5	1	<10	330	68	1320	<10	1.7	325
77	8.10	3.86	0.56	<5	1	140	550	470	1060	210	2.0	225

Appendix 9, continued

78	9.93	2.18	0.95	50	1	20	270	240	2540	90	2.7	120
79	6.39	4.36	2.43	<5	2	60	1130	480	810	60	2.4	1250
80	9.46	0.80	0.60	1500	2	40	820	805	1530	110	4.3	80
81	6.10	3.89	0.40	47	2	80	180	1900	1960	130	2.6	115
82	10.5	2.25	0.55	6	1	10	190	150	1640	20	1.9	240
83	12.2	1.80	0.63	7	1	10	890	176	1330	40	2.8	325
84	11.0	1.99	1.02	5	1	30	870	410	1240	20	2.8	365
85	11.0	2.10	0.22	<5	1	10	410	310	1140	10	3.9	325
86	12.70	2.00	0.32	5	1	<10	990	111	1330	<10	1.3	320
87	11.5	2.00	0.68	10	1	40	900	335	1390	80	3.1	420
88	8.67	3.10	0.93	<5	2	70	710	375	1390	150	2.3	695
89	8.75	3.77	0.18	13	1	<10	350	186	1140	20	2.3	540
90	10.9	2.34	0.24	<5	1	<10	350	300	1280	<10	2.3	430
91	12.4	1.03	0.29	330	16	10	2400	970	2080	70	32.0	190
92	8.00	1.11	0.87	56	15	310	1540	3680	1850	350	31.2	270
93	6.28	0.36	2.38	1600	2	720	1280	2300	1530	1130	2.9	40
94	10.00	1.59	0.34	1300	4	10	1830	725	2070	20	3.4	90
95	8.19	0.99	1.87	50	25	910	4400	2250	2260	2480	70.0	120
96	7.19	1.35	0.83	4300	13	290	2320	3020	1250	500	21.0	80
97	6.69	4.43	0.29	80	2	50	330	1160	1490	40	3.2	225
98	14.70	1.65	0.18	6	6	<10	2280	54	1720	10	3.9	265
99	7.84	4.64	0.23	280	2	<10	370	148	1090	10	2.9	190
100	6.04	3.32	0.83	76	2	120	430	290	2230	60	4.2	195
101	4.38	5.37	0.21	930	1	40	300	320	1100	30	3.4	55
102	9.88	3.32	0.29	35	2	<10	410	151	1480	<10	4.4	210
103	7.62	2.65	0.59	680	2	140	550	1060	1750	140	6.5	220
104	8.59	3.03	0.22	88	2	10	210	510	1720	10	2.5	170
105	8.00	2.81	0.33	300	2	10	240	700	1830	20	2.9	100
106	7.17	1.35	2.97	64	45	630	7800	3280	1670	2460	104.0	450
107	8.13	2.57	1.01	200	3	150	1210	2100	1910	460	5.1	240
108	9.23	1.20	0.82	2000	12	100	4600	1460	2020	210	5.6	170
109	4.95	1.19	3.28	1150	60	840	9000	2880	3780	3990	108.0	270
110	8.42	0.71	2.05	1350	6	430	1440	2920	1860	860	12.0	195
111	13.00	1.26	0.36	13	4	20	4750	1140	1790	60	3.2	280
112	14.5	0.74	0.13	5	8	10	1640	325	1730	10	6.8	275
113	9.07	2.80	0.13	<5	1	<10	220	675	1270	<10	3.6	285
114	8.28	2.95	0.56	6	2	170	240	525	1940	130	2.7	130
115	9.14	1.94	0.25	<5	2	10	180	130	1680	10	5.7	130
116	8.54	3.70	0.12	125	2	10	960	156	1630	<10	2.7	140
117	11.80	1.63	0.37	220	7	10	700	380	2510	20	3.4	305
118	8.81	2.25	0.19	45	2	<10	220	183	1770	10	4.2	170
119	6.45	3.71	0.10	11	2	<10	510	56	1380	<10	2.6	90

Appendix 9, continued

120	8.54	2.85	0.19	500	2	50	390	580	1800	40	4.6	140
121	9.26	2.97	0.26	35	4	30	800	94	1650	20	7.6	115
122	11.70	0.13	0.90	26	18	90	2420	2380	1320	480	41.5	250
123	8.85	0.71	1.73	740	12	940	6500	2220	1040	1440	28.2	155
124	10.70	0.43	1.30	1250	18	250	6500	2520	1960	890	46.5	175
125	12.90	0.48	0.81	15	16	150	3300	2480	1740	640	34.5	230
126	13.20	1.94	0.62	<5	2	10	390	191	1520	80	3.9	270
127	4.0	2.00	2.04	21	5	80	670	430	1940	90	8.0	215
128	6.67	2.02	0.29	<5	2	<10	430	129	2490	<10	3.0	190
129	8.47	2.35	0.31	13	2	10	330	150	1650	40	3.8	230
130	9.03	2.59	0.18	19	2	<10	220	110	1600	<10	2.8	135
131	9.26	2.89	2.80	130	5	130	90	365	580	470	2.8	480
132	8.04	3.40	0.22	540	2	<10	590	60	1330	<10	4.9	235
133	5.22	5.37	0.29	22	2	30	450	235	500	70	4.9	175
134	6.83	3.08	0.19	12	2	10	340	64	1840	<10	3.2	140
135	8.14	3.10	0.21	58	2	10	260	38	1440	<10	13.0	130
136	7.01	3.49	0.50	11	3	100	690	91	1810	40	4.2	100
137	7.98	3.29	0.20	5	2	20	190	76	1910	<10	3.9	40
138	8.78	2.85	0.18	9	2	<10	240	46	1700	<10	4.0	185
139	7.97	4.10	0.17	11	2	30	240	47	1650	<10	3.7	320
140	13.40	1.31	0.39	6	2	<10	1010	101	1300	<10	2.0	260
141	8.41	0.98	0.21	55	2	90	710	525	1190	40	3.3	35
142	5.13	4.93	0.81	14	3	70	660	410	580	60	2.7	1145
143	8.88	2.97	0.30	13	2	40	600	109	1020	20	2.8	235
144	11.40	2.13	0.59	5	2	80	900	260	1210	90	2.1	360
145	7.88	2.46	0.27	<5	2	<10	320	142	2290	<10	2.4	110
146	5.57	3.55	0.52	<5	2	40	350	162	1340	50	2.9	185
147	6.57	3.46	0.33	<5	2	10	230	78	2050	10	2.8	85
148	6.37	3.88	0.14	<5	2	<10	210	53	1690	<10	2.8	45
149	12.80	1.19	0.20	<5	1	10	590	295	1830	<10	3.0	155
200	12.60	1.11	0.36	57	14	20	4630	530	1950	60	11.0	600
201	8.77	2.21	0.63	300	2	190	963	255	1870	90	1.0	145
202	6.92	3.02	0.79	9	2	30	220	148	2180	30	1.2	185
203	4.45	5.35	0.19	640	<1	10	625	100	1060	<10	1.1	525
204	14.8	1.20	0.14	6	2	<10	910	88	1910	<10	2.3	275
205	13.6	1.04	0.18	10	1	<10	510	152	1550	<10	1.2	255
206	7.33	3.22	0.68	520	2	30	160	285	1700	30	0.8	90
207	13.30	1.89	0.46	11	2	20	1130	172	1770	10	2.3	315
208	9.79	1.64	0.80	<5	10	70	2690	1140	1450	180	20.4	335
209	13.70	0.62	0.24	11	1	10	590	355	1640	40	2.1	190
210	12.50	3.23	0.48	<5	2	10	1040	170	1240	<10	2.3	355
211	8.01	2.52	0.19	880	1	20	120	140	1240	<10	1.7	85

Appendix 9, continued

212	10.8	2.58	0.32	1300	1	10	830	94	1740	<10	2.9	225
213	9.14	2.07	0.12	<5	1	<10	90	148	1350	<10	1.6	70
214	8.78	2.01	0.19	4000	3	20	660	425	1300	20	4.7	55
215	8.35	3.04	0.42	310	1	20	130	116	1290	<10	1.4	205
216	9.22	2.00	0.14	<5	1	<10	110	136	1190	<10	1.7	150
217	9.19	2.40	0.13	<5	2	<10	160	330	1500	<10	1.6	115
218	10.80	1.66	0.19	300	1	10	230	480	1250	10	1.5	240
219	6.48	3.62	0.17	135	1	<10	120	116	1190	<10	2.1	295
220	14.10	0.51	0.37	205	15	10	2960	170	1600	50	13.2	22
221	9.48	1.66	0.30	15	2	30	720	790	1280	70	1.7	95
223	7.41	0.21	2.04	1850	12	1750	3920	2750	1900	2600	35.3	55
224	9.64	0.62	1.26	1950	12	800	5100	565	1130	1230	7.6	35
225	11.40	2.15	0.19	16	2	10	300	151	1230	10	1.8	265
226	4.37	4.40	0.15	1500	2	20	500	110	1280	10	2.7	160
227	7.89	3.89	0.29	900	2	<10	190	86	1380	<10	1.5	240
228	7.92	3.58	1.53	6	3	70	610	370	1120	50	3.1	410
229	7.63	1.16	0.13	410	2	30	210	395	1630	20	5.6	45
230	12.30	0.69	0.36	94	7	20	1580	185	1560	50	3.6	35
231	10.20	1.52	0.36	14	3	10	1300	162	1520	40	2.9	145
232	10.20	2.23	0.10	7	1	10	120	300	1310	10	1.9	200
233	2.23	4.97	0.27	84	4	40	270	119	750	40	5.9	60
234	7.08	2.62	0.32	18	2	60	280	310	1870	40	1.8	85
235	7.96	3.31	0.14	<5	2	<10	150	275	1740	10	1.4	170

Appendix 10. Partial Analyses and Selected Element Ratios of K-feldspar from Unit GLB-1 of Ghost Lake Batholith.

	B17-1	B17-2	C15-1	C20-1	C20-3	C26-1	D14-1	D14-6	D21-1	E11-4	E19-8	E21-3
Al ₂ O ₃												19.1
CaO									0.14	0.15	0.14	0.21
Na ₂ O									1.99	2.28	2.12	2.25
K ₂ O	13.2	12.9	13.0	12.7	13.3	13.6	13.4	12.5	13.1	12.5	12.6	12.5
Ba	910	80	1370	1530	1720	1790	1090	3680	1500	370	3310	500
CS	<5	5	<5	<5	<5	<5	7	<5	<10	<10	<10	10
Ga												13
Li	4	7	4	6	4	3	4	3	<3	<3	4	5
Rb	260	250	240	240	240	290	250	210	270	290	260	360
Sr	370	120	310	370	440	500	520	610	315	165	590	190
Pb												77
Ba/Rb 10 ⁻³	3.5	2.30	4.52	7.57	6.14	6.17	4.36	17.5	5.56	1.32	1.27	1.39
Rb/Sr	0.70	2.3	0.9	0.65	0.54	0.55	0.43	0.34	0.86	1.70	0.44	1.9
K/Ba	1.0	145	79	50	64	63	105	25.9	73	290	32	208
K/CS	>21912	1184	>21580	>21052	22075	>22576	16361	>21248	>10573	>10375	>10458	10375
K/Hb	4.21	14.2	355	4.19	394	352	455	506	402	371	402	295
Hb/CS	52	45	50	45	50	55	30	42	440	>25	>26	36

Appendix 10, continued

	E21-4A	E21-5	E23-1	E25-3	E25-5	E25-7	E25-11	E25-12	E27-1	F9-4	Average
Al ₂ O ₃				19.3	19.3		19.6		19.3		19.3
CaO	0.15	0.16	0.15	0.13	0.20	0.11	0.11	0.07	0.15	0.12	0.14
Fe ₂ O ₃	2.54	1.72	1.91	2.10	2.21	1.92	1.57	1.61	2.03	2.13	1.77
MgO	12.1	13.2	13.1	12.5	12.5	13.0	13.2	13.6	12.9	12.6	12.9
Ba	2000	1200	410	900	2040	1980	7790	260	1960	270	1655
Ca	10	<10	<10	10	<10	<10	<10	<10	<10	<10	<10
Li				13	10		5		10		11
La	5	<3	5	6	3	<3	<3	3	<3	6	<4
Hb	290	290	300	400	310	190	270	290	370	430	290
Sc	450	415	195	210	410	310	910	115	330	90	360
Pb				71	77		94		76		79
Ba/Rb x 10 ⁻³	6.90	4.29	1.37	2.25	6.58	7.26	2.58	9.29	5.10	0.12	6.24
Rb/Sc	0.64	0.67	1.54	1.9	0.76	0.61	0.30	2.4	1.12	4.8	1.19
K/Ba	50	91	265	115	51	78	14	434	55	387	181
K/Cs	1004.3	>10956	>10873	>10375	>10375	>10790	>10956	>11298	>10707	>10458	
K/Rb	346	391	362	259	335	568	406	403	289	243	344
Rb/Cs	45	>25	>30	40	>31	>19	>27	>24	>37	>43	

Appendix 11. Partial Analyses and Selected Element Ratios of K-feldspar from Unit G1B-3 of Ghost Lake Batholith.

	F7-1	F7-7	F9-1	F9-2	F10-1	F11-1	F11-2
CaO			0.05				0.08
K ₂ O	11.5	12.0	11.6	12.4	12.1	12.6	12.9
Na ₂ O			2.91				2.00
Ba	120	110	40	140	60	320	450
Cs	18	34	20	17	25	21	<10
Li	9	26	74	12	35	8	6
Rb	670	770	660	590	750	510	440
Sr	40	10	21	20	30	60	140
Ba/Rb	0.15	0.14	0.05	0.24	0.11	0.63	1.02
Rb/Sr	16.9	77	41	25.5	25	6.4	3.1
K/Ba	795	905	2407	735	134	327	238
K/Cs	5303	2929	4614	6054	4017	4990	>10707
K/Rb	159	129	112	174	134	205	243
Rb/Cs	74	23	43	35	30	24	>44

Appendix 11, continued

	F11-3	F11-4	F13-2	F16-1	F16-3A	F18-1	Average
CaO		0.05				0.08	0.07
K ₂ O	11.8	11.2	12.1	11.7	11.9	12.1	12.0
Na ₂ O		3.17				2.68	2.69
Ba	30	60	60	60	60	80	124
Cs	23	40	33	17	16	30	<24
Li		27	42	20	29	7	23
Rb	770	1030	770	660	640	880	719
Sr	<10	23	<10	30	30	45	38
Ba/Rb	0.04	0.06	0.08	0.09	0.09	0.09	0.22
Rb/Sr	>77	44.9	>77	22	21.3	19.6	>33
K/Ba	3265	1549	1674	1610	1646	1255	1273
K/Cs	4258	2324	3043	5712	5487	3348	4844
K/Rb	127	90	130	147	154	114	148
Rb/Cs	33	26	23	39	36	29	35

Appendix 12. Partial Analyses and Selected Element Ratios of K-feldspar from Units GLB-4 and -7 of Ghost Lake Batholith.

GLH Unit	C35-2	C35-3	D28-1	D34-7A	D34-9	D34-10	E21-4B	E21-4C	E29-1	E31-2
Al ₂ O ₃	4	4	4	4	4	4	4	7	4	7
CaO			0.21	0.07			0.12	0.13	0.11	0.19
K ₂ O	13.4	13.1	13.5	12.4	12.9	13.1	12.6	12.2	11.2	12.5
Na ₂ O			1.55	2.35			2.28	2.03	3.00	2.03
Ba	290	100	680	20	150	470	900	50	50	40
Cs	22	13	<10	30	13	7	<10	20	<10	350
Ga				25					17	
Li	5	7	5	9	15	5	4	3	<3	14
Pb	10			30					20	14
Rb	440	420	510	1000	630	490	360	770	900	990
Sr	110	110	115	30	60	240	205	30	20	--
Ba/Rb	0.66	0.24	1.33	0.20	0.24	0.96	2.50	0.065	0.05	0.044
Rb/Sr	4.0	3.8	4.4	3.3	10.5	2.0	1.9	25.7	45	--
K/Rb	304	1057	165	5140	714	231	116	2025	1859	2594
K/Cs	5055	8364	>11205	3430	5236	15534	>10455	5063	>9296	296
K/Rb	253	259	220	103	170	222	291	132	103	118
Rb/Cs	20	32	>51	33	48	70	>36	39.5	>90	2.5

Appendix 12, continued

	E31-3	E31-4	E31-7	E31-9A	E31-9B	E31-9A	E31-9B	E31-9A	E31-9B	E31-9C	E31-9D
GLB Unit	7	7	7	4	4	4	4	7	7	7	7
Al2O3	19.4	19.3	19.6	19.3	20.4	19.3	19.3	19.3	19.3		
CaO	0.11	0.07	0.10	0.13	0.82	0.05	0.08	0.07	0.07	0.05	0.05
K2O	11.6	12.0	11.9	12.4	7.52	12.9	12.3	13.0	13.0	13.2	13.2
Na2O	3.04	2.50	2.75	2.62	4.99	2.23	2.40	2.16	2.16	1.91	1.91
Ba	30	30	30	90	100	20	20	20	20	40	40
Cs	10	50	40	10	<10	1150	620	1100	1100	870	870
Ga		20	25	12	19	21	23				
Li	44	65	14	10	10	40	34	50	50	62	62
Pb	37	23	15	37	30	37	42				
Rb	2200	1220	1470	670	470	1930	1620	191	191	1790	1790
Sr	--	20	30	40	50	30	30	30	30	30	30
Ba/Rb	0.01	0.025	0.02	0.13	0.21	0.01	0.01	0.01	0.01	0.02	0.02
Rb/Sr	--	61	49	16.8	9.4	64.3	54	63.7	63.7	59.7	59.7
K/Ba	3209	3320	3292	1144	624	535	510	540	540	2739	2739
K/Cs	9629	1992	2469	10292	>6242	93	165	98	98	126	126
K/Rb	44	62	67	154	133	55.5	63	56.5	56.5	61	61
Rb/Cs	220	24.4	36.8	67	>47	1.7	2.6	1.7	1.7	2.1	2.1

Appendix 12, continued

GLB Unit	SM-1	SM-2	SM-3	SM-4	F27-4	F22-5	F24-2	Average	Average
Al ₂ O ₃	7	7		7	4	4	4	4	7
CaO	0.06	0.05	0.06	0.06	0.05	0.13	0.11	0.19	0.9
K ₂ O	13.2	11.9	12.7	12.5	12.0	11.5	11.7	12.1	12.5
Na ₂ O	1.79	2.62	2.21	2.19	2.35	1.13	2.88	2.55	2.29
Ba	30	<10	20	20	130	200	90	304	29
Ca	40	60	70	50	10	20	<10	14	341
Ga						11		17	22
Li	30	65	27	7	6	6	8	<7	35
Pb						53			
Rb	1530	1690	2600	1990	910	390	660	596	1678
Sr	30	25	40	30	25	50	50	87	32.5
Ba/Rb	16.4	<5.9	7.7	10.1	160.5	513	136	596	20.5
Rb/Sr	51	67.6	65	66.3	32.4	4.9	13.2	13.9	60.7
K/Ba	3652	>9577	5271	5155	766	477	1079	1314	4533
K/Cs	2739	1646	1506	2075	9960	4773	>5711	>8858	2119
K/Rb	80	55	40.5	52	123	245	147	186	69
Rb/Cs	45.6	>169	130	40	51	19.5	>66	>54	>53

Appendix 13. Partial Analyses and Selected Element Ratios of K-feldspars from Pegmatite Dykes Emplaced in Medium Grade Metasediments Adjacent to South Contact of the Ghost Lake Batholith.

	E29-4	F6-1	F7-3	F22-2	F24-5	F30-3	32-4-20	F24-3	F26-4	AVERAGE
Al ₂ O ₃	18.8									
CaO	0.10						0.07	0.07	0.03	0.07
K ₂ O	10.3	12.1	12.6	12.8	13.4	13.8	12.2	12.5	12.6	12.5
Na ₂ O	3.08						2.54	2.34	2.19	2.54
Ba	20	100	260	60	40	70	10	60	80	78
Cs	10	24	19	84	77	92	50	90	50	55
Ga	16									
Li	7	3	9	14	7	29	7	<3	14	<10
Pb	20									
Rb	920	530	670	1370	1320	1790	920	1890	1430	1204
Sr	20	70	40	30	10	10	20	35	35	30
Ba/Rb	0.02	0.02	0.04	0.04	0.03	0.04	0.01	0.03	0.06	0.09
Rb/Sr	46	7.6	16.8	45.7	132	179	46	54	40.9	63.1
K/Ba	4275	1004	409	1771	2781	1636	10126	1729	1307	2782
K/Cs	8549	4200	5600	1270	1400	1200	2025	1153	2092	3054
K/Rb	96	189	159	77	132	90	110	55	73	109
Rb/Cs	92	22	35	16	17	20	18	21	29	30

Appendix 14. Partial Analyses and Selected Element Ratios of K-feldspar from Pegmatite Dykes Related to Northeastern Contact of Ghost Lake Batholith and Emplaced in Medium Grade Host Rocks.

	C35-1	D33-1	D34-3	AVERAGE
Al ₂ O ₃	18.9			
CaO	0.13			
K ₂ O	13.8	12.8	13.2	13.3
Na ₂ O	1.13			
Ba	1040	410	480	643
Cs	10	11	17	9
Ga	20			
Li	<3	4	4	<3.7
Pb	78			
Rb	370	520	690	527
Sr	310	250	100	220
Ba/Rb	2.81	7.88	6.96	1.43
Rb/Sr	1.2	2.06	6.9	3.4
K/Ba	110	259	228	199
K/Cs	11454	9700	6500	9218
K/Rb	309	204	159	224
Rb/Cs	37	47	41	42

Appendix 15. Partial Analyses and Selected Element Ratios of K-feldspar from Pegmatite Dykes Related to Ghost Lake Batholith and Emplaced in High Grade Host Rocks.

	C11-1	C13-2	C13-6	E5-2	E9-1	E9-3	E11-1	E11-2	E13-3	E15-2	AVERAGE
Al ₂ O ₃					19.2				19.7		19.5
CaO					0.20			0.14	0.16		0.17
K ₂ O	12.2	12.4	12.6	13.1	12.4	13.6	13.1	12.2	6.73	12.4	12.3
H ₂ O					2.29			2.29	4.53		3.04
Ba	600	2050	310	5550	760	3140	1410	4470	1290	1920	2180
Cs	<5	<5	7	<5	12	<5	5	<10	10	7	<7
Ga					12				13		12
Li	3	4	7	2	4	3	6	<3	<3	3	<3.6
Pb					56				45		51
Rb	290	260	310	230	330	220	250	190	240	230	255
Sr	100	400	150	660	210	350	330	435	230	550	342
Ba/Rb	2.07	7.69	1.00	25.44	2.30	14.27	5.64	23.53	5.39	6.35	9.36
Rb/Sr	2.9	0.65	2.07	0.35	1.6	0.63	0.76	0.44	1.0	0.42	1.08
K/Ba	169	50	343	19	135	36	77	23	56	54	96
K/Cs	>20200	>20600	15200	>21700	6577	>22600	21700	>10126	7246	14700	>16265
K/Rb	349	396	343	473	312	513	435	533	302	447	410
Rb/Cs	>58	>52	44	>46	28	>44	50	>19	24	33	36

Appendix 16. Partial Analyses and Selected Element Ratios of K-feldspar from Pegmatites 1 and 2, Taylor Beryl Occurrence.

	PEGMATITE 1						AVERAGE
	2.8m	4.2m	5.9m	6.95m	KF1	KF2	
Al ₂ O ₃					19.1	19.1	19.1
CaO	0.05	0.06	0.07	0.04	0.11	0.14	0.09
Na ₂ O	2.27	2.08	2.58	2.03	2.01	2.25	2.20
K ₂ O	12.8	13.4	12.5	13.5	13.0	12.6	13.0
Ba	100	70	60	110	110	50	91
Cs	70	70	40	70	80	40	62
Ga					18	20	19
Li	48	42	45	28	18	25	34
Pb					24	15	20
Rb	1940	1990	1930	1890	2100	2030	1980
Sr	50	40	45	50	50	40	46
Ba/Rb	0.05	0.03	0.03	0.06	0.05	0.025	0.04
Rb/Sr	39	50	43	38	42	51	43.4
K/Cs	27.7	28.4	48.3	27	26.3	50.8	34.4
K/Ba	1062	1589	1729	1019	981	2092	1412
Rb/Cs	1518	1589	2594	1601	1349	2615	1878
K/Rb	55	56	54	59	51	52	54.5

	PEGMATITE 2					AVERAGE
	KF1	KF2	KF3	KF4	KF5	
Al ₂ O ₃	19.2	19.2	19.3			19.2
CaO	0.11	0.10	0.14	0.06	0.08	0.10
Na ₂ O	2.00	1.96	2.64	1.90	2.21	2.12
K ₂ O	13.2	12.9	11.8	13.5	12.8	12.8
Ba	150	110	150	70	80	112
Cs	210	290	360	1070	470	480
Ga	25	25	24			25
Li	22	18	20	24	14	20
Pb	22	24	25			24
Rb	2630	2880	2720	3710	2850	2954
Sr	40	60	60	75	60	59
Ba/Rb	0.06	0.04	0.06	0.29	0.16	0.12
Rb/Sr	66	48	45	49	48	51
K/Cs	12.5	9.9	7.6	3.5	6.1	7.9
K/Ba	730	973	653	1601	1328	1057
Rb/Cs	522	369	272	105	226	299
K/Rb	42	37	36	30	37	36

Appendix 17. Partial Analyses and Selected Element Ratios of K-feldspar from North Zone, Fairservice Spodumene Pegmatites.

	PEGMATITE 1										AVERAGE
	85cm	1.6m	5.2m	5.9m	7.3m	9.7m	10.3m	11.2m			
Al ₂ O ₃	19.4	19.4	19.6	19.3	19.5	19.3	19.4	19.5		19.4	
CaO	0.14	0.22	0.22	0.17	0.15	0.13	0.16	0.16		0.17	
Na ₂ O	1.51	1.40	1.49	1.10	1.45	1.52	1.31	1.30		1.39	
K ₂ O	13.1	13.4	13.4	13.3	13.4	13.3	13.6	13.4		13.4	
Ba	40	50	40	30	20	20	20	20		30	
Cs	150	130	110	140	120	110	180	140		135	
Ga	26	26	26	24	22	23	24	26		25	
Li	240	146	166	166	166	141	194	151		171	
Pb	10	10	<10	<10	<10	<10	11	10		<10	
Sr	120	120	50	80	50	70	90	60		80	
Rb	7720	6540	6810	7310	6960	6660	7120	7020		6955	
Ba/Rbx 10 ⁻³	5.5	7.6	5.9	4.1	2.9	3.0	2.8	2.8		4.3	
Rb/Sr	60	54.5	136	91	139	95	79	117		96.5	
Rb/Cs	44.1	50.3	61.9	52.2	58	60.5	39.6	50.1		70	
K/Ba	2718	2224	2791	3014	5561	5520	5644	5561		4228	
K/Cs	725	856	1011	918	927	1003	627	794		713	
K/Rb	15	17	16	16	16	17	16	16		16	

Appendix 17, continued

PIGMATITE 2

	Kf 1	Kf 2	Kf 3	Kf 4	Kf 5	Kf 6	Kf 7	AVERAGE
CaO	0.06	0.06	0.09	0.10	0.10	0.07	0.10	0.09
Na ₂ O	0.85	0.95	1.45	1.32	1.26	1.34	1.29	1.22
K ₂ O	14.1	13.5	13.5	13.5	13.4	13.7	13.4	13.6
H ₂ O	30	20	30	20	30	40	40	30
Cs	420	160	120	160	120	130	170	163
Li	65	162	104	132	162	141	147	130
Rb	9060	7490	7230	7720	7210	6770	8130	7661
Sr	165	175	140	150	145	150	155	157
Ba/Rb ^a 10 ⁻³	3.3	2.7	4.1	2.6	4.2	5.9	4.9	4.0
Rb/Sr	49	43	52	52	50	45	52.5	49.1
Rb/Cs	21.6	46.5	60.3	45.3	60.1	52.1	47.8	45.1
K/Ba	3901	5727	3735	5603	3707	2643	2780	4143
K/Cs	279	716	934	700	927	575	654	726
K/Rb	13	15	16	14.5	15	17	14	15

Appendix 17, continued

PEGMATITE 3				PEGMATITE 4							
	Kf 1	Kf 2	AVERAGE	Kf 1	Kf 2	Kf 3	Kf 4	Kf 5	Kf 6	Kf 7	AVERAGE
CaO	0.03	0.03		0.14	0.10	0.05	0.10	0.06	0.16	0.13	0.11
Mn2O	1.41	1.50		0.80	1.07	1.50	0.53	1.31	0.96	1.15	1.09
K2O	13.3	13.6		14.0	13.5	13.4	13.9	13.5	13.6	13.4	13.6
Ba	40	100	70	20	20	50	30	40	40	20	31
Cs	300	130	215	240	240	120	230	140	490	250	244
Li	52	34	43	139	132	27	117	26	191	164	114
Rb	7500	5550	6525	5740	5690	7070	5680	7570	9820	8910	8497
Sr	175	160	167.5	150	175	160	175	170	240	195	185
Ba/Rb x 10 ⁻³	5.1	17.1	11.1	2.3	2.3	7.0	3.4	5.3	4.1	2.2	3.4
Rb/Sr	44.6	37	40.4	49	50	44	50	44.5	41	46	46
Rb/Cs	26	45	35.5	36.4	36.2	55.9	37.7	54.1	20.0	35.6	39.6
K/Ba	2760	1129	1945	5510	5603	2224	3446	2801	2822	5561	4095
K/Cs	365	865	618	454	467	927	502	800	230	445	551
K/Rb	14	19	16.5	13	13	16	13	15	11.5	12.5	13.4

Appendix 17, continued

	PEGMATIC 5			PEGMATIC 6					
	KF 1	KF 2	AVERAGE	KF 1	KF 2	KF 3	KF 4	KF 5	AVERAGE
CaO	0.09	0.14	0.12	0.06	0.06	0.15	0.17	0.12	1.12
Na ₂ O	0.72	1.02	0.87	1.00	1.41	0.97	0.88	0.95	1.02
K ₂ O	13.9	13.9	13.9	13.5	13.5	13.5	13.8	13.9	13.6
Ba	30	30	30	50	40	40	30	30	38
Cs	400	190	295	290	160	240	400	210	260
Li	160	99	130	153	97	310	178	130	172
Rb	10400	5960	9680	9330	8020	8970	9950	9290	9110
Sr	230	185	208	195	165	210	220	220	202
Ba/Rb × 10 ⁻³	2.9	3.3	3.1	5.4	5.0	4.4	3.0	3.2	4.2
Rb/Sr	45	48	46.5	46	49	43	45	42	45.4
Rb/Cs	26	47.2	36.6	32.2	50.1	37.4	24.9	44.2	37.9
K/Ba	3846	3846	3846	2241	2901	2901	3819	3846	3101
K/Cs	255	607	447.5	386	700	467	296	549	479
K/Rb	11	13	12	12	14	12.5	11.5	12	12.4

Appendix 18. Partial Analyses and Selected Element Ratios of K-feldspar from South Zone, Fairservice Spodumene Pegmatites.

	PEGMATITE 9				PEGMATITE 10			PEGMATITE 11	
	KF 1	KF 2	KF 3	KF 4	AVERAGE	KF 1	KF 2m	AVERAGE	KF 2
CaO	0.13	0.04	0.06	0.06	0.07	0.08	0.10	0.09	0.04
Na ₂ O	0.75	1.22	0.64	0.72	0.93	1.53	1.03	1.28	1.30
K ₂ O	13.7	13.9	13.9	13.5	13.8	13.8	14.1	14.0	13.3
Ba	30	30	30	30	30	40	90	65	20
Cs	620	180	860	620	570	130	160	145	520
Li	165	56	166	565	319	5	4	4.5	7
Rb	10500	6220	10400	10300	9355	6360	7010	6685	8540
Sr	235	135	300	235	226	130	150	140	150
Ba/Rb × 10 ⁻³	2.8	4.8	2.9	2.9	3.4	6.3	12.8	9.6	2.3
Rb/Sr	45	46	35	44	42.5	49	47	48	57
Rb/Cs	16.9	34.6	12.1	16.6	20.0	48.9	43.4	46.4	16.4
K/Ba	37.0	3847	3847	3818	3876	2564	1300	2082	5520
K/Cs	183	641	134	185	286	861	731	806	212
K/Rb	11	18.5	11	11	12.9	18	17	17.5	13

Appendix 19. Partial Analyses and Selected Element Ratios of K-feldspar from East and Northeast Zones, Fairservice Spodumene Pegmatites.

	PEGMATITE 18				PEGMATITE 15	
	KF 1	KF 2	KF 3	KF 4	AVERAGE	KF 1
CaO	0.03	0.07	0.09	0.03	0.06	0.06
Na ₂ O	1.45	1.46	1.55	1.55	1.51	0.84
K ₂ O	13.3	13.5	13.3	13.5	13.4	13.9
Ba	410	470	460	330	415	40
Cs	260	200	210	230	225	560
Li	20	11	11	20	15.5	44
Rb	6460	5130	5730	6020	5635	9960
Sr	400	360	360	320	360	190
Ba, Rb x 10 ⁻³	63.5	21.6	40.3	54.5	72.6	4.0
Rb/Sr	16	14	16	19	13.5	52.5
Rb/Cs	24.5	25.7	27.3	26.2	26.0	17.5
K/Ba	26.9	235	240	340	272	2264
K/Cs	425	360	526	447	500	205
K/Rb	17	22	19	12	19.3	11.6

Appendix 20. Partial Analyses and Selected Element Ratios of K-feldspar from Magnetite-Biotite Granite Pegmatite Dykes Related to the Potassic Plutonic Suite.

Sample No.	A20-3	A20-4	A22-3	B13-2a	B19-1	C32-2	C32-6	D30-6	30-13-11	Average
Ba	140	910	3700	660	230	1160	920	980	1690	1052
Cs	6	<5	<5	<5	5	7	5	18	8	<8
Li	3	4	4	5	4	4	3	5	3	4
Rb	400	330	260	420	340	340	340	340	340	347
Sr	100	310	940	150	240	420	460	460	960	429
K ₂ O (%)	12.6	13.5	12.9	12.2	12.0	13.7	13.7	13.4	13.0	13.1
K/Ba	747	123	153	153	433	98	137	126	64	224
K/Cs $\times 10$	13.1	>22.4	>21.4	>20.2	12.45	16.2	22.4	6.18	13.5	>16.2
K/Rb	261	340	412	241	293	334	330	327	317	316
Rb/Sr	4	1.06	0.28	2.8	1.42	0.81	0.74	0.74	0.35	1.36
Ba/Rb	0.35	2.94	14.2	1.57	0.68	3.4	2.4	2.6	497	3.40

Appendix 21. Partial Analyses and Selected Element Ratios of K-feldspar from the Mica Point Pegmatitic Granite Dyke.

	KF 1	KF 2	KF 3	KF 4	KF 5	KF 6	AVERAGE
CaO	0.06	0.07	0.04	0.07	0.09	0.06	0.07
Na ₂ O	2.50	2.23	2.56	2.04	2.43	2.54	2.38
K ₂ O	12.2	12.6	12.4	13.5	12.6	12.5	12.6
Ba	70	20	60	70	210	50	92
Cs	30	90	30	120	70	40	63
Li	12	9	10	8	14	10	10.5
Rb	1090	1290	1080	1230	960	1220	1143
Sr	45	35	45	45	65	40	46
Ba/Rb × 10 ⁻³	64.2	15.6	55.6	56.9	216.8	41.0	75.4
Rb/Sr	24	37	24	27	15	31	26
Rb/Cs	36.3	14.2	36	10.3	13.7	30.5	26.5
K/Ba	1447	5229	1715	1601	498	2075	2094
K/Cs	3375	1162	3411	934	1494	2593	2165
K/Rb	93	52	95	91	109	65	92.5

Appendix 22. Partial Analyses and Selected Element Ratios of K-feldspar from the Gullwing Lake Pegmatitic Granite (North Zone).

	KF 1	KF 2	KF 3	KF 4	KF 5	KF 6	KF 7	KF 8	KF 9	KF 10	AVERAGE
Al ₂ O ₃	19.0	19.3	19.2	19.0	19.1	19.1	19.0	19.1	19.0	19.2	19.1
CaO	0.10	0.09	0.11	0.10	0.09	0.11	0.09	0.09	0.08	0.10	0.09
Na ₂ O	2.05	1.86	2.12	1.83	1.95	2.48	1.74	2.09	1.62	2.18	1.99
K ₂ O	12.5	12.7	12.4	12.9	12.6	12.0	13.0	12.5	12.7	12.5	12.6
Ba	30	20	30	20	20	70	10	10	10	40	26
Cs	160	330	410	250	320	80	430	210	570	150	293
Ga	29	26	28	25	28	28	27	28	28	27	28
Li	10	10	10	5	6	12	5	49	29	24	16
Pb	45	44	64	41	40	62	36	43	34	60	47
Rb	4240	5330	4660	4770	5490	2720	6260	4010	7680	2890	4865
Sr	60	70	70	60	70	50	90	60	110	50	69
Ba/Rbx 10 ⁻³	7.1	3.6	6.4	4.2	3.6	25.7	1.6	2.2	1.3	13.8	7.0
Rb/Sr	71	76	67	80	70	54	70	77	70	58	70
Rb/Cs	23.6	16.2	11.4	19.1	17.2	34	14.6	22	13.5	19.3	19.1
K/Ba	3458	5270	3431	5354	5229	1423	10790	10375	10541	2594	5847
K/Cs	576	319	2510	428	327	1245	251	494	185	692	703
K/Rb	24.5	20	22	22	19	37	17	22.5	14	36	23
Zone	S	S	S	S	S	K-Q	S	S	S	K-Q	

S = Spodumene - Quartz Zone
K-Q = K-feldspar - Quartz - Spodumene Zone

Appendix 23. Lateral and Vertical Variation of K_2O , Ba, Cs, Li, Rb and Sr and Selected Element Ratios in Blocky K-feldspar from the North Zone, Gullwing Lake Pegmatitic Granite.

LATERAL SECTION										
Sample No.	Distance from east contact (In feet)	K_2O	Sr	Rb	Cs	Ba	Li	Hb/Cs	K/Rb	K/Cs
	3.5'	12.7	40	2260	90	160	18	28	47	1318
	8.5'	12.6	40	2030	52	150	17	9	52	697
	20'	12.6	20	2640	98	120	18	27	40	1067
	25'	12.7	40	2670	91	110	20	31	40	1301
	30'	12.7	50	2590	97	70	17	27	41	1087
	35'	12.7	40	2460	59	150	18	42	43	1787
	40'	12.8	50	2450	60	150	17	41	43	1771
	45'	12.7	60	2360	42	200	20	56	45	2510
	50									
Average for North Zone			47	3399	167	171	17	33	38	1368
										733
										5.6
										69
VERTICAL SECTION										
Sample No.	Vertical Distance above lowest exposure (In feet)	K_2O	Sr	Rb	Cs	Ba	Li	Rb/Cs	K/Rb	K/Cs
KF-6	33.9	12.7	30	2200	46	150	22	48	48	2292
KF-7	27.9	12.8	40	2430	56	160	10	43	44	1897
KF-8	24.9	12.9	20	2570	62	120	17	41	41	171
KF-9	14.0	12.9	50	5110	320	230	10	16	21	335
KF-10	10.6	13.2	130	10590	970	450	21	11	10	113
KF-11	2.0	13.0	50	4090	140	150	12	23	26	599
KF-12	0	13.0	90	7570	320	320	30	13	14	193
										73
										61
										129
										102
										92
										82
										54

Sample No.	Vertical Distance above lowest exposure (In feet)	K_2O	Sr	Rb	Cs	Ba	Li	Rb/Cs	K/Rb	K/Cs
KF-6	33.9	12.7	30	2200	46	150	22	48	48	2292
KF-7	27.9	12.8	40	2430	56	160	10	43	44	1897
KF-8	24.9	12.9	20	2570	62	120	17	41	41	171
KF-9	14.0	12.9	50	5110	320	230	10	16	21	335
KF-10	10.6	13.2	130	10590	970	450	21	11	10	113
KF-11	2.0	13.0	50	4090	140	150	12	23	26	599
KF-12	0	13.0	90	7570	320	320	30	13	14	193
										73
										61
										129
										102
										92
										82
										54

Average for North Zone

Appendix 24. Variation of K_2O , Ba, Cs, Li, Rb, Sr and Selected Element Ratios in Blocky K-feldspars from Muscovite-K-feldspar-Quartz-Albite Core at Central Zone of Gullwing Lake Pegmatitic Granite.

Sample No.	K ₂ O	Sr	Rb	Cs	Ba	Li	Rb/Li	K/Rb	K/Cs	K/Ba	Ba/Rb $\times 10^2$
KF-1	12.5	40	2970	93	2820	29	32	34.9	1116	37	94.9
KF-2	12.5	40	3060	100	2290	28	31	33.9	1038	45	74.9
KF-3	12.5	40	3040	90	2670	17	38	38.9	1297	39	87.8
KF-4	12.6	30	3230	200	190	23	16	32.4	523	550	5.8
KF-5	13.4	90	7660	780	400	15	10	14.5	143	278	5.2
Average	12.7	48	3992	250	1674	22	25	30.9	823	190	53.7

Appendix 25. Partial Analyses and Selected Element Ratios of K-feldspar Selected at Intervals Normal to East Contact of Gullwing Lake Pegmatitic Granite (Central Zone).

	Biotite-Muscovite Pegmatitic Granite			Core Zone 9 m	Average for 2-Mica Pegmatitic Granite
	70 cm	2.4 m	4.7 m		
CaO	0.05	0.06	0.04	0.04	0.05
Na ₂ O	1.35	1.99	1.17	2.18	1.67
K ₂ O	10.0	13.1	14.4	12.8	12.6
Ba	110	160	50	20	117
Cs	40	60	180	60	93
Li	9	18	18	11	15
Rb	1720	2280	3380	2850	2460
Sr	50	70	65	50	63
Ba/Rbx 10 ⁻³	64.0	70.2	14.8	7.0	49.7
Rb/Sr	34	33	52	57	40
Rb/Cs	43	38	18.9	47.5	33
K/Ba	755	680	2390	5312	1275
K/Cs	2075	1812	664	1770	1517
K/Rb	48	48	35	37	43.7

Appendix 26. Partial Analyses and Selected Element Ratios of Blocky K-feldspar from South Zone of Gullwing Lake Pegmatitic Granite.

Sample No.	K ₂ O	Sr	Rb	Cs	Ba	Li	Rb/Cs	K/Rb	K/Cs	K/Ba	Ba/Rb $\times 10^{-3}$	Rb/Sr
KF-1	13.3	60	2420	92	230	5	26	46	1200	480	9.5	40
66'	12.8	30	2750	67	140	2	41	39	1586	759	5.1	92
70'	13.1	40	2670	81	210	1	33	41	1342	518	7.9	67
Average	13.1	43	2613	80	193	3	33	42	1376	586	7.5	66

Appendix 27. Partial Analyses and Selected Element Ratios of K-feldspar from Coates Pegmatite and Pegmatitic Dyke at Locality 30-17-1, 215 m SSW of Coates Pegmatite.

	KF1	KF2	KF3	AVERAGE	30-17-1
Al ₂ O ₃	19.0	19.0	19.0	19.0	
CaO	0.07	0.06	0.07	0.07	
Na ₂ O	1.43	1.48	1.65	1.52	
K ₂ O	13.8	13.7	13.3	13.6	13.2
Ba	90	80	70	80	160
Cs	60	50	170	93	76
Ga	28	29	45	34	
Li	24	6	6	12	4
Pb	142	147	106	132	
Rb	3280	3320	4210	3603	3680
Sr	50	50	70	56.7	40
Ea/Rb x 10 ⁻³	27.4	24.1	16.6	22.7	43.5
Rb/Sr	66	66	60	64	92
Rb/Cs	54.7	66.4	24.8	48.6	48.4
K/Ba	1273	1421	1577	1424	645
K/Cs	1909	2274	649	1611	1442
K/Rb	35	34	26	31.7	30

Appendix 28. Partial Analyses and Selected Element Ratios of K-feldspar from Tot Lake Pegmatite.

	150cm	190cm	220cm	250cm	290cm	355cm	380cm	490cm	510cm
Al ₂ O ₃	19.5	19.2	19.3	19.1	19.3	19.4	19.1	19.2	19.2
CaO	0.25	0.20	0.25	0.19	0.07	0.26	0.13	0.10	0.10
Na ₂ O	2.66	1.76	2.21	1.76	1.64	3.70	2.03	1.36	1.33
K ₂ O	11.0	12.1	11.5	12.2	12.4	9.13	12.4	12.7	12.9
Ba	40	30	30	40	50	30	50	50	50
Cs	660	1490	930	1320	1460	1420	1420	1650	1620
Ga	34	34	26	26	33	31	27	22	22
Li	29	170	31	37	40	39	36	46	50
Pb	36	65	27	66	63	50	49	48	50
Rb	6590	10700	10200	11300	11700	9560	10500	10500	11400
Sr	130	160	150	180	180	140	160	170	190
Ba/Rb $\times 10^{-3}$	4.5	2.8	2.9	3.5	4.3	3.1	4.8	4.8	4.4
Rb/Sr	66	69	66	63	65	68	66	62	60
Rb/Cs	13.5	7.2	11.0	8.6	8.9	6.7	7.4	6.4	6.3
K/Ba	2262	3346	3182	2532	2056	2526	2058	2108	2141
K/Cs	136	67	103	77	70	53	73	64	59
K/Rb	10.3	9.4	9.4	9.0	8.8	7.9	9.8	10.0	9.4

Appendix 28, continued

	KF1	KF2	KF3	KF4	KF5	KF6	KF7	KF8	KF9	KF10	AVERAGE
Al ₂ O ₃	19.5	19.2	19.3	19.2	19.1	19.2	19.1	19.2	19.1	19.2	19.2
CaO	0.17	0.04	0.05	0.05	0.06	0.05	0.09	0.09	0.06	0.08	0.07
Na ₂ O	1.37	1.18	1.39	1.83	1.51	1.57	1.64	1.61	1.63	1.49	1.52
K ₂ O	12.4	13.3	12.8	12.3	12.6	12.2	12.1	12.4	12.4	12.5	12.5
Ba	30	20	30	40	80	<10	20	10	40	30	36
CS	2410	1690	1170	740	940	1120	1220	1160	1160	660	1282
Ga	23	20	35	27	27	35	36	32	33	32	29
Li	220	50	44	26	37	18	34	36	26	27	53
Pb	85	51	57	23	69	18	56	57	53	40	51
Rb	11600	11500	11100	9940	10400	10900	11200	11100	11700	11000	10800
Sr	210	170	160	120	140	140	150	150	160	130	157
Ba/Rb x103	2.6	1.7	2.7	4.0	7.7	<0.9	1.5	0.9	3.4	2.7	3.3
Rb/Sr	55	65	69	83	74	75	75	74	73	95	70
Rb/Cs	4.5	6.5	9.5	13.4	12.4	9.7	9.2	9.6	10.1	16.7	9.4
K/Ba	3431	5520	3541	2552	1307	>10126	5022	10292	2573	3458	3697
K/Cs	43	65	91	135	125	90	52	89	99	157	95
K/Rb	5.9	9.6	9.6	10.3	10.0	9.3	9.0	9.3	5.5	9.4	9.4

Appendix 29. Analyses, Selected Element Ratios and Structural Formulae of Muscovite from Ghost Lake Batholith and Related Exocontact Dykes.

	F18-1	F26-4	E21-2	E25-7
SiO ₂	45.2	45.3	44.3	45.6
Al ₂ O ₃	36.2	35.3	36.5	36.1
Fe ₂ O ₃	1.06	1.16	0.93	1.06
FeO	0.37	0.80	0.37	0.44
MgO	0.61	0.60	0.64	0.55
CaO	0.04	0.04	0.04	0.03
Na ₂ O	0.35	0.46	0.32	0.34
K ₂ O	9.86	9.96	9.94	10.1
TiO ₂	0.05	0.04	0.14	0.22
P ₂ O ₅	0.00	0.01	0.01	0.01
MnO	0.02	0.02	0.01	0.02
CO ₂	0.20	0.20	0.32	0.20
H ₂ O ⁺	5.74	5.68	5.68	5.48
H ₂ O ⁻				0.52
TOTAL	99.7	99.6	99.2	100.7

Appendix 29, continued

	F18-1	F26-4	E21-2	E25-7
B	63	70	60	51
Ba	90	20	150	40
Be	11	17	7	6
Cs	140	260	60	80
F	1120	1970	330	440
Ga	105	145	74	130
Li	270	535	60	34
Nb	90	120	75	155
Rb	1300	2170	730	530
Sn	149	214	70.7	68.5
Sr	7	10	10	9
Ta	90	150	60	60
Zn	48	138	32	35
W	160	195	140	180
Ba/Rb $\times 10^{-3}$	69.2	9.2	205.5	75.5
Rb/Cs	9.3	8.3	12.2	6.6
Rb/Sr	185.7	217	73	58.9
K/Ba	909	4133	550	2096
K/Cs	585	318	1375	1048
K/Rb	63	38.1	113	158.2
Mg/Li	13.8	6.8	65	98.7
Nb/Ta	1.0	0.8	1.3	2.6

Appendix 29, continued

NUMBER OF IONS BASED UPON 24 (O, OH, F)

	<u>F16-1</u>		<u>F26-4</u>		<u>F21-2</u>		<u>F25-7</u>	
Si	5.925	8.000	5.961	8.000	5.861	8.000	5.952	8.000
Al	2.075		2.039		2.139		2.019	
Al	3.517	3.821	3.434	3.824	3.551	3.831	3.562	3.852
Ti	0.005		0.004		0.014		0.022	
Fe ³⁺	0.105		0.115		0.093		0.105	
Fe ²⁺	0.041		0.088		0.041		0.048	
Mg	0.119		0.116		0.126		0.108	
Mn	0.002		0.002		0.001		0.002	
Li	0.032		0.054		0.005		0.005	
Ca	0.006	1.756	0.006	1.817	0.006	1.772	0.004	1.786
Na	0.089		0.117		0.082		0.086	
K	1.649		1.672		1.677		1.690	
Cs	0.001		0.002		0.001		0.001	
Rb	0.012		0.020		0.007		0.005	
F	0.046	5.065	0.083	5.069	0.013	5.025	0.017	4.812
OH	5.020		4.986		5.013		4.795	
Fe/Fe+Mg	0.264		0.434		0.250		0.320	

Appendix 30. Partial Analyses and Selected Element Ratios of Muscovite from Ghost Lake Batholith.

	C35-3	D30-3	D34-7	E29-1	E31-1	E31-2	E31-3	E31-7
K ₂ O			10.7	10.9		9.72	10.1	9.91
Ba	40		160	40	180	<10	120	60
Be	4		7		16	23	9	8
Cs	20	10	100	15	350	340	30	190
Li	122		1120	290	775	194	950	675
Nb	80		220	150	155	130	175	135
Rb	710	420	1480	1250	1770	3180	1300	1610
Sn		6.7	566		364	684	70	150
Sr	10		14	50	23	35	16	15
Ta	<30	<30	120	11	50	70	<30	40
W	190		70		225	230	235	210
Ba/Rb x 10 ⁻³	56.3		106	32	102	<3.1	92	37
Rb/Cs	36	42	15	53	5	9	43	6.5
Rb/Sr	71		106	16	60	91	81	107
K/Ba			555	2262		>6068		1371
K/Cs			680	6031		237		433
K/Rb			60	72		25	65	51
Nb/Ta	>2.7	--	1.6	13.6	1.9	1.9	>5.8	3.4

Appendix 30, continued

	F31-5	F7-1	F7-6	F9-1	F10-1	F13-1	F16-3	F24-1	F27-4
K20	10.6	10.2	9.99		10.6	9.96	10.6		9.85
Ba	70	30	40		110	40	70	930	60
Be		7	11			23		11	7
Cs	16	80	70	30	66	690	56	30	80
Li	290	75	186			180	400	75	165
Hb	100	140	130		130	130	160	55	140
Hu	860	1060	1390	610	1060	1750	1020	620	1590
Sn	80						100	49	225
Sr		11	13			15		21	19
Ta	16	50	<30	<30	54	210	41	<30	<30
W		300	270			100		200	230
Ba/Rb $\times 10^{-3}$	61	26	29		104	23	69	1500	36
Rb/Cs	48	13	20	27	16	44	18	21	20
Rb/Sr		96	107			117		30	84
K/Ba	1257	2822	2073		800	2067	1257		1367
K/Cs	4888	1058	1185		1333	120	1571		1025
K/Rb	102	80	60		83	47	86		52
Nb/La	6.3	2.8	>4.3	--	2.4	0.6	3.9	>1.8	>4.7

Appendix 31. Partial Analyses and Selected Element Ratios of Muscovite from Pegmatite Dykes Related to Ghost Lake Batholith.

	D33-1	D34-3	F27-1	F22-6	F30-3	F22-1
K ₂ O	9.97	9.93				
Ba	20	20	110			130
Be	15	7	13	7		21
Cs	170	30	80	60	110	50
Li	18	75	880			1080
Nb	215	140	105	125		165
Rb	1370	1150	1060	1860	2450	1840
Sn	152	39	129			95
Sr	15	18		20		16
Ta	60	<30	<30	40	50	30
W	235	265	240	<50		100
Ba/Rb $\times 10^{-3}$	14.6	17.4	103.4			70.7
Rb/Cs	8.1	38	13.3	31	22.3	36.4
Rb/Sr	91	64		93		115
K/Ba	4138	4121				
K/Cs	480	27.7				
K/Rb	60	72				
Nb/Ta	3.6	>4.7	>3.5	3.2	--	5.5

Appendix 33. Analyses, Selected Element Ratios and Structural Formulae of Muscovite from Pegmatites 1 and 6, North Zone of Fairservice Spodumene Pegmatites.

	1.6m	3.7m	4.65m	5.5m	6.4m	7.3m	9.6m	10.3m	AVERAGE	PEGMATITE 6 HU 1
SiO ₂	45.0	44.4	45.2	45.0	44.7	45.0	45.1	45.2	45.0	45.0
Al ₂ O ₃	36.6	37.2	37.1	37.5	37.4	36.6	37.5	37.0	37.1	36.4
Fe ₂ O ₃	0.99	0.93	0.65	0.77	0.72	0.66	0.69	0.77	0.80	0.15
TiO ₂	0.44	0.44	0.66	0.44	0.51	0.44	0.51	0.37	0.45	0.37
MgO	0.22	0.22	0.16	0.12	0.17	0.17	0.13	0.16	0.17	0.16
CaO	0.03	0.04	0.03	0.03	0.03	0.04	0.03	0.03	0.04	0.04
Na ₂ O	0.29	0.21	0.25	0.15	0.25	0.35	0.36	0.25	0.28	0.29
K ₂ O	9.75	9.75	9.50	9.57	9.83	9.80	9.76	9.39	9.75	9.55
Li ₂ O	0.09	0.09	0.09	0.07	0.10	0.05	0.10	0.05	0.09	0.03
P ₂ O ₅	0.03	0.03	0.03	0.03	0.04	0.03	0.03	0.02	0.03	0.04
MnO	0.07	0.05	0.07	0.06	0.04	0.05	0.05	0.06	0.06	0.04
CO ₂	0.13	0.15	0.12	0.05	0.12	0.10	0.11	0.29	0.14	0.11
H ₂ O ₊	4.31	4.03	4.26	4.34	4.35	4.61	4.69	4.10	4.34	4.59
H ₂ O ₋	0.99	0.93	0.55	0.42	1.16	1.07	0.55	1.55	0.95	0.61
TOTAL	99.0	98.5	99.0	98.9	99.5	99.2	99.9	99.4	99.23	99.4

Appendix 33, continued

	1.2m	3.2m	4.55m	5.5m	6.4m	7.3m	9.6m	10.3m	AVERAGE	PEWITT 6 HUT
B	255	150	265	175	155	235	250	170	206	225
Ba	<10	11	<10	<10	<10	<10	<10	<10	<10	<10
Be	15	15	15	16	16	16	16	45	19	19
Cs	140	160	110	160	120	110	120	130	131	280
F	1930	1670	2260	1950	1950	1750	1530	1660	1795	1580
Ga	210	200	215	205	195	200	195	195	201	195
Li	660	705	535	480	580	495	425	440	540	610
Nb	95	75	110	90	95	115	105	130	109	80
Hb	7020	7940	7130	7350	6660	7120	7060	6860	7063	9620
Sn	710	670	673	734	752	639	710	722	729	734
Sr	50	55	55	60	50	55	55	65	56	60
Ti	50	20	40	50	50	40	40	80	69	90
Zn	240	210	260	240	220	210	230	230	230	240
W	140	135	145	155	150	170	150	150	149	155
Ba/Hb $\times 10^{-3}$	<1.4	1.5	<1.4	<1.4	<1.5	<1.4	<1.4	<1.4	<1.4	<1.1
Hb/Cs	50.1	45.9	64.6	45.9	55.5	64.7	58.7	52.6	54.8	33.6
Hb/Sr	140	133	120	123	133	130	126	105	128	116
K/Hb	>8117	7380	>8134	>8192	>8159	>8134	>8101	>7794	>8093	>7527
K/Cs	580	507	740	512	680	740	675	600	629	283
K/Hb	11.6	11.1	11.4	11.1	12.3	11.4	11.5	11.4	11.5	8.4
Hg/Li	2.0	1.9	1.8	1.5	1.8	2.1	1.9	2.2	1.9	1.6
Nb/Ta	1.9	0.6	2.8	1.0	1.1	2.9	2.6	2.4	1.9	0.9

Appendix 33, continued

NUMBER OF IONS ON THE BASIS OF 2% (O, OH, F)

	1.6m	3.7m	4.85m	5.5m	6.4m	7.3m	9.6m	10.3m	Pegmatite 6 MU 1
Si	6.051	6.009	6.049	6.013	5.989	6.024	5.979	6.092	5.953
Al	1.949	1.991	1.951	1.967	2.011	1.976	2.021	1.908	2.042
Al	3.850	3.942	3.899	3.918	3.895	3.798	3.837	3.968	3.949
Ti	0.009	0.009	0.009	0.007	0.010	0.008	0.010	0.008	0.003
Fe ³⁺	0.100	0.095	0.065	0.077	0.073	0.087	0.069	0.076	0.015
Fe ²⁺	0.049	0.13	0.074	0.049	0.057	0.049	0.057	0.042	0.041
Mg	0.044	0.044	0.032	0.024	0.034	0.034	0.026	0.032	0.032
Mn	0.006	0.009	0.007	0.005	0.006	0.006	0.007	0.004	0.007
Li	0.076	0.062	0.065	0.054	0.070	0.059	0.046	0.054	0.069
Ca	0.004	0.006	0.004	0.004	0.004	0.006	0.004	0.007	0.006
Na	0.076	0.055	0.073	0.039	0.065	0.099	0.098	0.073	0.074
K	1.677	1.62	1.673	1.517	1.640	1.671	1.650	1.614	1.613
Cs	0.001	0.001	0.001	0.001	0.001	0.011	0.001	0.001	0.002
Rb	0.067	0.070	0.067	0.070	0.063	0.067	0.066	0.065	0.069
F	0.051	0.041	0.097	0.065	0.063	0.076	0.063	0.060	0.067
OH	3.666	3.803	3.669	3.915	4.117	4.145	3.686	4.054	4.121
Fe/Fe+Mg	0.566	0.577	0.719	0.701	0.645	0.618	0.708	0.602	0.590

Appendix 34. Analyses, Selected Elements and Structural Formulae of Muscovite and Biotite from Gullwing-Tot Lakes Rare-Element Pegmatite Group.

GULLWING LAKE PEGMATITE											TOT LAKE PEGMATITE				
NORTH ZONE					CENTRAL ZONE					SOUTH ZONE					
MU14	MU15	MU16	MU17	MU18	MU19	MS16	MU2	MU5	BIOTITE 4472	MU 4471	MU1	MU4	MU1	MU2	
SiO2	45.2	46.7	45.3	45.5	46.3	44.4	41.1	44.1	36.8	40.6	41.5	42.6	45.4	46.9	
Al2O3	32.4	33.1	32.4	32.2	30.2	32.9	35.3	36.0	20.5	38.0	35.9	35.8	36.1	34.5	
Fe2O3	2.01	2.31	1.80	1.61	1.98	2.22	2.28	1.94	2.50	1.27	2.50	1.86	0.73	0.74	
FeO	2.04	1.53	1.83	2.12	1.97	2.04	1.65	1.10	14.1	1.23	1.10	0.89	0.37	0.32	
MgO	1.36	0.23	0.72	1.38	1.72	0.25	1.17	0.21	5.84	1.51	1.78	1.48	0.19	0.15	
CaO	0.04	0.04	0.04	0.03	0.03	0.04	0.08	0.05	0.00	0.13	0.11	0.07	0.17	0.15	
Mn2O	0.20	0.06	0.14	0.22	0.00	0.16	0.46	0.51	0.09	0.38	0.24	0.30	0.17	0.44	
K2O	10.0	9.83	10.0	10.2	10.3	9.87	8.93	9.05	9.06	8.84	9.17	9.18	9.50	9.41	
TiO2	0.19	0.08	0.15	0.18	0.21	0.11	0.24	0.12	0.96	0.22	0.24	0.27	0.02	0.02	
P2O5	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.02	
MnO	0.14	0.20	0.29	0.13	0.16	0.35	0.12	0.11	0.33	0.78	0.10	0.15	0.32	0.29	
CO2	0.13	0.17	0.28	0.18	0.12	0.21	0.19	0.24	0.17	0.23	0.19	0.14	0.20	0.30	
H2O+	3.85	4.32	4.16	4.03	4.15	4.10	3.81	4.40	4.87	3.46	2.47	2.36	4.90	3.75	
H2O-	0.75	0.98	1.04	0.73	0.49	1.20	0.66	0.39	0.59	0.23	0.41	0.42	0.46	1.35	
F	0.98	0.41	0.69	0.99	1.02	0.94	0.54	0.47	0.50	0.92	0.95	0.80	0.10	0.12	
Li2O	0.61	0.23	0.50	0.59	0.71	0.56	0.41	0.36	1.22	0.49	0.47	0.31	0.14	0.16	
Rb2O	0.55	0.71	0.59	0.44	0.47	0.75	-	0.93	0.75	0.21	1.19	-	1.19	1.69	
TOTAL	100.44	100.95	99.98	100.52	99.80	100.15	99.05	97.13	97.91	97.82	97.22	96.81	100.15	100.32	
TRACE ELEMENTS															
B	21	44	38	33	28	32	26	24	74	<5	50	66	440	39	20
Ba	20	20	20	20	20	<10	130	650	430	30	980	520	440	20	20
Be	11	14	15	7	13	13	13	9	14	7	8	12	11	-	-
Cs	270	470	330	140	250	550	-	-	540	630	125	-	-	1130	1056
F	9830	4080	6880	9930	10200	9360	5400	4700	5000	20400	9200	9500	8000	1420	-
Ga	210	190	195	170	165	200	180	170	250	96	160	170	-	265	200
Li	2840	1080	2300	2720	3280	2500	1890	1680	1260	5650	2270	2160	1450	475	560
Nb	220	155	185	220	215	150	415	425	220	1000	370	350	245	55	-
Rb	5000	6430	5360	4000	4260	6870	-	-	8500	6850	1900	-	-	10800	15400
Sn	259	171	232	370	197	132	413	475	163	85	150	150	163	224	130
Sr	9	24	25	9	7	50	30	50	75	30	35	45	50	100	-n
Ta	60	70	60	30	60	50	-	-	-	70	25	-	-	30	-
W	120	165	145	130	135	155	-	-	-	-	-	-	-	490	503
Zn	375	330	390	260	270	610	350	385	820	1760	290	320	280	155	-
Ba/Rb	0.004	0.003	0.004	0.005	0.005	<0.002	0.02	-	0.05	0.004	0.51	-	-	0.002	0.001
Rb/Cs	18.5	13.7	16.2	28.6	17.0	12.5	-	-	15.7	10.9	15.2	-	-	9.6	14.6
Rb/Sr	555	268	214	444	609	137	-	-	113	228	54	-	-	108	-
K/Ba	4150	4079	4150	4233	4275	<8192	630	114	175	2507	75	146	173	3943	3905
K/Cs	307	174	252	605	342	149	-	-	139	119	587	-	-	70	74
K/Rb	16.6	12.7	15.5	21.2	20.1	11.9	-	-	8.8	11	39	-	-	7.3	5.1
Mg/Li	2.9	1.3	1.9	3.1	3.2	0.6	4.0	4.2	1.0	6.2	4.0	4.9	6.1	2.4	1.6
Nb/Ta	3.7	2.2	3.1	7.3	3.6	3.0	-	-	-	14.3	14.8	7.8	4.9	1.8	-

Appendix 35. Variation of Ta, Nb, Rb, Cs, Ba, Li and Sn and Selected Element Ratios in Muscovite from Quartz-K-feldspar-Muscovite-Albite Core of Gullwing Lake Pegmatitic Granite (Central Zone).

Sample No.	K ₂ O	TiO ₂	Rb	Cs	Ba	Li	Ta	Nb	Sn	Rb/Cs	K/Rb	K/Cs	K/Ba	Ba/Rb $\times 10^{-3}$	Nb/Ta	Muscovite Colour
MUSC-1	10.6	0.15	5210	190	200	2300	99	430	960	27	16.9	454	440	3.8	4.3	Silver-white
MUSC-2	10.4	0.21	5530	230	200	2300	57	500	620	24	15.6	375	432	3.6	8.9	Silver-white
MUSC-3	10.4	0.16	6760	360	230	2800	82	390	480	19	30.8	240	375	3.4	4.8	Silver-white
MUSC-4	10.3	0.10	6130	550	290	1300	65	250	240	15	10.5	155	305	3.4	4.3	Silver-white, faint green hue
MUSC-5	10.2	0.10	9400	560	310	1600	62	270	190	15	10.1	151	273	3.7	4.4	Silver-white, faint green hue
Average	10.4	0.14	6806	374	244	2060	73	374	476	20	16.4	275	365	3.6	5.3	

Appendix 36. Variation of K_2O , TiO_2 , Ba, Cs, Li, Nb, Rb, Sn and Ta and Selected Element Ratios in Muscovite Samples across Gullwing Lake Pegmatitic Granite at South Zone.

	2.13m	8.1m	9.5m	13.4m	14.5m	16m	17.2m	21.9m	25.5m	30.3m	31.9m	33.7m
$K_2O\%$	10.8	10.8	10.8	10.8	10.8	10.6	10.8	10.8	10.8	10.7	10.7	10.7
$TiO_2\%$	0.20	0.20	0.20	0.20	0.20	0.19	0.18	0.19	0.20	0.26	0.24	0.19
Ba	200	210	210	220	230	200	210	220	230	240	220	220
Cs	160	330	320	120	100	110	120	120	100	880	410	110
Li	1300	2000	1800	1700	1800	2100	2100	2400	2100	2400	2100	1900
Nb	420	430	440	410	440	440	440	460	440	390	440	440
Rb	4340	4920	5370	4070	3820	4150	4320	4140	3780	5830	5670	4170
Sn	240	260	300	260	240	300	320	280	220	260	300	320
Ta	55	57	54	36	25	32	38	36	30	110	69	36
Ba/Rb	0.046	0.043	0.039	0.054	0.060	0.048	0.043	0.053	0.061	0.041	0.039	0.05
K/Ba	448	427	427	407	390	440	427	407	390	370	404	404
K/Cs	560	272	280	747	896	800	747	747	896	101	217	807
K/Rb	20.7	18.2	16.7	22	23.5	21.2	20.8	21.7	23.7	15.2	15.7	21.3
Rb/Cs	27.1	14.9	16.8	33.9	38.2	37.7	36.3	34.5	37.8	6.6	13.8	37.9
Nb/Ta	7.6	7.5	8.1	11.4	17.6	13.8	11.6	12.8	14.7	3.5	6.4	12.2

Appendix 37. Partial Analyses and Selected Element Ratios in Biotite from Ghost Lake Batholith.

	A17-1	C17-2	D29-1	E9-5	E19-8	E29-1	F7-3
Fe ₂ O ₃	11.8		12.6	10.6	9.40		
FeO	11.7	7.06	11.2	9.46	14.0	0.69	
CaO	0.11		0.05	0.03	0.03		
Na ₂ O	0.04		0.08	0.00	0.02		
K ₂ O	8.40		9.02	9.41	9.43		
TiO ₂	2.30		2.08	2.46	2.40		
Ba	760	2250	180	520	250		290
Be	2		2		1		2
Cs	50	30	90	40	60	190	270
Nb	120		175		90		460
Hf	840	560	1080	610	700	1510	1730
Sn	22.5	14.0	56.0	24.0	14.0		65.0
Sr	12		7		5		15
Ta	<30	<30	<30	<30	<30	50	<30
W	<50		<50		<50		<50
Li	700	98	1400	95	590		2050
Hf/Ta	>4	--	>5.4	--	>2.7	--	>15.3
GLB Unit	1	1	1	Exocontact dyke next to GLB1	1	4	Exocontact dyke next to GLB3

Appendix 38. Variation of Ta, Nb, Sn, Rb, Cs, Ba and Li and Selected Element Ratios in Biotite from Pegmatites of Gullwing-Tot Lakes Group.

	GULLWING PEGMATITE, SOUTH ZONE										Ba/Rb x10			Nb/Ta	
	K ₂ O	LiO ₂	Rb	Cs	Ba	Li	Ta	Nb	Sn	Rb/Cs	K/Rb	K/Cs	K/Ba		
I. GULLWING PEGMATITE, SOUTH ZONE															
B10-1	9.08	1.25	6050	540	310	2500	100	1220	120	11.2	12.5	140	243	5.1	12.2
B10-2	9.31	1.03	6390	510	310	14000	100	1050	120	7.8	12.1	95	249	4.9	10.5
B10-4	9.41	0.92	6800	610	290	3600	120	1120	140	11.2	11.5	128	279	4.1	9.3
B10-5	9.36	0.95	6660	650	310	11000	93	1050	120	10.3	11.7	120	251	4.7	11.3
4456	9.08	--	5930	520	270	6000	100	1250	80	11.4	12.7	150	279	4.6	12.5
4475	9.18	--	6540	620	290	5800	140	1170	60	10.6	11.7	123	272	4.3	8.4
II. GULLWING PEGMATITE, NORTH ZONE															
B10-6	8.79	1.12	6410	1600	310	10000	130	1010	190	4.0	11.4	46	235	4.9	7.8
B10-7	8.90	1.10	6500	960	270	10000	120	1020	200	6.7	11.4	77	274	4.2	8.5
III. RT PEGMATITE															
RT-B10	9.30	1.10	4520	600	240	16000	97	980	100	7.5	17.1	129	322	5.3	11.3
IV. MICA POINT PEGMATITE															
MP-B10	8.57	1.17	3740	690	230	--	200	1360	100	5.4	19.7	107	320	6.2	6.9

Appendix 39. Analyses of Columbite and Tantalite from the Mavis Lake Pegmatite Group.

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GEOLOGIST

ANALYSES BY C. K. ANDERSON
(1981, p. 113).

MAVIS LAKE PEGMATITE PLUMBILLS
LAST ZONE

FROM NORTH ZONE

NORTH ZONE

Pegmatite
16

Pegmatite
6

Pegmatite
1

Pegmatite Dye
No. 1

Pegmatite
16

Pegmatite
16

	58.42	98.94	58.20	100.94	96.96	100.79	102.14	101.61	100.05	100.00	98.60	97.99	59.99
Ta ₂ O ₅	39.13	12.00	44.63	69.86	54.70	73.53	45.87	57.47	77.34	77.03	23.41	73.69	77.40
HfO ₂	43.88	66.26	36.68	14.12	23.25	10.37	34.11	26.95	9.00	8.95	57.56	11.70	8.40
MnO	15.19	8.54	8.89	13.57	13.60	12.31	14.37	15.14	7.51	11.42	15.30	11.80	10.96
FeO	0.57	10.34	7.07	2.39	3.45	2.84	3.65	1.38	6.20	7.60	2.41	2.80	3.23
CaO	0.45	0.27	0.16	0.13	0.10	0.21	0.04	0.10					
TiO ₂	0.45	0.27	0.16	0.13	0.10	0.21	0.04	0.10					
SiO ₂	0.11	0.54	0.93	0.34	0.48	0.41	0.54	0.33					
MnO	0.11	0.54	0.93	0.34	0.48	0.41	0.54	0.33					
Y ₂ O ₃	0.09	0.59	0.44	0.12	0.08	0.10	0.15	0.21					
La ₂ O ₃	0.09	0.59	0.44	0.12	0.08	0.10	0.15	0.21					
Cr ₂ O ₃	0.09	0.59	0.44	0.12	0.08	0.10	0.15	0.21					

ELEMENT RATIOS BASED UPON 24 OXYGEN ATOMS

	58.42	98.94	58.20	100.94	96.96	100.79	102.14	101.61	100.05	100.00	98.60	97.99	59.99
Ta	2.53	0.76	3.56	5.80	4.47	6.28	3.72	4.44	6.84	6.76	1.52	6.36	6.84
Hf	5.52	7.13	4.39	1.98	3.16	1.47	4.11	3.47	1.48	1.32	6.20	6.36	1.24
Mn	3.42	1.80	2.02	3.56	3.46	3.78	3.25	3.65	2.08	3.12	3.08	3.16	3.00
Fe	0.13	2.06	1.59	0.62	0.87	0.75	0.81	0.32	1.58	0.72	0.48	0.72	0.88
Ca	0.03	0.05	0.05	0.04	0.03	0.03	0.01	0.03					
Ti	0.03	0.05	0.10	0.08	0.11	0.11	0.11	0.07					
Si	0.04	0.19	0.06	0.06	0.05	0.05	0.02	0.03					
Y	0.01	0.07	0.06	0.06	0.04	0.04	0.03	0.03					
La	0.01	0.07	0.06	0.06	0.04	0.04	0.03	0.03					
Cr	0.01	0.07	0.06	0.06	0.04	0.04	0.03	0.03					

Appendix 39, continued

Sample Descriptions

- 403 Manganotantalite from extensively albitized Fairservice Pegmatite No. 16. The mineral is euhedral $0.25 \times 0.5 \times 0.9$ cms. and intergrown with albite, lime-green sericite and irregular partially replaced quartz.
- 404 Manganotantalite from albitic replacement zone in Fairservice Pegmatite No. 6. This mineral is euhedral, platy, $1 \text{ mm} \times 1 \text{ cm} \times >4 \text{ mm}$ in size and is enclosed in quartz approximately 2 mm from an albite grain.
- 405 Zoned manganotantalite from spodumene zone of Fairservice Pegmatite No. 6. The mineral is euhedral, platy featured by distinct striations on (100) and to a lesser extent on (010)
- 406 Manganotantalite from an albite-rich corona surrounding spodumene situated in the quartz-rich core zone of Fairservice Pegmatite No. 1.
- 407 Manganotantalite from spodumene zone of Fairservice Pegmatite No. 6 and similar to 405.
- 408-A Mangarocolumbite, euhedral, 1×1.5 mm rectangular shaped crystal. Occurs in an ovoid 5 cm diameter albite-rich pocket in a primary beryl-tourmaline-quartz-albite assemblage. The small pocket also contains fine- to medium-grained black tourmaline, muscovite and possible chromian mica.
- 408-B Ferrocolumbite grain, $1 \times 1 \text{ mm}$, from same albite-rich pocket as in 408-A.

Appendix 40. Analyses of Columbite and Tantalite from the Western Part of the Gullwing-Tot Lakes Pegmatite Group.

Sample No.	SPODUMINE-COLUMBITE POLYDYMENITE PEGMATITE										COATES	
	GULLWING LAKE PEGMATITE										POLYDYMENITE	
	Albitization I Zone					Albitization II Zone					MICA POINT	
	391	393	395	397	398	394	396	398	399	400	30-13-1	Ghost Lake Road
Li2O	5.72	5.31	10.70	10.00	5.63	25.47	31.08	7.93	7.93	10.67	400	401
Mb2O5	73.6	73.6	69.52	67.88	72.67	54.60	50.31	69.49	69.49	68.14	44.55	44.55
MnO	5.96	6.11	10.76	7.02	7.07	7.85	7.91	4.23	4.23	5.80	3.48	3.48
FeO	12.88	13.08	9.26	12.38	11.57	10.77	10.28	16.77	16.77	13.28	12.53	12.53
CaO	0.06	0.04	0.13	0.05	0.05	0.41	0.41	0.11	0.11	0.11	0.25	0.25
Li2O2	1.14	1.18	0.71	1.39	1.13	0.60	0.77	1.18	1.18	1.33	1.40	1.40
SnO2	0.07	0.05	0.08	0.20	0.09	0.26	0.27	0.25	0.25	0.04	0.22	0.22
MgO	0.46	0.54	0.16	0.34	0.30	0.53	0.72	0.74	0.74	0.65	0.23	0.23
Y2O3	1.16	1.12	1.16	1.16	1.01	0.53	0.72	0.74	0.74	0.65	0.23	0.23
Li2O3												
Ce2O3				0.04			0.18					
TOTAL	100.05	101.03	102.48	99.52	99.52	101.08	101.32	100.59	100.59	100.06	100.09	57.55
ELEMENT RATIO BASED UPON 24 OXYGEN ATOMS												
Sample No.	391	393	395	397	398	394	396	398	399	400	401	
Li	0.35	0.33	0.66	0.64	0.20	1.78	2.11	0.50	0.50	0.68	0.82	0.82
Mb	7.57	7.90	7.17	7.12	7.64	6.10	5.68	7.23	7.23	7.18	1.14	1.14
Mn	1.44	2.47	2.08	1.36	1.39	1.64	1.67	0.82	0.82	1.15	0.20	0.20
Fe	2.43	1.17	1.77	2.40	2.25	2.23	2.15	3.23	3.23	2.59	0.71	0.71
Ca	0.02						0.11			0.03	0.02	0.02
Ti	0.19	0.2	0.12	0.25	0.18	0.11	0.14	0.50	0.50	0.23	0.07	0.07
Sn				0.02	0.01		0.01					
Mg	0.16	0.05	0.06	0.03	0.10	0.01	0.01	0.09	0.09	0.01	0.02	0.02
Y	0.10	0.03	0.14	0.04	0.13	0.07	0.07	0.09	0.09	0.08	0.031	0.031

Appendix 40, continued

SAMPLE DESCRIPTION

390. Ferrocolumbite from Mica Point pegmatitic granite. Euhedral crystal, 0.7 x 0.8 x 3.1 cm, is in contact with coarse intergrown biotite and muscovite.
391. Ferrocolumbite, North Zone of Gullwing Lake pegmatitic granite. Euhedral crystal, 0.5 x 1 x >1.1 cm, is associated with coarse albite. The ferrocolumbite is encased by a thin coating of yellow-green sericite. Slight striations on (010) parallel to the c-axis. Also some development of (021) crystal faces.
393. Manganocolumbite from bladed biotite unit of albitization I assemblage, North Zone, Gullwing Lake pegmatitic granite. Crystals are intergrown with albite, green muscovite and minor molybdenite.
394. Ferrocolumbite, from lithian muscovite-platy albite assemblage (albitization II zone), North Zone of Gullwing Lake pegmatitic granite. Euhedral crystal, 0.15 x 0.3 x 1.4 cm, occurs in platy albite which replaces part of a crystal face of light orange blocky K-feldspar. No sign of fine-grained sericitic coating which characterize columbites from the albitization I unit. Prismatic faces (110) and terminations 121 developed.
395. Manganocolumbite from columbite-muscovite-porphYROBLASTIC albite unit, North Zone, Gullwing Lake pegmatitic granite. Euhedral crystal, 0.7 x 0.8 x 1.6 cm, contacts albite and green muscovite.
396. Ferrocolumbite from muscovite-biotite pegmatitic granite located 1.1 m east of spodumene-K-feldspar-muscovite-quartz core unit and albitization I replacement zone. Euhedral crystal contacts coarse light green muscovite, quartz, K-feldspar and albite.
397. Ferrocolumbite from albitization I unit, Central Zone, Gullwing Lake Pegmatite. The largest columbite observed in Dryden Pegmatite Field. This specimen is 1.5 x 3 x >6mm, subhedral, contacts albite and quartz, and is coated by yellow-green sericite.
398. Ferrocolumbite, from albitization I unit, South Zone, Gullwing Lake Pegmatite. Euhedral crystal is 0.35 x 0.6 x 1.5 cm, contacts albite, and is coated by yellow-green sericite.
399. Samarskite, glossy black, brittle, 2.5 x 7 mm, euhedral crystal. Localized in albite replacement veins in blocky feldspar and accompanied by secondary molybdenite. Partial analysis only.
400. Ferrocolumbite, 2 x 3 cm on outcrop surface, from incipient albite replacement veins along the margins of large blocky K-feldspar crystals. Pegmatite also contains coarse muscovite and scant molybdenite, garnet and apatite.
401. Ferrocolumbite from muscovite-rich pegmatite near the southwest corner of Drope Township. Euhedral grain, 7 x 8 mm, is associated with green muscovite, quartz and albite.

Appendix 41. Analyses of Columbite and Tantalite from the Tot Lake Pegmatite.

Sample No.	Pollucite-Spodumene Zone (near Trench 2)	Columbite-Zone (near Trench 3)	Analyses of Ucauwun (1981, p.113). Locations not specifically known			
	407-A	407-B	408	5	7	8
Ta ₂ O ₅	67.11	55.46	13.10	36.20	61.51	74.61
Nb ₂ O ₅	18.64	28.04	64.58	45.51	21.59	11.50
MnO	14.86	15.3	7.65	16.73	12.89	11.19
FeO	0.45	0.51	10.70	1.65	2.02	2.70
CaO	0.07	0.02	0.03			
TiO ₂	0.16	1.20	0.24			
SnO ₂	0.08	0.26				
MgO	0.15	0.10	0.42			
Y ₂ O ₃	0.06	0.60				
TOTAL	101.05	100.96	97.32	100.09	97.98	100.0

ELEMENT RATIOS BASED UPON 6 OXYGENS

Ta	1.36	1.06	0.22	0.65	1.17	1.62
Nb	0.63	0.89	2.22	1.26	0.68	0.41
Mn	0.94	0.91	0.39	0.93	0.76	0.76
Fe	0.03	0.03	0.54	0.09	0.12	0.14
Ca	0.006	0.002	0.002			
Ti	0.01	0.06	0.01			
Sn	0.003	0.008				
Mg	0.02	0.01	0.04			
Y		.003	0.02			

Sample Description

407A-A. Part of euhedral, rectangular-shaped manganotantalite 4 x 10 mm x 10 mm from pollucite zone, at northeastern corner of Trench 2. This grain contacts a small pocket of grey cleavelandite situated between pink euhedral spodumene crystals and an anhedral altered pollucite, and anhedral grey quartz.

407-B. Smaller manganotantalite grain from above sample.

408. Columbite grain, 0.8 x 3mm on slab surface, almost enclosed by altered, coarse spodumene laths (see Plate 85). This rectangular-shaped tantalite apparently only contacts altered spodumene and a small cluster of green mica.

6 to 9. Columbite-tantalites analyzed by Ucauwun (1981, p.113). No locations or relationship with other minerals given.

Appendix 42. Some Chemical Features of the Average Metallogenetically Specialized Granite of Tischendorf (1977).

	<u>mean</u>	<u>Standard Deviation</u>
SiO	73.38	1.39
TiO ₂	0.16	0.10
Al ₂ O ₃	13.97	1.07
Fe ₂ O ₃	0.80	0.47
FeO	1.10	0.47
MnO	0.045	0.040
MgO	0.47	0.56
CaO	0.75	0.41
Na ₂ O	3.20	0.61
K	4.69	0.68

TRACE ELEMENTS

Be	13	±	6
F	3700	±	1500
Li	400	±	200
Mo	3.5	±	2
Rb	550	±	200
Sn	40	±	20
W	7	±	3

8

METRO
MULTIMEDIA

 

- Mineral Zones:** Labeled zones include 1b, 1c, 1e, 1f, 1g, 1h, 1i, 1j, 1k, 1l, 1m, 1n, 1o, 1p, 1q, 1r, 1s, 1t, 1u, 1v, 1w, 1x, 1y, 1z, 2a, 2b, 2c, 2d, 2e, 2f, 2g, 2h, 2i, 2j, 2k, 2l, 2m, 2n, 2o, 2p, 2q, 2r, 2s, 2t, 2u, 2v, 2w, 2x, 2y, 2z, 3a, 3b, 3c, 3d, 3e, 3f, 3g, 3h, 3i, 3j, 3k, 3l, 3m, 3n, 3o, 3p, 3q, 3r, 3s, 3t, 3u, 3v, 3w, 3x, 3y, 3z, 4a, 4b, 4c, 4d, 4e, 4f, 4g, 4h, 4i, 4j, 4k, 4l, 4m, 4n, 4o, 4p, 4q, 4r, 4s, 4t, 4u, 4v, 4w, 4x, 4y, 4z, 5a, 5b, 5c, 5d, 5e, 5f, 5g, 5h, 5i, 5j, 5k, 5l, 5m, 5n, 5o, 5p, 5q, 5r, 5s, 5t, 5u, 5v, 5w, 5x, 5y, 5z, 6a, 6b, 6c, 6d, 6e, 6f, 6g, 6h, 6i, 6j, 6k, 6l, 6m, 6n, 6o, 6p, 6q, 6r, 6s, 6t, 6u, 6v, 6w, 6x, 6y, 6z, 7a, 7b, 7c, 7d, 7e, 7f, 7g, 7h, 7i, 7j, 7k, 7l, 7m, 7n, 7o, 7p, 7q, 7r, 7s, 7t, 7u, 7v, 7w, 7x, 7y, 7z, 8a, 8b, 8c, 8d, 8e, 8f, 8g, 8h, 8i, 8j, 8k, 8l, 8m, 8n, 8o, 8p, 8q, 8r, 8s, 8t, 8u, 8v, 8w, 8x, 8y, 8z, 9a, 9b, 9c, 9d, 9e, 9f, 9g, 9h, 9i, 9j, 9k, 9l, 9m, 9n, 9o, 9p, 9q, 9r, 9s, 9t, 9u, 9v, 9w, 9x, 9y, 9z, 10a, 10b, 10c, 10d, 10e, 10f, 10g, 10h, 10i, 10j, 10k, 10l, 10m, 10n, 10o, 10p, 10q, 10r, 10s, 10t, 10u, 10v, 10w, 10x, 10y, 10z, 11a, 11b, 11c, 11d, 11e, 11f, 11g, 11h, 11i, 11j, 11k, 11l, 11m, 11n, 11o, 11p, 11q, 11r, 11s, 11t, 11u, 11v, 11w, 11x, 11y, 11z, 12a, 12b, 12c, 12d, 12e, 12f, 12g, 12h, 12i, 12j, 12k, 12l, 12m, 12n, 12o, 12p, 12q, 12r, 12s, 12t, 12u, 12v, 12w, 12x, 12y, 12z, 13a, 13b, 13c, 13d, 13e, 13f, 13g, 13h, 13i, 13j, 13k, 13l, 13m, 13n, 13o, 13p, 13q, 13r, 13s, 13t, 13u, 13v, 13w, 13x, 13y, 13z, 14a, 14b, 14c, 14d, 14e, 14f, 14g, 14h, 14i, 14j, 14k, 14l, 14m, 14n, 14o, 14p, 14q, 14r, 14s, 14t, 14u, 14v, 14w, 14x, 14y, 14z, 15a, 15b, 15c, 15d, 15e, 15f, 15g, 15h, 15i, 15j, 15k, 15l, 15m, 15n, 15o, 15p, 15q, 15r, 15s, 15t, 15u, 15v, 15w, 15x, 15y, 15z, 16a, 16b, 16c, 16d, 16e, 16f, 16g, 16h, 16i, 16j, 16k, 16l, 16m, 16n, 16o, 16p, 16q, 16r, 16s, 16t, 16u, 16v, 16w, 16x, 16y, 16z, 17a, 17b, 17c, 17d, 17e, 17f, 17g, 17h, 17i, 17j, 17k, 17l, 17m, 17n, 17o, 17p, 17q, 17r, 17s, 17t, 17u, 17v, 17w, 17x, 17y, 17z, 18a, 18b, 18c, 18d, 18e, 18f, 18g, 18h, 18i, 18j, 18k, 18l, 18m, 18n, 18o, 18p, 18q, 18r, 18s, 18t, 18u, 18v, 18w, 18x, 18y, 18z, 19a, 19b, 19c, 19d, 19e, 19f, 19g, 19h, 19i, 19j, 19k, 19l, 19m, 19n, 19o, 19p, 19q, 19r, 19s, 19t, 19u, 19v, 19w, 19x, 19y, 19z, 20a, 20b, 20c, 20d, 20e, 20f, 20g, 20h, 20i, 20j, 20k, 20l, 20m, 20n, 20o, 20p, 20q, 20r, 20s, 20t, 20u, 20v, 20w, 20x, 20y, 20z, 21a, 21b, 21c, 21d, 21e, 21f, 21g, 21h, 21i, 21j, 21k, 21l, 21m, 21n, 21o, 21p, 21q, 21r, 21s, 21t, 21u, 21v, 21w, 21x, 21y, 21z, 22a, 22b, 22c, 22d, 22e, 22f, 22g, 22h, 22i, 22j, 22k, 22l, 22m, 22n, 22o, 22p, 22q, 22r, 22s, 22t, 22u, 22v, 22w, 22x, 22y, 22z, 23a, 23b, 23c, 23d, 23e, 23f, 23g, 23h, 23i, 23j, 23k, 23l, 23m, 23n, 23o, 23p, 23q, 23r, 23s, 23t, 23u, 23v, 23w, 23x, 23y, 23z, 24a, 24b, 24c, 24d, 24e, 24f, 24g, 24h, 24i, 24j, 24k, 24l, 24m, 24n, 24o, 24p, 24q, 24r, 24s, 24t, 24u, 24v, 24w, 24x, 24y, 24z, 25a, 25b, 25c, 25d, 25e, 25f, 25g, 25h, 25i, 25j, 25k, 25l, 25m, 25n, 25o, 25p, 25q, 25r, 25s, 25t, 25u, 25v, 25w, 25x, 25y, 25z, 26a, 26b, 26c, 26d, 26e, 26f, 26g, 26h, 26i, 26j, 26k, 26l, 26m, 26n, 26o, 26p, 26q, 26r, 26s, 26t, 26u, 26v, 26w, 26x, 26y, 26z, 27a, 27b, 27c, 27d, 27e, 27f, 27g, 27h, 27i, 27j, 27k, 27l, 27m, 27n, 27o, 27p, 27q, 27r, 27s, 27t, 27u, 27v, 27w, 27x, 27y, 27z, 28a, 28b, 28c, 28d, 28e, 28f, 28g, 28h, 28i, 28j, 28k, 28l, 28m, 28n, 28o, 28p, 28q, 28r, 28s, 28t, 28u, 28v, 28w, 28x, 28y, 28z, 29a, 29b, 29c, 29d, 29e, 29f, 29g, 29h, 29i, 29j, 29k, 29l, 29m, 29n, 29o, 29p, 29q, 29r, 29s, 29t, 29u, 29v, 29w, 29x, 29y, 29z, 30a, 30b, 30c, 30d, 30e, 30f, 30g, 30h, 30i, 30j, 30k, 30l, 30m, 30n, 30o, 30p, 30q, 30r, 30s, 30t, 30u, 30v, 30w, 30x, 30y, 30z, 31a, 31b, 31c, 31d, 31e, 31f, 31g, 31h, 31i, 31j, 31k, 31l, 31m, 31n, 31o, 31p, 31q, 31r, 31s, 31t, 31u, 31v, 31w, 31x, 31y, 31z, 32a, 32b, 32c, 32d, 32e, 32f, 32g, 32h, 32i, 32j, 32k, 32l, 32m, 32n, 32o, 32p, 32q, 32r, 32s, 32t, 32u, 32v, 32w, 32x, 32y, 32z, 33a, 33b, 33c, 33d, 33e, 33f, 33g, 33h, 33i, 33j, 33k, 33l, 33m, 33n, 33o, 33p, 33q, 33r, 33s, 33t, 33u, 33v, 33w,

FIGURE 10

Ministry of Natural Resources
Hon. Alan W. Pope, Minister
John R. Sloan, Deputy Minister

ONTARIO GEOLOGICAL SURVEY

MAP P.2623

GEOLOGICAL SERIES-PRELIMINARY MAP

PRECAMBRIAN GEOLOGY

EAGLE RIVER- GHOST LAKE AREA

KENORA DISTRICT

Scale 1:31 600

NTS References 52 F 14 15

ODM GSC Aeromagnetic Maps 1155G 1163G

OGS Geological Compilation Map 2443

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LEGEND

PHANEROZOIC

CENOZOIC

QUATERNARY

PLEISTOCENE AND RECENT

Till, clay, sand, gravel

UNCLAYED CHERT

PRECAMBRIAN

LATE PRECAMBRIAN

MAGNETIC INTRUSIVE ROCKS

7 7 Deane

INTRODUCED CONTACT

EARLY PRECAMBRIAN

UNMETAMORPHOSSED FELIC TO INTERMEDIATE METAMORPHISM

MUTRIE TP.

SANFORD TP.

GHOST

AUBREY TP.

WABIGOON RIVER

WABIGOON RIVER

EAGLE RIVER

HIGHWAY 17

VILLAGE OF EAGLE RIVER

MINNITAKI

CREEK

AUBREY

Colour

- PRE-CAMBRIAN
ARCHEAN
PROTEROZOIC
DUBIE
UNMETAMORPHISED FELSIC TO INTERMEDIATE INTRUSIVE ROCKS
- 6 Unsubdivided
6a Biotite and magnetite biotite granite to alkali-feldspar granite massive fine to medium grained generally holocrystalline
6b Biotite and magnetite biotite granite to alkali-feldspar granite massive coarse grained to pegmatite generally holocrystalline
6c Porphyritic (K-feldspar) biotite granite to granodiorite massive to weakly foliated medium to coarse grained
- INTRUSIVE CONTACT
- METAMORPHISED FELSIC TO INTERMEDIATE INTRUSIVE ROCKS
- 5 5a Biotite and magnetite biotite hornblende to hornblende massive to weakly foliated
5b Biotite and hornblende biotite tonalite to quartz diorite foliated
5c Biotite and hornblende biotite tonalite well foliated to gneissic
5d Porphyritic (plagioclase) biotite hornblende massive
- INTRUSIVE CONTACT
- METAMORPHISED MAFIC AND ULTRAMAFIC INTRUSIVE ROCKS
- 4 4a Pentonite
4b Altered ultramafic rocks
4c Amphibolite dikes
- INTRUSIVE CONTACT
- METASEDIMENTS AND METACHRONIC
METASEDIMENTS
Metasediments, Derived Migmatite, and S-Type Granitoids
- 3 3 Unsubdivided
3a Wacke arkose
3b Siltstone argillite mudstone
3c Iron formation interbedded with wacke and rich metapelite
3d Proterozoic metasedimentary migmatite-foliated fine grained psammite and coarse grained pelitic psammite composed with up to 10% proterozoic leucosome (igneous) confined to pelitic horizons
3e Amphibolite biotite garnet rich metapelite
3f Metaleucite metasedimentary migmatite dark and pelitic psammite components with up to 50% in part in banded leucosome
3g Heterogeneous diorite coarse grained to pegmatite biotite muscovite granite and massive cordierite biotite granite leucosome associated by less than 5% metasedimentary leucosome and occasional large metapelite and amphibolite rich massive foliated
3h Homogeneous diorite coarse grained to pegmatite biotite muscovite granite and massive cordierite biotite granite leucosome associated by less than 5% metasedimentary leucosome and occasional large metapelite and amphibolite rich massive foliated
3i Homogeneous diorite coarse grained to pegmatite biotite muscovite granite and massive cordierite biotite granite leucosome associated by less than 5% metasedimentary leucosome and occasional large metapelite and amphibolite rich massive foliated

FEATURES OF RARE-METAL-BEARING PEGMATITES FROM THE

CHARACTERISTIC MINERAL ASSEMBLAGE	GEOCHEMICAL ASSOCIATION	DEGREE OF LATE STAGE REPLACEMENT ALBITIZATION
Quartz + Muscovite + Beryl + Albite + Quartz + Blochy K-feldspar (+ Emerald + Beryl)	B (Be)	Absent
Quartz + Tourmaline + Albite + Quartz + Graphite + Beryl + Blochy K-feldspar + (Columbite + Garnet + Muscovite + Green Apatite)	B Be (Sn Nb - Ta)	Inipient. Small pods of fine grained, saccharoidal apite (+ Chromium Mica + Quartz + Albite)
Quartz + Plagioclase + Biotite + Schistite + Muscovite + Beryl	W R (Li Sn)	Not Observed

TABLE 1 PETROLOGICAL AND GEOCHEMICAL FEATURES OF RARE-METAL-BEARING PEGMATITES FROM THE DRYDEN FIELD (continued)

AREA/OCCURRENCE	PEGMATITE STRUCTURE	CHARACTERISTIC PEGMATITE MINERAL ASSEMBLAGE	GEOCHEMICAL ASSOCIATION	DEGREE OF LATE STAGE REPLACEMENT ALBITIZATION
Coates Pegmatite, Webb Township	Unzoned, external potassic pegmatite	Muscovite + Albite + Quartz + Graphite K-feldspar + (Molybdenite + Pyrite + Garnet + Samarskite + Schistite + Chalcocopyrite + Beryl)	Mo-(Nb-Ta)	Inipient replacement by fine-grained albite along edges of K-feldspar. Most molybdenite and samarskite associated with albitized areas
Pegmatite 215 m Southwest of Coates Pegmatite, Webb Township	Unzoned, external potassic pegmatite	Muscovite + Albite + Quartz + Blochy K-feldspar + (Columbite-Tantalite + Garnet + Molybdenite)	Nb-Ta-(Mo)	Inipient, film-like albitization and associated columbite-tantalite in coarse blochy K-feldspar
Tot Lake Pegmatite, Archie Township	Complex external pegmatite. Contains pink type 2 spots.	Pink Spodumene + Pollucite + Quartz + Apatite + Beryl	Li-Cs Rb-Ta-Nb	Extensive albitization in muscovite + beryl + quartz + apatite + beryl + beryl

SYMBOLS

- Small bedrock outcrop
- Area of outcrop exposure
- Bedding top unknown (inclined vertical)
- Bedding top indicated by arrow (inclined vertical)
- Bedding top (arrow from gran granation) (inclined vertical, over steep)

ISOGRADS

- Isograd (muscovite + quartz + albite + beryl + samarskite + apatite + beryl)
- First Samarskite Isograd (muscovite + albite + samarskite + molybdenite)
- Garnet outcrop
- Apatite outcrop

- 3a Homogeneous diorite: coarse grained to peg-
matic: biotite, muscovite, orthoclase and quartz
abundant: biotite, granite, leucosome accompa-
nied by 5 to 30% metasedimentary paleosome in-
clusions and occasional large metachert; met-
asedimentary cuts massive to foliated.
- 3b Homogeneous diorite: coarse grained to peg-
matic: biotite, muscovite, orthoclase and quartz
abundant: biotite, granite, leucosome accompa-
nied by less than 5% metasedimentary paleosome
inclusions: massive to weakly foliated.
- 3c Homogeneous diorite: coarse grained to
pegmatic: biotite, muscovite, granite, leucosome
with less than 5% metasedimentary paleosome
inclusions: massive to weakly foliated.
- 3d Homogeneous diorite: fine to medium grained
grey muscovite, biotite, granite to granoblastic leu-
cosome with less than 5% metasedimentary pa-
leosome inclusions: massive.
- 3e Homogeneous diorite: medium grained, mus-
covite, biotite and biotite-muscovite granite leu-
cosome commonly containing khorite-muscovite
quartz metamorphic segregations in the Eagle
River Pegmatite Lake area: massive to irregular.
- 3f Homogeneous diorite: medium grained to peg-
matic: tourmaline, muscovite and tourmaline gar-
net, muscovite, granite, biotite, leucosome.
- 3g Intrusive granite: biotite.
- 3h Garnet.
- 3i Staurolite.
- 3j Andalusite.
- 3k Silimanite.
- 3l Muscovite.
- 3m Cordierite (may be partly to completely pseu-
morphed by chlorite and/or biotite).
- 3n Duranite.
- 3o Tourmaline.
- 3p Graphic quartz: tourmaline symplectite.
- 3q Plunose quartz: muscovite symplectite.
- 3r Apatite.
- 3s Metasomatic tourmaline replacement of metasedi-
ments.
- 3t Biotite.
- 3u Chlorite.
- 3v Quartz: tourmaline veins.
- 3w Garnet: biotite-quartz clots.
- 3x Cordierite: quartz clots.
- 3y Apatite.

METAVOLCANICS

Felsic to Intermediate Metavolcanics

2

- 2a Unsubdivided
2b Tuff and lapilli tuff
2c Rhyolite to dacite flows

Mafic to Intermediate Metavolcanics

1

- 1a Fine to medium grained flows
1b Medium to coarse grained flows
1c Pillow flow
1d Amphibolized mafic metavolcanics, foliated to
gneissic
1e Calcisilicate bearing mafic metavolcanics
1f Mafic breccia
1g Chlorite-carbonate mafic metavolcanics
1h Garnet bearing mafic metavolcanics

IF Iron Formation

SYMBOLS

- x Small rock outcrop
o Area of outcrop exposure
[] Bedding top (unknown or inclined vertical)
[] Bedding top indicated by arrow (inclined vertical)
[] Bedding top (known horizontal or vertical)

ISOGRADS

- Second Silimanite isograd (muscovite + quartz → alkali feldspar + silimanite + vapour or melt)
First Silimanite isograd (coexistence of muscovite + silimanite in metapelites)
Garnet in isograd
Andalusite in isograd

NOTES

- a. This is a preliminary legend which may be changed according to subsequent mapping.
b. Other units are shown in brackets and may contain varying degrees of metamorphism and may contain varying degrees of alteration.
c. Some units are shown in brackets and may contain varying degrees of alteration.
d. Most units are shown in brackets and may contain varying degrees of alteration.

MARGINAL NOTES

INTRODUCTION

The map area is bounded by latitudes 49°45'N and 49°56'N, and longitudes 92°37'W and 93°15'W, encompassing an area of 520 km². Access is readily accomplished in most parts of the area through Highways 11, 119, 594, 601 and 604 and intersecting township access roads.

PREVIOUS WORK

The early field work in this area was done by Moorhouse (1941), Satterly (1943), and Harding (1951). More recent work at reconnaissance and detailed scales have been published by Breaks, Bond, Westerman, and Hane (1976), Breaks, Bond, Hane, Westerman, and Desrochers (1976), Breaks (1980), and Page and Christie (1980).

MINERAL EXPLORATION

The first record of lithophile mineralization in the region was reported by Burwash (1939) who briefly described cassiterite associated with tourmaline pegmatite on Partridge (Portage?) Bay of Eagle Lake. Satterly (1943, 1955) briefly mentioned an occurrence of beryl in Zealand Township which is probably the Taylor Beryl Occurrence of this map. Mineral exploration for lithophile elements, however, occurred mainly during 2 periods:

1. from 1955 to 1964 for lithium and
2. in the late 1960s to present for tungsten and tantalum.

Discovery of spodumene pegmatites near Mavis Lake, Brownridge Township, in 1966, led to investigation by Lun Echo Gold Mines Limited. A program involving trenching and 40 diamond-drill holes totaling 1968 m established lithium mineralization over a strike length of 670 m and width varying between 1.5 and 11 m. The best intersection recorded 2.76% Li₂O over 2.9 m (Assessment Files Research Office, Ontario Geological Survey, Toronto (AFRO)).

During 1964, A. Kozowy and A. Leduchowski discovered a spodumene pegmatite bearing complex pegmatite near Tot Lake. J. Donner investigated the mineralization by geological mapping, a magnetometer survey, and diamond drilling (4 holes totaling 223 m) (AFRO). In 1973, A. Kozowy did a limited amount of trenching (AFRO).

In the late 1960s, tungsten mineralization was discovered by D. Petrunia near Sharpe Lake in Brownridge Township and initial evaluation was undertaken by Noranda Mines Limited (Blackburn and Halstone 1983). As of 1983, a 32 claim block is being assessed under option agreement by Sammine Exploration Incorporated who, in 1981, commenced a program of surface exploration and limited diamond drilling.

Interest in the tantalum potential of rare metal pegmatites of the Dryden area began in 1979 when Selco Mining Corporation Limited (Selco Incorporated) carried out geological mapping, diamond drilling, and a lithium lithochemical survey over the Mavis Lake spodumene pegmatite property (AFRO). In the same year, the Tantalum Mining Corporation of Canada Limited acquired property containing the Tot Lake complex rare metal pegmatite and undertook a limited diamond drill program (AFRO). In 1980, property containing the Gullwing Lake complex rare metal pegmatite was optioned from A. Kozowy by Selco and Doric Exploration Canada Limited. Both Selco-Doric options were terminated in 1982.

Tantalum Mining Corporation of Canada briefly optioned the Farservice Property in 1982 and terminated work following a magnetometer survey (AFRO).

GENERAL GEOLOGY

The Eagle River-Ghost Lake area is regionally situated along the south-west boundary of the English River Subprovince where an interdigitating interface with the Wabigoon Subprovince has been previously defined by Breaks et al. (1978). This portion of the Southern Plutonic Domain of the English River Subprovince contains an anomalously high proportion

of mafic and felsic rocks, the Zealand Sediments, and underlying Brownridge Volcanics, which are delineated by Satterly (1943).

The map area contains part of the Dryden Rare Metal Pegmatite Field. All known rare metal pegmatites are situated within metavolcanics of the Wabigoon Subprovince, very near its interface with the English River Subprovince. Geology of the area is dominated by the Ghost Lake Batholith, a 2 mica granitic complex which covers approximately 280 km². This plutonic mass extends for 40 km in an east-west direction with a breadth varying between 5 and 13 km.

Medium to high grade metawackes and mudstones of the Zealand Group envelop approximately three quarters of the Ghost Lake Batholith. These include metasedimentary migmatites which dominate the area northwest of the batholithic contact in Sandford and Elton Townships.

The remaining host lithologies consist of massive to foliated and locally gneissic biotite and magnetite biotite tonalite (Broadbent Lake Pluton) and mafic metavolcanics of the Brownridge Group. The latter, situated at the eastern end of the Ghost Lake Batholith, are highly important since rare metal pegmatites are almost exclusively contained within these metavolcanics. The youngest intrusion in the area is the Gullwing Lake Batholith which is situated along the northern periphery of the mapped area. This body consists of pink massive medium to coarse grained biotite and magnetite-biotite granite. Pegmatitic granites may also occur within this batholith either as gradational internal phases or as excontact dikes.

The overall geological setting and petrological characteristics are identical to various S-type granitoid masses which are prolifically developed within the Northern Supracrustal Domain of the English River Subprovince. These rocks are equivalent to the rhomogeneous and homogeneous diatexite-metasedimentary migmatite stages of Breaks et al. (1978, p. 18-19).

Rocks comprising the Ghost Lake Batholith and consanguineous Zealand Stock are distinguishable from all other plutonic masses in the region by virtue of their unique accessory mineral suite which includes any or all of muscovite, sillimanite (fibrolite), garnet, tourmaline, cordierite, and rare beryl and dumortierite.

Feldspar staining studies have established that homogeneous diatexite of most rocks of the Ghost Lake Batholith consist of nobleocrystic granitic or less commonly granodioritic compositions. Trondhjemite is relatively rare and confined mainly to excontact dikes emplaced in medium grade metasediments proximal to the southern batholithic contact.

Metasedimentary paleosome inclusions, mafic-rich restite, and larger metaleptitic rafts are ubiquitous, yet invariably constitute less than 5% of a given outcrop. Inhomogeneous diatexite is mainly distributed within 2 narrow, en echelon units situated within Elton and Wainwright Townships. These units appear to divide the Ghost Lake Batholith into 2 distinct portions. The area to the north contains granite typified by biotite as the chief accessory. They exhibit a paucity or absence of muscovite, and contain relatively widespread cordierite. The latter can occur as rectangular euhedral crystals up to 0.8 by 2 by 2.5 cm, variably pseudomorphosed by a dark blue-green aggregate of chlorite and muscovite. Granites in this northern area may represent a deeper structural level of the batholith given the difference in accessory mineralogy, the relatively low level of geochemical fractionation (Breaks in preparation), and gradational contact with high grade metasedimentary migmatites.

Younger intrusive granitic phases predominate within the south-central, southeastern, and eastern portions of the Ghost Lake Batholith. These phases are distinguished by noticeably higher accessory muscovite, tourmaline, and garnet contents, and commensurately lower biotite levels (Breaks in preparation) towards the eastern end of the batholith in the Ghost Lake-Dryden Airport area. These granitic rocks feature abrupt contacts with medium-grade supracrustals, very rare metasedimentary paleosome inclusions, ubiquitous potassic pegmatites, and abundant muscovite and tourmaline. These characteristics are typical of geochemically fractionated cupola zones within granitic batholiths parental to rare metal pegmatites (Varianoff 1972). This interpretation is reinforced by observation of sporadic rare metal mineralization contained within some internal potassic pegmatites; for example, at the outcrop situated near the southeastern end of the Dryden Airport main runway. Lime-green syngenetic beryl crystals measuring up to 3.0 by 8.5

cm were discovered there in 1979 by A.P. Pryskak of Selco Incorporated (A.P. Pryskak, Geologist, personal communication, 1980).

Regional Pegmatite Zonation

The regionally zoned pegmatite sequence initially delineated in the Dryden area by Breaks (1980, p. 61) has been modified to accommodate new data. As indicated in Table 1, 2 chemically and mineralogically distinct subfields have now been recognized: viz. the Mavis Lake Subfield and the Gullwing Lake-Tot Lake Subfield which is east of the map sheet. These subfields lie approximately 9.7 km apart. It remains to be determined whether these 2 areas of known rare-metal mineralization are physically connected or genetically related. This map deals mainly with the Mavis Lake Subfield. Detailed geochemical/petrological features of the Gullwing Lake-Tot Lake Subfield will be released at a later date (Breaks in preparation).

Mavis Lake Subfield

This rare-metal pegmatite subfield is confined to the northern limb of the west-plunging Thunder Lake Syncline, a macroscopic fold delineated by Satterly (1943). Dimensions of the subfield, which trends 090° to 105°, are 8 km in strike length and 0.8 to 1.4 km in width. Petrological and geochemical features of this subfield are summarized in Table 1. Noteworthy is the change in both primary mineralogical assemblages and degree of albitization in a west to east direction.

Pegmatites of the Mavis Lake Subfield are interpreted to have originated from a highly fractionated phase of the Ghost Lake Batholith, typified by outcrops at the eastern end of the Dryden Airport runway. Similar rocks occur within the Zealand Stock, which lies isolated in the core of the Thunder Lake Syncline and probably represents a phacolithic apophysis of the batholith. This consanguinity is substantiated somewhat by high (1 to 33 ppm) tin levels in both masses. Tin is also a characteristic ancillary element in most pegmatites of the Mavis Lake Subfield. A large bulk sample from spodumene pegmatite collected at the Main Zone of the Farservice Property recorded 100 ppm tin.

Most spodumene pegmatites in the Mavis Lake Subfield are confined to the Farservice Property and occur in 3 specific areas, herein termed the Main, South, and East Zones. Detailed mapping by Pryskak (1981) and the author have delineated 19 distinct pegmatite lenses, which are generally concordant to host rock foliation at surface. Contacts, however, typically undulate in vertical sections as evidenced by direct observation and by diamond drilling (A. Pryskak, Selco Incorporated, personal communication, 1981). Two pegmatite lenses in the Main Zone, for example, exhibit variable dip values of 42° to 75° N and 40° S to 80° N. Dimensions of most significant pegmatite lenses vary from 60 to 270 m in strike length and widths of 3 to 25 m.

Internal zonation within individual Mavis Lake Subfield pegmatites is only crudely developed and commonly obscured as late stage replacement by albite becomes more pervasive. The best example of internal zonation is provided by the Number 1 pegmatite from the Main Zone of the Farservice Property. This 12 by 76 m Li-Be-Ta-Nb pegmatite contains the following successively fractionated zones:

1. potassic feldspar-enriched spodumene pegmatite
2. spodumene-enriched albite quartz pegmatite
3. quartz-rich core

The quartz-rich core is discontinuously developed and typically, although not strictly, confined to the central part of the pegmatite. This zone comprises approximately 70 to 80% quartz with ancillary amounts of beryl, light green spodumene, blocky K-feldspar, green muscovite, and blue apatite. Secondary minerals related to late stage albitization include albite as coronas around spodumene (no associated eucryptite) and as discontinuous rims around blocky K-feldspar (which may contain associated tantalite), orange manganiferous garnet, lime-green sericite, and tourmaline.

The enveloping spodumene-enriched pegmatite contains the highest concentrations of spodumene observed in the Mavis Lake Subfield and also have relatively high quartz contents (Table 2). The K-feldspar-enriched spodumene pegmatite represents the outermost primary zone

redeveloped, and tends to be asymmetrically distributed in Pegmatite No. 1. This zone features markedly lower spodumene contents relative to the above mentioned unit. Petrologically, the K-feldspar-rich pegmatite is characterized by slender spodumene phenocrysts up to 1.0 cm in length and 12 to 19 cm in width containing granular quartz intergrowths, which comprise about 30% of the host rock. The spodumene quartz symplectites are typically enclosed by a thin shell of quartz and green muscovite, all of which is enveloped by a coarse K-feldspar aggregate. This potassic rich pegmatite only appears in several lenses of the Main Zone, but typically contains very little columbite-tantalite.

REPLACEMENT STAGE ALBITIZATION AND ENDOGENEOUS GEOCHEMICAL AUROLES

The degree of late stage albitization increases from west to east in the Mavis Lake Subfield such that thin (1 to 3 m wide) pegmatite sheets, exposed at the East Zone of the Farservice Property, consist essentially of secondary mineral assemblages, mainly white, sodic, beryl, green muscovite, + quartz + cleavelandite + tantalite. Primary relict spodumene, largely converted to cymalite, ragged blocky K-feldspar and quartz, are nevertheless still recognizable. Farther west at the Main Zone, albitization is much less pronounced, affecting only 20 to 50% of the pegmatites. In many places the albitization appears directly related to a tertiary mineral quartz-rich core as in Pegmatite No. 1.

Metasomatic release of Li, K, Sn, B, Rb, and F from albitized spodumene pegmatites can develop impressive endogenic dispersion patterns in surrounding mafic metavolcanics as exemplified by the Main and South Zones of the Farservice Property. These metasomatically introduced elements are responsible for the formation of secondary biotite, tourmaline, and holmquistite which are intensely developed within layers and pods in proximal mafic host rocks. Two analyses of these metasomatic assemblages (itemed glimmerite in Russian literature) from immediate contact areas around spodumene, beryl, tantalite and columbite beryl pegmatites, respectively, are presented in Table 3.

Be, Cs, F, Li, Rb, and Sn are all useful in lithochemical surveys. However, in reconnaissance surveys designed to locate rare metal pegmatite fields, lithium represents the element of greatest utility (Chernomirskiy 1976). Pryskak (1981) documented several impressive lithium auroles in the Main Zone, the largest measuring 60 to 185 m by 1270 m and containing maximum lithium concentrations of 4.095 ppm.

TUNGSTEN MINERALIZATION IN THE MAVIS LAKE SUBFIELD

Recent work by Sammine Exploration Incorporated in Brownridge and Zealand Townships has identified scheelite mineralization over a 14 km² by 488 m area (R. Redmond Consulting Geologist, personal communications, 1983). This zone, which parallels the 095° trend of the Zealand Volcanics-Zealand Sediments contact described by Satterly (1943), is situated within the beryl-bearing pegmatite zone of the Mavis Lake Subfield.

Scheelite mineralization occurs mainly in amphibolized, pillowed and mafic metavolcanics and sulphide horizons. Massive mafic flows, which occur on the property but generally do not appear to be favourable hosts for scheelite, which has so far been found in 2 distinct geological situations, namely:

1. in calcisilicate pods and layers contained within the deformation fabric of host rock, and
2. in tourmaline-rich sheets up to 0.3 m thick which variably truncate the albitation trends.

The calcisilicate domains which are composed mainly of quartz, calcite + diopside + epidote + garnet with minor zircon and apatite, contain relatively low levels of scheelite.

The highest concentrations of scheelite occur in the type 115, 116, 117 dip undulating biotite-tourmaline-rich sheets, which also contain minor quartz, plagioclase, pyrite, and rare holmquistite. This assemblage is also developed along several structurally associated ancillary fractures, with relatively steep dips (69 to 85°), and along 7 mm to 5 cm wide, low selvages, which have undergone intense replacement by tourmaline and biotite in the vicinity of the scheelite mineralization. Scheelite

GE ALBITIZATION AND GEOCHEMICAL AUREOLES

Albitization increases from west to east in the study area. In 3 m wide pegmatite sheets extending from the Fairservice Property consist essentially of quartz, mainly white, sodic, beryl + green epidote + tourmaline. Primary relict spodumene, also relict black K-feldspar and quartz are present. Further west at the Main Zone, albitization affects only 20 to 50% of the pegmatite. The albitization appears directly related to a nodular texture (Table 1).

Albitization is also found in 1 m wide albitized spodumene, epidote, and green epidote, dispersion pattern, and is exemplified by the Main Zone property. These metasomatically generated for the formation of secondary but not the albitization is intensely developed within the host rocks. Two analyses of these (metasomatically in Russian literature) from the Main Zone, beryl + tourmaline and columbite, are presented in Table 3.

Albitization in the Mavis Lake Subfield, however, is required to locate rare metal pegmatites. The element of greatest utility (Ouchinnikov, 1976) is the presence of lithium aureoles in the surrounding area in 195 m by 1270 m and concentration of 300 ppm.

ALBITIZATION IN THE MAVIS LAKE

Albitization is incorporated in Brownridge and Zealand, albitization mineralization over a 1525 m wide area. Geological personal communication (1986) reports the 1986 trend of the Brownridge contact described by Satterly (1943) as a pegmatite zone of the Mavis Lake.

Albitization in unaltered pillowed and pillow breccias. Massive mafic flows also present, do not appear to be favorable hosts for the brownridge, defined geological status.

Albitization is contained within the deformation fabric.

Albitization which variably intersected

which are composed mainly of quartz +

feldspar with minor vesuvianite and axinite

is typical.

Albitization occurs in flat lying (15 to 19

metre) sheets, which also contain minor

of the tourmaline. This assemblage is

typically associated with fracture sets

1 to 8 m wide and 2 mm to 5 cm wide

observed in the replacement by tourma-

lin. The scheelite mineralization. Scheelite

and tourmaline, beryl, crystals up to 5 by 7.5 cm, appears

along the main channel, weathered surfaces.

Analysis of a 3.5 m channel sample obtained from a flat lying biotite tourmaline sheet with an obvious high scheelite content is given below (in ppm unless otherwise stated).

W	Be	Cs	F	Li	Sn	Au
950	45	40	9600	246	19	<2 ppb

The value for tungsten is similar to those previously documented by Noranda Mines Limited, who reported channel samples of up to 0.1% WO₃ (Blackburn and Halstone, 1983, p. 12-13) and Selco Mining Corporation Limited who obtained grades of up to 0.08% WO₃ (A. Physiak, geologist, Selco Incorporated, personal communication, 1983).

The occurrence of scheelite has also been verified by this survey within the Main and East Zones of the Fairservice Property. At the Main Zone, low quantities of scheelite are disseminated within calcisilicate pods and layers situated up to 1 m from spodumene pegmatites. Occurrence of axinite in these calcisilicate domains strongly suggests that boron was derived metasomatically from nearby albitized spodumene pegmatites. Boron metasomatism is also evidenced by an increase in intensity of tourmaline replacement of mafic metavolcanics as the rare-metal pegmatites are approached. The East Zone contains sporadically distributed scheelite associated with albite-tourmaline veins connected with albitized spodumene pegmatite sheets. This establishes that tungsten mineralization is consanguineous at least with tourmaline-rich sheets associated with rare-metal pegmatites. The association of tungsten with anomalous trace levels of rare metals Li, Cs, and Be, and Sn, and F in tourmaline-rich sheets at the Petrukka Occurrence also tends to reinforce this genetic connection.

Holmquistite, which also occurs sporadically on this property, attests to the presence of Li in the tourmaline-rich fluid system which deposited the scheelite.

SUGGESTIONS FOR FUTURE EXPLORATION

The wide, but thus far sporadic, occurrence of scheelite in the Mavis Lake Subfield suggests that further exploration should be undertaken especially on the Fairservice Property and the area to the east. If continuity can be demonstrated between all known scheelite occurrences, a tungsten belt at least 6100 m in strike length will add to the economic attraction of this area.

Tourmaline appears to represent a reasonable prospecting guide because of its common association with scheelite in the study area and on a global basis. Charoy (1982) suggested that this association is due to the role of boron in the transport of tin and tungsten, although adequate experimental confirmation has not been undertaken.

Special attention should be paid to plagioclase-biotite-tourmaline sheets such as those found at the Petrukka Occurrence and East Zone of the Fairservice Property, especially where they are associated with replaced pillow selvages which suggest introduction of a fluid phase potentially containing tungsten and associated rare metals. Calcisilicate domains also widespread in the Brownridge Volcanics should be examined for scheelite.

The possibility of Ta-Nb mineralization in skarn assemblages in the Dryden area also should be investigated. Kudrn et al. (1967) described Ta-Nb-Bi mineralization in pyroxene-garnet scapolite vesuvianite skarns from Siberia which contains scheelite and molybdenite.

Cassiterite bearing float was discovered in the Earl Pit south of the Dryden Airport in the 1920s by E.G. Pidgeon (R. Pidgeon, prospector, Wabigoon, personal communication, 1981). No information exists on the tenor of the tin mineralization. It is probably significant (that) this boulder was found near the portions of Ghost Lake Batholith (and related Zealand Sluck) containing the highest tin contents. Tin concentrations there are up to 14 ppm and nearby rare-metal pegmatites contain up to 200 ppm tin. These levels are, respectively, 5 and 67 times higher than average abundance in granitic rocks (Bergner et al., 1970). It is possible that an unexposed cupola zone of the Ghost Lake Batholith exists in this part of Zealand. Potential, potentially containing low grade tin-tantalum + tungsten deposits.

In central Eton Township, granite with exceptionally high uranium content (up to 280 ppm) was encountered. This area should be further prospected.

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TABLE 1. PETROLOGICAL AND GEOCHEMICAL FEATURES OF RARE-METAL-BEARING PEGMATITES FROM DRYDEN FIELD

AREA OCCURRENCE	PEGMATITE STRUCTURE	CHARACTERISTIC PEGMATITE MINERAL ASSEMBLAGE ¹	GEOCHEMICAL ASSOCIATION ²	TEXTURAL FEATURES ³
Mavis Lake Subfield				
Dryden Airport Occurrence, Zealand Township	Unzoned, internal, usually barren potassic pegmatites	Garnet + Muscovite + Tourmaline + Albite + Quartz + Blocky K-feldspar + (Limegreen Beryl)	B (Be)	Albitized
Taylor Occurrence, Concession VII, Zealand Township	Unzoned potassic pegmatites developed marginal to Ghost Lake Batholith	Beryl + Tourmaline + Albite + Quartz + Graphitic K-feldspar + Blocky K-feldspar + Columbite + Garnet + Muscovite + Green Apatite	B (Be) (Sn) (Nb) (Ta)	Albitized, tourmaline, quartz, apatite, columbite, muscovite
Petrunka Occurrence, Brownridge and Zealand Townships	Unzoned, tourmaline-enriched fracture pillow selvage controlled replacement in Brownridge mafic metavolcanics	Quartz + Plagioclase + Biotite + Scheelite + Tourmaline + (Pyrite) + Holmquistite + Fluorite?	W-B (Li) (Sn)	Albitized
Fairservice Property, Brownridge Township				
Main and South Zones	Unzoned to crudely zoned external pegmatites. Contains randomly oriented green primary spodumene phenocrysts (Type 1 spodumene of Heinrich, 1975).	Beryl + Green Muscovite + Albite + Spodumene + Blocky K-feldspar + Quartz + (Blue Apatite) + Garnet + Tourmaline + Tantallite	Li-Be-Ta-Nb-Bi (Sn) (Rb)	Metasomatically albitized, tourmaline, quartz, apatite, columbite, muscovite
East Zone	Unzoned external pegmatites	Primary mineralogy largely replaced. Remnants of blocky K-feldspar, spodumene, and quartz still discernible.	Ta-Nb-Bi (Rb)	Metasomatically albitized, tourmaline, quartz, apatite, columbite, muscovite
Gulwing Lake-Tot Lake Subfield				
Mica Point, Gulwing Lake, Drope Township	Crudely zoned, external potassic pegmatite	Muscovite + Biotite + Albite + Quartz + K-feldspar + (Molybdenite + Columbite-Tantallite)	Nb-Ta (Ni)	Albitized
Gulwing Lake, Pegmatite, Webb Township	Complex external pegmatite. Most of dike consists of muscovite-biotite pegmatitic granite. Contains green Type 1 spodumene of Heinrich (1975).	Green Spodumene + Muscovite + Albite + Quartz + Blocky K-feldspar + Albite + Green Muscovite + Quartz + Blocky K-feldspar	Li-Ta-Nb (Ni) (Rb)	Albitized, tourmaline, quartz, apatite, columbite, muscovite

TABLE 1. PETROLOGICAL AND GEOCHEMICAL FEATURES OF RARE-METAL-BEARING PEGMATITES FROM THE DRYDEN FIELD (continued)

CHARACTERISTIC PEGMATITE MINERAL ASSEMBLAGE	GEOCHEMICAL ASSOCIATION	DEGREE OF LATE STAGE REPLACEMENT ALBITIZATION
Garnet + Muscovite + Tourmaline + Albite + Quartz + Blocky K-feldspar + (Emerald Green Beryl)	B (Be)	Absent
Beryl + Tourmaline + Albite + Quartz + Graphite + K-feldspar + Blocky K-feldspar + (Columbite + Garnet + Muscovite + Green Apatite)	B Be (Sn Nb-Ta)	Incipient. Small pods of fine-grained, saccharoidal apatite (± Chromium Mica + Quartz + Albite)
Quartz + Plagioclase + Biotite + Sphalerite + Tourmaline + (Pyrite + Fluorite + Fluorite?)	W B F (Li Sn)	Not Observed
Beryl + Green Muscovite + Albite + Spodumene + Blocky K-feldspar + Quartz + (Blue Apatite + Garnet + Tourmaline + Tantallite)	Li Be-Ta (Nb B Sn Rb)	Moderate development of albitization (± Garnet ± Tantallite + White Beryl + Green Muscovite + Quartz + Cleavelandite)
Primary mineralogy largely replaced. Remnants of blocky K-feldspar, spodumene and quartz still discernible.	Ta-Nb Be (Li)	Pervasive albitization. Most of pegmatite replaced by + White Beryl ± Tantallite + Green Muscovite + Quartz + Cleavelandite
Muscovite + Biotite + Albite + Quartz + K-feldspar + (Molybdenite + Columbite-Tantalite)	Nb-Ta? (Mo)	Incipient
Green Spodumene + Muscovite + Albite + Quartz + Blocky K-feldspar + Albite + Green Muscovite + Quartz + Blocky K-feldspar	Li-Ta-Nb (Mo Sc)	Moderate albitization developed in 2 stages: 1. early medium-grained, equigranular ± Biotite ± Tantallite + Muscovite + Quartz + Albite 2. later, fine-grained Lepidolite-Albite Complex ± Tantallite ± Beryl ± Garnet ± Lepidolite + Muscovite + Cleavelandite

TABLE 1. PETROLOGICAL AND GEOCHEMICAL FEATURES OF RARE-METAL-BEARING PEGMATITES FROM THE DRYDEN FIELD (continued)

AREA/OCCURRENCE	PEGMATITE STRUCTURE	CHARACTERISTIC PEGMATITE MINERAL ASSEMBLAGE ¹	GEOCHEMICAL ASSOCIATION ²	DEGREE OF LATE STAGE REPLACEMENT ALBITIZATION
Coates Pegmatite, Webb Township	Unzoned, external potassic pegmatite	Muscovite + Albite + Quartz + Graphite + K-feldspar + (Molybdenite + Pyrite + Garnet + Samarskite + Sphalerite + Chalcocopyrite + Beryl)	Mo-(Nb-Ta)	Incipient replacement by fine-grained albite along edges of K-feldspar. Most molybdenite and samarskite associated with albitized areas
Pegmatite 215 m Southwest of Coates Pegmatite, Webb Township	Unzoned, external potassic pegmatite	Muscovite + Albite + Quartz + Blocky K-feldspar + (Columbite-Tantalite + Garnet + Molybdenite)	Nb-Ta? (Mo)	Incipient, film-like albitization and associated columbite-tantalite in coarse blocky K-feldspar
Tot Lake Pegmatite, Webb Township	Complex external pegmatite. Contains pink Type 2 spodumene of Heinrich (1975)	Pink Spodumene + Polucite + Quartz + Albite + Blocky K-feldspar + (Fluorapatite + Garnet + Tourmaline)	Li-Cs-Rb-Ta-Nb-Be-F-P	Extensive albitization developed in 2 stages: 1. saccharoidal sodic apatite (Blue Apatite Tourmaline + Quartz + Albite) 2. Lepidolite-Albite Complex (± White Beryl ± Garnet ± Tantallite + Sencite + Lepidolite + Quartz + Cleavelandite)

¹Accessory minerals placed in brackets
²Rare elements placed in brackets

TABLE 2. MODAL ANALYSES IN VOLUME PERCENTAGE OF ZONES FROM SPODUMENE PEGMATITE NO. 1, FAIRSERVICE PROPERTY¹

ALBITE-SPODUMENE-QUARTZ PEGMATITE							
Spodumene	Quartz	K-feldspar	Albite	Green Muscovite	Sencite	Beryl	
29.4	47.3	5.0	8.8	3.8	5.0	0.8	
48.7	37.4	—	6.8	7.1	—	tr	
K-FELDSPAR-RICH SPODUMENE PEGMATITE							
6.7	23.2	54.6	12.9	2.6	—	tr	

¹Modes done over 2 by 4 m area on 20 cm grid
tr = Present in trace amounts

TABLE 3. SELECTED CHEMICAL ANALYSES OF METASOMATIC GLIMMERITE DEVELOPED AT CONTACTS OF SPODUMENE- AND BERYL-BEARING PEGMATITES FROM MAVIS LAKE SUBFIELD (in parts per million)

PEGMATITE NO. 2, FAIRSERVICE PROPERTY (tourmaline-holmquistite-biotite glimmerite)										
B	Be	Cs	F	Li	Rb	Sn	Ta	Nb	K	
450	22	920	7650	6050	3790	680	<30	13	3.69%	
PEGMATITE NO. 2, TAYLOR BERYL (biotite-rich glimmerite)										
B	Be	Cs	F	Li	Rb	Sn	Ta	Nb	K	
5	20	N.D.	7250	2300	2210	76	<30	70	4.58%	

N.D. = Not Determined

SYMBOLS

- Small bedrock outcrop
- Area of outcrop exposure
- Bedding top unknown, inclined vertical
- Bedding top indicated by arrow, inclined vertical
- Bedding top (arrow) from grain gradation, inclined vertical overturned
- Facing direction of lava flows as indicated by pillow shape and packing
- Foliation (strike only measurable dip direction known only vertical dip value measurable)
- Orientation of inclusions (inclined vertical)
- Layering in granitic rocks
- Mineral lineation (plunge known plunge value indeterminate)
- Minor S folds with plunge
- Minor Z folds with plunge
- Antiform
- Synform
- Geological boundary observed
- Geological boundary position interpreted
- Mineral occurrence

ISOGRADS

- Second Sillimanite isograd (muscovite-quartz + alkali feldspar + sillimanite + albite melt)
- First Sillimanite isograd (grossular + muscovite + sillimanite in metapelite)
- Garnet isograd
- Andalusite isograd

PROPERTIES AND OCCURRENCE

1. Taylor Beryl Occurrence (Fermontite, albitization incorporated)
2. Petronka Township Occurrence (Fermontite)
3. Main Zone Fairservice (Li Be-Ta-Nb)
4. South Zone Fairservice (Li Be-Ta-Nb)
5. East Zone Fairservice (Li Be-Ta-Nb)
6. Northeast Zone Fairservice (Li Be-Ta-Nb)

ABBREVIATIONS

- and = Andalusite
- ap = Apatite
- asp = Anorthoclase
- be = Beryl
- cb-la = Columbite-Tantalite
- cp = Chalcocopyrite
- Cu = Cuprite
- F = Fluorite
- gl = Glimmerite
- Li = Lithium
- mag = Magnetite
- Nb = Niobium
- Ni = Nickel
- po = Polucite
- py = Pyrite
- qv = Quartz
- Hb = Holmquistite
- S = Sillimanite
- Sn = Spodumene
- spd = Spinel
- Ta = Tantalum
- ta = Tantallite
- tour = Tourmaline
- U = Urtapelite
- ur = Urtapelite
- W = Wollastonite

SYMBOLS

- Taylor Beryl Occurrence (presently on property of Sanmine Exploration Incorporated)
- Petrunka Tungsten Occurrence (Sanmine Exploration Incorporated)
- Main Zone Fairservice Li-Be-Ta Property
- South Zone Fairservice Li-Be-Ta Property
- East Zone Fairservice Li-Be-Ta Property
- Northeast Zone Fairservice Li-Be-Ta Property
- Orientation of an isograd (inclined vertically)
- Faulting direction of local flows as indicated by pillow shape and orientation
- Unaltered or granitic rocks
- Mineral location (plunge known) (plunge value indicated numerically)
- Minor S. fold with plunge
- Minor N. fold with plunge
- Amphibolite
- Sillimanite
- Geological boundary observed
- Geological boundary system interpreted
- Mineral occurrence

ISOGRADES*

- Second Sillimanite isograd (muscovite + quartz → alkali feldspar + sillimanite + vapour or melt)
- First Sillimanite isograd (coexistence of muscovite + sillimanite in metapelites)
- Garnet in isograd
- Andalusite in isograd

PROPERTIES AND OCCURRENCES

- 1 Taylor Beryl Occurrence (presently on property of Sanmine Exploration Incorporated)
- 2 Petrunka Tungsten Occurrence (Sanmine Exploration Incorporated)
- 3 Main Zone Fairservice Li-Be-Ta Property
- 4 South Zone Fairservice Li-Be-Ta Property
- 5 East Zone Fairservice Li-Be-Ta Property
- 6 Northeast Zone Fairservice Li-Be-Ta Property

ABBREVIATIONS

and	Andalusite	po	Pyrrhotite
ap	Apatite	py	Pyrite
asp	Arsenopyrite	qv	Quartz vein
be	Beryl	rb	Rubidium
cb-ta	Columbite-Tantalite	s	Sulphides
cp	Chalcopyrite	sn	Tin
cu	Copper	spd	Spodumene
f	Fluorine	ta	Tantalum
gt	Garnet	ta	Tantalite
li	Lithium	tour	Tourmaline
mag	Magnetite	u	Uranium
nb	Niobium	ur	Yellow uranium staining
ni	Nickel	w	Tungsten

LOCATION MAP

Scale: 1:1 584 000

SOURCES OF INFORMATION

Base map derived from Forest Resources Inventory Maps of the Ontario Ministry of Natural Resources, Lands and Waters Group

Dryden-Vermilion Bay Sheet (P 1203) by Breaks, Bond, Westerman and Harris (1976) and Sandybeach-Route Lakes Sheet (P 1204) by Breaks, Bond, Harris, Westerman and Desnoyers (1976) Ontario Geological Survey, scale 1:63 360 or 1 inch to 1 mile

Dryden-Wabigoon Area, Ontario Department of Mines, Map 50e, by Saterly (1943), scale 1:63 360 or 1 inch to 1 mile

Eagle Lake Area, Ontario Department of Mines, Map 48d, by Moorhouse (1941), scale 1:63 360 or 1 inch to 1 mile

Assessment Files Research Office, Ontario Geological Survey, Toronto Regional and Resident Geologists' Files, Ontario Ministry of Natural Resources, Kenora and Sioux Lookout

ODM-GSC Aeromagnetic Maps 7105G, 7123G, scale 1:253 440

Geology is not tied to survey lines

Magnetic declination at the northwestern corner of the map sheet 5°05' E, 1975

Metric Conversion Factor: 1 foot = 0.3048 m

49°45'00\"/>

Mafic to Intermediate Metavolcanics

- 1a Fine to medium grained flows
- 1b Medium to coarse grained flows
- 1c Pillow flows
- 1d Amphibolized mafic metavolcanics, related to gneiss
- 1e Calcisilicate bearing mafic metavolcanics
- 1f Mafic breccia
- 1g Chlorite carbonate mafic metavolcanics
- 1h Garnet bearing mafic metavolcanics

IF Iron Formation

NOTES

- a This is basically a field legend which may be changed according to subsequent laboratory investigations
- b Often associated with rhyolite flows and may contain cumming tonite as exemplified by outcrops behind Ornden High S. fold
- c Comprise mappable units of the Ghost Lake Batholith and Zealand Block
- d Most common as excontact dikes and small masses emplaced in metapelites which envelop much of the Ghost Lake Batholith
- e Compositions of leucosome rock specimens obtained by hydrous leach and sodium cobaltinitrite method are indicated in squared brackets following the lithologic code as for example: 31(p)(GR)
- AGR - Alkali feldspar granite
- GR - Granite
- GD - Gneiss
- TR - Trondhjemite
- f Type of inclusions are given in rounded brackets following host rock code and in order of decreasing relative abundance, where applicable, for example: 16(1d, 3a)
- g Teeth on isograds point to higher grade zones

The letter 'C' preceding a code, e.g. C1a, refers to completed information

The letter 'D' placed in brackets following a mineral occurrence assembly, e.g. po py (D), refers to data compiled from diamond drill logs filed for assessment work credits. Ontario Geological Survey, Toronto Regional and Resident Geologists' Files, Ontario Ministry of Natural Resources, request freely situated at present and Sioux Lookout

CREDITS

Geology by F. W. Breaks, S. Kuehner and assistants, 1980 and 1983

Every possible effort has been made to ensure the accuracy of the information presented on this map; however, the Ontario Ministry of Natural Resources does not assume any liability for errors that may occur. Users may wish to verify critical information. Sources include both the references listed here, and information on file at the Resident or Regional Geologist's Office and the Mining Recorder's Office nearest the map area

Issued 1984

Information from this publication may be quoted if credit is given. It is recommended that reference be made in the following form:

Breaks, F. W. and Kuehner, S.
1984. Preliminary Geology of the Eagle River-Chapel Lake Area, Kenora District, Ontario Geological Survey, Map P 2623, Geological Notes Preliminary Map, scale 1:51 685 or 1 inch to 1/2 mile, 1980-1983

 

**NORTHEAST
ZONE**

**MAVIS
LAKE**

NORTH ZONE

**EAST
ZONE**

15

4

5

6

17

18

14

16

15

13

LOCATION OF FIGURE 19

PEGMATITE 2

FIGURE 17: DETAILED GEOLOGY AT SOUTH, EAST, NORTHEAST AND NORTH ZONES OF FAIRSERVICE PROPERTY LI-BE-TA PEGMATITES

AT SOUTH, EAST, NORTHEAST AND NORTH ZONES OF FAIRSERVICE PROPERTY Li-Be-Ta PEGMATITES

FIGURE 18 : DETAILED GEOLOGY OF NORTH ZONE OF FAIRSERVICE RARE-ELEMENT PEGMATITES AND L

GEOLOGY by F.W. BREAKS, 1983

4a

a. Quartz-rich core zone

4b

b. Spodumene-rich pegmatite

4c

c. K-feldspar-rich pegmatite

4d

d. Aplite

4e

e. Replaced pegmatite

4f

f. Quartz-tourmaline-albite-(beryl) veins and dykes

- a. Onyx
- b. Fine grained flows
- c. Medium to coarse grained, porphyroblastic massive flows
- d. Pillowed flows
- e. Mafic breccia
- g. Holmquistite-bearing
- h. Tourmaline-bearing
- i. Biotite-bearing
- j. Calc-silicate-bearing

3

CLASTIC METASEDIMENTS

a. Biotite wacke

OUTCROP

GEOLOGICAL CONTACT (observed, interpreted)

INTERNAL

FOLIATION (inclined, vertical dips)

MINERAL LINEATION (with plunge)

TRENCH

SURVEYED CLAIM LINES

LOCATION of PEGMATITE CHANNEL SAMPLE

LOCATION of LITHOGEOCHEMISTRY SAMPLE

LOCATION of PEGMATITE MINERAL GEOCHEMICAL SAMPLE (KF = K-feldspar, M = Muscovite) B = Beryl
CT = Columbite-Tantalite

JOINT DIP and STRIKE (inclined, vertical)

CONTACT of VEIN or DYKE (inclined, vertical dips)

DIAMOND DRILL HOLE (trend and plunge)

QUARTZ VEIN

9

RESOLUTION TEST CHART
METZ
RESOLUTION TEST CHART

FIGURE 21: GEOLOGY IN VICINITY OF GULLWING LAKE

30-17-1 IN VICINITY OF GULLWING LAKE PARE-ELEMENT PEGMATITE GRANITE DYKE

LEGEND

- 5 FELSIC TO INTERMEDIATE INTRUSIVE ROCKS
- 5a UNBAND
 - 5b DIABASE-ROSE PEGMATITE
 - 5c DIABASE-ROSE PEGMATITE
 - 5d MEDIUM-BYTE FINE GRAINED DIABASE
 - 5e MEDIUM-BYTE PEGMATITE
 - 5f ALUTE

- 4 CLASTIC METASEDIMENTS
- 4a UNBAND
 - 4b DIABASE, GARNET, MEDIUM-BYTE, MEDIUM-BYTE, MEDIUM-BYTE

- 2 INTERMEDIATE METAVOLCANICS
- 2a UNBAND
 - 2b PEGMATITE, PEGMATITE, TUFF, PEGMATITE
 - 2c TUFF, LARUL TUFF

- 1 MAFIC METAVOLCANICS
- 1a UNBAND
 - 1b TUFF TO MEDIUM GRAINED
 - 1c MEDIUM TO COARSE GRAINED
 - 1d FLOODED FLOW
 - 1e DIABASE-ROSE
 - 1f GARNET-ROSE
 - 1g GARNET-ROSE
 - 1h GARNET-ROSE
 - 1i FLOODED TO GARNET

○ OUTCROP AREA

x SMALL OUTCROP

/// GEOLOGICAL CONTACT (OBSERVED/INTERPRETED)

/// GRANITE PEGMATITE DYE (OF HORIZONTAL/VERTICAL)

/// FOLIATION STRIKE AND DIP (HORIZONTAL AND VERTICAL)

/// IMPERIAL LIGNITION WITH THERMO AND PLANE

MATERIAL OCCURRENCE

- 1a BERT
- 1b PLATINUM
- 1c DIABASE-ROSE
- 1d DIABASE-ROSE
- 1e DIABASE-ROSE
- 1f DIABASE-ROSE
- 1g DIABASE-ROSE
- 1h DIABASE-ROSE
- 1i DIABASE-ROSE
- 1j DIABASE-ROSE
- 1k DIABASE-ROSE
- 1l DIABASE-ROSE
- 1m DIABASE-ROSE
- 1n DIABASE-ROSE
- 1o DIABASE-ROSE
- 1p DIABASE-ROSE
- 1q DIABASE-ROSE
- 1r DIABASE-ROSE
- 1s DIABASE-ROSE

SCALE 1:400

GEODESY BY TALENT, AND P. HANLEY, 1981

1. Values of some pectinates have not been represented for map representation purposes. Actual width of some is placed in brackets following the map code.

GULLWING LAKE

COATES BAY STOCK

LEGEND

- 5
- FLCSC TO INTERMEDIATE INTRUSIVE ROCKS

- # 4 CLASSIC METASEDIMENTS

- 2 INTERMEDIATE METAVOLCANICS

- 1 MAFIC METAVOLCANICS

- [illegible]

COATES BAY STOCK

DROPE TWP No 2 to Mo
OCCURRENCE

DROPE TWP
WEBB TWP

FIGURE 2

A

NORTH ZONE

KF 6
KF 7
KF 120.66
KF 8
MUSC 6
MUSC 7
MUSC 8
MUSC 9, KF 9
LM-1, KFO
MUSC 10
MUSC 11
MUSC 12, KF 11
MUSC 13, KF 12, LM-2

CENTRAL ZONE

MUSC 1
MUSC 2
KF 1
KF 2
MUSC 3
MUSC 4
KF 4
KF 5
MUSC 5

FIGURE 23 : INT

ANA

23 : INTERPRETED LONGITUDINAL SECTION OF GULLWING LAKE DYKE AND LOCATION OF ANALYZED K-FELDSPAR (PLOTTED IN FIGURES 88 AND 89) AND MUSCOVITE SAMPLES.

0 Scale 1:120 50 feet
0 5 10 15 metres

Mafic Metavo

Muscovite-Bla

Blocky K-felds

Albitization Zo

MUSC 3 Location of Muscovite
KF=K feldspar,

FIGURE 30 DETAILED GEOLOGY of the TOT LAKE Li-Cs-Rb-Be-Nb-Ta RARE ELEMENT PEGMATITE DYKE

PRIMARY ASSEMBLAGES

- 1 Spodumene-Kspar-Albite-Quartz Pegmatite, randomly oriented bladed pink spodumene phenocrysts (transverse primary layering may be present)
- 2 Spodumene-Green Muscovite-Albite-Quartz Pegmatite, subhorizontally aligned pink spodumene in layers oriented parallel to dyke contacts
- 3 Spodumene-Albite-Green-Muscovite-Quartz, medium-grained, pink spodumene-poor layers grading into unit 2
- 4 Orange Kspar Zone Kspar-Quartz-Beryl-Spodumene
- 5 Pollucite Zone Pollucite-Spodumene-Kspar-Albite-Quartz-Green Muscovite

ALTERATION ASSEMBLAGES

- 6 Incipiently to Moderately Albitized Spodumene-Kspar-Albite-Quartz Pegmatite
Secondary albite-green muscovite-lepidolite-cookeite
- 7 Pervasively Albitized Spodumene-Kspar-Albite-Quartz Pegmatite
Secondary albite-green muscovite-alkali beryl
- 8 Columbite Zone Quartz-Albitized Kspar-Spodumene (albitic-green muscovitic alteration)-Alkali Beryl-Columbite-Spessartine
- 9 Saccaroidal Sodic Aplite Albite-Quartz-Tourmaline-Green Muscovite-Apatite
- 10 Holmquistite Veins
- 11 Quartz Veins Quartz-Spodumene-Kspar-Tourmaline

HOST ROCKS

- 12 Meta-ultramafic Rocks
 - 13 Fine-grained Foliated Mafic Metavolcanics
 - 14 Medium-to Coarse-grained Porphyroblastic Mafic Metavolcanics
- Outcrop Area
External, Internal Pegmatite Dyke Contacts
Strike and Dip of Host Rock Foliation
Strike and Dip of External Dyke Contacts
Mineral Lineation with Plunge
Trenched Areas

WHITE DYKE

LEGEND

5 FELSIC TO INTERMEDIATE INTRUSIVE ROCKS

5a GARNET-BIOTITE LAYERED APLITE PEGMATITE

GULLWING LAKE RARE-ELEMENT PEGMATITE
PRIMARY ZONES

BLOCK QUARTZ-K FELDSPAR ZONE

5B SPORING 4E JUNE

REPLACEMENT ZONES
ABUT ZONE UNITS

5C EX GRANULAR ALB 100

[illegible]

5F COLLEGE PARK, MARYLAND - HARRIS REPORT DATE 08-07-69
ALBUQUERQUE, NEW MEXICO

5E 1. MANUSCRIPTS - PLAYA DE ATENAS, CANTON

2 INTERMEDIATE METAVOLCANICS
2. 5-10% 100% 100% 100%

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1 MAFIC METAVOLCANICS

1. 1987年12月10日，在北京市召开的“1987年中国人口科学讨论会”上，与会代表一致通过《中国人口科学委员会章程》，并选举产生了中国人口科学委员会第一届理事会和第一届学术委员会。

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INTERNAL GEOLOGICAL CONTACT OF FOLLOWING ARE PERMANENT

FIGURE 49: PETROCHEMICAL VARIATION of B, Ba, Be, F, Ga, Li, Nb, Rb, Sn and Sr in GHOST LAKE BATHOLITH INCLUDING RELATED EXOCONTACT PEGMATITE DYKES

FIGURE 38: LOCATION MAP FOR SAMPLES AND FIGURES 15, 16 AND 17 IN THE DRYDEN AREA.

FIGURE 24 DETAILED GEOLOGY OF NORTH ZONE,
GULLWING LAKE RARE-ELEMENT PEGMATITE, WEBB
TOWNSHIP

- EXTERNAL GEOLOGICAL CONTACT OF GULLWING LAKE PEGMATITE
- INTERNAL GEOLOGICAL CONTACT OF GULLWING LAKE PEGMATITE
- FAULT
- MINERAL VEIN
- MINERALIZATION ZONE
- MINERALIZATION ZONE
- MINERALIZATION ZONE
- MINERALIZATION ZONE
- MINERALIZATION ZONE
- MINERALIZATION ZONE

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FIGURE 50: PETROCHEMICAL VARIATION OF Ba/Rb, Al/Ga, K/Ba, K/Rb, Ca/Sr, Rb/Sr, Na/K and Th/U in GHOST LAKE BATHOLITH INCLUDING RELATED EXOCONTACT PEGMATITE DYKES

FIGURE 70 : Exomorphic Chemical Aureoles (K₂O, B, Be, Cs, F, Li, Sn) Associated with Spodumene Pegmatites Nos. 1 to 5 from North Zone of Fairservice Property

FIGURE 78: VARIATION of Ba,Cs,Li,Rb,Sr,Ba/Rb,K/Ba,K/Cs,K/Rb and Rb/Sr in K-FELDSPAR from GHOST LAKE BATHOLITH and GULLWING LAKE BATHOLITH INCLUDING EXOCONTACT PEGMATITE DYKES (K-feldspars from potassic granitoid suite plutons are underlined)

Be

Cs

Li

Nb

Rb

Sn

Cs

Nb

Sn

Table 6. Mineral Assemblages of Metasediments in the Dryden- Eagle River Area.

Metamorphic Zone	Rock Type	Sample Number	Quartz	Plagioclase	K-feldspar	Chlorite	Muscovite	Biotite	Garnet	Cordierite	Staurolite	Andalusite	Sillimanite	Hornblende	Fe oxide
CHLORITE-BIOTITE	Wacke	32-9.1A	X	X	-	X	X	X	-	-	-	-	-	-	X
	Pelite	32-9.1B	X	X	-	X	X	X	-	-	-	-	-	-	X
ANDALUSITE	Wacke	F24-8A	X	X	-	X	-	-	X	-	-	-	-	-	X
		32-5.7A	X	X	-	-	X	X	-	X	-	X	-	-	X
	Pelite	32-8.14	X	X	-	-	X	X	-	-	-	X	-	-	-
		32-5.7B	X	X	-	-	X	X	-	X	-	-	-	-	-
		GB-5	X	X	-	X	X	X	-	-	-	-	-	-	-
GARNET	Wacke	F24-8B	X	X	-	-	-	X	X	-	-	-	-	-	-
		F27-4	X	X	-	-	-	X	-	-	-	-	-	-	-
	Pelite	F24-8C	X	X	-	-	X	X	-	-	-	-	-	-	-
STAUROLITE	Pelite	32-14.2	X	X	-	-	X	X	-	X	X	-	-	-	X
SILLIMANITE-MUSCOVITE	Pelite	GB-2	X	X	-	-	X	X	X	-	-	X	X	-	-
		F22-1	X	X	-	-	X	X	X	-	-	-	X	-	-
		F27-1	X	X	-	-	X	X	-	-	-	-	-	-	-
SILLIMANITE-K-FELDSPAR	Pelite	GB-1	X	X	X	-	-	X	X	X	-	-	X	-	-
		EG-6	X	X	X	-	-	X	-	-	-	-	-	-	-

in the Dryden- Eagle River Area.

Plagioclase	K-feldspar	Chlorite	Muscovite	Biotite	Garnet	Cardierite	Staurolite	Andalusite	Sillimanite	Hornblende	Fe-oxide	Cummingtonite	Epidote	Tourmaline
X	-	X	X	X	-	-	-	-	-	-	X	-	X	X
X	-	X	X	X	-	-	-	-	-	-	X	-	-	X
X	-	X	-	-	X	-	-	-	-	X	X	-	-	-
X	-	-	X	X	-	-	-	X	-	-	-	-	-	-
X	-	-	X	X	-	-	-	X	-	-	-	-	-	-
X	-	X	X	X	-	X	-	-	-	-	-	-	-	-
X	-	-	-	X	X	-	-	-	-	-	X	-	-	-
X	-	-	-	X	-	-	-	-	-	-	-	X	-	-
X	-	-	X	X	-	X	X	-	-	-	X	-	-	-
X	-	-	-	X	X	-	-	-	-	-	-	-	-	-
X	-	-	X	X	X	-	-	-	X	-	-	-	-	X
X	-	-	X	X	X	-	-	-	X	-	-	-	-	-
X	-	-	X	X	X	-	-	-	X	-	-	-	-	-
X	X	-	-	X	X	-	-	-	X	-	-	-	-	-
X	X	-	-	X	X	X	-	-	X	-	-	-	-	-

FIGURE 91: VARIATION OF Be, Cs, Li, Nb, Rb, Sn, Ta, W, K/Rb, AND Nb/Ta IN MUSCOVITE AND BIOTITE FROM GHOST LAKE BATHOLITH AND RELATED EXOCONTACT PEGMATITE DYKES (BIOTITE ANALYSES)

TABLE 13. Comparison of Regional Pegmatite Zoning in the Dryden Field to that of Intermediate Depth Zone Pegmatites in West Africa (after Varlamoff 1972).

DRYDEN PEGMATITE FIELD

| | <u>Mavis Lake Group</u>
<u>Ghost Lake Batholith</u> | <u>Gullwing-Tot Lakes</u>
<u>Group</u> |
|---|---|---|
| Type 9: Quartz veins containing cassiterite, wolframite + cassiterite, or wolframite + ferberite + scheelite | Not observed | |
| Type 8: Quartz veins with muscovite and cassiterite | Not observed | |
| Type 7-B: Quartz veins with blocky microcline, minor muscovite and cassiterite. Microcline commonly strongly albitized. | Not observed | |
| Type 7: Spodumene-Albite-Muscovite-Quartz Pegmatites. May contain minor green to white beryl, columbite-tantalite, cassiterite, and lepidolite. Strongly to completely albitized; subordinate greisenization may be present | East Zone, Fairservice Spodumene-White Beryl-Tantalite Pegmatite. Virtually complete albitization. | Tot Lake Spodumene-Pollucite-Albite-K-feldspar-Quartz Pegmatite. Zoned, approximately 60 percent albitized. Contains accessory white beryl, green muscovite, columbite-tantalite, garnet, tourmaline, and fluorapatite. |
| Type 6: Spodumene-Beryl Pegmatites. Commonly well zoned with quartz-rich core. May contain columbite-tantalite, microcline, green tourmaline, lepidolite, and coarse amblygonite. Significant albitization and/or greisenization. | North Zone, South Zone, Northeast Zone of Fairservice Spodumene-Beryl-Tantalite Pegmatites. Zoned to unzoned with quartz-rich core containing spodumene, beryl, blocky K-feldspar, muscovite. Strong albitization (50-60 percent) | Gullwing Lake Pegmatite: Spodumene-Blocky K-feldspar-Muscovite-Albite-Quartz Core Zone. Strong albitization. Accessory columbite, beryl, molybdenite, lithian muscovite, microcline |
| Type 5: Muscovite-Quartz-Microcline Pegmatites. May contain rare beryl. Commencement of albitization (local replacement) | Beryl-Muscovite-Tourmaline-Albite-K-feldspar Quartz Pegmatite (Taylor Occurrence). Local albitization (<10 percent of rock) | Coates Molybdenite-Albite-Quartz-K-feldspar Pegmatite. Locally albitized (less than 10 percent). Accessory beryl and senesite. |
| Type 4: Black Tourmaline - Muscovite Pegmatites | Tourmaline-Garnet-Muscovite-Albite-Quartz Potassic Pegmatite in GLB at Contact Beryl Occurrence and Dryden Airport Beryl Occurrence. Local albitization (less than 10 percent). Rare green beryl and columbite. | |
| Type 3: Biotite-Muscovite-Black Tourmaline Pegmatite Graphic intergrowths of feldspar-quartz | Pegmatitic Leucogranite (unit GLB-4) in central to eastern LGraphic intergrowths of feldspar-quartz parts of the Ghost Lake Batholith. | Mica Point Muscovite-Biotite-K-feldspar-Albite-Quartz Pegmatite. Local columbite and molybdenite. Negligible albitization (one percent) |
| Type 2: Biotite-Tourmaline Pegmatites. Good development of graphic intergrowths of feldspar-quartz and tourmaline-quartz. | Pegmatitic parts of GLB-1. May contain tourmaline-quartz intergrowths. | |
| Type 1: Biotite-Microcline-Plagioclase-Quartz Pegmatites | Exocontact Pegmatites in Migmatitic Metasediments west of Ghost Lake Batholith. May contain tourmaline-quartz intergrowths | |
| Type 0: Magnetite-Biotite-Microcline-Plagioclase-Quartz Pegmatites | Developed only in association with the Gullwing Lake Batholith | |

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