Uranium-lead geochronology study of Lynn Lake greenstone belt, Manitoba.

Jacek Stefan Jurkowski

*University of Windsor*

Follow this and additional works at: [https://scholar.uwindsor.ca/etd](https://scholar.uwindsor.ca/etd)

---

**Recommended Citation**


[https://scholar.uwindsor.ca/etd/4464](https://scholar.uwindsor.ca/etd/4464)
INFORMATION TO USERS

This manuscript has been reproduced from the microfilm master. UMI films the text directly from the original or copy submitted. Thus, some thesis and dissertation copies are in typewriter face, while others may be from any type of computer printer.

The quality of this reproduction is dependent upon the quality of the copy submitted. Broken or indistinct print, colored or poor quality illustrations and photographs, print bleedthrough, substandard margins, and improper alignment can adversely affect reproduction.

In the unlikely event that the author did not send UMI a complete manuscript and there are missing pages, these will be noted. Also, if unauthorized copyright material had to be removed, a note will indicate the deletion.

Oversize materials (e.g., maps, drawings, charts) are reproduced by sectioning the original, beginning at the upper left-hand corner and continuing from left to right in equal sections with small overlaps.

Photographs included in the original manuscript have been reproduced xerographically in this copy. Higher quality 6" x 9" black and white photographic prints are available for any photographs or illustrations appearing in this copy for an additional charge. Contact UMI directly to order.

Bell & Howell Information and Learning
300 North Zeeb Road, Ann Arbor, MI 48106-1346 USA
800-521-0600

UMI
NOTE TO USERS

Page(s) not included in the original manuscript are unavailable from the author or university. The manuscript was microfilmed as received.

This reproduction is the best copy available

UMI
U-Pb GEOCHRONOLOGY STUDY OF LYNN LAKE
GREENSTONE BELT, MANITOBA

by
Jacek Jurkowski

A Thesis
Submitted to the College of Graduate Studies and Research through School of Physical Sciences in Partial Fulfillment of the Requirements for the Degree of Master of Science in Geology at the University of Windsor

Windsor, Ontario, Canada
1999
© 1999 Jacek Jurkowski
The author has granted a non-exclusive licence allowing the National Library of Canada to reproduce, loan, distribute or sell copies of this thesis in microform, paper or electronic formats.

The author retains ownership of the copyright in this thesis. Neither the thesis nor substantial extracts from it may be printed or otherwise reproduced without the author’s permission.

L’auteur a accordé une licence non exclusive permettant à la Bibliothèque nationale du Canada de reproduire, prêter, distribuer ou vendre des copies de cette thèse sous la forme de microfiche/film, de reproduction sur papier ou sur format électronique.

L’auteur conserve la propriété du droit d’auteur qui protège cette thèse. Ni la thèse ni des extraits substantiels de celle-ci ne doivent être imprimés ou autrement reproduits sans son autorisation.

0-612-52583-X
NOTE TO USERS

Page(s) not included in the original manuscript are unavailable from the author or university. The manuscript was microfilmed as received.

This reproduction is the best copy available

UMI
ABSTRACT

The Lynn Lake greenstone belt in northwestern Manitoba is a Proterozoic belt and belongs to the Trans-Hudson orogen terrane of the Canadian Shield. The oldest rock units belong to the Wasekwan group, which is composed of mafic volcanic rocks, greywackes and rhyolites. Wasekwan rocks are overlain unconformably by the Sickle group made up of conglomerate, arkose and sandstone. Plutonism, gabbroic to granitic, constitutes the pre-Sickle and post-Sickle intrusives.

U-Pb zircon ages have been determined for the following pre-Sickle intrusives: the Lynn Lake gabbro 1871.5 ± 3.9 Ma, and 1870 ± 19 Ma; the Fraser Lake gabbro 1870 ± 6.2 Ma; the Eden Lake granite 1871 ± 10 Ma; the Fox Mine tonalite 1831 ± 3.7 Ma. Pb isotopic composition of feldspars from 8 units reveals genetic aspects of the rocks; both mantle and crustal signatures are identifiable. The three gabbros, stocks "A", "EL", and Fraser Lake define a Pb-Pb age of 1945 Ma which is interpreted as age of separation in the source region of the mantle, having μ values of 13.6, 13.3, and 15.7. Mantle μ values have been found in three rock units. The Pb in feldspars from two rhyolite units have an anomalous J-type composition which indicates a crustal source for this rhyolitic magma. The U-Pb age determined in this studies, together with 4 published zircon ages and 36 other isotopic ages for the Lynn Lake domain established the chronostratigraphy of the area. It is interpreted that the belt evolved between ca. 1945 to 1750 Ma in an island arc setting, but bulk of the igneous activity occurred between 1945 and 1830 Ma. The gabbros that host sulphide mineralization and barren ones are coeval.
ACKNOWLEDGMENTS

I would like to express my greatest appreciation to my advisor Dr. A. Turek, for his time and patience that helped me to finish this thesis. I also thank him for a great time during our trip to Manitoba to collect samples, and all advices he gave me during sample preparation in the laboratory at University of Windsor. I also thank Mrs. Ingrid Churchill for help during work in a laboratory with magnetic and heavy liquid mineral separation. My biggest thanks go to my wife Marta and my kids for the support that made this thesis possible.

This study was partially supported by an NSERC-Lithoprobe grant to Dr. A. Turek.
TABLE OF CONTENTS

ABSTRACT

ACKNOWLEDGMENTS

LIST OF TABLES

LIST OF FIGURES

CHAPTER

I. INTRODUCTION

II. GEOLOGY OF TRANS-HUDSON OROGEN
   Internal belts
   External belts

III. GEOLOGY OF LYNN LAKE BELT
   Supracrustal rocks
   Intrusive rocks
   Deformation and metamorphism
   Geochronology of the Lynn Lake area

IV. THE ISOTOPIC U-TH-PB METHODS OF DATING
   Basic theory
   The U-Pb concordia diagram

V. THE COMMON LEAD SYSTEM
   Basic theory
   Lead evolution models
   Dating of K-feldspar by the common lead method

VI. ANALYTICAL PROCEDURES
   Sample preparation
   Sample dissolution and ion-exchange
   Mass spectrometry
VII. RESULTS

VIII. DISCUSSION OF THE RESULTS

- U-Pb zircon ages
- Pb isotopic composition in feldspars
- Evolution of the Lynn Lake belt
- Regional geochronology

IX. SUMMARY AND CONCLUSIONS

REFERENCES

APPENDIX A: All collected samples description and location

B: Testing and pooling of ages

VITA AUCTORIS
<table>
<thead>
<tr>
<th>Table/Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Table of Precambrian formations for Lynn Lake area, Manitoba</td>
<td>14</td>
</tr>
<tr>
<td>2. Compilation of isotopic ages from Lynn Lake area</td>
<td>26</td>
</tr>
<tr>
<td>3. Numerical values of $e^{114t}$ - 1, $e^{12n}$ - 1, and the radiogenic $^{207}\text{Pb}/^{206}\text{Pb}$ ratio as a function of time (t)</td>
<td>37</td>
</tr>
<tr>
<td>4. Slopes of isochrons and corresponding model ages of Stacey and Kramers model (1975)</td>
<td>49</td>
</tr>
<tr>
<td>5. Sample description and location</td>
<td>59</td>
</tr>
<tr>
<td>6. Analytical data for zircons from Lynn Lake, Northwestern Manitoba</td>
<td>60</td>
</tr>
<tr>
<td>7. Analytical data for feldspars from Lynn Lake, Northwestern Manitoba</td>
<td>61</td>
</tr>
<tr>
<td>8. Zircon ages reported in this study and Baldwin et al. 1987</td>
<td>70</td>
</tr>
<tr>
<td>9. Chronostratigraphy of Lynn Lake area, Manitoba</td>
<td>73</td>
</tr>
<tr>
<td>Figure</td>
<td>Description</td>
</tr>
<tr>
<td>--------</td>
<td>-----------------------------------------------------------------------------</td>
</tr>
<tr>
<td>1</td>
<td>Location of Trans-Hudson orogen</td>
</tr>
<tr>
<td>2</td>
<td>Tectonic components of the Trans-Hudson orogen</td>
</tr>
<tr>
<td>3</td>
<td>Geological sketch map of the Lynn Lake area and the sample locations</td>
</tr>
<tr>
<td>4</td>
<td>Plot of radiogenic $^{207}\text{Pb}/^{206}\text{Pb}$ versus time</td>
</tr>
<tr>
<td>5</td>
<td>Concordia diagram illustrating a history of episodic Pb loss or uranium gain</td>
</tr>
<tr>
<td>6</td>
<td>Growth of lead isotope ratios with time</td>
</tr>
<tr>
<td>7</td>
<td>Graph of $^{206}\text{Pb}/^{204}\text{Pb}$ against $^{207}\text{Pb}/^{204}\text{Pb}$</td>
</tr>
<tr>
<td>8</td>
<td>Two-stage Pb evolution model of Stacey and Kramers</td>
</tr>
<tr>
<td>9</td>
<td>Concordia diagram for Eden Lake Granite (LL90-10)</td>
</tr>
<tr>
<td>10</td>
<td>Concordia diagram for Fraser Lake Gabbro (FL-11)</td>
</tr>
<tr>
<td>11</td>
<td>Concordia diagram for EL Gabbro (EL-1)</td>
</tr>
<tr>
<td>12</td>
<td>Concordia diagram for Fox Mine Tonalite (LL90-05)</td>
</tr>
<tr>
<td>13</td>
<td>Concordia diagram for EL Gabbro (LL90-03)</td>
</tr>
<tr>
<td>14</td>
<td>Concordia diagram for EL Gabbro (LL90-03 &amp; EL-1)</td>
</tr>
<tr>
<td>15</td>
<td>Pb evolution diagram of $^{206}\text{Pb}/^{204}\text{Pb}$ versus $^{207}\text{Pb}/^{204}\text{Pb}$ for eight feldspars, showing Stacey and Kramers (1975) Pb growth curve</td>
</tr>
<tr>
<td>16</td>
<td>Evolution of Lynn Lake and adjacent belts according to Gordon et al. 1990</td>
</tr>
<tr>
<td>17</td>
<td>General model of subduction zone operation</td>
</tr>
<tr>
<td>18</td>
<td>Frequency plot of isotopic ages in the Lynn Lake area</td>
</tr>
</tbody>
</table>
CHAPTER I
INTRODUCTION

The Proterozoic Lynn Lake greenstone belt is one of many in the Trans-Hudson orogen and is located east from the Manitoba and Saskatchewan border. It contains many economically important deposits of Cu, Ni, Co, Pb and Au.

Over the years there have been numerous studies of the Lynn Lake area. The definitive mapping was done in the 1960's by Milligan (1960) and later by Emslie and Moore (1961). The most recent detailed mapping of the supracrustal rocks was done by Gilbert et al. (1980). Isotopic ages for rocks in the Lynn Lake belt have been reported by Baldwin et al. (1987), Clark (1980), Moore et al. (1960), Lowdon (1961), Sangster (1978), and Turek (1967). Most of the rocks in this belt were dated by K/Ar and Rb/Sr methods and only a few by the U/Pb method. The reported ages range from 1.65 Ma to 2.0 Ma. In this thesis U-Pb zircon ages were determined for five samples from Lynn Lake area. In addition Pb isotopic composition of feldspars was measured for eight samples.

The purpose of this thesis was to measure U-Pb ages for the gabbroic and granitic plutons and thus to contribute to the chronostratigraphy of the area. Pb isotpic composition has been measured in order to obtain information on the source of magma. This thesis has nine chapters: Chapter I is introduction. Chapter II describes geology of the Trans-Hudson orogen. Most of this material comes from the papers by Paul Hoffman (Hoffman, 1981, 1987b, 1988). Presently, the Trans-Hudson orogen is topical and is focus of a Lithoprobe transect study. Chapter III presents geology and geochronology of the Lynn Lake greenstone belt. The Lynn Lake area is a base metal camp, and has been subject of many geologic investigations, but most of them are
proprietary mining company reports. The latest regional mapping is that of Gilbert (1980) which
deals only with the supracrustal rocks, as these are deemed economically significant. Chapters IV
and V describe the theoretical aspects of U-Pb and Pb-Pb methods of dating and explain the
concordia diagram and Pb growth curve. Chapter VI presents the sample preparation and
analytical procedure. It explains the statistical treatment of data. In Chapter VII the analytical
results are presented. Chapters VIII and IX are discussion, summary and conclusions. The ages
determined in this study are discussed in terms of all the isotopic ages available for the Lynn Lake
area.
CHAPTER II

GEOLOGY OF TRANS-HUDSON OROGEN

The Trans-Hudson orogen (Hoffman, 1981) is one of the best preserved and exposed Early Proterozoic orogenic belts in the world that stretches from the Arctic Islands in Canada to the subsurface of South Dakota in the USA. It is the dominant component of a more extensive orogenic system (Hoffman, 1988) that in North America includes younger Early Proterozoic belts, lying south of the Superior and Wyoming cratons (Fig. 1), obscured by Grenvillian orogen (1.0 Ga) and/or by Phanerozoic cover. They formed by Proterozoic crustal accretion and the collision of the pre-existing Archean continents. In case of Trans-Hudson orogen, the Archean continents were the Superior Province to the south and the Hearne-Rae provinces to the northwest. Available data suggest that this orogenic system evolved through a plate tectonic processes, which included, phases of initial continental rifting and passive margin development, ensuing subduction, arc generation and closure, juvenile terrane accretion, and terminal continental suturing at collisional plate boundaries.

In the Provinces of Saskatchewan and Manitoba Trans-Hudson orogen is ca. 500 km wide and is bounded by the Hearne province to the northwest and Superior province to the southeast (Fig. 1, 2). Southward, the orogen has been detected in the subsurface between the Wyoming and Superior provinces as far south as South Dakota (Green et al. 1985a, Klasner and King 1986, Thomas et al. 1987), where it is truncated by the Central Plains orogen (Sims and Peterman 1986). To the northeast, the main part of the orogen passes beneath the Paleozoic Hudson Bay basin (Gibb and Walcott 1971, Coles and Haines 1982, Gibb 1983, Sharpton et al. 1987), but its
Fig. 1. Location of Trans-Hudson orogen (from Hoffman 1988).

BH - Black Hills inlier
BL - Belcher belt
CH - Cheyenne belt
CS - Cape Smith belt
FR - Fox River belt
GL - Great Lake tectonic zone
GS - Great Slave Lake shear zone
KL - Killarney magmatic zone
KP - Kapuskasing uplift
KR - Keweenawan rift zone

LW - Lapland-White Sea tectonic zone
MK - Makkovik orogen
MO - Mistassini-Otish Basins
MRV - Minnesota River Valley terrane
SG - Saglek terrane
TH - Thompson belt
TS - Transscandinavian magmatic zone
VT - Vemery tectonic zone
WR - Winisk River fault
Fig. 2. Tectonic components of the Trans-Hudson orogen (modified after Hoffman 1988).
southeastern margin is discontinuously exposed south of Hudson Bay and along the coast and offshore islands of eastern Hudson Bay (Mukhopadhyay and Gibb 1981).

The available radiometric ages suggest that initial active-margin arc magmatism in much of the Trans-Hudson orogen occurred ca. 1.9 Ga, and that terminal collision and suturing of bounding continental platforms and intervening juvenile terranes occurred within a relatively short interval, ca. 1.85-1.80 Ga (Gordon et al. 1990). Trans-Hudson orogen comprises several internal belts of intra-oceanic rocks flanked by ensialic external belts (Stauffer 1984). Isotopic ages for the Trans-Hudson rocks in Manitoba have been compiled by Syme et al. (1993).

Internal belts

The zone of internal belts of the Trans-Hudson orogen is a complex of plutonic, metavolcanic, and metasedimentary rocks. The plutonic rocks are represented by the Baldock and the Wathaman-Chipewyan calc-alkaline batholiths of granodiorite and granite, with subordinate tonalite, diorite, and gabbro. The metavolcanic and metasedimentary rocks occur in La Ronge, Lynn Lake, Rusty Lake, Glennie Lake, Hanson Lake, Snow Lake, and Flin Flon belts of greenschist to lower amphibolite grade, and in Kinside and South Indian belts of upper-amphibolite grade.

Baldock and Wathaman-Chipewyan batholiths are located in the central part of the Trans-Hudson orogen. Both batholiths are connected in the eastern part and divided by the Rottenstone and South Indian gneissic belts in the western and central part (Fig. 2). Baldock batholith consists of granitic rocks which include two major phases and one minor phase as described by Lenton and Corkery (1981). One of the major phases is a pink, magnetite-bearing, megacrystic granite with minor amounts of hornblende and biotite. The second major phase is a pink, magnetite-bearing,
leucocratic, seriate, biotite granite. The minor phase is quartz monzonite which occurs locally along the northern margin of the Baldock batholith.

The Wathaman-Chipewyan batholith is a granitic pluton in which three major intrusive and two minor intrusive phases were identified (Lenton and Corker 1981). The predominant rock type is a magnetite-bearing, K-feldspar megacrystic granite. The second major rock type is megacrystic quartz monzonite, distinguished from the granite by its lower quartz content and higher magnetite, sphene, apatite, zircon, and fluorite contents. The contact between these two phases has not been observed. The third major phase is a leucocratic, seriate, biotite granite, which intrudes the other two major phases and is more homogenous and less foliated. Its abundance increases to the east, where it is the dominant phase at the contact with the Paleozoic cover. Two minor granitic phases are represented by rocks which vary in composition from granitic to granodioritic and were emplaced as dikes and sills into the three major phases.

The greenstone belts of greenschist to lower amphibolite grade metamorphism are located in the central (Lynn Lake, La Ronge, Rusty Lake) and southwestern (Hanson Lake, Flin Flon, Snow Lake, Glennie Lake) part of the Trans-Hudson orogen (Fig. 2). They contain structurally dismembered piles of mainly submarine, mafic and felsic metavolcanic rocks having the major- and trace element characteristics of Cenozoic island arcs (Stauffer et al. 1975, Gilbert et al. 1980, Syme 1985, 1987, Bailes and Syme 1987, Thom et al. 1987, Watters et al. 1987). They include tholeiitic basalts, high-alumina/low-titania basalts and basaltic andesites, and calc-alkaline andesite-dacite-rhyolite centered complexes. Mafic to felsic heterolithic breccias and redeposited volcanoclastic debris are common. The volcanic piles are intruded by synvolcanic gabbro, diorite, and tonalite sheets and stocks (Baldwin 1980, Barham and Froese 1986). Elsewhere, volcanic piles are intercalated with and grade laterally into volcanoclastic turbidites (Bailes 1980a).
The intercalated volcanic and sedimentary rocks and related plutons are overlain unconformably by alluvial conglomerate and sandstone (Milligan 1960, Mukherjee 1974).

Belts of upper-amphibolite grade include the Kisseynew gneiss belt located in the southeastern part of the Trans-Hudson orogen and the Rottenstone-South Indian belts located in the western and central parts of the belt respectively (Fig. 2). The transition into the Kisseynew gneiss belt from the bounding Flin Flon and Lynn Lake belts represent change from volcanic rocks to turbidites (Zwanzig 1976, Bailes 1980a, 1980b, Ashton and Wheatly 1986), and increase in a metamorphic grade from chlorite-biotite-muscovite schist to K-feldspar-cordierite-garnet paragneiss (Bailes and McRitchie 1978, Bailes 1980b, Jackson and Gordon 1985, 1986).

The Kisseynew belt contains varied plutonic rocks, and five supracrustal suites derived by high-grade metamorphism from adjacent lower grade rocks. The plutonic component comprises large intrusive bodies ranging from diorite to leucogranite, and abundant smaller masses of leucotonalite to leucogranite, probably representing anatetic melt fractions (Zwanzig 1990). The supracrustal components include the following amphibolites and felsic gneisses: quartz-rich Sherridan gneisses, graphite-bearing garnet-biotite gneisses, migmatites and amphibolites, all part of the Burntwood Suite; quartzofeldspathic gneisses of the Sickle Suite and; quartz-rich sedimentary gneiss and metavolcanic rocks of the Missi Suite.

Rottenstone and South Indian gneiss belts (Fig. 2) are lithologically similar to the Kisseynew belt. They were developed from a stratigraphic sequence comprising greywacke, volcanic and sedimentary rocks, overlain by arkosic rocks (Hoffman 1988).
External belts

The external belts of the Superior margin are exposed in four segments which include Belcher belt, Cape Smith belt, Fox River belt and Thompson belt (Figs. 1 and 2). The northwest marginal zone of the orogen includes the Wollaston and Seal River fold belts. The southern margin includes the Black Hills uplift in South Dakota.

The most complete stratigraphic sequence occurs in the Belcher belt in eastern Hudson Bay (Fig. 1). It is represented by east-trending rift-valley prism, 70 km wide, overlain by postrift shelf strata that onlap the basement north and south of the rift (Chandler 1984). The fold belt exposes at least 3.4 km of shelf carbonates and clastics containing a unit of plateau basalt up to 0.9 km thick (Ricketts and Donaldson 1981). The shelf strata are overlain paraconformably by foredeep deposits (Hoffman 1987b), comprising a transgressive sequence of quartz-arenite, ironstone, tholeiitic-komatiitic volcanics, and euxinic shale, followed by a regressive sequence of greywacke turbidites and fluvial arkose (Ricketts and Donaldson 1981, Baragar 1984).

The Cape Smith belt (Fig. 1) is an erosional remnant of a thin-skinned, south-vergent thrust-fold belt preserved as a doubly plunging synclinorium (St-Onge et al. 1986). The volcanic and sedimentary rocks record the evolution of the rifted north margin of the Superior province from continental, through transitional oceanic, to true oceanic crust (St-Onge et al. 1988). The continental rift sequence is represented by basal imbricates of semipelite, ironstone, and proximal to distal arkosic submarine-fan deposits, which pass stratigraphically upward into light-rare-earth-element-rich basalt, minor rhyolite, volcaniclastic deposits, and fault-scarp breccias (Hynes and Francis 1982). The transitional crust is dominated by pillowed basalt distinguished by a komatiitic to midoceanic-ridge-basalt-(MORB)-like of tholeiitic chemical composition (Francis et al. 1983). Both the continental and transitional oceanic crust sequences are intruded by layered peridotite-
gabbro sills. Structurally overlying the Chukotat Group is a metamorphosed ophiolite suite exposed in inverse stratigraphic order (St-Onge et al. 1988). From structural bottom to top, the ophiolite consists of thrust sheets composed of:

a. laminated graphitic pelite and semipelite;

b. pillowed basalt cut by mafic dikes, "sheeted" dikes, and gabbro cut by dikes;

c. layered mafic-ultramafic cumulates.

The east-trending Fox River belt (Fig. 2) and Sutton inlier south of Hudson Bay are poorly exposed but include steeply north-dipping thrust stacks of relatively thin, basal, shelf sediments and overlying foredeep flysch, the latter associated with differentiated tholeiitic-komatiitic sills and flows (Bostock 1971, Scoates 1981, Baragar and Scoates 1981). The 2-km-thick Fox River sill and the extensive northeast-trending Molson dike swarm of the northwestern Superior province are coeval at 1883 Ma (Scoates and Macek 1978, Heaman et al. 1986), but the tectonic significance of the magmatism is unclear. The north margin of the Fox River belt is a probable north-dipping thrust separating subgreenschist-facies rocks to the south from amphibolite-facies metasedimentary paragneisses similar to those of the Kisseynew belt (Fig. 2).

In the northeast-trending Thompson belt, equivalents of the Fox River belt strata are tightly infolded with Archean basement (Weber and Scoates 1978, Peredery et al. 1982). Foliations in the basement and cover dip very steeply to the southeast, and stretching lineations in mylonite zones are generally subvertical throughout the Thompson belt (Fueten et al. 1986). Kinematic indicators in the western part of the belt show that the Superior province has moved up relative to Trans-Hudson orogen. The bulk of the deformation postdates the 1.88 Ga Molson dikes and predates 1.79 - 1.77 Ga pegmatites. Postmetamorphic cooling is recorded by a concordant titanite age of 1.72 Ga (Hofmann 1988).
The northwest marginal zone of the orogen includes the Wollaston and Seal River fold belts (Fig. 2) of Saskatchewan and Manitoba. The cover sequence on the margin of the Hearne province is thought to include an early synrift assemblage, subsequent shelf quartzite, and late synorogenic arkosic deposits (Lewry and Sibbald 1980, Stauffer 1984). However, the stratigraphic order and its interpretation are uncertain because of poor outcrop, complex structure involving upright basement-cover folds superimposed on polyphase recumbent folds (Lewry and Sibbald 1980), and upper-amphibolite to granulite-facies metamorphism (Lewry et al. 1978, Schledewitz 1978). There are numerous granitic bodies, particularly in the Seal River belt (Schledewitz 1986). In the absence of U-Pb ages it is difficult to ascertain the relative proportions of granite related to the Wathaman-Chipewyan batholith (1.86-1.85 Ga) to the south, and the 1.75 Ga Rapakiwi suite of the Hearne province, as well as Archean basement inliers.

The Black Hills uplift in South Dakota (Fig. 1) exposes a fold belt marginal to the Wyoming province near the junction of the Trans-Hudson and Central Plains orogens (Sims and Peterman 1986). According to Redden et al. (1987), Archean basement is overlain by synrift clastic sediments that were intruded by a 2.17 Ga mafic sill, then deformed and eroded, prior to deposition of younger rift- and clastic shelf-facies sediments and mafic volcanics. The shelf quartzites, which contain a 1.97 Ga tuff bed, are laterally equivalent to and overlain by thick turbidites capped by 1.88 Ga alkalic tuffs (Redden et al. 1987). The entire sequence was recumbently folded about an easterly axis, then refolded about northwest-trending axis, and finally intruded by the 1.72 Ga Harney Peak granite (Hoffman, 1988).
CHAPTER III
GEOLGY OF LYNN LAKE BELT

The Proterozoic Lynn Lake Belt is a greenstone belt of which extends 130 km east from the Manitoba-Saskatchewan border and has a maximum width of 60 km (Fig. 2). The oldest supracrustal rocks are predominantly mafic volcanics and deformed sedimentary rocks. These have been intruded by mafic, intermediate and minor felsic plutons. These rocks have been deformed, eroded and are overlain by more felsic sedimentary rocks. Late felsic plutons have intruded the greenstone belt in such a way that they have split the belt into two parallel segments, consisting of a Northern Belt and a Southern Belt (Fig. 3).

The first detailed mapping of the Lynn Lake area is that of Milligan (1960). This was followed by Emslie and Moore (1961) who concentrated on the structure, stratigraphy and developed hypothesis for the Ni mineralization at Lynn Lake. Most recent mapping was done by Gilbert et al. (1980) who re-mapped this area in considerable detail. The following description of the geology of the area is based on reports by Gilbert et al. (1980) and Milligan (1960).

The supracrustal rocks in Lynn Lake area are represented by Wasekwan Group and Sickle Group which are separated by an unconformity (Table 1). The Wasekwan Group is subdivided into 9 units. The group comprises about 7000 m of mafic tholeiitic flows and fragmental units intercalated with subordinate felsic volcanic and clastic volcanogenic sedimentary units. The Sickle Group is subdivided into 2 units. It comprises about 4000 m of conglomerate, sandstone, schist and gneiss.

The intrusive rocks are represented by two groups called Pre-Sickle and Post-Sickle. Pre- Sickle Intrusives are subdivided into 4 units. They consist of mafic, through intermediate to felsic
plutons, with gabbro being the dominant rock type. Post-Sickle Intrusives are subdivided into 4 units represented by massive felsic plutons of tonalite to granite in composition.

The following is a more detailed description of the rock units. The stratigraphic sequence is presented in the table of formations (Table 1) and shown as geological sketch map in Fig. 3.

**Supracrustal rocks**

The Proterozoic supracrustal rocks are represented by two groups of mainly metavolcanic and meta-sedimentary rocks. The Wasekwan Group consists of nine units, and the Sickle of two units.

**Wasekwan Group**

Unit 1 includes greywacke, meta-siltstone, meta-mudstone and minor volcanic rocks (Table 1). Dark grey weathering beds of metagreywacke are the most prominent component of this unit. It is composed mostly of quartz, plagioclase and chlorite with small amounts of graphite. Beds of greywacke are massive or graded, 10-20 cm in thickness. Oval lenses of calc-silicate rock up to 30 cm long occur in some of the greywacke beds.

Dark grey and brown meta-mudstone contains quartz, feldspar and biotite as well as porphyroblasts of garnet, cordierite, biotite or muscovite. Graphite is abundant and tourmaline, apatite, zircon and inclusions of staurolite are generally present. Meta-mudstone occurs in 5-10 cm internally laminated beds, interlayered with greywacke.
Table 1. Table of Precambrian formations for Lynn Lake area, Manitoba. Modified from Gilbert et al. (1980). Unit numbers refer to map in Gilbert et al. (1980).

<table>
<thead>
<tr>
<th>Group</th>
<th>Supracrustal Rocks</th>
<th>Intrusive Rocks</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Unit</td>
<td>Rock type</td>
</tr>
<tr>
<td></td>
<td>11-12</td>
<td>Conglomerate with quartz-feldspar porphyry, sedimentary, volcanic and granitoid clasts, sandstone, schist and gneiss</td>
</tr>
<tr>
<td>Sickle</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sickle or</td>
<td>10</td>
<td>Conglomerate with sedimentary, volcanic, and granitoid clasts.</td>
</tr>
<tr>
<td>Wasekwan</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>8-9</td>
<td>Conglomerate, coarse- to fine-grained sedimentary rocks, paragneiss</td>
</tr>
<tr>
<td></td>
<td>5-7</td>
<td>Intermediate and felsic volcanic rocks, dacite, rhyolite and felsic gneiss</td>
</tr>
<tr>
<td></td>
<td>2-4</td>
<td>Aphyric basalt, porphyritic basalt, mafic and intermediate volcanic rocks, amphibolite</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>Greywacke, meta-mudstone, meta-siltstone</td>
</tr>
</tbody>
</table>

Unconformity
Fig. 3. Geological sketch map of the Lynn Lake area and the sample locations (modified after Baldwin et al. 1986). Samples FL1, FL2, 3002, and 3005 from Baldwin et al. 1986.
White meta-siltstone contains mostly quartz and feldspar with variable amounts of microcline and traces of graphite and pyrite. Rocks representing unit 1 occur in Laurie Lake - Dumphy Lake area.

Units 2–4 are represented by aphyric and porphyritic basalts, mafic and intermediate volcanics and amphibolite. Massive flows are commonly pillowed and intercalated with the pillowed facies which include: pillow breccia, mafic tuff, volcanogenic greywacke, and pebble conglomerate which is usually found at the top of the unit. Occurrence of pillowed flows indicate that the aphyric basalt was emplaced in a subaqueous environment. The maximum thickness of unit 2 is about 2000 m.

Porphyritic basalt (unit 3) is the same in age as the aphyric basalt (unit 2) but is chemically distinct from it. Other rock types of unit 3 are: include tuff, amphibolite and mafic porphyry. Massive porphyritic basalt flows 2 to 6 m thick are interlayered with aphyric flows and sills and with porphyritic basalt breccia. Quartz-filled amygaules occur locally in massive flows but are not common. Maximum thickness of unit 3 is about 1500 m.

Unit 4 comprises intermediate volcanic rocks and amphibolite. The intermediate volcanic rocks include: massive, porphyritic and aphyric basalt and andesite, pillowed basalt and andesite, autoclastic and polymictic breccia and mafic to intermediate tuff. Massive, porphyritic and aphyric basalt and andesite flows are the most abundant volcanic rocks in this unit. The porphyritic basalt is distinct from aphyric basalt by high content of Al₂O₃, high total FeO/MgO ratio and the low Ni and Cr content. Individual flows are clearly defined with sharp interflow contacts.

In pillowed basalt the pillows, 20 cm to 1 m long are rimmed by hornblende or epidote-
rich selvages. The composition of the flows is predominantly basaltic, but ranges up to felsic andesite. The majority of basalts are recrystallized to fine-grained amphibolites.

Mafic to intermediate breccias are interlayered with basaltic and andesitic flows and mafic to intermediate tuffs. The majority of the fragmental rocks are autoclastic but their source is not clear. The rocks are generally polymictic and have been interpreted as pyroclastic, hyaloclastic, or reworked breccias.

Garnetiferous amphibolite is fine- to coarse-grained, gneissic, and contains garnets (up to 1.5 cm in size) which may make up to 60% of the rock. It is interpreted as the recrystallized product of volcanoclastic rocks or interlayered mafic flows which were partly contaminated by the sediments.

Units 5-7 consist of intermediate and felsic volcanic rocks, dacite, rhyolite and felsic gneiss. Intermediate and felsic volcanics (unit 5) contain andesite, porphyric dacite, intermediate and lapilli tuff and pyroclastic breccia. Andesite flows in the lower part of the unit are aphyric and in the upper part massive, and are interlayered with intermediate felsic tuff.

Massive, homogenous dacite flows are characterized by euhedral plagioclase phenocrysts (0.5 - 2.5 mm) set in a groundmass of plagioclase microlites and amphibolite, biotite, quartz, and epidote.

Intermediate and lapilli tuffs occur predominantly within a lower portion of the unit. Pyroclastic breccia is clast supported and cemented by fine-grained, white, granular carbonate. Stratification is not developed at the base which indicates that the breccia was emplaced during the single event. The total thickness of unit 5 is about 2700 m.

Unit 6 comprises of massive eupyhr and porphyritic dacite, breccia, tuff, altered dacite and schist. Altered dacite and schist occur in the contact with the gabbro sills in Fox Mine area.
The alteration is about 200 m thick in the core of large bodies of dacite and is 4 km long. The layered structure of these schists suggests that they may have been tuffs and coarser pyroclastic deposits prior to alteration.

Unit 7 includes felsic volcanic rocks represented by massive aphyric and porphyritic rhyolite, porphyritic breccia, hyaloclastite and tuff. The rhyolite body consists of a succession of massive porphyritic rhyolite flows interlayered with massive aphyric rhyolite flows. All of the flows are massive, and homogeneous in composition and texture. Contacts between the flows are sharp to finely brecciated.

Porphyritic breccia occurs as concordant layers within volcanic rocks. It contain pebble-to small cobble-size fragments in a detrital crystal-tuff matrix. Fragments and matrix composition range from felsic to intermediate. The matrix is slightly darker than the majority of clasts. Hyaloclastite is composed of finely fragmental rhyolitic material in a biotite-rich matrix.

Fine-grained felsic and intermediate tuff is layered which is defined by moderate compositional contrast expressed by variation in colour, hardness, and intensity of foliation.

Units 8-9 include conglomerate, coarse- to fine-grained sedimentary rocks, and paragneiss. Conglomerate (unit 8) consists of quartz-pebble conglomerate, conglomerate with volcanic and sedimentary clasts and pebbly meta-mudstone. Clasts are rounded to subrounded felsic volcanic rocks and vein-quartz, angular and slabby laminated mafic greywacke, and subrounded aphyric basalt. The fine-grained mafic meta-mudstone matrix of the conglomerate consists of roughly equal amounts of hornblende and plagioclase, with minor quartz. Interbeds of mafic meta-mudstone are internally laminated and contain isolated pebbles.

Sedimentary rocks of unit 9 consists of greywacke, meta-siltstone and meta-mudstone, argillite, chert and iron formation. Greywacke, meta-siltstone and meta-mudstone rocks are
interlayered. The dark green or grey hornblende greywacke comprises massive beds which locally contain subordinate meta-siltstone laminae. Biotite greywacke and meta-siltstone are generally more layered than hornblendic greywackes. Mafic meta-mudstone occurs locally as sharply defined laminae in meta-siltstone beds. Dark grey argillaceous units contain biotite (up to 65%), quartz, feldspar, garnet and green hornblende porphyroblasts. They occur within feldspatic greywacke units. Chert is a relatively rare and is associated with iron-formation. Magnetite and/or pyrite-pyrrhotite-bearing iron formation occur between fine-grained sedimentary rocks and younger mafic volcanic sequences.

Sickle or Wasekwan Group

Unit 10, which may belong to the Sickle or Wasekwan group, either overlies or is equivalent to the upper part of the Wasekwan group. It consists of conglomerate with sedimentary, volcanic and granitoid clasts. This unit is similar to the Sickle conglomerate, it does not resemble the Wasekwan conglomerate.

The conglomerate is composed of pebbles consisting of quartzite, granite, gabbro, diorite and greywacke. The matrix is a brownish grey quartz-mica schist, with coarser clastic grains of quartz.

Sickle Group

Unit 11 is a polymictic conglomerate with clasts of quartzite, greywacke, arkose, chert, vein quartz, schist and granite. The matrix is usually dark grey, fine-grained and composed of quartz, muscovite, biotite and some feldspar. The conglomerate grades upwards into the overlying arkose. The thickness of unit 11 is about 1000 m.
Unit 12 is an arkosic sequence predominantly a medium-grained sandstone and minor pebbly sandstone derived from a granitic terrane. Rock contains 70-80% of quartz, 20% of brown or pinkish feldspar, and up to 10% of biotite and muscovite. Magnetite, hematite and apatite are accessory minerals.

The character of the sediments of the Sickle Group suggests that they were deposited in a shallow water basin. The total thickness of unit 12 is about 2000 m.

Intrusive rocks

Pre-Sickle Intrusives

Pre-Sickle intrusive rocks consist of 4 units. Units 13-15 consist of mafic and ultramafic intrusive rocks and units 16-17 consist of intermediate and felsic intrusive rocks. Gabbro, including minor amounts of norite and diabase are the dominant rock types of units 13 and 15. It occurs in the form of gabbro plugs intruded into rocks of the Wasekwan group. In the Lynn Lake area there are three important gabbroic plugs. The "A" plug and the "EL" plug contain Ni, Cu, Co - ore bodies. The mineralization is disseminated to massive. The third plug, the Fraser Lake gabbro is barren. Petrologically the three plugs are very similar. The common gabbro of this unit is a dark green to black, coarse-grained, with 45-65% of light green to dark bluish-green hornblende, 30-45% of feldspar and 1-5% of magnetite, epidote and chlorite. The hornblende found in the gabbro is probably secondary. It appears that the gabbro has been strongly uralitized, and small amounts of pyroxene are now present. In many places the gabbro shows a mineral assemblage characteristic of low grade metamorphism (chlorite and epidote).

Diorite (units 14 and 16) is composed of 40-60% of hornblende and biotite evenly distributed in a matrix of feldspar. Feldspar which comprises about 40-50% of the rock is white
to greyish white in colour and very fine-grained. It shows alterations with development of epidote, zoisite and carbonate. Quartz is present but rarely in amounts greater than 5%. It is clear, smoky or bluish in colour. Diorite is usually massive, with small bodies of gabbro and tonalite locally present. The dominant host for the sulphide mineralization is this diorite.

Tonalite (unit 16) cuts the diorite in many places. It is typically light brownish grey with the surface mottled by grains and irregular aggregates of hornblende and biotite in a matrix of quartz and feldspar. It is composed of about 25% mafic minerals, 25% quartz and rest is feldspar. Rock is generally massive and medium-grained. Alteration to epidote, zoisite, carbonate and sericite are common.

Unit 17 consist mostly of granite, granodiorite and aplite with minor pegmatite and syenite. Granite is medium- to coarse-grained and low in mafic minerals. Feldspar occurs as rounded or irregular grains, quartz may be clear, grey, or bluish, biotite is the usual dark mineral and is rarely over 10%. The granite is generally uniform and massive. Granodiorite is medium- to coarse-grained, massive or distinctly foliated and sometimes intrudes more mafic plutonic as well as volcanic and sedimentary rocks.

Aplite and pegmatite are associated with granite. Fine-grained aplite is bright pink and cuts granite as tabular dikes and large masses. The rock is composed mostly of quartz and orthoclase with small amounts of mica and dark, mafic minerals. Quartz-feldspar pegmatite, buff or pink in colour is also present in the granite in the form of dikes and masses, but is not common.

Post-Sickle Intrusives

Post-Sickle intrusive rocks are represented by units 18-22. They ranging in composition from gabbro and diorite to granite and intrude rocks of both the Sickle and Wasekwan groups.
Diorite (unit 18) is medium-grained, massive and slightly gneissic. It is composed of biotite, hornblende and quartz in a matrix of grey feldspars. Pyrite and magnetite (2-3%) are closely associated with dark minerals, and apatite is a minor accessory. Feldspar crystals show alteration to sericite, and some of the biotite has been formed by alteration from hornblende.

Tonalite (unit 19) is generally strongly gneissic with the foliation marked by streaks of biotite and hornblende. Quartz content maybe from trace to up to 40%. Feldspar crystals are altered to epidote and some sericite.

Granite (unit 20) is pink to brownish on weathered surfaces and contains irregular aggregates of black biotite, dark quartz, and pinkish feldspar. The feldspar grains are slightly altered to kaolin, and some contain finely divided hematite. Close to the margin the granite is more grey in colour and has 10-15% of biotite with a decreased amount of potash feldspar. Texture of the rock changes from aplitic and medium-grained to porphyritic. Granodiorite (unit 20) is leucocratic with orthoclase which is altering to kaolin. This rock may represent a marginal phase of early consolidation, or may be a separate intrusion.

Pegmatite, aplite and graphic granite (unit 21) occur in form as irregular dikes and granitic bodies and their occurrences are limited. Pegmatite is a coarse-grained quartz-feldspar rock that intrudes the Sickle group in form of narrow sills and dikes. Aplite which is associated with pegmatites and occurs as a dikes in the pegmatites. Graphic granite bodies are also associated with pegmatites.

Siliceous porphyry dikes (unit 22) which occur in vicinity of granitic bodies are represented by quartz-feldspar porphyry, quartz porphyry, feldspar porphyry, and felsite. There is no distinct contact between the porphyry and granite, and it is possible that they represent a chilled border phase of the granite.
Deformation and metamorphism

The Lynn Lake greenstone belt is part of the La Ronge - Lynn Lake Domain which is a linear belt of metamorphosed volcanic and volcanoclastic rocks, trending NE-SW. The volcanic and volcanoclastic rocks of the Wasekwan Group (Table 1) have been intruded by small subvolcanic plutons, and subsequently folded, faulted and intruded by larger mafic and felsic plutons (Gilbert et al. 1980). Sickle Group, composed of sandstone and conglomerate (Table 1), was deposited unconformably on the Wasekwan Group and Pre-Sickle intrusions. Regional metamorphism in the upper greenschist to upper amphibolite facies and extensive deformation and plutonism, at the margins of the greenstone belt, followed deposition of the Sickle Group.

There are essentially two phases of deformation and metamorphism in this area. The first deformation and plutonism in the Lynn Lake area took place during and after deposition of the Wasekwan Group. The early faulting, folding and intrusion of composite plutons was restricted to the greenstone belt. The Post-Sickle deformation and metamorphism involved the adjacent sedimentary belts and also greenstones and Pre-Sickle intrusives.

Wasekwan Group

Block faulting occurred during deposition of the Wasekwan sediments in the western part of the Lynn Lake belt. Tectonic depressions were filled by conglomeratic turbidities. Structural discontinuity exists along the northwestern boundary of the Lynn Lake belt, as indicated by abrupt facies change between mafic volcanic rocks and greywacke.

The Wasekwan Group is isoclinically folded. Folds are east-northeast trending and are steeply dipping. They are truncated by Pre-Sickle tonalite-granite intrusions. Large early folds
dominate the structure of the southern greenstone belt and occur in the eastern part of northern greenstone belt. Axial traces are generally curved and are parallel to the regional foliation. Major anticlines are developed in basaltic rocks of units 2 and 3. A syncline-anticline pair occurs in the intermediate and felsic volcanic rocks of units 5, 6 and 7. A syncline is present in greywacke of unit 9 and volcanic rocks of unit 4.

In the northern belt folding is present in the sedimentary section and is indicated by reversals in graded bedding. A large anticline occurs in the lower part of the sequence and is truncated by Pre-Sickle intrusive rocks.

The deformed Wasekwan Group rocks were intruded by gabbro, norite and minor ultramafic rocks of unit 15, and subsequently, large composite plutons of diorite, tonalite, granodiorite and granite of units 16 and 17 were intruded into the supracrustal rocks and now split the belt into the northern and southern greenstone belts.

The grade of metamorphism of Wasekwan Group rocks ranges from upper greenschist facies to middle amphibolite facies. Upper greenschist facies assemblages are developed in the eastern part of the southern belt and locally in the northern belt. Middle amphibolite facies assemblages are more common and comprise the greater portion of Lynn Lake belt whereas upper amphibolite facies assemblages are present in western part of the belt mainly in the Fox Mine area.

Sickle Group

The Sickle Group unconformably overlies the early plutons and deformed Wasekwan Group rocks. There are no major faults present in the rocks represented by this group, and deformation is related mostly to folding. North of Lynn Lake belt the distribution of sedimentary rocks is controlled by east-northeast trending folds, openly refolded about northeast
trending axial surfaces. The earlier structures are parallel to folds in the Wasekwan Group.

Structures developed in the Sickle Group within the Lynn Lake belt comprise north and east
trending folds, forming basins at eastern part of southern belt (Hughes Lake) and western part of
southern belt (Conglomerate Lake).

The Post-Sickle deformation and metamorphism produced the secondary foliation in the
Wasekwan Group, Sickle Group and early plutons. This regional foliation developed co-planar to
a Pre-Sickle foliation. The Wasekwan Group was not refolded during this later event. The early
folds become more appressed during the development of the foliation. Brittle structures were
formed locally.

Geochronology of the Lynn Lake area

A compilation of published isotopic ages for the Lynn Lake area which is relevant to this
study is shown in Table 2. Rocks are Proterozoic in age. Much higher scatter, ca. ± 100 Ma, in
K-Ar and Rb-Sr ages occurs because these methods of dating are more sensitive to tectonic
events compared to U-Pb ages.
Table 2. Compilation of isotopic ages from Lynn Lake area. Taken from Syme et al. (1993).

<table>
<thead>
<tr>
<th>Sample number and location</th>
<th>Method</th>
<th>Age (Ma)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Lynn Lake Domain</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>FL1</td>
<td>U-Pb</td>
<td>1915 + 7/-6</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td>FL2</td>
<td>U-Pb</td>
<td>1910 + 15/-10</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td><strong>Pre-Sickle Intrusive Rocks</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hughes Lake quartz diorite</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3002</td>
<td>U-Pb</td>
<td>1876 + 8/-7</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td>Norrie Lake tonalite</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3005</td>
<td>U-Pb</td>
<td>1876 + 8/-6</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td><strong>Hughes Lake, Sickle Lake</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(pre-Sickle plutons)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Granodiorite-tonalite</td>
<td>Rb-Sr</td>
<td>1940 ± 75</td>
<td>Clark (1980)</td>
</tr>
<tr>
<td><strong>Lynn Lake area</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(Wasekwan group)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rhyolite-andesite</td>
<td>Rb-Sr</td>
<td>1835 ± 75</td>
<td>Clark (1980)</td>
</tr>
<tr>
<td><strong>Pool Lake</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(pre-Sickle pluton)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Quartz diorite</td>
<td>Rb-Sr</td>
<td>1825 ± 99</td>
<td>Clark (1980)</td>
</tr>
<tr>
<td><strong>Lynn Lake, Gemmell Lake</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(Wasekwan group)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rhyolite</td>
<td>Rb-Sr</td>
<td>1790 ± 35</td>
<td>Clark (1980)</td>
</tr>
<tr>
<td><strong>Berge Lake</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(post-Sickle pluton)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Granodiorite</td>
<td>Rb-Sr</td>
<td>1765 ± 99</td>
<td>Clark (1980)</td>
</tr>
</tbody>
</table>
Table 2. (Continued)

<table>
<thead>
<tr>
<th>Sample number and location</th>
<th>Method</th>
<th>Age (Ma)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>3 km SW Gemmell Lake</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(Sickle group)</td>
<td>Arkose</td>
<td>Rb-Sr</td>
<td>1760 ± 85</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Rb-Sr</td>
<td>1855 ± 85</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Rb-Sr</td>
<td>1845 ± 99</td>
</tr>
<tr>
<td>Lynn Lake area</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(felsic plutons)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tonalite-granodiorite</td>
<td>Rb-Sr</td>
<td>1855 ± 85</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Rb-Sr</td>
<td>1845 ± 99</td>
<td></td>
</tr>
<tr>
<td>Lynn Lake</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Paragneiss</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B3875</td>
<td>K-Ar</td>
<td>1750</td>
<td></td>
</tr>
<tr>
<td></td>
<td>K-Ar</td>
<td>1730</td>
<td></td>
</tr>
<tr>
<td>Granite</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B3812</td>
<td>K-Ar</td>
<td>1700</td>
<td></td>
</tr>
<tr>
<td></td>
<td>K-Ar</td>
<td>1640</td>
<td></td>
</tr>
<tr>
<td>N Bay of Sickle Lake</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Granite gneiss</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GSC 60-76</td>
<td>K-Ar</td>
<td>1655</td>
<td></td>
</tr>
<tr>
<td>Fox Lake Mine, Lynn Lake</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SP-1785</td>
<td>Pb-Pb</td>
<td>1832</td>
<td></td>
</tr>
<tr>
<td>Lynn Lake</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>No. 1. Biotite</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>No. 2. Pyroxene</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>K-Ar</td>
<td>1716 ± 50</td>
<td></td>
</tr>
<tr>
<td></td>
<td>K-Ar</td>
<td>2045 ± 100</td>
<td></td>
</tr>
<tr>
<td>South Indian Domain</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Brochet/Big Sand</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Paragneiss</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Rb-Sr</td>
<td>1975 ± 99</td>
<td></td>
</tr>
<tr>
<td>Quartz diorite-tonalite</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Rb-Sr</td>
<td>1910 ± 50</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Rb-Sr</td>
<td>1878 ± 87</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Rb-Sr</td>
<td>1845 ± 65</td>
<td></td>
</tr>
</tbody>
</table>
Table 2. (Continued).

<table>
<thead>
<tr>
<th>Sample number and location</th>
<th>Method</th>
<th>Age (Ma)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Northern Indian Lake</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Granite</td>
<td>K-Ar</td>
<td>1800</td>
<td>Moore et al. (1960)</td>
</tr>
<tr>
<td>B3915</td>
<td>K-Ar</td>
<td>1750</td>
<td>Moore et al. (1960)</td>
</tr>
<tr>
<td>B3907</td>
<td>K-Ar</td>
<td>1720</td>
<td>Moore et al. (1960)</td>
</tr>
<tr>
<td><strong>Partridge Breast Lake</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Paragneiss</td>
<td>K-Ar</td>
<td>1720</td>
<td>Moore et al. (1960)</td>
</tr>
<tr>
<td>B3910</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Northern Indian Lake</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Granodiorite GSC 59-40</td>
<td>K-Ar</td>
<td>1730</td>
<td>Lowdon (1960)</td>
</tr>
<tr>
<td><strong>Veronica Lake</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Paragneiss</td>
<td>K-Ar</td>
<td>1610</td>
<td>Van Schmus and Schledewitz (1986)</td>
</tr>
<tr>
<td>GSC 60-75</td>
<td>U-Pb</td>
<td>1832 ± 11</td>
<td></td>
</tr>
<tr>
<td>Quartz diorite - MAN84-8</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Leaf Rapids Domain</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Rusty Lake Belt</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Karsakuwigamak Block rhyolite</td>
<td>U-Pb</td>
<td>1878 ± 3</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td>R1</td>
<td>U-Pb</td>
<td>1874 +/-7</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td>R4</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Rutan Mine, Leaf Rapids</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>SP-2318</td>
<td>Pb-Pb</td>
<td>1768</td>
<td>Sangster (1978)</td>
</tr>
</tbody>
</table>
CHAPTER IV

THE ISOTOPIC U–Th–Pb METHODS OF DATING

Basic theory

U–Pb and Th–Pb isotopic ages determinations derive from the decay of U and Th parents which decay to stable Pb daughter isotopes as per equation

$$\lambda t = \ln(D/P + 1)$$

where, $\lambda$ is decay constant, $D$ is daughter, $P$ is parent and "$t$" is time since formation or closure of system. The following review of U–Th–Pb systematics comes from several sources, but mostly from Faure (1986) and Doe (1970).

U has essentially two isotopes $^{235}\text{U}$ and $^{238}\text{U}$ which decay to $^{207}\text{Pb}$ and $^{206}\text{Pb}$ respectively. The ratio of $^{235}\text{U}/^{238}\text{U}$ is 138.88. There is (was) a third isotope $^{234}\text{U}$ which is extremely radioactive and has essentially disappeared. Th has only one isotope $^{232}\text{Th}$ which decays to $^{208}\text{Pb}$. Pb isotopes generated from radioactive decay add on to Pb isotopes present in the mineral as common Pb. So the three equations above have to be modified to subtract any common Pb present at time of formation of the mineral.

Ordinary lead has four naturally occurring isotopes: $^{206}\text{Pb}$, $^{207}\text{Pb}$, $^{206}\text{Pb}$, and $^{204}\text{Pb}$. Only $^{204}\text{Pb}$ is not radiogenic, although it is very weakly radioactive and decays to stable $^{200}\text{Hg}$ by alpha emissions with a half-life of $1.4 \times 10^{17}$ years (Holden and Walker, 1972). Because of its very long half-life, $^{204}\text{Pb}$ is used as a stable reference isotope. The isotopic composition of lead in minerals
containing uranium and thorium can be expressed in the form of three equations:

\[
\frac{^{207}\text{Pb}}{^{204}\text{Pb}} = \left( \frac{^{207}\text{Pb}}{^{204}\text{Pb}} \right)_0 + \frac{^{235}\text{U}}{^{204}\text{Pb}} (e^{\lambda_1 t} - 1)
\]  

(1)

\[
\frac{^{206}\text{Pb}}{^{204}\text{Pb}} = \left( \frac{^{206}\text{Pb}}{^{204}\text{Pb}} \right)_0 + \frac{^{238}\text{U}}{^{204}\text{Pb}} (e^{\lambda_2 t} - 1)
\]  

(2)

\[
\frac{^{208}\text{Pb}}{^{204}\text{Pb}} = \left( \frac{^{208}\text{Pb}}{^{204}\text{Pb}} \right)_0 + \frac{^{232}\text{Th}}{^{204}\text{Pb}} (e^{\lambda_3 t} - 1)
\]  

(3)

where \( \frac{^{206}\text{Pb}}{^{204}\text{Pb}}, \frac{^{207}\text{Pb}}{^{204}\text{Pb}}, \frac{^{208}\text{Pb}}{^{204}\text{Pb}} \) are the isotope ratios of lead in the mineral at the time of analysis, i.e. present day, and:

\[
\left( \frac{^{206}\text{Pb}}{^{204}\text{Pb}} \right)_0, \left( \frac{^{207}\text{Pb}}{^{204}\text{Pb}} \right)_0, \left( \frac{^{208}\text{Pb}}{^{204}\text{Pb}} \right)_0
\]  

are the initial isotope ratios of lead incorporated into the mineral at the time of its formation, i.e. at time \( t \),

\[
\frac{^{238}\text{U}}{^{235}\text{U}}, \frac{^{235}\text{U}}{^{204}\text{Pb}}, \frac{^{232}\text{Th}}{^{204}\text{Pb}}
\]  

are isotope ratios in the mineral at the time of analysis,

\( \lambda_1, \lambda_2, \lambda_3 \) are decay constants of \(^{238}\text{U}, ^{235}\text{U}, \) and \(^{232}\text{Th}, \) respectively, and

\( t = \) time elapsed since closure of the mineral to uranium, thorium, lead and all intermediate daughters.

In order to date uranium and thorium bearing minerals using Equations 1 to 3 the concentrations of uranium, thorium, and lead are usually measured by mass spectrometry using
together with measurements of the raw isotope dilution isotopic composition of lead. The equations can then be solved for "t" using assumed values for the initial lead isotope ratios:

$$t_{206} = \frac{1}{\lambda_1} \ln \left[ \frac{\frac{206\text{Pb}}{204\text{Pb}}}{\frac{206\text{Pb}}{204\text{Pb}}} \left( \frac{206\text{Pb}}{204\text{Pb}} \right)^{\lambda_1} + 1 \right]$$

(4)

The other equations are solved similarly, resulting in three independent ages based on the three separate decay series. These ages will be concordant and then represent the age of the mineral, provided the following conditions are satisfied.

1. The mineral has remained closed to uranium, thorium, lead, and all intermediate daughters throughout its history.

2. Correct values are used for the initial lead isotope ratios.

3. The decay constants of $^{238}\text{U}$, $^{235}\text{U}$, and $^{232}\text{Th}$ are known accurately.

4. The isotopic composition of uranium is normal and has not been modified by isotope fractionation or by the occurrence of a natural chain reaction based on a spontaneous fission of $^{235}\text{U}$.

5. All analytical results are accurate and free of systematic errors.

In many cases the ages so calculated for minerals containing uranium and thorium are not concordant. The reason for this is that most minerals are not closed systems, but may lose or gain lead, uranium, thorium, or intermediate daughters after crystallization. The effect of lead loss on
U–Pb ages can be minimized by calculating ages based on the $^{207}\text{Pb}/^{206}\text{Pb}$ ratio. This ratio is insensitive to lead loss, especially when the loss occurred recently and when the lead that was lost had the same isotope composition as the lead that remained, that is, when no isotope fractionation occurred. The relationship between the $^{207}\text{Pb}/^{206}\text{Pb}$ ratio and time results from the difference in the half-lives of their respective parents. The equation expressing this relationship can be derived by combining Equations 1 and 2:

$$\frac{^{207}\text{Pb}}{^{206}\text{Pb}} - \left(\frac{^{207}\text{Pb}}{^{206}\text{Pb}}\right)_0 = \frac{^{235}\text{U}}{^{238}\text{U}} \left[ \frac{e^{t_x} - 1}{e^{t_x} - 1} \right]$$

(5)

This equation involves the ratio $^{235}\text{U}/^{238}\text{U}$ which is a constant equal to 1/137.88 for all uranium of normal isotopic composition in the Earth, the moon and meteorites at the present time (Steiger and Jäger, 1977). Therefore, a $^{207}\text{Pb}/^{206}\text{Pb}$ age can be calculated without knowing the concentration of uranium or lead in the mineral.

$$\frac{^{207}\text{Pb}}{^{206}\text{Pb}} - \left(\frac{^{207}\text{Pb}}{^{206}\text{Pb}}\right)_0 = \left(\frac{^{207}\text{Pb}}{^{206}\text{Pb}}\right)^*$$

(6)
where \((^{207}\text{Pb}/^{206}\text{Pb})^*\) is the ratio of radiogenic \(^{207}\text{Pb}\) to radiogenic \(^{206}\text{Pb}\). This ratio can be calculated by subtracting an assumed initial \(^{207}\text{Pb}/^{206}\text{Pb}\) ratios from the measured values of these ratios. "\(^{207}\text{Pb}/^{206}\text{Pb}\) ages" can be calculated only by knowing the isotopic composition of lead in the mineral without having to know its concentration. The equation:

\[
\left( \frac{^{207}\text{Pb}}{^{206}\text{Pb}} \right)^* = \frac{^{235}\text{U}}{^{238}\text{U}} \left( \frac{e^{^{235}\text{U}}}{e^{^{238}\text{U}}} - 1 \right)
\]

(7)
cannot be solved for "t" by algebraic methods because it is transcendental. One possible way to solve it is to make a table of \((^{207}\text{Pb}/^{206}\text{Pb})^*\) ratios for selected values of "t" or to use tables such as those of Stieff et al. (1959) and Stacey and Stern (1973) to interpolate "t" for any value of the \((^{207}\text{Pb}/^{206}\text{Pb})^*\) ratio. A graph for compatible values of \((^{207}\text{Pb}/^{206}\text{Pb})^*\) and "t" are provided in Figure 4. Thus the U–Th–Pb system can provide 4 ages based on the following ratios: \(^{235}\text{U}/^{207}\text{Pb}\); \(^{238}\text{U}/^{206}\text{Pb}\); \(^{207}\text{Pb}/^{206}\text{Pb}\); \(^{232}\text{Th}/^{208}\text{Pb}\). The \(^{232}\text{Th}/^{208}\text{Pb}\) ages are typically incorrect, they are low due to loss of Th and hence are generally not determined.

Although uranium and thorium occur in a large number of minerals, only a few are suitable for dating by the U–Th–Pb method. To be useful for dating, a mineral must be retentive with respect to uranium, thorium, lead, and the intermediate daughters, and it should be widely distributed in variety of rocks. The mineral that most nearly satisfies these conditions is zircon.

Several other minerals are also useful: uraninite (pitchblende), monazite, sphene, apatite, and other accessory minerals in igneous rocks.

The concentrations of uranium and thorium in zircon range from a few hundred to a few thousand parts per million and average 1330 and 560 ppm, respectively. Zircons in pegmatites
may contain more uranium and thorium than those in ordinary igneous rocks. The presence of these elements in zircon can be attributed both to isomorphous replacement of Zr\(^{4+}\) (ionic radius = 0.87Å) by U\(^{4+}\) (1.05Å) and Th\(^{4+}\) (1.10Å) and to the presence of inclusions of thorite. The substitution of Zr\(^{4+}\) by U\(^{4+}\) and Th\(^{4+}\) is limited by differences in their ionic radii. Equally important is the fact that Pb\(^{2+}\), which has an ionic radius of 1.32Å, is excluded from zircon because of its large radius and lower charge. Therefore, zircon contains very little lead at the time of formation and has very high U/Pb and Th/Pb ratios which enhance its sensitivity as a geochronometer. For this reason, zircon is used most frequently for dating by the U–Th–Pb isotopic method (Silver and Deutsch, 1963; Steiger and Wasserburg, 1969; Doe, 1970; Allegre et al., 1974).

The U–Pb concordia diagram

The decay of the naturally occurring isotopes of uranium to lead gives rise to two independent geochronometers. When the mineral being dated has remained closed to uranium and all of its daughters and when appropriate corrections are made for the lead incorporated into the mineral at the time of its formation, the two geochronometers give concordant ages. Failure to yield ages indicates loss of daughter, parent or intermediate product. In this as in other age dating schemes (eg. Rb–Sr, K–Ar, Nd–Sn) the most likely problem is loss of daughter. The decay of \(^{238}\)U to \(^{206}\)Pb as a function of time is described by Equation 1, which can be rewritten as:

\[
\frac{^{206}\text{Pb}}{^{238}\text{U}} = e^{11t} - 1
\] (8)
where

\[
\frac{^{206}\text{Pb}}{^{238}\text{U}} = \frac{^{206}\text{Pb}}{^{204}\text{Pb}} \cdot \left( \frac{^{206}\text{Pb}}{^{204}\text{Pb}} \right)_0 \frac{^{238}\text{U}}{^{204}\text{Pb}}
\]  \tag{9}

The decay of \(^{235}\text{U}\) to \(^{207}\text{Pb}\) is expressed similarly:

\[
\frac{^{207}\text{Pb}}{^{235}\text{U}} = e^{21t} - 1
\]  \tag{10}

A uranium-bearing mineral that satisfies all of the assumptions of dating yields concordant ages by solution of Equations 8 and 10. We can also calculate compatible sets of \(^{206}\text{Pb}^{*}/^{238}\text{U}\) and \(^{207}\text{Pb}^{*}/^{235}\text{U}\) ratios for specified values of "t". These are the coordinates of points representing U–Pb systems that have concordant ages in coordinates of \(^{206}\text{Pb}^{*}/^{238}\text{U}\) (ordinate) and \(^{207}\text{Pb}^{*}/^{235}\text{U}\) (abscissa). Equations 8 and 10 therefore are the parametric equations of a curve that is the locus of all concordant U–Pb systems. Wetherill (1956) called this curve the "concordia." It can be plotted from the data compiled in Table 3 which contains values of \(e^{11t} - 1\) and \(e^{21t} - 1\) for different values of t. Such a concordia curve is shown in Figure 4.

Two scenarios may occur as the U–Pb system moves on the concordia diagram (Fig. 5).

1. At the time of crystallization, a U-bearing mineral contains no radiogenic lead and such a system plots at the origin point of the diagram. The U–Pb system will be represented by a point
that moves along the concordia curve as long as the system remains closed to U and all of its
daughters. The age of the system at any time after its formation is indicated by its location on the
concordia curve. As long as the system remains closed, it resides on concordia and its U–Pb ages
are concordant and indicate the age of the system.

2. System can experience an episode of lead loss (or gain of uranium) as a result of
metamorphism or chemical weathering. Loss of radiogenic lead changes the coordinates of the
point representing the system. If all of the radiogenic lead is lost, the point representing the
system returns to the origin. In this case the U–Pb geochronometers of the system are reset to
zero and all memory of the earlier history of the system is lost.

If the system loses only a fraction of the lead, it will be represented by a point somewhere
on the chord. All such systems on the chord have discordant ages, and the chord is called
discordia. A position of a U–Pb system on discordia depends of the fraction of radiogenic lead
that remains. A system that lost all of its radiogenic lead and returned to the origin will once
again move along concordia after it becomes closed to uranium and all of its daughters.

U-bearing minerals such as zircon, sphene, monazite, or apatite lose only a fraction of
their radiogenic lead. The lead loss is related to the size of the crystals, uranium concentration,
and to the radiation damage to the crystals. Thus the concordia diagram is an ideal mechanism to
evaluate U–Pb ages.
<table>
<thead>
<tr>
<th>$\times 10^9 y$</th>
<th>$e^{41%} - 1$</th>
<th>$e^{43%} - 1$</th>
<th>$\frac{^{207}\text{Pb}^*}{^{206}\text{Pb}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>0.0000</td>
<td>0.0000</td>
<td>0.04607</td>
</tr>
<tr>
<td>0.2</td>
<td>0.0315</td>
<td>0.2177</td>
<td>0.05014</td>
</tr>
<tr>
<td>0.4</td>
<td>0.0640</td>
<td>0.4828</td>
<td>0.05473</td>
</tr>
<tr>
<td>0.6</td>
<td>0.0975</td>
<td>0.8056</td>
<td>0.05994</td>
</tr>
<tr>
<td>0.8</td>
<td>0.1321</td>
<td>1.1987</td>
<td>0.06584</td>
</tr>
<tr>
<td>1.0</td>
<td>0.1678</td>
<td>1.6774</td>
<td>0.07264</td>
</tr>
<tr>
<td>1.2</td>
<td>0.2046</td>
<td>2.2603</td>
<td>0.08017</td>
</tr>
<tr>
<td>1.4</td>
<td>0.2426</td>
<td>2.9701</td>
<td>0.08886</td>
</tr>
<tr>
<td>1.6</td>
<td>0.2817</td>
<td>3.8344</td>
<td>0.09877</td>
</tr>
<tr>
<td>1.8</td>
<td>0.3221</td>
<td>4.8869</td>
<td>0.11010</td>
</tr>
<tr>
<td>2.0</td>
<td>0.3638</td>
<td>6.1655</td>
<td>0.12308</td>
</tr>
<tr>
<td>2.2</td>
<td>0.4067</td>
<td>7.7202</td>
<td>0.13790</td>
</tr>
<tr>
<td>2.4</td>
<td>0.4511</td>
<td>9.6296</td>
<td>0.15492</td>
</tr>
<tr>
<td>2.6</td>
<td>0.4968</td>
<td>11.9437</td>
<td>0.17447</td>
</tr>
<tr>
<td>2.8</td>
<td>0.5440</td>
<td>14.7617</td>
<td>0.19693</td>
</tr>
<tr>
<td>3.0</td>
<td>0.5926</td>
<td>18.1931</td>
<td>0.22279</td>
</tr>
<tr>
<td>3.2</td>
<td>0.6428</td>
<td>22.3716</td>
<td>0.25257</td>
</tr>
<tr>
<td>3.4</td>
<td>0.6946</td>
<td>27.4597</td>
<td>0.28690</td>
</tr>
<tr>
<td>3.6</td>
<td>0.7480</td>
<td>33.6556</td>
<td>0.32653</td>
</tr>
<tr>
<td>3.8</td>
<td>0.8030</td>
<td>41.2004</td>
<td>0.37232</td>
</tr>
<tr>
<td>4.0</td>
<td>0.8599</td>
<td>50.3878</td>
<td>0.42525</td>
</tr>
<tr>
<td>4.2</td>
<td>0.9185</td>
<td>63.5763</td>
<td>0.48651</td>
</tr>
<tr>
<td>4.4</td>
<td>0.9789</td>
<td>75.1984</td>
<td>0.55746</td>
</tr>
<tr>
<td>4.6</td>
<td>1.0413</td>
<td>91.7873</td>
<td>0.63969</td>
</tr>
</tbody>
</table>

Table 3. Numerical values of $e^{41%} - 1$, $e^{43%} - 1$, and the radiogenic $^{207}\text{Pb}^{206}\text{Pb}$ ratio as a function of time ($t$) (after Faure 1986).

Fig. 4. Plot of radiogenic $^{207}\text{Pb}^{206}\text{Pb}$ versus time. (after Faure 1986).
Fig. 5. Concordia diagram illustrating a history of episodic Pb loss or uranium gain (after Faure 1986). $\tau$ represents time elapsed since closure; $\tau_0$ represents time elapsed since original crystallization of the minerals (or rock); $\tau'$ represents an age of U-bearing system on concordia which lost lead; $Q$ represents a system that lost only fraction of its radiogenic lead.
CHAPTER V

THE COMMON LEAD SYSTEM

Basic theory

Common lead is any lead from a phase with low value of U/Pb and/or Th/Pb. Such phases are galenas and other sulphides, feldspars and K-feldspars in which Pb$^{2+}$ replaces K$^{+1}$. Isotope data on common lead may be used in determining ages and in the solution of genetic problems (Doe, 1970).

The end products of the $^{238}\text{U}$, $^{235}\text{U}$ and $^{232}\text{Th}$ radioactive series decay are the stable isotopes $^{206}\text{Pb}$, $^{207}\text{Pb}$ and $^{208}\text{Pb}$ respectively. A fourth stable isotope of lead, $^{204}\text{Pb}$, occurs naturally and has not been generated by any radioactive decay. The isotopes of $^{206}\text{Pb}$, $^{207}\text{Pb}$ and $^{208}\text{Pb}$ in any lead consist of two parts:

$$\text{Common Pb} = \text{Primeval Pb} + \text{Radiogenic Pb}$$

The isotopic composition of lead existing at the time when the Earth formed is termed primeval Pb and it appears to be the same for the Earth and meteorites. Radiogenic Pb consists of the isotopes generated by U and Th decay since the formation of the Earth. These radiogenic atoms, which are indistinguishable from primeval atoms, become mixed with the original lead. The parent radioactive nuclei of $^{238}\text{U}$, $^{235}\text{U}$ and $^{232}\text{Th}$ each produce another unstable nucleus so that a decay series results with the end product being a stable lead isotope.
\[ ^{235}\text{U} \rightarrow ^{207}\text{Pb} + 7 \alpha + 4\beta \]
\[ ^{238}\text{U} \rightarrow ^{206}\text{Pb} + 8 \alpha + 6\beta \]
\[ ^{232}\text{Th} \rightarrow ^{208}\text{Pb} + 6 \alpha + 4\beta \]

Number of radioactive atoms can be calculated from the equation:

\[ N = N_0 e^{-\lambda t} \quad (11) \]

where \( N_0 \) = number of atoms at the time \( t = 0 \)
\( N \) = number of radioactive atoms at any time
\( \lambda \) = decay constant

The common lead method is based on evolution of the lead since time "\( T \)“, when the Earth started to act as a closed system for \( \text{U}, \text{Th} \) and \( \text{Pb} \) to the time "\( t \)" when the lead was isolated from contact with radioactive parents. Under these conditions the number of \( ^{206}\text{Pb} \) atoms present at time "\( t \)”, in any such closed system is

\[ N(^{206}\text{Pb})_t = N(^{206}\text{Pb})_T + N(^{238}\text{U})_T - N(^{238}\text{U})_t \]

Both sides of the equation may be divided by the number of \( ^{206}\text{Pb} \) atoms, while the \( \text{U} \) abundances at times "\( T \)" and "\( t \)" may be expressed in terms of equation (1).
\[
\frac{N(^{206}\text{Pb})_t}{N(^{204}\text{Pb})} = \frac{N(^{206}\text{Pb})_0}{N(^{204}\text{Pb})} + \frac{N_0(^{238}\text{U})}{N(^{204}\text{Pb})} e^{\lambda_1 t} - \frac{N_0(^{238}\text{U})}{N(^{204}\text{Pb})} e^{\lambda_5 t}
\]

The equations for $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ become:

\[
(^{206}\text{Pb}/^{204}\text{Pb})_t = (^{206}\text{Pb}/^{204}\text{Pb})_0 + \mu (e^{\lambda_1 t} - e^{\lambda_5 t})
\]

(12)

\[
(^{207}\text{Pb}/^{204}\text{Pb})_t = (^{207}\text{Pb}/^{204}\text{Pb})_0 + \mu (e^{\lambda_1 t} - e^{\lambda_5 t})/137.88
\]

(13)

\[
(^{208}\text{Pb}/^{204}\text{Pb})_t = (^{208}\text{Pb}/^{204}\text{Pb})_0 + W(e^{\lambda_1 T} - e^{\lambda_2 t})
\]

(14)

where \((^{206}\text{Pb}/^{204}\text{Pb})_0 = a_0 = 9.30\)

\[(^{207}\text{Pb}/^{204}\text{Pb})_0 = b_0 = 10.294\]

\[(^{208}\text{Pb}/^{204}\text{Pb})_0 = c_0 = 29.479\]

These equations represent primeval Pb composition at time "T" (Tatsumoto, 1973), where:

\[T = \text{age of the Earth} = 4.570 \text{ Ga (Tilton, 1973)}\]

\[\mu = ^{238}\text{U}/^{204}\text{Pb}\]

\[W = ^{232}\text{Th}/^{204}\text{Pb}\]

\[137.88 = ^{238}\text{U}/^{235}\text{U}\]

\[\lambda_1, \lambda_5, \lambda_2 : \text{decay constants of} ^{238}\text{U}, ^{235}\text{U} \text{ and} ^{232}\text{Th respectively.}\]

\[t = \text{time when the Pb was isolated from its radioactive parents.}\]
Using equations (12) to (14) we can plot the growth of the Pb isotope ratios with time for given values of $\mu$ and $W$ (Fig. 6).

If the Pb was separated from the U and Th at time "t", the isotope ratios were frozen at the values ($^{206}\text{Pb}/^{204}\text{Pb}$)$_t$, ($^{207}\text{Pb}/^{204}\text{Pb}$)$_t$, ($^{208}\text{Pb}/^{204}\text{Pb}$)$_t$, and serve as fossil indicators of this event. A plot of $^{207}\text{Pb}/^{204}\text{Pb}$ against $^{206}\text{Pb}/^{204}\text{Pb}$ or $^{208}\text{Pb}/^{204}\text{Pb}$ against $^{206}\text{Pb}/^{204}\text{Pb}$ will define a series of growth curves, showing how the isotope ratios changed with time.

In the present-day ratio of $^{238}\text{U}/^{204}\text{Pb}$, $\mu$ may be eliminated between equations (12) and (13).

\[
\frac{\left(\frac{^{207}\text{Pb}}{^{204}\text{Pb}}\right)}{\left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)} - a_0 = \frac{1}{137.88} \times \frac{\left(e^{\text{i}5T} - e^{\text{i}5t}\right)}{\left(e^{\text{i}8T} - e^{\text{i}8t}\right)}
\]

(15)

On a graph of $^{206}\text{Pb}/^{204}\text{Pb}$ against $^{207}\text{Pb}/^{204}\text{Pb}$ this equation defines a straight line known as an isochron (Fig. 7).

\[
\left(\frac{^{207}\text{Pb}}{^{204}\text{Pb}}\right)_t = b_0 + \frac{1}{137.88} \times \frac{\left(e^{\text{i}5T} - e^{\text{i}5t}\right)}{\left(e^{\text{i}8T} - e^{\text{i}8t}\right)} \times \left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)_t - a_0
\]

(16)

All the leads that evolved between "T" and "t" in reservoirs with different $\mu$ values will lie on a $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ plot on a straight line passing through $P(a_0, b_0)$ (Fig. 7).
Fig. 6. Growth of lead isotope ratios with time (after Faure 1966).
Fig. 7. Graph of $^{206}\text{Pb}/^{204}\text{Pb}$ against $^{207}\text{Pb}/^{204}\text{Pb}$ (after Faure 1986).
Lead evolution models

There are two types of models used to evaluate the isotopic evolution of Pb. Single stage model postulates a continuous change in the isotope ratios from time "T". Two stage models assume Pb growth from time "T" to a subsequent time when there was a change in isotopic composition of the source region by mixing with Pb of different composition resulting from crustal evolution. There are also more complicated multistage models possible.

Holmes-Houtermans, single stage model

The Holmes-Houtermans model (1946) assumes that originally the Earth was fluid and homogeneous; at that time U, Th and Pb were uniformly distributed; the isotopic composition of the primeval Pb was everywhere the same. Subsequently the Earth became rigid, and small regional differences in the U/Pb ratio were developed and in any given region the U/Pb ratio changed only because of radioactive decay of U to Pb. At the time of formation of a common Pb mineral, the Pb was separated from U and Th and its isotopic composition has remained constant since that time. This will also apply to other minerals such as feldspars that do not contain U and Th. The equations (12) to (15) describing Holmes-Houtermans model can be used to determine the ages of common leads that have had single-stage histories. The model assumes that all samples of common Pb are mixtures of radiogenic Pb, which formed in closed source regions having differing values of P and W, with primeval Pb.
Cumming and Richards, single stage model

The Cumming and Richards model (1975) assumes that the values of $\mu$ and $W$ increase linearly with time:

$$\mu(t) = \mu(1 - e^t)$$

$$W(t) = W(1 - e^t)$$

$e' = 0.037 \times 10^{-9}/\text{yr}$.

$e = 0.050 \times 10^{-9}/\text{yr}$.

$T = 4.509 \text{ Ga}$

where $e'$ and $e$ are rate factors.

Cumming and Richards in their model force the growth curves through the data points of the Canyon Diablo lead and of the lead from Captains Flat (Australia), a stratiform Pb–Zn–Cu deposit. They choose the present day $\mu$ value to make a best fit for different pairs of "T" and $e$. The continuous linear increase of the $\mu$ and $W$ values can be related to continuous differentiation processes. According to this model, the Pb of lead deposits would be a mixtures of crustal and mantle, dominated by the former (Delaloye, 1979).

Stacey and Kramers, multistage model

The Stacey and Kramers model (1975) postulates that lead evolved between 4.57 Ga (T) and 3.70 Ga (T') in a reservoir having uniform $^{238}\text{U}/^{204}\text{Pb}$ ($\mu$) and $^{232}\text{Th}/^{204}\text{Pb}$ (W) ratios of 7.192 and 32.208, respectively. At time of 3.70 Ga, represented by point Q on the Pb evolution line (Fig. 8) the values of $\mu$ and $W$ were changed by geochemical differentiation to 9.735 and 36.837,
respectively. From 3.70 Ga to the Present reservoir remained unchanged. All leads that evolved in such a reservoir, and were removed from it at sometime in the past, must have isotope ratios that lie on the growth curve between points Q and P of Fig. 8. The time of separation from the reservoir can be calculated from the equation of the isochron.

For the Pb daughters of U, whose evolution is represented in Fig. 8., the equation is

$$\frac{^{207}\text{Pb} / ^{204}\text{Pb} - 12.998}{^{206}\text{Pb} / ^{204}\text{Pb} - 11.152} = \frac{1}{137.88} \times \left[ \frac{e^{4.5} - e^{4.3}}{e^{4.5} - e^{4.3}} \right]$$

(17)

where $T = 3.70$ Ga, is the starting time of the second stage of Pb evolution. The equation for Pb evolution in coordinates of $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ is

$$\frac{^{208}\text{Pb} / ^{204}\text{Pb} - 31.230}{^{206}\text{Pb} / ^{204}\text{Pb} - 11.152} = \frac{^{232}\text{Th}}{^{238}\text{U}} \times \left[ \frac{e^{4.5} - e^{4.3}}{e^{4.5} - e^{4.3}} \right]$$

(18)

If "t" is known from a solution of equation (17), then equation (18) can be used to calculate the $^{232}\text{Th}/^{238}\text{U}$ ratio (W/μ) of the second-stage reservoir.

The value of this ratio should be close to the value required by the Stacey and Kramer model which is 3.78. Equation (17) can be solved for "t" by iteration, Table 4 is an example of such iterative values and illustrates the slopes of isochrons (m) and corresponding model ages (t).

It may be used to date common leads that satisfy the following conditions: they evolved from
Fig. 8. Two-stage Pb evolution model of Stacey and Kramers (after Faure 1986).
primeval Pb between 4.57 and 3.70 Ga in a uniform reservoir having $\mu = 7.192$ and $W = 32.208$; at 3.70 Ga the $\mu$ and $W$ values of the reservoir were changed but remained constant after that until the Present time except for changes due to decay of U and Th to Pb; lead was removed from the reservoir at time "t" and was incorporated into a common Pb mineral; after separation from the reservoir, the isotopic composition of the Pb was not changed by mixing with other leads or by additions of radiogenic Pb formed by decay of U and Th.

Table 4. Slopes of isochrons (m) and corresponding model ages (t) of the Stacey and Kramers model (1975), based on equation (16).

<table>
<thead>
<tr>
<th>Age (t) Ga</th>
<th>Slope (m)</th>
<th>Age (t) Ga</th>
<th>Slope (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>0.34840</td>
<td>1.8</td>
<td>0.51781</td>
</tr>
<tr>
<td>0.2</td>
<td>0.36103</td>
<td>2.0</td>
<td>0.54764</td>
</tr>
<tr>
<td>0.4</td>
<td>0.37482</td>
<td>2.2</td>
<td>0.58079</td>
</tr>
<tr>
<td>0.6</td>
<td>0.38991</td>
<td>2.4</td>
<td>0.61769</td>
</tr>
<tr>
<td>0.8</td>
<td>0.40645</td>
<td>2.6</td>
<td>0.65884</td>
</tr>
<tr>
<td>1.0</td>
<td>0.42460</td>
<td>2.8</td>
<td>0.70482</td>
</tr>
<tr>
<td>1.2</td>
<td>0.44457</td>
<td>3.0</td>
<td>0.75630</td>
</tr>
<tr>
<td>1.4</td>
<td>0.46659</td>
<td>3.2</td>
<td>0.81404</td>
</tr>
<tr>
<td>1.6</td>
<td>0.49090</td>
<td>3.4</td>
<td>0.87892</td>
</tr>
</tbody>
</table>
Dating of K-feldspar by the common lead method

The feldspars of igneous and metamorphic rocks are enriched in Pb and depleted in U and Th. Pb concentrations may range from a few tens up to several hundreds ppm, whereas their U and Th concentrations are generally less than one ppm. Thus, the isotopic composition of Pb in feldspars remains unchanged for long periods of time and thus preserves a record of the Pb that was initially incorporated into igneous and metamorphic rocks. The isotopic composition of Pb in feldspars can therefore be used for dating by the common lead method and to derive information about the petrogenesis of the rocks. The interpretation of Pb in feldspars in terms of the common lead method of dating is subject to several assumptions, which are as follows:

1. Pb must have had a single-stage history including evolution in a closed U, Th, Pb system before its incorporation into the feldspar;
2. it must not be contaminated with other leads having different isotopic compositions during the formation, intrusion and crystallization of magma;
3. the feldspars must remain closed to the Pb and their Pb/U and Pb/Th ratios must be sufficiently large to make the isotopic composition invariant with time.

In practice these prerequisite conditions in many cases are not satisfied. A large number of case studies have shown that single-stage Pb model ages of feldspars are significantly lesser or greater than ages obtained by other methods. These differences typically imply the presence of excess radiogenic Pb that may have been accumulated during a multi-stage history or by contamination by radiogenic Pb into feldspar during episodes of thermal metamorphism following crystallization. The common lead technique, while related to U and Th constitutes a distinct specialty of isotope geology. There is extensive theory and mathematical models on this subject (eg. Doe 1970, Faure 1986) from which the above information is derived. The subject is also discussed by Jurkowski (1994).
CHAPTER VI

ANALYTICAL PROCEDURE

Sample preparation

Sixteen samples were collected in Lynn Lake area in Manitoba (Appendix A). Each sample consists of approximately 50 kg of fresh, unweathered rock. In the laboratory, the samples were crushed by a jaw crusher to about 0.5 to 1.0 cm fragments. Further size reduction to -70 Tyler mesh was done by using the Braun pulverizer. Preliminary separation of heavy minerals was done using the Wildfley table. After drying each sample was sieved through a -60 Tyler mesh screen to remove coarse material. Iron fillings from the pulverizer plates and highly magnetic mineral were removed by the Carpco magnetic separator. The nonmagnetic portion of the sample was processed by gravity separation in heavy liquids. Ethylene tetrabromide (SG = 2.96) was used to remove feldspars (saved for further analysis) and quartz and methylene iodide (SG = 3.1) was used to remove mafic minerals. The remaining concentrate contained zircons (SG = 4.7) and some other heavy minerals. The sample was then boiled in 50% HNO₃ to dissolve sulfide and any soluble minerals. The heavy minerals were then sieved through 70, 100, 200, and 325 nylon Tyler mesh screens and separated by magnetic susceptibility on the Frantz isodynamic magnetic separator. The different fractions (-70, -70+100, -100+200, -200+325, -325) were saved in small petri dishes for further separation. Abrasion of samples was done either before or after to the Frantz separation. Final separation of zircons was done by handpicking under binocular microscope, which achieves 99.9% purity of zircons. The clearest zircons were picked from the least magnetic fractions in an attempt to get the most concordant points on the concordia. Zircons were cleaned for about 4 hrs in 50% HNO₃ using ultrasound vibrations.

K-feldspars were separated from the same tetrabromide light fractions, saved above, by
another flotation using tetrabromo liquid having a SG of 2.72. The final separation was also done
by handpicking resulting in 99.8% pure concentrate.

Sample dissolution and ion-exchange

Reagents used in samples dissolution and separation were ultra-pure. Zircons were
dissolved in an ultra-pure bottle distilled acids (Mattison, 1972) using the hydrothermal technique
for zircons dissolution developed by Krough (1973) and Parrish (1987).

One to two milligrams of clean zircons was weighted out into each pre cleaned, self-
sealing Teflon pressure bombs. To dissolve the zircons 0.5 ml of concentrated HF and two drops
of concentrated HNO₃ was added to each bomb. The bombs were then sealed and encased in thin
plastic and thicker steel jackets and heated at 195°C for seven days. After that time bombs were
cooled down and disassembled. Samples were evaporated to dryness. A 0.5 ml of 6N HCl was
added to each bomb and the samples were heated at 190°C overnight. That process converted the
fluorides from HF dissolution, into chlorides which are more soluble. After this solution was split
into weighed aliquots. The first aliquot is spiked with a known quantity of a mixture of a $^{208}$Pb-
$^{235}$U tracer. This is called isotope dilution (ID). The second aliquot was left unspiked to measure
the isotopic composition (IC). Later both aliquots were dried down and were ready for ion-
exchange columns. In order to separate uranium and lead in the sample micro ion-exchange
columns were assembled each containing 0.5 ml of Dowex AG1-X8 200-400 mesh anion
exchange resin. Both aliquots were taken up in either 3N (ID) or 1N (IC) HCl and loaded into
the ion-exchange columns. Lead was eluted with 6N HCl into 5 ml Teflon beakers and uranium
in the spiked aliquot was eluded with water. A drop of 0.25N H₃PO₄ was added to each beaker
and the samples were evaporated to dryness on the hot plate. K-feldspar samples followed the
same procedure as described above for zircons. The feldspar dissolution was similar to that of the
zircons procedure. The feldspar fractions were cleaned using ultrasound with 25% HNO₃ acid. They were dissolved in pressure bombs with HF/HNO₃, converted to chloride form. The ion exchange separation was done using anion exchange resin. Due to the high content of alkali and alkali earth metals they were processed twice through the ion exchange columns. More detailed description of the analytical procedures is given in several theses, for example in Keller (1987).

Mass spectrometry

All U and Pb isotopic ratios were measured on a 10-inch, 90-degree Neir-type mass spectrometer at the University of Windsor, Earth Sciences. This is a triple-filament, solid source mass spectrometer capable of analyzing Rb, Sr, U, Pb, etc. Both Pb and U samples were run on the single center filament with an accelerating voltage of 3830 and 3010 volts for Pb and U, with magnetic field strength of 5228 gauss and 5192 gauss respectively. A vacuum of less than $1 \times 10^{-7}$ Torr was required for each run. Peak switching and data acquisition were done by an online computer.

Each of the dried samples was dissolved in 1 $\mu$L (0.001 ml) of 0.25N H₃PO₄. The outgassed rhenium filament was coated by silica gel, which acts as a binding agent for the sample. The whole procedure is described by Cameron et al. 1969. Mass fractionation corrections used on spiked and unspiked runs were 0.1% per a.m.u. (atomic mass unit). This mass fractionation/discrimination was established by separate measurements of Pb isotope ratios on NBS 981 common Pb standard. This standard was also used to monitor the mass spectrometer in the course of all measurements.

Determination of U-Pb ages requires three mass spectrometer runs on two aliquots of each sample (IC and ID). In the unspiked IC run isotope composition was measured as $^{208}\text{Pb}/^{206}\text{Pb}$, $^{207}\text{Pb}/^{206}\text{Pb}$, and $^{204}\text{Pb}/^{206}\text{Pb}$ ratios. The first two ratios were measured with the
background reading at mass 206.5 and the last ratio with the background readings taken at mass 203.5 and 204.5. Those three ratios give the isotopic composition of Pb in the analyzed zircon sample. The present day composition of U was constant and equal to 1/137.88 for $^{235}U/^{238}U$.

For the spiked isotope dilution (ID) runs U and Pb ratios were measured on same filaments. The Pb $^{206}Pb/^{206}Pb$ ratio was measured first with the background at mass 206.5. To measure U the filament temperature was increased and the $^{235}U/^{238}U$ ratio was measured with the background readings at mass 236.5. From these three runs, the IC Pb, the ID Pb and the ID U ratios the abundance of each isotope can be calculated.

The data reduction followed the procedure of Ludwig (1982). The statistical evaluation of results was done using the regression analysis of Ludwig (1997), which is specific to U-Pb concordia data. This regression provides the upper and lower concordia age intercepts and also provides two measures of goodness of fit: the mean square of weighted deviates (M.S.W.D.), which is a residual variance, and a probability of fit. If the probability of fit is $\geq 15\%$, it is model 1 age, i.e. within experimental error. But, if the probability of fit is $<15\%$ the analysis is outside the experimental errors and a model 2 age is calculated. The uncertainties reported with the ages are 95% confidence limits.

Equations used for the statistical testing and pooling of ages are given in Appendix B.
CHAPTER VII

RESULTS

In this study five zircon and eight feldspar samples were isotopically analysed. The location of the samples is shown on the geological map of the Lynn Lake area, in Fig. 3 and in the Table 5. The analytical data for the zircon and feldspar samples is reported in Table 6 and Table 7 and shown as concordia diagrams and Pb evolution diagram, Figs. 9-15. For convenience Tables 5-7 and Figs 9-15 are given at the end of this Chapter.

The rock units are dated as follows:

Eden Lake Granite (LL90-10)
Fraser Lake Gabbro (FL-11)
EL Gabbro (EL-1)
Fox Mine Tonalite (LL90-05)
EL Gabbro (LL90-03)
Combined EL-1 & LL90-03

Zircon samples

Eden Lake Granite (LL90-10)

The sample collected from this batholith is a K-feldspar porphyritic granite. There are two different types of zircons present in this sample. The first type are cloudy white and opaque crystals, elongated with 5:1 length/width ratio. The second type are orange-brown zircons Fe stained with 3:1 length/width ratio. All crystals are well developed, with only few showing fractures. No older cores are visible. Four fractions were selected (Table 6). Fraction A was abraded for 2 hrs and both types of zircons were selected. After abrasion ratio of orange to white
crystals remained unchanged, but white zircons fragmented. Fraction B is white in colour and zircons were hand picked and not abraded. Fraction C represents orange zircons which also were hand picked and not abraded. Fraction D was abraded for 15 minutes and hand picked. All fractions were boiled in nitric acid. Four analysed fractions (A, B, C, D) are collinear but discordant (Fig. 9). The data points are 6% to 22% discordant. The obtained age is 1871 ± 10 Ma and lower intercept is 644 ± 80 Ma. The minimum $^{207}\text{Pb} / ^{206}\text{Pb}$ age is 1853 Ma.

**Fraser Lake Gabbro (FL-11)**

This pluton has the same composition as the EL Gabbro but does not contain any known economic mineralization. Zircons are clear and transparent, and colourless to pinkish. Crystals are elongated and have 5:1 length/width ratio. Four analysed fractions (A, B1, CC, CX) are collinear and are 2% to 20% discordant (Fig. 10). Obtained age is 1870 ± 6.2 Ma and lower intercept is 303 ± 72 Ma. Minimum $^{207}\text{Pb} / ^{206}\text{Pb}$ age is 1865 Ma.

**EL Gabbro (EL-1)**

The EL plug is a composite intrusion which cuts quartz-feldspar porphyry flows and breccias 4 km southeast of the Lynn Lake. The intrusion has the form of pipe and is about 500 m in diameter on the surface (Pinsent, 1980). The EL plug contains a Ni-Cu ore body.

The zircons separated from this rock are clear and slightly yellow and pinkish in colour. Some zircon crystals contain possible cores or there are overgrowths; many are fragmented. Three fractions (A, B, C) were analysed, they are collinear, fraction C is concordant (Fig. 11). Two other fractions A, and B are 3% to 2% discordant respectively; fraction B plots above the concordia curve. The age of the sample is 1871.5 ± 3.9 Ma and the lower intercept is 271 ± 276 Ma. Minimum $^{207}\text{Pb} / ^{206}\text{Pb}$ age is 1874 Ma.
Fox Mine Tonalite (LL90-05)

The rock is medium-grained diorite to tonalite in composition. Separated zircons are transparent, short and rounded, and slightly pinkish in colour with 1:1 length/width ratio. Fe staining is present in some of the zircons. All fractions (Table 6) were boiled in 50% nitric acid and hand picked. Fractions B and C were strongly abraded. Four analysed fractions (B, C, D, E) are collinear and E fraction is concordant (Fig.12). Fractions B, C, and D are 3% to 20% discordant. Resulting age is $1831 \pm 3.7$ Ma with the lower intercept of $461 \pm 50$ Ma. Minimum $^{207}\text{Pb}/^{206}\text{Pb}$ age is 1830 Ma.

EL Gabbro (LL90-03)

This sample is from exactly the same location as sample EL-1, but was collected earlier. One population of zircons was found. They are pale yellow and pinkish in colour and very clear and transparent. Crystals are well developed with 2:1 length/width ratio. Many have prisms broken off, but whole zircons are long and narrow with defined terminations. Zircons were hand picked and pink crystals were removed. Two analysed fractions (A, B) are only 3-4% discordant (Fig.13). The indicated age is $1870 \pm 19$ Ma with the lower intercept of $268 \pm 648$ Ma. The minimum $^{207}\text{Pb}/^{206}\text{Pb}$ age is 1866 Ma. Combining the data for this sample with EL-1 yields an age of $1871.3 \pm 3.5$ Ma with the lower intercept of $291 \pm 183$ Ma (Fig.14).
Feldspar samples

Plagioclase and K-feldspars were separated from the following eight samples:

LL90-03 - plagioclase from Lynn Lake EL Gabbro;
LL90-04 - plagioclase from Fraser Lake Tonalite;
LL90-05 - plagioclase from Fox Mine Tonalite;
LL90-06 - plagioclase from Fox Mine Rhyolite;
LL90-07 - plagioclase from Lynn Lake A Gabbro;
LL90-08 - plagioclase from Fraser Lake Gabbro;
LL90-09 - K-feldspar from Cartwright Lake Rhyolite;
LL90-10 - K-feldspar from Eden Lake Granite.

The Pb isotopic composition is reported in Table 7 and represented as a plot of $^{206}\text{Pb}/^{204}\text{Pb}$ vs. $^{207}\text{Pb}/^{204}\text{Pb}$ in Figure 15. Three samples LL90-03, LL90-07, and LL90-08 plot above Pb evolution curve and are collinear within the experimental error. They define the secondary isochron with intercepts of 1945 Ma and 3852 Ma (Fig. 15). Samples 06 and 10 are rhyolites and their ages are anomalous. They contain J-type Pb which produces ages in the future. Samples 04, 05, 09 plot on or below the Pb growth curve and are mantle derived. The probability of fit is 56% and M.S.W.D. is 0.335. The Pb-Pb ages calculated for all the feldspar samples are reported in Table 7.
<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Description</th>
<th>Location (latitude/longitude)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EL-1</td>
<td>Gabbro EL Plug</td>
<td>56° 49' 15&quot; 101° 01' 55&quot;</td>
</tr>
<tr>
<td>LL90-03</td>
<td>Gabbro EL Plug</td>
<td>56° 49' 15&quot; 101° 01' 55&quot;</td>
</tr>
<tr>
<td>FL-11</td>
<td>Gabbro Fraser Lake</td>
<td>56° 47' 49&quot; 101° 05' 12&quot;</td>
</tr>
<tr>
<td>LL90-05</td>
<td>Tonalite Fox Mine</td>
<td>56° 38' 25&quot; 101° 38' 41&quot;</td>
</tr>
<tr>
<td>LL90-10</td>
<td>Granite Eden Lake</td>
<td>56° 41' 15&quot; 99° 57' 06&quot;</td>
</tr>
<tr>
<td>LL90-03</td>
<td>Gabbro EL Plug</td>
<td>56° 49' 15&quot; 101° 01' 55&quot;</td>
</tr>
<tr>
<td>LL90-04</td>
<td>Tonalite Fraser Lake</td>
<td>56° 48' 23&quot; 101° 06' 34&quot;</td>
</tr>
<tr>
<td>LL90-05</td>
<td>Tonalite Fox Mine</td>
<td>56° 38' 25&quot; 101° 38' 41&quot;</td>
</tr>
<tr>
<td>LL90-06</td>
<td>Rhyolite Fox Mine</td>
<td>56° 37' 59&quot; 101° 34' 41&quot;</td>
</tr>
<tr>
<td>LL90-07</td>
<td>Gabbro A Plug Lynn Lake</td>
<td>56° 51' 06&quot; 101° 02' 00&quot;</td>
</tr>
<tr>
<td>LL90-08</td>
<td>Gabbro Fraser Lake</td>
<td>56° 47' 49&quot; 101° 05' 12&quot;</td>
</tr>
<tr>
<td>LL90-09</td>
<td>Rhyolite Cartwright Lake</td>
<td>56° 47' 19&quot; 100° 38' 48&quot;</td>
</tr>
<tr>
<td>LL90-10</td>
<td>Granite Eden Lake</td>
<td>56° 41' 15&quot; 99° 57' 06&quot;</td>
</tr>
</tbody>
</table>
Table 6. Analytical data for zircons from Lynn Lake, Northwestern Manitoba

<table>
<thead>
<tr>
<th>Sample detail</th>
<th>Concentration (ppm)</th>
<th>Atomic ratios</th>
<th>Apparent ages (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>U</td>
<td>Pb</td>
<td>$^{206}$Pb/$^{206}$Pb</td>
</tr>
<tr>
<td>Eden Lake Granite (LL90-10)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A m1</td>
<td>114 2.3</td>
<td>618 215</td>
<td>0.00060</td>
</tr>
<tr>
<td>B m2</td>
<td>114 1.1</td>
<td>550 169</td>
<td>0.00053</td>
</tr>
<tr>
<td>C m2</td>
<td>114 0.9</td>
<td>778 246</td>
<td>0.00034</td>
</tr>
<tr>
<td>D m1</td>
<td>114 2.0</td>
<td>746 232</td>
<td>0.00024</td>
</tr>
<tr>
<td>Fraser Lake Gabbro (FL-11)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A m0</td>
<td>114 2.0</td>
<td>286 107</td>
<td>0.00085</td>
</tr>
<tr>
<td>B1 m4</td>
<td>114 1.3</td>
<td>472 169</td>
<td>0.00159</td>
</tr>
<tr>
<td>CC m0</td>
<td>62 8.0</td>
<td>350 107</td>
<td>0.00048</td>
</tr>
<tr>
<td>CX m0</td>
<td>62 2.7</td>
<td>451 156</td>
<td>0.00024</td>
</tr>
<tr>
<td>&quot;EL.&quot; Gabbro (EL-1)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A m0</td>
<td>114 0.3</td>
<td>76 42</td>
<td>0.01001</td>
</tr>
<tr>
<td>B m0</td>
<td>114 0.4</td>
<td>75 41</td>
<td>0.00785</td>
</tr>
<tr>
<td>C m0</td>
<td>114 0.2</td>
<td>122 46</td>
<td>0.00186</td>
</tr>
<tr>
<td>Fox Mine Tonalite (LL90-05)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B m0</td>
<td>62 1.4</td>
<td>367 135</td>
<td>0.00174</td>
</tr>
<tr>
<td>C m0</td>
<td>114 0.3</td>
<td>427 142</td>
<td>0.00108</td>
</tr>
<tr>
<td>D m0</td>
<td>62 3.6</td>
<td>494 163</td>
<td>0.00038</td>
</tr>
<tr>
<td>E m1</td>
<td>62 1.1</td>
<td>858 244</td>
<td>0.00036</td>
</tr>
<tr>
<td>&quot;EL.&quot; Gabbro (LL90-03)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>A m0</td>
<td>62 1.9</td>
<td>251 90</td>
<td>0.00011</td>
</tr>
<tr>
<td>B m0</td>
<td>114 1.1</td>
<td>225 82</td>
<td>0.00019</td>
</tr>
</tbody>
</table>
Table 6 (continued)

1Relative magnetic susceptibility of zircons is reported as m0 (nonmagnetic) to m4 (paramagnetic) and is related to the inclination of the Frantz isodynamic separator using maximum current of 2 A. Grain size is an average; sieves used were 85, 150, 75, 38 μm.

2Measured ratio.

3Blank corrected.

4Blank and nonradiogenic Pb corrected.

5Decay constants used: $\lambda^{238}_{\text{U}} = 1.55125 \times 10^{-10}$ year$^{-1}$; $\lambda^{235}_{\text{U}} = 9.8485 \times 10^{-10}$ year$^{-1}$ (Steiger and Jäger 1977).

---

Table 7. Analytical data for feldspars from Lynn Lake, Northwestern Manitoba

<table>
<thead>
<tr>
<th>Sample detail</th>
<th>$^{204}\text{Pb}/^{206}\text{Pb}$</th>
<th>$^{207}\text{Pb}/^{206}\text{Pb}$</th>
<th>$^{206}\text{Pb}/^{206}\text{Pb}$</th>
<th>$^{206}\text{Pb}/^{204}\text{Pb}$</th>
<th>$^{207}\text{Pb}/^{204}\text{Pb}$</th>
<th>$^{208}\text{Pb}/^{204}\text{Pb}$</th>
<th>Age (Ma)</th>
<th>$\mu$</th>
</tr>
</thead>
<tbody>
<tr>
<td>&quot;EL&quot; Gabbro (LL90-03)</td>
<td>0.05973</td>
<td>0.9594</td>
<td>2.1598</td>
<td>16.742</td>
<td>16.062</td>
<td>36.159</td>
<td>2003</td>
<td>13.6</td>
</tr>
<tr>
<td>Frazer Lake Tonalite (LL90-04)</td>
<td>0.06305</td>
<td>0.9663</td>
<td>2.2351</td>
<td>15.860</td>
<td>15.326</td>
<td>35.450</td>
<td>1628</td>
<td>9.6</td>
</tr>
<tr>
<td>Fox Mine Tonalite (LL90-05)</td>
<td>0.06141</td>
<td>0.9367</td>
<td>2.1877</td>
<td>16.284</td>
<td>15.253</td>
<td>35.624</td>
<td>1150</td>
<td>8.8</td>
</tr>
<tr>
<td>Fox Mine Tonalite (LL90-06)</td>
<td>0.05011</td>
<td>0.7907</td>
<td>1.8885</td>
<td>19.956</td>
<td>15.779</td>
<td>37.687</td>
<td>-609</td>
<td>10.2</td>
</tr>
<tr>
<td>&quot;A&quot; Gabbro (LL90-07)</td>
<td>0.06045</td>
<td>0.9659</td>
<td>2.2171</td>
<td>16.542</td>
<td>15.978</td>
<td>36.676</td>
<td>2033</td>
<td>13.3</td>
</tr>
<tr>
<td>Fraser Lake Gabbro (LL90-08)</td>
<td>0.05732</td>
<td>0.9457</td>
<td>2.1258</td>
<td>17.446</td>
<td>16.499</td>
<td>37.087</td>
<td>2054</td>
<td>15.7</td>
</tr>
<tr>
<td>Cartwright Lake Rhyolite (LL90-09)</td>
<td>0.05793</td>
<td>0.8742</td>
<td>2.0599</td>
<td>17.262</td>
<td>15.091</td>
<td>35.558</td>
<td>-99</td>
<td>7.7</td>
</tr>
<tr>
<td>Eden Lake Granite (LL90-10)</td>
<td>0.06459</td>
<td>0.9658</td>
<td>2.2148</td>
<td>15.482</td>
<td>14.952</td>
<td>34.290</td>
<td>1263</td>
<td>7.7</td>
</tr>
</tbody>
</table>
Fig. 9. Concordia diagram for Eden Lake Granite (LL90-10).
Fig. 10. Concordia diagram for Fraser Lake Gabbro (FL-11).
Fig. 13. Concordia diagram for EL Gabbro (LL90-03).
Fig. 15. Pb evolution diagram of $^{206}\text{Pb}/^{204}\text{Pb}$ versus $^{207}\text{Pb}/^{204}\text{Pb}$ for eight feldspars, showing Stacey and Kramers (1975) Pb growth curve.
CHAPTER VIII

DISCUSSION OF THE RESULTS

The six U-Pb zircon ages determined in this study together with the six zircon ages reported by Baldwin et al. (1987) (Table 8) provide a realistic set of reliable primary ages that define the chronostratigraphy of the Lynn Lake area. The problem in doing this study is that the rocks here are devoid of zircons. For this study seventeen samples were collected, zircons were found in only five of the samples. The same problem was encountered by Baldwin et al. 1987 who found zircons in four rock units (Zwanzig, per. comm., Zwanzig is one of the six authors of the Baldwin et al. 1987 paper). Moreover, the quantity of zircons found in the five samples dated for this study is very small (< 0.01 to 1 g from 50-70 kg samples). The sample from the Eden granite, was an exception; it yielded about 5 g of zircons.

U-Pb zircon ages

There are numerous gabbroic plugs and stocks in the Lynn Lake domain. In the Lynn Lake area there are three such bodies, "A" plug and "EL" plug which host Ni ore deposits and the Fraser Lake gabbro which is totally barren. Zircons from the Frazer Lake gabbro yield an age of 1870 ± 6.2 Ma. For the "EL" plug two samples were collected. Sample EL-1 yielded an age of 1871.5 ± 3.9 Ma, it is a very precise age with concordant data fractions. Sample LL90-03, collected prior to EL-1 from the same location, had very few zircons and only two points could be analysed hence the age of 1870 Ma has a large error of ± 19 Ma. These two data points
Table 8. Zircon ages reported in this study and Baldwin et al. (1987).

<table>
<thead>
<tr>
<th>Sample number and location</th>
<th>Age (Ma)</th>
<th>Lower Intercept (Ma)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Eden Lake Granite (LL90-10)</td>
<td>1871 ± 10</td>
<td>644 ± 80</td>
<td>this study</td>
</tr>
<tr>
<td>Fraser Lake Gabbro (FL-11)</td>
<td>1870 ± 6.2</td>
<td>303 ± 72</td>
<td>this study</td>
</tr>
<tr>
<td>Lynn Lake EL Gabbro (EL-1)</td>
<td>1871.5 ± 3.9</td>
<td>271 ± 276</td>
<td>this study</td>
</tr>
<tr>
<td>Fox Mine Tonalite (LL90-05)</td>
<td>1831 ± 3.7</td>
<td>461 ± 50</td>
<td>this study</td>
</tr>
<tr>
<td>Lynn Lake EL Gabbro (LL90-03)</td>
<td>1870 ± 19</td>
<td>268 ± 648</td>
<td>this study</td>
</tr>
<tr>
<td>Combined EL-1 &amp; LL90-03(1)</td>
<td>1871.3 ± 3.5</td>
<td>291 ± 183</td>
<td>this study</td>
</tr>
<tr>
<td>Lynn Lake Rhyolite (FL1)</td>
<td>1915 ± 6.5</td>
<td>524 ± 67</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td>Lynn Lake Rhyolite (FL2)</td>
<td>1910 ± 12.5</td>
<td>504 ± 264</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td>Hughes Lake Qtz Diorite (3002)</td>
<td>1876 ± 7.5</td>
<td>434 ± 154</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td>Norrie Lake Tonalite (3005)</td>
<td>1876 ± 7</td>
<td>396 ± 156</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td>Rusty Lake Rhyolite (R1)*</td>
<td>1878 ± 3</td>
<td>225 ± 149</td>
<td>Baldwin et al. (1987)</td>
</tr>
<tr>
<td>Rusty Lake Rhyolite (R4)*</td>
<td>1874 ± 7.5</td>
<td>-31 ± 98</td>
<td>Baldwin et al. (1987)</td>
</tr>
</tbody>
</table>

(1) samples EL1 and LL90-3 are from the same location

* samples are not shown on the Fig. 3, but Rusty Lake belt is shown on Fig. 2.
can be combined with the three points of sample EL-1 to yield a more precise age of 1871.3 ± 3.5 Ma. Clearly the "EL" gabbro and the Frazer Lake gabbro are of the same age. Therefore, there is no difference in age between the ore hosting and the barren gabbro. This is regarded as significant finding, it means that all gabbroic plugs in the Lynn Lake domain must be considered as potential hosts for Ni-Cu-Co mineralization. Statistically pooling (Appendix B) the three ages the average age is 1871.4 ± 2.8 Ma.

Zwanzig (per. comm.) postulated that the Lynn Lake gabbros should be 1876 Ma in age. This is the age obtained for the Norrie Lake tonalite (1876 ± 7 Ma), which they (Baldwin et al. 1987) considered as coeval with the gabbro. Statistically pooling the three gabbro ages the resulting average age is 1871.4 ± 2.8 Ma. This is not significantly different, at the 5% level, from the 1876 ± 7 Ma age for Norrie Lake tonalite. Nevertheless, the possibility that the gabbro is slightly younger than the tonalite and other similar age bodies should not be discarded.

The Eden Lake granite is a late high level intrusive and has been dated here at 1871 ± 10 Ma, again a very similar to age of the gabbros. Baldwin et al. (1987) were not able to date any granites for lack of zircons, they were able to measure an age for the Hughes Lake quartz diorite at 1876 ± 7.5 Ma.

Rhyolite samples collected for this study were totally devoid of zircons. Baldwin et al. (1987) were successful in finding zircons in two rhyolite samples from Lynn Lake which yielded ages of 1915 ± 6.5 Ma and 1910 ± 12.5 Ma. These are the oldest ages for the Wasekwan rocks in Lynn Lake area. At Rusty Lake (Fig. 2) Baldwin et al. (1987) obtained ages of 1878 ± 3 Ma and 1874 ± 7.5 Ma, clearly younger volcanism than at Lynn Lake.

Gordon et al. (1990) postulates that the Lynn Lake volcanic rocks developed first and were being eroded when the Rusty Lake volcanic arc was starting to form. Period of time
between 1871 - 1876 Ma is a plutonic event at Lynn Lake. Thus, when volcanism was building
the Rusty Lake greenstone belt, the Lynn Lake area was experiencing plutonic activity in the form
of gabbro and granite intrusions. A tonalite body at Fox Lake (sample LL90-05) yields an age of
1831 ± 3.7 Ma, which is the youngest U-Pb age of the pre-Sickle intrusive suite.

A chronostratigraphic sequence of rocks in the area is presented in Table 9.

Pb isotopic composition in feldspars

Isotopic composition of Pb in different rocks contains a record of the chemical
environments in which the Pb resided. These may include the mantle, crustal rocks, or Pb ores.
Each of these environments has a different U/Pb and Th/Pb ratios that affect the isotopic
composition of Pb. This composition may be changed both by decay of U and Th and by mixing
with Pb having different composition. As a result, the isotopic composition of Pb in rocks display
complex patterns of variations that reflect their geologic histories (Faure, 1986). Thus, the
common Pb method of dating may be used in determining ages, and more important, in the
solution of genetic problems (Doe, 1970).

The results of the Pb isotopic composition of the eight feldspars analysed in this studies
are plotted on a $^{206}\text{Pb}/^{204}\text{Pb}$ versus $^{207}\text{Pb}/^{204}\text{Pb}$ diagram Fig. 15, which also shows the Stacey and
Kramers (1975) growth curve for reference purposes. It is obvious from this plot that some
samples, those that plot above the evolution curve have a crustal signature, and those below the
curve have a mantle signature. Samples with normal Pb composition (single-stage leads) should
lie on the growth curve according to their ages. Anomalous leads are the results of the mixing of
a single-stage lead with radiogenic lead derived from minerals in the crust containing U and Th.
Three samples 03, 07, 08 which are Lynn Lake "EL", Lynn Lake "A", and Fraser Lake gabbros,
Table 9. Chronostratigraphy of Lynn Lake area, Manitoba.

<table>
<thead>
<tr>
<th>Group</th>
<th>Supracrustal Rocks</th>
<th>Intrusive Rocks</th>
<th>U/Pb Age</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Unit</td>
<td>Rock type</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Unit/Dated Rock</td>
<td>(Ma)</td>
</tr>
<tr>
<td>Post-Sickle Intrusives:</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>18-22</td>
<td>Diorite, tonalite, granite, granodiorite, pegmatite, porphyry</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sickle</td>
<td>Conglomerate with quartz-feldspar porphyry, sedimentary, volcanic and granitoid clasts, sandstone, derived shist and gneiss</td>
<td></td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>Conglomerate with sedimentary, volcanic, and granitoid clasts</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Unconformity</td>
<td></td>
</tr>
<tr>
<td>8-9</td>
<td>Conglomerate, sedimentary rocks, coarse- to fine-grained, paragneiss</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Wasekwan</td>
<td>Intermediate and felsic volcanic gneiss</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2-4</td>
<td>Aphyric basalt, porphyritic basalt, mafic and intermediate volcanic rocks, amphibolite</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Greywacke, mudstone, siltstone</td>
<td></td>
<td></td>
</tr>
<tr>
<td>17</td>
<td>Eden Lake granite</td>
<td>1871 ± 10</td>
<td></td>
</tr>
<tr>
<td>16</td>
<td>Fox Mine tonalite</td>
<td>1831 ± 3.7</td>
<td></td>
</tr>
<tr>
<td>16</td>
<td>Hughes Lake Qtz Diorite</td>
<td>1876 ± 7.5</td>
<td></td>
</tr>
<tr>
<td>16</td>
<td>Norrie Lake tonalite</td>
<td>1876 ± 7</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>Frazer Lake gabbro</td>
<td>1870 ± 6.2</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>EL gabbro</td>
<td>1871.5 ± 3.9</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>EL gabbro</td>
<td>1870 ± 19</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>Lynn Lake rhyolite</td>
<td>1915 ± 6.5</td>
<td></td>
</tr>
<tr>
<td>13-15</td>
<td>Gabbro, diorite, norite, diabase</td>
<td>1910 ± 12</td>
<td></td>
</tr>
</tbody>
</table>

* Note: at Rusty Lake, south-east from the study area, rhyolites are 1878 ± 3 and 1874 ± 7.5 Ma Baldwin et al. 1987
** Sources of the ages are reported in Table 7.
respectively, plot above the Pb evolution curve and are collinear within experimental error. They define secondary isochron with the upper intercept at 1945 Ma and the lower intercept at 3852 Ma. The upper intercept can be explained as the time at which gabbroic magma was separated from the source in the mantle. The lower intercept at 3852 Ma represents change in the source of the magma and generally agrees with the Stacey and Kramers (1975), two-stage model which postulates that the upper and lower mantle separated at 3700 Ma. Pb-Pb ages for these samples are 2003 Ma, 2033 Ma, and 2054 Ma, with the $\mu$ values of 13.6, 13.3, and 15.7 respectively. These $\mu$ values are much higher than the present $\mu$=9.735. The difference could be related to contamination of the gabbroic magma by crustal rocks which have high concentration of U and Th or from in situ decay of U and Th. These three gabbros were intruded about 1870 Ma ago (U-Pb zircon ages).

**Evolution of the Lynn Lake belt**

A possible sequence of events that formed the Lynn Lake belt is as follows.

1. At ca. 1945 Ma ago melting of the mantle produces basaltic magma which erupts on the surface forming early Wasekwan volcanic rocks. Portion of this melt remains in the mantle or lower crust as a gabbro.

2. At 1910 - 1915 Ma ago rhyolite volcanism occurs. Source of this magma is partial melting of a subducting ocean floor plate, though other sources are possible, as discussed below.

3. At 1871 Ma ago the gabbroic residue, produced before (point 1 above), and residing in the mantle or upper crust is remobilised and intruded to form the gabbro stocks. The sulphides disseminated and massive, are introduced at that time. The pyroxene crystals found in the massive sulphides register the age of the event at ca. 1945 Ma (point 1 above), they are survival ages. The
crustal signature of the gabbros is result of in situ accumulation of radiogenic Pb between 1945 and 1871 Ma.

4. At 1976 Ma ago tonalites and other granitoid rocks are introduced. The source of this magma is different than the gabbro magma. Note, there is no difference in age between 1871 Ma and 1876 Ma.

5. After 1871 Ma ago the above rocks are deformed and eroded.

6. Sickle metasedimentary rocks are deposited unconformably.

7. At 1835 Ma granitoid plutonism produces bodies such as the Fox Mine tonalite.

8. This 1835 Ma event is followed by regional metamorphism and granitoid plutons. This granitic formation probably terminates at 1780 Ma which is documented by Rb-Sr ages.

Lead isotopic composition of samples 06 and 09, Fox Mine and Cartwright Lake rhyolite, growth is typically called anomalous Pb or "Joplin" Pb. They have model ages of -609 Ma and -99 Ma respectively. These are future ages. Their U-Pb ages couldn’t be obtained because of the lack of zircons in the collected samples. However, there are two dated rhyolite (FL and FL) in the Lynn Lake area (Baldwin et al. 1987) and their ages are 1915 Ma and 1910 Ma respectively. The negative ages indicate that the rocks were probably contaminated by radiogenic lead from older crustal material. But, from where did that older crustal material come from?

There are three possible scenarios:

1. The Lynn Lake belt formed in the vicinity of a island arc settings next to an ocean - ocean subduction zone. Source of magma would be mantle and partial melting of the subducting plate. The subducting plate would probably be very old in order to produce the observed J-type Pb of the feldspar.

2. The Lynn Lake belt could have formed in a ridge or plume setting. The generation of early
basic magma presents no problems. Generation of rhyolite magma having a crustal signature is a problem. Partial melting of Wasekwan rocks, which are mafic volcanics and their degradation products would not have enough radiogenic Pb to produce the J-type Pb found in the feldspars. 3. Third possibility would be either island arc or ridge setting, but the rhyolite magma coming from a detached Archean plate in the mantle.

The remaining samples 04, 05, and 10 which are Fraser Lake, Fox Mine tonalitie, and Eden Lake granite plots on (04) or below (05, 10) the growth curve. Sample 04 has a normal isotopic composition and close to normal $\mu$ value of 9.6. Samples 05 and 10 have lower than normal $\mu$ values of 8.8 and 7.7 and are depleted in radiogenic lead. They are probably mantle derived. Their Pb-Pb ages of 1628 Ma, 1150 Ma, and 1263 Ma are lower than calculated U-Pb zircon ages. Since the $\mu$ values for these rocks are so different from the $\mu$ values for the gabbros it is obvious that the source of magmas for these rocks is different than the source of magma for the gabbros. Thus, these rocks are coeval but are not cogenetic. The Lynn Lake and Fraser Lake gabbros have the same age, at the 95\% level, as the Eden Lake granite (1871 Ma), Hughes Lake quartz diorite (1876 Ma), and Norrie Lake tonalite (1876 Ma). This rocks are, however, older that the Fox Mine tonalite (1831 Ma). These two sets of ages means that there were two plutonic events in the Lynn Lake area at 1871-1876 Ma and 1831 Ma. Both of these events are supported by Rb-Sr and K-Ar ages as discussed below.

Gordon et al. (1990) postulated an evolutionary scenario for the Lynn Lake and adjacent belts as shown in Fig. 16. Essentially the Lynn Lake belt formed first followed by Rusty Lake belt. But, he shows the Lynn Lake as forming in a rift setting, which is not likely because of the crustal signature of the rhyolites. The general model of subduction zone operation from (Fig. 17) is very close to what is proposed here.
Fig. 16. Evolution of Lynn Lake and adjacent belts according to Gordon et al. 1990.
Fig. 17. General model of subduction zone operation (from Kearey and Vine 1992).
Regional geochronology

In addition to the nine zircon ages discussed in this chapter there are also published and reported K-Ar, Rb-Sr, U-Pb, and Pb ages for the Lynn Lake area, and are listed in Table 2.

Fig. 18 is a plot of all the ages, including the ages determined in this study. This is a kind of frequency plot, and illustrates geological events. The ages are arranged in consecutive order. Each age has a frequency of one. Rocks with the same age will plot vertically above each other.

Fig. 18 is based on a population of forty ages, 13 K-Ar ages, 13 Rb-Sr ages, 12 U-Pb ages, and 2 Pb-Pb ages. These ages were assembled for this plot to show the continuum of igneous activity in the area. It is possible to analyse these ages as several populations which are labelled A, B, C, D in Fig. 18.

Population A contains 11 K-Ar ages which are from 1610 Ma to 1750 Ma. They may represent younger metamorphic event such as uplift and intrusion of dykes. Population B comprises three Rb-Sr ages, one K-Ar age, and one Pb-Pb age. They also represent probably plutonic and metamorphic events. Some of the Rb-Sr ages may be primary. Population C contains 12 U-Pb ages, 8 Rb-Sr ages, and one Pb-Pb age. These ages are primary ages and they represent main igneous rock forming period. Three events can be recognized: first at around 1830 Ma (Fox Mine tonalite is 1831 Ma old), second at around 1870 Ma (Lynn Lake and Fraser Lake gabbros are 1871 Ma old), and third at around 1910 Ma (Lynn Lake rhyolite is 1910 Ma old). Population D contains two Rb-Sr ages and one K-Ar age. First two ages probably belong to Population C because of their large uncertainty in age (± 75 Ma and ± 99 Ma respectively). Remaining K-Ar age (2045 Ma) is very interesting, because it is older than the rest of the ages and also because it was reported as the minimum age of the mineralization in Lynn Lake area. This is an age of a clinopyroxene from the massive sulphides of the "EL" plug (Turek, 1967).
Fig. 18. Frequency plot of isotopic ages in the Lynn Lake area. The ages are given in Table 2, plus one being reported in this study. Several distinct populations (A, B, C, D) are identified and discussed in text.
The pyroxene is cogenetic and coeval with massive sulphides and occurs as a large (1 to 5 cm) euhedral crystals in the massive sulphides. The sulphides are coeval and cogenetic with the gabbros, which are 1871 Ma. This K-Ar may be a survival age, the time the pyroxene and sulphides formed at depth but were then remobilised and together with the gabbroic melt were emplaced at present crustal level. Alternately the K-Ar age may be wrong due to presence of excess argon. While this is a possibility, it is difficult to accept as the pyroxene has normal level of K, and excess Ar is not normal in such situations.

Here is a possible scenario of the formation of the massive sulphide deposits and the emplacement of the gabbros at Lynn Lake area. At about 2 Ga melting in the upper mantle produced basaltic magma which erupts on the surface to produce the pillowed basalts of the Wasekwan series. Some of this magma remains in the mantle as semiliquid or solid. The Pb-Pb age of 1945 Ma obtained for the feldspars dates this separation from the mantle rock. At 1871 Ma this chamber was reactivated and the basaltic magma was emplaced as a gabbro in the crustal rocks formed at 2 Ga. The pyroxene present in the massive sulphides dates the crystallization of that mineral in the source region of the mantle, and not the emplacement of the gabbro and the sulphides.
CHAPTER IX
SUMMARY AND CONCLUSIONS

Isotopic ratios were measured on five zircon and eight feldspar samples separated from pre-Sickle intrusive rocks of the Lynn Lake greenstone belt, northwestern Manitoba.

U-Pb zircon ages determined for Lynn Lake and Fraser Lake gabbros (1871.3 ± 3.5 Ma and 1870 ± 6.2 Ma respectively) proved that both rocks have the same age within their error limits. The Lynn Lake gabbros host Ni-Cu-Co deposits while Fraser Lake gabbro is barren. This finding means that all gabbroic plugs in the Lynn Lake belt are possible hosts for Ni-Cu-Co mineralization.

A very similar age of 1871 ± 10 Ma was obtained for the Eden Lake K-feldspar granite. These ages indicate intense plutonic activity in this area at that time. The youngest rock dated was Fox Mine tonalite which has an age of 1831 ± 3.7 Ma. These ages together with 4 zircon ages published by Baldwin et al. (1987) established the geochronology of the Lynn Lake area.

Pb isotopic composition of feldspars reveal genetic history of the analysed rock units. Three gabbro samples ("A" plug, "EL" plug, Fraser Lake) define a secondary isochron which gives an age of 1945 Ma. This age was interpreted as an age of separation in the source region of the mantle.

Compilation of all available isotopic ages for the Lynn Lake area, a total of 40 ages, reveal a protracted igneous/metamorphic history for the region. The youngest group of ages is 1610 - 1750 Ma, these are 13 K-Ar ages and probably mark the wayning stages of the Hudsonian orogeny. A second group is 1750 - 1800 Ma, these are whole rock Rb-Sr, K-Ar, and Pb-Pb model ages and represent a plutonic and metamorphic event. A third group of ages is a major
rock forming event that is easily divisible into three subevents. First at ca. 1830 Ma was a
plutonic event defined by mixture of Rb-Sr, U-Pb, and Pb-Pb ages. Prior to that at ca. 1870 Ma a
group of 8 U-Pb zircon ages and one Rb-Sr age indicate a period of plutonic activity, time of
emplacement of many granites, tonalites and gabbros. A third subevent is defined by group of
ages indicating a felsic extrusive activity at ca. 1910 Ma. These ages would mark the end of
Wasekwan volcanism in the Lynn Lake area. The source of magma for this 1910 Ma felsic
volcanism is problematic on the basis of the Pb isotopic composition of feldspars a crustal source
for this magma is required. It could be generated by partial melting of Wasekwan volcanic and
sedimentary rocks. But, it would be more efficiently generated by melting of older Archean crust
except that the consensus of opinion is that there are no Archean rocks in the Lynn Lake area.
However, one could envisage a detached subducted plate in the mantle as a possible source.
REFERENCES


## APPENDIX A

All collected samples description and location.

<table>
<thead>
<tr>
<th>Sample No.</th>
<th>Rock unit</th>
<th>Rock description</th>
<th>Location (Latitude/longitude)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LL90-01</td>
<td>Cartwright Lake Rhyolite</td>
<td>Very fine-grained rhyolite, massive with minor plagioclase</td>
<td>56° 45' 58&quot; 100° 22' 45&quot;</td>
</tr>
<tr>
<td></td>
<td>(eastern zone)</td>
<td>phenocrysts. Interbedded with intermediate to felsic</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Porphyritic flows.</td>
<td></td>
</tr>
<tr>
<td>LL90-02</td>
<td>Lynn Lake A Plug Gabbro</td>
<td>Hornblende-quartz gabbro with minor quartz infill on joint</td>
<td>56° 51' 09&quot; 101° 02' 12&quot;</td>
</tr>
<tr>
<td>LL90-03</td>
<td>Lynn Lake EL Plug Gabbro</td>
<td>Medium to coarse-grained hornblende gabbro with minor</td>
<td>56° 49' 14&quot; 101° 01' 55&quot;</td>
</tr>
<tr>
<td></td>
<td></td>
<td>quartz.</td>
<td></td>
</tr>
<tr>
<td>LL90-04</td>
<td>Fraser Lake Tonalite</td>
<td>Foliated tonalite to granodiorite. Weathered surface</td>
<td>56° 48' 23&quot; 101° 06' 34&quot;</td>
</tr>
<tr>
<td></td>
<td></td>
<td>rusty buff to pink, fresh surface pinkish to white.</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Foliation define by 3-5% hornblende.</td>
<td></td>
</tr>
<tr>
<td>LL90-05</td>
<td>Fox Mine Tonalite</td>
<td>Medium-grained quartz diorite to tonalite. Moderately</td>
<td>56° 38' 25&quot; 101° 38' 41&quot;</td>
</tr>
<tr>
<td></td>
<td></td>
<td>foliated, defined by hornblende. Minor hematized quartz-</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>healed fractures.</td>
<td></td>
</tr>
<tr>
<td>LL90-06</td>
<td>Fox Mine Rhyolite</td>
<td>dacite flow. Very siliceous, interbedded with andesitic</td>
<td>56° 37' 59&quot; 101° 34' 41&quot;</td>
</tr>
<tr>
<td></td>
<td></td>
<td>to basaltic porphyritic flows. Biotitic, very few</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>fractures.</td>
<td></td>
</tr>
<tr>
<td>LL90-06a</td>
<td>Fox Mine Rhyolite</td>
<td>The same sample as above.</td>
<td>56° 37' 59&quot; 101° 34' 41&quot;</td>
</tr>
<tr>
<td>Location</td>
<td>Description</td>
<td>Coordinates</td>
<td></td>
</tr>
<tr>
<td>------------</td>
<td>-----------------------------------------------------------------------------</td>
<td>----------------------</td>
<td></td>
</tr>
<tr>
<td>LL90-07</td>
<td>Lynn Lake A Plug Gabbro. Medium to coarse-grained hornblende gabbro. Moderately foliated, well jointed. Minor quartz, and minor to trace pyrite.</td>
<td>56° 51' 05&quot; N, 101° 02' 00&quot; W</td>
<td></td>
</tr>
<tr>
<td>LL90-08</td>
<td>Fraser Lake Gabbro. Fine to medium-grained quartz-hornblende gabbro. Well jointed, minor quartz-calcite as fracture filling.</td>
<td>56° 47' 49&quot; N, 101° 05' 12&quot; W</td>
<td></td>
</tr>
<tr>
<td>LL90-09</td>
<td>Cartwright Lake Rhyolite (western zone). Sheared rhyolite, slightly fractured. Quartz and feldspar crystals 2-3 mm in diameter present. Weathered surface buff to white, fresh surface purplish brown.</td>
<td>56° 47' 19&quot; N, 100° 38' 48&quot; W</td>
<td></td>
</tr>
<tr>
<td>LL90-10</td>
<td>Eden Lake Granite. K-feldspar porphyritic granite. Massive blocky jointing. Quartz and feldspars 2-5 cm in size, 5-8% hornblende.</td>
<td>56° 41' 15&quot; N, 99° 57' 06&quot; W</td>
<td></td>
</tr>
<tr>
<td>FL-11</td>
<td>Fraser Lake Gabbro. Medium-grained quartz-hornblende gabbro. Some quartz present in the fractures.</td>
<td>56° 47' 49&quot; N, 101° 05' 12&quot; W</td>
<td></td>
</tr>
<tr>
<td>FL-12</td>
<td>Fraser Lake Gabbro. The same sample as above.</td>
<td>56° 47' 49&quot; N, 101° 05' 12&quot; W</td>
<td></td>
</tr>
<tr>
<td>LL-A</td>
<td>Lynn Lake A Plug Gabbro. Medium to coarse-grained hornblende gabbro. Minor quartz and pyrite.</td>
<td>56° 51' 09&quot; N, 101° 02' 12&quot; W</td>
<td></td>
</tr>
<tr>
<td>LL-A2</td>
<td>Lynn Lake A Plug Diorite. Medium to coarse-grained, quartz-hornblende diorite. The rock is grey and brownish grey on weathered surfaces.</td>
<td>56° 51' 09&quot; N, 101° 02' 12&quot; W</td>
<td></td>
</tr>
<tr>
<td>EL-1</td>
<td>Lynn Lake EL Plug Gabbro. Medium to coarse-grained hornblende gabbro with minor quartz.</td>
<td>56° 49' 14&quot; N, 101° 01' 55&quot; W</td>
<td></td>
</tr>
<tr>
<td>EL-A</td>
<td>Lynn Lake EL Plug Gabbro. The same sample as above.</td>
<td>56° 49' 14&quot; N, 101° 01' 55&quot; W</td>
<td></td>
</tr>
</tbody>
</table>
APPENDIX B

Testing and pooling of ages.

Formula for testing if ages are statistically different is:

\[
t_{\text{calc}} = \frac{|B_1 - B_2|}{\sqrt{[(\text{SE}_{B_1})^2 + (\text{SE}_{B_2})^2]}}
\]

where, \( B_1 \) and \( B_2 \) are the ages (or slopes for Rb-Sr isochrones); SE (standard error) are the SE of the age; \( t_{\text{calc}} \) is compared with \( t_{\text{table}} \) for the appropriate level of significance, usually 5%.

The degrees of freedom for above \( t_{\text{calc}} \) is \( K \) calculated from the following equation:

\[
\frac{[\text{SE}_{B_1}]^2 + [\text{SE}_{B_2}]^2}{K} = \frac{\text{SE}_{B_1}^4}{N_1 - 2} + \frac{\text{SE}_{B_2}^4}{N_2 - 2}
\]

where, \( N_2 \) are the degrees of freedom of the two ages.

The pooling of ages follows this procedure.

<table>
<thead>
<tr>
<th>Age (or slope)</th>
<th>SE</th>
<th>Weighted factor</th>
<th>Weighted B</th>
</tr>
</thead>
<tbody>
<tr>
<td>( B_1 )</td>
<td>( \text{SE}_{B_1} )</td>
<td>( W_1 = 1/(\text{SE}_{B_1})^2 )</td>
<td>( W_1 \times B_1 )</td>
</tr>
<tr>
<td>( B_2 )</td>
<td>( \text{SE}_{B_2} )</td>
<td>( W_2 = 1/(\text{SE}_{B_2})^2 )</td>
<td>( W_2 \times B_2 )</td>
</tr>
<tr>
<td>( B_3 )</td>
<td>( \text{SE}_{B_3} )</td>
<td>( W_3 = 1/(\text{SE}_{B_3})^2 )</td>
<td>( W_3 \times B_3 )</td>
</tr>
</tbody>
</table>

\[ T_1 \]

\[ T_2 \]

\[ \text{Mean } B = \frac{T_2}{T_1} \]

\[ \text{SE Mean } B = \left[ \frac{1}{T_1} \right]^{1/2} \]
The approximate degrees of freedom $K$, of the pooled means ($B_i$) are given by the equation:

$$\frac{B^2}{K} = \frac{B_1^2 (W_1/T_1)}{N_1 - 2} + \frac{B_2^2 (W_2/T_2)}{N_2 - 2} + \frac{B_3^2 (W_3/T_3)}{N_3 - 2}$$

The general equation is

$$K = \frac{(B \sum W_i)^2}{\sum (BW_i)^2 / N_i - 2}$$

The calculation is identical for averaging ages from different isochrons.
VITA AUCTORIS

Jacek Jurkowski was born in 1960 in Cracow, Poland. He graduated from Geological Technical School in 1980. From there he went to the University of Mining and Metallurgy where he started to study geology. In 1989 he moved to Canada and in 1994 obtained a B. Sc. degree in Geology from University of Windsor. In 1999 he obtained Degree of Master of Science in Geology.