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DEPOSITIONAL ENVIRONMENT OF THE SHERRIDON GROUP
AND RELATED MINERAL DEPOSITS NEAR SHERRIDON, MANITOBA

by

PETER ANDREW GOETZ, B.A., B.Sc.

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

Carleton University

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April, 1980

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ABSTRACT

At Sherridon, Manitoba, the Sherridon Group of the early Proterozoic Kisseynew sedimentary gneiss belt consists of biotite gneisses interpreted to be lithic arenites deposited in a deep to shallow water environment. Near the base of the Sherridon Group these arenites have the characteristics of turbidite deposits whereas in the upper part of the Group they have a lower mafic content, are thickly bedded and are intercalated with thick beds of marble.

Layers of amphibolite found in the lower part of the Sherridon Group have been recognized as meta-basalts, chemically similar to the low-K tholeiites of modern volcanic arcs. It is possible to distinguish both flows and volcanoclastic deposits. Felsic volcanic rocks, which were not previously recognized in the Kisseynew belt, occur near Singsing Lake and have calc-alkalic composition.

Three types of sulphide mineralization were recognized in the area: 1) volcanogenic distal massive sulphides, 2) sulphide facies volcanogenic iron formation and 3) Cu-Fe disseminations in pelitic rocks. It is concluded that these volcanogenic massive sulphides formed under a column of water 1.5 km deep, from brines carrying several hundred ppm of base metals and sulphur at a temperature of about 300°C. Some deposits overlie cordierite-anthophyllite rocks interpreted as metamorphosed, hydrothermally altered debris mechanically transported by the brines.

The area has undergone three periods of folding and one of faulting. Metamorphism which accompanied the deformation reached a peak of about 700°C and 6 kb between the first and second periods of folding.

A modern analogue of the depositional environment of the Sherridon Group might be the margin of a modern volcanic arc, particularly those of the South Pacific. The Osugi greenstone belt of Japan contains mineralization (the Besshi deposits) and an assemblage of sedimentary and volcanic rocks similar to the protoliths of sulphide deposits and enclosing rocks of the Sherridon Group.

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The Hudson Bay Exploration and Development Company also allowed the examination of their files on the geology of the Sherridon area. Comments by G. Bragg provided useful guidance in the examination of this material. D. Price contributed helpful discussion of the Flin Flon belt and collected samples from that area for this thesis.

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Chemical analyses were performed by the Geological Survey of Canada, Analytical Chemistry Section. Thin sections were provided by the Geological Survey of Canada and by R. Conlon and R. Taylor of Carleton University. X-ray identification of minerals was done by J. Baker and microprobe analyses were done by P. Morton. Special thanks are due L. J. Goetz for typing this thesis.

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TABLE OF CONTENTS

	<u>Page</u>
LIST OF TABLES	xi
LIST OF FIGURES	xiii
LIST OF PLATES	xvii
LIST OF APPENDICES	xviii
SYMBOLS	xix
MINERAL ABBREVIATIONS	xxi

Chapter

INTRODUCTION	1
I GEOLOGICAL SETTING OF THE SHERRIDON AREA	3
I-1 Geological Setting	3
I-2 Geochronology	7
I-3 Pole Position	12
I-4 Previous Work	14
I-5 Logistics	17
II LITHOLOGY, STRATIGRAPHY AND GEOCHEMISTRY	18
II-1 Introduction and Methods of Study	18
II-2 Local Setting and Stratigraphy	20
II-3 Nokomis Group (Units 1-4)	25
II-4 Identification of Sedimentary and Volcanic Rock in the Sherridon Group	28
II-5 Sedimentary Rocks of the Sherridon Group	45
II-5-a Introduction	45
II-5-b Biotite Gneiss (Lithic Arenite): Unit 5	46

TABLE OF CONTENTS cont'd

<u>Chapter</u>		<u>Page</u>
II-5-c	Quartzofeldspathic Gneiss (Arenite): Unit 6	52
II-5-d	Quartzitic Gneiss (Chert): Unit 7	56
II-5-e	Biotite-Garnet Schist (Graywacke): Unit 8	57
II-5-f	Meta-Pelite: Unit 9	58
II-5-g	Calc-Silicate Gneiss (Calcareous Rocks): Unit 11	59
II-5-h	Marble (Arenaceous Limestone): Unit 12	60
II-6	Volcanic Rocks in the Sherridon Group	61
II-6-a	Introduction	61
II-6-b	Amphibolites (Mafic Volcaniclastic and Flow Rocks): Units 13 and 14	62
II-6-c	Meta-Felsic Volcanics: Unit 15	76
II-7	Intrusive Rocks of the Sherridon Group	79
II-7-a	Introduction	79
II-7-b	Amphibolite (Meta-Gabbro): Unit 17	80
II-7-c	Meta-Granitic Rocks: Unit 18	83
II-7-d	Pegmatite: Unit 19	85
II-8	Altered Rocks in the Sherridon Group	86
II-8-a	Introduction	86
II-8-b	Rocks Altered by Non-hydrothermal Processes	86
II-8-c	Rocks Formed Through Hydrothermal Processes	90
III	ECONOMIC GEOLOGY	94
III-1	Introduction	94
III-2	Cherty Sulphide Horizons	94
III-3	Sulphide Disseminations in Pelites	97

TABLE OF CONTENTS cont'd

<u>Chapter</u>		<u>Page</u>
III-4	Subeconomic Massive Sulphide Deposits . . .	98
III-5	The Sherritt Gordon Deposit	103
III-5-a	General Geology	103
III-5-b	Sulphide Mineralogy	109
III-5-c	Metal Zoning	112
III-5-d	Interpretation and Comparison With Deposits of the Region	115
III-6	Summary of Field And Analytical Data Bearing on the Origin of the Sherritt Gordon Deposit	118
III-7	Construction of Solubility Diagrams . . .	119
III-7-a	Introduction	119
III-7-b	Calculation and Representation of the Fe-S-O Mineral Equilibria	119
III-7-c	Calculation and Representation of the Solubility Data	125
III-8	Application of Solubility Diagrams to the Deposits of the Sherridon Area	135
IV	STRUCTURE AND METAMORPHISM	150
IV-1	Structure: Introduction	150
IV-2	Description of Deformational Events . . .	150
IV-2-1	D ₁ : Formation of Regional Schistosity . .	150
IV-2-a	D ₂ : E-W Folding	153
IV-2-c	D ₃ : NNW-SSE Folding	154
IV-2-4	D ₄ : Faulting	156
IV-3	Strain	156
IV-4	Foliation	157

TABLE OF CONTENTS cont'd

<u>Chapter</u>		<u>Page</u>
IV-5	Lineation	158
IV-6	Metamorphism: Introduction	158
IV-7	Mineral Assemblages	159
IV-8	Metamorphic Conditions	163
V	INTERPRETATION AND CONCLUSIONS	170
V-1	Introduction	170
V-2	Depositional History of the Sherridon Group	170
V-3	Origin and Development of the Kisseynew Belt	179
	REFERENCES	184
	APPENDICES	
A	CHEMICAL ANALYSES OF ROCKS FROM SHERRIDON, MANITOBA	195
B	CHEMICAL ANALYSES OF ROCKS FROM A REGOLITH NEAR FLIN FLON, MANITOBA	226
C	MINERAL ASSEMBLAGES AND MODES FOR THIN SECTIONS OF ROCKS FROM THE SHERRIDON AREA, MANITOBA	229
D	MINERAL ASSEMBLAGES AND MODES FOR POLISHED THIN SECTIONS OF ROCKS FROM SULPHIDE OCCURRENCES AT SHERRIDON, MANITOBA	239
E	C.I.P.W. NORMS CALCULATED FOR THE IGNEOUS ROCKS OF APPENDIX A	242

LIST OF TABLES

<u>Table</u>		<u>Page</u>
I-1	Stratigraphic relationships of Kiseynew and adjacent belts	5
I-2	Age determinations of the Kiseynew and adjacent volcanic belts	8
II-1	Table of lithologies	22
II-2	Analysis of sedimentary and volcanic rocks bearing on the discrimination of ortho-gneiss and para-gneiss, Sherridon, Manitoba	31
II-3	Eigenvalues and eigenvectors from principal component analysis of major oxides of amphibolites	41
II-4	Average chemical composition of sedimentary rocks of the Sherridon area and vicinity	51
II-5	Average composition of some volcanic rocks at Sherridon with related rocks	74
II-6	Average chemical analyses of intrusive and related rocks, Sherridon	84
II-7	Average composition of altered rocks, Sherridon, and analyses bearing on their origin	89
III-1	Selected data for mines and sulphide occurrences from Central Manitoba	99
III-2	Equilibrium constants for reactions involving aqueous sulphurs species	121
III-3	Equilibrium conditions for volcanogenic fluids associated with massive sulphide formation and for seawater	123
III-4	Boundary conditions for volcanogenic fluids associated with massive sulphide formation and for seawater	126
III-5	Dissociation constants for metal associated chloride complexes	129
III-6	Equilibrium constants for aqueous reactions involving metallic sulphides and oxides	133

List of Tables cont'd

<u>Table</u>		<u>Page</u>
IV-1	Structure and metamorphism of the Sherridon area	151
IV-2	Compatible mineral assemblages	161
IV-3	Sphalerite Analyses	167
V-1	Structure of the Kisseynew gneisses	180

LIST OF FIGURES

Figure		Page
I-1	The location of Sherridon in relationship to the geology of northern Manitoba and Saskatchewan	4
I-2	Latitude maps of Laurentia for the interval -2300 to -1500 Ma. (from Roy and Lapointe, 1976)	13
II-1	Geology, Sherridon Area, Manitoba, Scale - 1:25,000	in pocket
II-2	Geology of Sherridon, Manitoba, Scale - 1" = 660'	in pocket
II-3	Stratigraphic column for the Sherridon Group	21
II-4	Chemistry of Unit 5, biotite gneiss, Sherridon, Man., plotted after methods of Van de Kamp et al, 1979. a) Niggli g vs. Niggli si b) Niggli si vs. ppm Zr	30
II-5	Chemistry of some amphibolites near Sherridon, Man. a) Niggli mg vs. Niggli c b) CaO vs. MgO	34
II-6	Plot of Sherridon amphibolites on "Igneous Spectrum Diagram" after Hughes (1973)	35
II-7	Ternary plots of some trace elements from amphibolites near Sherridon, Man. a) Co-Ni-Cr b) Co-Cu-Cr	37
II-8	Cluster analysis of amphibolites using major oxides	38
II-9	Cluster analysis of amphibolites using major oxides and minor elements	40
II-10	Principal component scores and clustering for amphibolites from Sherridon, Man. a) F_1 vs. F_2 ; b) F_1 vs. F_3 ; c) F_1 vs. F_4 ; d) F_1 vs. F_5	43

List of Figures cont'd

Figure		Page
II-11	Chemical composition of meta-sandstones in the Sherridon area; method of plot	53
II-12	Chemistry of some meta-sediments from Sherridon, Man., diagrams	54
II-13	Plot of Sherridon meta-volcanic rocks on: a) SiO_2 vs. $\text{Na}_2\text{O} + \text{K}_2\text{O}$ and b) SiO_2 vs. Al_2O_3 Diagrams	68
II-14	Normative Plagioclase vs. Normative Colour Index Diagrams; a) volcanic rocks b) plutonic rocks	69
II-15	Plot of discriminant functions	70
II-16	AFM Diagram	71
II-17	Chemical composition of Sherridon volcanic rocks compared with some recent trends plotted by Miyashiro (1974) a) SiO_2 vs. FeO/MgO ; b) FeO vs. FeO c) TiO_2 vs. FeO/MgO	72
III-1	Plot of Cu vs. Zn for sulphide occurrences in central Manitoba	100
III-2	Ore Tonnage vs. $\text{Zn/Zn} + \text{Cu}$ for sulphide occurrences in central Manitoba	101
III-3	Total base metal content vs. $\text{Zn/(Zn} + \text{Cu)}$ for sulphide occurrences in central Manitoba	102
III-4	Geology and level plans, Sherritt Gordon Mines, Sherridon, Manitoba	in pocket
III-5	Simplified section of the eastend of the east orebody	106
III-6	Geology and ore zoning, 3rd level, Sherritt Gordon Mine, Sherridon, Manitoba	in pocket
III-7	Geology of 'offshoot' orebodies; a) with pegmatite b) without pegmatite	108

List of Figures cont'd

Figure		Page
III-8	Petrology of some ore specimens from the Sherritt Gordon Mine	111
III-9	Longitudinal metal zoning, West Orebody	113
III-10	Longitudinal metal zoning, East Orebody	114
III-11	fO ₂ vs. pH Diagram illustrating the solubility of Cu, Pb and Zn at T = 300°C, $\Sigma S = 10^{-2}$ m and NaCl = 3 m	134
III-12a	fO ₂ vs. pH Diagrams at 150°C and 3 m NaCl	136
III-12b	fO ₂ vs. pH Diagrams at 200°C and 3 m NaCl	137
III-12c	fO ₂ vs. pH Diagrams at 250°C and 3 m NaCl	138
III-12d	fO ₂ vs. pH Diagrams at 300°C and 3 m NaCl	139
III-13	fO ₂ vs. T Diagrams at pH = 3 and 4 and 3 m NaCl	140
III-14	fO ₂ vs. T Diagram at 3 m NaCl modified to show the path of a cooling ore solution	144
IV-1	Axial Traces in the Sherridon Area	152
IV-2	Cross Sections through the Sherritt Gordon mine: a) 3500' North of West Shaft b) 2000' East of East Shaft	155
IV-3	Stereo plot of structural domains on West flank of Sherridon Basin	in pocket
IV-4	Mineral assemblages plotted on: a) AFM, and b) ACF diagrams	162
IV-5	Mineral assemblages shown on the Bi composition surface	164
IV-6	Mineral reactions at the amphibolite-granulite transition in Bi-bearing rocks	165

List of Figures cont'd

<u>Figure</u>		<u>Page</u>
IV-7	Peak metamorphic pressure at Sherridon, mean and range FeS in sphalerite shown	168
IV-8	Petrogenetic grid for pelitic rocks	169
V-1	Reconstruction of Sherridon-Flin Flon area during the deposition of Sherritt Gordon mineralization	177
V-2	Structural and metamorphic evolution of the Kiseynew and related belts	182
A-1	Chemical analyses and thin section locations	in pocket
B-1	Location of sample traverse for samples from regolith near Flin Flon, Manitoba	227

LIST OF PLATES

<u>Plate</u>		<u>Page</u>
II-1	Textures in Meta-sedimentary Rocks of the Sherridon Area	48
II-2	Textures in Meta-volcanic Rocks of the Sherridon Area	65
II-3	Textures in Metamorphosed Intrusive Rocks of the Sherridon Area	82

LIST OF APPENDICES

<u>Appendix</u>	<u>Page</u>
A Chemical Analyses of Rocks from Sherridon, Manitoba	195
B Chemical Analyses of Rocks from a Regolith near Flin Flon, Manitoba	226
C Mineral Assemblages and Modes for Thin Sections of Rocks from the Sherridon Area, Manitoba	229
D Mineral Assemblages and Modes for Polished Thin Sections of Rocks from Sulphide Occur- rences at Sherridon, Manitoba	239
E C.I.P.W. Norms Calculated for the Igneous Rocks of Appendix A	242

SYMBOLS

a	- Maier-Kelley heat capacity coefficient - activity
a_i	- activity of i^{th} species
q	- distance of closest approach of ions in solution
A	- temperature dependent Debye-Hückel coefficient
$X_{(\text{aq})}$	- aqueous species
b	- Maier-Kelley heat capacity coefficient
B	- temperature dependent Debye-Hückel coefficient
$\bar{B}$	- deviation function in extended Debye-Hückel equation
c	- Maier-Kelley heat capacity coefficient
f	- fugacity
f_i	- fugacity of i^{th} species
$X_{(\text{g})}$	- gaseous species
$\Delta H^\circ_{298.15}$	- standard enthalpy of reaction at 298.15°K
I	- true ionic strength
K	- equilibrium constant
L	- ligand
m	- molality
m_i	- molality of i^{th} species
M	- metal species
n	- number of exponents in law of mass action equation
R	- gas constant equal to 1.9872 cal.mole. deg
S	- solubility
ΣS	- moles of sulphur per kilogram of water
$\delta^{34}\text{S}$	- deviation of sulphur isotopes from Canyon Diablo standard

$\Delta S_{298.15}$

- standard entropy of reaction at 298.15°K

T

- temperature

Z

- charge

X

- vibration direction of fast ray

B

- vibration direction of intermediate ray
- solubility constant defined in terms of concentration

B_i

- solubility constant of i^{th} species in terms of concentration

B_n

- overall solubility constant

B'

- solubility constant defined in terms of activities

B'_i

- solubility constant of i^{th} species in terms of activities

γ_i

- activity coefficient of i^{th} species

ϕ

- chloride complexing factor in solubility equation

MINERAL ABBREVIATIONS

Silicate and Non-Sulphide Minerals

AB	Albite
AC	Actinolite
AL	Almandine Garnet
AM	Amphibole
AN	Anorthite
AP	Apatite
AT	Anthophyllite
BI	Biotite
BI ₁	Red Biotite
BI ₂	Green Biotite
C	Corundum
CC	Calcite
CH	Chlorite
CH ₁	Pennine
CO	Cordierite
CPX	Clinopyroxene
CU	Cumingtonite
DI	Diopside
EN	Enstatite
EP	Epidote
FA	Fayalite
FO	Forsterite
FS	Ferrosilite
GA	Garnet
GH	Gahnite
GR	Grossular Garnet
GT	Graphite
HB	Hornblende
HB ₁	Brown Hornblende
HC	Hercynite
HE	Hedenbergite
HM	Hematite

IL	Ilmenite
KF	Microcline
MT	Magnetite
MV	Muscovite
NE	Nepheline
OPX	Orthopyroxene
OR	Orthoclase
PH	Phlogopite
PI	Pinnite
Q	Quartz
SC	Scapolite
SE	Serpentine
SH	Sphene
SL	Sillimanite
SN	Spinel
SR	Sericite
ST	Staurolite
WO	Wollastonite

Sulphide Minerals

AS	Arsenopyrite*
CP	Chalcopyrite
CU	Cubanite
GL	Galena
PO	Pyrrhotite
PO ₁	Hexagonal Pyrrhotite
PO ₂	Low Temperature Pyrrhotite
PY	Pyrite
SP	Sphalerite

INTRODUCTION

The vicinity of Sherridon is the type area of the Sherridon Group in the Kisseynew gneiss belt, Manitoba. The last published geological works consist of a map with marginal notes on a scale of 1:63,360 (Bateman and Harrison, 1946) covering the western part of the Sherridon area; a memoir, which includes a map on a scale of 1:63,360 (Robertson, 1953) covering the eastern part of the area and a preliminary map on a scale of 1"=1,000' of the Sherritt Gordon mine area (Bateman, 1944). After the closing of the Sherritt Gordon mine in 1953, only sporadic interest was shown in the Kisseynew gneiss belt until the early 1970's when both the Manitoba Mines Branch and the Saskatchewan Department of Mineral Resources initiated major mapping projects. The vicinity of Sherridon represents a key area to an understanding of the Kisseynew gneisses and of the associated sulphide deposits. Consequently, petrological studies in the area were carried out by the Geological Survey of Canada (Project 740024, E. Froese) with the purpose of reconstructing the depositional and metamorphic environments. This thesis is concerned mainly with the first aspect.

The author was assigned the mapping of the western margin of the area at a scale of 1"=660' and the detailed mapping of the mine area at a scale of 1"=200'. General mapping of the eastern part of the area at a scale of 1:25,000 was also carried out, mainly by E. Froese and partly by the author.

A summary of the geological setting is presented in Chapter I. The stratigraphy and petrography of the gneisses mapped are

presented in Chapter II and on the accompanying maps. Rock samples for chemical analysis were collected in conjunction with the mapping; the chemical analyses are the basis of a geochemical study also presented in Chapter II.

Files of the Sherritt Gordon Mining Company and the Hudson Bay Exploration and Development Company were examined; material concerning the Sherridon orebodies was summarized and is presented in Chapter III, together with the petrography of the ore body. A geochemical model for the origin of the deposits is also presented. The metamorphism and structure are dealt with in Chapter IV. The mineral assemblages of the carbonate and pelitic rocks are described and an estimation of peak metamorphic pressure and temperature is given. The structural development of the Sherridon area and its relation to metamorphism is described.

Chapter IV presents a general discussion and concludes the thesis. A paleoenvironmental model based on the lithology, geochemistry and stratigraphic relationships of the various units is proposed. The timing of metamorphism and the succession of structural events in the Kisseynew belt are also examined.

CHAPTER I

GEOLOGICAL SETTING OF THE SHERRIDON AREA

I-1 Geological Setting

An area between Kississing Lake and Star Lake, known as the Sherridon structure, is underlain by rocks of the Sherridon Group, a stratigraphic unit of the Kiseynew gneiss belt. The belt occupies the southeast corner of the Churchill province in Manitoba (Fig. I-1), is 240 km long; 140 km wide and consists of 35,000 km² of high grade metamorphic terrain. The gneisses of the belt are complexly folded and about 80 percent of the belt is a migmatite complex derived from graywacke and shale (the Nokomis Group). The remainder of the gneisses are metamorphosed lithic arenites (the Sherridon Group) which stratigraphically overlie and occupy basins on the Nokomis migmatites. Minor amounts of gneiss derived from calcareous sediments and volcanic rocks occur in both groups. Granitic rocks which are considered to be anatectic, form a series of large plutons near the center of the belt. Table I-1 lists the stratigraphic relations of the rocks of the Kiseynew belt and the adjacent greenstone belts.

The Kiseynew belt is bounded on the west side by the northerly trending Tabernor fault zone and the rocks of the Glennie domain. The fault zone (Lewry et al., 1978) is considered to be a narrow element in a zone of Hadrynian mylonite. Metamorphic gradients across the fault zone are steep, marked by the transition of andalusite to sillimanite and the development of cordierite

TABLE I-1

STRATIGRAPHIC RELATIONSHIPS OF KISSEYNEW AND ADJACENT BELTS

Flin Flon-Snow Lake Belt*		Kisseynew Gneiss Belt		Lynn Lake Belt**	
<u>Post Missi Intrusive Group</u>		<u>Post Sheridon Intrusive Group</u>		<u>Post Sickie Intrusive Group</u>	
8 Minor intrusions related to 7		6 Pyroxenite, gabbro		8 Quartz porphyry	
7 Granodiorite, quartz diorite; largely gneissic		5 Granodiorite gneiss.		7 Granite, granodiorite and quartz diorite	
6 Gabbro and diorite with ultramafic phases		4 Quartz Monzonite			
		<u>Sheridon Group</u>		<u>Sickie Group</u>	
<u>Missi Group</u>		3 Lithic arenite and carbonate rocks		7 Lithic arenite, graywacke pillowed lavas and conglomerate	
5 Lithic arenite, graywacke quartzite and basal conglomerate		2 Pillowed and fragmental mafic volcanic rocks			
				<u>Post Wasekwan Intrusives</u>	
<u>Post Amisk Intrusives</u>				5 Massive granite and granodiorite	
4 Quartz-eye "granite"				4 Gabbro and diorite with ultra-mafic phases	
<u>Amisk Group</u>		<u>Nokomis Group</u>		<u>Wasekwan Group</u>	
3 Argillite, graywacke and tuff		1 Paragneiss derived from graywacke and unknown precursor		3 Felsic fragmental rocks and related sediments	
2 Rhyolite, dacite and quartz porphyry				2 Mafic to intermediate pillowed and fragmental volcanic rocks	
1 Mafic to intermediate pillowed and fragmental volcanic rocks				1 Graywacke, conglomerate and minor buffaceous rocks	

* Source: Bailes (1971).

**Source: McGlynn (1970)

and hypersthene. The Tabernor fault zone appears to be a significant crustal break.

On the east, the Kiseynew belt is bounded by the Setting Lake-Ospwagan Lake fault zone and the rocks of the Thompson-Pikwitonei region. Weber and Scoates (1978) state that the fault zone divides the Archean rocks of the Superior province to the east from the Proterozoic rocks of the Churchill province to the west. The primary relationship of the two provinces is considered to be an unconformity, remnants of which are preserved in the fault zone.

To the south and north of the Kiseynew belt lie the Flin Flon-Snow Lake and the Lynn Lake greenstone belts, respectively. Both belts are dominated by thick basal sequences of volcanic rocks overlain by graywacke and shales derived from them. Stauffer (1974) has demonstrated that the volcanic rocks of the Flin Flon-Snow Lake belt belong to the island arc tholeiitic series. Fox (1977) has demonstrated that a calc-alkalic suite overlies the tholeiitic series at the western margin of the Flin Flon-Snow Lake belt. In both belts, conglomerates and arenites unconformably overlie the volcanic rocks and the associated sediments. In the Flin Flon area, Mukherjee (1974) has interpreted these rocks as a piedmont alluvial fan deposit.

The transition from the greenstone belts to the gneiss belt is characterized by steep metamorphic gradients (Froese and Gasparrini, 1975; Bailes and McRitchie, 1978). There is no structural or stratigraphic discontinuity between the belts (Bailes, 1975, 1979). More recently, transitional lithofacies relationships have been noted between graywacke-shale paragneisses

of the Kiseynew belt and the volcanic rocks of the greenstone belts (Bailes, 1979; Zwanzig, 1976).

I-2 Geochronology

Radiometric dating was first carried out in the southern part of the Churchill province to determine whether the Lynn Lake and Flin Flon-Snow Lake belts are of Archean age. The east-west trend of these belts is similar to that of the Archean rocks of the Superior province although they are separated by the intervening Thompson-Pikwitonei region.

During the 1960's, K-Ar dates were determined for biotite and muscovite from paragneiss and granitic rocks of the volcanic and gneiss belts. With the exception of one date from Lynn Lake, all the analyses (Table I-2) fall between 1610 and 1960 m.y. with a mean of 1719 m.y. Ages from the Lynn Lake belt give the lowest ages with a mean of 1641 m.y. The Kiseynew gneiss belt is the next oldest with a mean age of 1712 m.y. and the Flin Flon-Snow Lake belt is the oldest with a mean age of 1750 m.y. It was soon realized that metamorphism allowed the escape of Ar gas, effectively resetting the K-Ar clock. Rather than giving the age of the rocks, the K-Ar date gives the time at which the cooling temperature falls below about 100°C for the last period of metamorphism.

To overcome this problem geochronologists switched to Rb-Sr whole rock dating in the early 1970's. The advantage of this method is that the composition of a hand specimen-sized sample is thought to remain constant despite the effects of metamorphism.

TABLE I-2

AGE DETERMINATIONS OF THE KISSEYNE GNEISSES AND ADJACENT VOLCANIC BELTS

K-Ar

REFERENCE	LOCATION	DESCRIPTION	MATERIAL DATED	AGE (M.Y.)
Flin Flon-Snow Lake Belt				
Moore et al 1960	Stall L.	Qz-Bt Paragneiss	Bt	1760
Lowden 1961	Reed L.	Granite cutting Amisk	Bt	1745
Lowden et al 1962	Hat L.	Meta-quartzite	Bt	1770
"	Reed L.	Granite cutting Amisk	Mv	1775
"	Kisseynew L.	Paragneiss	Bt	1735
"	Hat L.	Meta-quartzite	Mv	1620
Wanless et al 1964	Cliff L.	Granite (qz-eye)	Mv	1620
"	Wekusko L.	Monzonite	Bt	1960± 60
"	Phantom L.	Granodiorite	Bt	1865± 61
"	Crowduck Bay	Pegmatite	Mv	1790± 60
"	Crowduck Bay	Pegmatite	Bt	1610± 50
Kisseynew Gneiss Belt				
Lowden 1961	Kisseynew L.	Paragneiss	Bt	1735
"	Wood L.	Paragneiss	Bt	1630
"	Meagle L.	Paragneiss	Bt	1730
Lowden et al 1962	McCavock L.	Paragneiss	Bt	1720
"	McCavock L.	Paragneiss	Mv	1745
Lynn Lake Belt				
Lowden 1961	Veronica L.	Wasekwan gwk.	Bt	1610
"	Sickle L.	Granitic gneiss	Bt	1655
"	Granville L.	Meta-arkose	Bt	1690
Leech et al 1963	Granville L.	Paragneiss	Mv	2670
"	Granville L.	Paragneiss	Bt	1650

TABLE I-2 (cont.)

Rb-Sr					
REFERENCE	LOCATION	DESCRIPTION	METHOD	INITIAL RATIO	*AGE (M.Y.)
<u>Flin Flon-Snow Lake Belt</u>					
Flin Flon-Snow Lake Belt					
1970	Hanson L.	Metavolcanics	Isochron	.7012±.0005	2468±59
?	"	Granite	"	.7008±.0005	2394±16
?	"	Pegmatite	"	.6980±.0015	1761± 2
Mukherjee et al	Flin Flon	Amisk volcanics	"	.7039±.0013	1738±87
"	Amabel L.	Granodiorite	"	.7023±.0007	1766±14
1974	"	Amisk volcanics	"	.7020±.0002	1700±252
Josse et al	"	Morton Porphyry	"	.7029±.0001	1678±81
"	"	Haw's Pluton	"	.7019±.0005	1821±110
"	"	Pegmatite	Whole rock	.710 assumed	1704±54
"	"	Pegmatite	Whole rock	.710	1728±54
1977	Snow L.	Amisk volcanics	Isochron	.7020±.0003	1791±41
Moore	"	Amisk volcanics	"	.7018±.0010	1762±51
<u>Kisseynew Gneiss Belt</u>					
Anderson et al	Burntwood L.	Paragneiss	Isochron	.7032±.0013	1780±195
1973	File Gneiss Dome	Felsic Gneiss	"	.7038±.0008	1723±42
Josse et al	Burntwood L.	Hypersthene Intrusive	"	.7030±.0007	1786±78
Clark et al	"	Paragneiss	"	.7033±.0011	1811±78
1977	Triangle Dome	Granitoid Gneiss	"	.7017±.0014	1811±82
Moore	Squall Domé	Granitoid Gneiss	"	.7011±.0007	1772±62

*All ages have been recalculated using a decay constant of $^{87}\text{Rb} = 1.42 \times 10^{-11} \text{ yr}^{-1}$
 ?Correlation with Flin Flon Belt questionable.

TABLE I-2 (cont.)

REFERENCE	Pb/Pb	LOCATION	DESCRIPTION	MATERIAL DATED	AGE (M.Y.)	
					Stacey-Kramer	Cumming-Richards
<u>Flin Flon-Snow Lake Belt</u>						
Sangster		Flin Flon	Massive Sulphide	Galena	1888	1848
"		Osborne	"	Galena	1880	1846
"		Stall	"	Galena	1841	1788
"		Chisel	"	Galena	1752	1744
"		Schist	"	Galena	1675	1708
<u>Kisseynew Gneiss Belt</u>						
Sangster		Sherridon	Massive Sulphide	Sphalerite	1787	1854
<u>Lynn Lake Belt</u>						
Sangster		Brabant	Massive Sulphide	Galena	1899	1829
"		Fox	"	Galena	1979	1832
"		Ruttan	"	Galena	1794	1768
"		Will	"	Galena	1720	1778

In order to calculate the initial ratio of radiogenic to non-radiogenic Sr, a number of samples of the rock are analyzed to produce an isochron from which the age may be determined. Rb-Sr ages for the Flin Flon-Snow Lake and Kiseynew belts fall between 1566 and 1821 m.y. with a mean of 1744 m.y. The Kiseynew belt gives a mean age of 1781 m.y. and the Flin Flon-Snow Lake belt gives a mean age of 1726 m.y. The Archean ages from the Hanson Lake area have been discounted since they are now recognized (Lewry and Sibbald, 1978) to come from rocks which fall in the Glennie Domain to the west of the Tabernor fault zone and not in the Flin Flon belt as believed earlier (Coleman, 1970). In addition, the initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the Hanson Lake rocks are higher than those of the rocks to the east.

Pb-Pb dates have also been determined for galena and sphalerite from the massive sulphide deposits of the district. Acceptance of the Pb-Pb model has come gradually since the discovery of anomalous leads such as the J-type of Missouri. Nevertheless, Stanton and Russell (1959) and York and Farquhar (1972) have demonstrated that a remarkable consistency exists for the age relationships of leads from massive sulphide deposits. This regularity suggests that the history of conformable leads can be expressed by either the Stacey and Kramers (1975) two stage or the Cummings and Richards (1975) one stage model on which basis ages can be estimated. Pb-Pb ages range from 1675 to 1979 m.y. with a mean of 1822 m.y. for the two stage model and from 1708 to 1854 m.y. with a mean of 1800 m.y. for the one stage model (Sangster, 1978).

A number of U-Pb determinations have been carried out on zircons from various rocks of the Flin Flon area and ages of 1800 to 1900 m.y. were determined (MacQuarrie, 1977).

The ages from all types of determinations are compatible and suggest contemporaneous deposition of these three belts in the time range of 1900 to 1800 m.y. ago. Metamorphism and concomitant plutonism occurred during the time range of 1800 to 1700 m.y. ago. The age of the complex is Aphebian.

I-3 Pole Position

Park (1975) has calculated pole positions for the Flin Flon belt based on paleomagnetic orientations of rocks from the belt. He found two magnetizations. The earliest gave a pole position of 20°S and 46°W and is thought to correspond to an original thermoremanent magnetization of the rocks of the Boundary Intrusions 1800 m.y. ago. The second magnetization is present in all the rocks and gives a pole position of 24°N and 95°W . It is thought this magnetization corresponds to a viscous partial thermoremanent magnetization acquired during uplift of the Hudsonian orogeny 1600 to 1700 m.y. ago.

Fig. I-2 illustrates the position of the Laurentian continent during the period 2300 to 1500 m.y. ago. It would appear that the Flin Flon belt was laid down and "glued together" with Laurentia during the period 1700-1900 m.y. ago at which time it would have lain between the latitudes 30°N and 30°S and a subtropical climate would have prevailed.

FIG. 1-2. Latitude maps of Laurentia for the interval -2300 to -1500 Ma;
from Roy and Lapointe (1976)

I-4 Previous Work

Bell (1880) and Tyrrell (1902) mapped areas which are probably part of the Kiseynew gneisses. Dowling (1900) certainly crossed the gneisses when he descended the Burntwood and Kississing Rivers in 1889. During 1914 to 1918, Bruce (1917, 1918) carried out the first systematic mapping of northern Manitoba in the vicinity of what is now Flin Flon. He separated the rocks into three stratigraphic divisions: Amisk volcanics, Missi sediments, and Kiseynew gneisses. Interest in the area increased due to the discovery of the Flin Flon deposit in 1915 and the first production of metals from the Mandy Mine in 1917.

In 1922, an Indian, Mr. Phillip Sherlett, is reported to have staked the Sherritt Gordon deposit although a white trapper living in the vicinity is believed to have found several leached surface outcrops previously (Wright, 1930). In 1924, Sherlett's claims lapsed because of failure to report assessment work and were picked up by Mr. Carl Sherritt and his partner Mr. Dick Madole. In 1925, the deposit was optioned by Mr. J. P. Gordon who assigned it for development work to a number of different syndicates. One of these, Victoria Syndicate Limited, under Mr. E. L. Brown, formed Sherritt Gordon Mines Limited on July 26, 1927, and bought out the other interests. Shaft sinking operations were initiated and production commenced in 1931.

At the same time as development was initiated at the Sherritt Gordon mine site, geological mapping commenced. Wright (1929) carried out a reconnaissance survey of Kississing Lake and showed that the area was underlain by Kiseynew gneisses. He classified

the rocks as sediments and concluded that the amphibolites were derived from clayey arkoses. Bruce (1929) and Bruce and Matheson (1930) also accepted a sedimentary origin for the amphibolites. Derry (1942), on the other hand, concluded the amphibolites were derived from mafic lavas and the felsic gneisses from sediments. The first textural evidence of a volcanic origin was discovered by Bateman and Harrison (1946) who recognized fragments in some of the amphibolites and classified them as volcanic breccias. They also reported deformed pillow structures. The volcanic origin was accepted by Farley (1949). Later, Harrison (1951) decided that the Sherridon Group amphibolites were calcareous arenites. He was influenced by the presence of impure limestones in the Sherridon Group (Wright, 1930). Robertson (1953) compromised by recognizing amphibolites of volcanic and sedimentary origin in the Sherridon Group.

During this same period, a number of conflicting structural and stratigraphic interpretations of the rocks of the Sherridon area were presented. Bruce (1933) interpreted the Sherridon area as an overturned anticline. Farley (1948) considered the structure of the Sherridon area to be a synclinal basin folded by a later anticline. Bateman (1944) concluded that the structure was an anticlinal dome. He called the outer rocks in the dome the Sherridon Group. The rocks lying outside the dome were considered to be younger (Post-Sherridon Group) and the rocks at the center of the dome were considered to be older (Pre-Sherridon Group). Robertson (in Davies et al., 1962) reinterpreted the structure as a basin with a domed center. He retained the Sherridon Group but combined

the Pre- and Post-Sherridon Groups into the Nokomis Group after his type locality at Nokomis Lake just east of the structure (Robertson, 1953). He considered the rocks at the center of the dome and those outside the basin to be equivalent in age. This view that structural basins of Sherridon Group rocks overlie Nokomis Group rocks has been retained to the present (Pollock, 1964, 1965; Bailes 1971).

The age relations of the Kiseynew rocks with the neighbouring greenstone belts was summed up by Harrison (1951). Today, two main schools of thought prevail:

1. The Kiseynew gneisses are a complex of different ages probably including both Amisk/Wasekwan and Missi/Sickle strata as well as older formations.
2. The Kiseynew gneisses are the metamorphosed equivalents of supracrustal rocks of the greenstone belts (Bailes, 1971; McRitchie, 1974).

The validity of both hypotheses will be discussed in the final chapter.

The genesis of the Sherridon Group orebodies has not received much discussion, mainly because the ore reserves were depleted before the development of modern theories on massive sulphides. Bruce (1929), Wright (1930), Bateman (1944), and Farley (1949) postulated hydrothermal replacement based on the association of large amounts of pegmatites with the orebodies. Sangster (1972) included the Sherritt Gordon deposit on a map of Precambrian massive sulphide orebodies in his paper on volcanogenic massive sulphide deposits but failed to mention the deposit in the text. The origin of the Sherritt Gordon and related orebodies will be dealt with in detail in the chapter on economic geology.

I-5 Logistics

The area of the study is the west margin of the Sherridon area (Fig. II-2). It is predominantly open, wooded till plain with variable outcrop density. Two small towns, Sherridon and Cold Lake, lie at the south end of the map area. Access can be had by C. N. Railway and by float plane on numerous lakes. A number of abandoned bush roads cut by the Sherritt Gordon Mining Company give access to the immediate area. Transportation was by foot and canoe.

Geological data were collected by traversing with pace and compass using aerial photographs for control. Field work was carried out during the summers of 1975 and 1976, and briefly in the spring of 1977. Mapping was carried out under the direction of E. Froese. Geological assistants were D. K. Webster and S. J. Wilkinson during 1975, and H. E. Jamieson and J. Coppelmanns during 1976.

CHAPTER II

LITHOLOGY, STRATIGRAPHY AND GEOCHEMISTRY

II-1 Introduction and Methods of Study

The stratigraphic and lithologic descriptions are based on a combination of field observations, petrographic study and chemical analyses of rock samples. Chemical analyses of 195 rock samples collected from the Sherridon area are tabulated in Appendix A. Appendix B lists seven chemical analyses of rocks collected from a paleoregolith in the Flin Flon area by D. Price. Descriptions of the preparation, methods of analysis, reliability, accuracy and precision of the analyses are included in the introduction to Appendix A. The mineralogy and modes of the thin sections examined are listed in Appendix C. Appendix D contains the mineralogy and modes of the polished thin sections which were examined. C.I.P.W. norms were calculated for the igneous rock specimens analyzed and are contained in Appendix E.

The data generated from these studies were plotted on chemical and mineralogical diagrams, averaged and compared with possible ancient and modern analogues and subjected to multivariate statistical analyses for a fuller understanding of the rocks themselves and to provide guidelines for the analysis of similar terrains. The chemical and mineralogical plots used are all self-explanatory, as are the comparisons with analogues and will not be discussed further. The multivariate statistical procedures require a few additional words of explanation.

The two measures used are cluster analysis and principal component analysis. Cluster analysis is an efficient way of displaying complex relationships among objects based on sets of observations taken on the objects. At the present time, the theoretical underpinnings of cluster analysis are incomplete. Little is known of the statistical properties of numerical taxonomic methods and very few tests of significance are available. Clustering can be done using a variety of methods all of which could yield different results. Therefore, Davis (1973) suggested that in cluster analysis and principal component analysis, utility should be judged by performance and not theoretical considerations.

In order to carry out a cluster analysis on the chemical data gathered in this thesis, it was first necessary to standardize the data. Standardization consists of transforming the sets of observations so that they all have a mean of zero and a standard deviation of one. Once standardized, each set of observations will have equal weight in determining the similarity of the various objects. If standardization were not carried out on the data set, observations in ppm would be essentially ignored and SiO_2 would swamp the other observations. Standardization has a tremendous influence on the results of the analysis.

Principal component analysis consists of a linear transformation of initial variables to new variables, where the new variables are combinations of the initial variables and successively account for as much of the total variance as possible. The new variables or components are nothing more than the eigenvectors of a similarity matrix although they are uncorrelated, unlike the original variables. The principal components can be used to provide

insight into the multivariate data set by introducing parsimony of description by reducing the dimensionality of the data set since the bulk of the aggregate variance can usually be explained by relatively few principal components.

The computer programs which have been used to carry out the analysis of the data using the above outlined measures were taken from Davis (1973) with only minor adaptations.

II-2 Local Setting and Stratigraphy

As will be demonstrated in Chapter IV, the rocks of the Sherridon Group at Sherridon occupy a structural basin which overlies rocks of the Nokomis Group. The local geology at a scale of 1:25,000 (Froese and Goetz, 1979) is shown on Fig. II-1 (in pocket) and the detailed geology of the western flank of the basin at a scale of 1"=660' is shown on Fig. II-2 (in pocket). A modified stratigraphic column is shown on Fig. II-3 and a table of lithologic units is provided in Table II-1.

The rocks in the thesis area which have been assigned to the Nokomis Group are highly variable in structure and composition. They are represented by paragneiss, amphibolite, augen gneiss, migmatite and granitoid gneiss. The boundary between the rocks of the Nokomis and Sherridon Groups has been intruded by a gabbro sill which has obliterated their mutual relationships. Contacts of rocks from both groups with the sill are sharp.

Across the contact represented by the gabbro sill is a vast difference in metamorphic and tectonic style. The rocks of the Nokomis Group constitute a migmatite complex characterized by

FIG II-3.
Stratigraphic column for
the Sherridon Group

SYMBOLS

- unconformity
- massive sulphide horizon
- chert horizon

LEGEND

- lithic arenite
- graywacke
- calcareous arenite
- limestone
- basalt volcanoclastics
- basalt flows
- gabbro
- migmatite

TABLE II-1

TABLE OF LITHOLOGIES

	Observed Lithology	Interpreted Lithology
19	pegmatite	
	Sherridon Group	
18	biotite gneiss	porphyritic granite
17	amphibolite, layered	gabbro
17a	amphibolite	pyroxenite
16	sulphide occurrence	
15	felsic gneiss	felsic volcanic rocks
14	amphibolite with possible pillow structures	basaltic flow
14a	massive amphibolite	basaltic flow
13	amphibolite with clasts and layering	basaltic pyroclastic or debris flow
13a	amphibolite with layering	basaltic pyroclastic or debris flow
12	marble and calc-silicate	limestone
11	quartz rich hornblende gneiss and amphibolite	calcareous arenite
11a	diopside-bearing gneiss	calcareous argillite
10	anth ophyllite-garnet gneiss	magnesian sediments
10a	cummingtonite gneiss	paleo-regolith
10b	biotite-garnet-rich gneiss	paleo-regolith
9	biotite-garnet-silli- manite gneiss	argillite (pelite)
8	biotite-garnet schist	graywacke
7	quartzitic gneiss	chert
6	quartzofeldspathic gneiss	quartzofeldspathic arenite
5	biotite gneiss with ridges	lithic arenite
5a	sillimanite-bearing bio- tite gneiss	lithic arenite
5b	sillimanite nodule-bearing biotite gneiss	lithic arenite
5c	mafic biotite gneiss	lithic arenite

Table II-1 (cont'd)

		Observed Lithology	Interpreted Lithology
Nokomis Group			
	4	granitic rocks	
	3	stratified biotite gneiss	
	3a	migmatite complex	
	2	augen biotite gneiss	
	1	amphibolite	

partial melting, extreme lithological variation and abundant minor folding. The rocks of the Sherridon Group show almost no evidence of melting and very few minor structures. It has been suggested that the absence of melting and differences in tectonic style between the Sherridon and Nokomis Groups arises from compositional differences (Bailes, pers. comm., 1978). In the Sherridon area, however, the composition of both Groups overlaps considerably. This means that the rocks assigned to the Nokomis Group in the Sherridon area must have undergone high grade metamorphism and migmatization at some time before the deposition of the rocks of the Sherridon Group. The presence of an older basement group and an unconformity could account for the abrupt change in metamorphic style at the base of the Sherridon Group at Sherridon.

The rocks of the Sherridon Group are derived from sediments with intercalated lava flows, volcanoclastic rocks and sulphide-bearing horizons. The metasedimentary rocks can be divided into five informal formations (Fig. II-3), of which the author has mapped parts of the lower four. The divisions consist of:

- E) Upper Arenite: biotite gneiss derived from lithic arenite and truncated by the present day erosional unconformity
- D) Graywacke: biotite-garnet schist derived from a shale or mudstone
- C) Arenaceous Limestone: a marble unit derived from limestone and quartzofeldspathic arenite
- B) Lower Arenite: a thick unit of biotite gneiss derived from lithic arenite which becomes more thickly bedded and cleaner toward the top
- A) Calcareous Arenite: a thin calc-silicate layer derived from calcareous arenite and wacke

in order from top (E) to bottom (A).

Layers of amphibolite derived from volcanic flows and volcanoclastic rocks of basaltic composition are interspersed throughout formations (A) and (B) but do not occur in the upper part of the sedimentary succession. One small pod of felsic volcanic rocks was recognized. Mafic and felsic intrusions occur throughout. Cherty pyrrhotite horizons are associated with the calcareous formation (A) and (C). Economic massive sulphide lenses are associated with formation (B).

In other areas of Sherridon Group rocks only formations (A) and (B) have been recognized. Volcanic rocks are rare and massive sulphide deposits are apparently absent. Thus, the rocks of the Sherridon structure contain the most complete lithologic variation of the Sherridon Group extant and it is from these rocks that a complete picture of the Sherridon paleoenvironment must be constructed.

II-3 Nokomis Group (Units 1-4)

The rocks of the Nokomis Group in the thesis area could be divided into four units of which only Unit 3 is quantitatively important. Unit 3 consists of a uniform, foliated biotite gneiss with a grain size averaging about 0.5 mm. Subhedral to anhedral grains of mauve garnet are a striking constituent (Robertson, 1953). The garnets are slightly larger than the average grain size, measuring 0.5 to 2 mm in diameter. The biotite in this unit is brown and constitutes about 15 percent of the rock. Quartz is invariably the most common constituent and forms unstrained aggregates, some of which are elongated parallel with the direction

of foliation. Plagioclase (An 25-35) occurs both twinned and untwinned, and forms granoblastic polygonal aggregates with the quartz grains. Although not recognized in the thesis area, sporadic occurrences of sillimanite, microcline and graphite have been reported from the rocks of this unit (Robertson, 1953; Pollock, 1964).

This unit contains quartzofeldspathic pods and layers, some of which grade into pegmatite. The layers average 5 to 20 cm in thickness, show an increase in grain size of 2 to 5 mm and give the impression of bedding. However, the pods and bands are inferred to be a result of partial melting. Where the amount of quartzofeldspathic material and pegmatite are abundant, a migmatite subunit (3a) is recognized.

Granitic segregations within outcrops of migmatite vary considerably in style although segregations parallel to the foliation and arterial networks are the most common varieties. In a few instances, partial melting has been almost complete leaving fragments of more refractory gneiss in a granitic matrix. Mafic segregations of biotite and hornblende are common. Minor folds are much more abundant in outcrops of migmatite than in rocks which have not undergone migmatization. Late veins and dykes of pegmatite and aplite cut the migmatites and each other.

Chemically (Appendix A) the rocks of Unit 3 resemble graywacke. Outcrops of rocks similar to this unit at a lower grade of metamorphism near Nokomis Lake exhibit compositional layering which further supports the interpretation of graywacke protolith.

Within the framework of paragneiss and migmatite, horizons of amphibolite (Unit 1), augen gneiss (Unit 2), and granite (Unit 3) occur. The amphibolites are found as thin lenses or pods, which grade into the biotite gneiss of Unit 3 or have distinct margins. In outcrop, the amphibolites have been invaded by granitic material especially where they are associated with migmatites.

The most abundant mineral found in the amphibolites is green hornblende which forms a weak foliation or lineation. Diopside is found intergrown with the hornblende. Small red garnets and flakes of biotite are common. Quartz and plagioclase (An 50-60) make up the remainder of the amphibolite. Robertson (1953) reports calcite, epidote and sphene in amphibolites from the Batty Lake area. Chemically the rocks of this unit contain up to 15 percent CaO and can be considered to be a calc-silicate. The derivation of this unit is probably from calcareous layers in graywacke.

Augen gneiss, previously mapped as granite (Bateman and Harrison, 1946) was recognized in the vicinity of Singing Lake. The gneiss consists of porphyroblasts of pink microcline up to 2 cm in diameter, which are aligned to form a foliation in a matrix of quartz, plagioclase and biotite. The microcline porphyroblasts can form up to 80 percent of this unit. Thin layers of augen gneiss, alternating with other gneisses, form a stratigraphic horizon parallel with the foliation. Biotite gneiss of Unit 3 shows the development of pink microcline and patches of augen in the vicinity of augen gneiss layers. The contact relationships of

the gneisses are either gradational over a few metres or well-defined. The derivation of this unit is probably from K-rich sediments intercalated with the graywacke of Unit 3.

One small outcrop of granitic rocks was recognized in the thesis area. It was previously mapped as a granitoid gneiss or "granitized sediment" by Bateman and Harrison (1946). However, distinct angular xenoliths, up to 20 cm in diameter were seen in the outcrop which clearly indicates an igneous origin for this rock type. The rock is a granodiorite in composition and consists of plagioclase, microcline, quartz and biotite, in order of decreasing abundance.

II-4 Identification of Sedimentary and Volcanic Rock in the Sherridon Group

Metamorphism and tectonism have obliterated most of the relict features from the rocks of the Sherridon Group, making a genetic identification of the amphibolites and felsic gneisses difficult. As satisfactory, general criteria for the identification of these rocks are not available, it was necessary to develop a group of criteria for specific application to rocks of the Sherridon Group. This was done by using a combination of chemical and mineralogical plots, statistical evaluation and comparison with rocks from neighbouring belts containing relict features.

The dominant rock type of the Sherridon Group is a layered biotite gneiss which resembles the metamorphosed lithic arenites of the Missi Group of the Flin Flon belt. Because of the intimate association of the gneiss with massive sulphide deposits, however,

it could instead be interpreted as a felsic volcanic rock. Only two comprehensive, comparative studies on the chemical properties of felsic igneous rocks and felsic sedimentary rocks have been carried out (Shaw, 1972, and van de Kamp et al, 1975). The study by van de Kamp et al, shows that the discriminant functions devised by Shaw (1972) are inadequate for the discrimination of felsic volcanic and sedimentary rocks. Of the criteria developed by van de Kamp et al, (1975), only two can be used with rocks having the chemical composition of the biotite gneiss of the Sherridon Group:

1. Plots of Niggli mg vs Niggli si, which give a positive slope for sands and a negative slope for igneous rocks because of the concentration of fine mafic minerals in the high SiO_2 fraction of sands.
2. Plots of Zr vs Niggli si, which show an increase in the maximum value of Zr with an increase in si indicating a sedimentary origin because of the concentration of Zr-bearing minerals with the quartzitic fraction in sands.

Fig. II-4a which is a plot of Niggli mg vs Niggli si, does not show a negative slope. Fig. II-4b is a plot of Zr vs Niggli si and shows an increase in the maximum value of Zr with an increase in Niggli si. Figs. II-4a and II-4b suggest that the processes which formed the precursors of the biotite gneiss in the Sherridon Group were sedimentary in nature.

The biotite gneiss of the Sherridon Group can be compared chemically to the biotite gneiss derived from the high grade metamorphism of the Missi Group rocks in the File Lake area (Table II-2, Cols. 1, 2). This is expected because the Sherridon Group rocks are generally considered to be more highly metamorphosed equivalents of the Missi Group.

FIG II-4. Chemistry of Unit 5, biotite gneiss;

a) Niggli mg vs Niggli si

b) Niggli si vs ppm Zr

TABLE II-2 Analyses of Sedimentary and Volcanic Rocks Bearing on the Discrimination of Ortho-gneiss and Para-gneiss, Sherridon, Manitoba.

	1	2	3	4	5	6	7	8	9
MEAN									
SiO ₂	74.43	77.22	67.74	55.76	52.01	51.73	44.65	49.37	54.35
TiO ₂	0.21	0.41	0.70	0.78	0.56	0.64	2.32	1.20	0.70
Al ₂ O ₃	11.77	10.56	14.68	14.96	16.15	15.47	15.57	15.50	16.07
Fe ₂ O ₃	0.46	1.46	0.12	1.93	1.79	2.92	3.13	2.99	3.09
FeO	3.05	2.21	1.40	9.41	9.26	9.32	12.82	10.49	9.68
MnO	0.09	0.05	0.02	0.20	0.19	0.21	0.25	0.22	0.20
MgO	1.92	1.05	2.62	4.34	5.46	6.28	6.07	6.21	5.15
CaO	2.24	1.67	5.89	7.61	8.94	9.10	9.36	9.19	7.01
Na ₂ O	1.74	1.52	3.92	1.90	2.57	1.92	1.75	1.87	2.83
K ₂ O	2.36	2.53	1.69	1.37	0.70	0.46	0.37	0.43	0.57
P ₂ O ₅	0.08	0.07	0.59	0.32	0.14	0.14	0.26	0.18	0.17
CO ₂	0.52	0.22	0.64	0.64	1.59	0.83	1.62	1.09	
H ₂ O	1.13	0.77	0.68	1.47	1.39	1.32	1.60	1.42	
STANDARD DEVIATION									
SiO ₂	2.15	4.03	3.55	5.26	1.60	2.26	4.20	4.65	
TiO ₂	0.07	0.09	0.35	0.39	0.09	0.07	0.73	0.90	
Al ₂ O ₃	1.25	1.77	1.21	1.16	1.28	0.83	1.39	1.05	
Fe ₂ O ₃	0.48	0.74	0.12	1.13	0.92	1.03	1.23	1.11	
FeO	1.17	1.21	0.24	1.51	0.94	0.69	2.46	2.25	
MnO	0.04	0.02	0.01	0.03	0.03	0.02	0.05	0.04	
MgO	0.94	0.40	1.29	2.67	1.30	1.93	0.54	1.61	
CaO	1.09	0.74	3.26	1.28	1.76	0.87	0.82	0.87	
Na ₂ O	1.07	0.66	0.97	1.13	1.01	0.62	0.83	0.76	
K ₂ O	1.29	1.04	0.98	0.79	0.48	0.13	0.05	0.12	
P ₂ O ₅	0.03	0.04	0.41	0.19	0.09	0.06	0.06	0.08	
CO ₂	0.33	0.04	0.27	0.43	2.13	0.81	0.94	0.43	
H ₂ O	0.43	0.47	0.26	0.54	0.50	0.35	0.34	0.37	
SAMPLES	33	4	5	7	17	12	6	18	12

EXPLANATION FOR TABLE

- 1-Biotite gneiss, Sherridon, Unit 5
- 2-Lithic arenite, Missi Group, File Lake Area (Bailes, A.H., 1975, pers. comm)
- 3-Felsic volcanic rocks, Sherridon, Unit 15
- 4-Hornblende paragneiss, Sherridon, Unit 11 in part
- 5-Fragmental amphibolites, Sherridon, Unit 13/13a
- 6-Massive, low TiO₂ amphibolite, Sherridon. Parts of Unit 14
- 7-Massive, high TiO₂ amphibolite, Sherridon. Unit 14
- 8-All massive amphibolite, Sherridon. Unit 14
- 9-Basalt, Flin Flon area (from Stauffer et al, 1975)

Comparison of the biotite gneiss with the known felsic volcanic rocks in the Sherridon Group (Table II-2, Cols. 1, 3) shows a distinct dissimilarity. The volcanic rocks are lower in SiO_2 and total FeO . They are higher in Al_2O_3 , TiO_2 , and Na_2O and have a much smaller FeO/MgO ratio. While this difference does not confirm a sedimentary origin it does show that the biotite gneiss is chemically different from the known felsic volcanics in the area.

As will be discussed later in this chapter, the biotite gneiss is gradational with many of the definitely metasedimentary rock types in the area. Consideration of all these criteria show that the biotite gneiss of the Sherridon area is clearly of sedimentary origin.

It is more difficult to identify the origins of the amphibolites in the Sherridon area. Some, such as Unit 17 and the Cree Lake intrusions have relict textures revealing an igneous origin and others, such as Unit 12, contain abundant diopside + calcite indicating derivation from a calcareous sediment. The remainder are almost featureless. On the basis of field criteria, three divisions were made: 1) massive rocks were considered to be flows; 2) well layered rocks with recognizable fragments at a few localities were considered to be volcanoclastic rocks; and 3) rocks showing differential weathering of layers a few cm thick and recognizable biotite were considered to be meta-sediments.

Chemical analyses were made of samples from these groups to see if the field interpretation could be substantiated. Of the

attempts to distinguish between ortho-amphibolite and para-amphibolite by chemical means, only two studies are based on original research, those of Leake (1964) and van de Kamp (1968). Of the distinguishing methods developed, only one can be employed by this study:

1. Diagrams of Niggli c vs Niggli mg; which show a strong trend parallel to the mg ordinate for volcanic rocks and a large scatter with respect to c for sedimentary rocks.

Other methods rely on minor elements for which the rocks of this thesis have not been analyzed, or require element abundances outside the range found in Sherridon Group rocks.

Fig. II-5a is a plot of Niggli mg vs Niggli c and shows a trend parallel to the Niggli mg axis for the amphibolites though to be developed from a volcanic parent. However, diopside bearing gneiss (Unit 11a) added for comparison and the umphibolites of Unit 11 show a wide scatter with respect to the Niggli c axis.

Fig. II-5b is a simpler plot of CaO vs MgO. On this plot, the rocks thought to be developed from a volcanic parent form a tight cluster and the analyses of the diopside-bearing gneiss and amphibolites from Unit 11 show a large scatter.

Analyses from the amphibolites (Unit 11) have been plotted on a Hughes (1973) diagram (Fig. II-6) which discriminates relatively unaltered igneous rocks from altered igneous rocks and sediments. The rocks interpreted as having been formed from a volcanic parent all plot within the igneous spectrum with the exception of four fragmental amphibolites which fall just outside the spectrum. The fragmental amphibolites may have abnormal alkali ratios due to contamination by terrigenous material. All but one

FIG. II-5. Chemistry of some amphibolites near Sherridon, Man.;
 a) Niggli mg vs Niggli c, after Leake (1964)
 b) CaO vs MgO

□ Amph. Unit 11, in part	○ Frag. Amph. Unit 13/13a
■ Di Gns. Unit 11a	● Masv. Amph. Unit 14/14a

FIG. II-6. Plot of Sherridon amphibolites on an 'Igneous Spectrum Diagram';
 after Hughes (1973) modified by Stauffer et al. (1975)

of the amphibolites interpreted as paragneiss plot outside the igneous spectrum.

Cobalt, chrome and nickel contents of the rocks have been plotted on a ternary diagram (Fig. II-7a) after Franklin (1975). Franklin found that volcanic rocks which had undergone sedimentary transport became depleted in chrome and to a smaller extent in nickel. Fig. II-7a shows that the rocks thought to be paragneiss tend to have lower chrome contents than the rocks thought to be formed from a volcanic parent.

Sedimentary rocks in the Sherridon area have low cobalt, chrome and nickel contents with respect to their copper contents. For this reason, a copper-cobalt-chrome plot was constructed for the amphibolites (Fig. II-7b). It shows that the amphibolites interpreted as paragneiss are enriched in copper relative to the others.

It will be shown in the following section by means of a variety of plots that the mafic volcanic rocks of the Sherridon area were generated from a single magmatic series. This consistency of chemical variation in the rocks interpreted as mafic volcanics, as well as the results of the discrimination techniques used in this section, confirm the field classification.

Cluster and principal component analysis were both carried out on the chemical analyses of the amphibolites. The purpose of the cluster analysis was to test the field classification by using all of the chemical data rather than a few elements or oxides as is possible in simple chemical plots. Fig. II-8 is the result of a cluster analysis performed on the major oxides of the amphibolites. It was found that the highest degree of discrimination was

FIG. II-7 Ternary plots of some trace elements from amphibolites near Sherridon, Man.
 a) Co-Ni-Cr, after Franklin (1975)
 b) Co-Cu-Cr

FIELD CLASS.

- calcareous arenite
- mafic fragmental
- mafic flow
low TiO_2
- mafic flow
high TiO_2

SIMILARITY (r)

0 0.5 1.0

SAMPLES

CLUSTER #

EFFICIENCY

FIG. II-8. Cluster analysis of amphibolites using major oxides

achieved using the correlation coefficient (r) between oxides as variables to achieve a measure of similarity between rock specimens. The results show three groups corresponding mainly with the massive rocks, the rocks with fragments and the differentially weathering rocks. Some confusion in classification has occurred, particularly between massive and fragmental rocks and between fragmental rocks and rocks interpreted as paragneiss.

This confusion is interpreted as arising from the effects of weathering on volcanic flows or very short transport with minimal alteration of clastics in the first instance, and the incorporation of terrigenous material in clastic material during transport in the second.

Fig. II-9 is the result of a cluster analysis which includes trace elements as well as major oxides. The result is much the same except that the massive rocks have been divided into high TiO_2 and low TiO_2 groups. Examination of Fig. II-8 shows that the high TiO_2 rocks form a sub-cluster of the massive rocks on this diagram. This change in the clustering leads to a lower efficiency in the classification of the low TiO_2 rocks but a higher efficiency in classifying the fragmental rocks. Efficiency is defined as the number of analyses correctly clustered, divided by the total number of analyses within the cluster, based on the a priori field classification. In all cases, efficiency exceeds 50 percent and considering the probable effects of metamorphism, weathering, incorporation of terrigenous material, and cementing by carbonates, the results are good.

A principal components analysis was carried out on the data

FIELD CLASS.

- calcareous arenite
- mafic fragmental
- mafic flow low TiO_2
- mafic flow high TiO_2

SIMILARITY (r)

0 0.5 1.0

SAMPLES

CLUSTER #

EFFICIENCY

FIG. 11-9 Cluster analysis of amphibolites using major oxides and minor elements

TABLE II-3 Eigenvalues and Eigenvectors from Principal Component Analysis of Major Oxides of Amphibolites

	1	2	3	4	5	6	7	8	9	10
SiO ₂	.4534	-.1846	-.2019	.1458	-.1180	.0865	-.1489	.2067	.0199	-.0084
TiO ₂	-.4811	-.2451	-.0509	.0539	.783	-.2909	.2248	-.1565	-.2113	-.0576
Al ₂ O ₃	-.1781	.2144	-.2106	-.1287	.16525	.3240	.0152	-.0814	.5061	.1509
Fe ₂ O ₃	-.2658	-.2784	-.1180	-.2824	.0258	.4980	.3632	-.0031	-.4477	.0384
FeO	-.3780	-.2660	.0764	-.0486	.1038	-.3399	-.3296	-.1393	.2230	-.4350
MnO	-.4178	-.1636	-.0181	-.0514	-.980	.0579	-.4780	.4460	-.0797	.3759
MgO	-.1441	.2894	.5571	-.1663	-.2850	-.0215	.0227	-.3256	.0992	.4496
CaO	-.2229	.4251	-.3138	.1628	-.2996	.0364	.2080	-.2036	.0068	-.3082
Na ₂ O	.2639	.0171	.1219	-.2958	.4596	-.4879	.2087	.0418	-.2761	.1115
K ₂ O	.0874	-.1340	.3970	.4477	.3098	.3614	-.2729	-.3779	-.2832	.1398
P ₂ O ₅	-.0470	-.3980	-.0958	.5492	-.0141	-.1512	.3770	-.0941	.2931	.4292
CO ₂	-.1105	.4074	-.3661	.3205	.1796	-.2061	-.2624	-.0902	-.4341	.2902
H ₂ O	-.1925	.2789	.4096	.3579	-.1487	.0150	.2918	.6338	-.0224	-.2106

	1	2	3
SiO ₂	.1233	.2606	.7270
TiO ₂	-.4592	-.4255	.3916
Al ₂ O ₃	-.0581	-.0272	.2087
Fe ₂ O ₃	.0716	.4125	.0162
FeO	.1018	.5214	.0480
MnO	.3582	-.2484	.1331
MgO	-.0723	.2354	.3108
CaO	.5214	-.113	.2992
Na ₂ O	.4570	-.0080	.1857
K ₂ O	.1949	-.1640	.0887
P ₂ O ₅	.2307	.1428	-.1040
CO ₂	-.2052	.3280	-.1052
H ₂ O	-.0867	.1545	.0628

	1	2	3
1	3.6505	28.0812	28.0812
2	2.1838	16.7988	44.8800
3	1.6370	12.5926	56.4726
4	1.4219	10.9381	68.4107
5	1.1486	8.8356	77.2463
6	1.0892	8.3784	85.6247
7	.6409	4.9300	90.5547
8	.3551	2.7316	93.2863
9	.2942	2.2629	95.5492
10	.2424	1.8643	97.4136
11	.1885	1.4500	98.8636
12	.1224	.9414	99.8050
13	.0253	.1949	100.0000

Column 1 = Eigenvalues Column 2 = Percent of Trace
Column 3 = Cumulative Percent of the Trace

Principal Axis Matrix - Columns = Eigenvectors,
Rows = Variables

set in an attempt to see if the effective dimensionality of the 13 variable data set (the oxides) could be reduced with little loss of information. Table II-3 cites the results of this analysis, the eigenvectors being principal components and the eigenvalues being the variances of the new variables. Thus, the new variable $Y_1 = 0.4534 \text{ SiO}_2 - 0.4811 \text{ TiO}_2 - 0.1781 \text{ Al}_2\text{O}_3 - 0.2658 \text{ Fe}_2\text{O}_3 - 0.1925 \text{ H}_2\text{O}$ accounts for 28.0812% of the aggregate variance of the 13 oxides. Similarly the second component $Y_2 = -0.1846 \text{ SiO}_2 - 0.2451 \text{ TiO}_2 + 0.2789 \text{ H}_2\text{O}$ accounts for 16.7099% of the total variance and is uncorrelated with Y_1 . Not that the first six principal components account for 85.6247% of the total variance. These first six components can be used to summarize the variation in the 13 oxides with relatively small loss in information.

The principal component scores for the first principal component are plotted successively against the next four in Fig. II-10 and a clustering of the specimens attempted for each of the resulting scatter diagrams. Only one or two of the amphibolites can be distinguished in a single diagram since any two components account for less than half of the total variance. In retrospect it might have been desirable to attempt a cluster analysis based on the first six or seven principal components because these components are uncorrelated and a better clustering might have resulted.

The cluster analysis has confirmed the division of the amphibolites into three groups, one of which may be divided into two sub-groups. These groups correspond to the three groups of

FIG II-10. Principal component scores and clustering for amphibolites from Sherridon, Man.

a) F1 vs F2; b) F1 vs F3

- Bi Amph.
- Frag. Amph.
- Masv. Amph. - high TiO_2
- Masv. Amph. - low TiO_2

- Masv. Amph. - high TiO_2
- Masv. Amph. - low TiO_2
- Frag. Amph.
- Bi Amph.

FIG. II-10. cont.

c) F1 vs F4; d) F1 vs F5

amphibolites classified in the field. The results of the statistical techniques used in the early part of this section support the validity of the field classification and origin of the rock types as previously suggested.

The variation of SiO_2 , TiO_2 , CaO , and MgO among groups can be attributed to factors such as weathering, incorporation of terrigenous material in volcanoclastics during transport, cementation of volcanoclastics and sediments by carbonate. The variation between groups can be attributed to primary magmatic processes in the case of the volcanic rocks and clastic variation in the case of sediments. Environmental factors which contributed to the differences between groups and other factors such as local metasomatism may also be responsible for differences within groups.

II-5 Sedimentary Rocks of the Sherridon Group

II-5-a Introduction

The sedimentary rocks of the Sherridon Group have been classified into five informal formations. Each formation is characterized by a particular rock type, in which interbeds of broad compositional range are present. Gradation between rock types is common. Meta-lithic arenites vary from clean quartzofeldspathic rocks to varieties with 25 percent biotite and garnet. The meta-lithic arenites are transitional with meta-pelites and meta-carbonate rocks. The transition with the meta-carbonate rocks has produced a large variety of calc-silicate gneisses.

As a whole, the composition of the siliceous rocks is not aluminous and aluminosilicate minerals are rare. These rocks are

also low in alkali content so that they probably did not contain much muscovite at medium grade. A reconstruction of the sedimentary process which formed this succession of rocks will be discussed in the final chapter of this thesis.

II-5-b Biotite Gneiss (Lithic Arenite): Unit 5

The biotite gneiss of the Sherridon Group has been previously described by Wright (1929), Robertson (1953), and Pollock (1964). The gneiss is medium grained, weathers white to gray, has a distinctive surface texture of quartz ridges and a planar orientation of micas. The biotite gneiss forms the dominant rock type of the Sherridon area and constitutes about 80 percent of the rocks of the Sherridon structure. It is the characteristic rock type of the lower (B) and upper (E) arenite formations.

The biotite gneiss of the lower arenite (B) formation has a sharp contact with the rocks of the underlying calc-silicate formation. The rocks near the bottom have a compositional layering on the scale of tens of centimeters (Plate II-1a). This layering gradually becomes thicker upward to beds tens of metres thick at the top of the formation. Laminae of heavy minerals and crossbeds (Plate II-1b) are found in the parts of the formation with thick bedding. A gradual decrease in mafic content is accompanied by an increase in bed thickness. Biotite gneiss at the base contains an average of 25 percent biotite and garnet whereas the thickly bedded gneisses contain only ten percent biotite and garnet. Layers of amphibolite and calc-silicate gneiss are found in the biotite gneiss.

Plate II-1

- A) Layering in the lower part of formation B, near the Cambrian Hotel, Sherridon. Bedding is at a scale of tens of centimetres and individual beds are distinguished by variable mafic content or separated by calcareous laminae.
- B) Cross-beds in the upper part of formation B on Singing Lake. Compositional layering is not evident as in Plate II-1a and the gneiss has a much lower mafic content.
- C) Layering in the calc-silicate gneiss of formation A on Singing Lake. Differential weathering resulting from the variable distribution of carbonate minerals is well developed.

PLATE II-1. Textures in Meta-sedimentary Rocks of the
Sherridon Area

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Rocks of the upper arenite formation were not examined during the mapping for the thesis but Froese (pers. comm., 1979) has indicated that the rocks are very similar to the thick bedded gneiss in the rocks of the upper part of the lower arenite (B) formation.

Quartz and feldspar are the most common minerals of this rock type. Quartz invariably constitutes more than 50 percent of the rock and forms a granoblastic mosaic of weakly sutured grains with feldspar. The mineral grains show little strain except in a few localities where strain shadows are well developed in quartz. Plagioclase constitutes 20 to 30 percent of the rock. Many of the plagioclase grains are weakly sericitized. They show well-developed polysynthetic twinning and have a compositional range of $An_{25}-An_{40}$, of which sodic compositions are the most common. Small amounts of K-feldspar, invariably microcline, were found in two-thirds of the thin sections examined.

Biotite, the dominant ferromagnesian mineral was found to compose about ten percent of this gneiss except in the thin bedded gneiss at the base of formation (B) where it averages 20 percent. This mafic biotite gneiss has been given the sub-unit designation: Unit 5c. The biotite flakes show a well-developed planar fabric and exhibit simultaneous extinction. They are typically brown, although in about 20 percent of the thin sections examined, the pleochroic formula is α : light green; $\beta=\gamma$: dark green. The colour of the biotite can be related to the ferrous iron content of the rock. Samples with a large FeO/Fe_2O_3 ratio contain green biotite.

Pink garnet is a common constituent of this rock, forming up to five percent. There does not seem to be a correlation between biotite content and garnet content. Garnet occurs as both sub-hedral porphyroblasts and anhedral aggregates up to 2 cm in diameter, mantled in many cases by plagioclase. The biotite foliation is bent around garnets showing that tectonism outlasted garnet growth.

Sillimanite, in individual grains (lithologic Unit 5a), fibrous mats or nodules (lithologic Unit 5b), was found in discrete layers in the biotite gneiss. Comparison of the average chemical analysis of the sillimanite-bearing rocks with those of the other biotite gneisses of the Sherridon Group (Table II-4) shows that the various types of biotite gneiss are similar except for a relatively low alkali content in the sillimanite-bearing rocks. There is no difference between the composition of the sillimanite and sillimanite nodule-bearing rocks (Appendix A), suggesting that the development of sillimanite in nodules rather than individual grains is kinetic in nature and may possibly be related to the conditions which cause garnets to grow in aggregates rather than discrete porphyroblasts.

Hornblende is a common constituent of the biotite gneiss, found in association with diopside and calcite. For this reason, the hornblende-bearing variety of biotite gneiss is discussed with the other calc-silicate rocks later in this chapter. Other minerals present in the biotite gneiss include minor amounts of cordierite, magnetite, pyrite, pyrrhotite, and chalcopyrite.

TABLE II-4 Average Chemical Composition of Sedimentary Rocks from Sherridon and Vicinity

	1	2	3	4	5	6
MEAN						
SiO ₂	74.43	77.08	80.40	67.64	62.47	65.65
TiO ₂	0.21	0.19	0.16	0.37	0.63	1.08
Al ₂ O ₃	11.77	11.04	10.70	14.00	16.67	11.53
Fe ₂ O ₃	0.46	0.18	0.13	1.14	0.63	1.37
FeO	3.05	3.04	0.81	4.82	5.73	3.84
MnO	0.09	0.04	0.04	0.12	0.08	0.30
MgO	1.92	2.09	0.70	2.07	2.87	2.25
CaO	2.24	1.46	2.14	3.22	3.69	5.07
Na ₂ O	1.74	0.86	1.96	1.86	2.47	1.42
K ₂ O	2.36	1.75	2.19	3.07	2.24	2.17
P ₂ O ₅	0.08	0.06	0.06	0.13	0.51	0.16
CO ₂	0.52	0.26	0.31	0.44	0.23	1.11
H ₂ O	1.13	1.30	0.74	1.10	1.13	0.90
STANDARD DEVIATION						
SiO ₂	2.15	3.42	1.85	1.71	0.47	16.78
TiO ₂	0.07	0.05	0.06	0.07	0.12	3.31
Al ₂ O ₃	1.25	1.89	1.31	1.31	0.53	2.86
Fe ₂ O ₃	0.48	0.24	0.14	0.65	0.33	2.26
FeO	1.17	1.68	0.45	1.33	0.12	2.14
MnO	0.04	0.01	0.04	0.07	0.01	0.75
MgO	0.94	0.87	0.40	0.53	0.12	1.25
CaO	1.09	1.08	0.82	0.99	0.39	1.77
Na ₂ O	1.07	0.70	1.09	1.28	0.74	1.19
K ₂ O	1.29	0.77	1.09	1.66	0.14	1.24
P ₂ O ₅	0.03	0.02	0.02	0.03	0.13	0.23
CO ₂	0.33	0.31	0.16	0.30	0.17	0.57
H ₂ O	0.43	0.64	0.57	0.29	0.25	0.50
SAMPLES	33	5	7	10	3	15
	EXPLANATION OF TABLE					
	1-Biotite gneiss, Sherridon					
	2-Sillimanite-bearing biotite gneiss, Sherridon					
	3-Quartzofeldspathic gneiss, Sherridon					
	4-Nafic biotite gneiss, Sherridon					
	5-Biotite garnet schist, Sherridon					
	6-Nokomis gneiss, File Lake, Bailes (1979)					

The chemical composition of the siliceous sedimentary rocks have been plotted on Fig. II-11 and Fig. II-12, which are the only diagrams available for this purpose from the literature. In Fig. II-11, a $\text{Na}_2\text{O} - (\text{Fe}_2\text{O}_3 + \text{MgO}) - \text{K}_2\text{O}$ plot, the compositional range of the biotite gneisses occupies the entire lithic arenite field, with some overlap into the graywacke and arkose fields. Fig. II-12 which plots $\text{Log} (\text{Na}_2\text{O}/\text{K}_2\text{O})$ vs $\text{Log} (\text{SiO}_2/\text{Al}_2\text{O}_3)$ shows that most of the analyses of the biotite gneiss lie in the lithic arenite and arkose fields with some overlap of the graywacke field.

In inferring a protolith for these rocks, it should be recalled that they are chemically similar to the rocks of the Missi Group (Mukherjee, 1974). Mukherjee has shown that the rocks of the Missi Group contain little or no matrix and consist on the average of 45 percent rock fragments, 40 percent quartz and 15 percent plagioclase. Bailes (1979) has deduced a similar protolith for the Missi rocks of the File Lake area and has traced them into metamorphic equivalents of the Sherridon Group. For these reasons, the protolith of the Sherridon Group biotite gneiss is considered to have been a lithic arenite. Fig. II-12 also shows that the analysis of the biotite gneiss plot over a broad range of Na_2O and K_2O content but plot in discrete increments along the $\text{Log} (\text{SiO}_2/\text{Al}_2\text{O}_3)$ axis with respect to their field classification. This change probably reflects the increase in Al_2O_3 with increasing biotite content, the basis of the field classification.

II-5-c Quartzofeldspathic Gneiss (Arenite): Unit 6

The quartzofeldspathic gneiss unit, which includes those gneisses containing less than two percent mafic minerals, is rare.

FIG. 11-11. Chemical composition of meta-sandstones in the Sherridon area; Method of plot after Blatt et al. (1972)

FIG 11-12. Chemistry of some meta-sediments from Sherridon, Man.; diagrams after Pettijohn et al. (1972)

It is spatially related to the calc-silicate gneiss; Fig. II-2 shows that every occurrence of quartzofeldspathic gneiss large enough to be plotted is found near marble or calc-silicate. The quartzofeldspathic gneiss is interlayered with the major occurrence of marble near Found Lake but these layers are too thin to be shown.

In the vicinity of the Sherridon deposit quartzofeldspathic gneiss was found to grade both upward and downward into the biotite gneiss. Successive layers of biotite gneiss, 10 to 20 cm thick, contain progressively less biotite and ultimately become biotite-free. Near Found Lake and Camp Lake, the quartzofeldspathic gneisses are nearly massive and show sharp contacts with adjacent rocks.

The quartzofeldspathic gneiss consists almost entirely of quartz and feldspar in a granoblastic mosaic of weakly to moderately sutured grains which show little sign of strain. The grain size varies from medium to coarse and the rocks are invariably white. These rocks have been previously described as ortho-quartzites (Robertson, 1953) although they do not have a quartz content of over 75 percent.

Grains of both twinned and untwinned plagioclase ($An_{25}-An_{35}$) are present. Microcline, only slightly less common than plagioclase, was found in almost all of the thin sections examined. Mafic minerals recognized are biotite, hornblende and diopside. In some layers of the gneiss, minor amounts of calcite were found, especially near contacts with calc-silicates.

The gradational relationship with the biotite gneiss suggests they are a variation of this type of gneiss derived either through

a change in provenance or transportational processes. The relationship of the chemistry of the quartzofeldspathic gneiss and biotite gneiss, as plotted on Fig. II-11 and Fig. II-12, shows an affinity between the various gneisses and also suggests a related provenance.

II-5-d Quartzitic Gneiss (Chert): Unit 7

Layers of quartzitic gneiss were found in the carbonate rocks of the calc-silicate and marble formations. They consist of thin (2 to 3 mm) laminae of quartz with pronounced lateral continuity, which alternate with even thinner laminae of mica or calc-silicate minerals or pyrrhotite.

In the calc-silicate formation, the quartzitic gneiss grades downward into the underlying calc-silicate rock, as intervening layers of calc-silicate minerals become thicker. In some quartzitic layers pyrrhotite may constitute up to 30 percent and may persist down into the calc-silicate rock even where it is devoid of quartz laminae. This unit is in part overlain by a flow of basaltic composition.

With the exception of the minor Fe-sulphides and calc-silicate between quartz laminae these rocks consist of pure quartz. It is doubtful that orthoquartzites would form repeated laminae a few mm thick. Consequently, these rocks are probably derived from chert. The large amount of Fe-sulphides associated with these rocks suggests a mode of origin as a volcanogenic iron formation and this idea is evaluated more fully in the chapter on economic geology.

II-5-e Biotite-Garnet Schist (Graywacke): Unit 8

The biotite-garnet schist of the Sherridon area comprises the rocks of the graywacke formation (D) almost exclusively. This rock is very similar to the rocks of the Nokomis Group (Robertson, 1953; Bailes, 1979), and for this reason was thought to be a dome of such rocks pushed up through the rocks of the Sherridon Group near Cree Lake (Davies, 1962). No evidence could be found to support this view and it is concluded that the biotite-garnet schist is a conformable sedimentary layer of the Sherridon Group.

The most striking field feature of this rock is its extreme uniformity. It shows very little compositional layering and includes only a few, thin, rusty pelitic intercalations. The biotite-garnet schist consists of a fine grained groundmass of quartz and feldspar in which is developed a planar-oriented fabric of biotite flakes and garnet porphyroblasts. Quartz and plagioclase form a weakly sutured, granoblastic mosaic and constitute about 70 percent of the rock. Plagioclase has the composition $An_{30}-An_{35}$. Biotite is a brown variety and forms a well-developed lepidoblastic fabric. Garnet porphyroblasts are subhedral and vary in size from 3 to 5 mm. They are uniformly distributed throughout the rock.

The average chemical composition of this rock is similar to that of graywackes of the Nokomis Group (Table II-4). Chemical analyses have been plotted on Fig. II-11 and Fig. II-12 and in both instances were found to be in the graywacke field.

II-5-f Meta-Pelite: Unit 9

Pelitic rocks, like quartzofeldspathic rocks, are rare in the Sherridon area. Only four occurrences of pelitic rock are large enough to be shown on Fig. II-2 and three of these lie on the same stratigraphic horizon. The presence of distinct layers of pelitic rock has not been mentioned in previous studies of the Sherridon area.

By far the largest occurrence of pelitic rocks is in the vicinity of Singsing Lake where a layer reaches a maximum thickness of 100 m and persists for a strike length of several km. Internal layering cannot be recognized. The contacts of this unit are fairly sharp and it grades laterally into a mafic biotite gneiss, and thence into the common variety of biotite gneiss.

Pelitic beds of 2 to 10 cm thickness are interlayered with mafic clastic rocks above the east orebody of the Sherritt Gordon deposit. Here, the pelitic unit grades continuously downward into calc-silicate and marble.

The mineralogy and textures of the pelitic rocks are similar to those of the biotite gneiss except for generally larger grain size and amount of aluminous constituents. Needles of sillimanite reach several cm in length and garnets reach 3 to 4 cm in diameter. Cordierite was also recognized although not in this section. The texture is that of a granoblastic mosaic with a very crude planar orientation of biotite.

Quartz and plagioclase, which form the matrix of this rock, constitute about 50 to 10 percent respectively. Plagioclase has a composition of $An_{25}-An_{35}$. No K-feldspar was recognized in this

section. Biotite averages 20 percent of this rock and garnet and sillimanite each average 15 percent. Other minerals present are cordierite, magnetite, pyrite, pyrrhotite, and chalcopyrite. The pelitic rocks invariably contain a few percent sulphide minerals and for this reason have a rusty appearance in outcrop. The sulphide mineralogy is discussed in the chapter on economic geology.

Chemically, the pelitic rocks contain more FeO and MgO than the biotite gneiss. They also contain very little CaO and Na_2O (Appendix A; Fig. II-12). This suggests that the pelitic rock was developed from a feldspar-poor, clay-rich precursor.

II-5-g Calc-Silicate Gneiss (Calcareous Rocks): Unit 11 and Unit 11a

The calc-silicate gneisses are by far the most complex and varied rocks of the Sherridon area. Calcareous varieties of all the sedimentary and volcanoclastic rocks with the exception of the graywacke have been recognized. All varieties of the calc-silicate gneiss were seen to grade into calcite-bearing rocks, confirming a calcareous origin for these gneisses (Plate II-1c).

The most common calc-silicate gneisses are those related to the common biotite gneiss. They occur both as layers within the biotite gneiss and as layers of calcareous biotite gneiss. Compositionally, they are similar to the biotite gneiss except for additional CaO . The layers may be quite mafic and show differential weathering (Plate II-1d), in which case, they are difficult to distinguish from orthoamphibolites. These biotite gneisses grade into the assemblages with hornblende, diopside and very occasionally calcite.

An aluminous calc-silicate overlies part of the east ore-body. The addition of carbonate minerals to the original aluminous clays has resulted in a rock with grossularite, epidote, diopside, and hornblende. The grossularite occurs as subhedral porphyroblasts which are brownish-orange in thin section. The epidote is dark green and prismatic. It shows the pleochroism α : colourless; β - γ : greenish-yellow; and has a $2V_{\alpha}$ of 80° . The presence of epidote was confirmed by single crystal X-ray techniques. Mineralogically this is the most complex rock in the Sherridon area.

Chemical analyses of the rocks interpreted as having been produced by the metamorphism of calcareous arenites have been plotted on Fig. II-12b. The field produced by plotting the assemblages with calc-silicate minerals overlaps the field produced by plotting the biotite gneiss (Fig. II-12a) suggesting the calc-silicate rocks were formed by metamorphism of calcareous sediments.

Calcareous volcanoclastic layers are common. Fragment-bearing amphibolites can be seen to grade into diopside- and calcite-bearing amphibolites. Scapolite is very common in these rocks and does not coexist with plagioclase. Where sufficient calcite is present in these rocks, they show differential weathering.

II-5-h Marble (Arenaceous Limestone): Unit 12

The marbles in the Sherridon area are an impure variety and together with calc-silicate gneiss and quartzofeldspathic gneiss constitute the rocks of the arenaceous limestone, formation (C).

The marbles form about 50 percent of this formation, but are otherwise rare. Individual layers of marble range from one cm to tens of m in thickness, although the thick layers are invariably

well laminated. The distribution of calc-silicate minerals is extremely variable within individual layers and calcite content ranges up to a maximum of 70 percent. All marbles show pronounced differential weathering.

Texturally, the marble consists of a medium to coarse grained granoblastic polygonal mosaic of calcite, diopside, plagioclase and minor amounts of mica and quartz. The diopside is a dark green variety and occurs as stubby, subhedral porphyroblasts. It shows the pleochroism α = β : medium green; γ : pale green; has a $2V_Y$ of 50° ; and an extinction angle $\gamma_{\lambda 2}$ of 40° . Plagioclase is twinned and has a calcic composition ($\approx \text{An}_{70}$). These minerals invariably account for more than 90 percent of the marble. Other minerals present in small amounts are phlogopite, quartz, tremolite, green spinel, sphene, graphite, and pyrrhotite.

The chemical composition of these rocks (Appendix A) is as highly variable as the mineralogical composition. Of greatest interest is the MgO/CaO ratio which is small indicating that the initial carbonate mineral which formed these rocks was calcite assuming a minimum of diagenetic alteration. These rocks were originally arenaceous limestones and quartzofeldspathic arenites.

II-6 Volcanic Rocks in the Sherridon Group

II-6-a Introduction

Volcanic rocks were recognized at the outset of exploration in the Sherridon area, although there was some controversy as to which of a number of types of amphibolite was of volcanic or sedimentary origin (Harrison, 1951). Earlier in this chapter, specific

criteria were developed for the identification of volcanic rocks in the Sherridon area.

For the most part, volcanic rocks consist of mafic flow or volcanoclastic rocks which have been metamorphosed into amphibolite. A small body of felsic volcanic rocks was recognized in the vicinity of Singing Lake, and is the first occurrence of felsic volcanic rocks recognized in the rocks of the Sherridon Group. The body occurs in a thick layer of mafic volcanoclastic rocks.

Volcanic rocks are found in the rocks of the lower formation (A) and (B), but no higher in the stratigraphic succession. They usually form independent horizons and in very few instances do more than a few volcanic layers vertically overlap each other. The volcanic rocks constitute less than ten percent of the stratigraphic succession in the Sherridon area and a volcanic pile or volcanic feeder system such as a dyke complex or plug is not apparent. From these observations, it is inferred that the source of the volcanic rocks in the Sherridon area was outside of the Sherridon structure as it is presently preserved.

Alteration of volcanic layers by surficial weathering or by groundwater, dealt with later in this chapter, has led to a depletion of CaO and the development of a cumingtonite-bearing assemblage during metamorphism.

II-6-b Amphibolites (Mafic Volcanoclastic and Flow Rocks): Units 13 and 14

The mafic volcanic rocks of the Sherridon area have been divided into volcanoclastic and flow rocks. The division into these rock types is based on field criteria because both rock

types have similar chemistry and mineralogy. The rocks interpreted as flows occur in individual layers which vary in thickness up to a maximum of 200 m. They are massive and uniform and have sharp contacts with adjacent rocks. An interpretation as flows rather than hypabyssal or subvolcanic intrusions is preferred because these rocks do not exhibit cross-cutting relationships and were not seen to contain xenoliths. In a number of localities, they contain features which have been interpreted as pillow rims (Plate II-2a). These relict rims consist of curvilinear concentrations of hornblende and garnet. Harrison (1951) and Robertson (1953) both report having seen pillows in amphibolites of the Sherridon Group. Nevertheless, some of the completely featureless amphibolites might be subvolcanic intrusions.

Flows occur at the base of the Sherridon Group, where they are intercalated with rocks of the basal calc-silicate formation. Flows overlap each other and are also separated by intervals of sedimentary material. However, since many of these sedimentary interlayers are very thin, this relationship could not be shown on the geologic map of the thesis area (Fig. II-2), where the flow rocks are shown as a single continuous layer.

Flows in the arenites lying above the calc-silicate formation occur as individual layers. In some areas, small flows were found within mafic volcanoclastic layers but because of the scale of mapping could not be represented on Fig. II-2.

The rocks which form the volcanoclastic unit exhibit both gradational and sharp contacts. The rocks of this unit are persistently layered with individual layers varying in thickness from a few centimetres to tens of metres. Arenaceous or pelitic interbeds

Plate II-2

- A) Textures suggesting tectonically deformed pillows in an amphibolite near the Cambrian Hotel, Sherridon.
- B) Fragmental rocks from an outcrop northeast of the West Shaft. Fragments range from felsic to intermediate in composition and vary in size from 1 to 20 cm. The matrix is mafic in composition.
- C) Felsic fragmental rocks photographed on Singsing Lake. The fragments are angular and cemented by a calcareous matrix.
- D) A felsic block from an outcrop on Singsing Lake with textures suggesting vesicles.

PLATE II-2. Textures in Meta-volcanic Rocks of the
Sherridon Area

are common. In some places, the volcanoclastic rocks have been cemented by carbonate and have become gradational with calc-silicate rocks in a manner similar to the meta-lithic arenites.

Fragments of intermediate and felsic composition have been recognized (Plate II-2b). The first identification of volcanoclastic fragments in the Sherridon Group was by Bateman and Harrison (1946). Fragments range in size up to about 25 cm in the less deformed parts of the volcanoclastic layers. Many of the felsic fragments are rounded and appear similar in lithology to the enclosing arenites which suggests that some of the volcanoclastics were transported into the area as mudflows or lahars. Amphibolites containing these fragments are poorly layered. In areas with small scale layering and a lack of recognizable fragments the mode of origin was probably as an air-fall tuff of fine particles into the waters of the ancient Sherridon basin.

Mineralogically, the mafic flows and volcanoclastic rocks are similar. They consist of a granoblastic polygonal mosaic of hornblende and plagioclase in subequal proportions with minor quartz, biotite and garnet. The hornblende is found in euhedral grains 2 to 3 mm in diameter which exhibit the pleochroic formula α : yellowish-green; β : dark green; γ : dark bluish-green. Plagioclase is equigranular, 2 to 3 mm in diameter, and shows sutured boundaries with other plagioclase grains. It has the compositional range $An_{35}-An_{60}$ although compositions near An_{50} are most common.

Quartz is a common varietal mineral, forming up to 15 percent of the amphibolites. Pink garnet which forms both subhedral grains and anhedral masses is also very common. The subhedral

grains vary from 3 to 10 mm in diameter; the anhedral masses are up to 2 to 3 cm in diameter. A rim of plagioclase 1 to 2 mm thick surrounds many of the garnet porphyroblasts. Analyses of these garnets indicates they are almandine rich (Ribertson, 1953). They are particularly common in the features interpreted as pillow rims where garnet and hornblende may constitute 80 percent of the rock. The garnet is also common in biotite-rich layers in the rocks interpreted as volcanoclastics.

Biotite found in the amphibolites is the common brown variety. In the rocks interpreted as flows, it occurs as small (< 1 mm) flakes and may constitute as much as three percent of the rock. In the volcanoclastic rocks biotite may constitute up to ten percent of some layers, in which it develops a crude foliation. In areas where the volcanoclastic rocks have been cemented by carbonate, diopside and scapolite are present. These minerals increase in abundance as the volcanoclastic rocks grade into calc-silicates where calcite may also be present. Other minerals identified are magnetite, pyrite, and pyrrhotite.

The chemistry of the volcanic rocks has been plotted on Fig. II-13 to Fig. II-17 and the chemical composition of different units has been compiled in Table II-5. Fig. II-13a, an SiO_2 vs alkali diagram shows that the mafic volcanic analyses form a discrete group, probably derived from a single magmatic series which is subalkaline in nature, although a few points fall just inside the alkalic field. Fig. II-14a, a normative plagioclase composition vs normative colour index diagram shows that with a single exception all of the mafic rocks plot in the basalt field.

FIG. II-13. Plot of Sherridon meta-volcanic rocks on:
a) SiO₂ vs Na₂O + K₂O and b) SiO₂ vs-
Al₂O₃ Diagrams.
Diagrams after Irvine and Baragar, (1971)

FIG. II-14. Normative Plagioclase vs Normative Colour Index Diagrams; after Irvine and Baragar (1971);
a) volcanic rocks, b) plutonic rocks

FIG. II-15. Plot of Discriminant Functions; after Pearce (1976)
 WPB-within plate basalt, OFB-ocean floor basalt,
 CAB-calc-alkaline basalt, LKT-island arc thoeiite,
 SHO-shoshonite.

FIG. II-16. A FM Diagram; after Irvine and Baragar (1971)

ThB - field of tholeiite basalt

CAB - field of calc-alkaline basalt

FIG 11-17. Chemical composition of Sherridon volcanic rocks compared with some recent trends plotted by Miyashiro (1974): a) SiO_2 vs FeO/MgO ; b) FeO vs FeO/MgO .

FIG. II-17. cont.

c) TiO_2 vs FeO/MgO

2

TABLE II-5 Average Chemical Composition of Volcanic Rock from Sherridon and Vicinity

	1	2	3	4	5	6	7	8
MEAN								
SiO ₂	52.01	51.73	54.35	51.6	44.65	48.81	67.74	64.4
TiO ₂	0.56	0.64	0.70	0.81	2.32	2.47	0.70	0.44
Al ₂ O ₃	16.15	15.47	16.07	16.40	15.57	14.41	14.68	16.3
Fe ₂ O ₃	1.79	2.92	3.09	10.16	3.13	12.82	0.12	-0.4
FeO	9.26	9.32	9.68		12.82	12.00	1.40	3.1
MnO	0.19	0.21	0.20		0.25		0.02	
MgO	5.46	6.28	5.15	5.92	6.07	5.96	2.62	2.2
CaO	8.94	9.10	7.01	10.27	9.36	10.05	5.89	3.2
Na ₂ O	2.57	1.92	2.81	2.03	1.75	2.90	3.92	4.0
K ₂ O	0.70	0.46	0.57	0.43	0.37	0.95	1.69	2.4
P ₂ O ₅	0.14	0.14	0.17		0.26		0.59	
CO ₂	1.59	0.83			1.62		0.64	
H ₂ O	1.39	1.32			1.60		0.68	
STANDARD DEVIATION								
SiO ₂	1.60	2.62		2.32	4.20	2.40	3.55	
TiO ₂	0.09	0.07		0.39	0.73	0.83	0.35	
Al ₂ O ₃	1.28	0.83		2.15	1.39	2.16	1.21	
Fe ₂ O ₃	0.92	1.03			1.23		0.12	
FeO	0.94	0.69		2.33	2.46	1.78	0.24	
MnO	0.03	0.02			0.05		0.01	
MgO	1.30	1.93		1.84	0.54	1.25	1.29	
CaO	1.76	0.87		1.29	0.82	1.04	3.26	
Na ₂ O	1.01	0.62		0.47	0.83	0.68	0.97	
K ₂ O	0.48	0.13		0.28	0.05	0.48	0.98	
P ₂ O ₅	0.09	0.06			0.06		0.41	
CO ₂	2.13	0.81			0.94		0.27	
H ₂ O	0.50	0.35			0.34		0.26	
# SAMPLES	17	12	12	37	6	75	5	

EXPLANATION OF TABLE

- 1-Fragmental amphibolite, Sherridon
- 2-Low TiO₂ massive and pillowed amphibolite, Sherridon
- 3-Basalt, Flin Flon (from Stauffer et al, 1975)
- 4-Modern island arc tholeiite basalt (from Pearce, 1976)
- 5-High TiO₂ massive and pillowed amphibolite, Sherridon
- 6-Modern continental tholeiite basalt (from Pearce, 1976)
- 7-Meta-felsic volcanic rocks, Sherridon
- 8-Modern calc-alkaline dacite (from Condie and Moore, 1976)

In order to determine from which of the subalkaline series these rocks were derived, the chemical analyses were plotted on an Al_2O_3 vs SiO_2 diagram (Fig. II-13b) and on an AFM diagram (Fig. II-15a). Both of these diagrams indicate that the mafic volcanic rocks of the Sherridon area belong to the tholeiitic or iron-enrichment series. Plotting the analyses on Pearce discriminant function diagrams (Fig. II-16) shows that with few exceptions these rocks belong to the island arc tholeiitic series. The rocks which do not plot in the island arc tholeiitic field plot in the continental basalt field and correspond in part to the group of high TiO_2 basalts found in the cluster analysis.

These high TiO_2 basalts occur as flows generally intercalated with low TiO_2 flows, volcaniclastics and calc-silicates at the base of the Sherridon Group and the flow immediately overlying the Sherritt Gordon East and West orebodies. The other mafic volcanic and intrusive rocks are of the typical low- TiO_2 island arc tholeiite variety. Comparison of the mafic volcanic flow rocks of the Sherridon area with modern tholeiites on Fig. II-17 (after Miyashiro, 1974) shows two trends. One is the typical island arc tholeiite trend and the other is a trend more enriched in FeO corresponding to the high TiO_2 rocks.

Kuno (1968) has shown that considerable variation can be expected in the rocks from a single magmatic source, particularly as regards TiO_2 and FeO content. This variation is the result of variation in $f\text{O}_2$ and pH_2O which control fractional crystallization (Osborn, 1962). The association of island arc tholeiite basalts and tholeiite-related alkalic basalts has been reported from a

number of areas (Sugisaki et al, 1972; and Miyashiro and Shido, 1975).

The rocks of the Amisk Group in the Flin Flon belt show an island arc tholeiite trend (Stauffer et al, 1975; Fox, 1976) similar to the one found in the volcanic rocks of the Sherridon Group. Fox (1976) has shown that the volcanic rocks of the Flin Flon area progress in composition from olivine normative to quartz normative. This progression should in no way be compared to the change from high TiO_2 to low TiO_2 rocks in the Sherridon area. The volcanic rocks of the Sherridon Group should only be correlated with later stages of Amisk volcanism to the rocks of which they are chemically similar. Minor volcanic rocks have been recognized in the sediments of the Missi Group near Snow Lake (Shanks and Bailes, 1977), and mafic sills are also abundant (Harrison, 1951). Appendix E contains the normative mineralogy of the volcanic rocks of the Sherridon area; the unaltered low TiO_2 basaltic flows are quartz normative as are the basalts of the Flin Flon area.

Considering the excellent correlation of the composition of the sediments of the Sherridon Group with the sediments of the Missi Group and the similarity of the analyses of volcanic rocks of the Flin Flon and Sherridon areas, it is highly probable that the volcanics of the Sherridon Group are lateral equivalents of volcanic rocks occurring in the Flin Flon belt.

II-6-c Meta-Felsic Volcanics: Unit 15

Only one small area of felsic volcanic rocks was found during the mapping of the Sherridon area. This is a unit of white to

buff weathering pyroclastic rocks in the vicinity of Singing Lake. The felsic rocks occur for a distance of about 1,000 m in a layer of mafic volcanoclastic rocks and are 100 m thick. Fragments vary in size upward from about 10 cm and some are large blocks containing vesicles (Plate II-2c, 2d). The fragments are angular in nature and form an interlocking network about which is a matrix of fine mafic volcanoclastic material cemented by carbonate.

The mineralogy of these felsic rocks is very simple, consisting of about 70 percent quartz with subordinate amounts of plagioclase, biotite and a brown amphibole. Texturally, the rocks consist of a seriate mosaic of sutured quartz grains having a maximum size of five mm, in which occur randomly oriented grains of plagioclase, biotite and amphibole. Both twinned and untwinned plagioclase is present in the range $An_{25}-An_{30}$. Biotite is brown and occurs as fine flakes (1 mm in length) oriented along grain boundaries.

Porphyroblasts up to five mm in diameter of an amphibole having the pleochroic formula α : colourless; $\beta=\gamma$: light yellow brown are evenly distributed throughout the rock. The amphibole has an extinction angle $\gamma_A Z$ of 18° , and a $2V_\alpha$ of 85° . Based on the optical properties, the amphibole is probably intermediate in composition between common hornblende and pargasite. Other minerals recognized in the meta-felsic volcanic rocks are sphene, graphite, and pyrrhotite.

One small meta-rhyolite dyke was found intruding the calc-silicate rocks near the north end of Sherlett Lake. The dyke is three m wide where it crops out on a cliff face at an angle of about 25° to the stratigraphic trend. It is identical with the meta-felsic

volcanic rocks except for the absence of amphibole. The dyke contains biotite as its only mafic constituent.

The chemistry of these felsic volcanic rocks has been plotted on Fig. II-13 to Fig. II-16. An alkali vs SiO_2 diagram (Fig. II-13a) shows that the felsic volcanic rocks are subalkaline in nature. The normative plagioclase vs normative colour index diagram (Fig. II-14a) shows that these rocks fall into the andesite and dacite fields. However, the most felsic of the points on the diagram comes from the dyke and it is suggested that all the volcanic rocks should be of dacitic composition similar to this analysis. The spread could result from contamination by calcite cement, the CaO component of which was incorporated into the mineral assemblages during metamorphism while the CO_2 was evolved in the fluid phase. The high An contents of such felsic rocks and the presence of associated calc-silicate rocks supports this conclusion.

Fig. II-13b and Fig. II-16a suggest that the felsic volcanic rocks belong to a calc-alkaline magmatic series. The average analysis of the Sherridon Group felsic volcanic rocks is similar to a modern calc-alkaline dacite (Table II-5). Comparison of the analyses of the felsic volcanic rocks with trends of modern calc-alkaline series plotted by Miyashiro (1974) on Fig. II-17 show a great similarity, from which it is inferred that the felsic volcanic rocks of the Sherridon Group were generated from a calc-alkaline magma. Felsic volcanic rocks of both tholeiitic composition and calc-alkaline composition have been recognized in the Flin Flon belt (Stauffer et al, 1975; Fox, 1976; and Bailes, 1979).

II-7 Intrusive Rocks of the Sherridon Group

II-7-a Introduction

The intrusive rocks of the Sherridon Group are bimodal in composition, being either gabbros with ultramafic segregations or granites. With the exception of some pegmatites, all of the intrusive rocks have undergone the same metamorphic and tectonic history as the sedimentary and volcanic rocks.

Bateman (1944) described an intrusive complex near Cree Lake, consisting of metamorphosed anorthosite, pyroxenite, gabbro, and granite. Davies et al (1962) summarized the studies carried out on these rocks and concluded that only the pyroxenite and a small part of the granite were igneous in origin; the gabbro and anorthosite were considered to be calcareous sediments. Recent mapping by Froese (pers. comm., 1979) has shown that both the pyroxenite and the gabbro are igneous intrusions. He concludes, in agreement with Davies et al (1962), that a small amount of granite is present. The anorthosite has been remapped as marble (Unit 12 of this thesis).

Although the rocks of the Cree Lake intrusions are not in the thesis map area, they were examined and sampled so that they could be correlated with the mafic intrusions of the thesis area. A mafic sill (Unit 17), some small bodies of porphyritic granite (Unit 18) and pegmatite (Unit 19) constitute the intrusive rocks of the thesis area. Together with the Cree Lake intrusions, these rocks represent the full range of intrusive rocks in the Sherridon area.

II-7-b Amphibolite (Meta-Gabbro): Unit 17

Lying at the contact of the Sherridon Group with the Nokomis Group is a thick layer of amphibolite, which shows sharp contacts with the rocks of both Groups. It contains ultramafic accumulations near its contact with the Nokomis Group which are present in the form of dark continuous layers with sharp boundaries (Plate II-3a). Near Sheila Lake, a layer at the base of this amphibolite is 10 m thick. Grain size of these layers is coarse and, in one area near Cold Lake, graded beds of hornblende grains pseudomorphing pyroxenes are present (Plate II-3b). The relict pyroxene texture indicates an initial grain size of several cm. As the rock appears to have been composed almost entirely of pyroxene, the ultramafic rocks in the area are interpreted as pyroxenites. These ultramafic segregations and coarse grain sizes suggest that this layer is a sill and not a flow. No cross-cutting relationships were observed in the field.

The amphibolite consists of a coarse-grained, granoblastic polygonal mosaic of hornblende, plagioclase and quartz. The hornblende is a green variety and occurs as equant grains constituting 40 percent of the rock. Plagioclase has the composition $An_{40}-An_{50}$. A few small, randomly oriented flakes of biotite and a few pink almandine porphyroblasts are also present. Magnetite is commonly present in amounts ranging from 1 to 2 percent.

A pale amphibole having the pleochroic formula $\alpha=\beta$: colourless; γ : pale green was found to constitute almost 100 percent of the ultramafic rock of the Sheila Lake area. It has a $2V_{\alpha}$ of about 85° and an extinction angle $\gamma_{\lambda 2}$ of 18° . It is probably actinolite.

Plate II-3

- A) Layering in the mafic sills underlying formation A. Layering is sharp and variable in both composition and thickness. Photos taken near Cold Lake.
- B) Hornblende pseudomorphing pyroxenes which exhibit size gradation suggesting a cumulative origin in the base of the mafic sill near Cold Lake.

PLATE II-3. Textures in Metamorphosed Intrusive Rocks
of the Sherridon Area

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PAPIER DE COULEUR

Chemical analyses of the mafic sill and the Cree Lake intrusions have been plotted on Fig. II-14b. The rocks of the mafic sill have slightly less calcic plagioclase compositions than the mafic volcanic rocks. An average of these rocks including the Ultramafic layers would fall into the basalt (gabbro) field. Fig. II-16b is an AFM diagram which shows a pronounced differentiation trend for the intrusive rocks of the Sherridon area. The rocks of the mafic sill, the Cree Lake intrusions and the low TiO_2 mafic volcanic rocks are all very similar in composition (Table II-6). It is not unlikely that the magmas which formed the intrusive and volcanic rocks are related.

A few small amphibolite layers with coarse grained, uniform textures very similar to the textures of the gabbroic sill are found throughout the Sherridon area. Their chemical compositions have been plotted on Fig. II-14b and Fig. II-16b where they were found to overlap the composition of the gabbroic sill. It is probable that these amphibolite bodies represent hypabyssal or sub-volcanic intrusions.

II-7-c Meta-Granitic Rocks: Unit 18

A few small elongate bodies of meta-granite up to 200 m in length were recognized on a peninsula jutting into Sherlett Lake (Fig. II-2). These bodies have been intruded into both sedimentary and volcanic rocks; they contain amphibolite xenoliths and have cross-cutting relationships with an amphibolite layer.

The granitic rocks show a relict porphyritic texture and initially must have contained about 5 to 10 percent of quartz

TABLE II-6 Average Chemical Analyses of Intrusive and Related Rocks, Sherridon

	1	2	3	4	
MEAN					
SiO ₂	50.50	47.58	75.12	51.73	
TiO ₂	0.65	0.89	0.17	0.64	
Al ₂ O ₃	15.62	17.12	12.50	15.47	
Fe ₂ O ₃	2.68	2.10	0.52	2.92	
FeO	9.12	9.08	2.14	9.32	
MnO	0.19	0.22	0.08	0.21	
MgO	6.38	5.99	1.17	6.28	
CaO	8.79	10.45	1.72	9.10	
Na ₂ O	2.50	2.58	2.28	1.92	
K ₂ O	0.60	1.04	2.94	0.46	
P ₂ O ₅	0.10	0.44	0.18	0.14	
CO ₂	1.14	0.02	0.34	0.83	
H ₂ O	1.78	1.26	0.56	1.32	
STANDARD DEVIATION					
SiO ₂	3.05	4.49	2.28	2.62	
TiO ₂	0.24	0.21	0.07	0.07	
Al ₂ O ₃	1.05	0.29	1.56	0.83	
Fe ₂ O ₃	1.38	0.60	0.44	1.03	
FeO	1.26	1.70	0.49	0.69	
MnO	0.01	0.03	0.03	0.05	
MgO	1.24	1.14	0.35	1.93	
CaO	1.50	1.29	0.76	0.87	
Na ₂ O	1.02	0.41	0.70	0.62	
K ₂ O	0.13	0.27	1.46	0.13	
P ₂ O ₅	0.02	0.16	0.21	0.06	
CO ₂	0.87	0.04	0.21	0.81	
H ₂ O	0.56	0.29	0.36	0.35	
# SAMPLES	5	5	5	12	
					EXPLANATION FOR TABLE

1-Amfibolites which may represent subvolcanic intrusions, Sherridon

2-Gabbroic rocks, Sherridon

3-Granitic rocks, Sherridon

4-Low TiO₂ flows, Sherridon

phenocrysts up to 2 cm in diameter. The rocks have been metamorphosed and now have a weakly sutured granoblastic texture with a planar orientation of mica. The rock consists of 50 percent quartz, 10 percent biotite with the remainder being subequal amounts of plagioclase and microcline. The plagioclase composition is near An_{10} . A few garnets were also found in this rock.

The chemical composition of these rocks have been plotted on Fig. II-14b where they form a cluster in the granite field. The rocks have also been plotted on an AFM diagram on Fig. II-16b where they also form a discrete group. The granitic rocks are somewhat more enriched in FeO than the felsic volcanic rocks and it is doubtful if any correlation on either chemical or textural grounds can be made between the two rock types.

II-7-d Pegmatite: Unit 19

Equant masses and dykes of pegmatite are common throughout the Sherridon area. Grain size is highly variable and feldspar crystals range in size from 1 to 10 cm. Pegmatite bodies generally show a number of varieties based on grain size whereas dykes tend to be equigranular. The bodies of pegmatite do not exceed 200 m in diameter. The dykes are usually only a metre or two wide but several having a thickness of 10 m were found. The pegmatites are post-tectonic, except for one elongate body near the north end of Sherlett Lake which displays a weak flattening of feldspar and quartz grains.

The pegmatites are usually pink except for a few very pale pink and white pegmatites. The chief mineral of the pegmatite is

microcline which invariably constitutes 60 percent or more of the rock. Quartz and plagioclase make up the rest of the rock. Biotite is commonly present.

In a few instances, the pegmatites have remobilized large amounts of sulphide material, a relationship discussed in the chapter on economic geology.

II-8 Altered Rocks in the Sherridon Group

II-8-a Introduction

The term altered rocks in the context of this section is applied to those rocks which were altered in their pre-metamorphic history. These rocks fall into two categories: 1) rocks altered by weathering processes which formed metamorphic assemblages dominated by cummingtonite, and 2) rocks altered by hydrothermal solutions which formed metamorphic assemblages dominated by anthophyllite.

II-8-b Rocks Altered by Non-hydrothermal Processes

In addition to the hornblende-bearing amphibolites interpreted as mafic igneous rocks and calcareous sediments, there are a number of cummingtonite ($\pm$ biotite)-bearing amphibolites (Unit 10a) present in the Sherridon area. The cummingtonite-bearing amphibolites can be distinguished from the hornblende-bearing amphibolites in the field since the former have a gray-green waxy or greasy luster and the latter have a green-black vitreous luster on their fresh surfaces.

The cummingtonite-bearing amphibolites are found either as thin lenses or as the marginal zones of thick layers of hornblende-bearing amphibolite of igneous origin. The transition between the two types of amphibolite in these marginal zones may be abrupt or gradual. The marginal zones may be as much as 15 m thick. They are developed on both the upper and lower surfaces of the layers of hornblende-bearing amphibolite. Relict textures are not preserved in any of the cummingtonite-bearing rocks.

Typically, the cummingtonite-bearing amphibolites are massive. Grain size is highly variable with amphibole crystals ranging from 1 mm to several cm in length. The groundmass is generally medium to coarse grained having an average grain size of about 2 mm.

Cummingtonite, the characteristic mineral varies from 0 to 50 percent but averages about 25 percent in most of the well-developed occurrences of this rock type. In thin section, the amphibole is pale with the pleochroic formula $\alpha=\beta$: pale yellow to colourless; γ : pale brown. Multiple twinning on 100 is well-developed. Mineral grains have a $2V_{\gamma}$ of 75° and an extinction angle $\gamma \wedge Z$ of 20° . This suggests a chemical formula approximating $Mg_5Fe_2Si_8O_{22}(OH)_2$ for this mineral (Deer et al., 1967). The mineral habit is generally prismatic and idioblastic.

Hornblende, biotite, garnet, and anthophyllite form the mafic minerals associated with the cummingtonite in order of decreasing abundance. The hornblende has the pleochroic formula α : yellow-green; $\beta=\gamma$: medium to dark green. It has a $2V_{\alpha}$ of 85° and $\gamma \wedge Z$ of 22° . The mineral habit is generally idioblastic and

the grains occur as stubby prisms. Hornblende grains are found intergrown and in optical continuity with cummingtonite.

Biotite forms randomly oriented laths having the pleochroic formula α : reddish-brown; $\beta=\gamma$: dark brown. Like the hornblende, the biotite is found intergrown and in optical continuity with the cummingtonite. Biotite content varies from 3 to 10 percent except in a few instances where it constitutes as much as 75 percent of the rock (Unit 10b). Although this rock type has been given a sub-unit designation it is rare and interpreted as a biotite-rich variety of Unit 10a, the cummingtonite-bearing gneiss.

A poikiloblastic pink garnet, probably almandine in composition, forms aggregates up to 2 cm in diameter. It constitutes as much as 10 percent of the rock. Anthophyllite is rare, but where recognized forms as much as 30 percent of the rock. Its mineral habit is poikiloblastic acicular crystals forming sheaves. The mineral is almost colourless with the pleochroic formula $\alpha=\beta$: colourless; γ : pale tan. The crystals have $2V_{\gamma}$ of 70° , suggesting the aluminum rich variety gedrite.

The groundmass consists of a granoblastic polygonal aggregate of quartz and plagioclase. The composition of the plagioclase falls in the range of $An_{40}-An_{50}$. Traces of a pale green chlorite, magnetite, pyrrhotite, pyrite, and chalcopyrite were recognized.

Chemically, the cummingtonite-bearing gneiss is very similar in composition to the mafic volcanic amphibolites with which it is associated (Table II-7, Cols. 1, 2). The main difference between the two rock types is the reduction by about 50 percent of the CaO content and the doubling of the K_2O content. The relationship

TABLE II-7 Average Composition of Altered Rocks from Sherridon and Analyses Bearing on Their Origin

	1	2	3	4	5	6	7	8
MEAN	52.61	51.73	52.98	54.53	49.05	53.1	24.61	
SiO ₂	1.02	0.64	0.75	0.70	0.46	0.31		
TiO ₂	16.06	15.47	16.42	16.07	16.72	14.9	23.36	
Al ₂ O ₃	2.09	2.92	5.48	3.09	1.35	3.4	2.72	
Fe ₂ O ₃	9.82	9.32	6.42	9.68	9.87	15.48	20.93	
FeO	0.19	0.21	0.17	0.20	0.12	0.14	2.06	
MnO	6.62	6.28	4.17	5.15	15.16	7.3	14.15	
MgO	5.61	9.10	3.05	7.01	0.87	0.35	0.10	
CaO	2.91	1.92	2.27	2.83	0.27	0.44	0.28	
Na ₂ O	0.77	0.46	1.70	0.57	1.31	0.40	0.36	
K ₂ O	0.26	0.14	0.10	0.17	0.06	0.06		
P ₂ O ₅	0.75	0.83	2.63		0.25	0.11		
CO ₂	1.41	1.32	2.67		4.12	4.6	11.43	
H ₂ O								
STANDARD DEVIATION	3.65	2.62	5.37		9.66			
SiO ₂	0.63	0.07	0.10		0.20			
TiO ₂	0.92	0.83	1.61		2.96			
Al ₂ O ₃	0.99	1.03	3.49		0.61			
Fe ₂ O ₃	0.94	0.69	2.61		2.44			
FeO	0.04	0.02	0.07		0.03			
MnO	1.15	1.93	1.71		5.15			
CaO	1.79	0.87	2.47		0.89			
Na ₂ O	1.36	0.62	0.75		0.31			
K ₂ O	0.53	0.13	0.70		0.97			
P ₂ O ₅	0.14	0.06	0.10		0.02			
CO ₂	0.46	0.81	1.88		0.33			
H ₂ O	0.36	0.35	1.18		2.55			
SAMPLES	10	18	6	12	4	1	1	

EXPLANATION FOR TABLE

1-Cumingtonite gneiss, Sherridon

2-Low TiO₂ basaltic flows, Sherridon

3-Paleo-regolith developed on basaltic flows, Flin Flon

4-Basaltic flows, Flin Flon (from Stauffer et al, 1975)

5-Anthophyllite gneiss, Sherridon

6-Cordierite-anthophyllite gneiss, Coronation Mine, Manitoba (from Whitmore in Byers, 1969)

7-Chlorite, Shakani Mine, Japan (from Shirozu, 1974).

between these two rock types is similar to the relationship between the basalts of the Flin Flon area and weathering profiles developed on the basalts (Table II-7, Cols. 3, 4). These rocks are also identical except for a reduction by 50 percent of the CaO content and a doubling of the K_2O content.

The comparison with weathered volcanic rocks of the Flin Flon area suggests that the cummingtonite gneisses may be metamorphosed volcanic rocks which were altered by low temperature processes, either by weathering or through leaching by groundwater action.

II-8-c Rocks Formed Through Hydrothermal Processes

The anthophyllite-bearing rocks of the Sherridon Group were first recognized by Robertson (1953) and were the subject of a detailed petrographic study by Wilkinson (1976). They are found to form discontinuous layers over a broad area occupied by rocks of the Sherridon Group. In the eastern part of the area these layers are particularly well developed and reach a maximum thickness of 100 m and persist for a strikelength of several km. These layers have both gradational and sharp contacts with the lithic arenites which enclose them.

In the western part of the Sherridon area, the anthophyllite-bearing rocks are found as thin lenses, seldom more than 10 m thick and 50 m long. Localities of Sherridon Group rocks west of Sherridon have only a few small occurrences of anthophyllite. Layers of anthophyllite-bearing rock in the western part of the Sherridon area have sharp contacts with wall rocks and, in the case of the Smith-Pride showing, are interlayered with pelites.

gneiss, is almost invariably associated with Fe-sulphide as well as base metal mineralization. In particular, the east and west orebodies of the Sherritt Gordon deposit and Smith-Pride showing are underlain in a few localities by these rocks. Anthophyllite gneiss is associated with Cu and Zn sulphides near Trap Lake and at one occurrence near Elken Lake (Wright, 1931).

These rocks consist of anthophyllite and cordierite with minor quartz, biotite and garnet. Anthophyllite occurs in large, poikiloblastic radiating sheaves of crystals and may be as much as 10 cm in length. The anthophyllite is dark brown in hand specimen but shows the pleochroism $\alpha = \beta$: colourless; γ : light brown; and has a $2V_{\gamma}$ of 75° . Chemical analyses of this mineral from the Sherridon area show it to be the aluminous variety gedrite (Robertson, 1953). Cordierite is poikiloblastic with quartz inclusions. It was not seen to be polysynthetically twinned although Robertson (1953) reports this to be common in the Batty Lake area to the east. The cordierite is purple in hand specimen and has a $2V_{\alpha}$ of 85° . It is frequently altered to pinitite.

Large, poikiloblastic, pink garnets occur throughout this unit. Biotite and quartz make up much of the remainder of this rock. Magnetite and green spinel are intergrown in the cordierite-rich rocks and the minerals pyrite, pyrrhotite, chalcopyrite, and a pale green chlorite were also noticed.

Chemically, the anthophyllite-bearing rocks consist almost exclusively of SiO_2 , Al_2O_3 , FeO and MgO in order of abundance (Table II-7, Col. 5). Hypotheses for the origin of this rock have

been numerous. Robertson (1953) suggested that this unit is an altered tuff. Wilkinson (1976), noting the similarity of enrichment in MgO , FeO and Al_2O_3 in these rocks and alteration zones associated with massive sulphide deposits (e.g. Table II-7, Col. 6), suggested that this rock type was deposited as a chemical sediment on the sea floor.

As noted by Wilkinson (1976) this assemblage is very similar to the high-temperature assemblages found in metamorphosed volcanic alteration pipes. In fact, these are the only two types of occurrence for which this composition of rock has been recorded. In alteration pipes, this assemblage is the result of the conversion of the original rock assemblage to an assemblage consisting almost exclusively of Fe-Mg chlorite. For instance (Table II-6, Col. 7) chlorite from the Shakani Mine in Japan is rich in the same oxides as are plentiful in the anthophyllite layers at Sherridon and the alteration assemblage of the Coronation Mine, Flin Flon.

At lower temperatures such as are found associated with Red Sea brine pools and the weathering environment, the dominant clay formed is a Na-montmorillonite or smectite which has a composition rich in CaO and Na_2O (Bischoff, 1969; Slaughter and Earley, 1965). Thus, an origin related to high temperature hydrothermal alteration is to be preferred for the unmetamorphosed equivalent of anthophyllite-bearing gneiss.

The anthophyllite rocks near Sherridon could not have been formed as volcanogenic alteration pipes because they occur as thin widespread layers and do not show cross-cutting relationships. It

is doubtful that these rocks were precipitated as chemical sediments because most volcanogenic brines are very similar and layers of magnesian rocks would, therefore, be common in the stratigraphic record of volcanic accumulations. Direct precipitation of chlorite would require that geothermal brines be rich in Al^{+++} which is not true.

It is most probable that the protolith for the anthophyllite-bearing rocks at Sherridon was a chloritic mud deposited as a mechanical suspension along with sulphides precipitated from base metal bearing sulphides precipitated from base metal bearing solutions. The evidence supporting this view is presented with the discussion on the genesis of the sulphide deposits in the chapter on economic geology.

CHAPTER III

ECONOMIC GEOLOGY

III-1 Introduction

A number of discontinuous sulphide horizons occur throughout the Sherridon area and their stratigraphic position has been plotted in Fig. II-1. The sulphide zones present along these horizons fall into three types: 1) laminated, pyrrhotite-bearing, cherty calc-silicates; 2) disseminations of pyrite-pyrrhotite-chalcopyrite in pelites; and 3) economic to subeconomic, poly-metallic, massive sulphide lenses associated, in many cases, with anthophyllite-bearing gneiss. Sulphide occurrences of types (2) and (3) may be found on the same horizon, but the type (1) occurrence is found independently.

III-2 Cherty Sulphide Horizons

It was demonstrated in Chapter II that the pyrrhotite-bearing, graphitic, quartzitic calc-silicate horizons are the metamorphosed equivalents of Fe-sulphide bearing chert and calcium-magnesium carbonate sediments. Cherty sediments could have been formed by any of the following processes: 1) inorganic precipitation in a continental evaporite environment, 2) biogenic precipitation, and 3) as volcanic "exhalite" or sulphide facies iron formation.

The process of chert and carbonate precipitation in ephemeral lakes has been reviewed by Peterson and Borch (1965).

Essentially, the high pH of the lakes causes the dissolution of detrital silica grains and precipitation of chert occurs on lowering of pH by torrential flooding or saturation by drying. This environment is sufficiently alkaline to cause primary precipitation of high magnesium calcite and dolomite. The chert occurs as infillings around mud cracks in carbonate sediment and as irregular microscopic masses in the carbonates. The chert does not form well-developed, continuous laminations and the environment is too alkaline for the stability of iron sulphides so that it is unlikely that the pyrrhotite-bearing cherts of the Sherridon area were formed in a continental environment.

Assessment of a biogenic origin for these horizons is difficult because all biological evidence has been destroyed by metamorphism. It is possible to form a tentative conclusion by comparing these horizons to chert-carbonate association found in known modern and ancient biogenic environments. In the modern environment, biogenic chert occurs as masses and nodules in thick carbonate successions or as thin, laminated deposits in deep sea bottom sediments. Neither environment has abundant Fe-sulphides or rocks similar to the laminated chert protolith of the Sherridon area. Ancient biogenic precipitation resulted in the formation of "Superior-type" banded iron formations. The Kisseynew belt and related greenstone belts are younger than most biogenic banded iron formations and Stauffer (1974) stated that the Missi Group contains red beds indicative of an oxygenated atmosphere, a condition unfavourable to banded iron formation development. Banded iron formations generally are Fe-oxide rather than Fe-sulphide rich and are not typically associated with carbonates.

Sulphide facies volcanogenic iron formations or "exhalites" do resemble the Sherridon area pyrrhotite-chert horizons. Goodwin (1973, p. 917) gives the following description: "Lithologically the sulphide facies occur in the form of disseminated pyrite in black carbonaceous shale; pyrite and pyrrhotite mixed with siderite and carbonate in banded chert or quartz; and distinct pyrite or pyrrhotite beds with minor siderite. Thin interbanded chert may be present or absent." This is a very accurate description of the probable protolith for the silica-iron sulphide-graphite-carbonate assemblages found in the Sherridon area, although there is strong evidence to suggest that the underlying carbonates in the Sherridon area may have predated the solutions from which the cherty sulphides were precipitated. This evidence will be presented at the end of the chapter.

Sulphide facies iron formations have been described from the Kiseynew belt and from bordering volcanic belts. Gale and Koo (1978) have described the massive sulphide environment of the Lynn Lake belt. They found that most of the sulphide deposits occur within tuffs and siliceous sediments from a restricted stratigraphic interval. These siliceous sediments have been found to extend for several km, connect the massive sulphide lenses and contain graphite and disseminated iron sulphides.

Froese (pers. comm., 1979) observed thin cherty pyrite layers along several stratigraphic horizons in the Snow Lake area. Byers et al (1965) found a "barren sulphide group" in the Flin Flon area. Sulphide bodies carrying little or no economic mineralization were found to form a series of parallel zones along definite stratigraphic horizons where the associated rocks are

"silicified" and contain graphite. This type of occurrence is found in both the rocks of the Amisk Group and in the rocks of the Kisseynew belt.

The character of the cherty pyrrhotite horizons of the Sherridon Group and their similarity to the sulphide facies iron formations of the related rocks of central Manitoba leaves little doubt as to their genesis. The lack of iron formations in association with massive sulphide bodies in the Sherridon area is a problem which will be discussed later in this chapter.

III-3 Sulphide Disseminations in Pelites

The only large unit of pelitic rock occurs in the vicinity of Singsing Lake (Fig. II-2). In this unit, disseminated sulphides occur in sufficient concentration to have excited some economic activity. Small amounts of chalcopyrite were seen in trenches located on the claims of W. Shmon near Singsing Lake (Valray 1 showing). In most places, sulphide disseminations contain only iron sulphides.

This pelitic unit lies on the same stratigraphic horizon on which the Sherridon deposit and some subeconomic zones lie. Considering that there must have been a considerable hiatus in clastic sedimentation for the sulphide deposits to form, it is not surprising that an accumulation of clays would build up at that time. In the reducing environment associated with graphitic pelitic rocks, it is common for Fe-sulphides to form as accessory minerals. If Cu ions had been available to the Singsing Lake pelites it is probable that Cu-sulphides would have also been precipitated.

This possibility exists because large bodies of volcanogenic Cu-sulphides are found on the same horizon.

III-4 Subeconomic Massive Sulphide Deposits

A number of subeconomic polymetallic sulphide deposits are found in the Sherridon area. Their base metal content rarely exceeds 3 percent combined Cu and Zn, as shown in Fig. III-1, where the base metal content of the deposits in the Sherridon area is compared with that of others in central Manitoba. Additional statistics about the deposits of the Sherridon area and those of the Flin Flon, Lynn Lake, and Snow Lake districts can be found in Table III-1. Tonnages of the deposits of the Sherridon area are moderate when compared with others in central Manitoba (Fig. III-2), varying from 2 to 4 million tons.

Two of the larger deposits, the Jungle Lake and Park Lake zones, occur on the main ore-bearing horizon with the Sherritt Gordon orebodies. A smaller sulphide occurrence, the Fidelity showing is found on this horizon also. Near Elken Lake some small, uneconomic sulphide occurrences are found in association with anthophyllite gneiss (Wright, 1931).

The third of the larger deposits, the Bob Lake zone, is found on another mineralized horizon. Near Trap Lake, disseminated Cu sulphides in anthophyllite gneiss are found on this horizon.

All of the sulphide occurrences are thin, stratiform deposits rarely attaining a thickness greater than a few metres, and nowhere greater than 10 m. In some cases, an individual zone consists of a series of en echelon lenses, connected by disseminated sulphides. The horizons on which these sulphide deposits are

TABLE III-1 SELECTED DATA FOR MINES AND SULPHIDE OCCURRENCES FROM CENTRAL MANITOBA

MINE	% Pb	% Zn	% Cu	Tonnage (Millions of Tons)	Host Rock	Reference
<u>FLIN FLON DISTRICT</u>						
Flin Flon	-	4.4	2.2	64	fv	1
Schist Lake	-	7.01	4.22	2	fv	1
Coronation	-	0.24	4.25	1.4	mv	1
Centennial	-	2.60	2.06	1.4	fv	1
Westarm	-	0.6	4.63	0.71	fm	1
Cuprus	-	6.4	3.25	0.51	mv	1
White Lake	-	4.48	2.15	0.42	mv	1
Flexar	-	0.47	3.77	0.34	mv	1
Birch Lake	-	-	6.21	0.30	mv	1
North Star	-	-	6.11	0.27	fv	1
Mandy	-	15.0	8.0	0.15	mf	1
Don Jon	-	-	3.07	0.09	fm	1
<u>SNOW LAKE DISTRICT</u>						
Chisel	0.7	11.7	0.53	6	mv	1,3
Ghost	0.7	12.6	1.86	0.53	mv	1,3
Stall	-	0.6	4.79	5.4	fv	1
Osborne	-	1.5	3.75	3.2	fv	1
Anderson	-	0.1	3.81	3.1	fv	1
Dickstone	-	3.1	2.58	1.2	fv	1
Rod	-	2.88	7.1	0.30	fv	1
<u>LYNN LAKE DISTRICT</u>						
Ruttan	-	1.61	1.47	51	fv	4
Fox	-	2.7	1.84	13.1	fv	3
<u>SHERRIDON DISTRICT</u>						
Sherridon (East Ore-body)	-	2.97	2.45	8.5(Combined)	1a	2,5
Jungle Lake	-	1.1	1.42	3.7	1a	2
Park Lake	-	Similar to Jungle Lake	-	-	1a	(Hogg, pers. comm.)
Bob Lake	-	1.18	1.33	2.4	1a	2
Fidelity	-	Similar to Bob Lake	-	0.1	1a	(Shmon, pers. comm.)

Explanation to Table:
 fv - felsic volcanic
 mf - mafic volcanic
 fv - intermediate volcanic
 la - lithic arenite

References:
 (1) Price (1977)
 (2) Davies et al (1962)
 (3) Sangster (1972)
 (4) Langelaar (1975)
 (5) Farley (1949)

FIG. III-1. Plot of Cu vs Zn for sulphide occurrences in central Manitoba

FIG. III-2. Ore Tonnage vs $\frac{Zn}{Zn+Cu}$ for sulphide occurrences in central Manitoba

FIG. III-3. Total Base Metal Content vs $Zn/Zn+Cu$ for sulphide occurrences in central Manitoba

found, with the exception of the pelitic part, do not contain more than weakly disseminated sulphides. The horizons were recognized by the repeated occurrence of sulphide deposits at a similar position in the stratigraphic succession. All of the deposits occur in a lithic arenite host.

Recently, a small syncline containing Sherridon Group rocks was identified about 3 km west of the main basin of the Sherridon Group rocks (D. Baldwin, pers. comm., 1977). A small showing, the Smith-Pride, is found in the syncline and consists of a thin bed of Zn-rich sulphides. In addition to Zn, part of this bed is Pb-rich which is unusual for sulphide occurrences in the Sherridon area. The sulphide bed is overlain by thinly laminated pyrrhotite-bearing cherts and underlain by anthophyllite gneiss with disseminated sulphides (J. Ayer, pers. comm., 1979).

Discussion of this type of deposit is found in the next section (on the Sherritt Gordon deposit) because of the general similarity of all orebodies in the area.

III-5 The Sherritt Gordon Deposit

III-5-a General Geology

The historical background of the Sherritt Gordon deposit has been presented in Chapter I. The detailed geology of the two Sherritt Gordon orebodies is shown on Fig. III-4 (in pocket). Farley (1948, 1949) has produced the only other accounts of the geology of the Sherritt Gordon orebodies and the following statistics have been excerpted from them.

The orebodies had a combined length of 5,000 m of which 1,000 m was barren. The smaller east orebody had an average width of 4.7 m, a length of 1,450 m, and a maximum vertical extent of 75 m. The west orebody had an average width of 4.8 m, a length of 2,430 m and a maximum vertical extent of 460 m.

Level plans have been included on Fig. III-4. Mine records indicate mining was carried out from two levels and one shaft on the east orebody. The west orebody was mined from two shafts of which the main or west shaft was inclined 51° from the horizontal and the tenth and lowest level was at a depth of 413 m.

Final production statistics (Davies et al, 1962) record metal production as:

Cu	-	183,122 tons
Zn	-	74,480 tons
Au	-	101,026 ounces
Ag	-	3,218,324 ounces

with a total value of \$58,732,366 from 8,531,352 tons of ore milled. The production figures probably do not accurately reflect metal abundances in the mine because zinc was not recovered during the initial period of operation. Bateman and Harrison (1946) state that the east orebody (Table III-1) was more zinc-rich than the west orebody, which would indicate that overall grade for the mine was probably about 5 percent Cu + Zn with Cu being slightly more abundant than Zn.

The orebodies were thin, blanket like, stratiform layers which reached a maximum thickness of 15 m. Fig. III-4 shows that the west orebody dipped to the east and plunged to the north-east so that the vertical projection of its elongation has a trend

to the northeast. Fig. IV-2 shows that the dip of the orebody flattens and becomes essentially horizontal at a depth of less than 500 m. Bateman and Harrison (1946) and Farley (1949) thought that the pre-erosional sulphide body was continuous. The dissimilarity in the metal zoning of the two orebodies (which will be discussed later in this section) casts some doubt on this conclusion.

The sulphides of both orebodies are almost wholly contained in a lithic arenite host (Fig. III-4). Surface exposures of the west orebody show that anthophyllite gneiss underlies the orebody and siliceous quartzites and minor amounts of pelite overlie the orebody in places. Surface exposures of the east orebody reveal that the zinc-rich western part is found in siliceous, sillimanite-bearing, lithic arenite. The copper-rich east part of this orebody is contained in more diverse rocks (Fig. III-5) including a basal anthophyllite rock and several carbonate units. The level plans show that the association of the ore with lithic arenite, as revealed by surface exposure, persisted to depth in both orebodies.

The ore horizons are composed of about 70 percent sulphides and 30 percent gangue. Examination of polished thin sections of ore specimens (Appendix E) shows that the typical gangue associated with the ore is either siliceous arenite and/or chert. Additional types of gangue are composed of anthophyllite gneiss, pelitic gneiss or calc-silicates and carbonate. The gangue is found as rounded inclusions formed by the disruption of thin silicate layers during deformation (Fig. III-5). Layering is preserved only near the margins of the ore horizon where silicate layers were thick and resistant to deformation.

FIG. III-5. Simplified section through the east end of the east orebody.

Fig. III-6 (in pocket) illustrates a section of geology from the central part of the third level of the west orebody. On this figure is shown the sulphide distribution of the ore zone. The sulphides constitute almost 100 percent of the zone near its center but the amount diminishes to about 30 percent at the wall rock contact. Disseminations of 1 to 5 percent are common beyond this.

The commonplace occurrence of pegmatite is shown on Fig. III-6. Farley (1949) states that 20 to 25 percent of the mine production came from "offshoot" orebodies in which large amounts of sulphides were contained in pegmatite. It was inferred from this association that the ores were deposited as hydrothermal replacements from the pegmatites (Farley, 1949; Davies *et al.*, 1962).

Pegmatites, however, are found in no greater abundance with the orebodies than elsewhere in the area. The textures of the sulphides are typically metamorphic whereas the pegmatites are clearly post-metamorphic. Further, the pegmatites cross cut both sulphide and silicate layers, which means that the sulphides predate the pegmatite.

Fig. III-7a shows a typical offshoot orebody associated with pegmatite. The sulphides are wholly enclosed in the pegmatite, which has a cross-cutting relationship with respect to the stratigraphy. Offshoot orebodies are not restricted to an association with pegmatite. Fig. III-7b shows an offshoot orebody without pegmatite. In this type of offshoot, the metallic sulphides form lenses, which penetrate the wall rock and disrupt the stratigraphy. Ore minerals have also been remobilized into folds in both the hanging wall and footwall (Farley, 1949).

FIG. III-7. Geology of "Offshoot" orebodies; a) with pegmatite, b) without pegmatite: diagram modified from records of the Sherritt Gordon Co.

III-5-b Sulphide Mineralogy

The sulphide mineralogy of the orebodies is very simple. Farley (1949) stated that the sulphides consist almost entirely of pyrite, pyrrhotite, chalcopyrite, and sphalerite, in order of decreasing abundance. Minor amounts of galena with flecks of gold fill fractures on the hanging wall. Traces of cubanite and arsenopyrite were also recorded. The sulphide assemblages found from examination of polished thin sections in this thesis are recorded in Appendix E and the results of this study follow.

In the stratiform parts of the orebodies pyrrhotite occurs as granoblastic polygonal aggregates in association with subhedral pyrite cubes and masses of chalcopyrite and sphalerite. The dominant type of pyrrhotite is hexagonal although in a number of instances there is some alteration to a low temperature polymorph, particularly along fractures and at grain boundaries (Fig. III-8a). Farley (1949) states that pyrrhotite was found in a 1:2 ratio with pyrite whereas Bruce (1929) states that pyrrhotite was the dominant Fe-sulphide. The surface exposures of the mine show that pyrrhotite and pyrite occur in nearly equal amounts, pyrrhotite predominating. The offshoot orebodies in pegmatite contained pyrite as the exclusive Fe-sulphide which may account for the discrepancy in the estimates.

Pyrite occurs as euhedral to subhedral cubes in the stratiform part of the deposit (Fig. III-8b). These cubes range in size from a few millimetres to several centimetres. The edges of the cubes are irregular, show abundant embayments and many are coated with chalcopyrite. Numerous fractures within the cubes contain chalcopyrite and/or sphalerite. Where pyrite forms the dominant sulphide it shows a

c) Cp blebs exsolved from
Sp in Po free specimen
from "offshoot"

b) subhedral cubes of Py
in massive Po-Cp-Sp

a) rounded inclusions of
gangue and alteration
of hexagonal Po

1 Cm

FIG. III-8. Petrography of some specimens, from the
Sherritt Gordon Mine.

crushed texture of subhedral cubes in masses of disaggregated pyrite bits. Frequently, the larger grains are polygonized. These features probably result from late strain which deformed and flattened the orebody and caused rotation of grains.

Pyrite found in the offshoot orebodies forms irregular masses of large grains. Euhedral grains are very rare and much of the pyrite is associated with chalcopyrite and sphalerite (Fig. III-8c),

Chalcopyrite occurs as coarse-grained patches throughout the network of Fe-sulphides. It commonly coats pyrite grains and is found as a filling in fractures in pyrite and pyrrhotite. Chalcopyrite also forms disseminations with the Fe-sulphides in the siliceous and magnesian rocks underlying the massive sulphides, where it coats the grain boundaries of silicates and fills fractures and spaces at triple junctions.

The sphalerite is a dark, purplish, coarse-grained variety. It occurs as coarse-grained patches and also in aggregates with chalcopyrite and pyrrhotite. In some grains, there are small blebs of exsolved chalcopyrite or pyrrhotite (Fig. III-8c). These blebs have a well-defined orientation suggesting exsolution along crystallographic planes. In the offshoot orebodies, it is found in massive patches together with chalcopyrite and pyrite. Sphalerite is also found as disseminations with or without Fe-sulphides at the lateral edges of the orebodies (in plan). This is particularly evident in surface exposures at the west end of the east orebody.

No galena was seen in polished section but a number of fractures in the amphibolite near the hanging wall of the east

orebody were found filled with coarse-grained cubic aggregates of this mineral. Cubanite was found to form rectangular laths in the chalcopyrite from the east orebody; it is extremely rare. With the exception of a few rounded grains of magnetite these were the only metallic minerals found in the ores of the Sherritt Gordon mine.

III-5-c Metal Zoning

The distribution of minerals in the Sherritt Gordon mine was never recorded. However, partial records of assays and ore reserves were made available by Sherritt Gordon Mines Limited. From these, it is possible to plot the distribution of Cu and Zn, which is directly related to the distribution of chalcopyrite and sphalerite. Actual locations, values, and number of samples are confidential.

Fig. III-6 (in pocket) illustrates the Cu and Zn relationships in the thick central part of the west orebody. The total amount of Cu and Zn in the zone is proportional to the amount of sulphides in the zone. The Cu/Zn ratio varies from footwall to hanging wall irrespective of the sulphide content, a pronounced lateral zonation. The footwall is rich in Cu with a Cu/Zn ratio > 5 . This decreases to a Cu/Zn ratio < 1 at the hanging wall. Longitudinal metal distribution is depicted in Fig. III-9. Only weak longitudinal zoning is apparent on either of the seventh or eighth levels which suggests that zoning in the west orebody was restricted to the lateral dimension.

Fig. III-10 shows well developed zoning in the east orebody. The Cu/Zn ratio increases from < 1 at the west end to almost

FIG.III-9. Longitudinal Metal Zoning, West Orebody (based on reserve estimates)

FIG. III-10 Longitudinal Metal Zoning, East Orebody (based on reserve estimates)

3 at the east end, a pronounced longitudinal zonation. Once again, the Cu/Zn ratio is independent of the amount of sulphides in the zone. No data are available for lateral zonation in the east orebody but the detailed examination of surface exposures indicate it to have been negligible. This distinct dissimilarity in zoning between the east and west orebodies suggests they were separate orebodies at the time of formation, and that different mechanisms may have been responsible for their precipitation and their zoning.

III-5-d Interpretation and Comparison With Deposits of the Region

The mineralogy, metal content, morphology and associated rocks of the Sherritt Gordon and related subeconomic orebodies suggest three possible modes of origin: 1) hydrothermal replacement, 2) syngenetic or epigenetic precipitation in the sedimentary environment, 3) precipitation as a distal volcanogenic massive sulphide deposit. The proximal volcanogenic environment can be discounted because these deposits do not sit on thick piles of volcanic rocks and lack pipe-like alteration zones.

As has already been discussed, the fabric of the ores and wall rocks is similar and the pegmatites show cross-cutting relationships with respect to the ore. The well-developed layering of the orebodies at their margin and restriction to a stratigraphic horizon indicates a conformable genesis of the deposit and makes an origin by hydrothermal replacement unlikely.

The choice between a stratiform sedimentary and a volcanogenic deposit is more difficult to make because of the deformation and metamorphism of the orebodies. Stratiform sedimentary deposits

generally belong to one of three types: 1) carbonate-hosted Pb-Zn, 2) Cu-Pb-Zn deposits of the Kupferschiefer type, and 3) Cu or Cu-Pb sandstone-type deposits. Both the carbonate-hosted Pb-Zn and Kupferschiefer-type deposits are associated with thick accumulations of evaporites and shale whereas the deposits of the Sherridon area are associated with sandstones. The Cu-Pb sandstone-type deposits have a different metal association than the deposits of the Sherridon area. These types of deposits also do not show well-developed vertical zoning.

Figs. III-2, 3, 4 compare the Sherridon deposits with volcanogenic sulphide deposits in other Proterozoic rocks of central Manitoba. Fig. III-4 shows that the Cu/(Zn/Cu) ratios of the Sherridon deposits are well within the range of volcanogenic deposits for this region. Although distal volcanogenic deposits tend to be Zn-Pb rich, Cu/Zn ratios > 1 are not improbable (as will be discussed in the following sections). Both Geco and Manitouwadge, which are considered to be distal volcanogenic deposits (Sangster and Scott, 1976), contain substantial quantities of Cu.

Tonnages (Fig. III-3) of the Sherridon deposits are also well within the range of Proterozoic volcanogenic deposits in Manitoba. Base metal contents (Fig. III-2) tend to be on the low side, possibly because there are no values to plot for the sub-economic deposits in the Flin Flon area.

Sulphur isotopes are of great importance in deducing the origin of base metal deposits and have been determined for Sherritt Gordon (D. Sangster, pers. comm., 1978). Sulphur has four stable

isotopes of which the two most abundant, ^{32}S and ^{34}S differ in mass by about 6 percent and can be determined to a precision of 0.02 percent. The theory of sulphur isotopes and their application to sulphide deposits has been reviewed in detail by Sangster (1976).

Normally, sulphur isotopes are reported as permil (‰) with reference to the Canyon Diablo standard (0.00 ‰):

$$\delta^{34}\text{S} \text{ ‰} = \frac{(\frac{^{34}\text{S}}{^{32}\text{S}})_{\text{sample}} - (\frac{^{34}\text{S}}{^{32}\text{S}})_{\text{standard}} \times 1,000}{(\frac{^{34}\text{S}}{^{32}\text{S}})_{\text{standard}}}$$

Biologic action and the oxidation of sulphide to sulphate in sea-water tend to fractionate sulphur isotopes. Sedimentary deposits which are formed from sulphide produced through the action of sulphate-reducing bacteria tend to have wide ranges of $\delta^{34}\text{S}$ values. Magmatic deposits and volcanogenic deposits which are formed directly from magmatic sulphides tend to have much smaller variations. Metamorphism does not appreciably affect $\delta^{34}\text{S}$ values.

Sangster (pers. comm., 1976) reports that for Sherridon an average of 22 samples gave a $\delta^{34}\text{S}$ of 0.2 ‰ and a standard deviation of 0.8 ‰. This is an amazingly small range because the average standard deviation of volcanic deposits is 6.5 ‰ and of sedimentary deposits is 11.8 ‰ (Sangster, 1976).

Sangster also reports that the volcanogenic deposits of the Flin Flon belt have a mean of 0 ‰ and a standard deviation of 2.0 ‰ for their sulphur isotopes. These and the previous data suggest that the Sherridon deposit is not only a volcanogenic massive sulphide deposit but may be related to the events which were

responsible for the deposition of sulphides in the Flin Flon-Snow Lake belt.

II-6 Summary of Field and Analytical Data Bearing on the Origin of the Sherritt Gordon Deposit

From the point of view of field and analytical evidence, the Sherritt Gordon orebodies have the aspect of volcanogenic massive sulphide deposits. To summarize the main points:

1. The structure and stratigraphy of the deposit are consistent with a stratiform deposit.
2. The mineralogy, metal abundances and zoning are similar to known volcanogenic deposits.
3. Sulphur isotopes favour a volcanogenic origin.
4. Metal abundances, tonnages and isotopic characteristics are similar to deposits of the Flin Flon-Snow Lake belt.
5. There is no evidence supporting a magmatic, sedimentary or epigenetic origin for the deposit.

Further, the ore deposits of the Sherridon area are similar to deposits which have been described from the rocks of the Osugi belt in Japan (Mitchell and Bell, 1973) and the Phillipines (Bryner, 1969). In both volcanic belts, three types of mineralization, constituting the Besshi type of deposits are present:

1. Cu-Zn massive sulphide layers,
2. Banded cherts with Fe-sulphides, and
3. Bedded Cu-Fe ore, typically associated with pelitic schists.

The host rocks of the mineral deposits are mafic tuffs, volcanogenic sediments and thin limestone layers. The similarity of setting, host rock lithology, and sulphide mineralization suggests a similar origin for the Sherridon Group and its mineral deposits.

However, the lack of underlying volcanic rocks and the occurrence of the deposit in sedimentary rocks complicates the interpretation. For this reason, a geochemical model of metallic sulphide deposit formation is constructed in the following sections and tested against the evidence of the Sherritt Gordon deposit to determine whether it is of volcanogenic origin.

III-7 Construction of Solubility Diagrams.

III-7-a Introduction

In order to propose a model for the deposition of the Sherridon area sulphide deposits, it was necessary to calculate a series of diagrams on which solubility data and mineral equilibria could be represented. Log fO_2 vs pH and log fO_2 vs T diagrams were chosen for this purpose. Anderson (1973, 1975, 1977) used fugacity vs pH diagrams to illustrate the solubility of sulphides. His work was concerned with carbonate-hosted Pb-Zn deposits. Large (1977) used this type of diagram as well as log fO_2 vs T diagrams to illustrate depositional processes in volcanogenic massive sulphide deposits. The metal-carrying potential of brines at various temperatures as well as the metal ratios of the sulphide ores being precipitated can be estimated from these diagrams.

III-7-b Calculation and Representation of the Fe-S-O Mineral Equilibria

Before calculating the solubilities of the base metal sulphides and plotting them, it is necessary to represent the mineral equilibria of the coexisting Fe-sulphides and oxides on the diagram.

Experimental work depicting the equilibria of Fe-sulphides and oxides is available for only a limited range of the conditions of interest (Crerar et al, 1978). It was, therefore, necessary to use thermochemical data to calculate the equilibria of the Fe-S-O minerals.

Also, to carry out a thermodynamic study of an aqueous system it must be adequately defined. For example, the phase relations of the metallic oxides and sulphides in an aqueous system are controlled by the variables of temperature, pressure, pH and the oxygen and sulphur fugacities. The effect of pressure is of minor importance in the range of pressure associated with subaqueous volcanogenic sulphide emplacement and can be neglected. The sulphur fugacity will be determined if the number of moles of sulphur (ES) per kg of water is fixed and the other parameters are known. Thus, it is necessary to consider the variables T, ES, fO_2 and pH in the evaluation of the stability fields of the Fe-S-O minerals.

The effects of ES and T can be considered by producing a series of plots of $\log fO_2$ vs pH at selected values of ES and T. $\log fO_2$ vs T diagrams were created graphically since sufficient $\log fO_2$ vs pH diagrams are available for the temperature range under consideration.

The first step in establishing the Fe-S-O mineral equilibria is to represent the equilibria of the aqueous sulphur species present. This graphical representation requires balanced equations between the ionic and molecular species present in terms of O_2 , H^+ and H_2O (Table III-2). Reactions involving only ionization are functions of pH and reactions involving only changes in the

TABLE III-2 EQUILIBRIUM CONSTANTS FOR REACTIONS INVOLVING AQUEOUS SULPHUR SPECIES

▼ Data are from Helgeson (1969)
 ** Data are from Ellis and Gfgenbach (1971)

	Log K (T°C)					
	50	100	150	200	250	300
▼ $\text{H}_2\text{S} = \text{H}^+ + \text{HS}^-$	-6.77	-6.63	-6.72	-6.96	-7.35	-8.06
** $\text{HS}^- = \text{H}^+ + \text{S}^0$	-	-	-12.8	-12.5	-12.2	-12.0
$\text{HSO}_4^- = \text{SO}_4^{2-} + \text{H}^+$	-2.27	-2.99	-3.74	-4.49	-5.41	-7.06
$\text{H}_2\text{S} = 2\text{O}_2 = 2\text{H}^+ + \text{SO}_4^{2-}$	113.71	94.46	79.53	67.54	57.61	49.0
$\text{HS}^- + 2\text{O}_2 = \text{SO}_4^{2-} + \text{H}^+$	120.48	101.09	86.25	74.50	64.96	57.1
$\text{H}_2\text{S} + 2\text{O}_2 = \text{H}^+ + \text{HSO}_4^-$	115.98	97.45	83.27	72.03	63.02	56.1
$\text{S}^0 + 2\text{O}_2 = \text{SO}_4^{2-}$	-	-	99.05	87.00	77.16	69.1
$\text{S}(\text{s}) + 3/2 \text{O}_2 + \text{H}_2\text{O} = \text{HSO}_4^- + \text{H}^+$	82.76	69.80	59.98	52.30	46.31	42.16
$\text{S}(\text{s}) + 3/2 \text{O}_2 + \text{H}_2\text{O} = \text{SO}_4^{2-} + 2\text{H}^+$	80.49	66.81	56.24	47.81	40.90	35.1
$\text{S}(\text{s}) + \text{H}_2\text{O} = \text{H}_2\text{S}(\text{a}) + 1/2 \text{O}_2$	-33.33	-27.76	-23.38	-19.83	-16.81	-13.94
$\text{S}(\text{s}) + \text{H}_2\text{O} = \text{HS}^- + \text{H}^+ + 1/2 \text{O}_2$	-40.10	-34.39	-30.10	-26.79	-24.16	-22.0
$2\text{S}(\text{s}) = \text{S}_2(\text{g})$	-12.24	-9.47	-7.37	-5.73	-4.41	-3.33
$\text{S}_2(\text{g}) + \text{O}_2 + 2\text{H}_2\text{O} = 2\text{HSO}_4^- + 2\text{H}^+$	175.27	145.88	123.40	105.63	91.42	80.5
$\text{S}_2(\text{g}) + \text{O}_2 + 2\text{H}_2\text{O} = 2\text{SO}_4^{2-} + 4\text{H}^+$	173.00	142.89	119.66	101.14	86.01	73.4
$\text{S}_2(\text{g}) + 2\text{H}_2\text{O} = 2\text{H}^+ + 2\text{HS}^- + \text{O}_2$	-67.95	-59.30	-52.83	-47.84	-43.91	-40.7
$\text{S}_2(\text{g}) + 2\text{H}_2\text{O} = 2\text{H}_2\text{S}(\text{a}) + \text{O}_2$	-54.41	-46.04	-39.39	-33.92	-29.21	-24.6
$\text{S}_2(\text{g}) + 2\text{H}_2\text{O} = 4\text{H}^+ + 2\text{S}^0 + \text{O}_2$	-	-	-78.43	-72.84	-68.31	-64.7

* Unless otherwise specified, all equilibrium constants are calculated from the data of Helgeson (pers. comm. to G. Skippin, 1978). All reactions involving H_2S were estimated by addition of the similar reaction involving HS^- and the reaction $\text{H}_2\text{S} = \text{H}^+ + \text{HS}^-$.

oxidation state are functions of fO_2 . Reactions involving both oxidation and ionization are functions of both fO_2 and pH.

The boundaries between the various sulphur species demarcate fields of predominance and are drawn where the two particular species involved have equal activities. Inside these fields, except very close to the edge, the molality of the particular sulphur species is approximately equal to ΣS (Barnes and Kullerud, 1961; Large, 1977). Once the fields have been established it is possible to contour the diagram in fS_2 , which is necessary if the Fe-S-O minerals are to be plotted. A series of reactions can be written where the sulphur species in the field of interest reacts to form S_2 (Table III-2). It has also been assumed that the activity of the dominant species approaches ΣS in the field of interest (i.e. activity coefficients have been neglected). Substituting an activity of the sulphur species provides equations in $\log fO_2$, pH and $\log fS_2$ which can be solved for values of fS_2 to be plotted on the diagram (Garrels and Naeser, 1959).

Equilibria between the minerals of the Fe-S-O system in terms of the variables fO_2 and fS_2 can now be written (Table III-3). Since the diagrams have fO_2 as the ordinate and are contoured in fS_2 , these equilibria can be plotted directly (Barnes and Kullerud, 1961).

The calculation of the equilibrium constants used to plot the preceding data was carried out using thermochemical information. To a large degree, this was done using the enthalpies of formation, entropies and Maier-Kelley heat capacity coefficients from Helgeson (personal communication to G. B. Skippen, 1977) and a mathematical expression from Skippen and Carmichael (1978):

TABLE III-3 EQUILIBRIUM CONSTANTS FOR REACTIONS INVOLVING IRON SPECIES

	Log K T°C					
	50	100	150	200	250	300
$4\text{Fe}_3\text{O}_4 + \text{O}_2 = 6\text{Fe}_2\text{O}_3$	65.62	54.99	46.07	40.47	35.28	30.99
$\text{FeS} + 1/2 \text{S}_2 = \text{FeS}_2$	15.53	12.59	10.33	8.54	7.08	5.87
$3\text{FeS}_2 + 2\text{O}_2 = \text{Fe}_3\text{O}_4 + 3 \text{S}_2$	48.95	44.22	40.62	37.78	35.49	33.61
$2\text{FeS}_2 + 3/2 \text{O}_2 = \text{Fe}_2\text{O}_3 + 2\text{S}_2$	43.57	38.62	34.82	31.81	29.37	27.45
$3\text{FeS} + 2\text{O}_2 = \text{Fe}_3\text{O}_4 + 3/2 \text{S}_2$	95.54	81.99	71.61	63.41	54.75	51.22
$3\text{Fe} + 2\text{O}_2 = \text{Fe}_3\text{O}_4$	162.66	138.47	120.01	105.48	93.74	84.07
$\text{Fe} + \text{S}_2 = \text{FeS}_2$	37.91	31.42	26.47	22.57	19.42	16.82
$\text{Fe} + 1/2 \text{S}_2 = \text{FeS}$	22.38	18.83	16.13	14.02	12.33	10.95

* Equilibrium constants calculated from data of Helgeson (pers. comm. to G. Skippen, 1978) except for iron data (Kelley, 1960).

$$\ln k = (\Delta H_{298.15}^{\circ} - \Delta S_{298.15} T + \Delta a [T - 298.15 - (T \ln T / 298.15)] \\ - \Delta b [T^2 / 2 + (298.15^2 / 2) - 298.15 T] \\ - \Delta c [\frac{1}{2} T - (1 / 298.15) + T / 298.15^2] / RT$$

The application of this expression requires thermodynamically consistent data for all the phases employed. The Helgeson data have been tested by using entropy-volume data to calculate a Clapeyron slope for a reaction. In all cases, the data yielded internally consistent results (Skippen and Carmichael, 1977).

As heat capacity coefficients are not available for ionic species at temperatures in excess of 250°C or for the species H_2S (aq) the equilibrium constants at these conditions and for reactions involving the species H_2S (aq) had to be carried out in a different manner. The equilibrium constants for reactions involving the species H_2S (aq) were estimated by combining the data for the reaction $H_2S = H^+ + HS^-$ with reactions involving HS^- from Table III-2. The equilibrium constants for the reactions involving ionic species at temperatures in excess of 250°C were extrapolated graphically by plotting values of $\log K$ vs T .

The dissociation data for H_2S and HS^- were taken from experimental sources because of the discrepancies associated with known thermochemical data (Anderson, 1977). Equilibrium constants for the reaction:

were taken from Helgeson (1969), which is in good agreement with the data of Ellis and Giggenbach (1971). Equilibrium constants for the reaction:

are taken from Ellis and Giggenbach (1971) as suggested by Anderson (1977).

III-7-c Calculation and Representation of the Solubility Data

Before considering how the metal solubilities were calculated for the metallic sulphides it is first necessary to review a few additional constraints.

The findings of Ellis and Giggenbach (1971) show that the solubility expression for a metallic sulphide MS should be either

or

Anderson (1977) has summarized the data showing that the solubilities of the metallic sulphides in pure water are much too small to account for ore deposits. Large (1977), summarizing the available experimental work, concluded that the complexing of Cu, Pb, Zn, and Fe with chlorides in saline, weakly acid solutions containing $ES < 10^{-2}$ m is responsible for ore deposits. Bisulphide complexes become important in neutral to alkaline solutions below 250°C and carrying $ES > 10^{-1}$ m. At temperatures below 50°C , transport of metals could be achieved by carbonates, sulphates and hydroxyl species as well as organic ligands and colloids. Fluid inclusions in massive volcanogenic sulphide deposits have been shown to be saline in the range of 0.5 to 3 molal and neutral to weakly acid (Table III-4). Temperatures estimated from fluid inclusions are in the range of 170 to 350°C and sulphur concentrations are of the order of 10^{-2} to 10^{-3} m. For these reasons, all

TABLE III-4 BOUNDARY CONDITIONS FOR VOLCANOGENIC FLUIDS ASSOCIATED WITH MASSIVE SULPHIDE FORMATION
AND FOR SEAWATER

	Temp. Range °C	Avg. Temp. °C	pH	log fO ₂	-log ΣS m	Cl ⁻ m
Black Ore (stratiform sphalerite)	175-225	200	4-4.5	-45	3-3.5	3
Yellow Ore (stratiform chalcopyrite)	225-275	250	3.5-4	-40	2.5-3.5	3
Siliceous Ore (stockwork chalcopyrite)	275-350+	300+	3-3.5	-35	2-3	3
Seawater		20	8	highly variable	2	0.5

Estimates of temperatures are taken from Urabe and Sato (1978); remaining estimates are extrapolated from Lambert and Sato (1974).

of the complexing associated with volcanogenic hydrothermal brines is considered to occur with the chloride ion.

The addition of chloride to the hydrothermal system means that in addition to T , fO_2 , pH , and $a_{H_2S(aq)}$, the solubility of metallic sulphides will also depend on the activities of the species of aqueous metallic chlorides involved.

The metal solubilities can now be calculated taking into account the complexing action of the chlorides. A general set of equations for calculating the solubility of sulphides in solutions has been developed by Nriagu (1971) and their use demonstrated by Anderson (1977).

If a cation M forms a complex with a ligand L , the complex forming reaction can be written:

The charges on the ligand and the complex have been omitted but in this model $L = Cl^-$. By considering the mass balance for the cation which exists in the above n linear simultaneous equations, it can be shown that:

$$\begin{aligned} m_{M(\text{total})} &= m_M (1 + B'_1 m_L + B'_2 m_L^2 + \dots + B'_n m_L^n) \\ &= m_M \sum_{n=0}^n B'_n m_L^n; \end{aligned} \quad B'_0 = 1 \text{ by definition}$$

where

$$\begin{aligned} B'_2 &= m_{ML} / (m_M \cdot m_L) \\ &= B'_1 (\gamma_M \cdot \gamma_L) / \gamma_{ML} \end{aligned}$$

and similarly for the other B' values.

The theory can be expanded to include complexes with more than one ligand but Nriagu (1971) indicates that mixed species

are of no geological interest.

The solubility which is the sum of the metallic ions in solution plus the metal associated with the chlorides is given by:

$$m_{M(\text{total})} = m_M + m_{MCl} + m_{MCl_2} + m_{MCl_3} + m_{MCl_4}$$

$$= m_M \cdot \phi \quad \phi = \sum_{n=0}^n B'_n (m_L)^n$$

ϕ is the factor which accounts for all the metal chloride complexing and converts the simple metal ion concentration to the total metal concentration. It was previously demonstrated how ϕ is calculated so that it is only necessary to show how to substitute for m_M in the above equation to solve for $m_{M(\text{total})}$.

In the case of a bivalent metal ion, this is done as follows:

$$K = \frac{[a_{M^{++}}][a_{H_2S}]}{[a_{MS}][a_{H^+}]^2}$$

$$K = m_{M^{++}} \cdot m_{H_2S} \cdot \gamma_{M^{++}} \cdot \gamma_{H_2S} / a_{H^+}^2$$

or

$$m_{M^{++}} = K \cdot a_{H^+}^2 / m_{H_2S} \cdot \gamma_{M^{++}} \cdot \gamma_{H_2S}$$

and substituting into the equation for solubility:

$$m_{M(\text{total})} = K \cdot a_{H^+}^2 \cdot \phi / m_{H_2S} \cdot \gamma_{M^{++}} \cdot \gamma_{H_2S}$$

This is the final form of the equation needed to calculate the solubilities of the simple metallic sulphides to plot on the diagrams.

Thermochemical data used to make the actual calculations come from a number of sources. Values for B_n , the dissociation constants for the chloride complexes are listed in Table III-5.

TABLE III-5 DISSOCIATION CONSTANTS FOR METAL ASSOCIATED CHLORIDE COMPLEXES

		Log K (T°C)			
		150	200	250	300
PbCl ⁺	Pb ⁺⁺ + Cl ⁻	-1.88	-2.1	-2.4	-3.0
PbCl ₂	Pb ⁺⁺ + 2Cl ⁻	-2.29	-2.6	-3.1	-3.9
PbCl ₃ ⁻	Pb ⁺⁺ + 3Cl ⁻	-2.50	-3.0	-3.6	-4.6
PbCl ₄ ²⁻	Pb ⁺⁺ + 4Cl ⁻	-2.57	-3.2	-4.0	-5.3
ZnCl ⁺	Zn ⁺⁺ + Cl ⁻	-2.78	-3.9	-4.8	-6.0
ZnCl ₂	Zn ⁺⁺ + 2Cl ⁻	-3.19	-4.4	-4.4	-6.9
ZnCl ₃ ⁻	Zn ⁺⁺ + 3Cl ⁻	-3.34	-4.8	-6.0	-7.7
ZnCl ₄ ²⁻	Zn ⁺⁺ + 4Cl ⁻	-3.35	-5.0	-6.4	-8.3
Cu ⁺	Cu ⁺ + Cl ⁻	-2.2	-2.4	-2.2	-2.7
CuCl ₂ ⁻	Cu ⁺ + 2Cl ⁻	-5.35	-5.8	-6.5	-7.4
CuCl ₃ ²⁻	Cu ⁺ + 3Cl ⁻	-5.77	-6.4	-7.2	-8.3

All data from Helgeson (1969) except where indicated
 * Greer and Barnes (1976)

Most of the values come from Helgeson (1969) who computed them from experimental data. No heat capacity coefficients are known for chloride complexes in aqueous solution.

The activity coefficients used to convert the B to B' values were calculated using the extended Debye-Hückel equation plus a deviation function:

$$\log \gamma = -AZ^2I^{1/2}/(1 + aBI^{1/2}) + BI$$

where I is the true ionic strength and B is the deviation function, both taken from tables for NaCl solutions in Helgeson (1969). A and B are temperature dependent Debye-Hückel coefficients from Helgeson (1967). a is the distance of closest approach of ions in solution for which γ is calculated. a values for Pb^{++} , Cu^{+} and Zn^{++} are from Garrels and Christ (1965), and an a value of 4.2×10^{-8} for Cl^{-} comes from Anderson (1977). The activity coefficient of all neutral ions is assumed to be that of CO_2 in NaCl solution (Helgeson, 1969). Activity coefficients of all singly charged ions are assumed equal to $\gamma_{Cl^{-}}$ and all doubly charged ions to $\gamma_{M^{++}}$. The rationale for this treatment is that very few a values exist for complex ions and it is assumed that the behaviour of ions with most similar charges will be most similar. Anderson (1977) has shown that for solubility calculations involving these assumptions agreement is excellent at low chloride concentrations. The calculations are sensitive to errors in a but, considering all possible sources of error, not unreasonable. Anderson (1977) indicates solubilities are correct to a factor of 4 or 5 considering all possible errors. With the refinements in the equilibrium constants used in this thesis solubilities are as good or better.

As stated earlier, these solubility calculations work for the simple metallic sulphides such as ZnS or PbS. However, in volcanogenic massive sulphides Cu is precipitated not as the simple sulphide CuS but as the complex sulphide CuFeS_2 (Lambert and Sato, 1974). In the case of chalcopyrite the solubilities of both Cu and Fe must be taken into account:

and

$$K = \frac{[a_{\text{Cu}^{++}}][a_{\text{Fe}^{++}}][a_{\text{H}_2\text{S}}]^2}{[a_{\text{CuFeS}_2}][a_{\text{H}^+}]^4}$$

In volcanogenic massive sulphide deposits, one or another of the Fe-S-O minerals pyrite, pyrrhotite, magnetite, or hematite is always found in association with the Cu sulphide. From this, it is known that the Fe content of the solution must have been at saturation. Thus, it is possible to write the following reactions:

or

or

or

any one of which is valid depending on the coexisting Fe-S-O mineral. It is interesting to note that the reactions involving pyrite, magnetite, and hematite all exhibit a dependency on $f\text{O}_2$.

This technique avoids the problem of dealing with two ions in solution simultaneously. These equations are valid only below about 150°C because Cu^{++} is replaced by the more stable ion Cu^+ above this temperature. This necessitates rewriting the above reactions in terms of the Cu^+ ion:

In this case, the reaction involving pyrite shows a dependency on fO_2 but the reaction involving hematite does not. The reaction involving hematite has an infinite slope, the pyrite reaction has a negative slope and the pyrrhotite and magnetite reactions have positive slopes on an $\log \text{fO}_2$ vs pH diagram. It is only necessary to determine the Cu solubility at one point to construct the diagram since the slopes of the reactions are constant.

The equilibrium constants for the aqueous reactions involving metallic sulphides can be found in Table III-6. A sample diagram is provided (Fig. III-11) to illustrate a completely constructed $\log \text{fO}_2$ vs pH diagram. Contours of fS_2 have been omitted from succeeding diagrams because they clutter the diagrams and make them difficult to interpret.

The solubilities of the base metal sulphides were not calculated for the sulphate and bisulphate fields because it has already been demonstrated that the volcanogenic fluids are reducing.

TABLE III-6 EQUILIBRIUM CONSTANTS FOR AQUEOUS REACTIONS INVOLVING METALLIC SULPHIDES AND OXIDES

	Log K (T°C)			
	150	200	250	300
$\text{PbS} + \text{H}^+ = \text{Pb}^{++} + \text{HS}^-$	-10.79	-9.77	-8.94	-8.22
$\text{ZnS} + \text{H}^+ = \text{Zn}^{++} + \text{HS}^-$	-9.73	-9.26	-8.85	-8.49
$\text{PbS} + 2\text{H}^+ = \text{Pb}^{++} + \text{H}_2\text{S}$	-4.07	-2.81	-1.59	-0.16
$\text{ZnS} + 2\text{H}^+ = \text{Zn}^{++} + \text{H}_2\text{S}$	-3.01	-2.30	-1.50	-0.43
$\text{CuFeS}_2 + \text{H}^+ + \text{H}_2\text{O} = \text{Cu}^+ + \text{FeS}_2 + \text{H}_2\text{O}$	4.49	3.95	3.60	3.48
$\text{CuFeS}_2 + \text{H}_2\text{O} + \text{H}^+ = \text{Cu}^+ + \text{FeS} + \text{H}_2\text{S} + \text{H}_2\text{O}$	-25.54	-21.55	-18.04	-14.80
$\text{CuFeS}_2 + \text{H}_2\text{O} = \text{Cu}^+ + \text{FeS} + \text{HS}^- + \text{H}_2\text{O}$	-32.26	-28.51	-25.43	-22.86
$3\text{CuFeS}_2 + 4\text{H}_2\text{O} + 3\text{H}^+ = 3\text{Cu}^+ + \text{Fe}_3\text{O}_4 + 6\text{H}_2\text{S} + \text{H}_2\text{O}$	-64.41	-94.45	-42.15	-31.2
$3\text{CuFeS}_2 + 4\text{H}_2\text{O} = 3\text{Cu}^+ + \text{Fe}_3\text{O}_4 + 6\text{HS}^- + 3\text{H}^+ + \text{H}_2\text{O}$	-104.73	-52.69	-86.25	-79.56
$2\text{CuFeS}_2 + 3\text{H}_2\text{O} + 2\text{H}^+ = 2\text{Cu}^+ + \text{Fe}_2\text{O}_3 + 4\text{H}_2\text{S}$	-35.13	-28.38	-22.22	-15.63
$2\text{CuFeS}_2 + 3\text{H}_2\text{O} = 2\text{Cu}^+ + \text{Fe}_2\text{O}_3 + 4\text{HS}^- + 2\text{H}^+$	-62.00	-56.22	-51.62	-47.87
$\text{FeS} + \text{H}^+ = \text{Fe}^{++} + \text{HS}^-$	-6.30	-6.34	-6.34	-6.31
$\text{FeS} + 2\text{H}^+ = \text{Fe}^{++} + \text{H}_2\text{S}$	0.42	0.62	1.01	1.75
$\text{FeS}_2 + \text{H}_2\text{O} = \text{Fe}^{++} + 2\text{HS}^- + \text{H}_2\text{O}$	-43.06	-38.84	-35.43	-32.63
$\text{FeS}_2 + \text{H}_2\text{O} + 2\text{H}^+ = \text{Fe}^{++} + 2\text{H}_2\text{S} + \text{H}_2\text{O}$	-29.62	-24.92	-20.73	-16.51
$\text{Fe}_2\text{O}_4 + 6\text{H}^+ = 3\text{Fe}^{++} + 3\text{H}_2\text{O} + \text{H}_2\text{O}$	-11.26	-10.71	-10.04	-9.25
$\text{Fe}_2\text{O}_3 + 4\text{H}^+ = 2\text{Fe}^{++} + 2\text{H}_2\text{O} + \text{H}_2\text{O}$	-15.32	-13.89	-12.57	-11.34

FIG. III-11. fO_2 vs. pH Diagram illustrating the solubilities of Cu, Pb and Zn at $T = 250^\circ C$, $XS = 10^{-2}$ m and $Cl^- = 3$ m
 ----Cu ———Pb - · - · Zn

The solubility contours are, therefore, truncated at the sulphate and bisulphate field boundaries.

Figs. III-12a, b, c, and d show the solubilities of Zn, Pb, and Cu on $\log fO_2$ vs pH diagrams for the range of $T = 150, 200, 250,$ and 300°C and for the range of $ES = 10^{-1}, 10^{-2}, 10^{-3},$ and $10^{-3.5}$ m. Fig. III-13 shows the solubilities of Zn, Pb, and Cu on $\log fO_2$ vs T diagrams for the range of $pH = 3, 4$ and $ES = 10^{-3}$ m. For ease of calculation, a molality of $Cl^- = 3$ rather than a molality of $NaCl = 3$ was used in this thesis. As Na^+ and Cl^- tend to associate at higher temperatures, the relationship in $NaCl = mCl^-$ holds approximately to 250°C .

III-8 Application of Solubility Diagrams to the Deposits of the Sherridon Area

In order to apply the data of the solubility diagram it is first necessary to understand how various environmental factors affect solubility. The effects of T and ES are shown on the series of $\log fO_2$ vs pH diagrams: Fig. III-12a-d. Increase in temperature is responsible for an increase in the solubility of base metals and a decrease in the stability of pyrite. Below temperatures of 150 to 200°C , it is doubtful if enough base metal can be carried to form a volcanogenic-type deposit.

An increase in ES is responsible for a decrease in the solubility of base metals and an increase in the stability of pyrite. Thus, at high T and low ES much base metal can be carried but the pyrite field vanishes and at low T and high ES the pyrite field is large but almost no base metal content can be carried in solution. (It should be remembered that pyrite is almost the sole Fe-mineral

FIG. III-12. $f\text{O}_2$ vs pH diagrams for conditions of interest.

FIG. III-12. cont.

FIG. III-12. cont.

FIG. III-12. cont.

FIG. III-13. fO_2 vs T diagrams for conditions of interest

found in unmetamorphosed volcanogenic base metal deposits).

The solubility of Cu is more sensitive to changes in ΣS and T than either Zn or Pb. For instance, at 300°C and ΣS of $10^{-3.5}$ m, Cu is several orders of magnitude more soluble than either Zn or Pb. An increase of ΣS to 10^{-1} m changes the ratio of the solubilities so that Zn and Pb are several orders of magnitude more soluble than Cu. A decrease in T to 250°C will have the same effect. Zn and Pb respond similarly to changes in ΣS and T , with the exception that at temperatures less than 200°C , Zn is slightly more soluble than Pb and the reverse applies at higher temperatures.

The effect of pH is shown on the $f\text{O}_2$ vs T diagrams on Fig. III-13. This figure shows that the fields of the Fe-S-O minerals have a slight dependency on pH. The solubilities of the base metals are also dependent on pH, Zn being more sensitive than Cu.

Molality of NaCl, although not illustrated on any diagram, has an effect on the solubility of the base metals. Calculations (which are in agreement with Anderson, 1977) shows that the solubility of base metals decreases an order of magnitude with a reduction from 3 m to 1 m NaCl.

These observations lead to a number of basic conclusions about ore deposits which can be used to test the validity of the solution model as illustrated on the diagrams. At high temperatures ($> 300^{\circ}\text{C}$) such as found in the porphyry environment the dominant base metal should be Cu and minor amounts of Pb and Zn would only be expected in the cool periphery of the intrusion.

Ore solutions would be expected to carry several thousand ppm of base metal. At moderate temperatures (200 to 300°C) such as found in the volcanogenic environment, base metal deposits should consist of subequal amounts of Cu, Zn, and Pb. Because of the transition from Cu dominance to Zn and Pb dominance in the temperature range, zoning would be expected. Both moderate and high temperature solutions carry enough sulphide with base metal to form a deposit. Molalities of NaCl > 1 would be expected to carry enough metal to form these deposits.

At low temperatures (< 200°C), Pb and Zn would be expected to form the majority of deposits. The amount of base metal carried in solution would probably be < 10 ppm which implies large volumes of solution and significant periods of time to form a deposit. It is doubtful if sufficient metal and sulphide could be carried simultaneously to form a deposit in this temperature range.

These conclusions appear to be valid for porphyry deposits which have a Cu-rich core and a Zn-rich halo or peripheral Pb-Zn veins. Crerar *et al* (1976) have summarized the fluid inclusion data from porphyries and found that the fluid inclusions from the Cu-rich core contain several thousand ppm of metals. Fluid inclusions from the periphery of this type of deposit contain less than one hundred ppm of metal. The amount of sulphide carried by these solutions appears to have been variable in the range of 10^{-1} to $10^{-3.5}$ m. Salinities for these conditions were much greater than 3 m NaCl (Moore and Nash, 1974).

The model also works well for volcanogenic sulphide deposits, which generally contain subequal amounts of Cu and Zn. Massive

sulphide deposits have well-developed zoning as predicted by this model. The amounts of metal and sulphide carried by the solutions which form this type of deposit have not been determined. Fluid inclusion studies indicate salinities up to 2.5 m NaCl (Sato, 1973).

Application of the model to low temperature deposits is problematic because the role of chloride complexing at these temperatures is not well established. Nevertheless, most low temperature base metal deposits are shale and carbonate-hosted Pb-Zn deposits, so that the model probably has some validity. Roedder (1972) has demonstrated through fluid inclusion studies that these deposits were formed from ore solutions with salinities of about 5 m NaCl. Carpenter *et al* (1974) in their study of oil field brines found base metal concentrations to be of the order 10 to 100 ppm.

Having established the effects of environmental conditions, as well as a reasonable degree of validity for the model, it is possible to apply it to the Sherritt Gordon deposit. For this purpose, Fig. III-14 was constructed, which describes the cooling path of a typical, saturated, volcanogenic ore fluid based on the data contained in Table III-4. Saturated solutions are used because a temperature decrease of less than 50°C will bring a 10 percent saturated solution to 100 percent saturation. It is probable that volcanogenic fluids have cooled several hundred degrees before they flow out onto the sea floor, suggesting they are at saturation. It has already been established that the Cu/Zn ratio of the Sherritt Gordon deposit was about 1 or a little greater,

FIG: III-14. Log f_{O_2} vs. T diagram modified to show the path of a cooling ore solution

thus the solutions which formed this deposit must have had a temperature of 300°C or a little less when they first came to rest on the sea floor. The solutions must have also been under a hydrostatic pressure equal to a column of seawater of at least 1.5 km deep to keep from boiling (Guber and Ohmoto, 1978).

Traditionally, distal volcanogenic deposits were thought to form from a volcanic fluid which had cooled as it moved slowly away from its source (Large, 1977). Recent work by Turner and Gustafson (1979) shows that this view is not necessarily true. Dense brines can move at rates of tens of km per hour and travel distances of 100 km. Brines hold large quantities of heat (White *et al*, 1971) so that they need not cool appreciably while moving. The composition of a distal deposit may, therefore, be controlled by the geothermal gradient of its source area. The higher the thermal gradient of the source area, the hotter and the more Cu-rich the brine flowing onto the sea floor. However, most distal deposits are Zn-rich, indicating that the conditions which allowed for the Cu-rich (i.e. high temperature) solutions flowing onto the sea floor to form the Sherritt Gordon deposit are probably rare.

A high geothermal gradient leads to above-average instability in a geothermal system as well as increased energy available to the system. These conditions also lead to intense fracturing of the rocks of the source area as well as hydrothermal explosive activity (Muffler *et al*, 1971). Following a suggestion by Turner and Gustafson (1978) it is here proposed that altered material could be carried as a suspension in a brine and precipitated when the brine comes to rest. Henley and Thorneley (1979) report altered rocks in polyolithic transported ores.

Once the ore solution had come to rest it would begin to cool following the path shown on Fig. III-14. Cooling from 300 to 250°C would precipitate almost all of the Cu but only part of the Zn. Further cooling would precipitate the remainder of the Zn. This would lead to a progressive change in metal ratio in the deposit from a Cu-rich bottom to a Zn-rich top such as found in the west orebody. Chloride dilution, as suggested by Crerar (1974) could also aid this process.

Unfortunately, to form a deposit with an average thickness of 15 m from a brine pool carrying about 1,000 to 10,000 ppm of combined metals would require that the pool be between 3,000 and 30,000 m deep. In all probability the pool which formed the Sherritt Gordon deposit was not more than a few hundred metres thick. Therefore, the brine pool must have slowly been replenished by fresh fluids over time; considered by many geologists (J. Franklin, pers. comm.) to be about 1,000 years for massive sulphide deposits. Also, the Cu/Zn ratio of the mineralizing fluid must have become progressively smaller with time.

Changes in the metal ratio of a hydrothermal system can be accounted for in a number of ways (Sangster and Scott, 1976), of which the most probable is cooling of the ore-bearing solution at the source area with time. Cathles (1975) estimates that geothermal areas have average lifetimes of less than 5,000 years or about the same order of magnitude. It is thought necessary to form a volcanogenic sulphide deposit. Geothermal systems cool with time so that a decrease in the Cu/Zn ratio of their mineralizing fluids is highly probable. This change would account for the metal zoning in the west orebody.

The east orebody is longitudinally zoned, probably as a result of a pH gradient set up by the carbonate rocks at the east end of this orebody. The brine pool would have become more alkaline because of interaction with the carbonate rocks. Because Zn is much more sensitive to a shift in pH than Cu (Fig. III-14), Zn would have tended to precipitate before it reached the high pH environment near the carbonates. This would lead to longitudinal zoning with a Cu-rich zone near the carbonate rocks and a Zn-rich zone on the periphery.

The Fe-S-O minerals precipitated with the base metal sulphides must also be considered. Fig. III-15 shows that above 225°C magnetite would be the stable Fe-S-O mineral and below 225°C pyrite would be stable with the base metal sulphides. In unmetamorphosed deposits, however, pyrite is essentially the only Fe-S-O mineral associated with base metal sulphides (Sangster and Scott, 1976). This inconsistency indicates that the available thermochemical data might be in error.

Recent experimental work by Crerar et al (1978) is of no help in explaining this problem. Their data suggest that the pyrite field is very small at 250°C and vanishes at less than 300°C leading to one or the other of two conclusions: 1) that the available thermochemical and experimental data are in error, or 2) that there is a rapid equilibration of the Fe-S-O minerals with the surrounding fluid as the temperature decreases. Fe-S-O minerals are known for their rapid equilibration in the metamorphic environment, therefore, the second conclusion is preferred.

The Sherritt Gordon and related subeconomic deposits contain about 50 percent pyrrhotite, which is undoubtedly the result of

metamorphism of the original pyritic assemblage (Sangster and Scott, 1976). The absence of pyrrhotite from the pegmatite-hosted orebodies can be explained by the equilibration of the pyrrhotite in the pegmatitic fluids to form pyrite upon cooling of the pegmatitic body in much the same manner as hypothesized for the equilibration of the original sulphide assemblage during deposition.

The distribution of sulphide facies iron formation in the Sherridon area is the last factor which must be considered. Normally, the cooled solutions of a brine pool which are steadily being replaced by fresh, hot dense brine should spill out of the pool and flow out onto the surrounding area to form volcanogenic iron formations. It is apparent in the Sherridon area that iron formations do not connect the various base metal deposits on individual stratigraphic horizons. Spent solutions must have been diluted with seawater in such a way that excess SiO_2 , Fe and S made available by cooling were absorbed by the seawater without precipitation.

Only in the presence of carbonate rocks are sulphide facies iron formations present. In this case, the high pH conditions created by the presence of the carbonate rocks must have been sufficient to cause the final precipitation of Fe, S, and SiO_2 remaining in cooled solutions or from a saturation of the general environment (Iwao, 1970).

To summarize, the Sherritt Gordon and related base metal sulphide deposits were formed from hot, dense brines entering the basin from an area of high geothermal gradient. Mechanical debris settled from the solutions upon their slowing and coming to rest

in hollows on the basin floor. Base metal sulphides were then precipitated by either cooling, change of pH or dilution of the solution. Coexisting F-S-O minerals deposited with the base metal sulphides equilibrated with the solution as it cooled. Spent solution mixed with seawater in such a way as to prevent the formation of volcanogenic iron formations except in the high pH zones associated with the carbonate rocks.

CHAPTER IV

STRUCTURE AND METAMORPHISM

IV-1 Structure: Introduction

An understanding of the folding within the Kisseynew belt is complicated by the lack of marker horizons and stratigraphic top indicators. Modern structural studies have been carried out by Pollock (1964, 1965) and Bailes (1975, 1979). A common deformation pattern in the Kisseynew belt appears to consist of early isoclinal folds followed by two or more periods of folding giving rise to complex interference patterns.

IV-2 Description of Deformational Events

Age relationships among fabric elements and major structures indicate four periods of deformation in the Sherridon area. These are listed in Table IV-1 together with the succession of metamorphic periods. The traces of the axial planes and the plunge directions of the folds in the Sherridon area are shown on Fig. IV-1.

IV-2-1 D_1 : Formation of Regional Schistosity

The first deformational event recognized in the rocks of the Sherridon area comprises the formation of the regional schistosity (S_1) which is everywhere parallel to the compositional layering of the gneisses (S_0). Rootless isoclinal folds (F_1) have a penetrative axial planar fabric of biotite flakes. Minor F_1 folds are rare and no major F_1 fold was recognized in the Sherridon

Table IV-1 Structure and Metamorphism of the Sherridon Area

D ₁	Intrafolial isoclinal folds (F ₁) with regional axial planar schistosity (S ₁)	M ₁	High grade regional metamorphism producing upper amphibolite facies mineral assemblages. M ₁ persists through D ₂ .
D ₂	East trending isoclinal folds (F ₂) with axial planes dipping moderately to the north and axes plunging easterly.		
D ₃	Northwest trending, neutral, isoclinal folds (F ₃) with shallow east plunging axes containing axial surfaces dipping moderately to the east.		
D ₄	North-northeast and northwest trending cataclastic faults and fractures.

FIG. IV-1. Axial Traces in the Sherridon Area

area although nappes and recumbent folds of this generation have been recognized by McRitchie (in Bailes, 1975) to the north and by Bailes (1975) to the southeast. Tuckwell (1979) tentatively identified a macroscopic F_1 fold on a small scale map of the Sherridon area. This author favours an interpretation lacking F_1 folds since neither a closure nor lithologic symmetry can be recognized.

IV-2-a D_2 : E-W Folding

The second deformational event occurred after the peak of metamorphism, but while the rocks were still ductile enough to create tight to isoclinal folds (F_2). A weak S_2 fabric was recognized in only one minor fold nose, otherwise F_2 folds do not show an axial planar fabric. McRitchie (in Bailes, 1975) states that an axial planar fabric was found locally in F_2 fold hinges to the north of the Sherridon area. Fabrics folded by D_2 are recrystallized and show little sign of strain.

Axial planes of the F_2 folds are overturned to the south and dip about 45° N. Where they have not been reoriented by later periods of deformation, F_2 folds plunge shallowly to the east (Stauffer, 1974). F_2 minor folds are infrequent but not as rare as F_1 minor folds. They vary in wavelength from a few cm to 10 m and have axes parallel to the mineral lineation (L_1).

One major F_2 fold occurs in the area and this is the entire structure occupied by the Sherridon Group rocks. Before being refolded by D_3 the Sherridon structure was a syncline or elongated basin. McRitchie (in Bailes, 1975) states that F_2 folds form elongated domes and basins in the area north of Sherridon because F_2 fold axes are not cylindrical. This inhomogeneity may reflect

the high temperature and ductile nature of the rocks during deformation (Ramsay, 1967).

Drilling data of the Sherritt Gordon Company in the vicinity of their orebody has been plotted on Fig. IV-2. Fig. IV-2a shows that the rocks near the west orebody on the west side of the structure dip to the east and flatten with depth. Fig. IV-2b shows that the rocks near the east orebody on the east side of the basin dip to the east near surface, but pass through the vertical to dip to the west, becoming flat lying at depth. This geometry implies a closure under the structure, which means it is a synform.

Disturbances probably produced by the D_2 event have been plotted and found to show considerable dispersion (discussion follows separately in Section IV-5).

IV-2-c D_3 : Folding

The third deformational period folds D_2 structures isoclinally and produces what appears to be a macroscopic type 2 interference pattern. Where measured, a few poorly defined fold axes about refolded F_2 folds were found to plunge shallowly to the east, coaxial with associated F_2 fold axes. The author cannot provide an explanation for this but considering the complexity of this area it should not negate a large scale type 2 interference pattern.

One major F_3 fold, the Sheila Lake fold (Stockwell, 1950), occurs in the Sherridon area, refolding the F_2 isoclinal syncline

FIG. IV-2. Cross sections through the Sherritt Gordon mine:
 a) 3500' North of West Shaft; data from company records
 b) 2000' East of East Shaft; after Farley (1949)

about a tight to isoclinal, neutral fold surface. An S_3 fabric was not recognized.

IV-2-4 D_4 : Faulting

D_4 faults comprise narrow, continuous, linear zones filled with cataclastically crushed rocks. These zones may be up to 10 metres thick. Only one major fault was recognized in the area, the Molly Lake fault, which severs the Sherridon structure. This fault is sinistral, trends about 25° , and Froese (pers. comm., 1979) estimates the separation to be about 1 km. A few subsidiary faults trending 25° and 100° with separations of less than 50 m occur in the Transit Lake area (Fig. II-2).

IV-3 Strain

Strain in rocks can be measured by the deformation of primary features such as clasts and pillows. A comprehensive study could not be carried out in the Sherridon area because of the paucity of strain markers, but sufficient evidence was gathered to draw a few conclusions.

Clasts in a mafic fragmental unit lying to the east of the West Orebody show an enormous variation in strain over a strike length of about 1 km. Elongation ratios of clasts in two dimension vary from 1:1 to about 200:1. This large variation suggests that strain is highly variable throughout the area, due to the complexity of the folding which has occurred. It also suggests that strain reached very large proportions.

IV-4 Foliation

Foliation here refers to all planar structural features of metamorphic origin, including schistosity, gneissosity, and planar mineral orientations. Most of the rocks of the Sherridon area show evidence of foliation, although in granitic gneisses and amphibolites the structure can be indistinct. Only the pegmatites are consistently non-foliated. Most of the foliation is produced by planar, preferred orientation of single crystals, mainly biotite. In places it is accompanied by flattened quartz or feldspar grains or relict features such as clasts.

Layering in the Nokomis Group could not be assigned to a metamorphic or pre-metamorphic origin because of the intense migmatization present.

Layering of pre-metamorphic origin, such as bedding or igneous layering is present in many parts of the Sherridon Group. These primary structures have been modified by both tectonic and metamorphic processes so that they are invariably parallel to the foliation.

Foliations were plotted and contoured on a Schmidt equal area net by use of the Carleton University Department of Geology computer program "Stereo" developed by W. Yzerdraat. In order to produce discrete groups of foliations, it was necessary to divide the thesis area into three domains corresponding to three major fold closures (Fig. IV-3, in pocket). Domains 1 and 3 correspond to F_3 folds and domain 2 corresponds to an F_2 fold. Plots for domains 1 and 3 give the same direction of plunge although the fold axes of domain 1 have a slightly steeper plunge. The

reoriented F_2 fold in domain 2 has an axis very similar to the F_2 folds of domains 1 and 3.

IV-5 Lineation

Two varieties of lineation can be distinguished in the Sherridon area: mineral lineations caused by elongate crystals or mineral aggregates and the axes of minor folds. Almost all of the lineations measured are the long axes of amphibole crystals, quartz aggregates, sillimanite nodules or relict clasts which have been elongated. Well-developed lineations are rare. Folds are even less abundant.

Lineations were also plotted by the computer program "Stereo." Lineations from all three domains plot on what appears to be a small circle. Lineations plotting on a small circle have been interpreted as resulting from flexural slip folding of an early (F_2) lineation. Pollock (1964) found a similar distribution of lineations in rocks to the west of Sherridon.

IV-6 Metamorphism: Introduction.

Metamorphic zones in the Kiseynew gneiss belt are shown by Bailes and McRitchie (1978). More detailed studies of the southern part of the belt have been carried out by Robertson (1953), Bailes (1975, 1979), and Froese and Gasparrini (1975). The results of these studies indicate that nearly uniform conditions of high grade metamorphism exist throughout the belt, decreasing with steep metamorphic gradient into the medium and low grade rocks of the adjacent volcanic belts.

The Nokomis gneisses display evidence of partial melting, in some instances up to 50 percent. Over large areas, they have been converted into a migmatite complex and in other areas into featureless granitoid gneisses. At Sherridon, the Sherridon gneisses in stratigraphic distance of a few metres from the Nokomis migmatites exhibit only incipient melting of 1 or 2 percent. In other areas (Bailes, 1979; Froese and Gasparrini, 1975), the transition between Sherridon and Nokomis rocks is not marked by a discontinuity in metamorphic grade. Apparently, the Nokomis Group contains rocks differing widely in age, a basement gneiss complex at Sherridon and an Amisk Group equivalent at File Lake.

Retrogression is a common feature in most of the rocks examined. Feldspars have been altered to sericite and epidote and ferromagnesian minerals converted to chlorite. In a few places well-developed flakes of muscovite, which exhibit parallel extinction with biotite flakes, look primary. It is thought that these flakes formed during the initial decrease in metamorphic grade, which fell into the muscovite field.

IV-7 Mineral Assemblages

Metamorphic assemblages were developed during D_1 as evidenced by the primary schistosity S_1 . Metamorphism must have persisted through D_2 and D_3 because the fabrics deformed by these tectonic events have been annealed. The poor fabrics developed during D_2 and the lack of a fabric associated with D_3 suggests that metamorphic conditions were declining (Table IV-1).

A great number of mineral assemblages, observed in a variety of rock types throughout the Sherridon area are consistent with a uniform grade of metamorphism (Table IV-2). These assemblages can be represented on a number of diagrams, some of which are more effective than others in illustrating metamorphic conditions. In the presence of quartz, plagioclase of constant composition, magnetite and ilmenite, many of these assemblages can be represented in the system (Froese, 1978):

Neglecting the effects of other components, phase relationships within this system are determined at fixed values of P, T, and $P_{\text{H}_2\text{O}}$. Mineral assemblages not including K-bearing minerals can be represented on the AFM face of an AKFM tetrahedron (Fig. IV-4a). The classical ACF diagram (Fig. IV-4b) illustrates phase relationships in grossular bearing assemblages which cannot be shown on the AFM diagram.

Mineral assemblages including K-bearing minerals can be projected onto the AFM face of an AKFM tetrahedron. Alternatively, in biotite-bearing assemblages, phase relations among these minerals can be shown on the biotite composition surface (Froese, 1978). Changes in mineral compatibilities brought about by each univariant reaction are seen on the biotite surface by a change in topology. This allows phase relations in K-feldspar-bearing as well as K-feldspar-free rocks to be shown in a convenient manner.

Quartzo-feldspathic and
pelitic gneisses; amphibolitesGarnet-
Anthophyllite
Rocks

Calc-Silicate rocks

Q	Pl	Kf	Bi	Sl	At	Al	Hb	Cu	Co	Cpx	Ep	Gr
Q	P1		Bi									
Q	P1		Bi			Al						
Q	P1		Bi	Sl								
Q	P1		Bi	Sl		Al						
Q	P1		Bi						Co			
Q	P1		Bi					Cu				
Q	P1		Bi			Al		Cu				
Q	P1		Bi				Hb	Cu				
Q	P1		Bi			Al	Hb	Cu				
Q	P1		Bi			Al	Hb					
Q	P1		Bi				Hb					
Q	P1		Bi				Hb					
Q	P1		Bi				Hb					
Q	P1		Bi			Al	Hb					
Q	P1	Kf	Bi									
Q	P1	Kf	Bi				Hb					
Q	P1	Kf	Bi			Al						
Q	P1	Kf	Bi			Al	Hb					
Q	P1	Kf	Bi	Sl								
Q			Bi	Sl		Al			Co			
Q			Bi		At	Al			Co			
Q			Bi		At	Al			Co			
Q	P1		Bi		At	Al	Hb	Cu				
Q	P1		Bi		At	Al		Cu				
Q	P1		Bi									
Q	P1		Bi			Al	Hb			Cpx		
Q	P1		Bi							Cpx		
Q	P1		Bi			Al	Hb			Cpx		
Q	P1		Bi			Al				Cpx		
Q	P1		Bi				Hb			Cpx		
Q	P1		Bi				Hb			Cpx		
Q	P1		Bi				Hb			Cpx		
Q	P1	Kf	Bi			Al	Hb			Cpx		
Q	P1	Kf	Bi				Hb			Cpx		
Q	P1	Kf	Bi				Hb			Cpx		
Q	P1	Kf	Bi				Hb			Cpx		
Q	P1	Kf	Bi				Hb			Cpx		
Q	P1	Kf	Bi			Al	Hb			Cpx		
Q	P1		Bi							Cpx	Ep	Gr
Q	P1		Bi							Cpx	Ep	Gr
Q	P1		Bi							Cpx	Ep	Gr
Q	P1		Bi							Cpx	Ep	Gr

FIG. IV-4. Mineral assemblages plotted on: a) AFM and b) ACF diagrams.

Of the assemblages compatible with biotite, all but five could be shown on one topology of the biotite composition surface. Certain assemblages indicate two reactions which give rise to new topologies:

These reactions did not go to completion. Slight variations in metamorphic conditions and/or plagioclase composition are sufficient to account for these additional assemblages. Thus, the metamorphic grade within the area is essentially uniform and mineral assemblages do not define isograds.

The occurrence of Fe-sulphide minerals as accessories in quartz-rich, pelitic, and calc-silicate gneisses can be related to the coexisting silicates. Thus an AFM or biotite composition surface diagram can be divided into two fields, each characterized by the stability of either pyrite or pyrrhotite (Froese, 1976). Mineral compatibilities observed in sulphide-silicate rocks essentially agree with the phase relations shown in Fig. IV-4a and Fig. IV-5. The chief iron sulphide in most of the rocks observed is pyrrhotite although some pyrite is present as well. The occurrence of pyrite in relatively iron-rich rocks appears to be anomalous and it is interpreted as a retrograde mineral.

IV-8 Metamorphic Conditions

The mineral assemblages of the Sherridon Group can be plotted on biotite surfaces which occupy fields around an invariant point: cu-al-at-hb-opx, shown on Fig. IV-6 (Froese, pers. comm., 1979). Muscovite is not stable and partial melting has

FIG. IV-5. Mineral assemblages shown on the Bi composition surface. after Froese (1978, 1979).

FIG. IV-6. Mineral reactions at the amphibolite-granulite transition in Bi-bearing mafic rocks; plus Qz, Pl of constant comp., Mt and Il. E. Froese (pers. comm., 1979)

commenced in a few areas, which suggests a minimum metamorphic temperature of 650°C (Kerrick, 1972). It is unlikely that metamorphic conditions exceeded 700°C , otherwise partial melting would have been more prominent in the Sherridon Group.

Estimates of pressure can be made using the sphalerite geobarometer. Scott (1973) has investigated the equilibrium among pyrite, hexagonal pyrrhotite and sphalerite at various temperatures and pressures. He proposed that the variation in mole fraction of FeS in sphalerite was independent of temperature.

Electron microprobe analyses of 17 coexisting sphalerite-pyrite-pyrrhotite assemblages from the Sherridon mine have been compiled on Table IV-3. They yield a mean concentration of FeS in sphalerite of 0.136 percent. The mean composition and range of the FeS in sphalerite have been plotted on Fig. IV-7 for 650°C , the minimum temperature derived from consideration of the silicate assemblages. This gives a pressure of 6.0 ± 1 kb. A value of 700°C would give a pressure of 5.5 ± 1 kb.

In Fig. IV-8, part of a petrogenetic grid is shown (D. M. Carmichael, pers. comm., 1978). A path of metamorphic conditions compatible with the assemblages found in the Sherridon area and the ferromagnesian mineral stabilities plotted on the grid is shown reaching a peak of 675°C and 6 kb just after crossing the curve representing the beginning of melting for a system saturated in H_2O .

TABLE IV-3
SPHALERITE ANALYSES

		SAMPLE E				
		1	2	3	4	5
WEIGHT %	S	33.38	33.93	33.55	33.37	33.63
	Zn	60.48	59.78	60.79	59.88	59.41
	Fe	7.44	7.54	6.94	7.54	7.87
TOTAL		101.30	101.25	101.28	100.79	100.92
ATOMIC %	S	1.041	1.058	1.046	1.041	1.049
	Zn	.925	.915	.930	.916	.909
	Fe	.133	.135	.124	.135	.141
	FeS	.126	.129	.118	.128	.134

		SAMPLE F				
		1	2	3	4	5
WEIGHT %	S	33.56	33.79	33.91	34.13	34.33
	Zn	58.70	59.70	59.13	58.44	59.20
	Fe	7.51	7.30	7.38	7.54	7.53
TOTAL		99.77	100.79	100.42	100.11	101.06
ATOMIC %	S	1.047	1.054	1.058	1.065	1.071
	Zn	.898	.913	.904	.894	.906
	Fe	.135	.131	.132	.135	.136
	FeS	.131	.125	.127	.131	.131

		SAMPLE G				
		1a	2	3a	3b	5
WEIGHT %	S	33.80	33.90	33.96	33.93	33.95
	Zn	59.47	58.95	59.93	58.70	58.23
	Fe	7.68	7.42	7.09	8.08	8.68
TOTAL		100.95	100.27	100.97	100.72	100.87
ATOMIC %	S	1.054	1.057	1.059	1.058	1.059
	Zn	.910	.902	.917	.898	.891
	Fe	.138	.133	.127	.145	.156
	FeS	.132	.129	.122	.139	.149

FIG. IV-7. Peak metamorphic pressure at Sherridon, mean and range FeS in sphalerite shown: diagram from Scott (1973).

CHAPTER V

INTERPRETATION AND CONCLUSIONS

V-1 Introduction

This chapter presents an interpretation of the depositional environment of the rocks of the Sherridon Group and a synopsis of tectonic events in the Kisseynew belt. Salient characteristics of the rocks of the Sherridon Group are summarized as required in support of inferences.

V-2 Depositional History of the Sherridon Group

The rocks of the Kisseynew gneiss belt have been correlated with the rocks of the Flin Flon belt (Bailes, 1979) and the depositional environment of the Sherridon area can in part be deduced from the depositional environment of the low grade rocks immediately to the south.

The Amisk volcanic belt near Flin Flon is formed from 5 km of mafic island arc tholeiitic flow rocks extruded in a sub-aqueous environment (Stauffer, 1974) with some subaerial extrusion during late stages of volcanism (Ayres, 1978). Felsic volcanic rocks, which are abundant near the top of the volcanic section compose 20 percent of the Amisk Group. The felsic rocks are in general tholeiitic but late felsic calc-alkaline domes have been recognized in the Files Lake area (Bailes, 1979) and a calc-alkaline volcanic complex is present near Amisk Lake (Fox, 1976).

There is a general upward increase in the amount of pyroclastic rocks, heterolithic laharic breccias and mafic volcanoclastic

the Sherridon area in the latter part of Chapter IV in this thesis.

The Sherridon Group and the slopes of the island arcs of Japan and the South Pacific show a great similarity in type, distribution and composition of volcanic rocks and of the sulphide mineralization. Flows, layers of volcanic ash, and laharic mudflows occur as isolated horizons in an otherwise sedimentary sequence (Mitchell, 1970). The volcanic rocks of island arcs show a progression in composition from island arc tholeiitic rocks at the base to calc-alkaline rocks at the top (Miyashiro, 1974). Calc-alkaline and tholeiitic rocks can be found intercalated at the top and on the margin of volcanic arcs. Only one occurrence of alkaline volcanic rocks in an island arc has been recorded, at the margin of the Japanese arc in a regime associated with rifting and back-arc basin formation (Sugisaki *et al.*, 1972). The volcanic rocks of the Sherridon Group show all the characteristics of volcanic rocks found on the margins of volcanic arcs. Frequently, intrusive rocks compositionally related to the volcanic rocks occur within the shelf and slope sediments (Mitchell, 1970).

Thus, the Sherridon Group, because of its physical relationship to the Flin Flon volcanic belt, and its type of sedimentation, volcanism, and sulphide mineralization is considered to represent environmental conditions as found in the transition from slope to shelf of a prograding volcanic arc succession.

Fig. V-1 is a cross-section illustrating conditions in the Flin Flon-Sherridon area at the time of the deposition of the Sherritt Gordon massive sulphide deposits. It illustrates the position of the Sherridon area relative to the Flin Flon volcanic

sediments; this feature has been attributed to a shallowing of water and progressive change in magmatic activity (Bailes, 1979). Mass wasting of this debris has produced submarine fan deposits with layers 5 cm to 11.5 m thick. The beds are well graded and usually consist of the entire Bouma sequence (parts A to E). The volcanic debris forming these layers is quite immature, consisting on the average of 10 percent quartz, 25 percent plagioclase, and 69 percent rock fragments and matrix; the grain size is 0.5 to 1.0 mm. These rocks have been interpreted as graywacke turbidites (Bailes, 1979). Rarely, lenses of carbonate-rich rocks are intercalated with the graywacke.

The Missi Group, 1.5 to 3.5 km thick, overlies the Amisk Group with an erosional unconformity marked by a basal conglomerate. The Missi Group is characterized by thick poorly sorted beds, some of which exhibit cross-stratification with individual sets between 10 to 30 cm thick. The rocks consist of immature volcanic debris with an average composition of 40 percent quartz, 15 percent feldspar, and 45 percent lithic fragments. These rocks are lithic arenites and have been interpreted as an alluvial deposit (Mukherjee, 1974).

Attention has been drawn to the similar chemical composition of rocks of the Missi and Sherridon Group in an earlier chapter and in this thesis. However, it is unlikely that the depositional environment of the two groups was the same. The Sherridon Group near Sherridon lacks conglomerates, includes marbles and contains volcanogenic massive sulphide deposits. The presence of the sulphide deposits places a minimum on the depth of water during deposition, because the ore-bearing solutions must have been below

their boiling point. In Chapter IV, the temperature of a 3 m NaCl was estimated to be 250 to 300°C. A depth of 1.5 km of water is necessary to keep such a solution from boiling. (This depth is compatible with the depth of deposition of many volcanogenic sulphide deposits in the Japanese volcanic arc--Guber and Ohmoto, 1978.)

A depth of 1.5 km of water during the deposition of the lower part of the Sherridon Group rules out a subaerial environment as proposed for the Missi Group. Instead, rocks of the lower part of the Sherridon Group display characteristics of the distal turbidites of the Amisk-Nokomis graywackes. The rocks of both groups were deposited in beds 5 cm and 11.5 m thick and are interbedded with carbonates. Carbonate as partings or as thin beds up to 1 cm thick is found between many of the beds of lithic arenite in the Sherridon area. Mitchell (1970) stated that this association is a typical sedimentary facies in the island arc turbidites. Metamorphism has obscured any graded bedding, if it was ever present. Compositional variation within individual layers is not apparent; it would not be expected in a turbidite composed of immature lithic fragments and a small amount of matrix.

The upper parts of the Sherridon Group arenites show thick uniform beds, in places with heavy mineral laminae and cross-bedding. The thickness of beds and size of cross-beds is about the same as for the Missi Group. This indicates a shallowing of water and change in depositional environment during the sedimentation of the Sherridon Group. The development of a more feldspathic arenite in the shallow water environment may indicate the operation of depositional processes favouring the production of a

more mature arenite. Dickinson (1970, 1974) states that a change in provenance to the felsic part of a volcanic pile or to eroded plutonic roots may also be responsible for this phenomenon.

The upper part of the Sherridon Group includes thick units of limestone and feldspathic arenite. This association of rock types as well as the absence of conglomerate suggest a shallow water depositional environment and preclude a subaerial depositional environment as has been suggested for the Missi Group. The change from deep water and the deposition of turbidites to shallow water deposits takes place over a stratigraphic thickness of 1 km. Considering the highly variable nature of isostatic adjustments in volcanic belts (Dickinson, 1974), a change of this nature over a stratigraphic thickness of 1 km is not unreasonable.

The carbonate rocks of the Sherridon Group probably represent deposition in various environments. The hornblende- and diopside-bearing calc-silicates at the base of the Sherridon Group must have been formed from the metamorphism of clay-rich carbonate rocks which are indicative of a low energy environment (Plumley et al, 1962). - A low energy environment is consistent with the interpretation of the associated arenites as deep water turbidites. The carbonate rocks of the limestone formation overlying the shallow water arenites are interbedded with calc-silicates and quartzofeldspathic arenites. This assemblage is indicative of a variable energy environment such as a beach or lagoon (Plumley et al, 1962). A possible modern analogue can be found in the Laguna Madre area along the Texas coast (Rusnak, 1960).

The interpretation of the limestone formation as a lagoonal facies need not imply a biogenic origin. Inorganic carbonates do

occur among the rocks of volcanic belts as by-products of fumarolic activity (Sangster and Scott, 1975). However, since the paleo-latitude of the Sherridon area during the deposition of these rocks was about 30° , a biogenic origin in shallow water is not precluded.

Carbonate deposition must have been continuous during the deposition of much of the Sherridon sequence because of its ubiquitous presence throughout the rocks of the Group. During periods of quiescence, co-deposition of carbonates and clays would produce calcareous rocks like those at the base of the Sherridon Group. Deposition during periods of turbidite emplacement would account for the carbonate partings between individual turbidite beds. Carbonate material entrained in turbidites would account for some of the calc-silicates which occur in thicker horizons; it is doubtful that the calcareous rocks were produced by movement of carbonate material during diagenesis because of the stratigraphic nature of the layers. Mitchell (1970) observed these modes of occurrence of carbonate rocks in an island arc of the South Pacific area.

The basal calc-silicate rocks of the Sherridon area appear to be transgressive in that to the east they are underlain by lithic arenites. In contrast to the view expressed by Robertson (1953) and Davies *et al* (1962), it is not inferred that these calc-silicates mark a transitional environment between the deposition of the Sherridon and Nokomis Groups.

The carbonate rocks of the limestone formation are overlain by a graywacke similar in composition to the graywackes of the Nokomis Group but similar in depositional style to the shallow water arenites with which it is associated. Dickinson (1974)

stated that isostatic compensation in volcanic belts is variable because of the constant adjustments required to keep pace with volcanic activity. This isostatic instability may lead to the exhuming of older sediments and their redeposition in a new sedimentary environment. Such a process is thought to be responsible for placing a rock similar in composition to the deepwater graywackes of the Amisk Group among the shallow water arenites and carbonates of the upper part of the Sherridon Group.

The presence of cherts and sulphide deposits at several stratigraphic positions within the Sherridon Group supports the conclusions which have been drawn concerning its depositional environment. The precipitation of volcanogenic massive sulphide deposits and of bedded chert deposits must take place in a sub-aqueous environment. Both bedded cherts and massive sulphide deposits have been described from the rocks of the Osugi volcanic belt in Japan (Mitchell and Bell, 1973) and the Philippine volcanic arc (Bryner, 1969). Three types of mineralization are present, typical of the so-called Besshi deposits: 1) Cu-Zn massive sulphide layers, 2) banded cherts with Fe-sulphides, and 3) bedded Cu-Fe ore, often in association with pelitic schists (Kanehira and Tatsumi, 1970). In all cases, these deposits are found in a succession of mafic tuffs and volcanoclastic sediments with minor limestone interlayers, deposited on the shelves and slopes of volcanic belts. These three types of mineralization are remarkably similar to the three types of mineralization observed in the Sherridon area. These three types of mineralization have been related to volcanogenic processes in modern volcanic belts (Mitchell and Bell, 1973) and have been related to volcanogenic processes in

the Sherridon area in the latter part of Chapter IV in this thesis.

The Sherridon Group and the slopes of the island arcs of Japan and the South Pacific show a great similarity in type, distribution and composition of volcanic rocks and of the sulphide mineralization. Flows, layers of volcanic ash, and laharic mudflows occur as isolated horizons in an otherwise sedimentary sequence (Mitchell, 1970). The volcanic rocks of island arcs show a progression in composition from island arc tholeiitic rocks at the base to calc-alkaline rocks at the top (Miyashiro, 1974). Calc-alkaline and tholeiitic rocks can be found intercalated at the top and on the margin of volcanic arcs. Only one occurrence of alkaline volcanic rocks in an island arc has been recorded, at the margin of the Japanese arc in a regime associated with rifting and back-arc basin formation (Sugisaki *et al.*, 1972). The volcanic rocks of the Sherridon Group show all the characteristics of volcanic rocks found on the margins of volcanic arcs. Frequently, intrusive rocks compositionally related to the volcanic rocks occur within the shelf and slope sediments (Mitchell, 1970).

Thus, the Sherridon Group, because of its physical relationship to the Flin Flon volcanic belt, and its type of sedimentation, volcanism, and sulphide mineralization is considered to represent environmental conditions as found in the transition from slope to shelf of a prograding volcanic arc succession.

Fig. V-1 is a cross-section illustrating conditions in the Flin Flon-Sherridon area at the time of the deposition of the Sherritt Gordon massive sulphide deposits. It illustrates the position of the Sherridon area relative to the Flin Flon volcanic

belt and shows the distribution and relationships of the various rock types. From the general discussion of the Flin Flon area by Bailes (1971) and the more detailed studies of this thesis, the following sequence of events was deduced.

Deposition began with the deposition of the linear volcanic edifice of the Flin Flon belt, the base of which is composed of subaqueous extruded mafic flow rocks. As the volcanic edifice rose to wave base and volcanism became more explosive in nature, its upper part was eroded and reworked. The eroded material was redeposited as the graywackes of the Amisk and Nokomis Groups on the margin of the volcanic arc. Felsic volcanism commenced after the volcanic belt had become well established. Reworked volcanic material formed alluvial, fluvial, and deltaic deposits of the Missi Group. Some of this material was transported to the edge of the volcanic belt where it was deposited as the Sherridon Group, on the shelf and slope overlying the graywacke turbidites of the Nokomis Group.

Felsic volcanism produced the source material of the Missi and Sherridon Groups giving rise to the formation of both subaerial and submarine deposits overlying and adjacent to the volcanic rocks of the Flin Flon belt, and also the lithic arenites of the Sherridon Group. It is inferred that deposition of the Missi and Sherridon Groups occurred concurrently with late stage Amisk volcanism, rather than during post-Amisk time.

Postulated basement rocks below part of the Sherridon Group could have been present during the initial formation of the belt, as in the Japanese arc (Minato et al., 1969), or may have been emplaced tectonically during the deformation and orogeny following

the deposition of the Sherridon Group. A choice between the alternatives cannot be made in the Sherridon area because of the mafic intrusions along the base of the Sherridon Group.

V-3 Origin and Development of the Kisseynew Belt

The Kisseynew meta-sedimentary belt lies between two volcanic belts from which its sediments were derived (Bailes, 1971; McRitchie, 1974). Dickinson stated (1974) that a much larger amount of sediments can be derived from a volcanic belt than the volcanic rocks preserved in the belt. This is true of the Kisseynew belt, which contains at least 10 times the amount of sediments as there are volcanics preserved in the Lynn Lake and Flin Flon belts. The entire area has a striking resemblance to modern back-arc or inter-arc basins. It has about the same size, relationship to volcanic areas, distribution of volcanic rocks and sediments as modern arc-related basins described by Karig (1970, 1971, 1972) and Nino and Emer y (1961). From these relations, it could be inferred that the volcanic and sedimentary belts of central Manitoba and Saskatchewan could have been created by the process of plate tectonics. The mechanisms of such processes are discussed by Karig (1972.) Low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the meta-sedimentary and meta-volcanic belts imply a short crustal history (Anderson et al, 1973; Josse, 1974).

The structural and metamorphic history of the Kisseynew area and adjacent volcanic belts has been compiled in Table V-1. From this table, Fig. V-2 was constructed to show the evolution of the Kisseynew area.

Table V-1

STRUCTURE OF THE KISSENEW GNEISSES

SNOW LAKE - FILE LAKE
(After Barlow, 1975)FLIN FLOW
(After Stauffer, 1974)GUAY - WIMAPEDI LAKES
(After Barlow, 1975)

SHERRIDON AREA

FLIN FLOW (After Stauffer, 1974)	SNOW LAKE - FILE LAKE (After Barlow, 1975)	GUAY - WIMAPEDI LAKES (After Barlow, 1975)	SHERRIDON AREA
D1 Tight E-W trending folds (F1) with axial planes dipping moderately to the north	D1 Large recumbent folds (F1) with axial planar schistosity (S1) - regional F1 not recognized in the Snow Lake Area but S1 present	D1 Recumbent folds (F1) and axial planar regional schistosity (S1)	D1 Intrafolial isoclinal folds (F1) with regional axial planar schistosity (S1)
D2 N-S trending, tight, upright folds (F2) with endogenously developed axial planes and regional schistosity (S1)	D2 Major NNE trending open shallow plunging flexural folds (F2) becoming steeply plunging in the File Lake Area where they deform F1 structures	D2 Open, NNE trending shallow plunging, flexural folds (F2) Local foliation (S2) in fold hinges	D2 East trending isoclinal folds (F2) with axial planes dipping moderately to the north and axes plunging easterly
D3 NE trending open folds (F3)	D3 East trending, up-right quasi-flexural folds (F3) in the South and middle amphibolite facies mineral assemblages in the north	D3 North dipping, steeply plunging quasi-flexural folds (F3) related to diapiric uprise and east trending cataclastic zones	D3 NW trending isoclinal folds (F3) with shallow east plunging axes contained in a neutral axial surface dipping moderately to the east
D4 Northerly trending fault zones	D4 Late kinematic and brittle faults and fractures trending NNE	D4 Late kinematic and brittle deformation producing NNE trending fractures and faults	D4 NNE and NW trending cataclastic faults and fractures
	D5 Local retrograde greenschist facies mineral assemblages in D4 structures	D5 Local retrograde greenschist facies metamorphism in fault zones	
	D6 Major NNE trending open shallow plunging flexural folds (F2) becoming steeply plunging in the File Lake Area where they deform F1 structures	D6 Major NNE trending open, flexural slip folds (F4)	
	D7 East trending, up-right quasi-flexural folds (F3) in the South and middle amphibolite facies mineral assemblages in the north	D7 East trending isoclinal folds (F2) with axial planes dipping moderately to the north and axes plunging easterly	
	D8 NE trending open folds (F3)	D8 NE trending, open, flexural slip folds (F4)	
	D9 Local retrograde greenschist facies mineral assemblages in D4 structures	D9 Local retrograde greenschist facies metamorphism in fault zones	

Table V-1

STRUCTURE OF THE KISSKEMEY GNEISSES

PIKATOWAGAN - WILSON MOUNTAINS (After McRitchie, in Bailey, 1975)	NORTHERN SASKATCHEWAN (After Pearson, 1972)	LYNN LAKE (After Mulligan, 1966 and McGlynn, 1970)	SOUTHERN INDIAN LAKE (After Schledervitz, 1972)
<p>D1 Nappes and recumbent folds (F1) with formation of regional schistosity (S1)</p> <p>M1 High grade regional metamorphism producing upper amphibolite facies mineral assemblages. M1 is continuous into M2</p>	<p>D1 Recumbent isoclinal (F1) folds with axial planes dipping toward the core of the gneiss complex - Generally NE</p> <p>M1 High grade regional metamorphism producing upper amphibolite facies mineral assemblages</p>	<p>D1 Tight E-W trending folds (F1) with moderate plunges and axial planes dipping steeply to the South</p> <p>M1 Progressive metamorphism forming Greenschist facies mineral assemblages to the North and amphibolite facies mineral assemblages to the South</p>	<p>D1 Easterly trending recumbent isoclinal folds (F1) with regional axial planar schistosity (S1)</p> <p>M1 Regional metamorphism producing upper amphibolite (Barrovian) mineral assemblages.</p>
<p>D2 Synchronous NNE and East trending folds (F2) forming dome and basin interference pattern</p> <p>M2 High grade regional metamorphism slightly lower pressure than M1</p>	<p>D2 Tight North to NNW trending folds (F2)</p>	<p>D2 NW trending folds (F2)</p>	<p>D2 Tight to isoclinal, upright East trending folds (F2a) and localised cross folding (F2b) causing dome and basin interference pattern</p> <p>M2 Regional metamorphism producing upper amphibolite facies (Abukuma) mineral assemblages</p>
<p>D3 NNE trending folds (F3), mainly re-oriented F2 structures</p> <p>Local zones of cataclastic M3 deformation</p>	<p>D3 Open flexural folds (F3) with a northerly to easterly trend</p>	<p>D3 Local zones of cataclasis</p>	<p>D3 Local zones of retrograde assemblages in fault zones</p>
<p>D4 NE trending flexure folds (F4)</p>	<p>D4 Northerly trending fault zones</p>	<p>D4 Steep to vertical NF trending open folds (F4)</p>	<p>D4 Retrograde Greenschist facies mineral assemblages in D5 structures</p>

EVOLUTION

182

Volcanic Belt | Sedimentary Belt | Volcanic Belt
S Flin Flon Snow Lake Sheridan Lynn Lake N

D1 formation of thrusts and nappes about easterly axes
deformation and metamorphism confined to sedimentary belt and edges of volcanic belts
lateral flow predominant

D2 one or more episodes of co-axial cross-folding about easterly axes
formation of gneiss domes and complex interference patterns
deformation and metamorphism completely affect volcanic belts
vertical flow predominant

D3 open, upright folding about northerly axes
metamorphism waning

D4 late brittle fractures

green - greenschist facies
amph - amphibolite facies
gran - granulite facies

PRESENT FORM

FIG. V-2. Structural and metamorphic evolution of the Kisseynew and related belts, Manitoba; diagrams modified from Sibbald et al. (1976) and Harrison (1951).

Individual periods of tectonism and metamorphism have affected the whole region. Initially, D_1 compression of the belt caused crustal shortening and overturning towards the edges of the belt in a mushroom-like fashion. As metamorphism progressed, vertical tectonics (D_2) about easterly co-axial trends became dominant and resulted in diapirism and the formation of complex interference patterns of folding. Large bodies of granitic material were formed and emplaced at this time. With the waning of metamorphism, some north to south open folding (D_3) and brittle fracturing (D_4) occurred. This style of tectonism is what would be expected if the Kisseynew inter-arc basin and its enclosing volcanic arcs had been caught in a continental collision between the Archean rocks of the Wollaston belt to the northeast and the Superior Province to the southwest.

REFERENCES

- Anderson, G. M. (1973) The hydrothermal transport and deposition of galena and sphalerite near 100°C: *Econ. Geol.*, V. 68, p. 480-492.
- _____. (1975) Precipitation of Mississippi Valley-type of *Econ. Geol.*, V. 70, p. 937-942.
- _____. (1977). Thermodynamics and sulphide solubilities: in *Short course in application of thermodynamics to petrology and ore deposits*, ed. H. J. Greenwood, Min. Assoc. Canada, Short Course Handbook, V. 2, p. 136-150.
- Anderson, R. K., Clark, G. S. and McRitchie, D. (1973) Rubidium-strontium age determinations in the Burntwood Lake area, west-central Manitoba: in *U. of Man. Centre Precam. Studies*, 1973 Ann. Rept., p. 89-90.
- Ayres, L. D. (1978) A transition from subaqueous to subaerial eruptive environments in the Precambrian Amisk Group at Amisk Lake, Saskatchewan--a progress report: in *U. of Man., Centre Precam. Studies*, 1977 Ann. Rept., p. 36-51.
- Bailes, A. H. (1971) Preliminary compilation of the geology of the Snow Lake-Flin Flon-Sheridon area, Manitoba: *Man. Mines Br., Geol. Paper* 1/71.
- _____. (1975) Geology of the Guay Wimapedi Lakes area, Manitoba: *Man. Mineral Res. Div., Geol. Serv. Br., Pub.* 75-2.
- _____. (1979) Sedimentology and metamorphism of a Proterozoic volcaniclastic turbidite suite that crosses the boundary between the Flin Flon and Kiseynew belts, File Lake, Manitoba, Canada. *U. of Manitoba PhD. thesis*.
- Bailes, A. H. and McRitchie, W. D. (1978) The transition from low to high grade metamorphism in the Kiseynew sedimentary gneiss belt, Manitoba: in *Metamorphism in the Canadian Shield*, ed. J.A. Fraser and W. W. Heywood, *Geol. Surv. Canada Paper* 78-10, p. 155-178.
- Barnes, H. L. and Kullerud, G. (1961) Equilibria in sulfur containing aqueous solutions, in the system Fe-S-O, and their correlation during ore deposition: *Econ. Geol.*, V. 55, p. 648-688.
- Barton, P. B. and Toulmin, P. (1966) Phase relations involving sphalerite in the Fe-Zn-S system: *Econ. Geol.*, V. 61, p. 815-849.
- Bateman, J. D. (1944) Sherritt Gordon mine area: *Geol. Surv. Canada, Paper* 44-4.

- Bateman, J. D. and Harrison, J. M. (1946) Sherridon Manitoba: Geol. Surv. Canada, Map 862A, with descriptive notes.
- Blatt, H., Middleton, G. and Murray, R. (1972) Origin of sedimentary rocks: Prentice Hall, New Jersey.
- Bell, R. (1880) Report on explorations on the Churchill and Nelson Rivers, and around God's and Island Lakes: Geol. Surv. Canada, Rept. of Prog. 1878-79 pt. C.
- Bischoff, J. L. (1969) Red sea geothermal deposits: their mineralogy, chemistry and genesis: in Hot brines and recent heavy metal deposits in the Red Sea, ed. E. T. Degens and D. A. Ross. Springer-Verlag, New York, p. 368-401.
- Bruce, E. L. (1917) Schist Lake and Wekusko Lake area, northern Manitoba: Geol. Surv. Canada, Sum. Rept. 1916, p. 159-169.
- _____. (1918) Amisk-Athapapuskow Lakes district: Geol. Surv. Canada, Mem. 105.
- _____. (1929) The Sherritt Gordon copper-zinc deposit, Northern Manitoba: Econ. Geol., V. XXIV, p. 457-469.
- _____. (1933) Mineral deposits of the Canadian Shield: MacMillan, Toronto.
- Bruce, E. L. and Matheson, A. F. (1930) The Kisseynew gneiss of northern Manitoba and similar gneisses occurring in northern Saskatchewan: Trans. Roy. Soc. Canada, 3rd ser., 24, Sec. 4, p. 119-132.
- Bryner, L. (1969) Ore deposits in the Phillippines--an introduction to their geology: Econ. Geol., V. 64, p. 644-666.
- Byers, A. R. and Dahlstrom, C. D. A. (1954) Geology and mineral deposits of the Amisk-Wildnest Lakes area, Saskatchewan: Sask. Dept. Min. Res., Rept. 14.
- Byers, A. R., Kirkland, S. J. T. and Pearson, J. W. (1965) Geology and mineral resources of the Flin Flon region: Sask. Dept. Min. Res., Rept. No. 62.
- Cathles, L. M. (1975) An analysis of the cooling of intrusives by ground-water convection which includes boiling: Econ. Geol., V. 72, p. 804-826.
- Carpenter, A. B., Trout, M. L. and Pickett, E. (1974) Preliminary report on the origin and chemical evolution of lead- and zinc-rich oil field brines in central Mississippi: Econ. Geol., V. 69, p. 1191-1206.
- Clark, G. S., Anderson, R. and McRitchie, D. M. (1974) Total rock Rb-Sr ages from the Burntwood area: in U. of Man. Centre Precam. Studies, 1974 Ann. Rept., Ser. 2, p. 51-52.

Coleman, L. C. (1970) Rb-Sr isochrons for some Precambrian rocks in the Hanson Lake area, Saskatchewan: *Can. J. Earth Sci.*, V. 7, p. 338-345.

Colley, H. (1976) Classification and exploration guide for Kuroko-type deposits based on occurrences in Fiji: *Trans. Instn. Min. Metall. (Sect. B: Appl. earth sci.)*, V. 85, p. 141-148.

Crerar, D. A. and Barnes, H. L. (1976) Ore solution chemistry. V. Solubilities of chalcopyrite and chalcocite assemblages in hydrothermal solution at 200° to 350°C: *Econ. Geol.*, V. 71, p. 772-794.

Crerar, D. A., Susak, N. J., Borcsik, M. and Schwarz, S. (1978) Solubility of the buffer assemblage pyrite + pyrrhotite + magnetite in NaCl solutions from 200 to 350°C: *Geochim. Cosmochim. Acta*, V. 42, p. 1427-1437.

Cumming, G. L. and Richards, J. R. (1975) Ore lead isotope ratios in a continuously changing earth: *Earth Planet. Sci. Lett.*, V. 28, p. 155-171.

Davies, J. F., Bannatyne, B. B., Barry, G. S. and McCabe, H. R. (1962) *Geology and mineral resources of Manitoba*: Man. Dept. Mines and Nat. Res., p. 92-103.

Davis, J. C. (1973) *Statistics and data analysis in geology*: J. Wiley and Sons, New York.

Deer, W. A., Howie, R. A. and Zussman, J. (1966) *An introduction to the rock forming minerals*. Longman, London.

Derry, D. R. (1942) *The Sherritt Gordon mine: in Ore deposits as related to structural features*: Princeton Univ. Press.

Dewey, J. F. and J. M. Bird (1970) Mountain belts and the new global tectonics: *Jour. Geophys. Res.*, V. 75, p. 2625-2647.

Dickinson, W. R. (1974) Sedimentation within and beside ancient and modern magmatic arcs: *in Modern and ancient geosynclinal sedimentation*, ed. R. H. Dott and R. M. Shaver. *Soc. Econ. Paleon. Miner., Spec. Pub.* 19, p. 230-239.

_____. (1970) Relation of andesites, granites, and derivative sandstones to arc-trench tectonics. *Rev. Geophys. Space Phys.*, V. 8, p. 813-860.

Ellis, A. J. and Giggenbach, W. (1971) Hydrogen sulfide ionization and sulphur hydrolysis in high temperature solution: *Geochim. Cosmochim. Acta*, V. 35, p. 247-260.

Farley, W. J. (1949) *Geology of the Sherritt Gordon orebody*: Bull. Can. Inst. Min. Met., V. 42, p. 25-30.

- _____. (1948) Sherritt Gordon mine: Structural Geology of Canadian Ore Deposits: Can. Inst. Min. Metall., p. 292-295.
- Fox, J. S. (1976) Some comments on the volcanic stratigraphy and economic potential of the West Amisk Lake area, Saskatchewan: Circular 9, Geol. Div. Sask. Res. Council.
- Franklin, J. M. (1976) Role of lahatic breccia in genesis of volcanogenic massive sulphide deposits: Geol. Surv. Canada, Paper 76-1a, p. 293-300.
- Froese, E. (1976) Applications of thermodynamics in metamorphic petrology. Geol. Surv. Canada; Paper 75-43.
- _____. (1978) The graphical representation of mineral assemblages in biotite-bearing granulites: in Rept. Act., Geol. Surv. Canada, Paper 78-1A, p. 323-325.
- _____. (1979) A reaction grid for biotite-bearing granulites: in Rept. Act., Geol. Surv. Canada, Paper 79-1A, p. 83-85.
- Froese, E. and Gasparri, E. (1975) Metamorphic zones in the Snow Lake area, Manitoba: Can. Min., V. 13, p. 162-167.
- Froese, E. and Goetz, P. A. (1979) Geology, Sherridon area, Manitoba. Geol. Surv. Canada, Open File 623.
- Gale, G. H. and Koo, J. (1977) Evaluation of massive sulphide environments: in N.R.E.P., 2nd Ann. Rept., 1976-1977, p. 43-62.
- Garrels, R. M. and Christ, C. L. (1965) Solutions, minerals and equilibria: Harper and Row, New York.
- Garrels, R. M. and Naeser, C. R. (1958) Equilibrium distribution of dissolved sulphur species in water at 25°C and 1 atm total pressure: Geochim. Cosmochim. Acta, V. 15, p. 113-130.
- Goodwin, A. M. (1973) Archean iron formations and tectonic basins of the Canadian Shield: Econ. Geol., V. 68, p. 915-933.
- Guber, A. L. and Ohmoto, H. (1978) Deep sea environment of Kuroko formation as indicated by the benthic foraminifera from the Hokuroku district, Japan: Min. Geol., V. 28, p. 245-256.
- Harrison, J. M. (1951) Precambrian correlation and nomenclature, and problems of the Kisseynew gneisses, in Manitoba: Geol. Surv. Canada, Bull. 20.
- Helgeson, H. C. (1967) Solution chemistry and metamorphism: in Researches in geochemistry, V. 2, ed. P. H. Abelson, John Wiley and Sons, New York, p. 362-404.
- Helgeson, H. C. (1969) Thermodynamics of hydrothermal systems at elevated temperatures and pressures: Am. J. Sci., V. 267, p. 729-804.

- Henley, R. W. and Thoinley, P. (1979) Some geothermal aspects of polymetallic massive sulfide formation: *Econ. Geol.*, V. 74, p. 1600-1612.
- Hughes, C. J. (1973) Spilites, keratophytes and the igneous spectrum: *Geol. Mag.*, V. 6, p. 513-527.
- Irvine, T. N. and Baragar, W. R. A. (1971) A guide to the chemical classification of the common volcanic rocks: *Can. J. Earth Sci.*, V. 8, p. 523-548.
- Iwao, S. (1970) Clay and silica deposits of volcanogenic affinity: in Japan in volcanism and ore genesis, ed. T. Tatsumi, Univ. Tokyo Press, Tokyo, p. 267-283.
- Josse, G. and Bailes, A. (1974) Rb/Sr ages from the File-Morton-Woosey Lakes area of the Flin Flon volcanic belt: in U. of Man. Centre Precam. Studies, 1974 Ann. Rept., Sec. 2, p. 40-50.
- Kalliokoski, J. (1953) Interpretations of the structural geology of the Sherridon-Flin Flon region, Manitoba: *Geol. Surv. Canada, Bull.* 25.
- Kanehira, K. and Tatsumi, T. (1970) Bedded cupriferous sulphide deposits in Japan: a review: in Volcanism and ore genesis, ed. T. Tatsumi, Univ. Tokyo Press, Tokyo, p. 51-76.
- Karig, D. E. (1970) Ridges and basins of the Tonga-Kermadec island arc system: *Jour. Geophys. Res.*, V. 75, p. 239-254.
- _____. (1971) Origin and development of marginal basins in the western Pacific: *Jour. Geophys. Res.*, V. 76, p. 2542-2561.
- Karig, D. E. (1972) Remnant arcs: *Bull. Geol. Soc. Am.*, V. 83, p. 1057-1068.
- Kelley, K. K. (1960) Contributions to the data on theoretical metallurgy, XIII. High temperature heat content, heat capacity and entropy data for the elements and inorganic compounds: *U.S. Bur. Mines Bull.* 584.
- Kerrick, D. M. (1972) Experimental determination of muscovite + quartz stability with $P_{H_2O} < P_{Total}$: *Am. J. Sci.*, V. 272, p. 946-958.
- Kornik, L. J. (1968) Geology of the Guay Lake area (west half): *Man. Mines Br., Pub.* 64-6.

- Kuno, H. (1968) Differentiation of basaltic magmas: in The Roldervaart treatise on rocks of basaltic composition. Wiley Interscience, New York, p. 623-688.
- Lambert, I. B. and Sato, T. (1974) The kuroko and associated ore deposits of Japan: A review of their features and metallogenesis: *Econ. Geol.*, V. 69, p. 1215-1236.
- Langelaar, J. and Pauliw, P. (1975) Ruttan mine--a geophysical discovery: Paper presented to March 1975 Meeting of the Prospectors and Developers Assoc.
- Large, R. R. (1977) Chemical evolution and zonation in volcanic terrains: *Econ. Geol.*, V. 72, p. 549-572.
- Leake, B. E. (1964) The chemical distinction between ortho- and para-amphibolites: *J. Pet.*, V. 5, p. 238-254.
- Leech, G. B., Lowden, J. A. and Stockwell, C. H. (1963) Age determinations and geological studies: *Geol. Surv. Canada*, Paper 63-17.
- Lewry, J. F., Sibbald, T. I. I. and Rees, C. J. (1978) Metamorphic patterns and their relation to tectonism and plutonism in the Churchill Province in northern Saskatchewan: in *Metamorphism in the Canadian Shield*, eds. J. A. Fraser and W. W. Heywood, *Geol. Surv. Canada*, Paper 78-10, p. 139-154.
- Lowden, J. A. (1961) Age determinations by the Geological Survey of Canada: Rept. 2, Isotopic Ages: *Geol. Surv. Canada*, Paper 61-17.
- Lowden, J. A., Stockwell, C. H. and Wanless, R. K. (1962) Age determinations and geologic studies: Including isotopic ages, Rept. 4: *Geol. Surv. Canada*, Paper 62-17.
- MacQuarrie, R. (1977) Uranium-lead dating of igneous events in the Flin Flon region, Manitoba and Saskatchewan. PhD. thesis, U. of Sask. at Saskatoon.
- McRitchie, W. D. (1974) The Sickle-Wasekwan debate: a review: *Man. Dept. Mines Res. Env. Mang.*, *Geol.* Paper 1/74.
- Minato, M., Gorai, M. and Hunuhashi, M., eds. (1965) The geological development of the Japanese islands: Tsukiji Shokan, Tokyo.
- Mitchell, A. H. G. (1970) Facies of an early Miocene volcanic arc, Malekula Island, New Hebrides: *Sedimentology*, V. 14, p. 201-243.
- Mitchell, A. H. G. and Bell, J. D. (1973) Island-arc evolution and related mineral deposits: *J. Geol.*, V. 81, p. 381-405.

- Miyashiro, A. and Shido, F. (1975) Tholeiitic and calc-alkalic series in relation to the behavior of titanium, vanadium, chromium and nickel: *Am. J. Sci.*, V. 275, p. 265-277.
- Miyashiro, A. (1974) Volcanic series in island arcs and active continental margins: *Am. J. Sci.*, V. 274, p. 321-355.
- Moore, J. M., Hart, S. R., Barnett, C. C. and Hurley, P. M. (1960) Potassium-argon ages in northern Manitoba: *Bull. Geol. Soc. Am.*, V. 71, p. 225-229.
- Moore, J. M. (1977) Orogenic volcanism in the Proterozoic of Canada: in Volcanic regimes in Canada, ed. W. R. A. Baragar et al, G.A.C. Spec. Paper No. 16, p. 127-148.
- Moore, W. J. and Nash, J. T. (1974) Alteration and fluid inclusion studies of the porphyry copper ore body at Bingham, Utah: *Econ. Geol.*, V. 69, p. 631-645.
- Muffler, J. P., White, D. E. and Truesdell, A. H. (1971) Hydrothermal explosion craters in Yellowstone National Park: *Geol. Soc. Amer. Bull.*, V. 82, p. 723-740.
- Mukherjee, A. C. (1974) Some aspects of sedimentology of the Missi Group and its environment of deposition: *Can. J. Earth Sci.*, V. 11, p. 1018-1019.
- Mukherjee, A. C., Stauffer, M. R. and Baadsgaard, H. (1971) The Hudsonian orogeny near Flin Flon Manitoba: A tentative interpretation of Rb-Sr and K-Ar ages: *Can. J. of Earth Sci.*, V. 8, p. 939-946.
- Niino, H. and Emery, K. O. (1961) Sediments of shallow portions of East China Sea and South China Sea: *Bull. Geol. Soc. Am.*, V. 72, p. 731-762.
- Nriagu, J. O. (1971) Expressions for calculating the solubilities of metal sulfides in hydrothermal solutions: *Can. J. Earth Sci.*, V. 8, p. 813-819.
- Osborn, E. F. (1962) Reaction series for subalkaline igneous rocks based on different oxygen pressure conditions: *Am. Min.*, V. 47, p. 211-226.
- Park, J. K. (1975) Paleomagnetism of the Flin Flon-Snow Lake greenstone belt, Manitoba and Saskatchewan: *Can. J. Earth Sci.*, V. 12, p. 1272-1290.
- Pearson, D. E. (1972) The location and structure of the Precambrian Kiseeynew gneiss domain of northern Saskatchewan: *Can. J. Earth Sci.*, V. 9, p. 1235-1249.

Peterson, M. N. A. and von der Borch, C. C. (1965) Chert: Modern inorganic deposition in a carbonate-precipitating locality: Science, V. 149, p. 1501-1503.

Pettijohn, F. J., Potter, P. E. and Siever, R. (1972) Sand and Sandstone, Springer-Verlag, New York.

Plumely, W. J., Risley, G. A., Graves, R. W. and Kaley, M. E. (1962) Energy index for limestone interpretation and classification: in Classification of Carbonate Rocks, Am. Assoc. Pet. Geol., Mem. 1, p. 85-107.

Pollock, G. D. (1964) Geology of the Duval Lake area: Man. Dept. Mines Nat. Res., Pub. 61-6.

_____. (1965) Geology of the Russick Lake area: Man. Dept. Mines Nat. Res., Pub. 63-2.

Price, D. P. (1978) Geology and economic potential of the Flin Flon-Snow Lake area: in U. of Man., Centre Precam. Studies, 1977 Ann. Rept., p. 52-83.

Ramsay, J. G. (1967) Folding and Fracturing of Rocks: McGraw-Hill, Toronto.

Ridge, J. D. (1974) Note on boiling of ascending ore fluids and the position of volcanic-exhalative deposits in the modified Lindgren classification: Geology, V. 5, p. 287-288.

Robertson, D. S. (1953) Batty Lake map area, Manitoba: Geol. Surv. Canada, Mem. 271.

Roedder, E. (1972) Composition of fluid inclusions: Data of Geochemistry: U.S. Geol. Surv. Prof. Paper 440-J.

Rusnak, G. A. (1960) Sediments of Laguna Madre, Texas: in Recent sediments, northwestern Gulf of Mexico, eds. F. P. Shepard, F. B. Phleger and T. H. van Andel., Am. Assoc. Pet. Geol., p. 117-152.

Sangster, D. F. (1972) Precambrian volcanogenic massive sulphide deposits in Canada: A review: Geol. Surv. Canada, Paper 72-22.

_____. (1976) Sulphur and lead isotopes in strata-bound deposits: in Handbook of strata-bound and stratiform ore deposits, V. 2, Geochemical Studies, ed. K. H. Wolf, Elsevier, New York, p. 219-266.

_____. (1978) Isotopic studies of ore-leads of the circum-Kisseynew belt of Manitoba and Saskatchewan: Can. J. Earth Sci., V. 15, p. 1112-1121.

- Sangster, D. F. and Scott, S. D. (1976) Precambrian, strata-bound, massive Cu-Zn-Pb sulfide ores of North America: in Handbook of strata-bound and stratiform ore deposits, V. 6, Cu, Zn, Pb and Ag deposits, ed. K. H. Wolf, Elezévir, New York, p. 129-222.
- Sato, T. (1973) A chloride complex model for Kuroko mineralization: *Geochem. J.*, V. 7, p. 245-270.
- Scott, S. D. (1973) Experimental calibration of the sphalerite geobarometer: *Econ. Geol.*, V. 68, p. 466-474.
- Shanks, R. J. and Bailes, A. H. (1977) "Missi Group" rocks, Wekusko Lake area. *Man. Dept. Mines Res. and Env. Man., Report Act.*, 1977, p. 83-87.
- Shaw, D. M. (1972) The origin of the Apsley gneiss, Ontario: *Can. J. Earth Sci.*, V. 9, p. 18-35.
- Shirozu, H. (1974) Clay minerals in altered wall rocks of the Kuroko-type deposits: *Japan Min. Geol., Spec. Iss.*, No. 6, p. 303-310.
- Sibbald, T. I., Munday, R. J. G. and Lewry, J. F. (1976) The geological setting of uranium mineralization in northern Saskatchewan: in *Uranium in Saskatchewan*, ed. C. E. Dunn, *Sask. Geol. Soc., Spec. Pub.* 3, p. 51-98.
- Skippen, G. B. and Carmichael, D. M. (1977) Mixed volatile equilibria: in *Short course in application of thermodynamics to petrology and ore deposits*, ed. H. J. Greenwood, *Min. Assoc. Canada, Short Course Handbook*, V. 2, p. 109-125.
- Slaughter, M. and Earley, W. (1965) Mineralogy and geological significance of the Mowry bentonites, Wyoming: *Geol. Soc. Am. Spec. Paper* 83.
- Snarely, P. D., MacLeod, N. S. and Wagner, H. G. (1968) Tholeiitic and alkalic basalts of the Eocene Siletz River volcanics, Oregon Coast Range: *Am. J. Sci.*, V. 266, p. 454-481.
- Stacey, J. S. and Kramers, J. D. (1975) Approximation of terrestrial lead isotope evolution by a two-stage mode: *Earth Planet. Sci. Lett.*, V. 26, p. 207-221.
- Stanton, R. and Russell, R. D. (1959) Anomalous leads and the emplacement of lead sulphide ores: *Econ. Geol.*, V. 54, p. 588-608.
- Stauffer, M. R. (1974) Geology of the Flin Flon area: A new look at the sunless city: *Geosci. Canada*, V. 1, p. 30-35.
- Stauffer, M. R., Mukherjee, A. C. and Koo, J. (1979) The Amisk Group: An Aphebian island arc deposit: *Can. J. Earth Sci.*, V. 12, p. 2021-2035.

Stockwell, C. H. (1950) The use of plunge in the construction of folds: Proc. Geol. Assoc. Canada., V. 3, p. 97-121.

_____. (1960) Flin Flon-Mandy area, Manitoba and Saskatchewan: Geol. Surv. Canada, Map 1078A, with descriptive notes.

Sugisaki, R., Mizutani, S., Hattori, H., Adachi, M. and Tanaka, T. (1972) Late Paleozoic geosynclinal basalt and tectonism in the Japanese Islands. Tectonophysics, V. 14, p. 35-56.

Tuckwell, K. (1979) Stratigraphy and mineral deposits of the Sherridon Area. Man. Min. Res. Div., Rept. Act., p. 42-45.

Turner, J. S. and Gustafson, L. B. (1978) The flow of hot saline solutions from vents in the sea floor--some implications for exhalative massive sulfide and other ore deposits: Econ. Geol., V. 73, p. 1082-1100.

Tyrrell, J. B. (1902) Report on explorations in the northeastern portion of the District of Saskatchewan and adjacent parts of the District of Keewatin: Geol. Surv. Canada, Ann. Rept. 1909, V. XI, pt. F.

Urabe, T. and Sato, T. (1978) Kuroko deposits of the Kosaka mine, northeast Honshu, Japan--products of submarine hot springs on Miocene sea floor: Econ. Geol., V. 73, p. 161-179.

van de Kamp, P. C. (1968) Geochemistry and origin of meta-sediments in the Haliburton-Madoc area, southeastern Ontario: Can. J. Earth Sci., V. 5, p. 1337-1370.

van de Kamp, P. C., Leake, B. E. and Senior, R. (1975) The petrography and geochemistry of some Californian arkoses with application to identifying gneisses of sedimentary origin: J. Geol., V. 84, p. 195-212.

Wanless, R. K., Stevens, R. D., Lachance, G. R. and Rimsaite, R. Y. H. (1964) Age determination and geologic studies, Part 1--Isotopic ages, Rept. 5: Geol. Surv. Canada, Paper 64-17.

Weber, W. and Scoates, R. F. J. (1978) Archean and Proterozoic metamorphism in the northwestern Superior province and along the Churchill-Superior boundary, Manitoba: in Metamorphism in the Canadian Shield, ed. J. A. Fraser and W. W. Heywood, Geol. Surv. Canada, Paper 78-10, p. 5-16.

White, D. E., Muffler, L. J. P. and Truesdell, A. H. (1971) Vapour dominated hydrothermal systems compared with hot-water systems: Econ. Geol., V. 66, p. 75-97.

Whitmore, D. R. E. (1968) Geology of the Coronation copper deposit: in Symposium on the geology of Coronation mine, Saskatchewan, ed. A. R. Byers, G.S.C., Paper 68-S, p. 37-54.

Wilkinson, S. J. (1976) The petrography of the anthophyllite rocks at Sherridon, Manitoba: Unpub. BSc. thesis, Univ. Western Ont.

Wright, J. F. (1929) Kississing Lake area, Manitoba: Geol. Surv. Canada, Sum. Rept. 1928, pt. B, p. 73-104.

_____. (1930) Crystalline limestone in the Kiskeynew of northern Manitoba: Can. Min. Jour., V. 51, p. 762.

_____. (1931) Geology and mineral deposits of a part of north-west Manitoba: Geol. Surv. Canada, Sum. Rept. 1930, pt. C, p. 1-125.

York, D. and Farquhar, R. M. (1972) The earth's age and geochronology: Pergamon Press, Toronto.

Zwanzig, H. V. (1976) Laurie Lake area: in Man. Mineral Res. Div., Geol. Surv. Canada, Rept. Act. 1976, p. 26-32.

APPENDIX A

CHEMICAL ANALYSES OF ROCKS FROM SHERRIDON, MANITOBA

The results of 195 chemical analyses of rocks from the Sherridon area are compiled in this appendix. Sample locations are given on Fig. A-1 in the pocket.

Hand specimens of about 2 kg mass were collected for a chemical analyses and thin-sectioning. All weathered surfaces were removed, the sample was sawn in half; one half was saved for making a thin section and the other half was cleaned. All saw marks were removed with a hammer. The sample was then crushed and submitted to the Geological Survey of Canada Analytical Chemistry Section for analysis. CO_2 , H_2O and FeO were determined by chemical methods and the remaining oxides and elements were determined by X-ray fluorescence on the 12B DR system.

The estimated reliability of the major oxides of these analyses based on control standards and reported by the Geological Survey of Canada are given in Table A-1. The estimated reliability of the minor elements is reported to be about 10 percent of their concentration in the rocks?

Replicate analyses of a calc-silicate (1, 1a), an amphibolite (94, 94a), and a biotite gneiss (158, 158a) are also reported in Table A-1. These analyses are given for the purpose of estimating precision. Also the Geological Survey of Canada standard LU-1 was submitted in order to gain an estimate of the accuracy of these analyses. The precision and accuracy were found to be about ± 5 percent of the reported concentration except for the alkalis. The

Fe/Fe₂O₃ ratio does not show good precision although total Fe as Fe₂O₃ is much better. Precision and an accuracy of K₂O are about ± 10 percent of the reported value and for Na₂O about ± 20 percent of the reported value.

Table A-1

Oxides	*Reliability of Analyses as Reported by the Geological Survey							Reported Standard	Analysis of Standard
	1	1a	94	94a	158	158a			
SiO ₂	1	31.1	30.2	53.6	52.6	71.9	71.2	49.3	50.7
TiO ₂	1	0.51	0.49	0.7	0.69	0.32	0.30	.75	0.77
Al ₂ O ₃	1	5.5	9.5	15.7	16.1	12.8	13.2	19.6	15.4
Fe ₂ O ₃	1	0.9	1.1	4.3	3.4	1.0	0.4	1.3	1.2
FeO		3.7	3.7	9.3	9.8	3.9	4.3	6.2	6.6
MnO	2	0.12	0.12	0.22	0.22	0.09	0.10	0.19	0.20
MgO	1	5.99	6.18	5.24	5.69	1.31	1.36	4.29	4.20
CaO	1	33.2	33.1	9.14	8.84	3.14	3.06	10.31	10.39
Na ₂ O	2	0.0	0.0	1.1	1.7	2.0	2.5	1.9	1.6
K ₂ O	2	0.44	0.59	0.25	0.34	2.02	2.09	0.45	0.49
P ₂ O ₅	1	0.12	0.13	0.11	0.15	0.11	0.13	0.08	0.07
CO ₂		18.4	18.1	0.0	0.0	0.5	0.5	6.2	6.4
H ₂ O		0.5	0.5	1.2	0.9	1.2	1.0	3.2	2.5
TOTAL									

* ± percent of concentration

Rock Type BI GNEISS: UNIT 5

Specimen No.	4	5	7	16	19	23	33
Major oxides							
SiO ₂	73.7	75.2	69.8	76.7	72.1	76.8	74.7
TiO ₂	0.18	0.27	0.28	0.18	0.29	0.18	0.18
Al ₂ O ₃	11.5	13.3	14.0	11.7	13.4	10.8	10.1
Fe ₂ O ₃	-	-	0.3	-	0.3	1.1	0.1
FeO	0.8	2.5	5.5	2.7	2.7	3.4	3.0
MnO	0.03	0.02	0.01	0.06	0.07	0.11	0.10
MgO	3.39	2.28	2.51	0.91	1.41	1.27	1.86
CaO	5.45	1.92	3.18	2.81	3.20	3.47	2.59
Na ₂ O	1.8	1.1	1.5	2.4	2.1	0.2	0.9
K ₂ O	0.64	0.80	1.71	1.16	3.28	1.25	2.52
P ₂ O ₅	0.15	0.10	0.08	0.06	0.09	0.10	0.10
CO ₂	1.0	0.3	0.4	0.7	0.7	0.6	0.8
H ₂ O	0.9	1.0	1.1	1.4	1.0	1.2	1.8
Total	99.54	100.46	98.79	100.78	100.64	100.48	98.72

Minor Elements

Ba	.011	.020	.014	.022	.048	.011	.040
Co	-	.0012	.0015	-	.0011	-	.0011
Cr	-	-	-	-	-	-	-
Cu	.0012	.0014	.0019	.0045	.0052	.0091	.0012
Ni	-	.0015	.0010	.0013	-	.0013	.0014
Rb	-	-	.003	.001	.003	-	.001
S	.07	.05	.09	.13	.07	.10	.05
Sr	.0170	.0260	.0140	.0110	.0080	.0055	.0072
V	-	-	.0038	-	-	-	-
Y	-	.0035	.0027	.0032	.0023	.0039	.0048
Zn	-	.011	.010	.013	.006	.006	.013
Zr	.0049	.0190	-	.0110	.0091	.0068	.0130

- not detected

Rock Type

Specimen No.	40	43	44	57	61	62	65
<u>Major oxides</u>							
SiO ₂	73.2	77.5	75.0	76.1	74.3	75.1	75.3
TiO ₂	0.20	0.26	0.24	0.15	0.26	0.13	0.20
Al ₂ O ₃	13.2	11.1	11.0	11.9	12.2	11.8	10.6
Fe ₂ O ₃	0.4	0.5	-	-	0.2	-	1.64
FeO	2.8	2.6	5.2	2.9	2.8	3.0	-
MnO	0.16	0.06	0.09	0.06	0.06	0.09	0.08
MgO	1.51	0.65	1.44	0.74	1.27	1.50	4.00
CaO	3.03	1.98	1.82	2.22	2.83	2.11	2.85
Na ₂ O	1.8	2.4	0.8	2.8	2.9	2.7	1.3
K ₂ O	2.21	2.23	3.45	1.87	2.02	1.45	1.8
P ₂ O ₅	0.06	0.09	0.08	0.04	0.10	0.05	0.09
CO ₂	0.7	0.3	0.4	0.4	0.6	0.2	0.6
H ₂ O	1.2	8.6	1.1	0.6	0.8	0.7	1.0
Total	100.47	100.27	99.82	99.78	100.34	98.83	99.46

Minor Elements

Ba	.035	.052	.043	.049	.043	.026	.036
Co	.0012	.0023	.0019	-	.0031	.0011	.0016
Cr	-	.0007	.0007	-	.0009	-	.0007
Cu	.0011	.0071	.0017	.0029	.0110	.0039	.0071
Ni	.0015	.0018	.0014	-	.0017	-	.0013
Rb	-	.009	.002	.001	.003	-	.002
S	.04	.04	.05	.14	.16	.06	.10
Sr	.0082	.0080	.0100	.0170	.0100	.0083	.0110
V	-	.0020	-	-	-	-	-
Y	.0043	.0055	.0042	.0036	.0049	.0034	.0038
Zn	.015	.017	.020	.006	.020	-	.013
Zr	.0089	.0170	.0150	.0130	.0150	.0100	.0130

Rock Type

Specimen No.	83	96	97	99	108	112	122
Major oxides							
SiO ₂	74.6	74.8	78.2	72.5	78.0	71.2	77.3
TiO ₂	0.14	0.24	0.19	0.28	0.17	0.37	0.17
Al ₂ O ₃	12.3	11.1	10.0	11.9	10.5	11.7	11.5
Fe ₂ O ₃	0.4	1.0	0.6	0.6	0.2	1.2	0.6
FeO	2.5	3.5	2.1	3.7	2.2	4.2	1.8
MnO	0.06	0.08	0.06	0.10	0.06	0.12	0.07
MgO	2.70	1.74	2.15	1.87	2.57	1.61	1.29
CaO	2.63	2.08	1.80	1.45	1.28	2.08	3.06
Na ₂ O	0.6	0.8	0.1	-	1.7	4.1	1.7
K ₂ O	2.51	2.46	3.86	7.21	2.27	0.60	1.57
P ₂ O ₅	0.04	0.07	0.08	0.09	0.07	0.14	0.05
CO ₂	0.5	0.7	0.4	0.3	0.3	0.3	0.7
H ₂ O	2.3	1.3	1.0	1.2	1.1	0.4	1.2
Total	101.28	100.67	100.54	101.20	100.42	98.32	101.61

Minor Elements

Ba	.031	.071	.040	.034	.027	.038	.024
Co	-	.0051	.0015	-	-	.0031	.0012
Cr	-	.0032	-	-	-	.0031	-
Cu	.0021	.0068	.0060	.0019	.0024	.0040	.0140
Ni	.0011	.0029	.0013	.0013	.0012	.0027	.0015
Rb	.003	.002	.003	.006	.001	-	.001
S	.38	.06	.10	.05	.07	.07	.05
Sr	.0150	.0090	.0061	.0062	.0064	.0180	.0083
V	-	.013	-	-	-	.004	-
Y	.0025	.0044	.0036	.0031	.0026	.0053	.0044
Zn	.014	.015	-	.015	.023	.017	.014
Zr	.0130	.0066	.0100	.0140	.0120	.0096	.0110

Rock Type

Specimen No.	132	134	140	144	148	152	156
Major oxides							
SiO ₂	72.2	74.0	76.2	76.3	70.1	75.3	71.9
TiO ₂	0.32	0.22	0.08	0.26	0.26	0.12	0.12
Al ₂ O ₃	12.9	13.3	11.5	10.7	13.9	9.1	13.5
Fe ₂ O ₃	-	0.2	-	-	0.8	0.8	1.6
FeO	4.4	2.4	1.6	4.2	3.2	4.6	1.8
MnO	0.07	0.06	0.08	0.16	0.12	0.11	0.04
MgO	1.77	0.71	1.42	2.53	1.35	2.79	1.62
CaO	2.17	3.39	1.58	1.12	3.28	2.33	2.60
Na ₂ O	1.6	3.4	1.0	1.4	1.6	0.1	2.60
K ₂ O	3.73	3.53	4.07	2.51	4.18	1.37	1.58
P ₂ O ₅	0.13	0.08	0.07	0.04	0.10	0.05	0.06
CO ₂	0.2	-	0.4	0.1	1.0	1.0	-
H ₂ O	1.1	0.4	1.0	1.0	0.8	2.5	0.8
Total	100.59	100.69	99.00	100.32	100.69	100.14	98.22

Minor Elements

Ba	.045	.048	.120	.054	.090	.022	.016
Co	.0019	.0012	-	.0022	.0020	.0019	-
Cr	-	.0011	-	.0006	.0007	.0007	-
Cu	.0020	.0056	.0061	.0057	.0014	.0013	.0900
Ni	.0015	.0010	-	.0018	.0016	.0014	-
Rb	.004	-	.002	-	.005	.002	-
S	.70	.05	.15	.20	.06	.24	.72
Sr	.0093	.0097	.0087	.0062	.0080	.0065	.0170
V	.0032	.0024	-	-	-	-	.0027
Y	.0035	.0034	.0028	.0061	.0032	.0047	-
Zn	.012	-	-	.022	.021	.018	-
Zr	.0095	.0084	.0100	.0097	.0095	.0055	.0120

Rock Type	SL-BI GNEISS UNIT 5a						
Specimen No.	157	158	159	160	161	34	81
<u>Major oxides</u>							
SiO ₂	74.4	71.2	74.2	74.2	74.0	73.0	75.8
TiO ₂	0.11	0.30	0.15	0.25	0.24	0.24	0.13
Al ₂ O ₃	9.4	13.2	11.8	11.8	11.8	13.9	12.3
Fe ₂ O ₃	-	0.4	0.7	0.1	1.3	0.3	-
FeO	4.6	4.3	3.0	3.6	3.1	1.9	2.0
MnO	0.06	0.10	0.16	0.21	0.16	0.04	0.04
MgO	5.06	1.36	2.64	2.43	1.15	2.21	2.54
CaO	0.04	2.50	0.80	0.06	0.07	2.82	2.68
Na ₂ O	0.30	3.06	2.52	2.93	3.27	1.50	1.80
K ₂ O	1.17	2.09	2.57	2.44	1.61	1.10	1.23
P ₂ O ₅	0.04	0.13	0.06	0.07	0.06	0.10	0.05
CO ₂	-	0.5	0.8	0.9	1.5	0.8	0.4
H ₂ O	1.5	1.0	1.5	1.4	1.2	2.4	1.6
Total	96.69	99.94	100.50	100.39	99.46	100.31	100.57
<u>Minor Elements</u>							
Ba	.013	.028	.036	.027	.041	.018	.031
Co	.0022	.0018	.0013	.0011	.0011	-	-
Cr	.0006	.0007	-	-	.007	-	-
Cu	.0100	.0052	.0022	.0014	.0030	-	.0021
Ni	-	.0017	.0030	-	.0014	-	.0021
Rb	-	.001	.001	.001	.001	.001	.001
S	.68	1.3	.07	.08	.06	.08	.06
Sr	.0034	.0099	.0067	.0077	.0092	.0230	.0150
V	-	.0027	-	-	-	-	-
Y	.0039	.0041	.0045	.0077	.0042	.0037	.0025
Zn	1.30	-	-	.004	.013	.012	.017
Zr	.0160	.0094	.0084	.0080	.0120	.0200	.0130

Rock Type	SL-BI GNEISS UNIT 5b			QUARTZO- FELDSPATHIC GNEISS UNIT 6			
Specimen No.	126	123	131	15	25	77	110

Major oxides

SiO ₂	75.6	83.2	77.8	82.1	81.3	81.5	76.6
TiO ₂	0.16	0.16	0.26	0.16	0.16	0.10	0.14
Al ₂ O ₃	10.5	8.4	10.1	8.6	11.2	9.9	12.8
Fe ₂ O ₃	-	0.6	-	-	-	0.1	0.2
FeO	3.6	1.6	6.1	1.4	0.3	0.4	0.5
MnO	0.04	0.03	0.05	0.03	-	0.01	0.13
MgO	3.28	1.76	0.66	1.17	0.43	0.48	1.41
CaO	0.79	0.80	0.19	1.61	1.34	1.81	3.98
Na ₂ O	0.8	0.2	-	1.2	2.8	1.4	0.1
K ₂ O	3.00	2.30	1.14	2.18	2.49	4.51	2.47
P ₂ O ₅	0.07	0.04	0.04	0.07	0.07	0.04	0.03
CO ₂	-	0.1	-	0.5	-	0.4	0.3
H ₂ O	1.1	0.7	0.7	1.0	0.3	0.4	2.0
Total	98.94	99.89	97.04	100.02	100.39	100.05	100.66

Minor Elements

Ba	.030	.039	.041	.026	.040	.092	.046
Co	-	-	.0047	-	-	-	-
Cr	-	-	.011	-	-	-	-
Cu	.0023	.0057	.0017	.0027	.0025	.0052	.0072
Ni	.012	.012	.017	-	-	-	-
Rb	.003	.002	-	.003	.001	.002	.001
S	.04	.05	.210	.07	.05	.07	.08
Sr	.0065	.0077	.0054	.0062	.0110	.0120	.0110
V	-	-	-	-	-	.003	-
Y	.0035	.0027	.0045	.0070	-	-	.0025
Zn	.017	-	.037	.009	-	.006	.012
Zr	.0062	.0032	.0140	.0061	.0080	.0130	.0110

Rock Type				QUARTZITIC GNEISS UNIT 7		MAFIC BI GNEISS UNIT 5c	
Specimen No.				68	74	59	78
Major oxides							
SiO ₂	82.1	80.1	79.1	80.1	80.4	66.8	68.5
TiO ₂	0.08	0.22	0.25	0.31	-	0.39	0.35
Al ₂ O ₃	9.9	10.7	12.0	6.1	0.4	12.8	14.8
Fe ₂ O ₃	0.4	0.2	-	3.9	5.6	1.3	0.2
FeO	1.0	1.5	0.6	-	0.1	5.7	3.2
MnO	0.02	0.07	0.01	0.01	0.13	0.29	0.05
MgO	0.27	0.71	0.40	1.16	2.40	2.84	2.37
CaO	2.25	2.33	1.66	2.72	4.11	3.53	2.35
Na ₂ O	2.2	2.3	3.7	1.8	-	1.9	0.4
K ₂ O	1.29	1.23	1.16	0.62	0.12	1.83	6.14
P ₂ O ₅	0.03	0.07	0.08	0.03	0.32	0.08	0.12
CO ₂	0.4	0.2	0.4	1.2	0.4	0.5	0.2
H ₂ O	0.3	0.8	0.4	0.4	0.3	1.3	1.5
Total	99.31	100.63	99.76	99.96	98.28	99.26	100.18
Minor Elements							
Ba	.020	.015	.015	.009	-	.110	.048
Co	-	.0018	-	.0008	.0023	.0036	.0016
Cr	.0060	-	-	.0019	.0012	.0012	.0005
Cu	.0018	.0014	.0023	.0016	.0036	.0024	.0016
Ni	.0013	.0013	-	.0049	.0017	.0016	.0011
Rb	-	-	-	-	-	.001	.005
S	-	.07	.01	1.1	.85	.09	.08
Sr	.0600	.0110	.0061	.0120	.0041	.0070	.0140
V	-	-	-	.0100	-	.0055	.0039
Y	.0021	.0043	-	-	.0037	.0046	.0029
Zn	.016	.013	-	.020	.026	.019	.014
Zr	.0140	.0120	.0110	.0076	-	.0130	.0100

Rock Type

Specimen No.	80	118	120	121	127	153	165
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Major oxides

SiO ₂	67.6	69.7	68.6	69.7	66.5	67.8	63.5
TiO ₂	0.35	0.33	0.42	0.28	0.40	0.37	0.54
Al ₂ O ₃	13.4	13.2	13.7	12.8	16.9	12.9	13.8
Fe ₂ O ₃	2.4	1.2	0.1	1.2	1.6	1.3	1.5
FeO	3.8	4.3	6.0	5.1	3.2	5.6	7.5
MnO	0.09	0.09	0.14	0.06	0.08	0.20	0.15
MgO	1.68	1.51	2.47	1.83	1.21	1.94	2.88
CaO	1.60	4.20	2.68	2.13	3.46	4.75	4.43
Na ₂ O	1.2	1.3	2.7	0.9	4.7	0.7	1.4
K ₂ O	5.25	1.28	2.34	4.97	1.38	3.08	2.01
P ₂ O ₅	0.15	0.13	0.17	0.09	0.15	0.16	0.10
CO ₂	-	0.4	0.4	0.9	-	0.8	0.8
H ₂ O	1.2	1.0	1.1	1.2	0.4	0.9	1.4
Total	98.72	98.64	100.82	101.16	99.98	100.50	100.01

Minor Elements

Ba	.089	.031	.023	.035	.160	.064	.012
Co	.0025	.0024	.0006	.0130	.0170	.0028	.0032
Cr	.0012	.0009	.0016	-	.0005	.0007	.0014
Cu	.0013	.0063	.0180	.0160	.0010	.0085	.0160
Ni	.0019	.0017	.0032	.0012	.0014	.0018	.0020
Rb	.006	.001	.005	.005	.003	.006	.003
S	.06	.07	.11	.07	.04	.16	.23
Sr	.0011	.0170	.0260	.0063	.0220	.0110	.0150
V	.0040	-	.0320	.0024	.0036	-	.0140
Y	.0046	.0042	.0035	.0021	.0041	.0043	.0040
Zn	.023	.016	.011	-	.016	.024	.019
Zr	.0150	.0090	.0059	-	.0040	.0058	.0037

Rock Type		BI-GA SCHIST UNIT 8			PELITE UNIT 9		
Specimen No.	179	173	174	175	102	177	178
Major oxides							
SiO ₂	67.7	61.8	62.8	62.8	76.2	78.0	69.8
TiO ₂	0.27	0.79	0.50	0.60	0.20	0.13	0.18
Al ₂ O ₃	15.7	16.7	17.3	16.0	11.6	10.2	12.1
Fe ₂ O ₃	0.6	0.7	0.2	1.0	0.8	1.4	4.1
FeO	3.8	5.9	5.7	5.6	3.2	4.4	6.2
MnO	0.08	0.09	0.08	0.07	0.23	0.07	0.09
MgO	1.94	2.99	2.71	2.90	3.39	2.44	2.90
CaO	3.05	4.21	3.27	3.60	0.65	0.39	0.42
Na ₂ O	3.4	1.6	3.4	2.4	2.2	0.2	0.2
K ₂ O	2.38	2.05	2.30	2.38	1.69	1.28	1.44
P ₂ O ₅	0.17	0.40	0.45	0.69	0.06	0.06	0.15
CO ₂	0.4	0.3	0.4	-	-	0.1	0.1
H ₂ O	1.0	1.4	1.2	0.8	0.6	0.9	0.9
Total	100.44	99.63	100.31	98.84	100.82	99.57	99.58

Minor Elements

Ba	.048	.071	.041	.067	.028	.031	.054
Co	.0138	.0021	-	.0014	.0011	.0011	.0014
Cr	.0089	.0148	-	.0148	-	-	-
Cu	.0013	.0014	.0027	.0033	.0070	.0276	.1890
Ni	.0029	.0034	-	.0034	.0011	-	-
Rb	.008	.004	.004	.004	-	-	-
S	.10	-	.07	.03	.08	.09	.07
Sr	.0268	-	.0061	-	.0033	.0021	.0010
V	.0136	.0170	-	-	-	-	-
Y	.0044	-	.0045	-	.0030	.0051	.0059
Zn	-	-	-	-	-	-	.040
Zr	.0279	.0271	.0175	.0250	.0067	.0158	.0277

Rock Type	AT-GA GNEISS UNIT 10				CU GNEISS UNIT 10a		
Specimen No.	115	130	141	154	6	29	89
<u>Major oxides</u>							
SiO ₂	43.6	63.4	51.6	37.6	49.8	49.3	55.6
TiO ₂	0.49	0.24	0.77	0.34	2.01	1.64	0.64
Al ₂ O ₃	16.2	12.6	17.2	20.9	16.1	18.1	16.6
Fe ₂ O ₃	1.9	0.6	0.9	2.0	1.7	2.4	3.4
FeO	12.9	10.4	10.1	6.1	11.3	8.2	8.9
MnO	0.17	0.10	0.12	0.08	0.22	0.16	0.17
MgO	17.50	8.36	12.70	22.10	5.06	7.56	6.41
CaO	0.47	0.49	2.38	0.12	6.96	6.19	2.72
Na ₂ O	0.1	0.2	0.8	-	3.5	2.2	4.2
K ₂ O	2.97	0.86	0.52	0.88	0.34	0.28	0.64
P ₂ O ₅	0.07	0.04	0.10	0.04	0.25	0.42	0.21
CO ₂	-	0.2	0.8	-	1.3	1.4	-
H ₂ O	4.2	1.4	2.7	8.2	1.2	1.2	0.7
Total	100.57	99.89	100.09	98.66	99.74	99.95	100.19
<u>Minor Elements</u>							
Ba	.014	.008	.011	.004	.002	.003	.009
Co	.0038	.0038	.0060	-	.0061	.0065	.0054
Cr	.0020	.0014	.0020	-	.0017	.0120	.0028
Cu	.0110	.2000	.0025	.2300	.0018	.0130	.0027
Ni	.0029	.0026	.0036	.0014	.0058	.0083	.0032
Rb	-	-	-	-	-	-	-
S	.11	1.20	.06	.84	.36	.11	.06
Sr	.0026	.0050	.0054	.0030	.0026	.0110	.0064
V	.0030	-	.0033	-	.0250	.0230	.0017
Y	.0039	.0052	.0049	.0052	.0044	.0060	.0047
Zn	.042	.013	.016	.051	-	.019	.017
Zr	.0130	.0220	.0067	.0200	.0120	.0200	.0086

Rock Type							BI GNEISS UNIT 10b
Specimen No.	90	98	101	142	155	166	22
Major oxides							
SiO ₂	54.2	53.9	53.5	55.5	43.8	54.7	55.8
TiO ₂	0.72	0.58	0.66	0.57	2.20	0.52	0.64
Al ₂ O ₃	16.7	15.9	15.7	14.6	16.1	15.9	14.9
Fe ₂ O ₃	1.4	2.2	0.8	1.8	4.2	1.0	2.0
FeO	9.5	9.9	10.9	8.9	10.6	10.4	9.6
MnO	0.15	0.23	0.20	0.18	0.23	0.15	0.26
MgO	5.31	6.90	7.19	5.42	7.69	8.78	5.89
CaO	4.27	5.48	4.34	7.03	8.98	3.65	6.52
Na ₂ O	5.1	3.2	3.0	3.1	1.4	3.4	-
K ₂ O	0.84	0.43	1.39	0.50	0.43	0.75	2.08
P ₂ O ₅	0.09	0.38	0.16	0.32	0.54	0.09	0.16
CO ₂	0.5	0.4	0.7	1.0	1.1	0.1	1.0
H ₂ O	1.3	1.1	2.0	1.5	1.7	1.7	1.7
Total	100.08	100.60	100.54	100.42	98.97	101.14	100.55
Minor Elements							
Ba	.019	.006	.018	.011	.002	.003	.047
Co	.0045	.0072	.0070	.0060	.0069	.0061	.0050
Cr	.0017	.0032	.0026	.0020	.0170	.0049	.0035
Cu	.0130	.0290	.0054	.0025	.0089	.0120	.0031
Ni	.0022	.0044	.0043	.0032	.0110	.0040	.0031
Rb	-	-	-	.001	-	-	.001
S	.07	.15	.07	.22	.08	.08	.10
Sr	.0170	.0056	.0085	.0110	.0190	.0091	.0065
V	.0270	.0031	.0350	.0130	.0250	.0280	.0150
Y	.0036	.0052	.0061	.0047	.0043	.0040	.0042
Zn	.013	.020	.023	.017	.012	.016	.016
Zr	.0059	.0075	.0100	.0049	.0130	.0032	.0054

QZ RICH HB GNEISS
UNIT 11

Specimen No.	9	14	18	28	31	32	37
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Major oxides

SiO ₂	64.9	76.3	72.0	73.4	60.5	70.5	69.6
TiO ₂	0.31	0.14	0.10	0.27	0.54	0.30	0.24
Al ₂ O ₃	13.6	11.3	9.6	12.8	12.5	11.0	12.2
Fe ₂ O ₃	-	-	1.6	0.7	2.7	1.9	0.4
FeO	4.7	1.6	5.7	1.0	7.9	3.8	4.3
MnO	0.12	0.04	0.20	0.04	0.18	0.20	0.08
MgO	1.91	0.75	3.00	0.84	4.72	1.48	1.78
CaO	7.53	4.37	5.30	4.84	6.32	5.64	5.11
Na ₂ O	1.7	1.7	0.5	2.5	0.6	0.2	0.1
K ₂ O	1.66	1.61	2.10	2.38	0.86	1.35	2.08
P ₂ O ₅	0.11	0.06	0.20	0.09	0.12	0.17	0.09
CO ₂	2.5	1.2	0.6	1.1	1.1	0.9	1.5
H ₂ O	0.8	1.0	0.9	0.4	1.8	0.9	1.5
Total	99.84	100.07	101.70	100.34	99.84	98.34	98.98

Minor Elements

Ba	.018	.039	.023	.032	.018	.021	.030
Co	.0020	-	.0030	.0018	.0032	.0019	.0026
Cr	-	-	.0011	-	.0008	-	.0007
Cu	.0150	.0043	.5600	.0029	.0130	.0065	.0014
Ni	.0016	-	.0030	.0014	.0014	.0015	.0014
Rb	.004	.001	-	.001	-	-	.003
S	.19	.06	.20	.08	.69	.05	.06
Sr	.0200	.0062	.0053	.0270	.0094	.0069	.0053
V	.0110	-	.0060	.0280	.0059	-	.0033
Y	.0034	.0029	.0051	.0042	.0033	.0044	.0041
Zn	.011	-	-	.011	-	.016	.017
Zr	.0056	.0100	.0240	.0030	.0037	.0067	.0049

Rock Type

Specimen No.	46	56	58	79	87	100	103
<u>Major oxides</u>							
SiO ₂	73.4	67.8	62.6	65.3	56.5	72.0	59.9
TiO ₂	0.11	0.26	0.56	0.42	0.66	0.21	0.43
Al ₂ O ₃	9.3	12.5	14.4	13.1	14.7	10.6	14.6
Fe ₂ O ₃	-	1.2	0.7	1.0	2.1	0.7	1.5
FeO	1.9	4.6	4.4	5.6	10.2	5.4	5.9
MnO	0.04	0.18	0.11	0.20	0.15	0.15	0.17
MgO	2.25	1.18	2.36	2.94	3.40	2.25	4.17
CaO	5.37	5.51	7.46	7.17	4.52	4.43	6.87
Na ₂ O	0.2	1.5	2.3	0.2	2.0	1.1	0.7
K ₂ O	3.84	2.02	2.26	1.19	3.35	2.21	4.12
P ₂ O ₅	0.05	0.16	0.11	0.20	0.17	0.06	0.08
CO ₂	2.2	1.1	1.2	0.9	1.1	0.7	1.2
H ₂ O	0.5	0.7	1.8	1.6	1.6	0.8	1.0
Total	99.03	98.08	99.16	100.02	100.45	100.61	100.64

Minor Elements

Ba	.060	.045	.019	.026	.002	.040	.061
Co	.0012	.0019	.0028	.0023	.0075	.0021	.0029
Cr	-	.0011	.0006	.0007	.0021	-	.0012
Cu	.0027	.0037	.0044	.0029	.0180	.0054	.0020
Ni	.0010	.0016	.0015	.0012	.0029	.0011	.0016
Rb	-	.001	.001	-	.005	-	.002
S	.05	.17	.08	.07	.16	.13	.07
Sr	.0130	.0099	.0140	.0190	.0055	.0110	.0110
V	-	-	.0140	.0028	.0160	-	.0220
Y	.0060	.0045	.0035	.0033	.0049	.0050	.0280
Zn	.017	.007	.015	.031	.016	.014	.011
Zr	.0120	.0070	.0073	.0064	.0076	.0110	.0045

Rock Type

MARBLE/CALC-SILICATE
UNIT 12 / UNIT 11a

Specimen No.	106	139	146	162	1▼	2▼	3▼
Major oxides							
SiO ₂	78.3	77.5	61.9	75.5	30.2	37.0	25.1
TiO ₂	0.19	0.16	0.40	0.17	0.49	1.39	0.35
Al ₂ O ₃	8.6	12.3	14.8	12.2	5.5	17.6	3.6
Fe ₂ O ₃	0.8	-	0.9	0.3	1.1	2.8	0.6
FeO	1.4	1.6	6.6	2.5	3.7	2.0	0.8
MnO	0.08	0.04	0.14	0.07	0.12	0.08	0.09
MgO	1.06	0.76	3.31	1.20	6.18	4.12	1.23
CaO	6.43	2.83	3.77	3.40	33.1	21.2	37.8
Na ₂ O	0.1	2.5	0.8	2.5	-	1.2	-
K ₂ O	0.86	2.41	5.55	2.30	0.59	1.53	1.02
P ₂ O ₅	0.07	0.03	0.11	0.07	0.13	0.55	0.10
CO ₂	1.1	0.3	0.5	0.2	18.1	8.8	28.5
H ₂ O	0.7	0.3	1.1	0.4	0.3	1.3	-
Total	99.71	100.73	99.88	100.85	99.48	98.04	99.15

Minor Elements

Ba	.015	.042	.066	.045	.007	.039	.012
Co	.0013	-	.0030	.0014	.0064	.0071	.0018
Cr	-	-	.0018	.0006	.0068	.0032	.0039
Cu	.0120	.0023	.0016	.0180	.0056	.0120	.0038
Ni	.0012	.0013	.0025	.0012	.0037	.0069	.0091
Rb	.001	.002	.006	.003	.002	.002	-
S	.06	.08	.07	.06	.40	.91	.39
Sr	.0062	.0110	.0110	.0130	.0160	.0410	.0110
V	-	.0027	.0082	.0230	.0120	.0320	.0056
Y	.0047	.0032	.0039	.0041	.0030	.0045	-
Zn	.014	-	.015	.012	.005	.008	.001
Zr	.0079	.0064	.0057	.0070	.0043	.0160	-

Rock Type

Specimen No.	8	11	12	26	35	38	39
Major oxides							
SiO ₂	51.5	53.8	48.0	57.6	19.5	60.9	50.4
TiO ₂	0.20	0.53	0.49	0.43	0.02	0.26	0.53
Al ₂ O ₃	13.0	16.3	16.3	15.3	3.9	12.7	14.4
Fe ₂ O ₃	0.4	0.6	0.8	1.5	-	0.5	1.0
FeO	4.1	8.7	10.2	5.9	0.7	4.4	8.0
MnO	0.14	0.21	0.27	0.14	0.07	0.16	0.21
MgO	1.95	3.22	4.62	3.10	1.84	1.64	5.80
CaO	19.2	12.80	13.70	11.10	40.5	13.5	15.0
Na ₂ O	0.5	1.3	0.8	2.8	-	0.3	0.6
K ₂ O	0.64	0.69	0.54	0.32	0.19	0.45	0.25
P ₂ O ₅	0.09	0.08	0.08	0.17	0.05	0.10	0.26
CO ₂	6.8	2.1	2.4	1.3	33.2	3.3	2.9
H ₂ O	0.3	1.0	1.9	0.7	0.4	0.5	1.2
Total	98.82	101.33	100.10	100.36	100.37	98.71	100.55

Minor Elements

Ba	.015	.008	.007	.013	.006	.020	.003
Co	.0043	.0060	.0034	-	.0024	.0043	
Cr	.0008	.0011	.0014	-	.0074	.0035	
Cu	.0014	.0140	.0600	.0026	.0007	.0020	.0035
Ni	.0015	.0028	.0030	.0028	-	.0018	.0034
Rb	-	-	-	-	-	-	.002
S	.06	.05	.40	.13	.06	.05	.04
Sr	.0190	.0140	.0260	.0290	.0190	.0021	.0190
V	.0065	.0320	.0360	.0160	-	.0068	.0200
Y	.0043	.0034	.0045	.0042	.0022	.0045	.0039
Zn	.018	.007	.008	.005	.003	.020	.014
Zr	.0073	.0037	.0054	.0058	.0077	.0083	.0026

Rock Type

Specimen No.	50	55	64	67	69	71	75
Major oxides							
SiO ₂	51.4	42.9	47.4	46.0	45.4	45.3	40.1
TiO ₂	0.61	0.58	0.57	0.64	0.63	0.90	-
Al ₂ O ₃	13.9	16.8	13.9	17.5	12.1	15.3	0.7
Fe ₂ O ₃	0.7	-	-	1.0	0.7	1.1	16.5
FeO	9.0	3.9	9.7	10.0	1.3	8.7	12.3
MnO	0.16	0.09	0.11	0.17	0.03	0.20	0.40
MgO	4.37	5.23	6.69	5.08	2.97	6.11	9.02
CaO	12.8	19.7	14.8	14.8	19.8	16.1	16.9
Na ₂ O	1.5	0.1	2.2	1.6	3.0	1.6	-
K ₂ O	1.73	2.52	0.51	0.75	2.00	0.39	0.11
P ₂ O ₅	0.27	0.09	0.20	0.18	0.17	0.09	0.57
CO ₂	2.2	4.3	1.1	1.4	1.9	1.5	0.9
H ₂ O	0.9	3.0	0.9	1.6	0.4	1.5	0.6
Total	99.54	99.21	98.08	100.72	100.40	98.79	98.09

Minor Elements

Ba	.021	.052	.003	.009	.046	.009	-
Co	.0036	.0055	.0042	.0073	.0043	.0059	.0052
Cr	.0013	.0055	.0019	.0020	.0005	.0037	.0036
Cu	.0160	.0085	.0076	.0180	.0043	.0110	.0200
Ni	.0046	.0087	.0029	.0046	.0033	.0120	.0062
Rb	.002	.002	-	-	.001	-	-
S	.07	.06	.91	.02	.34	.25	.92
Sr	.0380	.0120	.0210	.0330	.0450	.0170	.0140
V	.0330	.0300	.0160	.0350	.0350	.0380	-
Y	.0048	.0023	.0043	.0054	.0036	.0058	.0066
Zn	.061	.010	.016	-	.003	.013	.017
Zr	.0080	.0061	.0610	.0093	.0076	.0100	.0083

Rock Type

Specimen No.	109	114	116	125	128	135	137
<u>Major oxides</u>							
SiO ₂	71.7	56.1	56.5	47.6	55.5	47.1	77.2
TiO ₂	0.10	0.44	0.09	0.44	0.69	0.64	0.17
Al ₂ O ₃	9.7	10.5	9.8	12.5	18.2	17.2	11.0
Fe ₂ O ₃	0.4	3.4	0.3	1.3	0.4	0.4	-
FeO	0.4	5.4	1.4	8.3	5.3	10.3	1.8
MnO	0.12	0.62	0.14	0.21	0.13	0.21	0.04
MgO	1.13	4.27	2.47	6.09	2.36	4.91	0.46
CaO	9.2	13.9	18.0	18.6	10.00	14.10	5.7
Na ₂ O	0.2	-	0.1	0.4	3.9	2.0	1.3
K ₂ O	2.31	0.22	0.21	0.64	1.95	0.28	1.60
P ₂ O ₅	0.03	0.13	0.07	0.14	0.18	0.18	0.07
CO ₂	4.5	3.5	10.1	2.7	1.5	2.3	0.9
H ₂ O	0.6	1.5	1.2	0.8	0.4	1.4	0.2
Total	100.61	99.98	100.38	99.72	100.51	101.02	100.44

Minor Elements

Ba	.100	.006	.012	.018	.023	.009	.043
Co	.0016	.0011	.0010	.0062	.0050	.0054	-
Cr	-	.0012	-	.0013	.0079	.0030	-
Cu	.0027	.0160	.0110	.0180	.0110	.0120	.0029
Ni	.0011	.0038	.0013	.0034	.0031	.0025	-
Rb	-	-	-	-	.001	-	-
S	.08	.48	.11	.25	.12	.13	.05
Sr	.0087	.0110	.0130	.0170	.0140	.0076	.0140
V	-	.0150	-	.0200	.0280	.0026	-
Y	.0026	.0052	.0053	.0045	.0042	.0050	.0032
Zn	.013	.030	.014	.018	.016	.015	.010
Zr	.0090	.0089	.0200	.0026	.0040	.0046	.0075

Rock Type			AMPHIBOLITE PART OF UNIT 11				
Specimen No.	138	150	49	82	88	133	143
Major oxides							
SiO ₂	57.4	52.9	62.7	45.5	57.7	60.8	52.5
TiO ₂	0.41	0.59	0.54	0.39	0.58	0.83	0.69
Al ₂ O ₃	13.5	15.4	12.9	16.6	14.3	14.1	15.4
Fe ₂ O ₃	1.3	1.5	1.1	2.1	2.4	1.2	0.6
FeO	7.6	8.4	6.3	9.8	9.0	9.5	10.8
MnO	0.17	0.23	0.16	0.22	0.21	0.20	0.24
MgO	4.15	3.69	2.09	9.73	3.83	2.10	6.66
CaO	11.10	13.00	7.6	9.79	7.90	5.97	5.89
Na ₂ O	2.1	0.8	2.8	0.6	1.9	3.7	2.4
K ₂ O	0.36	0.30	0.38	1.58	0.66	0.66	1.62
P ₂ O ₅	0.11	0.17	0.16	0.04	0.21	0.50	0.32
CO ₂	1.6	1.9	1.4	0.5	0.8	-	0.9
H ₂ O	0.9	1.0	0.7	2.3	1.3	0.9	1.6
Total	100.70	99.98	98.81	99.15	100.79	100.46	99.62

Minor Elements

Ba	.003	.009	.016	.016	.036	.034	.046
Co	.0035	.0039	.0040	.0086	.0013	.0040	.0064
Cr	.0012	.0025	.0011	.0130	-	.0011	.0017
Cu	.0120	.0130	.0160	.0190	.0024	.0300	.0120
Ni	.0022	.0024	.0024	.0064	.0012	.0025	.0030
Rb	-	-	-	.001	-	.001	-
S	.07	.10	.19	.05	.19	.07	.09
Sr	.0150	.0200	.0180	.0120	.0031	.0076	.0120
V	.0150	.0150	.0160	.0230	-	.0026	.0290
Y	.0044	.0035	.0038	.0035	.0046	.0050	.0045
Zn	.016	.015	-	.020	.010	.015	.018
Zr	.0064	.0033	.0061	.0026	.0150	.0046	.0063

Rock Type			LAYERED AMPHIBOLITE UNIT 13				
Specimen No.	164	167	13	27	51	52	63
<u>Major oxides</u>							
SiO ₂	55.1	56.0	53.3	53.6	51.9	50.6	51.5
TiO ₂	1.69	0.72	0.49	0.43	0.65	0.47	0.57
Al ₂ O ₃	15.5	15.9	15.9	16.5	17.5	16.5	13.8
Fe ₂ O ₃	4.3	1.8	1.0	1.7	1.0	1.2	0.7
FeO	11.4	9.1	8.9	8.0	11.3	9.0	8.9
MnO	0.18	0.18	0.17	0.19	0.18	0.18	0.22
MgO	2.14	3.86	4.47	5.46	5.08	5.09	6.93
CaO	8.56	7.57	8.77	8.44	6.92	9.84	11.70
Na ₂ O	1.7	0.2	2.2	2.4	3.8	2.3	0.9
K ₂ O	1.90	2.80	0.80	0.29	0.39	0.63	0.45
P ₂ O ₅	0.63	0.36	0.12	0.06	0.10	0.10	0.46
CO ₂	0.2	0.7	1.9	1.6	0.2	2.2	1.2
H ₂ O	2.1	1.4	2.1	1.7	0.9	2.1	1.5
Total	101.12	100.59	100.12	100.41	99.92	100.21	98.83
<u>Minor Elements</u>							
Ba	.019	.017	.017	.005	.007	.007	.005
Co	.0037	.0035	.0048	.0056	.0052	.0040	.0033
Cr	.0009	.0013	.0042	.0029	.0014	.0013	.0016
Cu	.0130	.0039	.0095	.0110	.0130	.0130	.0079
Ni	.0042	.0022	.0030	.0034	.0028	.0021	.0026
Rb	.003	.003	-	-	-	-	-
S	1.30	.08	.05	.09	.09	.25	.22
Sr	.0170	.0100	.0180	.0120	.0096	.0120	.0270
V	.0061	.0013	.0150	.0260	.0370	.0300	.0078
Y	.0062	.0043	.0037	.0046	.0039	.0028	.0039
Zn	.011	.012	.009	.008	-	.002	.012
Zr	.0160	.0040	.0053	.0050	.0056	.0040	.0076

Rock Type

Specimen No.	66	107	111	113	119	124	129
Major oxides							
SiO ₂	52.9	52.6	51.7	52.7	52.8	49.6	51.3
TiO ₂	0.64	0.65	0.62	0.43	0.67	0.50	0.64
Al ₂ O ₃	15.5	18.4	17.6	15.3	17.1	16.1	16.4
Fe ₂ O ₃	0.7	1.0	3.3	1.8	3.4	1.5	1.8
FeO	10.3	9.6	8.8	7.8	9.2	9.1	11.3
MnO	0.17	0.22	0.23	0.25	0.18	0.17	0.22
MgO	6.38	2.52	4.36	6.13	4.87	5.91	4.64
CaO	7.40	13.20	9.18	10.70	8.04	10.40	8.42
Na ₂ O	3.0	1.2	2.1	0.5	2.7	3.0	3.2
K ₂ O	0.74	0.57	0.66	0.69	0.54	0.53	0.42
P ₂ O ₅	0.06	0.17	0.13	0.14	0.13	0.10	0.17
CO ₂	0.9	0.4	0.5	1.6	-	2.0	1.2
H ₂ O	1.7	0.6	1.3	1.7	1.0	1.6	1.0
Total	100.39	101.93	100.48	99.74	100.63	100.51	100.71

Minor Elements

Ba	.007	.013	.010	.034	.050	.040	.004
Co	.0067	.0047	.0056	.0038	.0027	.0061	.0067
Cr	.0031	.0014	.0020	.0035	.0009	.0020	.0050
Cu	.0033	.0120	.0085	.0085	.0035	.0061	.0066
Ni	.0041	.0030	.0031	.0029	.0018	.0037	.0046
Rb	-	.001	-	-	-	.001	-
S	.18	.06	.10	.30	.08	.04	.09
Sr	.0140	.0230	.0200	.0120	.0160	.0140	.0160
V	.0320	.0280	.0310	.0200	.0042	.0310	.0200
Y	.0042	.0047	.0046	.0061	.0048	.0049	.0059
Zn	.016	.014	.017	.015	.018	.019	.021
Zr	.0069	.0079	.0059	.0100	.0086	.0048	.0130

Rock Type	MASSIVE AMPHIBOLITE UNIT 14						
Specimen No.	145	147	149	163	168	10	20
<u>Major oxides</u>							
SiO ₂	51.5	50.9	50.1	56.6	50.6	38.1	55.1
TiO ₂	0.53	0.58	0.67	0.54	0.42	3.56	0.68
Al ₂ O ₃	15.6	16.6	17.5	13.7	14.5	17.5	16.2
Fe ₂ O ₃	2.2	1.5	3.9	2.0	1.8	5.7	3.8
FeO	9.4	8.5	9.7	8.6	9.1	15.6	8.3
MnO	0.19	0.15	0.23	0.13	0.21	0.34	0.20
MgO	5.27	5.51	6.42	4.95	8.89	5.46	4.46
CaO	8.50	8.75	7.00	8.78	5.93	9.12	8.83
Na ₂ O	4.0	3.6	2.2	2.5	4.1	0.3	1.5
K ₂ O	0.69	0.53	0.85	0.67	2.53	0.37	0.46
P ₂ O ₅	0.10	0.16	0.17	0.13	0.09	0.22	0.20
CO ₂	1.2	1.4	0.6	1.0	0.9	1.0	-
H ₂ O	1.5	1.4	1.2	1.4	1.9	1.3	0.8
Total	100.68	99.58	100.54	101.00	100.97	98.57	100.53
<u>Minor Elements</u>							
Ba	.008	.003	.010	.008	.016	.002	.017
Co	.0058	.0040	.0060	.0054	.0045	.0130	.0070
Cr	.0027	.0012	.0031	.0021	.0080	.0110	.0034
Cu	.0098	.0091	.0150	.0061	.0092	.0100	.0069
Ni	.0038	.0028	.0037	.0033	.0071	.0110	.0040
Rb	-	-	.001	-	-	-	-
S	.10	.09	.09	.09	.07	.12	.06
Sr	.0140	.0250	.0180	.0210	.0140	.0120	.0230
V	.0280	.0260	.0290	.0180	.0220	.0390	.0320
Y	.0042	.0033	.0041	.0050	.0021	.0091	.0056
Zn	.016	.011	.015	.016	-	.019	.009
Zr	.0034	-	.0048	.0066	-	.0180	.0056

Rock Type

Specimen No.	21	24	30	45	48	54	76
Major oxides							
SiO ₂	52.9	52.2	47.8	49.2	53.0	51.6	47.2
TiO ₂	0.69	0.65	1.50	0.49	0.59	1.41	0.69
Al ₂ O ₃	16.0	16.0	15.9	15.3	14.4	12.8	13.7
Fe ₂ O ₃	3.8	2.0	2.8	4.1	2.8	2.6	1.1
FeO	9.3	9.4	8.4	7.8	10.0	15.7	9.3
MnO	0.22	0.25	0.16	0.20	0.23	0.28	0.19
MgO	5.33	4.73	6.08	7.43	4.90	5.33	11.8
CaO	8.62	9.70	9.21	9.83	8.57	7.89	10.80
Na ₂ O	2.2	1.0	2.7	1.2	1.7	1.3	1.8
K ₂ O	0.55	0.27	0.37	0.44	0.66	0.43	0.38
P ₂ O ₅	0.14	0.16	0.37	0.08	0.19	0.20	0.07
CO ₂	-	1.4	1.7	1.6	-	-	1.1
H ₂ O	1.0	1.7	1.7	1.7	1.2	1.0	1.9
Total	100.75	100.46	98.69	99.37	98.24	100.54	100.03

Minor Elements

Ba	.016	.002	.004	.004	.005	.005	.004
Co	.0072	.0060	.0050	.0059	.0061	.0100	.0088
Cr	.0037	.0027	.0098	.0071	.0027	.0036	.0059
Cu	.0057	.0140	.0067	.0014	.0038	.0065	.0760
Ni	.0044	.0031	.0071	.0046	.0035	.0072	.0200
Rb	-	-	-	-	-	-	.001
S	.04	.07	.11	.05	.05	.21	.07
Sr	.0220	.0270	.0310	.0160	.0240	.0059	.0160
V	.0330	.0280	.0170	.0290	.0310	.0480	.0240
Y	.0051	.0042	.0043	.0040	.0038	.0110	.0041
Zn	.008	.007	.010	.008	.006	.015	.013
Zr	.0062	.0030	.0140	.0057	.0061	.0150	.0091

Rock Type

Specimen No.	84	85	86	92	93	94	95
Major oxides							
SiO ₂	43.3	48.3	44.0	51.3	49.3	52.6	53.6
TiO ₂	2.43	0.56	2.35	0.60	0.70	0.69	0.66
Al ₂ O ₃	15.8	16.3	15.6	15.8	16.2	16.1	16.1
Fe ₂ O ₃	2.6	1.9	1.8	2.6	3.6	3.4	3.9
FeO	12.4	9.7	12.7	9.3	10.3	9.8	8.8
MnO	0.25	0.20	0.23	0.30	0.21	0.22	0.21
MgO	6.68	5.78	6.14	5.59	6.30	5.69	5.44
CaO	10.20	10.20	10.40	7.69	9.44	8.84	8.73
Na ₂ O	1.7	3.3	2.7	2.9	2.0	1.7	2.1
K ₂ O	0.39	0.57	0.38	0.57	0.66	0.34	0.38
P ₂ O ₅	0.26	0.07	0.24	0.14	0.12	0.15	0.14
CO ₂	2.4	2.2	1.7	1.7	0.4	-	-
H ₂ O	1.9	1.5	1.9	1.6	1.4	0.9	0.9
Total	100.31	100.58	100.14	99.99	100.63	100.43	100.96

Minor Elements

Ba	.004	.003	.002	.007	.015	.003	.004
Co	.0069	.0062	.0062	.0071	.0067	.0065	.0051
Cr	.0150	.0032	.0032	.0041	.0039	.0036	.0032
Cu	.0057	.0110	.0066	.0120	.0140	.0062	.0070
Ni	.0072	.0040	.0076	.0043	.0042	.0041	.0027
Rb	.003	-	-	-	-	-	-
S	.09	.07	.12	.12	.07	.07	.05
Sr	.0480	.0280	.0033	.0170	.0190	.0210	.0240
V	.0260	.0340	.0300	.0290	.0300	.0300	.0290
Y	.0041	.0041	.0050	.0051	.0042	.0047	.0028
Zn	-	.015	.011	.016	.015	.014	-
Zr	.0110	.0048	.0150	.0093	.0061	.0060	.0047

Rock Type			MASSIVE AMPHIBOLITE UNIT 17				
Specimen No.	151	191	17	41	42	60	70
Major oxides							
SiO ₂	43.1	56.1	50.8	47.0	52.1	55.2	47.4
TiO ₂	2.69	0.71	0.48	0.53	0.55	0.58	1.12
Al ₂ O ₃	15.8	14.5	16.3	16.9	15.3	15.8	13.8
Fe ₂ O ₃	3.3	1.0	1.0	3.6	2.5	4.8	1.5
FeO	12.1	9.9	9.9	9.3	9.1	6.8	10.5
MnO	0.27	0.17	0.19	0.21	0.19	0.20	0.20
MgO	6.74	7.96	6.15	6.29	6.91	4.35	8.18
CaO	9.33	7.96	9.28	9.47	6.10	8.50	10.60
Na ₂ O	1.8	1.7	1.8	1.9	4.5	1.9	2.4
K ₂ O	0.27	0.26	0.37	0.69	0.77	0.70	0.59
P ₂ O ₅	0.29	0.27	0.08	0.08	0.08	0.13	0.13
CO ₂	2.9	1.6	1.7	2.5	0.8	-	0.7
H ₂ O	1.8	1.3	2.5	2.2	1.8	0.9	1.5
Total	100.39	98.80	100.55	100.67	100.70	99.76	98.62
Minor Elements							
Ba	.002	.008	.006	.008	.006	.012	.007
Co	.0085	.0026	.0069	.0062	.0062	.0047	.0082
Cr	.0140	.0020	.0062	.0026	.0027	.0017	.0180
Cu	.0074	.0052	.0150	.0063	.0120	.0039	.0016
Ni	.0091	-	.0046	.0046	.0046	.0027	.0100
Rb	-	-	-	-	-	-	-
S	.24	-	.08	.12	.06	.05	.15
Sr	.0410	.0304	.0240	.0280	.0110	.0260	.0150
V	.0320	.0169	.0270	.0330	.0270	.0180	.0340
Y	.0072	.0044	.0044	.0048	.0044	.0041	.0052
Zn	.016	-	.006	.002	.004	.019	.018
Zr	.0089	.0142	.0070	.0080	.0074	.0032	.0014

Rock Type AMPHIBOLITE, META-GABBRO SILL
UNIT 17

Specimen No. 53 72 73 183 184 185 195

Major oxides

SiO ₂	49.4	51.9	51.3	45.5	39.8	44.2	48.5
TiO ₂	0.74	0.72	0.73	1.07	1.21	1.25	0.18
Al ₂ O ₃	17.1	17.6	17.2	17.0	16.7	11.9	3.2
Fe ₂ O ₃	1.6	2.1	1.5	2.1	3.2	2.8	3.3
FeO	8.4	7.3	7.7	10.1	11.9	10.6	5.2
MnO	0.20	0.19	0.18	0.25	0.26	0.25	0.18
MgO	5.92	4.50	5.25	6.45	7.85	12.6	21.3
CaO	10.40	8.78	9.77	10.60	12.7	11.3	11.7
Na ₂ O	2.7	3.0	2.7	2.7	1.8	0.8	0.1
K ₂ O	1.18	1.29	1.31	0.76	0.68	0.88	0.19
P ₂ O ₅	0.29	0.29	0.39	0.52	0.71	0.11	0.12
CO ₂	-	-	-	-	0.1	0.4	1.8
H ₂ O	1.2	1.0	1.0	1.3	1.8	2.2	2.4
Total	99.13	98.67	99.03	98.65	98.70	99.29	98.17

Minor Elements

Ba	.032	.059	.040	.025	.016	.016	.004
Co	.0043	.0037	.0065	.0038	.0040	.0053	.0062
Cr	.0064	.0035	.0043	.0054	.0062	.0354	.1720
Cu	.0081	.0110	.0031	.0090	.0055	.0015	.0180
Ni	.0033	.0026	.0041	.0035	.0037	.0102	.0520
Rb	.001	-	.001	-	-	-	-
S	.07	.10	.04	.09	.12	.03	.05
Sr	.0590	.0670	.0830	.0490	.0414	.0093	.0096
V	.0230	.0180	.0300	.0430	.0444	.0511	.0092
Y	.0034	.0031	.0054	.0044	.0049	.0047	-
Zn	.012	.010	.012	-	-	-	-
Zr	.0072	.0100	.0120	.0082	.0082	.0093	.0062

Rock Type	CREE LAKE INTRUSIVES				META-FELSIC VOLCANICS UNIT 15		
Specimen No.	186	187	188	189	34	180	181
Major oxides							
SiO ₂	47.2	48.7	43.1	43.0	70.4	68.0	65.9
TiO ₂	0.60	0.54	0.23	0.32	0.29	0.88	1.0}
Al ₂ O ₃	14.8	14.3	2.6	7.8	14.1	14.7	12.8
Fe ₂ O ₃	1.2	1.1	5.8	5.7	0.3	-	-
FeO	7.5	7.1	5.9	5.7	1.3	1.4	1.7
MnO	0.15	0.15	0.17	0.15	0.02	0.02	0.04
MgO	12.90	11.90	30.10	22.50	3.64	1.61	3.80
CaO	10.5	11.9	3.99	10.10	2.51	7.66	6.73
Na ₂ O	0.7	2.0	0.2	0.3	2.9	4.9	2.6
K ₂ O	0.95	0.67	0.12	0.17	2.28	0.80	0.65
P ₂ O ₅	0.08	0.08	0.05	0.07	0.06	0.95	1.0
CO ₂	0.1	0.1	2.2	1.4	0.3	0.5	1.0
H ₂ O	2.9	1.6	4.9	4.3	1.0	0.5	0.7
Total	99.58	100.14	99.36	99.91	99.10	101.92	97.93

Minor Elements

Ba	.021	.010	.001	.002	.017	.015	.016
Co	.0028	.0038	.0117	.0070	.0014	.0013	.0036
Cr	.1120	.0410	.2950	.2270	-	-	-
Cu	.0010	.0093	.0042	.0052	.0039	.0091	.0081
Ni	.0152	.0136	.1450	.0704	.0013	-	-
Rb	.001	-	-	-	.001	-	-
S	-	.10	.22	.53	.03	.07	-
Sr	.0080	.0116	.0088	.0024	.0170	.0116	.0210
V	.0256	.0291	.0076	.0159	.0027	.0061	.0148
Y	-	-	-	-	.0026	-	.0041
Zn	-	-	-	-	.020	-	-
Zr	.0074	.0085	.0072	.0076	.009	.0076	.0085

Rock Type			META-GRANITIC ROCKS UNIT 18				
Specimen No.	182	194	47	91	104	105	190
Major oxides							
SiO ₂	62.1	72.3	78.4	75.2	75.9	74.8	71.3
TiO ₂	1.07	0.26	0.06	0.18	0.19	0.17	0.27
Al ₂ O ₃	16.4	15.4	10.4	12.3	11.7	13.0	15.1
Fe ₂ O ₃	0.2	0.1	-	0.6	0.7	1.2	0.1
FeO	1.6	1.0	1.3	2.6	2.6	1.9	2.3
MnO	0.03	0.01	0.04	0.09	0.07	0.12	0.06
MgO	3.45	0.58	0.64	1.58	1.50	0.93	1.21
CaO	10.60	1.94	1.44	0.83	1.39	1.83	3.10
Na ₂ O	4.5	4.7	1.6	2.9	1.3	3.0	3.1
K ₂ O	1.45	3.27	3.33	4.30	3.81	3.12	2.6
P ₂ O ₅	0.80	0.13	0.03	0.14	0.08	0.06	0.14
CO ₂	0.9	0.5	0.4	0.1	0.3	0.2	0.6
H ₂ O	0.9	0.3	0.7	0.9	0.9	0.3	0.7
Total	103.70	100.49	98.34	101.72	100.44	100.63	100.53
Minor Elements							
Ba	.029	.077	.096	.085	.070	.084	.088
Co	.0020	-	.0012	.0012	.0016	.0011	-
Cr	-	-	-	-	-	.0006	.0012
Cu	.0091	.0033	.0017	.0047	.0071	.0020	.0008
Ni	-	-	.0014	.0011	.0013	.0013	-
Rb	-	.006	.001	.002	.003	.001	.002
S	.08	.06	.06	.07	.06	.09	.01
Sr	.0210	.0300	.0100	.0061	.0067	.0076	.0384
V	.0068	-	-	-	.0022	-	.0066
Y	-	-	.0042	.0023	.0037	.0023	-
Zn	-	-	.018	-	.011	.016	-
Zr	.0083	.0152	.0150	.0092	.0110	.0086	.0230

Rock Type	NOKOMIS PARAGNEISS UNIT 3					NOKOMIS AMPHIBOLITE UNIT 1
Specimen No.	170	171	172	192	193	169
Major oxides						
SiO ₂	69.6	69.5	68.7	64.6	56.8	49.8
TiO ₂	0.25	0.26	0.32	0.66	1.15	0.84
Al ₂ O ₃	15.9	15.8	15.8	15.8	15.0	15.5
Fe ₂ O ₃	0.2	0.5	-	2.0	3.6	1.3
FeO	2.2	2.5	2.4	3.3	6.1	7.3
MnO	0.06	0.09	0.04	0.07	0.16	0.18
MgO	1.11	1.25	1.24	1.52	4.69	6.55
CaO	3.83	3.28	3.70	3.78	6.90	15.80
Na ₂ O	3.5	2.3	3.9	4.3	2.5	1.3
K ₂ O	1.50	2.13	2.02	1.70	0.60	0.54
P ₂ O ₅	0.23	0.22	0.33	0.19	0.19	0.17
CO ₂	-	-	-	0.5	-	0.1
H ₂ O	0.4	0.4	0.3	0.5	0.8	1.0
Total	98.78	98.23	98.75	98.92	98.60	100.38
Minor Elements						
Ba	.046	.061	.096	.028	.025	.019
Co	-	-	-	-	.0035	.0049
Cr	.0007	.0008	.0009	.0006	.0016	.0340
Cu	.0017	.0008	.0011	.0024	.0665	.0048
Ni	-	-	-	-	.0028	.0092
Rb	.004	.005	.004	.007	-	-
S	.07	-	.07	.06	-	.09
Sr	.0179	.0201	.0422	.0219	.0219	.0390
V	.0041	.0049	.0059	.0078	.0105	.0250
Y	.0054	.0043	-	-	.0046	.0039
Zn	-	-	-	-	-	.012
Zr	.0416	.0334	.0370	.0135	.0181	.0072

APPENDIX B

CHEMICAL ANALYSES OF ROCKS FROM A REGOLITH NEAR FLIN FLON, MANITOBA

The results of seven chemical analyses of samples taken from a regolith near Flin Flon are presented in Appendix B. The location of the sample traverse from which the samples were taken is shown on Fig. B-1. Samples were collected, cleaned in the field of all weathered surface and shipped to Ottawa courtesy of Dr. D. Price of the Hudson Bay Exploration and Development Co. Ltd. All samples were analyzed in the same way as the samples in Appendix A.

The geology of the area from which these samples were taken is an outlier of Missi conglomerate on flows of the Amisk Group. Underlying the conglomerates and developed on the flows is a zone of blackening and cellular development similar to modern zones of spheroidal weathering. This zone is considered to be a paleo-regolith developed on the Amisk rocks prior to the deposition of the Missi rocks. The zone is about 20 metres thick.

The depths of the samples taken below the Amisk-Missi contact are as follows:

<u>Sample No.</u>	<u>Depth Below Contact (Metres)</u>
1	2
2	2.5
3	6
4	10
5	8
6	20
7	30

FIG. B-1. Location of sample traverse for samples from Amisk regolith, Flin Flon, Manitoba. (geology from Stockwell, 1960)

Rock Type FLIN FLON REGOLITH

Specimen No.	FFR1	FFR2	FFR3	FFR4	FFR5	FFR6	FFR7
<u>Major oxides</u>							
SiO ₂	58.0	50.5	49.6	62.5	47.7	49.6	54.2
TiO ₂	.79	.73	.70	.96	.71	.63	.96
Al ₂ O ₃	19.5	17.4	15.1	14.9	16.2	15.4	15.0
Fe ₂ O ₃	11.0	8.0	2.8	7.3	1.8	2.0	1.8
FeO	2.0	5.2	7.2	5.3	9.7	9.1	10.5
MnO	.06	.21	.28	.11	.18	.18	.18
MgO	1.20	4.81	5.85	2.48	5.30	5.37	4.09
CaO	-	1.93	3.96	0.55	4.83	7.04	4.00
Na ₂ O	1.9	2.6	3.5	1.0	2.2	2.4	3.1
K ₂ O	2.40	2.29	1.98	1.48	1.76	.29	.14
P ₂ O ₅	.04	.06	.07	.32	.05	.06	.18
CO ₂	-	2.9	4.8	.2	4.1	3.8	3.8
H ₂ O	2.9	1.5	2.3	1.1	4.4	3.8	2.6
Total	99.39	98.13	98.14	98.2	98.43	99.67	100.55

Minor Elements

Ba							
Co							
Cr							
Cu							
Ni							
Rb	.004	.001	-	.002	-	-	-
S	-	-	-	-	-	-	-
Sr							
V							
Y							
Zn	.012	.027	.033	.031	.019	.009	.013
Zr							

APPENDIX C

MINERAL ASSEMBLAGES AND MODES FOR THIN SECTIONS OF ROCKS FROM THE SHERRIDON AREA, MANITOBA

The mineral assemblages and modes for 195 thin sections of the chemically analyzed rocks from the Sherridon area are listed in this appendix. Sample locations are plotted on Fig. A-1 in the pocket. All estimates are in volume percent. A list of mineral abbreviations can be found at the beginning of the thesis. All thin sections were produced by the Geological Survey of Canada.

BI GNEISS: UNIT 5

SPEC. NO.	4	5	7	16	19	23	33	40	43	44	57	61	62	65	83	96	97	99	108	112	122	132	134	140
Q	50	60	55	60	50	65	60	60	60	50	60	60	70	55	80	60	65	65	65	75	60	50	40	45
PL	35	15	15	25	15	20	20	20	25	10	20	25	20	30	30	20	15	20	15	30	25	5	8	
SC																								
KF	20	10	5	20	9	10	4	4	20	10	5						5	25			5	40	35	
BI1	5	5	13	7	10	5	8	10			5	7		11	10	7	10	10	15	10	20	10	9	
BI2									7	14			7											
CH			TR				TR	2										TR			TR			
CH1																								
PI																								
MV	TR				TR	1	2	TR	1	1	5	3				5	5	TR	TR		TR	TR		
PH																								
SE																								
H8																								
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CM																								
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DI																								
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OL																								
AL			7		3			5	3	5	TR		3	2		3						3	3	
GR																								
HC																								
SN																								
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GT																								
MT													TR		TR	TR	TR	TR						
PY																								
PO																								
CP																								

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	SL-BI GNEISS				QUARTZ- FELDSPATHIC GNEISS				QUARTZITIC GNEISS														
	UNIT 5a				UNIT 5b				UNIT 6				UNIT 7										
Q	144	148	152	156	157	158	159	160	161	34	81	126	123	131	15	25	77	110	117	136	176	68	74
PL	50	50	60	50	70	45	55	55	50	50	60	65	70	65	65	60	60	70	70	65	75	80	80
SC	25	15	30	40	30	30	30	30	30	30	35	5	10	10	25	5	10	25	23	23	23	5	5
KF	20											5				20	25	27	2	7			
BI1	13	10	10	8	13	13	13	15	10	10	15	10	7	5	5	3	2	3	3	3	2		
BI2						18						20											
CH																							5
CH1																							
PI																							
MV									3						5	2	3						5
PH																							
SE																							
HB																							
HB1																							
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TR																							
CM																							
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DI																							5 10
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AL	2						5	2	2	2													
GR																							
HC																							
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AMPHIBOLITE
PART OF UNIT 11 LAYERED AMPHIBOLITE: UNIT 13

	109	114	116	125	128	135	137	138	150	49	82	88	133	143	164	167	13	27	51	52	63	66
Q																						
PL																						
SC																						
KF																						
B11																						
B12																						
CH																						
CH1																						
PI																						
MV																						
PH																						
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	MASSIVE AMPHIBOLITE UNIT 17						CREE LAKE INTRUSIVES						AMPHIBOLITE, META-GABBRO SILL: UNIT 17									
	92	93	94	95	151	191	17	41	42	60	70	186	187	188	189	53	72	73	183	184	185	195
Q																						
PL	12	15	15	20	5	30	5	TR	TR	20	TR					15	10	10	TR			
SC	35	25	35	30	30		25	40	45	20	40	30	40			35	40	40	50	30	10	
KF										20												
BI1																						
BI2	TR	3					TR	3		5							3	TR				
CH																						
CH1														10								
PI																						
MV																						
PH																						
SE																						
HB																						
HB1	50	45	35	40	45	30	50	50	50	30	60	70	60	30	15	42	45	50	50	70	90	
AC																						95
TR														55	75							
CH																						
AT																						
DI																						
EP																						
OL															5							
AL	3	12	13	8	12	10	20	5														
GR																						
HC																						
SN																						
CO																						
SL																						
CC																						
SH	TR								TR													
GT																						
MT	TR	TR	2	2				2		5												5
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PO																						
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	META-FELSIC VOLCANICS: UNIT 15			META-GRANITIC ROCKS: UNIT 18			NOKOMIS PARAGNEISS: UNIT 3			NOKOMIS AMPHIBOLITE: UNIT 1						
	34	180	181	182	194	47	91	104	105	190	170	171	172	192	193	169
Q																
PL	45	50	60	60	50	50	45	50	55	50	50	50	50	40	20	5
SC	40	40	30	30	40	10	20	40	20	35	35	30	35	40	45	35
KF					5	30	20		15	5				TR		
B11	12	5	2	5		10	13	10	8	10	15	15	15	17	5	
B12																
CH																
CH1																
PI																
MV																
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HB																
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APPENDIX D

MINERAL ASSEMBLAGES AND MODES FOR POLISHED THIN SECTIONS OF ROCKS FROM SULPHIDE OCCURRENCES AT SHERRIDON, MANITOBA

The mineral assemblages and modes for 25 polished thin sections of rocks from sulphide occurrences at Sherridon are listed in this appendix. Sample locations are given on Fig. 111-4. All estimates are in volume percent. A list of mineral abbreviations can be found at the front of this thesis. Polished thin sections were made by both the Geological Survey of Canada and the Carleton University Geology Department.

	RS-A	RS-B	RS-C	RS-E	RS-F	RS-G
% SULPHIDES						
CP	80	90	90	80	80	85
CU	15	15		20	20	5
P01					TR	
P02	50	70	14	45	19	60
PY			1		1	
MT			45	5	25	115
SP	5	5	15	10	5	5
% SILICATES						
Q	20	10	10	10	20	15
PL	15	10	10	10	10	10
BI	5	TR	TR	TR	TR	TR
MV		TR	TR	TR	10	5
SL						
GH						
DI						
CC						

APPENDIX E

C.I.P.W. NORMS CALCULATED FOR THE IGNEOUS ROCKS OF APPENDIX A

C.I.P.W. norms were calculated for 62 samples from Appendix A which were interpreted as having an igneous origin. All calculations were made using the Carleton University Geology Department computer program LEMNORM 9. All values given in this appendix are in weight percent. A summary of this program's capabilities may be found in the Departmental "Guide to the Use of Computer Programs Often Required by the Department of Geology." (W. Yzerdraat)

MAFIC FRAGMENTALS UNIT 13

	13*	27	51	52	63	66	107	111	113	119	124	129
Q	8.52	9.06		3.65	8.61	2.13	9.32	6.58	13.96	5.06		.90
C	4.92	1.77	2.33	3.88	2.77	4.47	3.37	4.08	4.22	3.21	3.23	2.52
OR	19.36	20.91	32.52	20.26	7.91	25.93	10.14	18.58	4.38	22.95	26.17	27.47
AB	32.40	34.39	29.88	34.19	33.55	27.22	43.07	38.32	38.79	33.10	29.82	29.57
AN												
NE	4.94	3.43	1.34	6.54	11.91	3.96	5.57	2.95	7.57	2.25	9.86	4.03
DI	5.24	2.71	1.72	6.26	8.89	3.62	12.24	3.51	4.91	2.58	8.57	5.51
HE	9.47	12.41	10.09	10.17	12.49	14.40	3.68	9.99	12.29	11.13	5.69	9.86
EN	12.53	11.24	14.89	11.16	10.62	15.09	9.27	13.61	9.15	14.64	5.67	15.49
FS			1.46								3.45	
FO			2.38								3.79	
FA												
WO												
MT	1.51	2.54	1.47	1.81	1.06	1.04	1.45	.30	7.93	3.16	2.24	2.65
IL	.99	.84	1.25	.93	1.13	1.24	1.23	1.23	4.49	1.28	.98	1.23
AP	.41	.45	.42	.43	.53	.40	.51	.56	13.90	.42	.41	.52
PO	.14	.25	.25	.71	.63	.50	.16	.29	7.87	.22	.11	.25

* SPEC. NO.

MAFIC FLOWS UNITS 14a AND 14b

	145	147	149	163	168	10	20	21	24	30	45	48
Q			1.71	10.95		.57	14.15	6.52	10.94	.53	4.87	10.59
C			.792									
OR	4.16	3.24	5.07	4.06	15.24	2.27	2.73	3.27	3.96	2.30	2.71	4.03
AB	34.53	31.46	18.79	21.66	27.07	2.64	12.74	18.69	8.71	23.96	10.59	14.83
AN	23.03	28.48	33.54	24.77	13.94	45.5	36.25	32.29	38.39	31.65	36.57	30.63
NE					4.47							
DI	7.85	6.87		8.16	7.99		2.45	3.64	3.69	7.64	6.18	4.41
HE	7.70	5.82		7.57	4.60		2.91	4.04	4.81	4.78	4.54	5.35
EN	4.79	8.06	16.14	8.84		12.26	10.02	11.64	10.43	12.34	16.42	10.54
FS	5.38	7.84	18.70	9.41		17.83	13.63	14.80	15.58	8.86	13.59	14.67
FO	3.48	2.05			13.20	1.31						
FA	4.31	2.19			9.62	2.10						
WO												
MT	3.00	2.25	3.17	2.97	2.66	7.62	3.17	3.19	2.61	4.26	3.01	3.12
IL	1.03	1.14	1.29	1.05	.81	7.02	1.30	1.32	.49	2.99	.97	1.16
AP	.45	.36	.54	.31	.21	.53	.47	.51	.38	.39	.48	.55
PO	.28	.26	.25	.25	.20	.34	.17	.11		.32	.14	.14

	54	76	84	85	86	92	93	94	95	151	191
Q	9.55					3.12	.47	8.4	8.17		17.97
C											
OR	2.55	2.32	2.40	3.48	2.33	3.49	3.95	2.02	2.25	1.67	1.61
AB	11.04	15.69	14.99	28.81	20.82	25.37	17.14	14.47	17.78	15.90	15.05
AN	27.92	29.03	35.78	28.87	30.37	29.38	33.71	35.49	33.41	35.73	32.62
NE					1.54						
DI	3.13	13.99	6.73	9.5	8.80	3.74	5.1	2.91	3.49	5.03	2.30
HE	4.91	6.17	5.90	8.73	9.09	3.50	5.16	3.05	3.64	3.81	4.10
EN	11.87	7.70	6.83	.33		12.66	13.53	12.90	11.94	10.76	7.85
FS	21.34	3.90	6.86	.34		13.60	15.68	15.51	14.29	9.36	16.04
FO		11.28	5.17	7.09	8.24					3.10	
FA		6.29	5.74	8.23	10.75					2.97	
WD											
MT	3.78	1.64	3.93	2.84	2.70	3.15	3.23	3.19	3.14	4.99	1.52
IL	2.69	1.35	4.81	1.09	4.62	1.18	1.34	1.32	1.25	5.33	.54
AP	.65	.45	.61	.48	.55	.48	.49	.51	.49	.65	.41
PO	.58	.20	.26	.20	.20	.34	.14	.19	.14	.69	

HYPABYSSAL INTRUSIVES UNIT 14a

CREE LAKE INTRUSIVES

	17	41	42	60	70	186	187	188	189
Q	5.17			13.1					
C									
OR	2.27	4.28	4.64	3.60	3.61	5.82	4.02	.77	1.07
AB	15.8	16.77	38.82	16.30	21.01	6.13	17.18	1.84	2.69
AM	36.63	37.08	19.65	33.27	26.01	35.65	28.49	6.35	20.62
NE									
DI	4.23	4.50	4.77	3.17	13.14	10.48	18.49	9.44	21.55
HE	3.94	4.20	3.11	3.65	9.06	3.40	5.97	1.64	3.41
EN	13.93	11.87	6.76	9.52	4.50	18.56	3.43	28.28	10.94
FS	14.90	12.72	4.41	12.60	3.56	6.90	1.27	5.64	1.98
FO		1.66	7.59		7.34	6.90	12.68	34.24	26.99
FA		1.96	4.95		6.40	2.83	5.17	7.52	5.34
WO									
MT	1.50	3.07	3.03	3.06	2.25	1.80	1.62	2.73	2.86
IL	.95	1.05	1.07	1.12	2.20	1.18	1.04	.48	.65
AP	.46	.51	.45	.47	.48	.36	.35	.43	.37
PO	.29	.34	.17	.14	.43		.28	.66	1.54

BOUNDARY GABBRO UNIT 17

	52	72	73	183	184	185	195
Q		2.13					
C							
OR	7.22	7.81	7.96	4.62	5.60	5.39	1.20
AB	23.63	25.98	23.45	19.93	3.21	7.01	.90
AN	32.13	31.47	31.77	32.88	28.27	27.20	8.23
NE							
DI	9.13	5.31	7.36	8.56	16.97	11.32	35.41
HE	7.46	4.48	6.55	7.53	7.56	6.43	6.15
EN	3.86	9.01	9.59			9.31	29.87
FS	3.62	8.73	9.78			5.76	5.95
FO	5.02		.29	8.78	18.14	12.36	7.18
FA	5.18		.34	9.75	10.22	8.43	1.58
WO							
MT	2.36	3.12	2.23	3.13	4.29	2.63	2.80
IL	.15	1.40	.14	2.09	2.55	.49	.36
AP	.05	.45	.43	.60	.62	3.00	.44
PO	.20	.11	.11	.25	.35	.09	.15

101°05

55°10

100° 55'

304

OPEN FILE
DOSSIER PUBLIC

623

GEOLOGICAL SURVEY
COMMISSION GÉOLOGIQUE
OTTAWA

FIG. II-1

GEOLOGY
SHERRIDON AREA
MANITOBA

SCALE 1:25 000

55°

1c

1c

2

1c

2

1c

2

2

2

2

2

Star Lake

2

5

2

P

JUNGLE LAKE

LEGEND

- 12 Pegmatite
- 11 Granodiorite
- 10 Pyroxenite
- 9 Gabbro

4 of

SHERRIDON GROUP

- 8 Massive amphibolite (metamorphosed flows)
- 7 Layered amphibolite with fragments (f) (metamorphosed volcaniclastic rock)
- 6 Impure marble and calc-silicate rocks
- 5 Calc-silicate rocks
- 4 Cordierite-anthophyllite rocks
- 3 Pelitic schist
- 2 Biotite-garnet schist
- 1 Quartz-rich gneisses

NOKOMIS GROUP

- 1 Quartzofeldspathic gneisses (la), migmatites (lb) and granitoid gneisses (lc)

Geological boundary (approximate, assumed)

Foliation (inclined, vertical)

55°10

101°00

5

Impure marble and calc-silicate rocks

Calc-silicate rocks

Cordierite-anthophyllite rocks

4

Pelitic schist

3

Biotite-garnet schist

2

Quartz-rich gneisses

1

NOKOMIS GROUP

Quartzofeldspathic gneisses (1a), migmatites (1b) and granitoid gneisses (1c)

Geological boundary (approximate, assumed)

Foliation (inclined, vertical)

Lineation (inclined)

Fault (defined)

Sulphide orebody

Sulphide mineralization

Stratigraphic horizon locally mineralized

Geology by E. Froese and P.A. Goetz, 1975-1976.

Index Map

Scale: 1" = 50 miles

100° 55"

LATE AND
POST APHEBIAN

20f

EBIAN

GEOLOGY OF SHERRIDON MANITOBA

Scale: 1" = 660'

LEGEND

No Stratigraphic Order Implied With

19 Pegmatite

SHERRIDON GROUP

18 Meta-Granitic Rocks, porphyritic

17 Meta-Gabbro, in part layered
17a Meta-Pyroxenite, associated with Gabbro

16 Sulphide occurrence, massive, disseminated

15 Meta-Felsic Volcanics, flows and pyroclastics undifferentiated

14 Amphibolite, with pillow structures; Basaltic Flows
14a Massive Amphibolite; probable, Basaltic Flows or Subvolcanic Intrusives

13 Amphibolite, with clasts and layering; Basaltic Pyroclastic or Lahar
13a Layered Amphibolite; probable Basaltic Pyroclastic or Lahar

12 Marble, Calc-silicate; Carbonate Bank

POST APHEBIAN

EBIAN

GEOLOGY OF ERRIDON MANITOBA

Scale: 1" = 660'

LEGEND

No Stratigraphic Order Implied Within Groups

ocks, porphyritic

part layered
e, associated with Gabbro

nce, massive, disseminated

onics, flows and pyroclastics undifferentiated

pillow structures; Basaltic Flows
bolite; probable Basaltic Flows or Subvolcanic Intrusives

clasts and layering; Basaltic Pyroclastic or Lahar
bolite; probable Basaltic Pyroclastic or Lahar

cate; Carbonate Bank

40

5 of 7

55° 10"

6 of

ARCHEAN?
AND
APHEBIAN

APHEBIA

55°10"

- 12 Marble, Calc-silicate; Carbonate Bank
- 11 Hb Gneiss $\pm$ Ga, with Qz ridges; Calcareous Lithic Arenite, Hb $< 15\%$
 11a Di-bearing Amphibolite $\pm$ Bi, weathers differentially; Calcareous Argillite or Arenite
- 10 Al-Ga Gneiss $\pm$ Co; Mg Rich Sediments
 10a Cm Gneiss $\pm$ Ga; } Ca and/or
 10b Bi-Ga Gneiss; Bi $> 75\%$ } Na Depleted Rocks
- 9 Bi-Ga-Sl Gneiss (Pelite); Mudstone
- 8 Bi-Ga Schist (Semipelite); Graywacke
- 7 Quartzitic Gneiss, in part Chert, Qz $> 90\%$
- 6 Quartzo-feldspathic Gneiss; Quartzo-feldspathic Arenite, C.I. $< 5\%$
- 5 Bi Gneiss $\pm$ Ga, with Qz ridges; Lithic Arenite; Bi $< 15\%$
 5a Sl-bearing Lithic Arenite
 5b Sl-nodule bearing Lithic Arenite
 5c Mafic Lithic Arenite, Bi $> 15\%$

NOKOMIS GROUP

- 4 Granitic Rocks, undifferentiated
- 3 Stratified Gneiss (Paragneiss)
 3a Migmatite Complex; Stratified Gneiss with $> 20\%$ felsic segregations and pegmatite
- 2 Augen Bi Gneiss
- 1 Amphibolite, undifferentiated

SYMBOLS

■ Shaft (abandoned)

□ Outcrop

--- Geological Contact

with Qz ridges; Calcareous Lithic Arenite, Hb < 15%
Biolite ± Bi, weathers differentially; Calcareous Argillite or Arenite

Mg Rich Sediments
75% } Ca and / or
No Depleted Rocks

Pelite); Mudstone

mpelite); Graywacke

in part Chert, Qz > 90%

ic Gneiss; Quartzo-feldspathic Arenite, C.I. < 5%

th Qz ridges; Lithic Arenite; Bi < 15%
Arenite
g Lithic Arenite
ite, Bi > 15%

Differentiated

Paragneiss)
ex; Stratified Gneiss with > 20% felsic segregations and pegmatite

erentiated

SYMBOLS

80%

55°10"

Don L.

9a/

55°10"

Cree L.

10 of

Found L.

Base map compiled from N.A.P.L. airphotos
 Geology by P. A. Goetz 1975-1976

44

OF / DE

ogical Contact

ing, with Foliation; inclined, vertical

ion; inclined, vertical

tion; plunging, horizontal

ay

line

and Stream

ase map compiled from N.A.P.L. airphotos
Geology by P.A. Goetz 1975-1976

12 of 1

13 a/f

Kississing L.

Camp L.

Cold
Lake

Portage L.

Ken L.

C.N.R.

17 of

18 of

100° 55"

20 of 29

FIG. III-4

4 of

FIG. III-4
GEOLOGY A

5 of

Level P
Base ma
Surface

AND LEVEL PLANS, SHERRITT GORDON MINE, SHERRIDON MANITOBA

LEGEND

5. Bi Gneiss $\pm$ Ga, Lithic Arenite
5c Mafic Lithic Arenite

6 Quartzo-feldspathic Gneiss (Arenite)

9 Bi-Ga-Si Gneiss (Pelite), Mudstone

10 Ga-An Gneiss $\pm$ Co, Mg Rich Sediment

11 Hb Gneiss $\pm$ Ga, Calcareous Arenite

11a Di-bearing Amphibolite, Calcareous Arenite

12 Marble, Calc-silicate

13 Amphibolite, Basaltic

14 Amphibolite, Basaltic

16 Massive Sulphides

19 Pegmatite

DBA

illicate

asaltic Pyroclastic

asaltic Flow

es

SYMBOLS

- Outcrop
- Geological Contact
- Building, Shaft
- Ore Specimen, in situ, float
- Lake and Stream
- Road

SYMBOLS

Outcrop

Geological Contact

Building, Shaft

Specimen, in situ, float

Lake and Stream

road

807

bearing Amphibolite, Calcareous Arenite,

modified from compilations of Sherritt-Gordon Mining Co., Projected vertically to surface
compiled from N. A. P. L. airphotos
by P. A. Goetz 1976

10 of 1

5th

11th

Central Shaft

Camp Lake

11 of 1

shaft

1st

S

LEVEL

VERTICAL DEPTH (Ft)

1	148
2	265
3	382
4	494
5	616
6	733
7	888
8	1043
9	1193
10	1343

12 of

200 100 0 200 400 600 ft

13 of

14 of

U

FIG. III-6
GEOLOGY AND ORE ZONING, 3RD LEVEL, SHERRITT

LEGEND

☐ 5 Bi Gneiss ± Ga, Lithic Arenite	☐ 14 A
☐ 5b Si-bearing Lithic Arenite	☐ 16 A
☐ 5c Mafic Lithic Arenite	☐ 19 F
☐ 11 Hb Gneiss, Calcareous Arenite	☐ 13 Amphibolite, Basaltic Pyroclastic

% Sulphides ☐ 0-20

% Cu ☐ < 2

RD LEVEL, SHERRITT-GORDON MINE, SHERRIDON MAN.

SYMBOLS

- 14 Amphibolite, Basaltic Flow
- 16 Massive Sulphides
- Disseminated Sulphides
- 19 Pegmatite

- Geological contact
- Drill hole
- Drift

Data compiled from records of the Sherritt-Gordon Mining Co.

% Sulphides 0-20 20-40 40-60 60-80 80-100

% Cu <2 2-3 3-4 4-5 >5

FIG. 11-6
GEOLOGY AND ORE ZONING, 3RD LEVEL, SHERRITT-GO

LEGEND

☐ 5 Bi Gneiss $\pm$ Ga, Lithic Arenite	☐ 14 Amphibolite
5b. Sl-bearing Lithic Arenite	☐ 16 Massive
5c. Mafic Lithic Arenite	☐ Dissem.
☐ 11 Hb Gneiss, Calcareous Arenite	☐ 19 Pegmatite
☐ 13 Amphibolite, Basaltic Pyroclastic	

% Sulphides ☐ 0-20

% Cu ☐ < 2

% Zn ☐ < 1

Cu / Zn Ratio ☐ < 1

Zoning compiled from assay
plans of drill holes.

34

0 LEVEL, SHERRITT-GORDON MINE, SHERRIDON MAN.

SYMBOLS

- 14 Amphibolite, Basaltic Flow
- 16 Massive Sulphides
- Disseminated Sulphides
- 19 Pegmatite
- Geological contact
- Drill hole
- Drift

Data compiled from records of the Sherritt-Gordon Mining Co.

2af

55 09"
100 59"

191

FIG. A-1
Chemical Analysis and Thin Section Locations
Scale: 1" = 1320'

4 of 1

186

194

137 138
26 27 139

24

23

33 28
31 29
32 30

17 49 50 51 52 11 9 7
16 15 14 13 12 10 8 6 5 4

19

20

21

131

130

136

141

140

142

145

147

148

149

150

152

153

154

155

156

157

158

159

160

161

162

163

164

165

166

167

168

169

170

171

172

Lost L.

Camp L.

SHERRIDON

89 90

86

87

88

85

84

83

82

81

80

79

78

77

76

75

74

s and Thin Section Locations

= 1320'

52

1

2 of 1

FIG. IV-3
Stereoplot of Structural Domains on West Fl

Domain 1

200 Poles to Foliation
* C.I. - 2.5
** M.V. - 15:0

Domain

Domains on West Flank of Sherridon Basin

(Schmidt Projection)

34

Domain 1

*Contour Interval %

**Maximum Grid Value %

6 Lineations

C.I. - 16.67

M.V. - 50.00

Domain 2

*C.I. - 2.5

**M.V. - 15.0

Domain 2

369 Poles to Foliation

C.I. - 1.9

M.V. - 10.30

Domain 3

197 Poles to Foliation

C.I. - 3.55

M.V. - 19.29

All Domains

766 Poles to Foliation

C.I. - 0.91

M.V. - 4.96

Domain 3

59

C.I. - 16.67
M.V. - 50.00

Domain 2

24 Lineations
C.I. - 4.17
M.V. - 16.67

Domain 3

32 Lineations
C.I. - 6.25
M.V. - 21.88

All Domains

62 Lineations
C.I. - 3.23
M.V. - 14.52

6 of 6

END

1 7 1 2 8 1

FIN